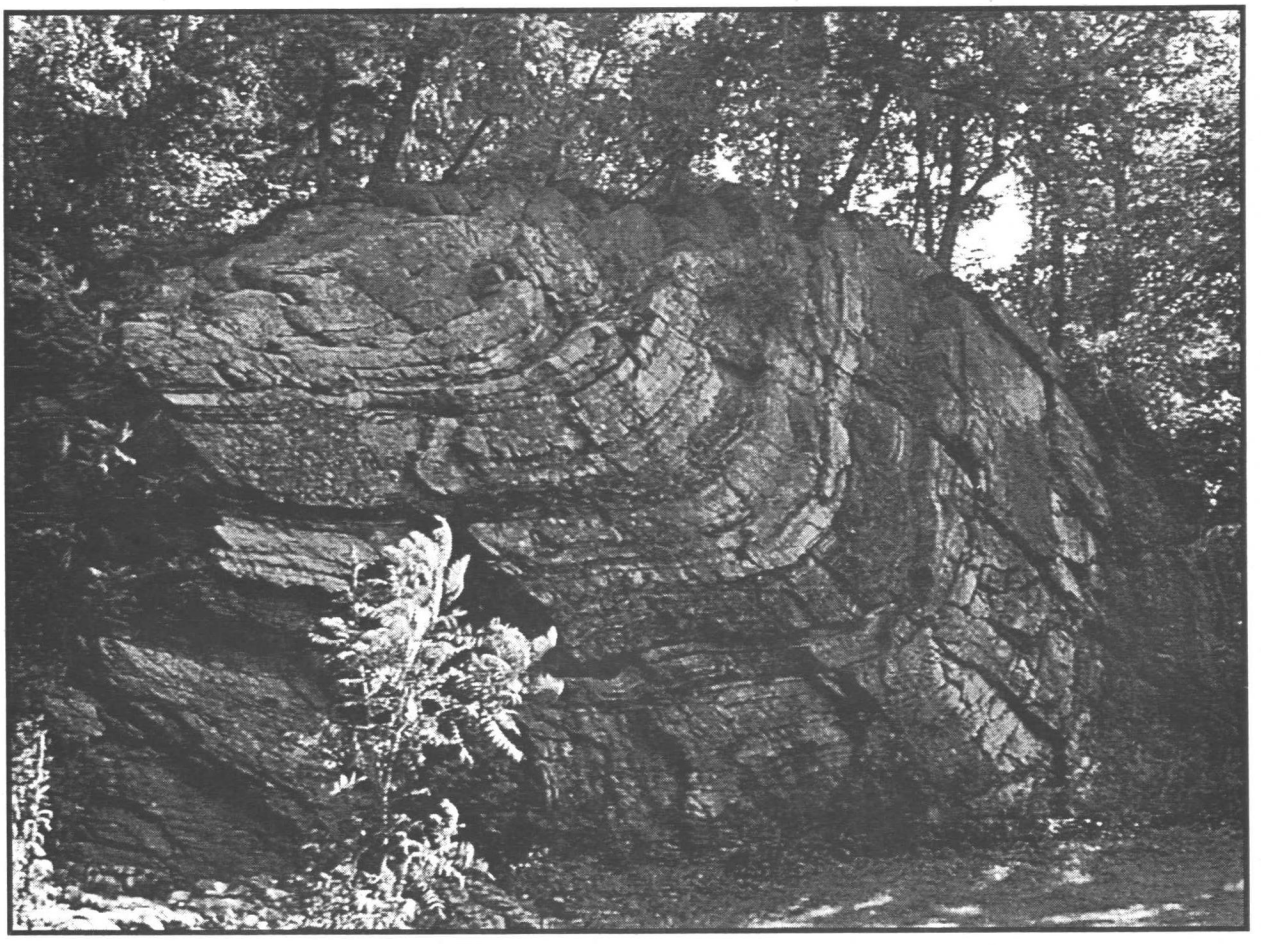
1997
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Guidebook to Field Trips in Vermont and Adjacent New Hampshire and New York

Organizers and Editors:
Timothy W. Grover, Helen N. Mango, Edward J. Hasenohr



View of the Glen Lake Syncline

Meeting hosted by Castleton State College

89TH ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE
GEOLOGICAL CONFERENCE

Guidebook to Field Trips in Vermont and Adjacent New Hampshire and New York

September 19, 20, and 21, 1997
Killington - Pico Region,
Vermont

Organizers and Editors:
Timothy W. Grover,
Helen N. Mango,
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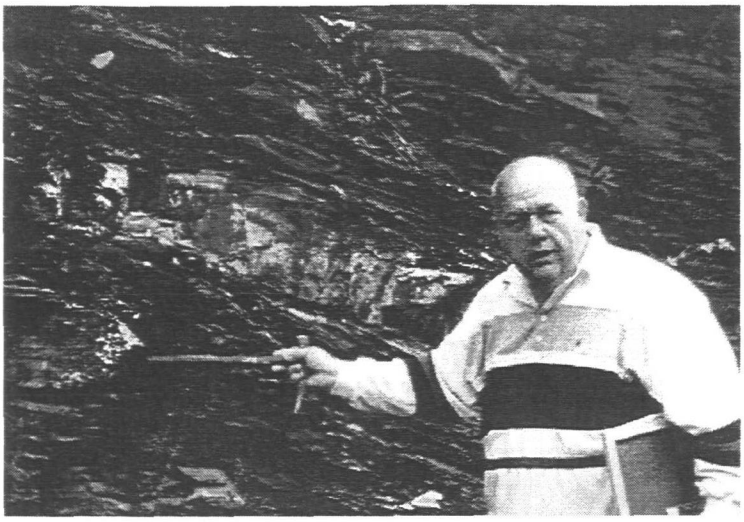
Hosted by
Castleton State College

A DEDICATION



Andrew V. Raiford
1937-1993

Andrew Raiford taught geology at Castleton from 1970 to 1993. During this time he built the geology program from a few course offerings to a full-fledged major, teaching everything from the introductory courses to structure, igneous-metamorphic-sedimentary geology, hydrogeology and tectonics, even including a seminar course every semester for upperclass majors. He was more than a one-man department; he was the friend and mentor for twenty years of geology majors. For many, his office was their second home.

Andy was born and raised in New Orleans, received his Bachelor's degree in geology from Louisiana State University, and his Master's degree in geophysics from Tulane University. He taught earth science in the Jefferson Parish School District, and then worked for Boeing Aerospace, designing the silk-screen process that put the U.S. flag on the Saturn booster rocket. Vermont seemed an unlikely place for someone who had always lived in the deep South, but after moving up here Andy didn't want to live anywhere else. He never lost his Louisiana roots, however. Every semester he would host a geology party at his home, and the students would pore over oil company seismic profiles while eating his famous red beans and rice. When Castleton's newly-designated sexual harassment officer came to the department to talk about dealing with a variety of different students, including how *not* to address them, Andy said "fine, instead of using "girls" and "guys," I'll just say "y'all" like I'm used to saying!".

Andy always believed that getting out in the field and looking at rocks was the only way to understand geology. As long as he was the sole geology professor, he required his students to go to *two* field camps. And he always took his students on field trips for his courses, and for any field conferences available; a vanload of Castleton students with Andy at the wheel attended virtually every Vermont Geological Society and NEIGC meeting. There are stories from geology graduates of his December field trips to the Catskills, complete with brooms to brush snow off the outcrop. He reprimanded the new geology hire in 1991 for suggesting that she should plan for rain dates. His enthusiasm for his students and for his science affected a generation of students, who remember him as a great teacher and a kind and decent human being.

Andy Raiford died of cancer on December 4, 1993.

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C2	Stratigraphy and Structure of the Orwell 15' Quadrangle, West-Central Vermont *Helen Mango*	C2-1:13
C3	The Potsdam Sandstone of the Southern Champlain Valley: Sedimentology, Stratigraphy and Diagenesis *Bruce Selleck*	C3-1:16
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Welcome

Welcome to the 89th annual meeting of the New England Intercollegiate Geological Conference hosted by Castleton State College. It has been 10 years since the NEIGC was last held in Vermont and 38 years since it was held in the Rutland-Killington-Pico region, hosted then by E-an Zen and others. We felt it was time for NEIGC to return to the area.

The bedrock geology of New England is a complex collage resulting largely from plate boundary processes, including multiple, overprinting phases of orogenesis and rifting, spanning the last billion years or so of Earth history. Geological research in New England has a long and storied history (as is evidenced by the history of the NEIGC) and has provided a solid foundation for our understanding of New England geology. However the picture is far from complete (nor universally agreed upon!). Recent research, presented here, continues to add to our knowledge of this geologically complicated region. In addition to the bedrock geology presented at this conference, there are also several trips pertaining to the surficial geology and hydrology of the area, topics of recent emphasis in the geological sciences.

Acknowledgements

We would like to thank everyone who assisted us in hosting the NEIGC. The success of this meeting is due to the many people who volunteered their services above and beyond the call of duty. We would especially like to thank:

- All the field trip leaders for their efforts in preparing their field guides and their willingness and enthusiasm to share the results of their research.

- Mark Van Baalen for sharing his knowledge and expertise gained from hosting last year's NEIGC

- Rose Harvey, Karen Puente, and Trudy Perry for their assistance in preparing all the NEIGC mailings and the guidebook.

- Lisa Myette and Angela Boisvert from Killington, and Murray McGrath from the Inn at Long Trail, for their assistance in arranging and preparing the banquet and welcoming reception.

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MEETINGS OF THE NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

1901 Westfield River Terrace, MA (Davis)
1902 Mount Tom, MA (Emerson)
1902 West Peak, Meriden, CT (Rice)
1904 Worcester, MA (Emerson)
1905 Boston-Nantasket, MA (Johnson and Crosby)
1906 Meriden-East Berlin, CT (Gregory)
1907 Providence, RI (Brown)
1908 Long Island, NY (Barrel)
1909 Northern Berkshires, MA (Crosby and Warren)
1910 Hanover, NH (Goldthwait)
1911 Nahant-Medford, MA (Lane and Johnson)
1912 Higby-Lamentation Blocks, CT (Rice)
1915 Waterbury-Winsted, CT (Barrell)
1916 Blue Hills, MA (Crosby and Warren)
1917 Gay Head, Martha's Vineyard, MA (Woodworth and Wigglesworth)
1920 Hanging Hills, Meriden, CT (Rice and Foye)
1921 Attlebora, MA (Woodworth)
1922 Amherst, MA (Antes)
1923 Beverly, MA (Lane)
1924 Providence, RI (Brown)
1925 Waterville , ME (Perkins)
1926 New Haven, CT (Longwell)
1927 Worcester, MA (Perry, Little and Gordon)
1928 Cambridge, MA (Billings, Bryan and Mather)
1929 Littleton, NH (Crosby)
1930 Amherst, MA (Loomis, Grodon)
1931 Montreal, PQ (O'Neil et al.)
1932 Providence-Newport, RI (Brown)
1933 Williamstown, MA (Cleland, Perry and Knopf)
1934 Lewiston, ME (Fisher and Perkins)
1935 Boston, MA (Morris,Pearsall and Whitehead)
1936 Littleton, NH (Billings et al)
1937 NYC-Dutchess Co., NY (O'Connell et al.)
1938 Rutland, VT (Bain)
1939 Hartford, CT (Troxell et al.)
1940 Hanover, NH (Goldthwait et al.)
1941 Northampton, MA (Balk et al.)
1946 Mt. Washington, NH (Billings)
1947 Providence, RI (Quinn)
1948 Burlington, VT (Doll)
1949 Boston, MA (Nichola et al.)
1950 Bangor, ME (Trefethen and Raisz)
1951 Worcesterm MA (Lougee, Little)
1952 Williamstown, MA (Perry et al.)
1953 Hartford, CT (Flint, et al.)
1954 Hanover, NH (Elston et al.)
1955 Ticonderoga, NY (Rodgers el al.)
1956 Portsmouth, NH (Novotny, et al.)
1957 Amherst, MA (Bain)
1958 Middletown, CT (Rosenfeld et al.)

1959 Rutland, VT (Zen)
1960 Rumford, ME (Griscom and Milton)
1961 Montpelier, VT (Doll)
1962 Montreal, PQ (Clark et al.)
1963 Providence, RI (Quinn)
1964 Chestnut Hill, MA (Skehan)
1965 Brunswick, ME (Hussey)
1966 Katahdin, ME (Caldwell)
1967 Amherst, MA (Robinson et al.)
1968 New Haven, Ct (Orville)
1969 Albany, NY (Bird)
1970 Rangeley Lakes-Dead River, ME (Boone)
1971 Concord, NH (Lyons and Stewart)
1972 Burlington, VT (Doolan and Stanley)
1973 Fredericton, NB (Rast et al.)
1974 Orono, ME (Osberg)
1975 Great Barrington, MA (Ratcliffe)
1976 Boston, MA (Cameron)
1977 Quebec City, PQ (Beland and LaSalle)
1978 Calais, ME (Ludman)
1979 Troy, NY (Friedman)
1980 Presque Isle, ME (Roy and Naylor)
1981 Kingston, RI (Boothroyd and Hermes)
1982 Storrs, CT (Joesten and Quarrier)
1983 Greenville-Millinocket, ME (Caldwell and 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 Amherstm MA (Robinson and Brady)
1993 Boston, MA (GSA: Cheney and Hepburn)
1994 Millinocket, ME (Hanson)
1995 Brunswick, ME (Hussey and Johnson)
1996 Mr. Washington, NH (Van Baalen)

A1
Deformation of the Taconic Sequence, western Vermont and eastern New York

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INTRODUCTION

The Taconics are a series of thrust sheets which lie along the western edge of the northern Appalachians in western Vermont, Massachusetts and Connecticut and eastern New York (figure 1). The thrust sheets contain metamorphosed Cambro-Ordovician argillaceous rocks with minor arenites and limestones. Along the western edge of the Taconics in the lowest thrust sheet, the metamorphic grade is low and the rocks contain a prominent slaty cleavage. To the east and in higher thrust sheets the slaty cleavage is overprinted by crenulation cleavages, the metamorphic grade is higher and phyllites and schists are common. The Taconic thrust sheets sit on autochthonous to para-autochthonous Cambro-Ordovician carbonates or middle Ordovician flysch. The geology of the Taconics has been used as a model of relating plate tectonics to the evolution of mountain belts (Bird and Dewey, 1970; Rowley and Kidd, 1981). Despite some differences, most workers agree on the basic plate tectonic setting of the Taconic orogeny (e.g. Rowley and Kidd, 1981, Stanley and Ratcliffe , 1988). Rocks contained within the Taconic klippe were deposited on the slope and rise of the early Paleozoic passive margin of North America from Cambrian through middle Ordovician time. In the middle Ordovician, the North America passive margin reached the west-facing Taconic subduction zone and collision produced the deformation now seen within the thrust sheets as well as creating the thrust sheets themselves. This trip will examine the details of Taconic deformation within the notrthern part of the thrust sheets (figure 1) and, thus, can be viewed as an investigation of deformation associated with arc-continent collision.

Taconic Stratigraphy. The stratigraphy of the Taconics was originally established by Zen (1961) and was later redefined by Rowley et al. (1979). As subdivided by Rowley et al. (1979), the general stratigraphy within the northern Taconics consists of six formations. The lowest formation, the Bomoseen, is largely clastic with overlying formations generally composed of slates and silty-slates. Limestones and sandstones are locally present, but volumetrically insignificant. The slatey formations are distinguished principally by color. A regional angular unconformity occurs at the top of the Taconic stratigraphy with a shale-greywacke

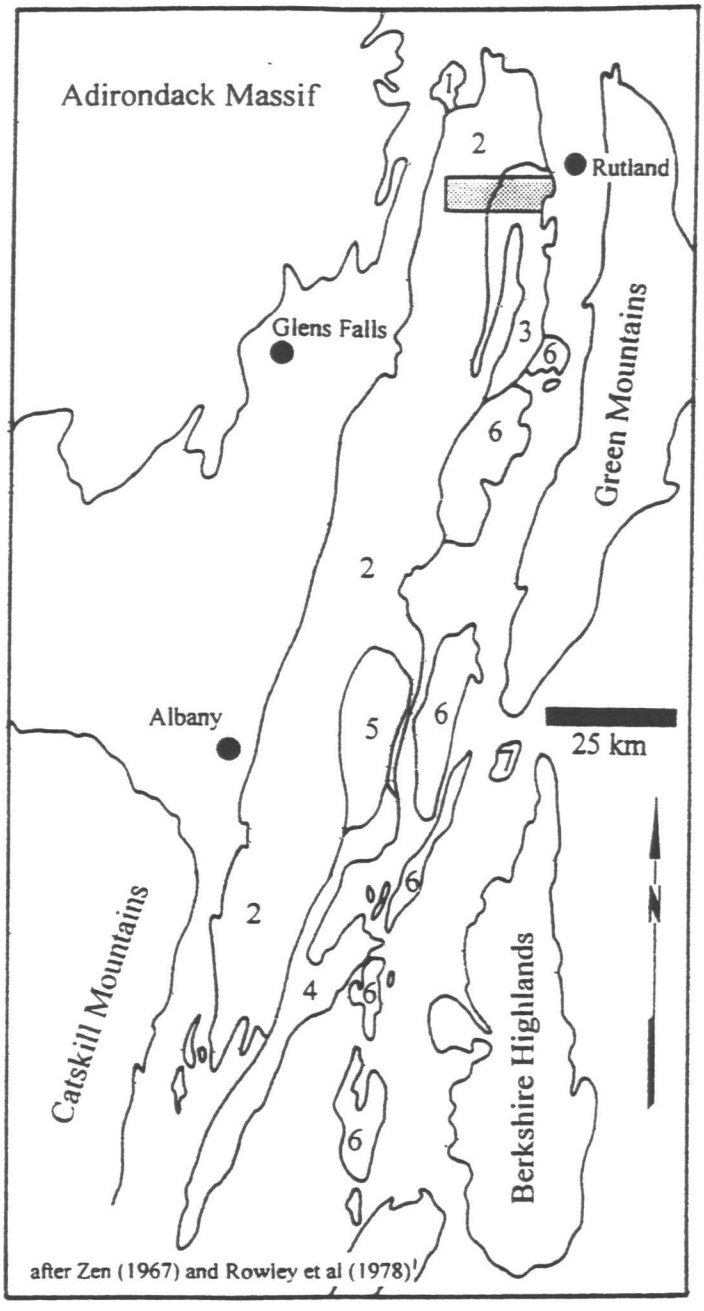

Figure 1. Map showing locations of the major geological features of eastern New York and western Vermont and Massachusetts. Taconic thrust sheets are indicated by numbers; 1-Sunset Lake slice; 2-Giddings Brook slice; 3-Bird Mountain slice; 4-Chathan slice; 5-Rensselaer Plateau slice; 6-Dorsett Mountain slice; 7- Greylock slice.

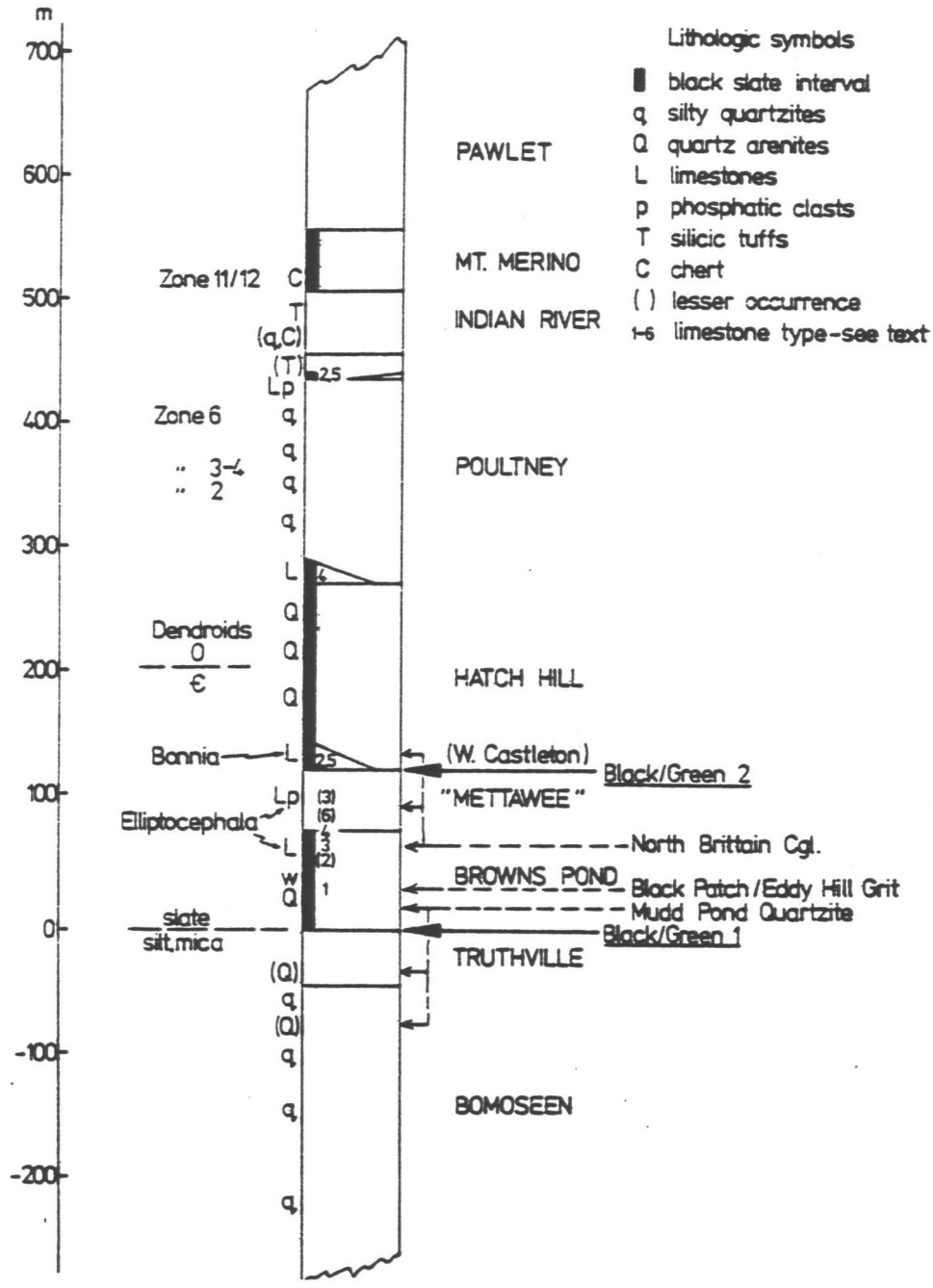

Figure 2. Stratigraphy of the Giddings Brook slice, from Rowley et al, 1979

unit (Pawlet fm) overlying it. The formations are laterally continuous and have been used successfully for mapping over large distances.

General Structural Geology. The most prominent structures of the Taconic klippe and folds and thrust faults (figure 3). The most prominent thrust faults bound the different klippe and minor thrusts within klippe are locally present. In the region of this field trip, and for most of length of the Taconics, the lowest thrust slice is the Giddings Brook slice (figure 1) bounded on bottom by the Giddings Brook fault and above by the Bird Mountain fault. In the northern Taconics, the next highest slice is the Bird Mountain slice. Both the Bird Mountain and Giddings Brook faults truncate isoclinal folds which dominate the structures within thrust sheets (figure 3). The largest folds have wavelengths of 1 - 1.5 km and amplitudes of 1.5 - 3 km. Fold axes trend roughly N-S and plunge shallowly to the north and south (figure 4). Axial planes strike roughly N-S and dip eastward at angles from 20-60°. The hinges are very tight and are only rarely exposed. The folds have a prominent axial plane slaty cleavage which makes the rocks commercially usable. Because the folds are isoclinal and have an axial plane cleavage, bedding and cleavage are parallel on limbs and transect each other at or close to fold hinges.

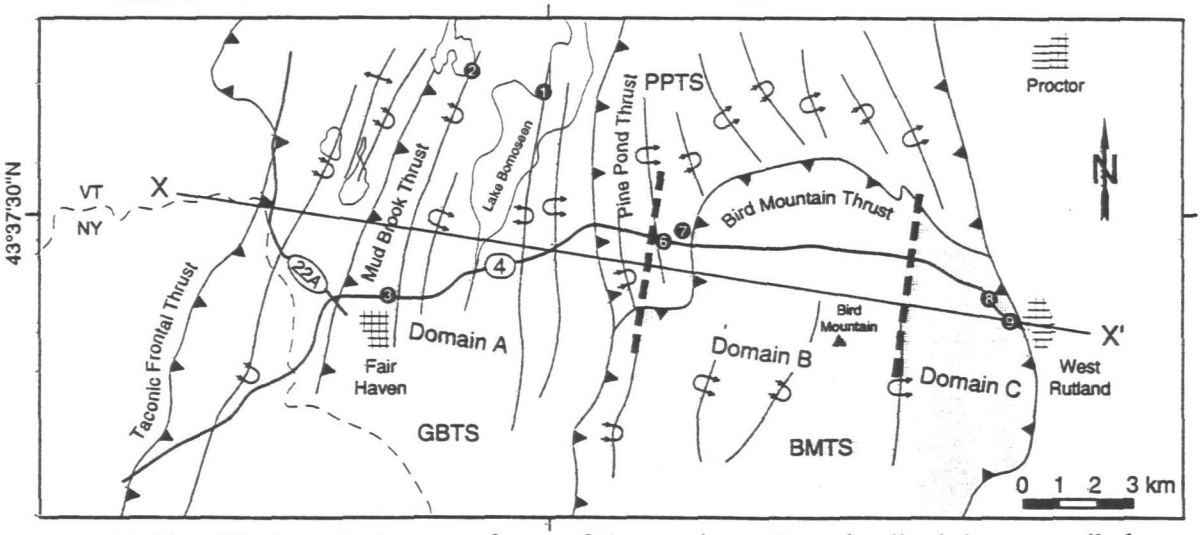

Figure 3. Simplified geologic map of part of the northern Taconic allochthon compiled from Zen (1961) and Fisher et al. (1970). Some stop locations are shown by numbered black circles.

Timing of Deformations. A reasonably precise relative timing of various structural elements can be constructed from basic field relations and cross-cutting relationships. The earliest structures recognized are early thrust faults (folded by isoclinal folds) and an early cleavage (S_1). Early thrusts are rare and evidence for the S_1 cleavage is seen almost exclusively in thin section. The origin of S_1 is a matter of speculation. We are confident, however, that its origins are tectonic rather than compactional as it commonly is not parallel to bedding. Folds are the next youngest structures. Locally, early thrust faults

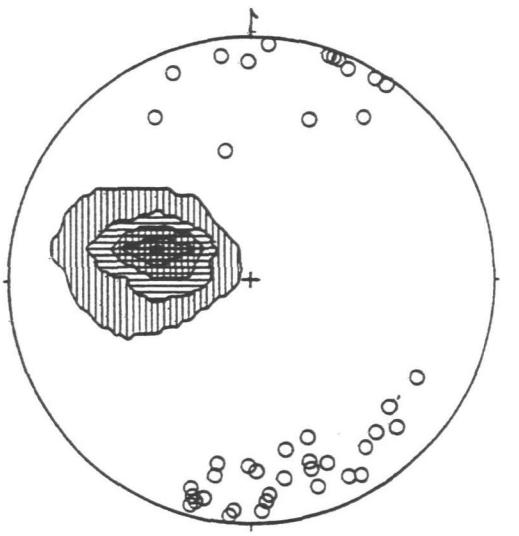

Figure 4. Lower hemisphere, equal-area projection of 600 poles to slaty cleavage (S_2, contours) and 42 measured fold axes and bedding-cleavage intersections (open circles). Contours are 1, 5, 10, 15 and 20% per 1% area

are folded, revealing the timing between the early faults and the folds, but any possible relationship between fold formation and the S_1 cleavage has yet to be determined. Thrust faulting continued after folding as is evidenced by both the Bird Mountain and Giddings Brook faults truncation of folds. The Bird Mountain fault, however, is older than the Giddings Brook fault based on relationships between the faults and the main slaty cleavage (S_2). The Bird Mountain fault is cut by S_2 and, thus, must pre-date it. The imposition of this cleavage on previously formed folds is most likely responsible for tightening the folds to their isoclinal geometry. The finite strain associated with cleavage formation is discussed below. Unlike the Bird Mountain fault, the Giddings Brook fault cuts the S_2 cleavage and last motion along it is the youngest event evidenced in the northern Taconics. Locally within the Giddings Brook slice and pervasively in its eastern portion and within the Bird Mountain slice late crenulation cleavages are present. Two principal sets occur with the youngest being present only in the easternmost part of the northern Taconics. The interpretation of these cleavage is discussed below.

For the most part, veins are uncommon in the northern Taconics but are present within the northern Taconics and will be visited by this trip. Three different vein families are present, although the timing of vein formation and its significance is a matter of continuing study. Some veins are clearly early as they are found folded and boudinaged. Other veins form in en-echelon sigmoidal arrays and may represent more than one period of veining. Some of these veins may be syn-folding whereas others are clearly post-folding. Some of these veins arrays are indicative of dip-slip ductile shearing whereas others indicate strike-slip deformation. Finally, one can find meter-scale bull quartz veins which cut all other structures and fabrics and are the latest veins to have formed in the northern Taconics.

FINITE AND INCREMENTAL STRAIN
(Goldstein and Pickens)

Considerable effort has been expended in analysis of both finite and incremental strain, especially in the Giddings Brook slice, which contains many strain indicators. Finite strain has been investigated using reduction spots (Goldstein et al, 1995a; Hoak, 1992 and Wood, 1974) and recently by measuring deformed graptolites (Goldstein et al, in press). This trip will visit localities used for both studies. In addition, incremental strains have been investigated using pyrite framboid pressure shadows (Cushing and Goldstein, 1990; Chan and Crespi, 1995) and pyrite porphyroblast pressure shadows (Pickens, 1993). Whereas the finite strain results are quite consistent, there are considerable differences in both the form and interpretation of pressure shadows observed by different workers, suggesting that none of us has fully determined what the complete incremental strain history is for these rocks. The implications of incremental strains are most significant with respect to the origin of the main slaty cleavage (S_2) and one view is expressed below. This topic is one which is still being debated by the authors.

Finite strain results. Pale green spots in the Mettawee fm. allow the measurement of finite strain at a number of localities (Goldstein et al, 1995a; Hoak, 1992; Wood, 1974). Spots occur in two general forms; nearly perfect ellipsoids and irregular blotches. The ellipsoids lie with their XY planes parallel to the plane of the main slaty cleavage (S_2) and must have begun as nearly perfect spheres or ellipsoids. Goldstein et al. (1995a) measured reduction spots are 7 localities in the Giddings Brook slice and another three sites were measured by Hoak (1992) and Wood (1974). All results agree well, lying in the field of apparent flattening on a flinn diagram and are arrayed along a horizontal trend (Fig 5). This pattern indicates that the difference in strain state between sites with lower strain and sites with higher strain lies in the ratio between Y and Z whereas the ratio between X and Y remains largely unchanged. Because reduction spots yield only strain ratios rather than absolute strains, there is an inherent ambiguity in their interpretation. This is especially true when it comes to determining volume changes which may have accompanied slaty cleavage development. One may interpret the variation in strain as reflective of constant volume deformation or as indicating considerable loss of volume. We do not favor the constant volume interpretation as it requires considerable elongation in Y to maintain a nearly constant XY ratio. None of the incremental strain work has yielded any indication of elongation in Y. Thus, our favored interpretation is that the shortening in Z was accommodated by loss of material through dissolution. Our calculations (Goldstein et al, 1995a) suggest that an average of 55% of the original solid volume of rock was lost at the most highly strained site. This number is a minimum because of assumptions which had to be made before the calculation could be done.

Reduction spots are inherently ambiguous, as noted above, because they yield strain rations rather than absolute strains. Absolute strains can be measured, however, through study of deformed graptolites (Goldstein et al, in press). Graptolites were

colonial planktonic organisms which were abundant during lower Paleozoic time. For some species, certain dimensions were constant so that, in the deformed state, one can determine strain by measuring those dimensions and using the formula:

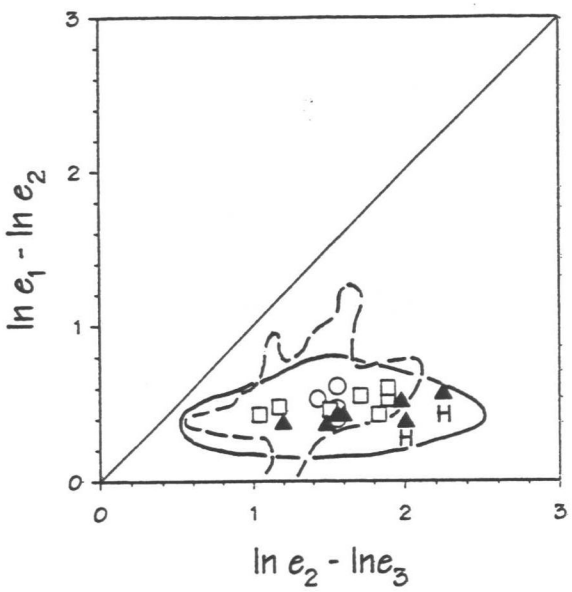

Figure 5. Summary of reduction spot strain measurements from the northern Giddings Brook slice. Black triangles are site averages measured by Goldstein et al. (1995a); open squares are from Wood (1974) and open circles are from Hoak (1992). The solid line shows the range of data measured by Hoak (1992) and the dashed line shows the range of values from slates in the Taconics and Wales measured by Wood (1974).

$$e=l_f-l_o/l_o$$

in which l_f represents the final length of a line (dimension in deformed state) and l_o represents the original length of a line (dimension in the undeformed state). For graptolites, the spacing between thecae is used (figure 6). By measuring enough graptolites which lie at different angles to principal strains one can define the two dimensional strain for the bedding plane, the plane which contains the graptolites. By measuring graptolites on the limbs of folds, in which bedding and cleavage are parallel, one derives the X and Y components of strain and measuring at hinges of folds, in which bedding and cleavage are perpendicular, one derives the Z component of strain. We have done this for five sites at limbs of folds and two at hinges, with another two sites lying between limb and hinge. e_1 (X) ranges from 110% to 30% with an average of 55%, e_2 (Y) ranges from 0% to -40% with an average of -22% and e_3 (Z) ranges from -60% to -80% with an average of -70%. Thus, we find that the absolute finite strain in these slates is constrictional (one elongation and two shortenings). The principal strains can be used to determine volume changes from the following relationship:

$$\ln (1+\Delta) = \ln (1+e_1) + \ln(1+e_2) + \ln(1+e_3)$$

in which Δ is the volume change (positive for volume gains and negative for losses) and e_i is a principal strain. Volume changes based on graptolite strains and using this

relationship lie between 37% and 76% losses, similar to the volume loss determinations based on reduction spots. Thus, an even more precise definition of the finite strain would be constrictional plus a volume loss.

Strains from reduction spots and graptolites agree well, once one takes into account the effects of sedimentary compaction. Graptolites lie in bedding planes and do record the effects of compaction. However, if reduction spots formed early in the depositional history of the Mettawee fm, they would record compaction as well as tectonic strains. We can evaluate these variation on a Flinn diagram (figure 7) which shows that if we add a reasonable amount of compaction, the finite strain recorded by graptolites is essentially coincident with reduction spots and both are strongly indicative of volume losses. of compaction on graptolites (arrow). Note that if a compaction is added on to the graptolite strains they are essentially coincident with reduction spot strains.

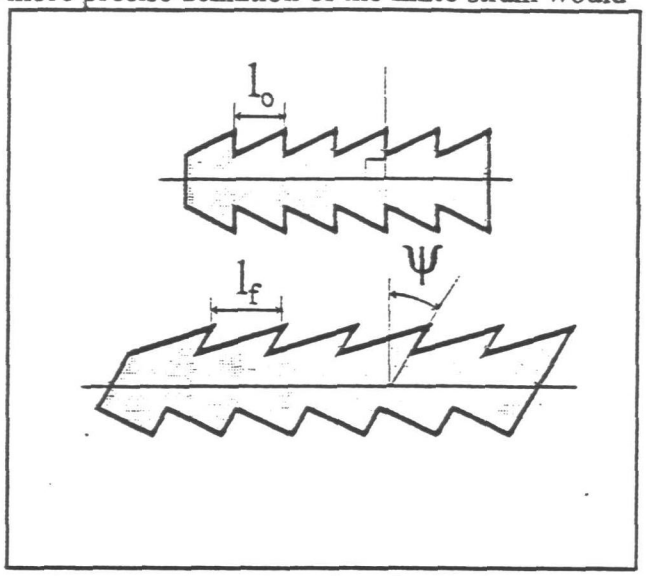

Figure 6. Deformation of a schematic Orthograptus Whitfieldii showing changes in thecal spacing and angle between the thecal aperture and the long axis of the stipe.

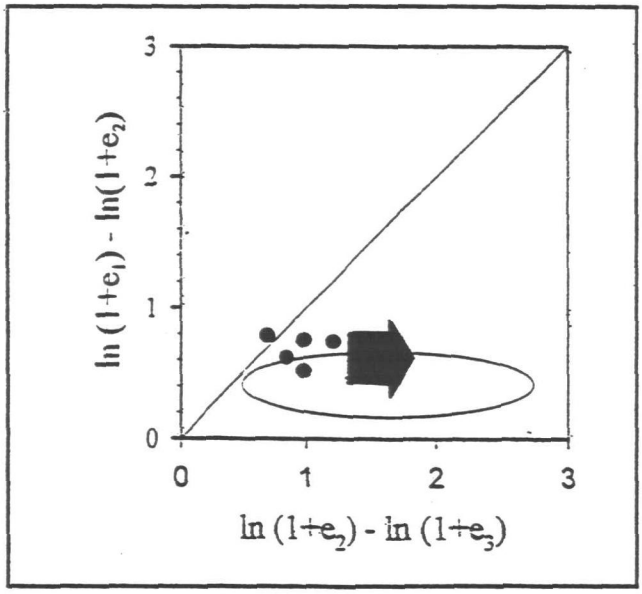

Figure 7. Logarithmic Flinn diagram illustrating the state of strain from graptolites (dots), the range of reduction spots (ellipse) and the effect of adding a compaction on graptolite strain

Other Evidence for Volume Loss. The volume loss determinations for the Taconics agree well with other determinations of volume change from slates (e.g. Wright and Platt, 1982; Beutner and Charles, 1985; Henderson et al, 1986 and Wright and Henderson, 1992; Jenkins, 1988) which typically report between 40% and 60% loss. There is ample evidence that volume loss has affected Taconic slates beyond the finite strain results. Pressure solution was a dominant deformation mechanism during slaty cleavage formation. However, very small volumes of material were redeposited as veins and pressure shadows. Further, whole rock

geochemical studies and micro-scale geochemistry also indicate large volume losses (Goldstein et al, 1995b).

Incremental Strain Results. Whereas the finite strain recorded by graptolites and reduction spots reflects the sum total of all strain events which have affected the Taconics, studies of pyrite framboid and porphyroblast pressure shadows indicate the incremental history of strain. Pickens (1993) studied the porphyroblasts pressure shadows and Cushing and Goldstein (1993) and Chan and Crespi (1995) studies framboid pressure shadows. Unfortunately, all three studies resulted in somewhat different results suggesting that we have yet to fully understand the strain history of the Taconics. Incremental strain analyses of fiber growth around cubic pyrite porphyroblasts (Pickens, 1993) indicate a complex non-coaxial strain history for the slates of the Giddings Brook slice. Pyrite porphyroblasts with associated fibrous strain shadows are locally present throughout the klippe. In some samples, fiber growth parallels changes in cleavage development, providing insight into the transformation of one cleavage into the next. Strain shadow morphology and type vary considerable, but fiber growth can generally be divided into three phases (figure 8). It is important to emphasize, however, that these growth phases should be thought of as continuum of fiber development and not as discontinuous segments. In short, the incremental data demonstrate that the direction of principal extension began parallel to the S_1 foliation (Phase I), rotated (counterclockwise) towards this cleavage (Phase II), and eventually came to lie in the X^Y plane of the S_2 cleavage (Phase III). With the exception of Phase III growth, all increments were plane strain, with no growth in the Y-axis direction. Only the last increment of growth forms a crown, parallel to S_2. These data and observations demonstrate the continuous nature of fabric development during the Taconian deformation.

At one locality with unusually large pyrite framboids, Cushing and Goldstein (1990) described pressure shadows which were composed of segments of long straight fibers, indicating co-axial deformation, separated by short increments of highly curved fibers, indicating a non-coaxial transition between co-axial strain increments. They found four dominant co-axial increments of fiber growth with only minor non-coaxial strains. They concluded that all strain increments were superposed on a pre-existing small fold as none of the fiber segments were rotated between opposing limbs of the fold.

Chan and Crespi (1995) found framboid pressure shadows entirely different from those studies by Cushing and Goldstein (1990) and Pickens (1993). They describe pressure shadows with smoothly curving fibers suggesting a non-coaxial strain history for the Taconics. Below, they argue that these observations can be used to construct a model of S_2 formation as a result of non-coaxial strain. Considering the lack of agreement between the three studies, we suggest that more work needs to be addressed at the strain history of the Taconics.

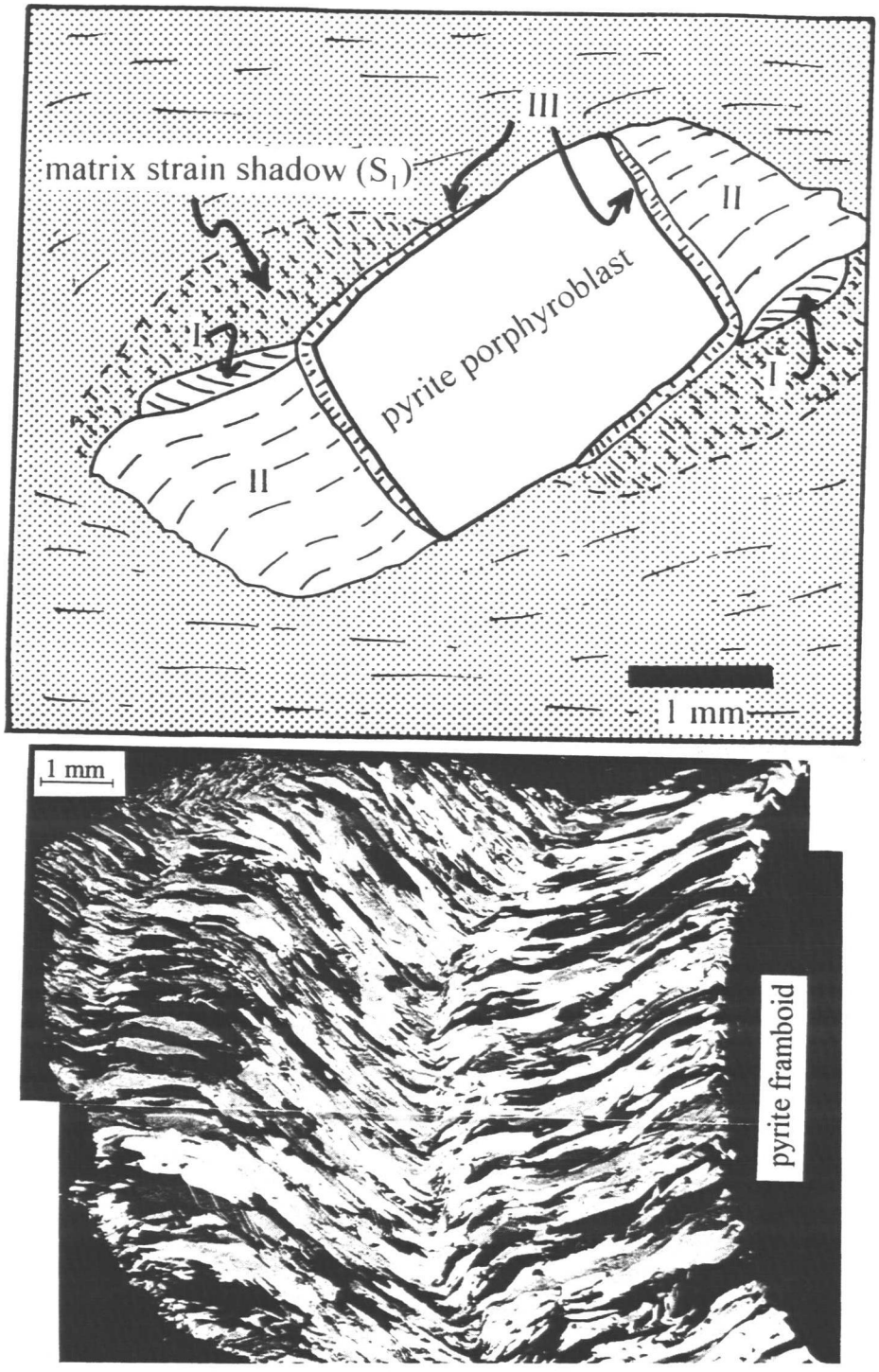

Figure 8. Examples of fibrous pressure shadows developed around pyrite porphyroblasts and framboids. A. Sketch of pyrite porphyroblast showing three stages of fiber growth plus matrix strain shadows which preserve S_1. B. Photograph, in cross polarized light, of large pyrite framboid and pressure shadow showing different stages of largely co-axial fiber growth.

CLEAVAGE DEVELOPMENT
(Chan and Crespi)

At most locations the slaty cleavage observed in the Taconics appears to be an extremely well defined penetrative foliation. In thin section, rather than the single dominant cleavage one might expect, a composite of fabrics can be observed. In the Giddings Brook and Bird Mountain slices, at least two cleavages (S_1 and S_2) can be observed in thin section, with a third (S_3) and fourth (S_4) locally overprinting the earlier fabrics. As discussed below, in the eastern portions of the northern Taconics, S_3 becomes the dominant cleavage and S_2 is barely recognizable.

As one might expect, the slates exhibit a very well developed p- (phyllosilicate), q- (quartz) domainal structure. These domains exist at a variety of scales ranging from single detrital grains to layered concentrations of p and q minerals visible in hand specimen. Where present, the S_1 cleavage is typically at a high angle to S_2, wrapping into the later cleavage along S_2 folia. In examples where this transition in fabrics is preserved, a progressive deformation of S_1, involving grain rotation, strain solution and eventually mineral growth, can be observed.

Development of multiple cleavages Multiple cleavages in rocks are commonly observed in collisional orogenic belts. Studies of their geometry, spatial distribution, and kinematics can contribute significantly to our understanding of the structural processes that occur in orogenic belts and the response of rocks to deformation at depth. The purpose of this section is to provide descriptions of the multiple cleavages observed in the northern Taconic Allochthon and to discuss the kinematics associated with individual cleavages.

Previous work has resulted in the recognition of one locally developed cleavage (S_1) and two regionally developed cleavages (the S_2 slaty cleavage and the S_3 crenulation cleavage of this text) within the northern Taconic Allochthon. The two regionally developed cleavages have been referred to as the Cl1 and Cl2 cleavages (Steuer and Platt, 1981) or, alternatively, the S_2 cleavage and D3-related crenulation cleavage (Bosworth and Rowley, 1984). Recent mapping indicates a regionally developed cleavage (S_4), which was previously unrecognized, is also present in the easternmost part of the Allochthon.

On the basis of the distribution and intensity of the S_2, S_3, and S_4 cleavages, the northern Taconic Allochthon in our study transect (at a latitude of about 43°37'N) can be subdivided into three structural domains. These structural domains are defined from west to east as follows (Fig. 3). Domain A, which extends from the Taconic Frontal thrust to the middle of the Pine Pond thrust sheet, contains a well developed S_2 and a weakly developed S_3. Domain B, which ranges from the middle of the Pine Pond thrust sheet to just east of the peak of Bird Mountain, contains a strongly overprinted S_2, a well developed S_3, and a very weakly developed S_4. Domain C, which extends eastward from domain B, contains a strongly developed S_3 and a moderately to strongly developed S_4. Although the domains are distinct, the domain boundaries represent gradational rather than abrupt changes in the cleavage distribution.

The three regionally developed cleavages differ in geometry, distribution, and intensity and are shown schematically in a geologic cross section across the study transect (Fig. 9). Kinematic studies suggest that S_2 and S_4 developed as a result of west-northwest-directed noncoaxial flow, whereas the flow type for S_3 development is not conclusive based on available data. These conclusions are supported by cleavage-related strain determinations and kinematic indicators associated with individual cleavages. In the following, we describe briefly the morphology, geometry, and distribution of individual cleavages, and we examine kinematic indicators for individual cleavages.

S_1 cleavage. The S_1 cleavage is locally preserved in Domain A of the northern Taconic Allochthon. S_1 is observed to both lie at an angle to bedding and to lie parallel to bedding (Passchier and Urai, 1988). Due to limited preservation, the regional significance of S_1 is unclear although preliminary results suggest that S_1 is related to either early folding or early thrust faulting..

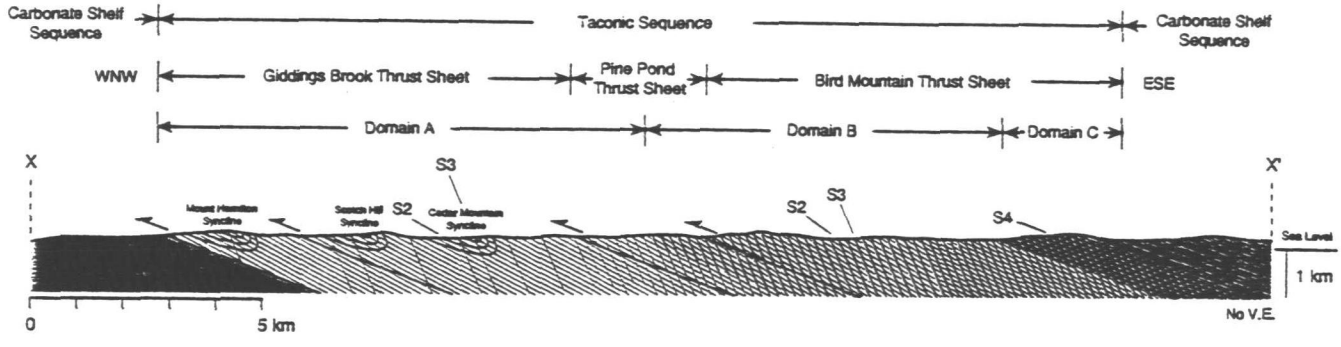

Figure 9. Schematic Cross section of the northern Taconic allochthon showing geometry, intensity and distribution of regionally developed cleavages. Line XX' is shown in fig. 3.

S_2 cleavage .The S_2 cleavage is usually referred to as the regional slaty cleavage because of its pervasiveness within the Taconic Allochthon. In coarse-grained rocks, the S_2 cleavage is a rough to smooth disjunctive cleavage, whereas in finer-grained rocks, it is a fine, continuous cleavage. Textures indicative of dissolution and reprecipitation are commonly observed for S_2. Dissolution is suggested by offset veins and bedding and by truncated detrital grains, whereas reprecipitation is evidenced by strain fringes and veins. The S_2 cleavage is fairly homogeneously distributed and shows minimal variation in orientation across the study transect. In Domains A and B, S_2 strikes on average north-northeast and dips moderately to the east (Fig. 4 and 10). In Domain C, however, measurement of S_2 in the field is not possible because it is strongly overprinted by the S_3 and S_4 cleavages. A graph of the dip of S_2 versus distance from the Taconic Frontal thrust (Fig. 11) shows that the mean dip of S_2 for individual localities ranges only between about 20°E and 40°E. A strong mineral lineation (L_2), which is defined by the alignment of platy minerals (chlorite and white micas) on S_2 surfaces, is a penetrative feature of S_2. This

lineation is generally observable only when a hand lens is used. The mapping of L_2 was conducted largely in Domain A because L_2 is least disturbed by later cleavages in this domain. The results show a consistent east-southeast trend for L_2 (Fig. 10c) similar to that observed by Pickens (1993). Microstructural evidence suggests that S_2 may have

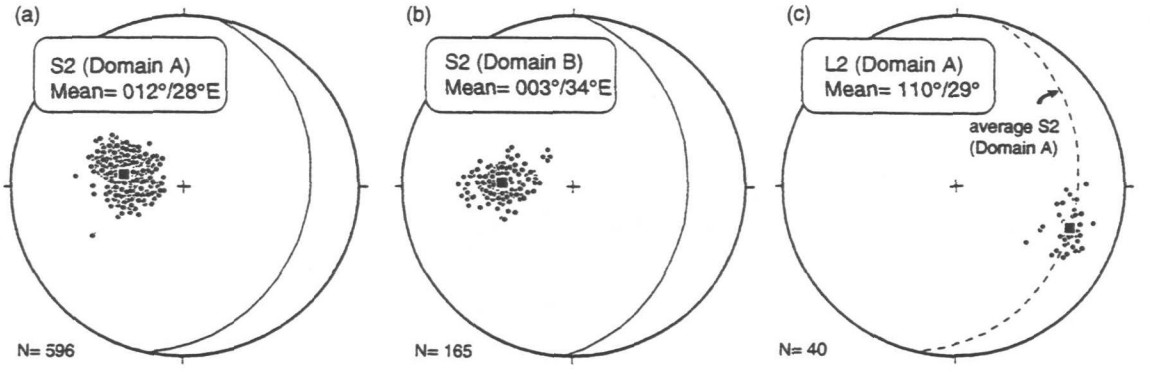

Figure 10. Lower hemisphere, equal area projections of poles to S_2 in domains A and B and L_2 in domain A. Squares show mean orientations.

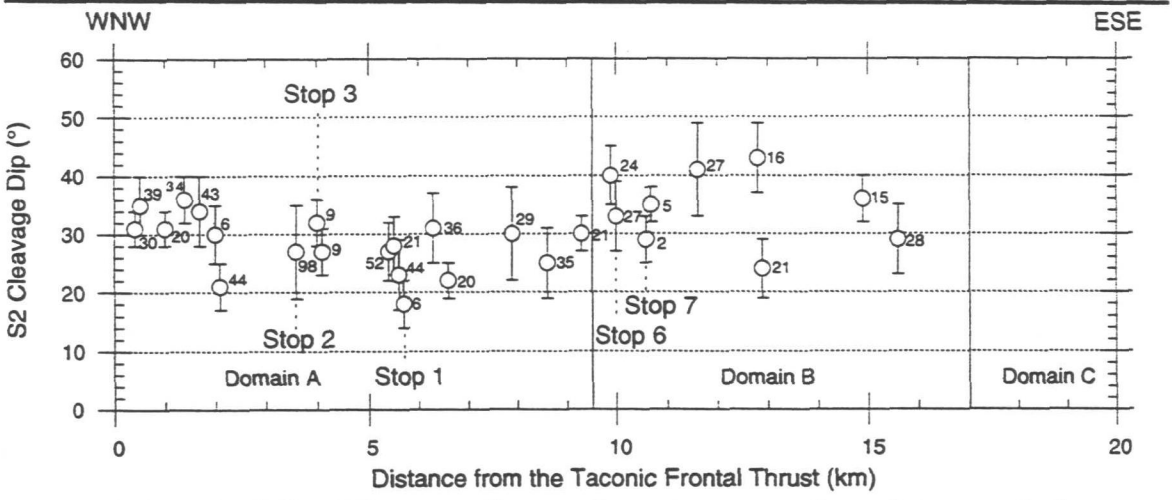

Figure 11. Graph of dip of S_2 versus distance from the taconic frontal thrust. Circles are mean values, lines are one standard deviation and numbers of measurements are shown adjacent to the corresponding symbol.

developed in a noncoaxial flow regime, although this is not a universally held opinion. This evidence comes from kinematic indicators including strain fringes around pyrite framboids and syn-slaty cleavage vein fibers. In addition, basic field relations and evaluation of the relation between S_2 and folding in the core of a regional-scale fold indicates S_2 was superposed on existing asymmetric folds formed by flexural folding.

Evidence for noncoaxial development of S_2: Strain fringes around pyrite framboids Microscale pyrite framboids are commonly observed in the black slates of the northern Taconic Allochthon. The pyrite framboids usually have well defined strain fringes, which provide evidence for the strain history associated with S_2. Several lines of evidence suggest that the strain fringes around the pyrite framboids formed during S_2 development. First, the fibers within the strain fringes show smooth curvature without abrupt changes in orientation or kinks, suggesting that the strain fringes are related to only one deformational event. Second, the S_3 crenulation cleavage is not present or, at most, very poorly developed in the samples used for strain-fringe analysis. Therefore, fiber growth related to the S_3 cleavage is unlikely. Third, the trace of the fibers in the strain fringes in XY sections is parallel to the phyllosilicate lineation in the matrix which formed during S_2 development. This parallelism strongly suggests that the strain fringes are S_2 related.

In XZ sections, the strain fringes around pyrite framboids display smooth clockwise curvature from oldest to youngest fibers when viewed to the NNE, and the youngest fibers, which are adjacent to the framboids, lie at an angle to S_2. The geometry, relative length, and sense of curvature of the fibers in the strain fringes are remarkably consistent across Domain A. On the basis of the similarity in fiber curvature on long and short limbs as well as the similarity in fiber curvature across the ten-kilometer width of Domain A, the fiber geometry is interpreted to reflect a component of top-to-the-west-northwest simple shear. Results of the strain-fringe measurements from five localities in Domain A are shown in Fig. 12, which illustrates the relation between the value determined for the finite elongation using the technique of Durney and Ramsay (1973) and the distance from the Taconic Frontal thrust. The values determined for the finite elongation for the five localities range from 1.1 to 1.4.

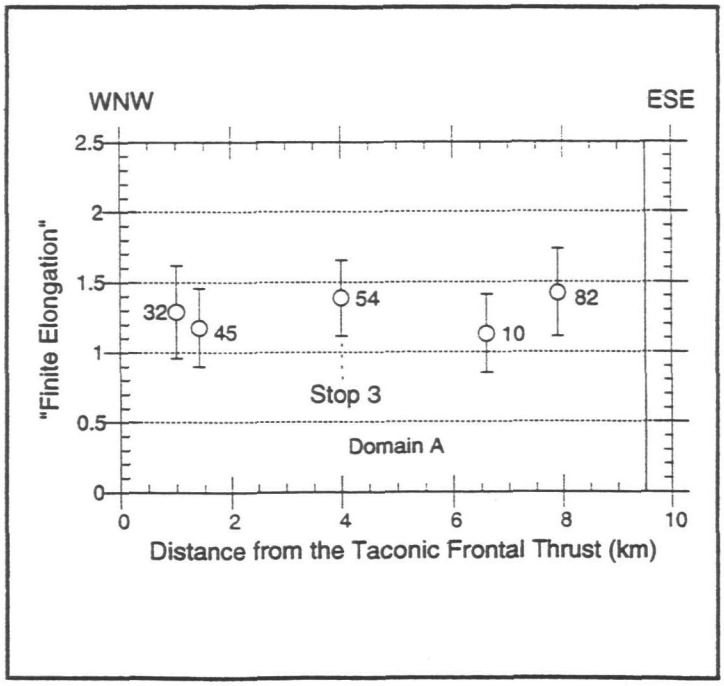

Figure 12. Graph of finite elongation determined from strain fringes around pyrite framboids versus distance from the Taconic frontal thrust. Circles are mean values, error bars are one standard deviation and numbers of measurements are shown.

Evidence for noncoaxial development of S_2: Syn-slaty cleavage vein material. Veins in the northern Taconic Allochthon also record evidence for

top-to-the-west-northwest noncoaxial flow during slaty cleavage development. Exposures of black slate in Domain A, which bear the syn-S_2 strain fringes around pyrite framboids, typically contain two main sets of veins, one striking on average north-south and the other striking on average east-west (Fig. 13). The veins are deformed, being either folded or boudinaged depending on their orientation with respect to S_2. In sandstone and carbonate horizons, the veins lie at approximately right angles to bedding, whereas in slate and silty slate, they lie at low to moderate angles to bedding. Two ages of fill are recognized in the veins that lie in slate and silty slate (Crespi and Chan, 1996). The pre-S_2 vein fill displays blocky as well as fibrous textures and lies within the central part of the veins. Although predominantly composed of calcite, the blocky pre-S_2 vein fill may also contain minor amounts of quartz. In contrast, the syn-S_2 vein fill is composed consistently of fibrous calcite and lies along the margins of the veins. Veins in the north-south striking set more commonly contain both ages of fill, indicating that these veins were preferentially oriented for reactivation during S_2 development.

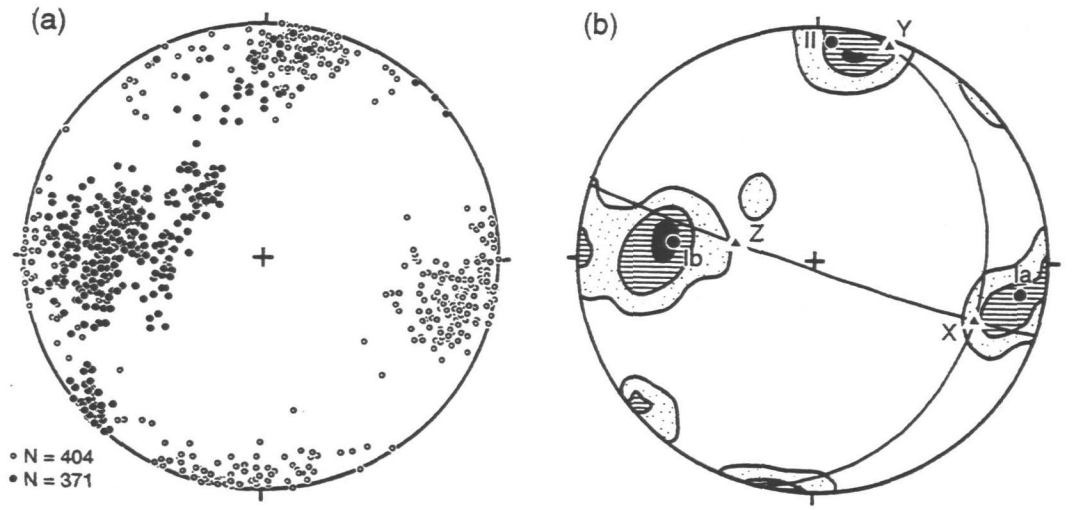

Figure 13. Lower hemisphere, equal area projections of poles to earlt layer-parallel extension veins in domain A. Solid circles are poles to veins in slate. Open circles are poles to veins in limestones and sandstones. X, Y and Z are principal strain axes for S_2 cleavage. Contour interval is 2% per 1% area.

The orientation of the youngest syn-S_2 vein fibers with respect to S_2 can be used to determine whether S_2 formed in a coaxial or noncoaxial flow field. For coaxial flow, the youngest syn-S_2 vein fibers should lie parallel to S_2, whereas for noncoaxial flow, they should lie at an angle to S_2, with the disposition of the fibers to S_2 giving the sense of shear. This assumes the fibers grew parallel to the maximum instantaneous stretching axis, which is likely the case because offset bedding in the wall rock can be connected across a vein by following the trace of the fibers. Because fibers in boudin necks commonly display complex geometries, only non-boudinaged veins in slate or silty slate were analyzed to ensure measurement of the youngest syn-S_2 fibers. Evidence for antitaxial fiber growth

indicates the youngest syn-S_2 vein fibers lie immediately adjacent to the veins walls.

Only four veins were analyzed because of the rarity of non-boudinaged veins in the slate and silty slate. However, in each case, the youngest syn-S_2 vein fibers are at an angle to S_2 such that the fibers lie clockwise from S_2 when viewed to the north-northeast; mean angles range from 26° to 34°. Like the results from the analysis of the syn-S_2 strain fringes around pyrite framboids, these results are consistent with top-to-the-west-northwest noncoaxial flow during S_2 development.

Relation of slaty cleavage development to folding. The relation between slaty cleavage development and folding in Domain A was evaluated at an outcrop of the Scotch Hill syncline. The outcrop is unusual because it exposes the core of a regional-scale fold and because it contains small-scale structures, in particular veins, that may be related to fold development. On the subvertical short limb of the syncline, en echelon, sigmoidal veins are "Z"-shaped when viewed to the north-northeast, indicating east-side-down displacement parallel to bedding. On the gently east-dipping long limb, sigmoidal veins are "S"-shaped when viewed to the north-northeast, indicating top-to-the-west displacement parallel to bedding. These observations suggest that this asymmetric fold formed by flexural folding and that pinning was along the hinge, at least from the time of initial formation of the veins. Thus, if hinge migration was involved in the development of fold asymmetry, it must have occurred early in the history of folding. Bedding-parallel veins at the outcrop are also consistent with flexural folding and a pinned hinge. The two bedding-parallel veins lie on the long limb of the fold and die out at similar distances from the hinge. In addition, inclined layer-silicate microstructures (Crespi, 1993) in the veins indicate top-to-the-west displacement. The absence of bedding-parallel veins on the short limb of the fold may be a result of the outcrop geometry, which preserves a greater length of the long limb.

The preceding observations together with the results of the strain-fringe and vein-fiber analyses imply that the slaty cleavage is not related to folding but was superposed on the fold shapes that developed as a result of the flexural folding process.

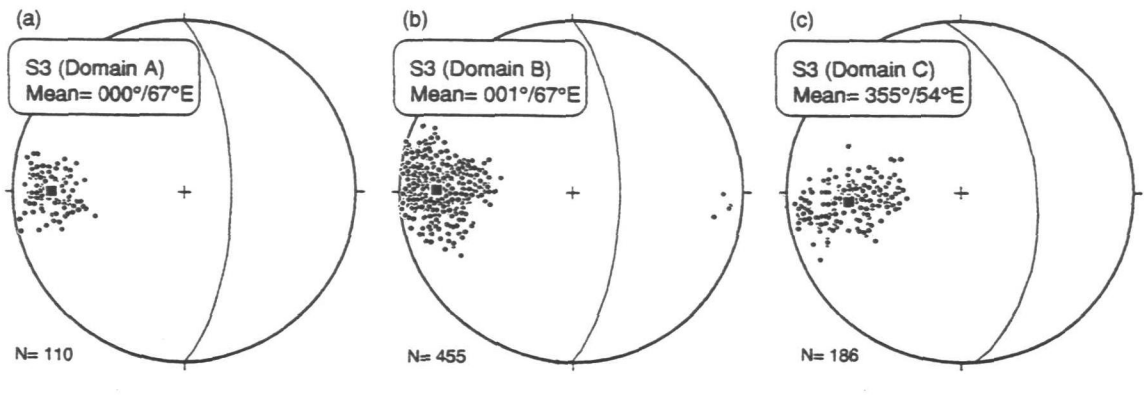

Figure 14. Lower hemisphere, equal-area projections of poles to S_3 in domains A, B and C. Squares show mean orientations.

Although data were not obtained for the short limb of the Scotch Hill syncline because of the lack of suitable markers, strain-fringe and vein-fiber data for the short limb of the Mount Hamilton syncline record top-to-the-west-northwest displacement, which is inconsistent with flexural folding and pinning along the fold hinge. The isoclinal geometry of the folds in the northern Taconic Allochthon, therefore, appears to be a result of flattening of flexural folds during regional-scale, top-to-the-west-northwest noncoaxial flow and slaty cleavage development.

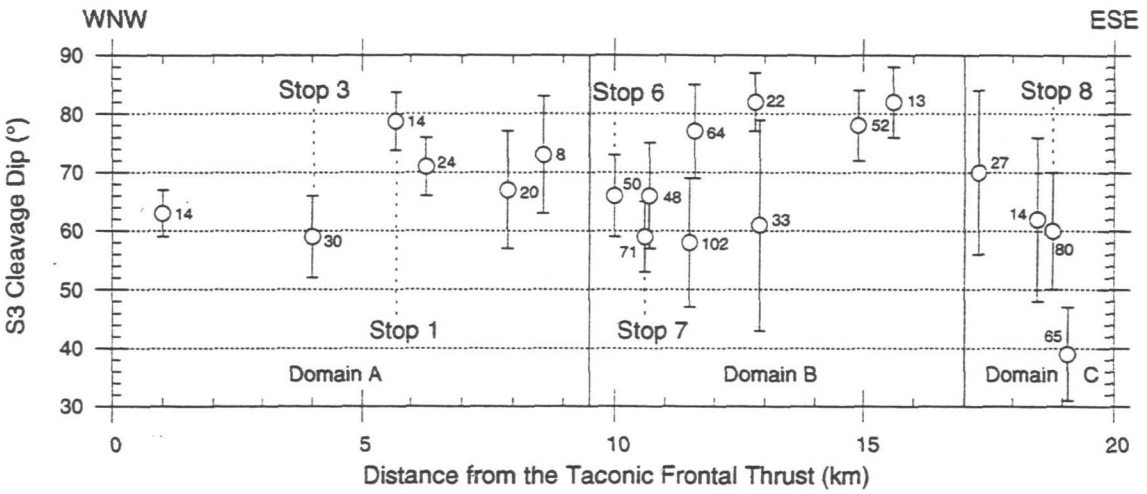

Figure 15. Graph of dip of S_3 versus distance from the Taconic frontal thrust. Circles are mean values, error bars are one standard deviation and numbers of measurements are shown adjacent to symbols.

S_3 cleavage. The S_3 cleavage is a crenulation cleavage and is dominated by pressure solution. The S_3 cleavage is mainly a zonal crenulation cleavage characterized by gradational boundaries between the cleavage and microlithon domains, although it also appears as a discrete crenulation cleavage characterized by abrupt boundaries between microlithon domains. From west to east across the study transect, the spacing of the S_3 cleavage decreases, and the amount of crystallization of new micas along cleavage domains increases. The S_3 cleavage is only locally present in

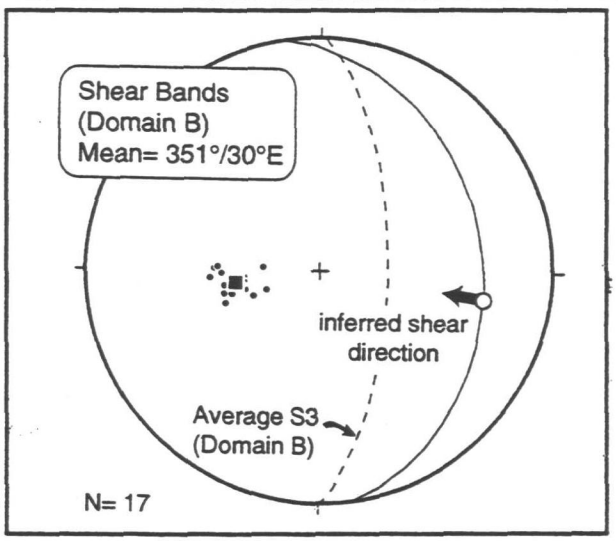

Figure 16. Lower hemisphere, equal area projection of poles to shear bands in domain B. Square is mean orientation

Domain A and becomes more pervasive in Domains B and C. S_3 strikes on average north and dips steeply to the east (Fig. 14). A graph of the dip of S_3 versus distance from the Taconic Frontal thrust (Fig. 15) shows that the mean dip of S_3 ranges only between about 60°E and 80°E. Although the dip of the S_3 cleavage is relatively constant across the study transect in Domain B, it decreases systematically from west to east in Domain C. Shallowly east-dipping extensional shear bands are locally present in Domain B. These shear bands are commonly traceable for less than one meter and have a spacing of about one decimeter to a few decimeters. All the observed extensional shear bands truncate S_3 in their centers; however, their tips curve smoothly into near parallelism with S_3. This suggests that these extensional shear bands were developed late in the history of S_3. The geometry of the extensional shear bands gives a top-to-the-west-northwest sense of shear (Fig. 16).

S_4 cleavage. The S_4 cleavage is a previously unrecognized cleavage that crosscuts S_3. Similar to S_3, pressure solution is the dominant deformation mechanism, although S_4 shows stronger crystallization of new micas along the cleavage domains. The S_4 cleavage domains are dominated by muscovite and chlorite; minor minerals include quartz, chloritoid, albite, paragonite, and rutile. The thickness of the cleavage-domain phyllosilicate grains ranges from about 10 to 100 microns, which is significantly greater than that observed for the phyllosilicate grains in the cleavage domains of S_2 and S_3. The S_4 microlithon domains contain dominantly quartz, albite, muscovite, chlorite, and rutile.

The S_4 cleavage is weakly developed in the easternmost part of Domain B; however, it is pervasively developed in Domain C. The intensity of the S_4 cleavage increases from west to east. S_4 strikes on average north and dips gently to the east (Fig. 17a). Mesoscopically, S_4 is characteristically lustrous and has a mineral lineation (L_4)

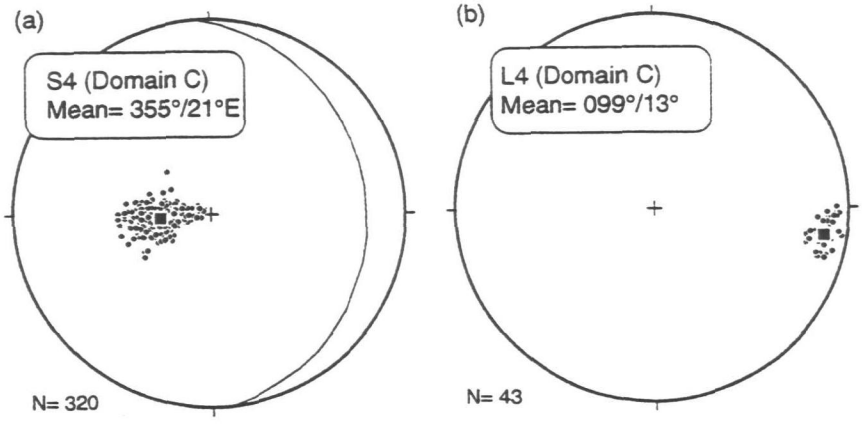

Figure 17. Lower hemisphere, equal area projections of poles to S4 and L4 in domain C. Squares are mean orientations.

which is readily observed and measured in the field (Fig. 17b). A graph of the dip of S_4 versus distance from the Taconic Frontal thrust shows that the mean dip of S_4 gradually decreases from about 30°E to 10°E from west to east (Fig. 18).

The microlithon domains of S_4 commonly contain syntectonic albite porphyroblasts. These albite porphyroblasts generally have a long dimension less than 0.25 mm and a short dimension less than 0.12 mm. The porphyroblasts are characterized by inclusion-rich albite cores and inclusion-free albite overgrowths. The kinematics of the S_4 cleavage is indicated by the asymmetry of the portion of the albite overgrowths that is related to the development of S_4. The asymmetry of the albite overgrowths suggests that the S_4 cleavage developed noncoaxially with a top-to-the-west-northwest sense of shear.

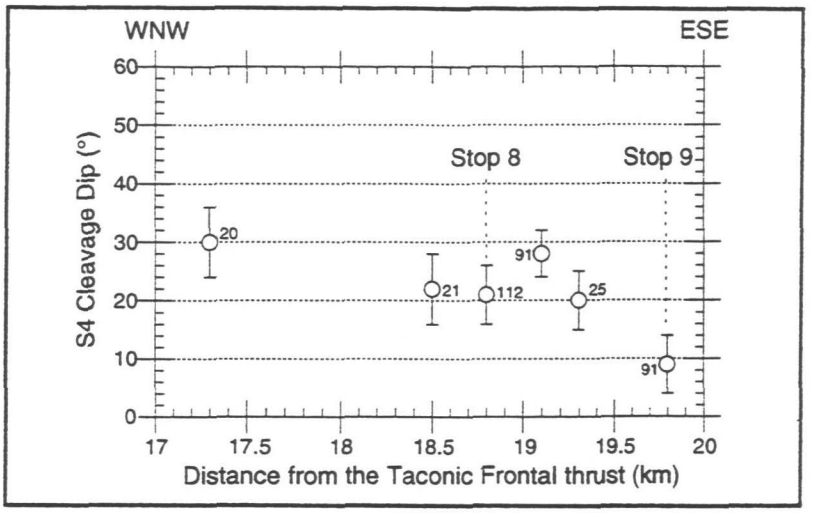

Figure 18. Graph of dip of S4 vs. Distance from the Taconic frontal thrust. Circles are mean values, error bars are one standard deviation and numbers of measurements are shown.

Discussion. An important question is the age and tectonic significance of the S_2, S_3, and S_4 cleavages. Harper (1968) dated six samples of slate and phyllite from the northern Taconic Allochthon using conventional K-Ar whole-rock methodology. His results were recalculated by Sutter et al. (1985), who applied newer constants for the K-Ar radiometric system. The recalculated apparent ages range from 468±8 to 381±8 Ma, i.e., from middle Ordovician to middle Devonian. Although Sutter et al. (1985) argue against a possible Acadian age for the cleavages in the Taconic Allochthon, the six apparent ages correspond very well with the deformational complexity at Harper's sample locations. The two oldest ages (468±8, 453±8 Ma) are from locations where S_2 is dominant, and the two youngest ages (394±8, 381±8 Ma) are from locations where the rocks are severely overprinted by S_3. Because the peak temperature for the formation of slate and phyllite is very likely below the closure temperature for argon retention in white mica, the radiometric ages of white mica growing in cleavage domains may represent the formation age of the white mica, and, thus, the cleavage. For samples with more than one cleavage, the apparent age may represent a mixed age such that the obtained age is a maximum estimate for the youngest cleavage generation in the sample. Therefore, the results of Harper's dating suggest that S_2 is Taconic in age, whereas S_3 and S_4 are younger and possibly Acadian in age.

The available radiometric ages, however, are open to question because of uncertainty with respect to the assumption that the white mica growing in the cleavage domains is the main potassium-bearing phase in the slate and phyllite. For example,

detrital mica, detrital potassium feldspar, and diagenetic mica are all possible potassium-bearing minerals in these rocks. Thus, a better dating technique should be applied. We (Chan and Crespi) are currently dating samples in which a single cleavage (either S_2, S_3, or S_4) is dominant using the laser probe ^{40}Ar-^{39}Ar method. This advanced dating method will help in constraining the age of specific cleavage domains in a slate or phyllite with good precision, and the results should provide a better estimation of the cleavage ages. Samples from the study transect as well as Harper's original sample locations are included in this geochronologic study.

In addition to the available geochronologic data, differences in orientation data for S_2 and S_3/S_4 also suggest that S_2 and S_3/S_4 may have developed during separate tectonic events. First, S_2 strikes on average 012^0 (Domain A), whereas S_3 and S_4 strike on average 001^0 (Domain B) and 355^0 (Domain C), respectively. Second, L_2 trends on average 110^0, whereas L_4 trends on average 099^0. (L_3, which is less well developed than L_4, trends about parallel to L_4.) Third, S_2 is relatively homogeneous in terms of cleavage spacing, whereas the spacing of S_3 and S_4 decreases from west to east. Although not conclusive, these differences are consistent with at least two tectonic events for S_2 and S_3/S_4. It remains unclear, however, whether there is a significant time gap between the development of S_3 and S_4 without additional radiometric ages.

S_2 is interpreted to have formed during the WNW-directed emplacement of the northern Taconic Allochthon. In addition, structural data for S_2 can be explained relatively easily using the cleavage trajectory model proposed by Sanderson (1982), which has been applied to a number of other slate belts (e.g., Gray and Willman, 1991; Sanderson, 1979; Tillman and Byrne, 1995). In this model, cleavage curves asymptotically toward the base of a regional-scale thrust sheet (Fig. 19). In the upper portion of the thrust sheet, cleavage development is characterized by coaxial flow, whereas in the lower portion, cleavage development is characterized by noncoaxial flow. If Sanderson's cleavage trajectory model applies, the kinematics and geometry of S_2 indicate the study transect lies within the lower portion of a thrust sheet and at the same

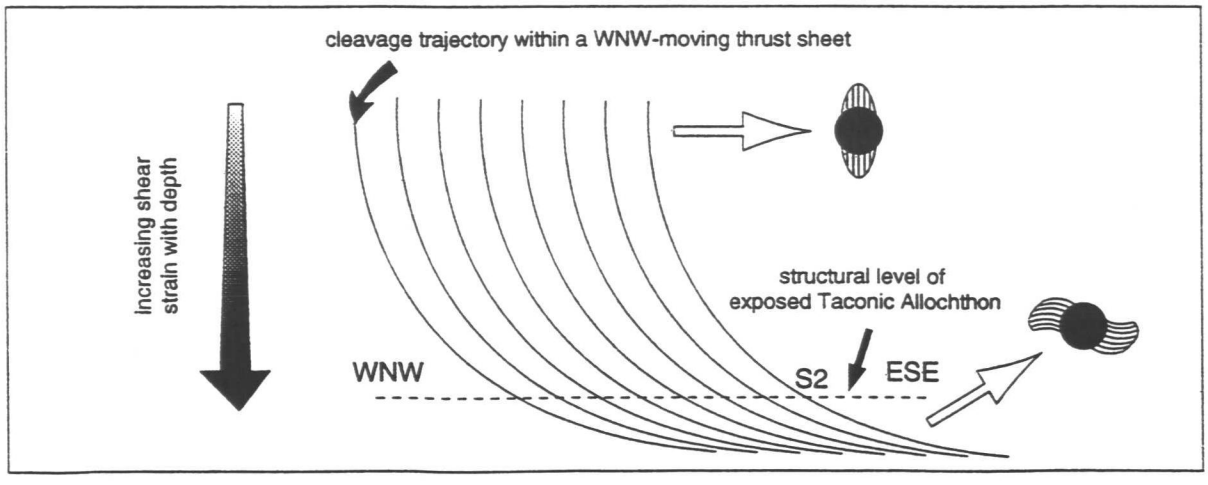

Figure 19. Schematic diagram showing cleavage-trajectory model of the interpretation of S_2.

structural level. This interpretation is suggested by (1) the lack of variation in S_2 orientation, (2) the lack of variation in strain magnitude as determined from syn-S_2 strain fringes around pyrite framboids, and (3) the consistent curvature of the syn-S_2 strain fringes.

In contrast to S_2, the tectonic significance of S_3 is less straightforward because of the lack of reliable microscale kinematic indicators for S_3. Based on available structural data related to S_3, possible tectonic interpretations for S_3 include (1) development of S_3 as an axial planar cleavage in a coaxial flow during formation of a regional-scale fold and (2) development of S_3 as a footwall cleavage in a noncoaxial flow during a regional-scale thrusting event. The increase in intensity of S_3 from west to east, which suggests an increase in strain magnitude, is consistent with both tectonic interpretations (i.e., an increase in strain from an upright to an overturned limb and an increase in strain toward a thrust plane). The late-S_3 extensional shear bands dip in only one direction, suggesting S_3 may have formed in a noncoaxial flow. However, because the extensional shear bands are only found at one locality, they may not be representative for the study transect. It is also possible that S_3 formed in a coaxial flow with a possible post-S_3 noncoaxial event being responsible for the shear bands.

Insight into the tectonic significance of S_4 can be gained by comparing the geometric, morphologic, and kinematic characteristics of S_4 with those of S_2 and S_3. First, S_4 dips relatively gently to the east in contrast to the steeply dipping S_3. Second, S_4 shows a rapid transition from weakly to strongly developed in contrast to the relatively gradual transition for S_3 and the homogeneously developed S_2. Third, the stretching lineation on S_4 is very strong. Fourth, S_4 contains microscale albite porphyroblasts, which indicate S_4 developed in a noncoaxial flow. Together, these observations suggest that S_4 formed in a shear zone that is characterized by enhanced noncoaxial flow and high strain magnitude. In such a zone, the angle between the shear plane and the developed cleavage is very low. Thus, the development of S_4 may have been related to motion along a regional-scale, WNW-directed thrust.

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Road Log

Directions from Pico

Meet at parking area of Pico. Drive east on Rt. 4 towards Rutland.
Intersection of Rt. 4 and Rt. 7. Turn Left to stay on Rt. 4 east.
Turn right to remain on Rt. 4.
Exit to Rt. 30 south. Drive to Intersection with with Rt. 4A. Turn right onto Rt. 4A at Castleton Corners.

Mileage begins at Intersection of Rt 30 with Rt 4A

1.2	Turn right (west) onto Creek Road. Roadcuts of Mettawee fm. On the left as you drive along the west shore of Lake Bomoseen.
5.1	Entrance to Lake Bomoseen State park on right; road curves sharply to left...continue straight onto dirt road which follows lake shore. Continue until road ends.
6.4	Park so as to not block in other drivers or residents.

STOP 1. Cedar Point Quarry

The Cedar Point quarry lies on the hinge of a first-order syncline developed in the Mettawee fm. This structure, typical of the Taconics (fig. 20), is extremely well displayed within the quarry. As one walks up into the quarry, a folded bedding surface is seen on the right which defines the hinge of this syncline. Along this surface one can see bedding cleavage intersections which parallel the fold axis. Within the quarry itself the same hinge is well displayed and the axial plane orientation of the slaty cleavage (S_2) can be seen. This site is one which was used for the study of reduction spots (Goldstein er al, 1995a) and the spots are easily seen here. Note that some spots are nearly perfect ellipsoids with their XY planes parallel to the cleavage. Other spots are irregular and were not measured for strain analyses. One can also observed pyrite porphyroblasts with pressure shadows and S_3 crenulation cleavage at this locality.

	Turn around and retrace the route to Creek Road.
7.6	Turn right onto Creek Road
7.7	Creek Road curves to left, continue straight onto dirt road.
7.8	Turn left onto another dirt road (Moscow Road ?) Park along right side of road so as to not block traffic. The stop lies at the intersection of the dirt road and Scotch Hill Rd.

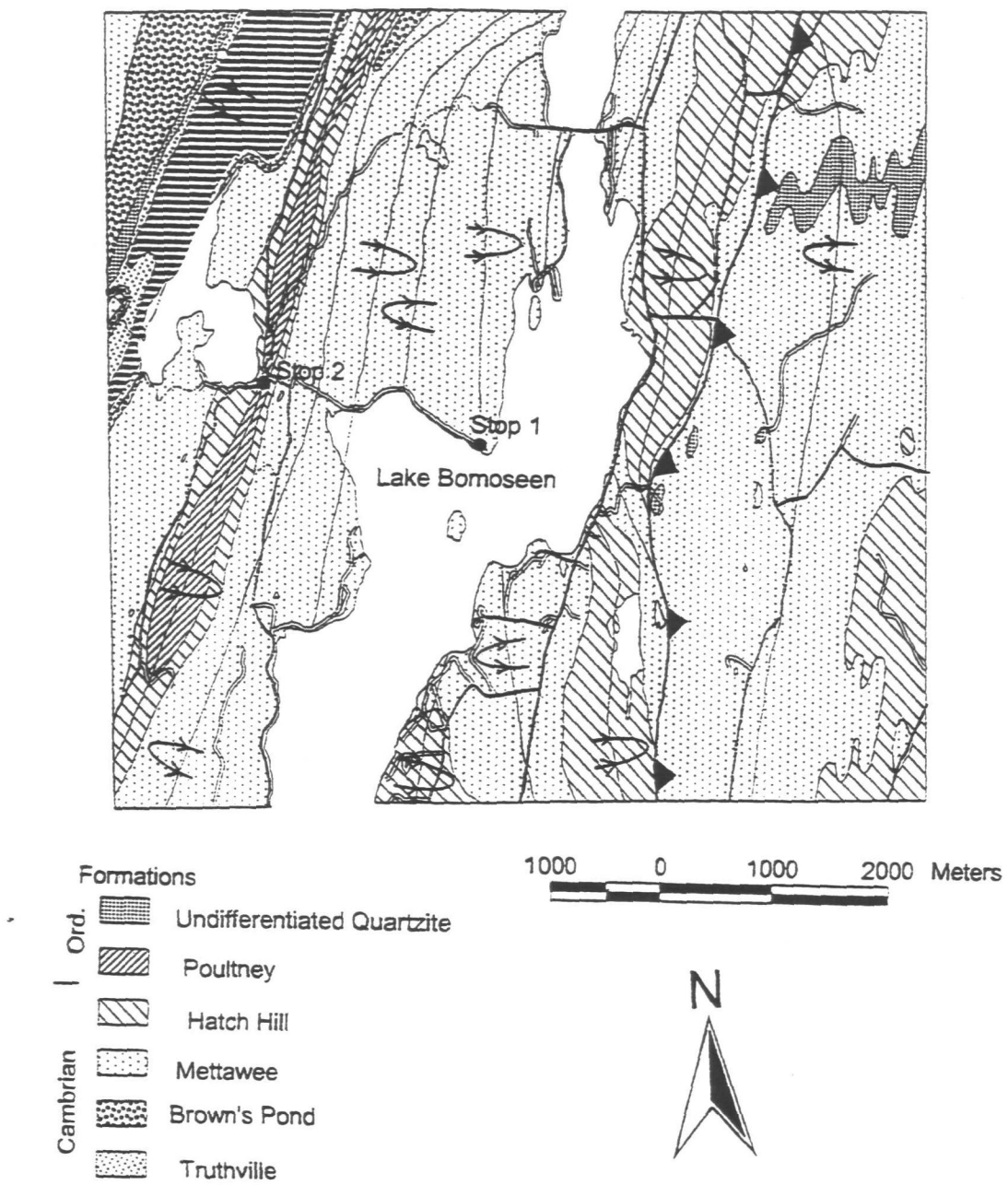


Figure 20. Geologic map of the southern portion of the Bomoseen 7 ½ minute Quadrangle showing locations of stops 1 and 2.

STOP 2 Scotch Hill Syncline

The purpose of this stop is to show some aspects of the pre-slaty cleavage deformation history of the Taconic sequence. In particular, we will examine two generations of veins, one related to early, layer-parallel extension and the other related to fold development, and we will evaluate evidence for fold development by flexural folding. The locality exposes the core of the Scotch Hill syncline, a regional-scale, west-vergent asymmetric fold. The strata at this location are within the Poultney fm. and are composed of cm. scale alternating layers of black and gray silty slate and brown-weathering dolomitic meta-arenites, which are bedded at the meter scale.

Two sets of veins can be observed in the lowermost slate horizon. The older set, which formed during a period of early, pre-folding, layer-parallel extension, appears to be approximately parallel to the slaty cleavage on the profile plane of the fold. Bedding-parallel surfaces show, however, that the veins lie at an angle to the slaty cleavage and are noncoaxial with the fold. Good three-dimensional exposures of the veins are visible on the short limb of the fold. These pre-folding veins also show little variation in orientation across the fold, being essentially parallel on the long and short limbs. This results from a combination of rotation of the fold limbs and distributed shear parallel to layering in the siltstone during flexural folding. Cross-cutting relations between the early, pre-folding, layer-parallel extension veins and a set of younger, en echelon, sigmoidal veins are visible on both the long and short limbs of the fold. The en echelon, sigmoidal veins are coaxial with the fold and formed during fold development. On the short limb, the veins are "Z"-shaped, indicating east-side-down sense of shear along bedding. Veins that display a straightforward sigmoidal geometry on the long limb are "S"-shaped, indicating top-to-the-west sense of shear along bedding. These veins provide evidence for distributed shear along bedding during flexural folding and indicate pinning was along the fold hinge.

Another set of veins, consisting of two bedding-parallel veins, is present in the exposure and provides evidence for discontinuous shear parallel to bedding during flexural folding. These veins lie in the center of the two dolomite units and only exist on the long limb of the fold.

Finally, the approximately axial planar nature of the slaty cleavage (S_2) is well shown at this locality. S_2 is inferred to have been superposed on the fold subsequent to vein development and flexural folding.

7.8	Turn right onto Scotch Hill Road
11.8	Intersection with Rt. 4; park along right side of road. Walk down exit ramp towards Rt. 4 east.

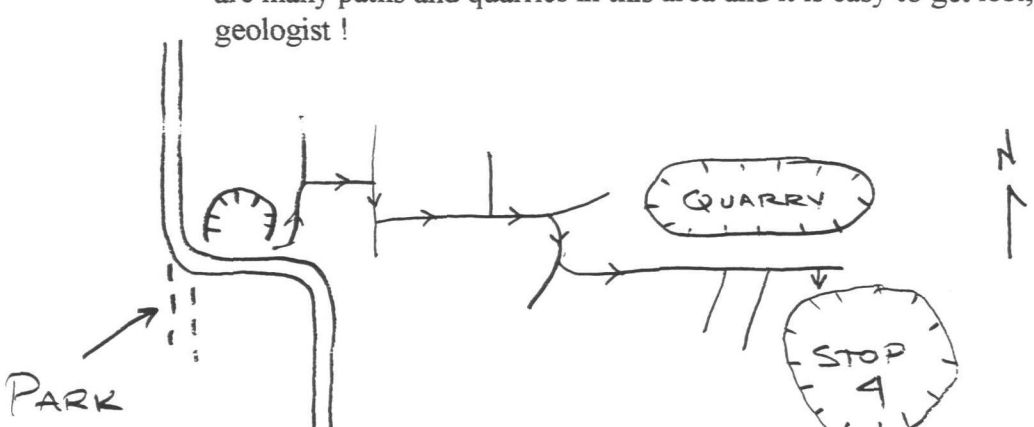
STOP 3

The purpose of this stop is to show meso- and microscale structures that are characteristic of Domain A and to present evidence for top-to-the-west-northwest

noncoaxial flow during slaty cleavage (S_2) development. In addition, some aspects of the history of vein development will be noted.

The locality exposes the long limb of the Scotch Hill syncline, and the strata are composed of slate and silty slate interbedded with limestone and limestone conglomerate. As is typical of Domain A, the slaty cleavage (S_2) is the main cleavage at this locality, and it dips on average 30¡ to the east-southeast. A crenulation cleavage, which dips on average 59¡ to the east, is also locally present. Although not visible in the field, a penetrative mineral lineation (L2) defined by the preferred orientation of phyllosilicates is present on the S_2 surfaces. Also prominent at the exposure are early layer-parallel extension veins which have been variably reactivated during slaty cleavage development. The veins, which are primarily composed of calcite, lie in two main sets, the veins in one striking on average north-south and the veins in the other striking on average east-west. Good examples of boudinaged veins belonging to the north-south striking set and folded veins belonging to the east-west striking set are present within the slate layers throughout the exposure. Pre-S_2 and syn-S_2 vein fill can be discerned for some of the veins in the field using a hand lens, although microscale observations are generally needed to confirm the age of the vein fill. Reactivation of the veins during S_2 development has resulted in complex vein intersection geometries in the slate layers (Crespi and Chan, 1996); some good examples are visible on the top of the exposure between Rte. 4 and the exit ramp. Finally, syn-S_2 strain fringes from this locality provide evidence for top-to-the-west-northwest noncoaxial flow during S_2 development.

	Continue straight on Scotch Hill Rd.
12.3	Flashing light; continue straight towards Fair Haven, Vt. Drive through town of Fair Haven.
13.0	Turn left onto River St.
14.1	Turn right onto Evergreen Rd. Quarries in Mettawee Fm. On left. These are active quarries and should not be entered without permission.
15.3	Road curves sharply to left, park on Sawmill Rd. which continues straight. Walk back to Evergreen Rd. And follow path to right of pile of rock dust. Follow sketch map to find the quarry which is the next stop. Note: there are many paths and quarries in this area and it is easy to get lost, even for a geologist !



STOP 4 : LUNCH

This quarry is located on the upright limb of one of the isoclinal folds and displays parallel bedding and cleavage. Our purpose here is to examine another more reduction spots as well as pyrite porphyroblasts with pressure shadows and late stage faulting and veining. The quarry sits astride the contact between the Mettawee and Hatch Hill (black, rusty-weathering slate) formations and the contact is exposed on the upper bench of the quarry. Veining is abundant as large sigmoidal veins filling late brittle faults and as veins localized within meta-arenites in the hatch Hill fm. The sigmoidal veins fill early faults whcih were then further deformed in a ductile shear environment to yield the sigmoidal shape.

This site was also used of an oxygen isotope study. Early (folded) quartz veins, syn-tectonic quartz mineralization in boudin necks and the late veins were all analyzed for $\delta^{18}O$ in the stable isotope lab at the University of Wisconsin. All the isotopic ratios are nearly identical varying between 18.57 and 19.10. Further, no zoning of $\delta^{18}O$ within single veins was observed. We interpret these results as indicating that fluid flow was pervasive and intragranular resulting in homogenization of oxygen isotopes between fluids and rock.

	Continue south on Evergreen Rd.
16.4	Turn right onto Saltis Rd. The quarries near this intersection take stone from the Mettawee Fm., are active and you must receive permission to enter them.
16.6	Turn right onto Bolger Rd.
18.2	Turn left onto Rt. 22A
28.9	Turn left onto Rt. 24; Middle Granville. Cross steel deck bridge
29.0	Turn left onto Depot Rd.
29.8	Turn left onto Stoddard Rd.
30.3	Park on right side of road.

STOP 5 Deformed Graptolites

This stop exposes the principal graptolite-bearing stratigraphic unit within the Taconics. Rowley et al (1978) defined this rusty weathering black slate at the top of the Mt. Merino fm and named it the Stoddard Rd. Member. Because it occurs at the very top of the Taconic stratigraphy, the Stoddard Rd. Member is only found in the deepest synclines, which is the structural setting for this stop. The outcrop located on the road lies on the limb of this syncline as bedding and cleavage are parallel (note: bedding can be defined as the plane on which graptolites lie. Along the road, graptolites lie on cleavage indicating bedding-cleavage parallelism). Three species were used for strain measurement: *Orthograptus Calcaratus, Orthograptus Whitfieldii and Climacograptus Bicornis*. All three species are abundant, are easily recognized and have constant thecal spacing in the

undeformed state. For this site, graptolite measurement indicates that X is 32% elongation and Y is 40% shortening.

Across the road in the beaver pond is another exposure which lies on the hinge of the syncline. Bedding-cleavage intersection varies from 60° to 90°. The same species can be found in these rocks and their measurement indicates 60% shortening perpendicular to cleavage.

	Turn around and retrace the route.
30.9	Turn right onto Depot Rd.
31.6	Turn right onto Rt. 24, cross steel deck bridge in Middle Granville
31.7	Turn right onto Rt. 22A.
43.7	Center of Fair haven, Vt.
43.9	Follow Rt. 22A left around village green, stay on Rt. 22A.
44.8	Turn right to enter Rt. 4 east.
50.4	Park on shoulder of Rt. 4 by roadcut. It is legal to stop on this highway but vehicles must be pulled well off the road and should have flashers.

STOP 6

The purpose of this stop is to show the geometric and cross-cutting relations between the moderately east-dipping S_2 cleavage and the steeply east-dipping S_3 cleavage at the western edge of Domain B. Key features to observe include (1) the gradational change in the intensity of S_3 from a zone in which S_2 is dominant to a zone in which S_3 is dominant and (2) dissolution and folding of pre-existing layering and veins. Small lozenge-shaped blocks are commonly observed because of the angular relation between S_2 and S_3. In addition, asymmetrical folding of S_2 in which differentiation has resulted in coarser-grained hinges and finer-grained limbs can be observed.

	Proceed east on Rt. 4
50.7	Exit 5 from Rt. 4
50.9	Turn left onto Hubbarton Rd.
51.3	Park on right side of road.

STOP 7

The purpose of this stop is to show the main characteristics of the S_3 crenulation cleavage, where it is well developed in Domain B. Crenulation folds and evidence for pressure solution are well illustrated at this outcrop. Key features to observe include (1) differentiation associated with S_3 illustrated by coarse-grained fold hinges and fine-grained fold limbs, (2) the down-dip mineral lineation on the S_3 surface, defined mainly by chlorite and muscovite, and (3) dissolution and folding of pre-existing layering and veins. The spacing of the S_3 cleavage ranges from millimeters to centimeters.

	Turn around and retrace the route
51.6	Enter Rt. 4 east
57.1	Park well off the roadway.

STOP 8

The purpose of this stop is to show the geometric and cross-cutting relations between the S_3 and S_4 cleavages in Domain C. The S_4 cleavage is characterized by dissolution and newly developed phyllosilicate minerals in the cleavage domains. Key features to observe include (1) the relatively small spacing of the S_3 cleavage compared with that of the previous outcrop, (2) the spaced S_4 cleavage that crosscuts the steeply east-dipping S_3 cleavage, and (3) the lustrous S_4 surfaces and associated mineral lineation.

STOP 9 (Optional)

The purpose of this stop is to show the well developed S_4 cleavage. At this locality, the steeply dipping S_3 crenulation cleavage is almost completely overprinted. Key features to observe include (1) lustrous S_4 surfaces with a strong mineral lineation defined by phyllosilicates and (2) strongly folded and boudinaged pre-existing veins.

NAPPES, GNEISS DOMES AND PLUTONIC SHEETS OF WEST-CENTRAL NEW HAMPSHIRE

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Introduction

The area between Keene and Hanover in west-central New Hampshire (NH; Figure 1) is home to some "classic" geology of the Acadian Appalachians. This geology includes (1) fold nappes which transported highly metamorphosed deep-basin sediments over less-metamorphosed shelf sediments, (2) gneiss domes which subsequently deformed (and metamorphosed) these nappes, and (3) large anatectic plutonic sheets, whose emplacement may have been intimately involved with the formation of the nappes. These rocks are encompassed within the Central Maine Terrane (CMT; Eusden & Lyons, 1993), and astride the Bronson Hill Anticlinorium (BHA; Billings, 1956). On this trip, we are primarily interested in the relationships between and among rocks on the BHA and those further east in central NH. The relationship of these rocks (NH sequence) with those to the west (Vermont sequence) is the subject of other trips at this conference (A6, Armstrong *et al,* 1997.*;* and C1, Thompson *et al.*, 1997*).*

Metasedimentary and Metavolcanic Stratigraphy

An important part of understanding the geology in any region is the ability to recognize the relative sequence in which the rock units were deposited or intruded, from principles of super-position and cross-cutting relationships, but also from fossil assemblages and radiometric dates. On the basis of fossils discovered in the Littleton area, Billings and others (1934, 1935, 1937) were able to define a stratigraphic sequence for the metavolcanic and metasedimentary rocks of the BHA (Figure 2). Once a sequence is established, mapping in adjacent areas can commence by making correlations back to the known rocks. Billings, his students and others proceeded to map much of the rest of NH, with reference to the Littleton section, leading to the publication of a NH state geologic map in 1956. Most of the rocks throughout NH (with a few notable exceptions) are, however, devoid of the fossils needed to corroborate one's position in the stratigraphic sequence, thus correlations must be made primarily on the basis of lithologic characteristics. Given a high degree of structural complexity, a wide range of metamorphic changes and degrees of weathering, large expanses of intervening plutonic rocks, and significant original sedimentological variations, such correlations (upon which structural and tectonic interpretations hinge) are often problematic.

As the state map was being compiled, Thompson (1954, 1956) discovered structural complications on the BHA, arising from the recognition of distinct stratigraphic units (based on fossil occurrences) in rocks that had previously been mapped as belonging to one formation. This lead to a second round of mapping in the region, still referencing the Littleton section, that culminated in the oft-cited 1968 paper by Thompson and others, establishing the current paradigm outlined above.

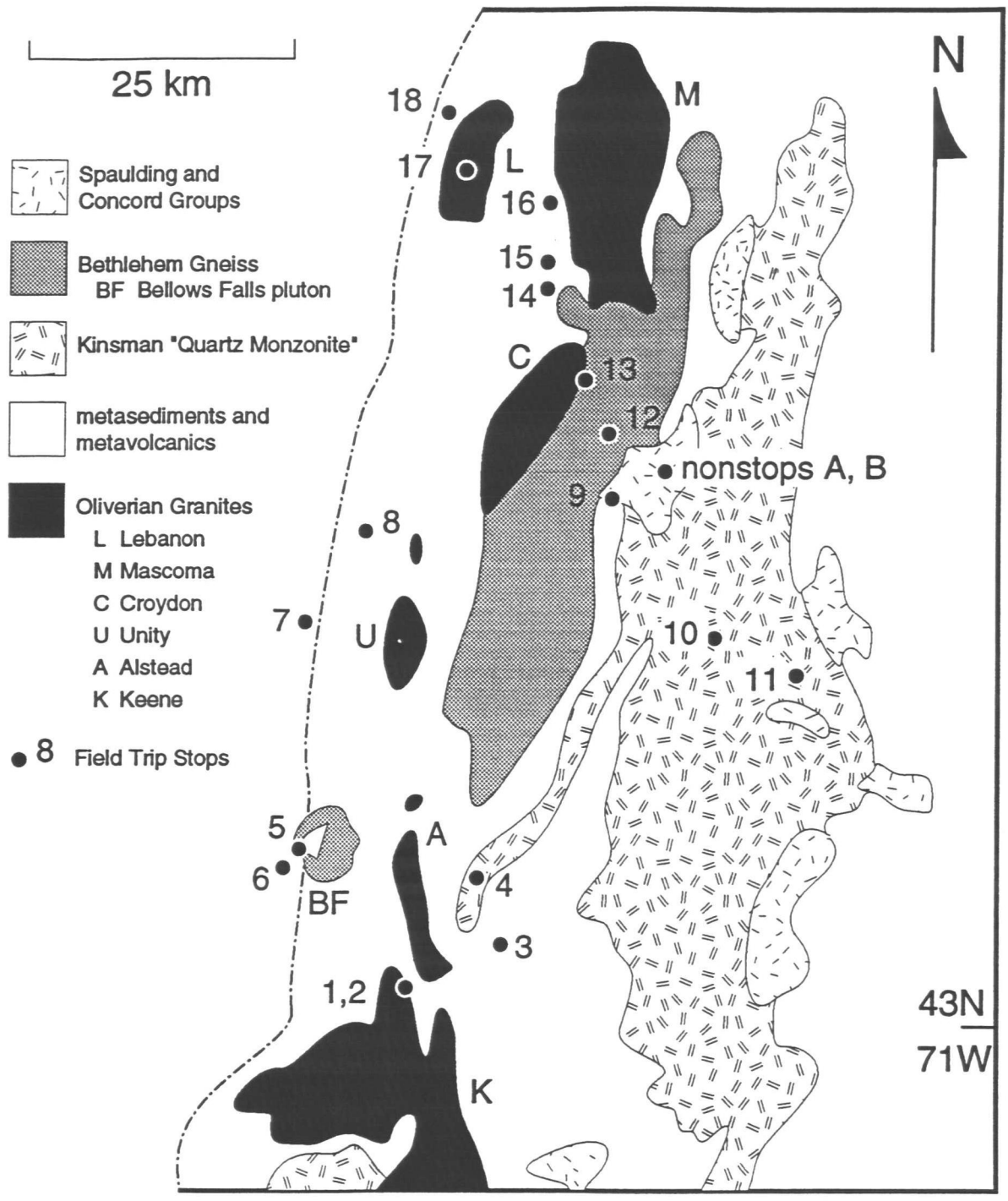
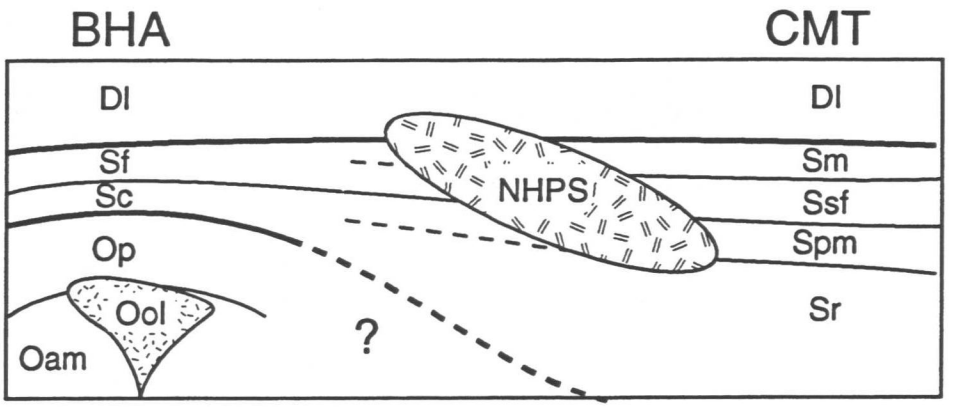


Figure 1: Schematic geologic map of west-central New Hampshire, showing approximate locations of field trip stops



Bronson Hill Anticlinorium

Dl: Devonian Littleton formation, an aluminous marine turbidite (good graded bedding) with quartzites and minor volcanics. This rock is regarded as a flysch deposit preceding the Acadian orogeny, and was derived from a source terrane to the east.

Sf: Silurian Fitch formation, a shelf-type dirty limestone or limey mudstone

Sc: Silurian Clough formation, a clean quartzite and quartz pebble conglomerate, with minor amounts of dirty limestone and greywacke or aluminous shales; locally bearing fossils (Boucot and Thompson, 1958, 1963). The Silurian rocks unconformably overlie the Ordovician rocks, including the Oliverian Granites.

Ool: Oliverian Gneiss, foliated calc-alkaline granites, tonalites and trondjhemites with ages of approximately 442-456 Ma (Naylor, 1969; Zartman & Leo, 1985; Aleinikoff c.f. Lyons et al., 1996).

Op: Ordovician Partridge formation, a carbonaceous and sulfidic black shale, with significant volcanogenic material (abundant coticules). The upper volcanics are probably coeval with the Oliverian granites.

Oam: Ordovician Ammonoosuc Volcanics, bimodal volcanics including basalt and rhyolite flows, tuffs, and pyroclastics. At ~468 Ma old (Aleinikoff c.f. Moench et al., 1995) they have been intruded by the Oliverian granites.

New Hampshire Plutonic Series

Dcg: Concord Granite, a two-mica granite: 354±5 Ma (Harrison et al., 1987)

Dsg: Spaulding Group, diorite to granite in composition: 392 Ma (Lyons & Livingston, 1977)

Dbg: Bethlehem Gneiss, foliated granite to tonalite, 407±5 to 410±5 Ma (Moench et al., 1995)

Dkqm: Kinsman Quartz Monzonite, tonalite to granite, 413±5 Ma (Barriero & Aleinikoff, 1985)

CMT metasediments

Dl: Devonian Littleton formation, as described at left

Sm: Silurian Madrid formation, a dirty "ribbon" limestone, with occasional black shales, sulfide rich zones, and arenaceous (sandy) rocks containing calcareous concretions.

Ssf: Silurian Smalls Falls formation, includes quartzites, argillites (shales), and limestones, generally black in color, with abundant pyrrhotite and pyrite.

Spm: Silurian Perry Mountain formation, typically impure but well bedded sandstones, sometimes with thinly interbedded (cyclic) shales (rarely showing graded beds). Non-calcareous and usually non-sulfidic.

Sr: Silurian Rangeley formation, includes all kinds of rocks including graphitic and sulfidic black shales, polymict conglomerates (matrix dominant), and arenaceous lenses with calcareous concretions. The Rangeley has been described as an olistostrome (Eusden et al., 1996a, b; Allen, 1996a), and was deposited in a rapidly subsiding basin from a source terrane to the northwest (Moench, 1979).

Figure 2: stratigraphy and rock descriptions for the field trip area

Meanwhile, Lyons, his students, and others working in central NH found that the BHA stratigraphy didn't fit very well with rocks exposed there, and developed a new informal lithostratigraphy for their work (*c.f.* Duke *et al.,* 1988). Billings and others (1941, 1946, 1975) had earlier wrestled with this problem in the Mount Washington-Gorham area further north, and in revisiting this area Hatch and others (1983; Duke *et al.,* 1988) extended to central NH a stratigraphic sequence established in the Rangeley area of Maine (Figure 2; Boone *et al.,* 1970). A key component of this interpretation is the recognition of a change in depositional environments, from a shelf-type on the BHA to a deep closed basin in the CMT. A third generation of mapping was undertaken in western NH to determine the relationship between the stratigraphy of rocks on the BHA and that to the east in central NH (Chamberlain *et al.,* 1988). While the rocks themselves have remained essentially unchanged for several hundred million years, their exposures (too often too few and far between) and our interpretations of them seem to change continuously. The release of a new NH state geologic map is imminent (Lyons *et al.,* 1997), but many interesting problems persist.

Intrusive Igneous Rocks

Billings' state map (1956) also showed several suites of intrusive igneous rocks in this region (Figure 1), including the Oliverian magma series and the New Hampshire Plutonic Series (NHPS), including the so-called Kinsman Quartz Monzonite (KQM), Bethlehem Gneiss, Spaulding Group and Concord Group. At that time, the details of age relationships, the sources of these magmas, and their modes of emplacement were uncertain. Since then, the application of radiometric dating techniques, petrologic and isotopic studies, geophysical and structural investigations, and thermal modelling have vastly improved our understanding of the Oliverian and NHPS rocks.

The Oliverian suite is comprised of granites, tonalites and trondjhemites ranging in age from 442–456 Ma (Naylor, 1969; Zartman & Leo, 1985; Aleinikoff *c.f.* Lyons *et al.,* 1996) which form the core rocks of the *en echelon* mantled gneiss domes making up the BHA, as well as a few discordant intrusions just off the axis of the BHA to the west. These rocks have an internal gneissic foliation paralleling their contacts with wall rocks, giving rise to their (sometimes asymmetric) domal structures. Structures within the mantling strata, including Silurian and Devonian rocks substantially younger than the core gneisses, also exhibit this doming. In addition, metamorphic isograds in these strata commonly wrap around the domes, with the grade of metamorphism increases toward the center of the dome. It is thought that the domes might have been created by diapiric rise (Thompson *et al.,* 1968), resulting from a density inversion between the core materials and the mantling strata (Fletcher, 1972), during the early Devonian Acadian orogeny. Such a model has been demonstrated by scaled gravity tectonic experiments (Ramberg, 1973). Lyons and others (1996) used gravity measurements to confirm the geometry of these domes, and their mushroom-shaped models support reactivated diapiric rise of the core rocks. The metamorphic patterns might be the result of focussed heat flow due to contrasts in thermal conductivity (Allen & Chamberlain, 1989). The cores of the domes may also represent relict volcanic islands of an oceanic island arc (Hitchcock, 1883, 1890 *c.f.* Thompson *et al.,* 1968)—an interpretation which is not incompatible with diapiric rise. One school of thought suggests that the

docking of the Bronson Hill island arc with the Laurentian continent was responsible for the Taconic orogeny.

The NHPS were emplaced during and after the early Devonian Acadian orogeny. Of these, the Mount Clough pluton and the Cardigan pluton are the most voluminous. Both of these plutons are heterogeneous granitoids, ranging from granite (*sensu stricto*) to quartz diorite but being predominately granodiorite. Although compositionally very similar (Billings & Wilson, 1964), the rocks of the these plutons do differ in texture and mineralogy and have been identified as separate units, the Mount Clough pluton being comprised of Bethlehem Gneiss, and the Cardigan pluton of KQM (although no longer considered a quartz monzonite).

Characteristic of the KQM are abundant potassium feldspar megacrysts in a course groundmass of quartz, plagioclase, biotite, muscovite and garnet. In some places, garnet becomes the dominant mineral suggesting zones of restite entrained within the magma. The Bethlehem Gneiss is a more homogeneous micaceous granodiorite distinguished from the KQM by a prominent gneissic foliation and the lack of potassium feldspar megacrysts and restites.

Lathrop and others (1994, 1996) made detailed Sr, Nd and O isotopic studies of the Mount Clough and Cardigan plutons to address questions about the nature of origin and source for these rocks. Their results show that, while isotopically heterogeneous and complex, the Bethlehem Gneiss and KQM magmas formed from anatectic melting of the adjacent metasedimentary rocks found in the BHA and the CMT, with very little if any contribution from the mantle. The original sedimentary material in the CMT was enriched in heat producing elements, which when the crust was thickened during the Acadian orogeny, would have provided a sufficient heat source for melting *in situ* (Chamberlain & Sonder, 1990). However, the rocks of the BHA are not so enriched, and so would have required an external heat source or deeper burial for melting.

In addition to the Mount Clough and the Cardigan plutons, several smaller satellite plutons of Bethlehem Gneiss (or KQM) have also been mapped, for example the Bellows Falls pluton. Geophysical investigations, combined with geologic structural constraints, indicate that these plutons are floored by metasedimentary rocks and were probably emplaced as large sill-like bodies, not much thicker than about 2000 meters (Nielson *et al.,* 1976; Kruger & Linehan, 1941). It is thought that several of these plutons may have at one time been connected, covering the entire area; the Bellows Falls Pluton perhaps being an outlier of the Mount Clough Pluton, for example.

It should further be noted that the Bethlehem Gneiss and the KQM cannot be distinguished from one another on the basis of Nd, Sr, or O isotope compositions (Lathrop *et al.,* 1996), nor their major element compositions (Billings & Wilson, 1964). Thus it has been suggested that these two magmas were originated from the same parent material, with their mineralogical differences the result of different metamorphic conditions during their crystallization, or perhaps post-crystallization metamorphism (Chamberlain & Lyons, 1983). Differences in texture could be related to the degree to which the plutons were involved in nappe-stage deformation, the Mount Clough pluton of Bethlehem Gneiss being the further west on the eastern flanks of the BHA, while the Cardigan pluton is within the CMT. Allen proposed (1984, 1985) that the upper-most Fall Mountain nappe may have been aided in its transport to the west by the intrusion of Bethlehem Gneiss as a sheet along its sole, such that the Fall Mountain outlier and the Bellows Falls pluton

are a package. To the east, the Huntley Mountain Spur of the Cardigan pluton is mapped along the sole of the Fall Mountain nappe (Chamberlain *et al.,* 1988).

The Spaulding Group rocks, also syn-orogenic, and the post-orogenic Concord Group, may be associated with migmatite zones (Allen & Chamberlain, 1992; Allen, 1994, 1996a, b) found extensively along the axis of the Central NH Anticlinorium (Eusden *et al.,* 1987; Eusden & Lyons, 1993). Thermal relaxation following the crustal thickening of the Acadian orogeny could have lead to melting of CMT rocks to produce the post-orogenic Concord Group plutons (Chamberlain & England, 1985).

Structure and Metamorphism

Thompson and others (1968) mapped a series of several west-vergent isoclinal fold nappes lying on and lapping over the BHA. These were recognized on the basis of repeats and inversions of the stratigraphy as well as structural features. The nappes were then deformed and backfolded by the subsequent diapiric-rise of the Oliverian domes (Robinson & Hall, 1979), greatly complicating the structure of the region. Further to the south, thrust faulting was recognized as an intermediate stage between emplacement of the fold nappes and subsequent doming (Thompson *et al.,* 1987). Clearly, not all of the nappes are strictly fold nappes, the upper-most Fall Mountain nappe probably being more of a thrust structure (Allen, 1984, 1985, Chamberlain *et al.,* 1988), with intimate involvement of rocks from the NHPS, as discussed above. On the eastern flanks of the CMT, to the east of the Central NH Anticlinorium, a somewhat similar set of early nappes are recognized, there being east-vergent (Eusden *et al.,* 1987). The Central NH Anticlinorium is interpreted as a "dorsal zone" or "pop-up" structure forming the core of the Acadian orogeny (Eusden *et al.,* 1987; Eusden & Lyons, 1993).

The succession of nappes emplaced over the BHA explained the inversion of metamorphic isograds observed in the region (Chapman, 1953). Each nappe, rooted to the east of the BHA in the CMT, brought deeper, hotter rocks westward with it, over colder previous nappes and autochthonous rocks. Thus rocks of high metamorphic grade were correlated to high tectonic level, in this case later nappes (Thompson *et al.,* 1968). Spear and others (1990, 1992; Kohn *et al.,* 1992) have correlated this scenario with pressure-temperature-time (P-T-t) histories of metamorphism found in rocks from the various nappes, wherein the nappe or thrust emplacement of hot rocks drives metamorphism of underlying colder rocks as the nappe cools. Indeed, they have gone on to postulate the existence of additional nappes (or thrust sheets), that are not exposed anywhere, on the basis of P-T-t histories. Certainly, within the CMT east of the BHA, the metamorphic pattern shows localized regions of high- or low-grade metamorphism (Chamberlain & Lyons, 1983). Chamberlain (1986) has explained these as the result of metamorphism enhanced by sequential episodes of intersecting folds.

Conclusion

The geology seen here in west-central NH appears to be characteristic of many orogenic belts. Thompson and others (1968) drew analogies between their interpretation of nappes and gneiss domes with similar structures observed in the western Alps, as further developed by Thompson and others (1990, 1993). Mantled gneiss domes are present in many deeply eroded mountain belts,

including the Alps, Caledonides, Canadian Cordillera and the Appalachians (Eskola, 1949; Allen & Chamberlain, 1989), as well as the Himalaya and Karakorum (Allen & Chamberlain, 1991; Searle *et al.,* 1989; Kundig, 1989). Perhaps their previous experience in west-central NH influenced their thinking, but Allen & Chamberlain (1991) recognized in the Karakorum of Pakistan a very similar scenario to that outlined here: fold-and-thrust nappes inverting the crust and thus metamorphic isograds, with subsequent deformation by gneiss domes of re-mobilized earlier granitic intrusives (Allen, Smith, Zeitler & Chamberlain, unpublished data).

Acknowledgements

I would like to express my sincere appreciation to C. Page Chamberlain, James B. Thompson, Jr., and John B. Lyons for first showing me these outcrops, and introducing me to and involving me in the study of the geology of my home stomping grounds. Any errors or omissions are mine alone. I'm sure that I have overlooked some important references, and haven't adequately represented all of those cited.

Road Log

We will depart at 8:00 AM from the intersection of Routes 12 and 12A (Maple Avenue) in northwest Keene, NH. Please carpool, bring a lunch, and have a full tank of gas—enough for at least 150 miles of driving. The best reference map I can suggest would be a NH state highway map—we're going to cover a lot of ground!

Mileage

0.0 Set trip meter to zero at the Rt. 12 overpass over Rt. 12A.
0.7 Stoplight, continue on Rt. 12A towards Surry.
4.6 roadside outcrops of Oliverian granite in the Keene Dome, view of Surry Mountain (see stops 1 & 2 below) to the east.
7.0 Park on the right side, as far over as possible. Leave hazard indicators flashing.

STOP 1: Border Gneiss of the Keene Dome (10 minutes) The Oliverian rocks of the Keene Dome are non-peraluminous granodiorite to quartz monzonite, here strongly laminated in the border region of the dome, adjacent to Surry Mountain—a septum of metasediments between the Keene and Alstead domes, in a vertical syncline that folds the axial plane of the nappes. Within the border gneiss, one finds thin mafic layers and quartz "eyes." Return to the vehicles and continue north on 12A.

7.2 Turn right onto River or Gilsum road
7.4 Park on the right side, as far over as possible. Leave hazard indicators flashing.

STOP 2: Surry Mountain Septum (10 minutes) Exposed here are some of the metavolcanic and metasedimentary rocks of the Surry Mountain septum (see stop 1 above). We may find gedrite-cordierite assemblages in a "garbenschiefer" texture (better exposed up on Surry Mountain); the rock is also reported to contain Na-micas with talc intergrowths (J. B. Thompson, Jr., pers. comm., 1992). Return to the vehicles and continue east.

11.4 Stop sign across stone arch bridge, turn left onto Rt. 10 north through Gilsum

15.0 Park on the right side, as far over as possible. Leave hazard indicators flashing. Take care in crossing the highway.

STOP 3: High-Grade Rangeley Formation (15 minutes) This is Stop 9 of Chamberlain and others (1988) and Stop 10 of Spear (1992). The metamorphic mineral assemblage is garnet + cordierite + biotite + K-feldspar + sillimanite + plagioclase, and represents a localized metamorphic "hot spot." Chamberlain and others (1988) mapped these rocks as part of the root zone for the Fall Mountain nappe, and Chamberlain (1986) interpreted the high grade of metamorphism to be the result of metamorphism enhanced by sequential episodes of intersecting folds (e.g. syncline + syncline). Spear (1992) mapped these rocks as the upper plate of the Chesham Pond nappe, and interprets the P-T-t path as cooling following thrust emplacement and backsliding. Return to the vehicles and continue north on 10.

18.6 Turn left onto Rt. 123 west at Marlow
19.9 Park on the right side, as far over as possible. Leave hazard indicators flashing.

STOP 4: Huntley Mountain Spur (10 minutes) This is a spur of KQM extending southwest from the Cardigan pluton, mapped as representing the sole of the Fall Mountain nappe (Chamberlain *et al.*, 1988). The rocks are highly foliated. Continue west on 123.

23.1 Flashing yellow light, East Alstead. Continue straight through on 123 west
24.3 Mill Hollow. Continue through on 123 west
27.2 Rt. 12A comes in from the left, continue on 123 west
28.0 Rt. 123A comes in from the right, continue on 123 & 12A through Alstead village
28.6 Rts. 12A and 123 diverge, turn left staying on 123 west
30.6 Stop sign, Rt. 123 turns left, continue straight across onto Cold River Road
31.2 (optional stop) outcrops in the river are of a window to the floor of the Bellows Falls pluton
31.9 outcrop on far bank of the river are of the Bellows Falls pluton
32.5 another river outcrop of the floor under the Bellows Falls pluton, Fall Mountain looms to the right.
33.3 Stop sign, intersection with Rt. 123, bear right on 123 west
33.5 Stop sign, intersection with Rt. 12, turn right onto 12 north
34.5 Stop lights, continue through to just past the railroad yard
34.6 Park on the right side, as far over as possible. Leave hazard indicators flashing. Take care in crossing the highway and scrambling down to the outcrops in the river.

STOP 5: Fall Mountain Nappe and Bellows Falls Pluton (20 minutes) This is Stop 1 of Chamberlain and others (1988) and Stop 5 of Spear (1992). Gray schists of the Rangeley Formation from the CMT, containing characteristic calc-silicate and amphibolite "pods," is here in contact with the Bethlehem Gneiss of the Bellows Falls pluton. Note the sillimanite pseudomorphs after andalusite ("turkey tracks") in the schist, and garnets in the Bethlehem Gneiss associated with assimilated blocks of schist. The CMT rocks on Fall Mountain were clearly of higher metamorphic grade than the BHA rocks surrounding and underneath the Bellows Falls pluton, which implies significant horizontal transport from the root zone in Marlow. Much of this transport and shortening may have been taken up within the Bellows Falls pluton (Allen, 1984, 1985; Spear, 1992). The metamorphic evolution of Fall Mountain shows cooling following emplacement of hot rocks over cold (Allen, 1984); Spear (1992) develops a more complex scenario. The hike up Fall Mountain (Mount Kilburn) via the tower access road is interesting (Stop 2 of Chamberlain *et al.*, 1988). Return to the vehicles and continue north on Rt. 12.

35.1 Stop light, turn left and cross the bridge to Vermont
35.3 Stop sign at end of bridge, turn left onto Rt. 5 south
35.6 Downtown Bellows Falls, continue south on 5
35.9 Stoplight, bear right uphill on Rt. 121 west
36.9 Turn left around monument onto Gage St.
37.1 Continue straight ahead onto dirt road
37.3 Park on the right or left sides, as far over as possible.

STOP 6: Littleton Formation in the Saxton's River (20 minutes) This is Stop IV-2 of Zen (1989) and Stop 3 of Spear (1992). ***No hammering, please.*** The rocks here are a staurolite schist and quartzite showing spectacular graded bedding (grade reversed due to metamorphism) characteristic of a turbidite deposit. The rocks are isoclinally folded and otherwise deformed, with inverted topping directions and several hinges apparent. Much of the deformation may well have occurred during deposition, when these rocks were still soft sediment. These rocks are part of the Skitchewaug nappe, structurally below the Bellows Falls pluton and the Fall Mountain nappe. P-T paths from other outcrops in this section show heating with loading (Spear, 1992), consistent with the emplacement of the hot Fall Mountain rocks over these colder rocks. After getting your photographs, return to the vehicles and continue along the dirt road.

37.9 Rejoin Rt. 121 headed back into town
38.4 Stoplight at bottom of hill, turn 90° to the left onto Rt. 5 North (to I-91)—*not* back through downtown
39.2 Stop sign, continue north on 5
42.0 Rt. 5 turns off to the right, Rt. 103 ahead to the left—go straight up the on-ramp to I-91 north
46.2 first view of Mount Ascutney, a remnant volcano associated with the White Mountain Magma Series, straight ahead For more information, see Schneiderman (1988).
48.6 Exit at Springfield
48.9 Stop sign at end of ramp, turn left to go north on Rt. 5 & east on Rt. 11
49.3 Turn left onto Rt. 5 north, just before the toll bridge
50.6 Park in the fork on the right.

STOP 7: Fossiliferous Clough Quartzite of the Skitchewaug Nappe (10 minutes) This is Stop IV-1 of Zen (1989). ***No hammering, please.*** These rocks have been highly deformed, and metamorphosed to the staurolite zone, and yet calcareous bands within the quartzite contain recognizable fossils and fossil fragments, including crinoid stems, corals, and brachiopods (Thompson, 1954; Boucot & Thompson, 1963). Similar fossils have been found in several other locations within the Clough Quartzite throughout the region, such that it is a good Silurian marker and very useful in deciphering the stratigraphy and structure of the BHA. Continue north on 5.

54.7 Rt. 143 comes in from the left, continue north on 5
60.3 Yield sign, turn right towards Claremont
61.2 Stoplight at intersection with Rt. 12A, continue straight towards Claremont
63.6 Turn left, following sign for Newport and Sunapee (*not* downtown)
64.2 Stoplight, continue straight through
64.6 Park in turnout to the right, just before the stoplight at intersection with Rt. 120. This may well be our Lunch Stop. Take care in crossing the highway to roadcuts at the intersection.

STOP 8: Blue quartz eyes in pyroclastic volcanics (15 minutes) The blue quartz eyes, from the exsolution of very fine rutile crystals within the quartz, are indicative of very high temperatures, but these rocks have only been metamorphosed to biotite zone (low grade). This rock represents a high-temperature volcanic ignimbrite, with lenticles of pumice, lapilli, etc...; probably equivalent to

the Post Pond Volcanics in Vermont. The coarse pyroclastic sequence is best visible on a water polished surface (cross back to the riverside park). A slaty cleavage has been superposed, creating a lineation in the rock. Look down-dip to see these features. (J. B. Thompson, Jr., pers.comm., 1992) Return to vehicles and continue towards Washington Street.

64.8 Stoplight, turn left onto Rts. 11 & 103 east (Washington Street, Claremont). Continue through all the stoplights along this strip.
71.7 outcrops of Bethlehem Gneiss
73.7 Stoplight, go straight
74.1 Stoplight at intersection with Rt. 10 (Main Street, Newport), turn left to follow Rts. 11 & 103 east
74.2 Turn right following sign for Rts. 11 & 103 east
74.4 Yield sign, continue onto Rts. 11 & 103 east
77.3 Rts. 11 & 103 diverge (Guild), bear right to stay on Rt. 11 east toward Sunapee
79.7 Rt. 103B on the right, continue straight
79.8 Flashing yellow for Sunapee Harbor, continue straight
80.0 Turn left onto Sargent Road, park on dirt track by Sunapee Fire Station. Take care in crossing the highway.

STOP 9: Spaulding Quartz Diorite and pegmatite autoliths (15 minutes): The Spaulding, dated at 392 million years, somewhat foliated, cross-cuts the KQM and the Bethlehem Gneiss, thus a late tectonic intrusive. Here it is cut by pink pegmatites (with perthitic feldspar). The pegmatite and the Spaulding are magmatically related, and here you can see "autoliths" of the pegmatite within the Spaulding and vice-versa. At the north end of the outcrop are true xenoliths (as opposed to autoliths) of the country rock, including the KQM. In some places the KQM has become disaggregated, with incorporation of feldspar xenocrysts into the Spaulding. This represents an igneous breccia, with flow foliation, resulting from explosion venting of fluidized magma. (J. B. Lyons, pers.. comm., 1988). Return to the vehicles and enter Rt. 11 heading west from the fire station.

80.3 Flashing yellow for Sunapee Harbor, continue straight
80.4 Turn left onto Rt. 103B south
81.0 small quarry operation, with antique steam tractor on left
82.8 View of Mount Sunapee Ski Area ahead through trees.

The ridge extending south-southwest from Mount Sunapee to Lempster Mountain, and north from Sunapee to Mount Cardigan, represents a septum of metasediments (interdigitated with the Huntley Mountain Spur) caught between the Cardigan pluton to the east and the Mount Clough pluton to the west. Mapped by Dean (1976) based on the BHA stratigraphy, a new look at these rocks with insights from the CMT and other areas, may help tie together the stratigraphy and structure of southwestern NH with that to the north and west, as well as furthering our understanding of the relationships between the Cardigan and Mount Clough plutons.

83.8 Stop sign, proceed across and around the Mount Sunapee Traffic Circle, leaving on Rt. 103 east.
84.7 outcrops of KQM injected with multiple dikes of aplite
85.1 outcrops of KQM with rafts of metasediment
86.2 Newbury Harbor
86.3 Intersection with Rt. 103A

Alternate Route (shortcut):
86.3 Turn left onto R.t 103A north

88.7 Pull off on the LEFT side of the road to park just before Ramblewood Circle. Leave hazard indicators flashing, and take care in crossing the highway. Across the road is an interesting outcrop of the KQM.
89.2 Rt. 103A bears to the left, continue following it to the north.
91.3 Turn right onto King Hill Road, towards I-89 southbound
92.3 Bog Road takes off the left, continue on King Hill Road
92.5 The author's childhood home is on the left
93.0 View of Mount Kearsarge ahead (see comments at mile 99.0).
93.7 Under the highway, turn left on to entrance ramp for I-89 northbound. Pick up the road log at mile 109.7, and subtract about 16 miles from subsequent mileage readings.

Original Route (the long way):
86.3 Rt. 103A goes off to the left, continue east on Rt. 103
86.8 more outcrops of KQM
87.1 Mountain Road on the right, continue on 103 east. K-Ar dating of kaolinite gouge from the Newbury Fault zone exposed here yielded an age of 160 Ma (J. B. Lyons, pers. comm., 1988)
89.1 Flashing yellow, South Newbury
91.4 Park on right side of road, as far over as possible. Leave hazard indicators flashing. Take care in crossing the highway.

STOP 10: Restites in the Kinsman Quartz Monzonite (15 minutes) This is Stop 10 of Lyons (1988) and Stop IV-5 of Zen (1989) Garnetite (garnet-cordierite-biotite-plagiogclase-quartz) pods and layers within the KQM represent restites of xenoliths of Littleton schist undergoing anatexis, anatectic melting of CMT and BHA metasediments being the source of the magma making up the pluton (Lathrop *et al.,* 1994, 1996). The KQM is here a granodiorite with the characteristic K-feldspar megacrysts and a mineral assemblage including sillimanite (fibrolite) as well as garnet and cordierite The Cardigan pluton is at its maximum preserved thickness of 2.5 km here, based on gravity studies (Nielson *et al.,* 1976; Lyons, 1988). The Cardigan pluton to the south of here was briefly considered as a potential host for the disposal of high-level radioactive wastes (Department of Energy, 1986, p. 3-435). Return to the vehicles and continue east on 103.

91.7 Stoplight at intersection with Main Street, Bradford. Continue straight on Rt. 103 east
92.4 Stoplight at intersection with Rt. 114, continue straight on Rt. 103 east to Warner
98.7 Park on right just before onramp to I-89 south, as far over as possible. Leave hazard indicators flashing. Take care in crossing the highway.

STOP 11: Kinsman Quartz Monzonite with garnets (15 minutes) This is Stop IV-6 of Zen (1989). Here were are near the eastern edge of the Cardigan pluton, in a garnetiferous phase of KQM, with garnet rimmed or replaced by biotite, and K-feldspar megacrysts partly sericitized and mantled by myrmekite. The Cardigan pluton does not appear to have been initially homogeneous melt (Clark & Lyons, 1986; Lathrop *et al.,* 1994, 1996). The occurrence of garnet within the pluton has been interpreted as both metamorphic (Chamberlain & Lyons, 1983) and igneous (Clark & Lyons, 1986), with evidence now favoring igneous garnet (Lyons, 1988). Return to vehicles and continue east on 103 under the interstate.

99.0 Turn left on to ramp for I-89 north, and take in the view of Mount Kearsarge

Mount Kearsarge (Lyons, 1988), Mount Monadnock to the south (P. J. Thompson, 1988) and Mount Washington to the north (Eusden *et al.,* 1996a, b), are all composed of highly erosion-resistant members of the Littleton formation, noted on each summit for the occurrence of "turkey

track" sillimanite pseudomorphs after andalusite and dramatic cyclic graded bedding of quartzites and schists. These peaks essentially mark the trace of the Kearsarge-Central Maine synclinorium.

105.9 View of Mount Kearsarge over right shoulder (exit 10, continue N on I-89)
109.7 outcrop of abundant pegmatite (exit 11, continue N on I-89)
111.0 Slow down and look out the vehicle to the left.

NONSTOP A: "Mount Radon," New London Over the years since the interstate was built, weathering had destabilized the high roadcuts ahead, at NONSTOP B. Several years ago, the NH Department of Transportation undertook a blasting program to re-stabilize this roadcut. It was discovered after the roadcut was first opened that the rock was highly radioactive, as the joint surfaces were covered with ore-grade Uranium deposits. Most of this "ore" had been used as fill for the roadbed. During the re-stabilization effort, great care was taken to keep the dust down and the material removed from the cut was landfilled here in the median strip. This location was chosen in part because it was hoped that any Uranium leached from the rock would be trapped in the highly-reducing environment of the organic sediments of the Messer Pond - Clark Pond bog system down gradient (off to the right; we pass over "Bog Road" ahead). However, much of the Uranium had already been leached from these rocks due to the lowering of the water table resulting from the initial road cut (Dick Lane & Gene Boudette, oral comm., 1995). The author's childhood home is about 3/4 of a mile through the woods to the left!

111.3 overpass over Bog Road
113.0 Slow down and look out the vehicle to the right.

NONSTOP B: Sunapee pluton, New London Note that when the roadcuts were re-stabilized several years ago, the entire height of the cut was blasted in a single lift, for which special permission was needed from the Federal Highway Administration (Dick Lane & Gene Boudette, oral comm., 1995). The rocks are two-mica granite of the Concord Group. This outcrop is famous for the heavy coating of joint faces with Uranium phosphate minerals. This mineralization is presumably represents late hydro-thermal mobilization, and is focussed on fractured zones—it has now largely been weathered away. Zircons from this rock show little evidence of inheritance, but Uranium loss, with an age of crystallization of 354±5 Ma (Harrison *et al.*, 1987), about the right timing to be the result of anatectic melting following thermal relaxation after tectonic loading (Chamberlain & England, 1985). The pluton intrudes both the Mount Clough pluton and the Cardigan pluton, cutting off the Sunapee Septum of metasediments. As a young boy, the author used to accompany his father who would practice rock-climbing on these roadcuts before the Interstate was officially opened.

114.3 outcrop of Bethlehem Gneiss
115.6 Take Exit 12A for Georges Mills
115.8 Stop sign at end of ramp, turn left under highway
116.0 Turn right onto Stoney Brook Road
116.5 Park on left, just under the highway.

STOP 12: Bethlehem Gneiss of the Mount Clough pluton (10 minutes): A syn-kinematic intrusion inside the Skitchewaug Nappe, and/or below the Fall Mountain Nappe. Granite to tonalite in composition, although highly variable. The rock has a foliation that may be due to flow, and may also be related to its interactions with the nappes. Continue northwest on Stoney Brook Road.

118.3 View ahead of the white ledges over Grantham

These ledges, well-viewed from the interstate, are outcrops of the Clough Quartzite in the mantle of the Croydon dome. Samples from these outcrops contain well preserved brachiopod and other fossils (J.B. Thompson, Jr., pers. comm., 1984). Contained within the boundaries of the private Blue Mountain hunting reserve, they are now essentially inaccessible.

120.9 Stop sign, bear to the left on the Grantham–Springfield Road
122.0 Stop sign, turn right onto Rt. 10 north
122.6 Pass under I-89, continue straight ahead on Old Route Ten
122.9 View to the left of the gray ledges, opposite the entrance to the Eastman development
125.1 Turn left onto Miller Pond Road, under the highway
125.2 Park just past the underpass in class VI road.

STOP 13: Mantle of the Croydon Dome (10 minutes) The rocks here are Ammonoosuc Volcanics exposed at the north end of the Croydon dome, rich in sulfur (pyrrhotite, sphalerite has been found within the Blue Mountain preserve). The core rocks of the Croydon dome, exposed in interstate road cuts a few hundred feet to the south are a true trondhjemite—a high-sodium granitoid, completely lacking in K-feldspar—whatever K there is is tied up in biotite and/or hornblende (J.B. Lyons, pers. comm., 1988). Backtrack to Old Route Ten.

125.3 Turn left to continue north on Old Route Ten
127.7 Turn right and pass under highway
127.9 Turn left up on ramp to I-89 north
128.1 outcrops on the right of Bethlehem Gneiss
128.4 outcrops on the right of Bethlehem Gneiss
128.7 outcrop on the right of the core of the Skitchewaug Nappe
130.0 Exit off I-89 at Montcalm
130.2 Stop sign at end of ramp, turn left under highway
130.3 T-intersection, turn left (now heading south)
131.4 Park in turnaround at dead end by the Humane Society

STOP 14: Skitchewaug Nappe (30 minutes). This is Stop 1 of J.B. Thompson, Jr. (1988), and your are referred to that source for detailed information about these rocks. We'll make a traverse along the bike path, looking at outcrops of the BHA stratigraphic sequence (Figure 3). The outcrops of Bethlehem Gneiss where we got onto I-89 at Exit 14 are an outlier of the Mount Clough pluton, separated by the Grantham Fault. It appears to be folded into the core of the Skitchewaug Nappe. With a sharp eye, fossils can be observed in the glacially polished surface at the top of northernmost outcrop of the Clough (Stop 14-A in Figure 3). *No hammering please!* Return to the vehicles and head back the way we came.

132.6 Turn right to pass under the highway
132.7 Turn left on to the ramp to I-89 north, park on right side (as far over as possible) before merging with highway. Leave hazard indicators flashing.

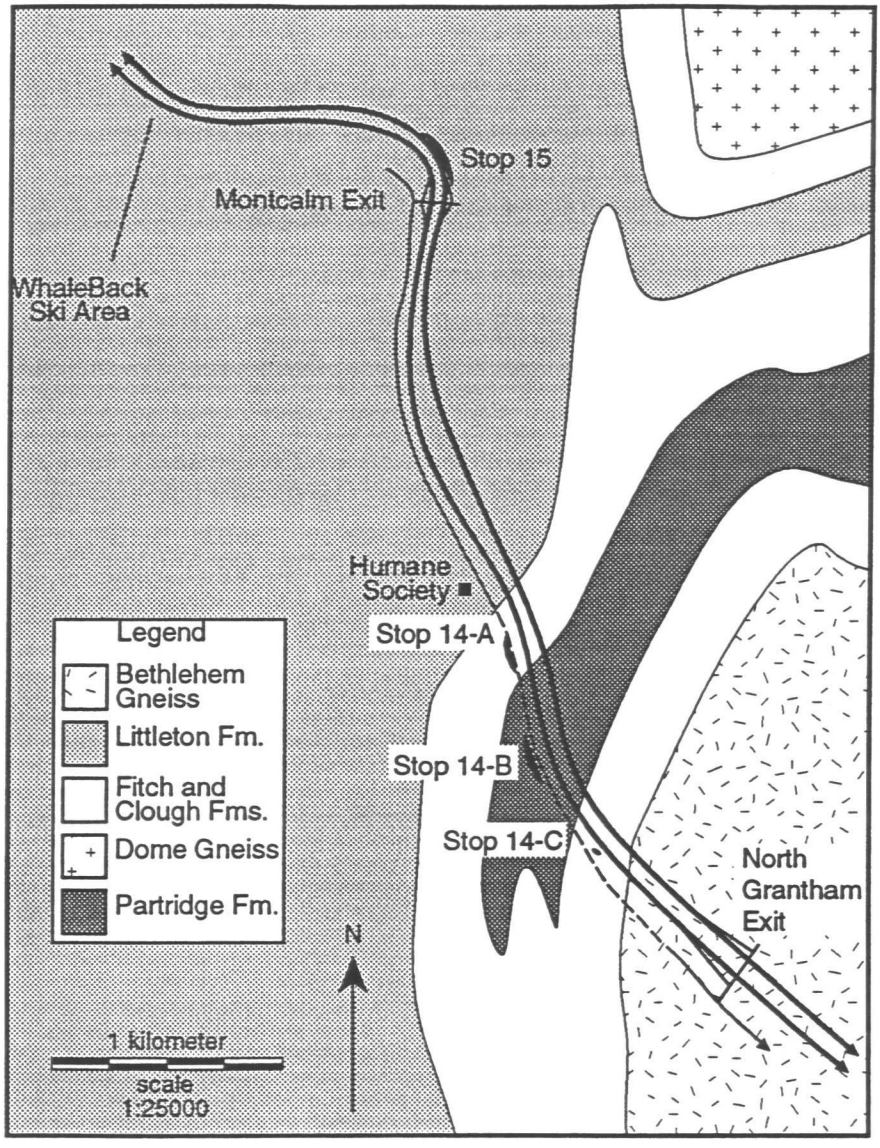


Figure 3: Geology of the Skitchewaug Nappe between North Grantham and Montcalm, after J. B. Thompson, Jr. (1988), showing the locations of Stops 14 and 15 of this field trip.

STOP 15: Littleton Formation (10 minutes) This is Stop 2 of J. B. Thompson, Jr. (1988). A thick section of Littleton schist (Figure 3), rich with staurolite crystals—some with cruciform twins—giving the rock a lumpy appearance. Few of the crystals are very fresh, most showing incipient alteration. Bedding is obscure, although the rock has good foliation and a dome-stage mineral lineation. Continue onto I-89 northbound.

136.0 If planning to visit optional stop 16, exit off I-89 at Rt. 4 east for Enfield, otherwise continue on I-89 northbound.

STOP 16: (optional) Mantling Strata of the Mascoma Dome on Moose Mountain

136.3 Yield sign at end of ramp, bear right onto Rt. 4 east

137.9 Intersection with Rt. 4A, continue straight ahead on 4 east across the Mascoma River

138.9 Turn left onto Rudsboro Road

139.8 Turn around and park in turnout on left

This is Stop 6 of J. B. Thompson, Jr. (1988). We will hike uphill through the woods to fields, across the fields, through a sugarbush and more woods to open ledges along the ridge top. Along the way, we will cross over gray schists of the basal Littleton and calc-silicates of the Fitch. Fossils have been found here in the Clough. The ledges are pink quartzite, and from them we can look out to the east over a steep drop down to the Oliverian granite of the Mascoma dome. Return back down to the vehicles, and return down Rudsboro Road towards Mascoma Lake.

140.7 Stop Sign, turn right onto Rt. 4 west
141.8 Rt. 4A again, continue of Rt. 4 west
143.2 Bear right on to ramp to I-89 northbound; rejoin field trip route and add about 7 miles to subsequent mileage readings.

Original Route:
138.2 Exit off I-89 for Lebanon & Hanover
138.5 Yield sign at end of ramp, bear right onto Rt. 120 north
138.7 Stoplight, continue straight on 120 north
139.1 Stoplight, continue straight on 120 north
140.0 Park on right side of road, as far off as possible, or just ahead at the tire dealer. If on Rt. 120, leave hazard indicators flashing and beware of traffic.

STOP 17: Oliverian Granite of the Lebanon Pluton (10 minutes) The Lebanon pluton is a discordant pluton of Oliverian granite in the core with an envelope of granodiorite to tonalite. It is found west off the main axis of the BHA, intruded into Partridge Volcanics. Structural and gravity studies show that it is an inverted cone shape plunging to the north-northwest, deforming the Cornish nappe (Lyons *et al.,* 1996). Gravity anomalies suggest that the pluton must be underlain by a sizable body of subjacent metagabbro—represented by three small bodies of metagabbro which outcrop along the western margin of the pluton. Metamorphism in the adjacent Partridge Formation has been attributed to contact metamorphism associated with original magma intrusion (Spear *et al.,* 1984); focussed heat flow due to contrasting thermal conductivities (Allen & Chamberlain, 1989); and upward domal transport of deeper, hotter rocks (Lyons *et al.,* 1996; Allen & Chamberlain, 1989). Here, the Lebanon granite includes mafic enclaves and pyrite. Once quarried for building stones, the pyrite lead to rust stains on the resulting blocks, and quarrying was abandoned. Return to the vehicles and continue north on 120.

140.5 Stoplight at top of hill (entrance to DHMC), continue straight on 120 north
141.4 Stoplight (entrance to DHMC and Jesse's), continue straight on 120 north
141.8 Stoplight, bear to the left staying on 120 into Hanover
142.6 Stoplight at the Hanover Food Coop, bear to the right staying on 120 north
143.1 Stoplight, turn left onto Rt. 10 south (East Wheelock St)
143.2 Turn right up Dartmouth driveway
143.4 Dartmouth Observatory
143.5 Park wherever there is a spot. Climb down to outcrops alongside building housing Dartmouth College's Department of Earth Sciences and the Kresge Physical Science Library.

STOP 18: Metavolcanics in mantle of Lebanon Dome— a green chlorite-amphibolite, with well-developed foliation, apparently meta-basalt.

End of Trip

To get to Pico from here, go down and around the Dartmouth Green to the head of Main Street. Turn right onto West Wheelock Street and follow this across the Ledyard Bridge to Norwich. Take I-91 south to I-89 north to Route 4 west. It is approximately 45 miles.

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Middle Ordovician Section at Crown Point Peninsula

Charlotte Mehrtens, University of Vermont
Bruce Selleck, Colgate University

Location

All the stops for this trip are located within the Crown Point Reservation State Historic Site. From the west, take NY Route 22 north from Ticonderoga, continuing north through the Village of Crown Point. Turn east approximately five miles north of the Village of Crown Point, following signs to the 'Bridge to Vermont'. From the east, take VT Route 22A north from Fairhaven, or south from Burlington area, and follow signs to 'Bridge to New York'. Stop locations are keyed to the aerial photo. Stop 1 is in the ditch and wall of a small outpost fort (the Redoubt) on the east side of Route 8, immediately across the highway from the historic site entrance road. We will enter the historic site and park near the main entrance, then walk back to stop 1. No collecting is permitted in the historic area.

Chazy Group

The internal stratigraphy of the Chazy Group in the Lake Champlain Valley was defined by Oxley and Kay (1959). Welby (1962) provides a comprehesive synopsis of stratigraphic relationships between exposures in New York and Vermont. Hoffman (1963) presented a regional lithostratigraphic framework for the Montreal-Ottawa area. Fisher (1968) presents descriptions of Chazy Group strata in the northern Lake Champlain Valley. Biostratigraphic correlations of the Chazy Group have been attempted using brachiopods (Cooper, 1956), trilobites (Shaw 1968) and conodonts (Raring, 1973). Speyer and Selleck (1986) summarize major lithostratigraphic trends in the Chazy Group and the appropriateness of biostratigraphic correlation.

The Chazy Group is nearly 250 meters thick in the northern Champlain Valley (Oxley and Kay, 1959) but thins to the south (approximately 90 meters at Crown Point peninsula; less than 15 meters at Ticonderoga, absent at Whitehall). In the type area in the northern Champlain Valley, a three-part subidivision into Crown Point, Day Point and Valcour Formations is applied. The classic biostromes largely occur within the Day Point Formation. Significant biostromal buildups are absent in the southern exposures. In the vicinity of Crown Point, the entire Chazy Group is placed within the Crown Point Formation (Oxley and Kay, 1959), although there is considerable lithologic variation.

The Chazy Group marks the resumption of shallow marine sedimentation following partial emergence and subaerial erosion of the underlying Beekmantown Group during early Medial Ordovician time. In the southern Lake Champlain Valley, the basal Chazy units are clearly in unconformable contact with tilted, eroded upper Beekmantown Group strata. In the northern Champlain Valley, basal Chazy Group beds are in apparently conformable contact with the Providence Island Dolostone of the Beekmantown Group (Speyer , 1982). Rapid north-to-south thickness and facies changes, and variations in the nature of the basal contact suggest that the Chazy was deposited during or just after a period of block faulting which disrupted earlier Beekmantown and older rocks in some areas. The presence of coarse, angular quartz and feldspar grains within the Chazy Group suggests that Proterozoic basement may have been exposed close to the areas of Chazy deposition (Selleck, et al, 1985). In the Ottawa area, basal

Chazy consists of coarse sandstones and pebble conglomerates of braided stream origin, indicating uplifted source terrains in that area (Hoffman, 1963).

The Chazy Group contact with the overlying Black River Group is abrupt, but evidence for significant erosion is generally absent. At Chazy, New York, the contact is apparently confomable (Fisher, 1968; Raring, 1973). The contact is well-exposed on the Crown Point Peninsula and is marked by a coarse quartz sandstone with scattered large, angular feldspar grains.

The Chazy Group at Crown Point consists largely of fossiliferous bioclastic wackestones, packstones and grainstones, with varying degrees of post-depositional dolomitization. Shaley, nodular limestones are present in the sequence, but are rarely exposed at the surface. Environments of deposition varied from subtidal, storm-dominated shelf settings to inshore sand shoals and lower tidal flats. Muddy peritidal facies are generally absent, but intervals of penecontemporaneous cementation and karstic erosion may mark intervals of subaerial exposure (Selleck, 1983).

Cements within the Chazy Group at Crown Point typically consist of an early equant to prismatic lo-Mg calcite followed by later coarse calcite spar. Dissolution and chert replacement of aragonitic bioclasts is common. Dolomitization of Crown Point carbonates is widespread, and is highly selective in some facies. Variations in primary mineralogy (lo-Mg calcite vs. aragonite) appear to have controlled the dolomitization of some bioclastic materials; grain size, sorting, porosity, intensity of burrowing and distribution of early cements (and thus permeability of materials during burial diagenesis) seem to best explain the highly variable patterns of dolomitization (Selleck, 1988).

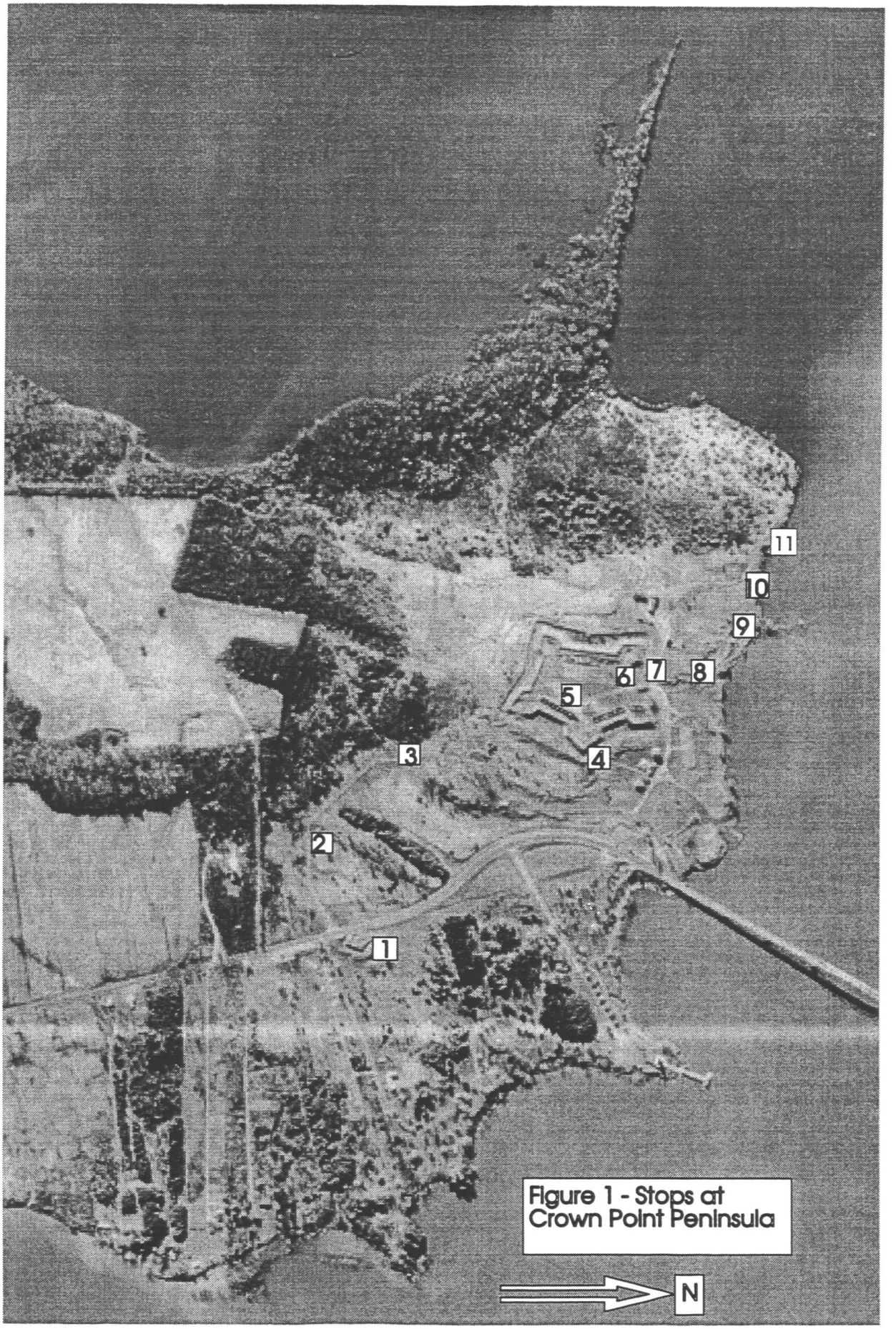


Figure 1 - Stops at Crown Point Peninsula

Stop 1- Redoubt

Approximately 6 m of variously burrowed, slightly dolomitic, thin to medium bedded bioclastic packstone and grainstone is exposed in this section. Some beds are relatively well-sorted grainstones with sharp bases and are interpreted as storm deposits. Note the large intraclasts in the base of one unit in the low ledge at the southeast corner of the ditch. These indicate rip-up of cemented grainstone, apparently by a storm. Abundant "*Girvanella*" algal oncolites are present in beds approximately three meters from the base of the section. Rounded, dark calcite grains (abraded gastropod fragments?) form the cores of the oncolites, and are scattered in other beds. Fossils are relatively abundant, and are best seen on slighly weathered bedding surfaces. Trilobite fragments, brachipods, bryozoans, pelmatozoan plates, nautiloids and large *Maclurites magnus* are present. Dolomite occurs in shaley weathering laminae and in burrow fills.

The relatively high faunal diversity, general bioturbation, and storm-related sedimentation point to a 'low energy', shallow subtidal environment at depths slightly below normal wave base. The abundant algal oncolites and discrete calcarous algal fossils (e.g. *Hedstromia*) suggest depths well within the photic zone. A possible modern analogue is found in the mixed mud and sand shelf to the west of the emergent tidal flats of Andros Island, Bahamas, as described by Bathurst (1971) and Purdy (1963).

The wavy, irregular dolomite laminae result from dolomitization of lime mud, followed by compaction and pressure solution of calcite that produced irregular, clay- and dolomite-rich stylocumulate seams. Preferential dolomitization of burrows is due to contrasts in permeability of burrow-fill versus burrow-matrix sediment. The burrow-fill material retained permeability longer during diagenesis and allowed more pervasive dolomitization. In similar facies exposed on Bullwagga Bay (west shore of peninsula), nodular limestone with shaley dolostone seams and stringers are present (photo 1). The limestone nodules appear to have been cemented prior to significant burial compaction, whereas the shaley dolostone material was compacted around the cemented limeston. The early-cemented limestones were resistant to dolomitization. This sort of fabric selective dolomitization is common in the Chazy and Black River Groups throughout the Champlain Valley. Can you find other evidence of early cementation in this outcrop. Are there hardgrounds?

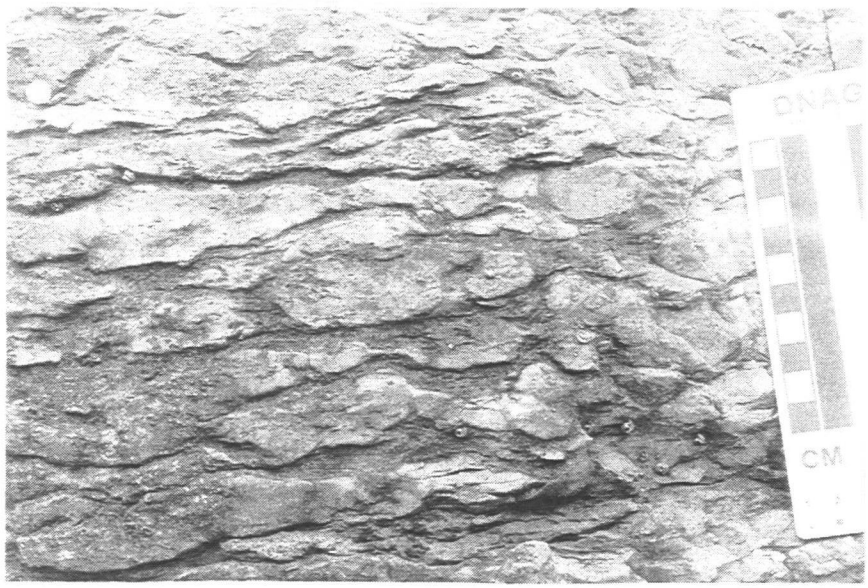


Photo 1: Nodular limestone in shaley dolostone matrix.

Stop 2 – Gate

Cross-stratified coarse bioclastic grainstones are well-exposed near the main gate along the entrance road and adjacent ridge. Foreset cross-strata show bipolar dip directions. Angular quartz and feldspar grains are concentrated in certain laminae. The carbonate particles are dominately sub-rounded, abraded pelmatozoan plates with gastropod and brachiopod fragments. Large *Maclurites* fragments and grainstone intraclasts are present on the upper bedding surfaces of the ledges east of the entrance road.

We envision the environment of deposition of this facies as a shallow subtidal wave and current reworked sand shoal. Active transport of abraded grains may have been accomplished by tidal currents (as suggested by bipolar cross-strata), or by storm-generated currents that produced complex, anastomosing patterns of cross-strata and intervening reactivation surfaces. The lack of burrowsing and well-preserved whole-shell body fossils may be due to the inhospitable shifting sand substrate. This environment likely resembled the unstable sand shoal environment of the Bahamas Platform (Bathurst, 1971; Ball, 1967). The scale and style of cross-stratification here are similar to that predicted by Ball (1967) from study of the bedfroms present on Bahamian Platform sand bodies. Similar Chazy Group facies in the northern Lake Champlain Valley are oolitic (oxley and Kay, 1959).

Note that these grainstones are essentially undolomitized. Does this indicate early cementation or diagenetic stabilization prior to deeper burial?

Stop 3. Low ledges adjacent to entrance road (Picnic Pavilion Ridge) approximately 50 meters north of Stop 2.

Brown-weathering, slightly shaley dolostone exposed here contains lenses and stringers of fossiliferous lime packstone and wackestone. As at Stop 1, trilobites, small brachiopods and *Maclurites* are common. The environment of deposition is assumed to be subtidal shelf, with less storm activity than at Stop 1.

Note that some of the fossils are almost entirely encased in dolomite, which is assumed to be of replacement origin here. Why are some of the fossils so well-preserved, and not dolomitized?

Stop 4 - SE moat of Fort Crown Point:

Approximately 3 meters of thickly laminated limestone and dolostone are exposed in the southeast 'moat' of the British Fort. The dominant facies here is alternating 0.5-2.0 cm thick laminae of limestone and dolostone – commonly termed 'ribbon rock'. The limestone ribbons are very fine peloid grainstones or 'calcisiltites' and appear blue-grey on slightly weathered surfaces, and are indentations on more deeply weathered surfaces. The dolostone ribbons weather tan-brown, and consist of an interlocking mosaic of 20-300 micron dolomite crystals of replacement origin. Quartz silt grains are present in the dolostone ribbons, versus medium to fine quartz sand in the limestone, suggesting that the limestone ribbons were slightly coarser than the dolostone when orginally deposited.

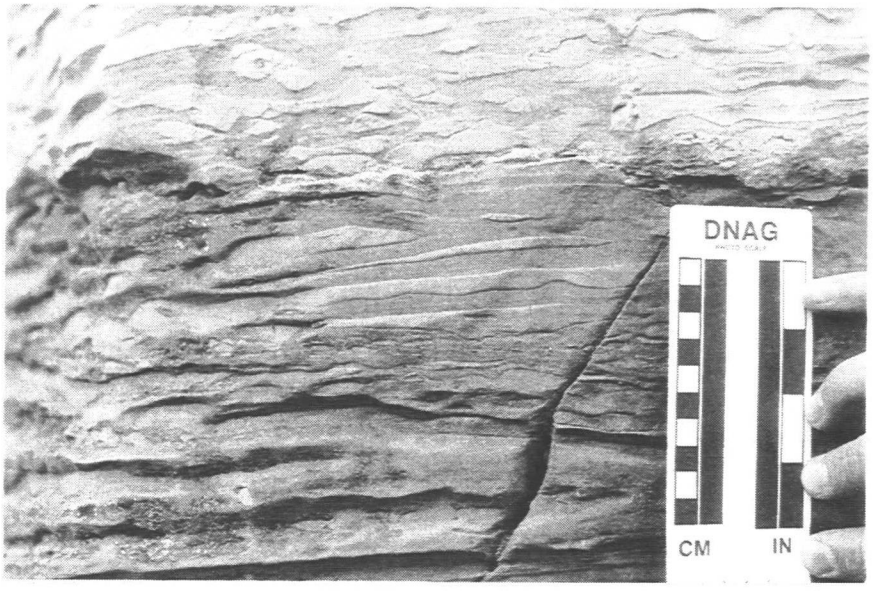


Photo 2: Typical 'ribbon' fabric. Light-weathering thick lamina are limestone; dark-weathering are dolostone.

An erosional surface with 10-20 cm relief is exposed near the base of the south wall. Similar erosional surfaces occur within this facies in other exposures, and appear to represent micro-karstic solution surfaces on a tidal rock platform that developed during subaerial exposure of cemented limestone. (See photo 3). Typically, the rock below the surface is mostly calcite limestone, suggesting that cementation and diagenetic stabilization of the limestone occurred prior to development of the erosional surfaces. Overlying rock contains more dolomite. *Maclurites* shell hash can be found in pockets on the erosional surface, suggesting wave transport of shells onto the rocky platform of the erosional surface. Dolomitized burrows cut across the limestone ribbons in some parts of the outcrop, and trough cross-strata that appear to fill low scours are also visible.

On the less weathered prominence on the SE corner of the moat, shallow scours containing brachiopods and gastropod debris are seen. Intraclasts or pseudoclasts of limestone in dolostone are also present. Some 'clasts' appear to be cored by dolomitized burrows.

We interpret this sequence as a tidal flat to shallow subtidal facies. The alternating limestone/dolostone 'ribbon rock' may represent rhythms of slightly coarser (limestone) and slightly finer (dolostone) sediment deposited on the lower reaches of a tidal flat, similar to the bedding described by Reineck and Singh (1980) from the mud/sand tidal flats of the North Sea. These coarse-fine alternations might also reflect storm-related, ebb-surge deposition. Early cementation of the slightly coarser limestone ribbons made this lithology less susceptible to later dolomitization, which affected the finer, muddy ribbons that are now dolostone. Variations in intensity of burrowing reflect subtle differences in duration of subaerial exposures of the flat and/or extent of reworking by tidal currents. Limited *in situ* faunal diversity is expected in a stressed tidal flat setting. The absence of mudcracks and evaporite minerals may indicate that only the lower portion of a humid climate tidal flat system is preserved here.

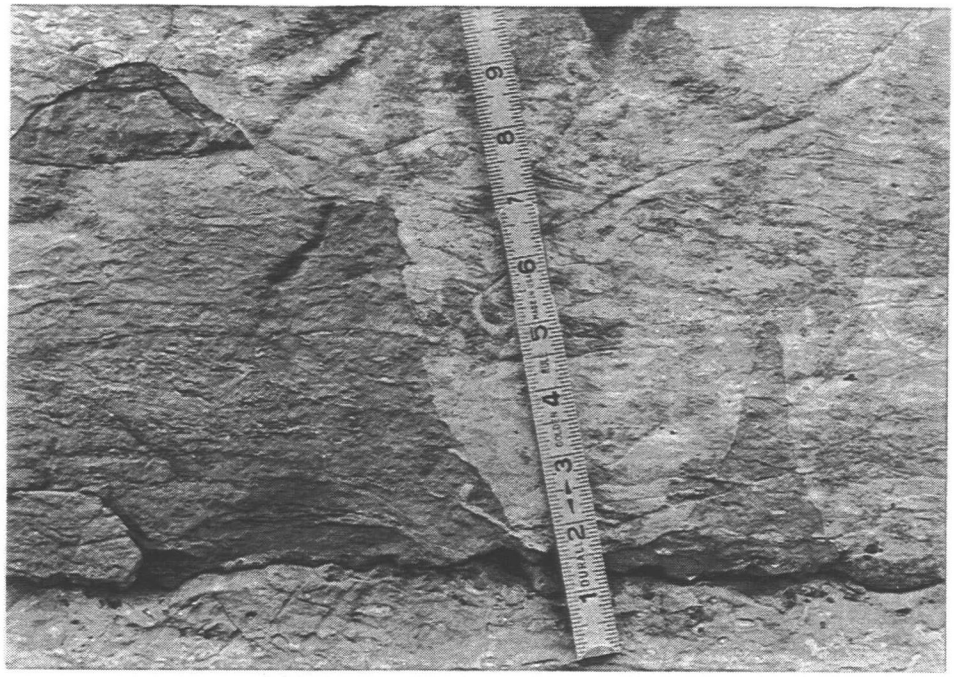


Photo 3: Limestone (dark) forming sharp, scalloped erosional surface beneath dolostone above. The limestone was cemented prior to deposition of overlying sediment. Later dolomitization altered only burrow-fill material in dark limestone, whereas sediment above was throughly dolomitized.

Stable carbon and oxygen isotope analysis of limestone and dolostone ribbons from this facies are presented in Figure 2. The oxygen isotopic signatures of the dolomite, and later calcite spar, are consistent with precipitation of these phases from relatively light waters (meteoric) at low temperatures (10-30 degrees C), or from isotopically heavier waters (seawater or saline brines) at elevated temperatures (100-200 degrees C). Preliminary fluid inclusion analyses on late coarse calcite spar cements indicate temperatures of precipitation in the range of 150-170 degrees C, with salinities 10-15 wt. % NaCl equivalent (2-3x seawater salinity). This suggests that some cementation, and perhaps dolomitization, occurred during later deep burial of these rocks.

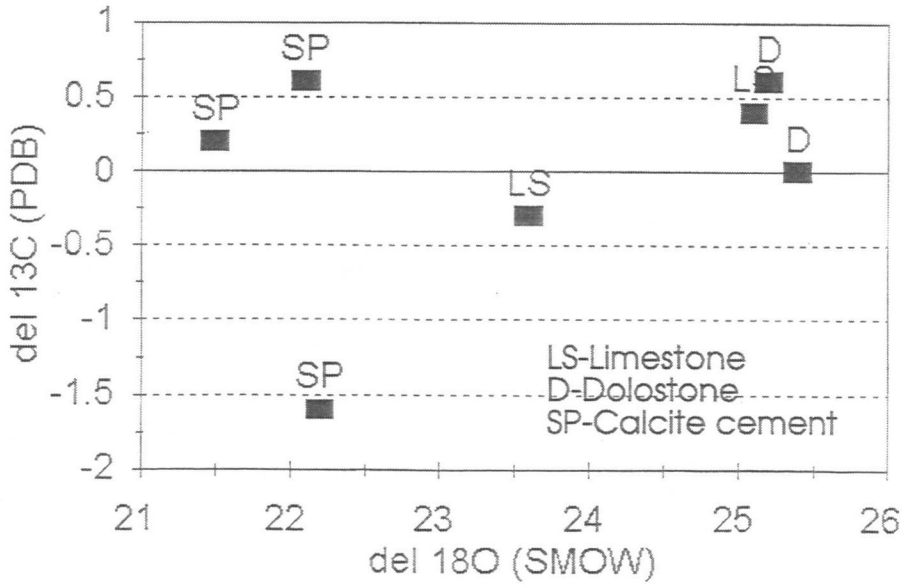


Figure 2: Carbon and oxygen isotope signatures of 'ribbon rock' carbonates

Stop 5: Parade Grounds near Barracks:

As we enter the parade ground from the southwest corner of the moat, note the array of carbonate rocks used in construction of the barracks. Chazy, Black River, and rare Trenton lithologies can be identified. Restoration of the barracks began in 1916. More recently, starting in 1976, the New York State Division for Historic Preservation undertook protection and stabilization of the ruins.

The low rock pavement just north of the barracks is within the upper part of the 'ribbon rock' unit seen at stop 4. Immediately up-section, cross-stratified grainstone beds are visible. Coarse quartz and feldspar sand is easily seen on weathered surfaces. Trough cross-strata and 'herringbone' co-sets of planar-tabular cross-strata are visible on the low vertical faces. Large, angular clasts of slightly dolomitic grainstone and *Maclurites* are present on bedding surfaces. We interpret this facies as a current-dominated sand shoal environment similar to that seen at stop 2.

Note the polygonal pattern on some outcrop surfaces. Are these mudcracks?

Black River Group

The Black River Group in the Champlain Valley is a relatively thin unit (85-90 feet thick) consisting of massively-bedded wackestones to packstones representing deposition in lagoonal to shallow subtidal environments. The gradual deepening characterizing this unit and continuing into the Trenton Group and overlying shales is interpreted to represent foreland basin subsidence during the Taconic Orogeny. This gradual deepening was punctuated by both shallowing and deepening events on macroscopic (meters) as well as microscopic (cm) scales, the latter visible only in thin section. The Black River Group is so lithologically variable in New York/Ontario/Vermont that stratigraphic names have proliferated, however the Pamelia, Lowville (House Creek and Sawyer Bay Members) and Chaumont Formations can be recognized in the Champlain Valley. Bechtel (1993) summarizes the evolution of nomenclature applied to this unit.

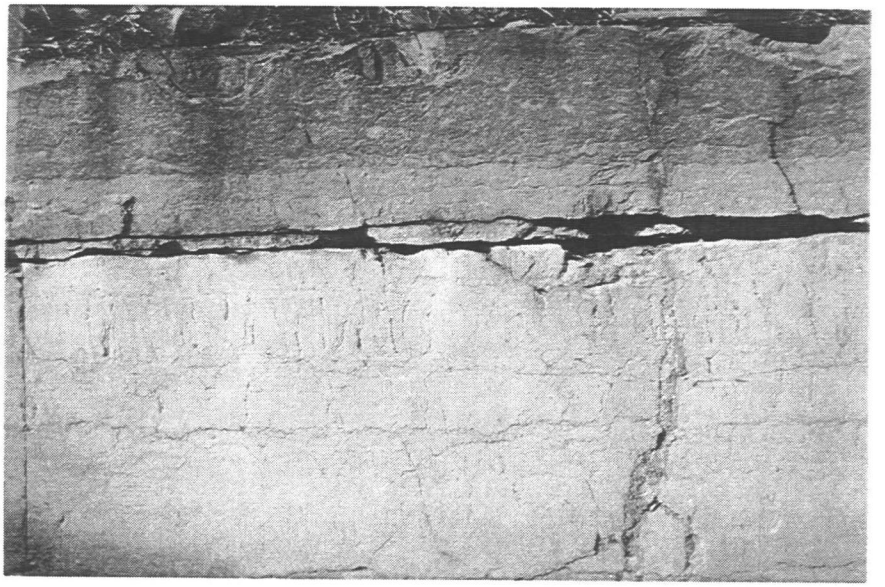
Stop 6 W parade grounds

Bechtel and Mehrtens (1993) suggested that the sandstone unit in the westernmost parade ground is the basal sandstone of the Black River Group, an interpretation which differs from that of Speyer and Selleck (1988) who suggested that this unit was part of the underlying Chazy Group. In thin section this sandstone is a quartz arenite in composition, very poorly sorted, containing fewer lithic fragments and phosphatic fragments than Chazy sandstones. Visible at the very easternmost portion of this ridge an overlying buff-colored dolomite bed containing pockets of quartz sand (burrow infills?) is exposed. The sandstone and dolomite lithologies are very similar to those described by Walker (1972) for the Pamelia Formation at the type locality in New York. Alternatively, placement of the sandstone within the Chazy Group is consistent with the common presence of coarse quartz and feldspar sand within the Chazy here, whereas siliciclastic material in the Black River Group at Crown Point is mainly fine silt and clay. Whatever the stratigraphic placement of this unit, it marks an interval when sands were transported from a nearby (Adirondack?) source area. This period of subaerial exposure of the shelf was followed by marine reworking of the sand, and deposition of the muddy carbonates of the basal Black River Group.

Stop 7 Northeastern moat

There is approximately 4 feet of covered interval between the parade grounds and moat exposures.

The wall of the northeasternmost moat exposes several feet of the lower Black River Group (Lowville Formation, House Creek Member). At the base of this exposure a series of stylolitized gastropod-bearing (*Liospira*) wackestone beds are overlain by thick beds of Phytopsis-burrowed aphanitic mudstones. This sequence can be interpreted as an example of a "classic" shallowing-up cycle consisting of subtidal to peritidal lagoonal muds. Examine the sharp contact of the aphanitic mudstone with the overlying wackestones, a contact which in thin section appears to be a firmground (Bechtel, 1993).



Examination of *Phytopsis* burrows in thin section reveals that many are filled with graded (fining-up) geopedal silt, evidence of cementation in the meteroic vadose zone (see text photo).

At other localities in the Champlain Valley, several of these cycles comprise the base of the Black River. Of the three motifs of repetitious bedding in the Black River Group, these cycles occur at the largest, macroscopic scale and are interpreted to represent 4th order (10,000 to 100,000 years) or smaller cycles.

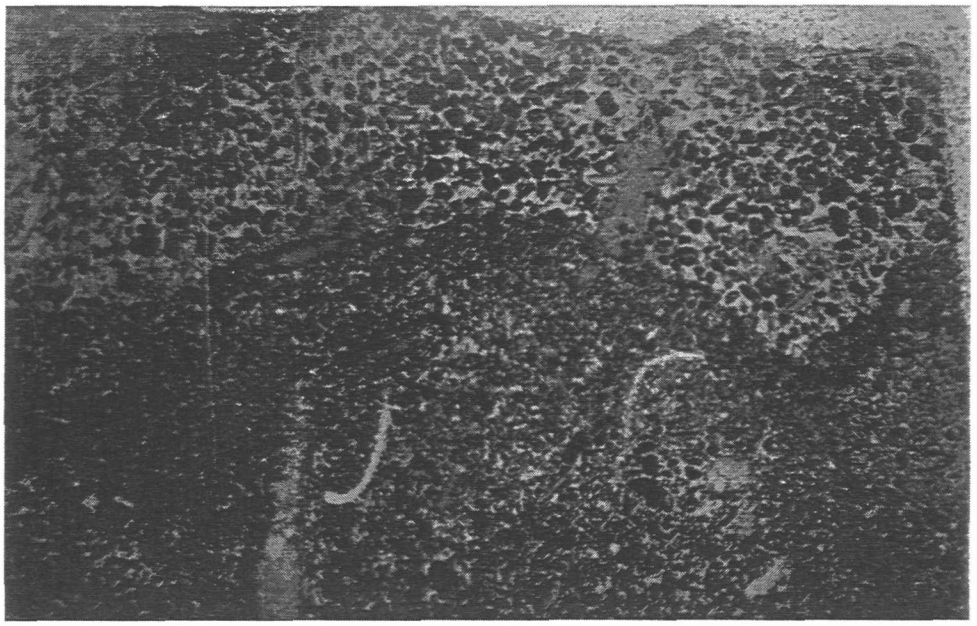
Continuing upsection, several thick packstone beds are exposed. More detailed examination of these beds reveals that they consist of alternating one to three inch thick intraclast and oncolite-rich packstone horizons interpreted as tempestites, interbedded with fossiliferous wackestone/packstone horizons. The tempestites, or storm-generated deposits, consist of graded and crudely imbricated intraclasts and skeletal fragments. Note the nature of the upper and lower contacts of these horizons. Horizons of tempestite beds within in situ fossiliferous muds is a second motif which occurs throughout much of the Black River Group.

It is instructive to spend several moments sketching the basal six feet of the moat exposure. Your sketch could illustrate the basal SUC's as well as the upper and lower contacts and internal fabric of the tempestite-rich beds.

SKETCH HERE

The uppermost third of the outcrop appears to be a massive bed of limestone, however closer examination also reveals small scale "cycles" of alternating wackestone/packstone and grainstone, the third motif of bedding in the Black River. These are characterized by a base of thinly laminated or cross laminated grainstone horizons 1 to 2 inches thick, overlain by fossiliferous wackestones and packstones. In thin section the bases of the grainstones can be identified as firmgrounds, recognizable by the truncations of allochems and cements in the underlying mud.

The photograph below illustrates the nature of the contact between a peloid-rich grainstone of the base of a cycle and the top of the underlying packstone. The scalloped surface and truncated grains and allochems are characteristic of these contacts.



The very top of this exposure (best seen at the next outcrop) exhibits a burrow mottled fabric with selected dolomitization of many burrows. Tetradium occurs in live position in these horizons.

Stop 8 East-west ridge

A black chert layer near the top of Stop 7 provides the correlation to Stop 8, the outcrop across the service road.

The limestone beds on this ridge commonly consist of alternating wackestone/packstone and planar to cross laminated grainstone beds as seen at Stop 7, however bedding plane exposures permit identification of many fossils in these, the most faunally diverse beds in the Black River. Specimens of gastropods (*Liospira, Lophospira, Hormotoma*), *Lambeophyllum, Tetradium*, stromatoporoids, the bivalve *Cyrtodonta*, *Strophomena* sp. and cephalopods are recognizable. This ridge is most notable for its bedding plane exposures of *Tetradium* and *Lambeophyllum* and is interpreted as recording a wave baffle margin lithofacies described by Walker (1972) at the type section.

A sketch of the third type of cycle, laminated grainstone overlain by bioturbated wackestone, with special attention paid to the nature of the contacts between cycles, would be appropriate. These sequences are interpreted to represent smaller scale 4th order 'micro-cycles' that could be the result of facies mosaicing and/or small scale base level changes.

SKETCH HERE

There are at least two distinct types of chert occurences in the Black River. One appears to be fabric selective: infilling horizontal burrows, for example (many other burrows are dolomitized). The other chert occurrence is less frequent and consists of broad, bedding parallel sheets. The uppermost chert horizon on this ridge, traceable down to the shoreline, is of this latter variety. Clearly, there was a significant source of silica available for chert formation, perhaps a combination of silica derived from sponges (*Tetradium*?) and bentonite alteration (the Ordovician sequence in the Champlain Valley is notorious for the paucity of bentonites compared to equivalent strata in central New York and Quebec). In thin section the chert cross cuts all previous cements, including late fracture-filling calcite, and is therefore the youngest, latest example of diagenesis in these rocks.

Stop 9 Quarry

Be extremely careful around the quarry - the thick algal scum in the quarry water obscures where the grass begins and the quarry wall drops off.

The older, weathered south walls of the quarry show, by color differentiation, two cycles. Closer examination of the more accessible north wall reveals more occurences of the third motif of Black River bedding: 8 to 10 inch thick beds of planar laminated skeletal and peloidal hash overlain by burrowed wackestones overlain by an intraclast-rich horizon. Interbedded with these cycles are also tempestite couplets of mudstone/wackestone and fossil hash layers in which brachiopod-rich layers are abundant (look for 'nested' pockets of shells).

Sketch the interbedding of the two cycles:

Before leaving the quarry area, note the numerous quarried blocks stacked between the quarry and the shoreline. See if you can recognize cycles, and from these, topping directions.

Stop 10- Shoreline

Uppermost horizons in the quarry can be traced down to the shoreline to the north where the Black River section continues (Chaumont Formation of the Black River Group) with small covered intervals up to the Trenton Group limestones. Shoreline bedding planes exhibit horizontal burrows of *Chondrites*, opercula of *Maclurites*, and polished surfaces also reveal large intraclasts, in other words the same internal stratigraphy that could be viewed in cross section on the quarry walls.

Cement Stratigraphy of the Black River Group

There are multiple types of cements present within the Black River limestones which record a complex diagenetic history. The general cement stratigraphy pattern records early nonluminescent cement associated with precipitation in oxidizing waters of the shallow meteoric phreatic zone. With increasing reducing conditions, bright and dull luminescent cements represent precipitation in shallow burial conditions. Ferroan calcite with dull to nonluminescence represents precipitation in a late burial situation from high temperature burial fluids. Early marine Black River Group micritic cement is ferroan and very dull luminescence representing deposition in a reducing, lagoonal environment. Subsequent cementation took place in the shallow meteoric phreatic zone, with nonluminescent cements with bright rims representing oxidizing conditions becoming slightly more reducing with burial. These observations are consistent with those of Mussman, et al. (1988) who interpreted such patterns to be related to a cratonward-dipping meteoric water lens beneath tidal flats. Tectonic uplift would lead to stagnation of the aquifer and increasingly reducing conditions. Within this general pattern, however, there are many variations in the Black River limestones which record frequent base level changes associated with sea level fluctuations and block fault movements in the Taconic foreland basin. These base level changes have produced numerous firmgrounds (at all Black River localities) as well as beachrock (at Arnold Bay) and paleo-karst (at Arnold Bay, Chippen Point and Sawyer Bay localities) horizons.

Fractures are common throughout the Black River and their cements record evolving burial conditions. The following two photos illustrate some of the observed patterns. In the first thin section two cement events are visible in the fracture. The first consists of nonferroan scalenohedral crystals extending outward from the fracture wall. These are interpreted to have been precipitated in the meteoric phreatic zone. The later large ferroan equant blocky crystals in the center of the fracture represent a late burial cement precipitated under reducing conditions (field of view 1.8cm).

In the second photomicrograph, taken under cathodoluminescence (field of view 0.5mm), the zoning of rhombohedral crystals infilling a fracture can be seen. The very symmetrical zoned patterns starts (from the interior outward) with a nonluminescent nonferroan core, a dull rim, a bright orange rim, another dull rim, to another bright rim and fading to nonluminescent outer rims. The nonferroan to ferroan zonation is indicative of increasing reducing conditions during cementation. The cement stratigraphy of the fractures indicates that their formation occured throughout the diagenetic history of the Black River, from early syndepositional events associated with karst and beachrock formation, through to deep burial.

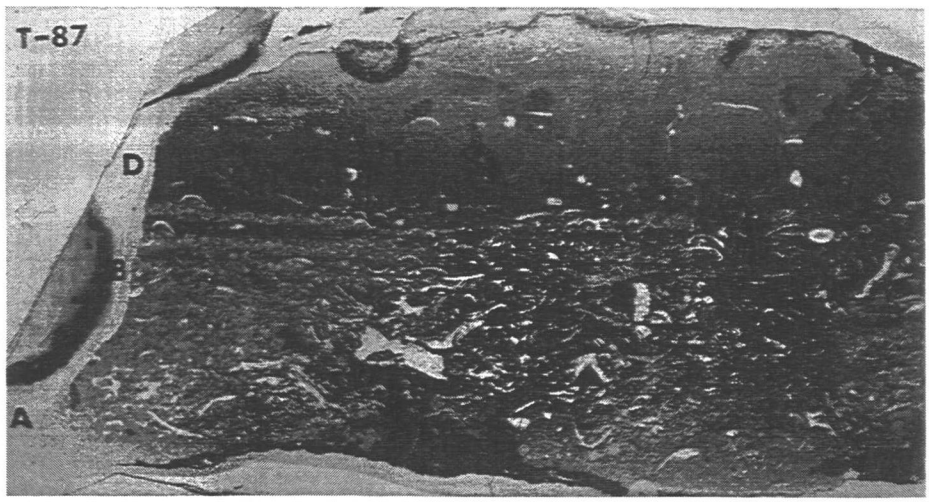
Trenton Group

The contact between the Trenton and Black River Groups is covered at most localities in the Champlain Valley, probably because the thinner-bedded and finer-grained Trenton is easily weathered. At Arnold Bay, to the northeast of Crown Point, the contact is exposed and is interpreted to be a disconformity. The dark gray colored, massive homogenous beds of the Black River are in sharp contact with the rubbly, laterally discontinuous beds of the Trenton. Here at Crown Point the Trenton is thinner than elsewhere in the Valley (28 feet), which MacLean (1987) suggested might reflect deposition on a down thrown block in the Taconic foreland basin. MacLean measured 50 feet of Glens Falls in the section at Button Bay (a few miles to the north of Crown Point) which, because it includes both the upper and lower contacts with the Black River Group and Cumberland Head Formation, respectively, represents the only complete exposure of this unit in the Champlain Valley. Bechtel (1993) summarized the variable nature of the Black River/Trenton contact around the Champlain Valley, New York and Ontario and noted that the regional variation seen would be expected in a foreland basin actively undergoing syndepositional block faulting

Stop 11- Shoreline of eastern Bulwaga Bay

Continuing up the shoreline from the uppermost outcrops of Black River a thin covered interval (4') occurs before the basal beds of the Glens Falls Formation. The Glens Falls is characterized by thin, nodular to wavy bedded wackestones, mudstones and rare grainstones, a very different type of bedding style and faunal assemblage from the underlying Black River. Bedding planes along the shoreline contain mostly *Chondrites* and *Helmenthopsis* burrows, however as one moves up section recognizable pieces of *Cryptolithus*, *Isotelus*, orthid brachiopods, *Stictopora*, and *Prasopora simulatrix* can be found, the latter is important because it permits the correlation of the lower Glens Falls here in the Champlain Valley to the lower Denley Limestone at the Trenton type section in central New York.

MacLean (1987) interpreted the lithofacies of the basal Glens Falls to represent sedimentation in a shallow subtidal environment periodically influenced by storm activity. In thin section the nodular and wavy bedded wackestones appear thoroughly bioturbated, a process which would influence and inhance subsequent differential compaction. Grainstone beds exhibit more planar bases with basal skeletal fragment lags or finely crushed debris of brachiopod, trilobite and crinoidal material capped by mud (see thin section photo below). MacLean interpreted these as tempestite deposits.



Based on lithologies, fossil assemblages and sedimentary structures, or the lack thereof, the Glens Falls Formation is interpreted to represent sedimentation in a deeper water, subtidal setting relative to the Black River deposits. For those familiar with lower Trenton localities in central New York the paucity of fossiliferous bedding planes here at Crown Point is noteworthy. The overall fine grain size and ichnofauna suggest that bathymetry increased significantly and rapidly from the Black River into the Glens Falls, a transition which might reflect not only rising sea level but base level changes as well. The sedimentologic and faunal transitions from the Glens Falls to the overlying Cumberland Head Argillite and Stony Point Shale is much more gradational that that of the Black River/Glens Falls contact.

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WILSON – CARBON CYCLE LINKAGES:

THE IAPETAN CYCLE AND EARLY PALEOZOIC CLIMATE

by

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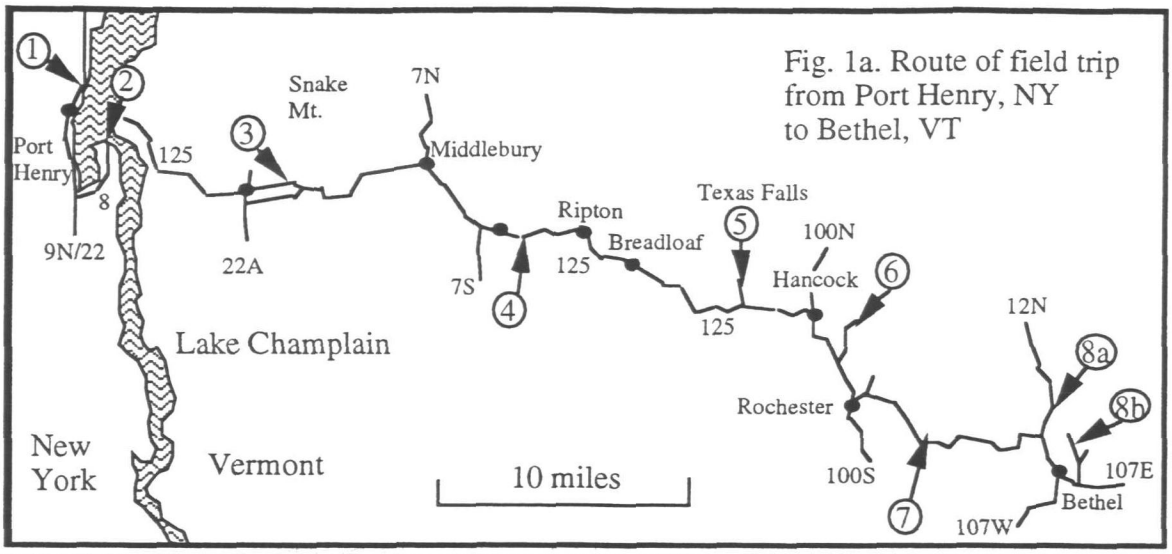
INTRODUCTION

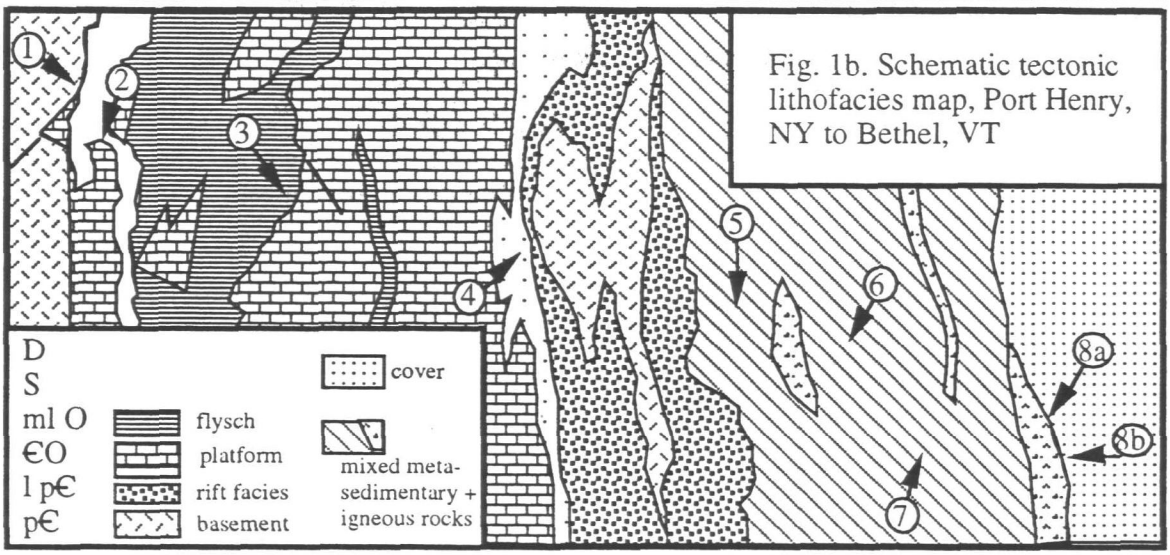
Viewing the Earth as a system raises a number of intriguing questions about relationships between tectonic and surficial processes. The geochemical carbon cycle illustrates particularly well some of the linkages among the different subsystems of Earth. Through the course of a Wilson cycle of seafloor growth and destruction, tectonic processes rearrange the global pattern of C sources and sinks. Consequently, variations in the partial pressure of atmospheric carbon dioxide (pCO_2) may influence climate. A major effort is underway to establish the Cenozoic climate record and to determine the role of tectonic processes in climate change. This effort is based largely on marine cores. Whereas Paleozoic records are less complete, they are accessible and easily viewed where preserved on cratons or in orogenic belts. These records pose a variety of problems that should stimulate research leading to a better understanding of the earth system.

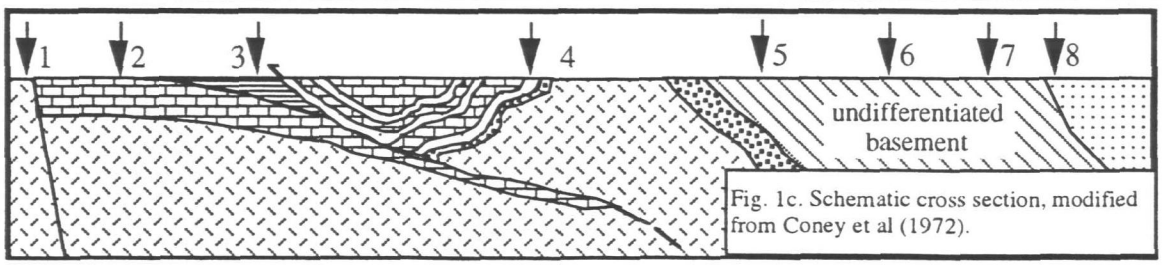
A pertinent example of a tectonic event with ramifications for climate is the Taconic orogeny (Chamberlin, 1899; Raymo, 1991; Kump et al., 1995; Gibbs et al., 1997; Reusch and Maasch, 1997). This large-scale disturbance may have played a key role in Ordovician global change by turning CO_2 sources (subduction systems) into CO_2 sinks, as calcium and magnesium-rich silicates within marine arc and ophiolite assemblages were uplifted, exhumed, and weathered. The weathering of these materials consumes atmospheric CO_2 when Ca and Mg ions are leached, transported to the ocean, and deposited in carbonate minerals. In a speculative scenario, the attempted subduction of the Laurentian passive margin led to decreased degassing, increased availability of soluble mafic lithologies for reaction with CO_2, and increased burial of organic C. The implied CO_2 drawdown suggests that a decreased greenhouse effect caused the Late Ordovician ice age.

Effects of climate on tectonic processes have also been noted (Hoffman and Grotzinger, 1993; Sleep, 1995). Precipitation influences the tectonic style of an orogen by enhancing erosional fluxes. Ranges subjected to high rainfall are rapidly eroded, expose high grade metamorphic rocks, and are flanked by their detritus. Arid ranges stand high, in contrast, and mass flows outward via thrust faults. Other factors equal, metamorphic rocks from wet climates should record steeper dP/dt (change in pressure through time) and dT/dt (change in temperature through time) than equivalent rocks from dry climates. The absence of a wide Taconic fold and thrust belt may imply a maritime climate and extensive erosion. High erosion rates, in turn, enhance CO_2 consumption by making silicates available for weathering.

This field trip across the Taconic orogen from Port Henry, NY to Bethel, VT will focus on materials and reaction sites relevant to the global CO_2 budget (Figs 1 and 2; Table 1). The trip should provide an overview of both Taconic geology and the geochemical (long term) C cycle. It will not present specific results of detailed local investigations but should provide a general context for future research on the role of tectonics in the C cycle, climate, and other aspects of global change. More questions will probably be raised than answered.







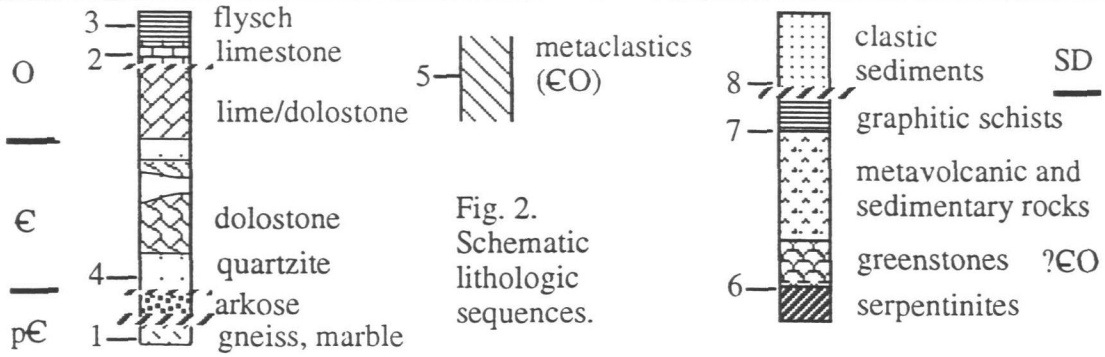


Fig. 2. Schematic lithologic sequences.

	Location	Name	Lithologies	Age	Tectonic significance	Significance for CO_2
1	Port Henry	Grenville	calc-silicate, graphite marbles	Middle Proterozoic	continental crust	thermal degassing
2	Crown Point	Glens Falls Orwell Valcour	carbonates	Middle Ordovician	top of carbonate platform	arc-continent collision as CO_2 switch (on/off)
3	Snake Mountain	?Iberville	shales	Middle Ordovician	foreland basin	silicate weathering; organic C burial
4	Ripton Gorge	Cheshire	ortho-quartzite	Early Cambrian	base of passive margin	spurt of sea-floor spreading? (degassing)
5	Texas Falls	Pinney Hollow	pelitic schists	Early Paleozoic	metamorphic core	implications of exhumation for CO_2
6	Verde Antique quarry		serpentinite, carbonate	Cambro-Ordovician	ophiolite suture	degassing; silicate weathering
7	Rochester Mt.	Stowe	chlorite and graphitic schists	Cambro-Ordovician	basement relative to Stop 8	degassing and organic C burial
8	Bethel	Shaw Mt.	calcareous quartzite	Silurian	successor basin	silicate weathering; organic C cycle

Table 1. Field trip stops.

Historical background

More than 150 years ago, major insights into the geochemical cycling of C were made by the French chemist and mining engineer Jacques Joseph Ebelmen (Ebelmen, 1845; Berner and Maasch, 1996). Based largely on comparisons of unaltered and weathered materials, Ebelmen inferred the principal inorganic and organic reactions governing the C cycle. Ebelman grasped the significance of these reactions for atmospheric composition. Not long after Agassiz had publicized evidence for past climate change, Ebelmen hinted at how atmospheric composition might affect climate.

Before the end of the 19th century, A.G. Högbom developed the concept of the geochemical C cycle and Svante Arrhenius quantified the role of CO_2 in warming the atmosphere via the "greenhouse effect" (Högbom, 1894; Arrehenius, 1896; Berner, 1995). The eclectic American geologist T.C. Chamberlin noted a correlation between ice ages and orogenies (1899). His prototype working hypothesis, which has yet to be rejected, states that mountain building leads to global cooling by increasing the availability of silicates for weathering and consumption of CO_2.

In the 20th century, stable isotope geochemistry has aided in the quantification of the C cycle (Appendix A). A number of geochemists have modeled the C cycle (see Berner, 1991 or 1994 for references), and at least some of the model results suggest that pCO_2 may play an active role in long-term climate change. In the Berner-type models, silicate weathering serves as an important negative (stabilizing) feedback, considered by some to be the principal planetary thermostat (Walker et al., 1981; Volk, 1987).

The past decade has witnessed intense interest in the earth system and global change, and a plethora of pCO_2-forced climate change hypotheses have been proposed. Raymo (1988, 1991)

suggested that erosion-driven consumption of CO_2 by silicate weathering forced the Cenozoic and other global cooling events (also Bickle, 1996). Kerrich and Caldeira (1993) related warm climates to increased subduction and degassing of carbonates; Selverstone and Gutzler (1993) arrived at just the opposite conclusion by noting a correlation between Cenozoic cooling and long-term removal of sedimentary C by subduction. The organic C cycle has been invoked as both a cause of cooling (Derry et al., 1992; Raymo, 1994) and of warming (Beck et al., 1995).

CARBON CYCLE

The C cycle refers to the movement of C through different rock and fluid reservoirs. The atmosphere, oceans, cryosphere, and biosphere have interacted extensively with the lithosphere through earth history. Carbon plays important roles in each of these subsystems. In the form of CO_2, it behaves as a greenhouse gas, elevating the temperature of the atmosphere. Carbon is the fundamental element of the biosphere. In natural waters, inorganic C regulates pH and organic C exercises a strong influence on redox conditions. Carbonate rocks and hydrocarbon deposits constitute quantitatively significant parts of the solid Earth.

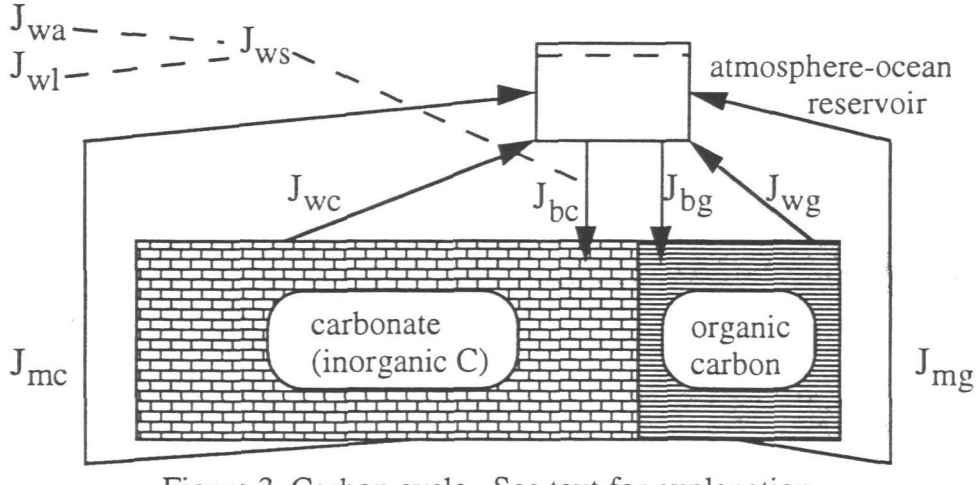


Figure 3. Carbon cycle. See text for explanation.

The vast majority of C in the Earth (outside of the mantle) exists in carbonate minerals (5000×10^{18} moles) and as organic C (1250×10^{18} moles) (Berner et al., 1983). Organic C occurs chiefly as kerogen in sediment, but also as coal, oil, and gas. In comparison, the combined atmosphere-ocean reservoir contains only 3×10^{18} moles, and the miniscule atmosphere reservoir contains approximately 1/60th of the amount of C in the ocean reservoir. Thus, on a million year time scale, the atmosphere-ocean reservoir behaves more as a conduit than a reservoir, because the residence time of C in this small reservoir is negligible. The relationship between the partial pressure of atmospheric CO_2 (pCO_2) and the net rate at which C flows to and from rock reservoirs is complex (e.g., Berner, 1994). Inputs of C to the atmosphere include volcanic degassing and oxidative weathering of reduced organic C. The rates of the two output processes, silicate weathering and organic C burial, are dependent on atmospheric pCO_2, which results in important stabilizing (negative) feedbacks. For example, volcanic degassing tends to increase pCO_2, which enhances silicate weathering through the direct effect of CO_2, augmented by increased temperature, precipitation, and plant activity.

The C cycle operates because of the differential movement of C-bearing phases and their respective stability fields. Generally, carbonates and organic C compounds are not stable at high temperature, carbonates are not stable at low pH, and reduced C is not stable at high pe (high electron activity or redox level). When tectonic, climatic, or biologic forces cause environmental parameters to change, the C phases may react and cause a flux between reservoirs. Three highly generalized reactions account for the principal C fluxes.

$$CO_2 + H_2O = CH_2O + O_2 \quad (1)$$

$$CO_2 + CaCO_3 + H_2O = Ca^{++} + 2HCO_3^- \quad (2)$$

$$CO_2 + CaSiO_3 = CaCO_3 + SiO_2 \quad (3)$$

Each reaction has two or more fluxes associated with it, depending on the direction and cause of the reaction. Fig. 3 shows how the six main fluxes link the three reservoirs.

J_{bg} = burial of organic carbon (eq. 1 left to right);

J_{wg} = oxidative weathering of organic carbon (eq. 1 right to left);

J_{mg} = thermal oxidation of organic carbon (eq. 1 right to left);

J_{bc} = burial of carbonate (eq. 2 right to left);

J_{wc} = dissolution of carbonate (eq. 2 left to right);

J_{mc} = thermal decarbonation (eq. 3 right to left).

In addition, silicate weathering is the reverse of thermal decarbonation (J_{mc}).

J_{ws} = silicate weathering (eq. 3 left to right).

With respect to the rock reservoirs, these C fluxes can be thought of as additions at the water-sediment interface (J_{bc} and J_{bg}), subtractions via weathering from the top of the reservoirs (J_{wc} and J_{wg}), and subtractions from below (J_{mc} and J_{mg}). With respect to the atmosphere-ocean reservoir, just the opposite is true. The level of C in the atmosphere-ocean reservoir is generally assumed not to have shifted wildly through time since evidence for either a runaway icehouse or runaway greenhouse has not been recognized in the geologic record (with the exception of a widespread hiatus in Precambrian strata that could signify a frozen planet [Cesare Emiliani, 1995, pers. com.]). This assumption makes it possible to model the geochemical C cycle as a succession of steady states. The sum of inputs, $J_{bc} + J_{bg}$, therefore equals the sum of outputs, $J_{wc} + J_{wg} + J_{mc} + J_{mg}$. The mass balance for C in the atmosphere-ocean reservoir can be further simplified because the total carbonate burial flux (J_{bc}) equals the sum of the carbonate and silicate weathering fluxes ($J_{wc} + J_{ws}$). Thus, on a million year time scale, the sources of atmospheric CO_2 (metamorphic degassing and organic C weathering) must closely match the sinks (silicate weathering and organic C burial). Check the algebra! Some parameters used to calculate C fluxes in C cycle models are shown in Table 2.

C flux	sea-floor s.rate	pelagic carbon ate	land area	relief	runoff	vegeta-tion	temper-ature, CO_2	product ivity	anoxia	mineral surface area
J mg	X									
J mc	X	X								
J bg							X	X	X	X
J wg			X	X	X					
J wc			X		X	X	X			
J ws			X	X	X	X	X			

Table 2. Parameters multiplied to obtain C flux estimates in C cycle models (e.g., Berner, 1990).

Different tectonic settings favor different C fluxes, for example active magmatic arcs are sites of degassing and rifted margins are sites of C accumulation. Table 3 shows C fluxes that may be dominant in particular settings. The settings are arranged in order of the Wilson (supercontinent) cycle. Note that periods of earth history dominated by seafloor spreading and subduction may be accompanied by high degassing rates, high pCO_2, and warm climate. The culminations of Wilson cycles, characterized by collisions, may be accompanied by high weathering rates, low pCO_2, and cool climate.

Tectonic setting	Thermal degassing (J_{mc} and J_{mg}) **source**	Organic carbon weathering (J_{wg}) **source**	Ca, Mg silicate weathering (J_{ws}) **sink**	Organic carbon burial (J_{bg}) **sink**
rift				x ?
spreading center	X			
subduction zone	X			
arc-continent collision zone		x (early)	X	X
continent-continent collision zone			x	x

Table 3. Carbon fluxes thought to be quantitatively significant in specific tectonic settings.

GEOLOGIC RECORDS AND INTERPRETATIONS

The Iapetan cycle

The Iapetan cycle began in latest Precambrian time with the breakup of the hypothesized short-lived supercontinent Pannotia (Powell, 1995; Dalziel, 1997). Pannotia was the successor to Rodinia, the supercontinent formed as a result of the Grenville Orogeny at 1 Ga. Subsequently, during the Late Proterozoic, Rodinia broke apart resulting in the Cordilleran margin of Laurentia. Pannotia formed as a consequence of the Pan-African consolidation of East and West Gondwana, with Laurentia attached to West Gondwana through this interval. The prominent rise in marine $^{87}Sr/^{86}Sr$ ratios that lasted into the Cambrian has been attributed to sialic crust exposed in Pan-African orogens (Kaufman et al., 1993). During the latest Proterozoic, Laurentia separated from the present west coast of South America (in this reconstruction), breaking Pannotia apart and resulting in the Iapetan margin of Laurentia.

By earliest Cambrian time, a transition from rifting to full-fledged seafloor spreading was underway (Williams and Hiscott, 1987). Through the Cambrian, Iapetus grew into a wide ocean basin. The Laurentian margin, located at low latitude, accumulated orthoquartzites and later an extensive carbonate platform (Fig. 4). Across Iapetus lay the Avalonian microcontinent, possibly a peri-Gondwanan landmass. The Avalon platform also accumulated orthoquartzites during the earliest Cambrian. Its later development differed substantially, however, as it became a site of siliciclastic rather than carbonate deposition.

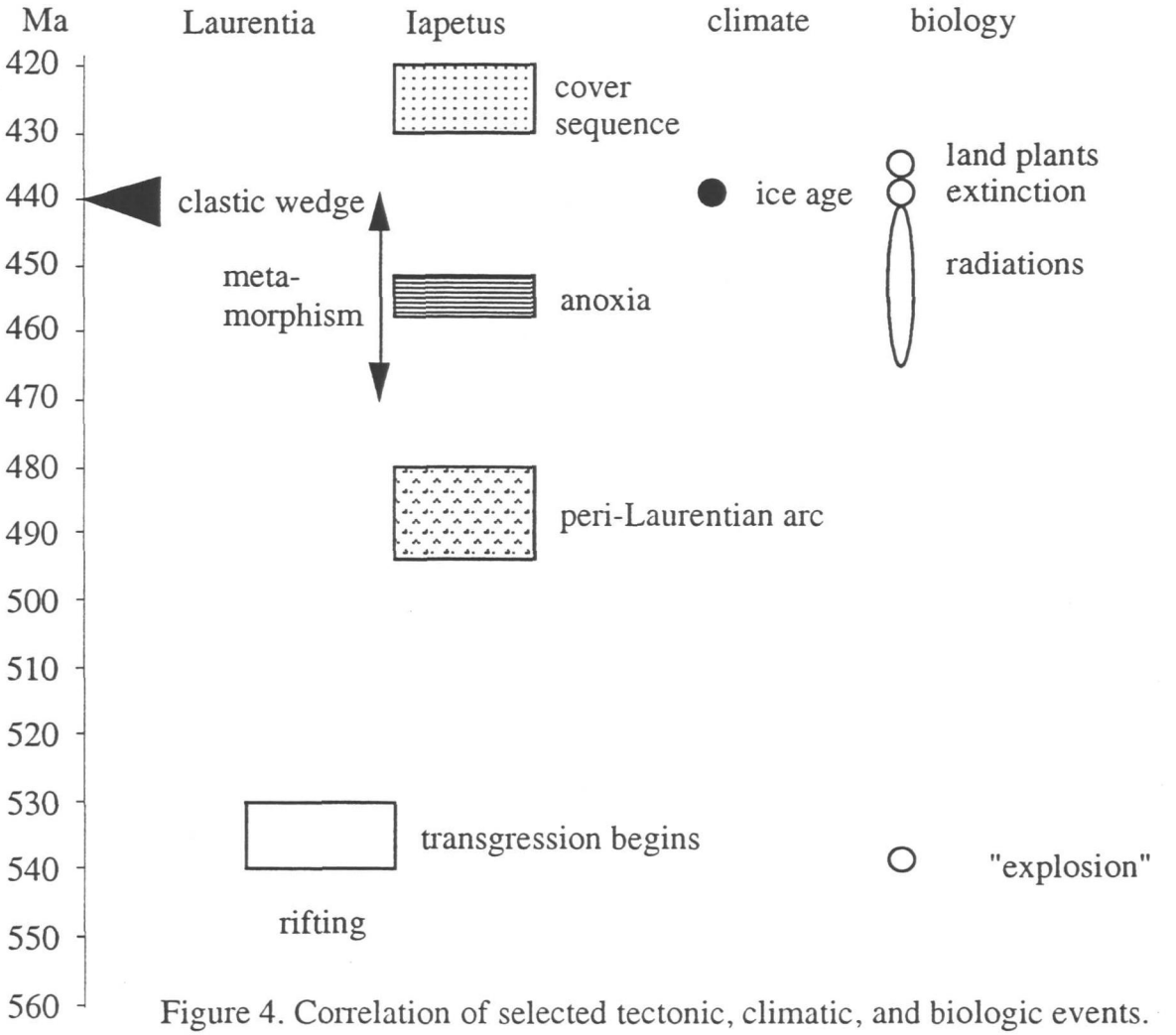


Figure 4. Correlation of selected tectonic, climatic, and biologic events.

Toward the end of the Cambrian, Iapetus began to self-destruct, which led to a series of collisions. Ophiolitic melanges mark the subduction zones. The earliest collisions occurred internally, for example the Penobscot orogeny in which the Boundary Mountains (Chain Lakes) and Gander terranes were sutured together (Boone and Boudette, 1989).

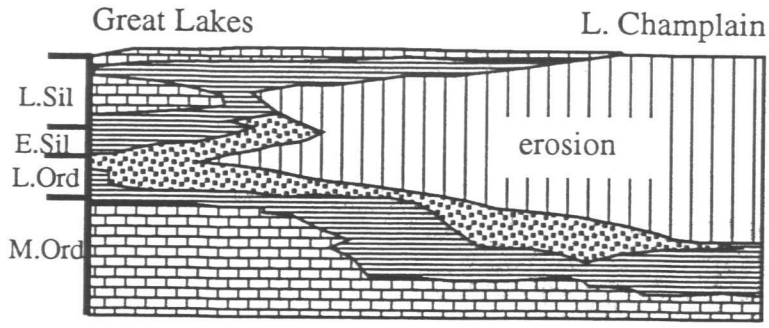


Figure 5. Queenston clastic wedge (modified from Rodgers, 1971). This clastic wedge extended from the Taconic orogen westward a third of the way across the continent. It reached its maximum extent during the Late Ordovician. A prominent glaciation occurred on Gondwana during this interval.

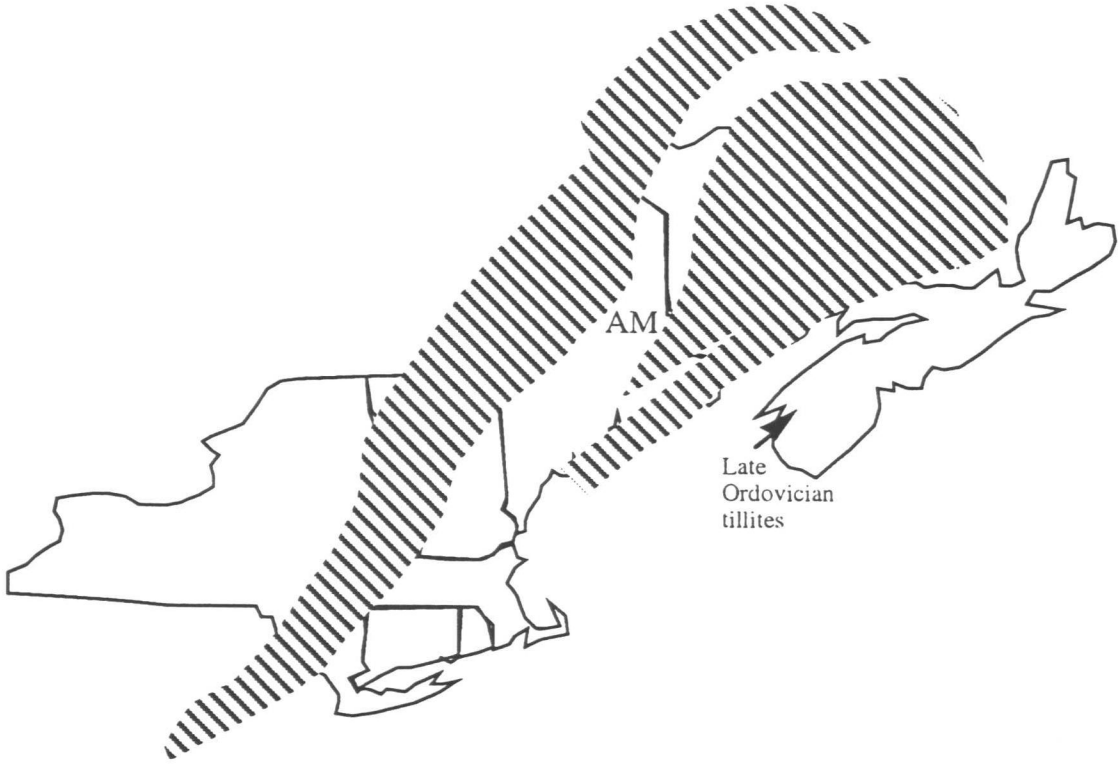


Figure 6. Silurian unconformity. Most of the Silurian strata in the northern Appalachian Orogen were deposited unconformably on deformed Ordovician marine facies. One exception is the Aroostook-Matapedia Belt (AM), where sedimentation was continuous through the interval of Taconic orogeny. Modified from Pavlides, 1968.

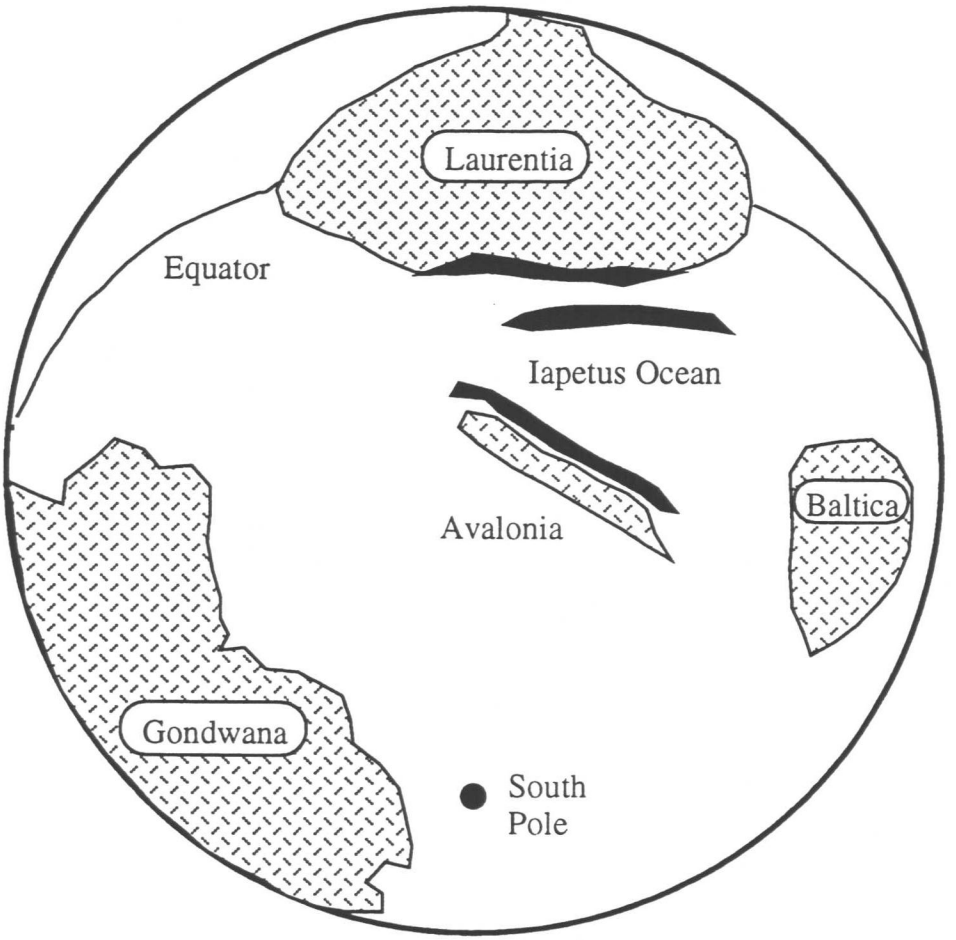


Figure 7. Reconstruction of world during Middle to Late Ordovician time (modified from Mac Niocaill et al., 1997). In the following reconstruction for Late Ordovician - Silurian, Avalonia is shown as having collided with Laurentia. Arcs shown in black.

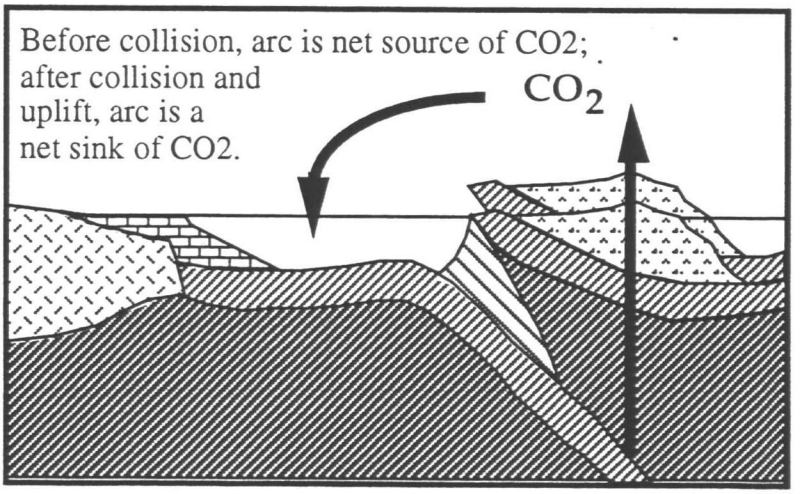


Figure 8. Passive margin - arc collision (e.g., Hamilton, 1988). The arrows suggest how the plate boundary evolves from a net source to a net sink of CO_2.

During the Middle Ordovician, the Laurentian margin, from Newfoundland to the southern U.S., entered a subduction zone. By Late Ordovician time, clastic wedges extended westward over the Laurentian craton (Fig. 5). By Silurian time, continental or shallow marine sediments began to accumulate on formerly marine tracts which, in most places, had been deformed, uplifted, and eroded (Williams, 1967; Fig. 6). Locally, sedimentation was continuous through this interval (e.g., Aroostook-Matapedia belt of northern Maine and the Maritimes; New World Island in NF, Reusch, 1987). In Maine, Silurian unconformities occur on both sides of Iapetus. The post-Taconic evolution of the Appalachian Orogen is not at all straightforward, but by Devonian time, the eastern Avalonian continent was firmly welded onto Laurentia.

Unconformities are of particular interest with respect to the C cycle because they represent former surfaces of interaction with the atmosphere. The composition of the basement, before it was covered, influenced the composition of the contemporaneous atmosphere-ocean. The field trip traverses four Paleozoic unconformities, shown in Table 4.

Name (Stop no.)	Age of oldest cover	Basement	Cover	Comments
Taconic unconformity (8)	Silurian	variable	quartzite	major angular unconformity
forebulge (?) unconformity (2)	Middle Ordovician	carbonates	limestone	disconformity
rift-drift transition (4)	Cambrian	rift deposits	orthoquartzite	weak angular unconformity
Great Unconformity (4)	late Precambrian	Grenville basement	rift deposits	major nonconformity

Table 4. Unconformities. These erosional surfaces constituted part of the "lining" of the atmosphere. The basement composition is pertinent to the composition of the contemporaneous atmosphere-ocean.

The record of surficial systems and possible tectonic linkages

The latest Precambrian (Vendian) period set the stage for the Cambrian "explosion" of skeletalized animals (Kaufman et al., 1993). Widespread glaciations occurred at 600 Ma (Young, 1995). These Varanger glaciations have been ascribed to various mechanisms such as 1) widespread uplift contemporaneous with supercontinent rifting and 2) Taconic-style arc-passive margin collisions (initial Pan-African orogenies) during the assembly of Gondwana (Reusch and Maasch, 1997). Later in the Vendian, the first metazoa (Ediacaran fauna) evolved and flourished. Derry et al. (1992) attribute high $\delta^{13}C$ values during this interval to preservation of organic C due to high sedimentation rates (perhaps related to Pan-African orogenesis). They further suggest a related increase in pO_2, which may have enabled the Ediacaran metazoa to evolve. Kaufman et al. (1993) stress that the prominent increase in marine $^{87}Sr/^{86}Sr$ ratios follows the Varanger glaciations; i.e., increasing marine $^{87}Sr/^{86}Sr$ ratios can not be linked to a causal mechanism for global cooling as suggested for Phanerozoic glaciations by Raymo (1991).

Generally, Cambrian through Middle Ordovician climates were warm. No important tillites are known from this interval, despite Gondwana having occupied a polar position. The basal Cambrian transgression and a contemporaneous decrease in marine $^{87}Sr/^{86}Sr$ ratios suggest a surge of seafloor spreading at this time (Nicholas, 1996). Higher spreading rates, in turn, may have sent the planet into greenhouse mode. The pattern is analogous to the beginning of the Mesozoic-Cenozoic supercycle. As in the Cretaceous, the deep ocean may have become warm and anoxic. Near the base of the Middle Cambrian, the Avalon platform hosts magnesium carbonate deposits (Anderson, 1987; Emiliani, 1992), which perhaps may have precipitated when Mn-saturated bottom waters upwelled onto the platform and were oxidized.

During Middle Ordovician time, black shales were deposited all across Iapetus and in other parts of the world (Leggett, 1978). The cause of this event is unclear. Similar shales in the early to middle Miocene Monterey Formation were deposited during a warm interval before the middle Miocene cooling event. Positive shifts in $\delta^{13}C$ values coincide with both intervals of organic C deposition. Evidence for Middle Ordovician climate is meager, but a cooling trend may have begun by the late Middle Ordovician (Patzkowsky et al., 1997). A second relevant comparison is with the early Eocene warming and negative shift in $\delta^{13}C$ values that Beck et al. (1995) linked to uplift, erosion, and weathering of organic C-rich sedimentary margins during the first phase of Himalayan orogenesis. From these examples, it appears that the initial uplift of passive margins (Laurentia during the Ordovician, India during the Eocene, and Australia during the Miocene) may perturb the organic C cycle but in different ways.

Widespread marine invertebrate radiations took place through the Middle and Late Ordovician (Miller and Mao, 1995). These radiations have been related to the development of diverse new habitats, specifically foreland basins, created during the Taconic collision (Fig. 7). Much of this biologic creation was, however, undone, by the mass extinction at the end of the Ordovician.

At the end of the Late Ordovician, quite suddenly, the world cooled for a short period of time, probably less than a million years (Brenchley et al., 1994; Kump et al., 1995). Tillites were deposited on Gondwana (some of these may be found in southwestern Nova Scotia in the Meguma terrane of Gondwanan affinity). The Late Ordovician ice age has been explained by the passage of Gondwana over the south pole. However, the rapidity of the cooling is inconsistent with typical plate velocities. An alternative explanation invokes a drop in atmospheric pCO_2 (Fig. 8). Model results suggest that pCO_2 was generally high through the early Paleozoic (Berner, 1990), but recent work suggests that it may have decreased enough to trigger glaciation (Gibbs et al., 1997; Crowley and Baum, 1995). Maxima in C isotopic records (closest samples from Anticosti Island, Quebec) occur at the same level, and Brenchley et al. (1994) suggest that organic C burial may have been the cause rather than an effect of the cooling. A silicate weathering mechanism for CO_2 drawdown is also compatible with the tectonic record and may have been necessary to obtain sufficiently low background pCO_2.

Following the Late Ordovician ice age, climate warmed abruptly. During Silurian time, plants colonized the continents. An important question is why did they wait until the Silurian to invade this new habitat? A highly speculative possibility is that the increased oxygen production implied by the latest Ordovician C isotope anomalies led to increased stratospheric ozone and decreased ultraviolet radiation, a limiting factor for land plants.

OVERARCHING QUESTIONS

In closing, the merging of traditional geology with earth system science prompts a variety of questions that can be asked about the geologic record of the Appalachian Orogen.

- How much of the Laurentian passive margin sequence, including geochemical variations, was influenced by processes of a global nature? (Stop 4)

- Do Taconic-style passive margin-arc collisions cause global cooling by transforming the convergent plate boundary from a CO_2 source into a CO_2 sink? (Stops 2, 3, 5, 6, 7, 8)

- What is the relationship between exhumation of the Taconic metamorphic core, the Late Ordovician Queenston clastic wedge, contemporaneous orographic precipitation, structural style of the Taconic foreland, and CO_2 sinks (silicate weathering and organic C burial)? (Stops 3 and 5)

- Do arc-ophiolite lithologies play a disproportionate role in the C cycle because 1) their high temperature genesis is intimately associated with degassing phenomena and 2) their Ca and Mg silicate-rich compositions make them potent CO_2 sinks when exhumed and weathered? (Stops 6 and 7)

- What is the relationship between the widespread Silurian unconformity of the Appalachian Orogen and the specific mechanism of Taconic arc-continent collision restricted to the western part of the orogen? (Stop 8)

- How much C passes from the sedimentary realm into the metamorphic realm? (Stop 1)

ACKNOWLEDGEMENTS

Thanks to Mike Solari of Vermont Verde Antique International for permission to visit the Rochester quarry. Heather Short battled through a first draft. Tim Cronan helped with Stops 5 through 8; he located Stop 8B. Don Burns provided accommodations while in western Vermont. Martin O'Connell inspired the possibility of a C cycle connection to land plant evolution. Kirk Maasch, Kelly Rust, Dan Lux, Barry Doolan, and Dave Reusch made comments on later drafts.

ROAD LOG FOR TRIP A-4

Assembly time and place: Friday, September 19, 1997, 8:00 AM. We will assemble at the public boat launch and picnic area in Port Henry. This location is reached by taking Beach Road, which intersects NY 9N/22 0.3 miles north of the center of Port Henry (junction of NY 9N/22 and Broad Street). This is the same location described by McHone (1987).

Following a brief overview of the geochemical C system, Iapetan/Taconic cycle, and possible tectonic-climate linkages we will visit the first stop, located about a mile to the north.
Significance of stops is indicated by bold font, descriptions and explanations by plain font, and *questions for discussion by ➡ symbols and italicized font.*

MILEAGE

0.0 Turn right out of parking lot and return via Beach Road to NY 9N/22.
0.2 Turn right on Rte. 9N/22.
0.6 Entrance to Craig Harbor Campground. Turn right and proceed down steep hill. Bear left on first unpaved side road to campsite by cliff.

0.8 **STOP 1: GRENVILLE MARBLE, Port Henry, NY.**

Some of the sedimentary C carried to depth by orogenic processes returns to the external realm of Earth as a consequence of thermally enhanced reactions. The remainder resides in the internal realm for long durations.

The Grenville basement exposed here samples mid-crustal levels. The cliff is a scarp along one of the normal faults that mark the boundary between the Champlain Lowlands and Adirondack Massif (note patches of mafic dike that intruded along the fault). Differential weathering is well developed due to the high solubility of calcite. Most of the cliff is coarse-grained graphite-bearing marble with large (up to meter-size) fragments of calc-silicate material and gneiss scattered through it. Calc-silicate minerals, also disseminated in the marble along with quartz and pyrite, include diopside and garnet. A pocket of graphite occurs near the base of the outcrop. Towards the east is a steeply-dipping, well-foliated hornblende-rich gneiss. Note the 20 cm thick reaction zone along the contact with the marble.

➡ *How long have the C atoms in the calcite and graphite, within these metasediments, resided in the crust?*

➡ *Generally, how are rocks that crystallized in the mid-crust exhumed?* ➡ *Through time, what has been the relative importance here of erosion vs. tectonic denudation?*

➡ *What fraction of sedimentary C successfully makes it through the "calc-silicate" isograd from the external realm of Earth to deeper levels?* ➡ *For calc-silicate reactions, what limits the reactions: water, silica, or another factor?* ➡ *Why hasn't the quartz been entirely consumed?* ➡ *For the C now present in graphite, what factors limited the release of C from precursor organic compounds?* ➡ *Is all of the graphite derived from organic C in the parent sediment?*

The modern weathering surface is instructive for estimating relative solubilities. ➡*Which proceeds at a faster rate, oxidative weathering of graphite or dissolution of calcite?* ➡*How might physical processes influence chemical weathering of minerals with different hardnesses?* Grenville basement was exposed twice during the Iapetan cycle, during the late Precambrian rifting event and following the Taconic collision. ➡ *During the rifting event, what type of climate is likely to have prevailed in the interior of the Pannotian supercontinent?* ➡ *What implications would this climate have had for chemical weathering?*

➡ *Is the sialic crust exposed during orogenic events typical of average continental crust, which is thought to be generally andesitic (Rudnick, 1995)?* (For silicate weathering, only Ca and Mg silicates are relevant to the C cycle.) ➡ *For those with isotopic vision, what is the $^{87}Sr/^{86}Sr$ composition of the calcite here?* Edmond (1992) has argued that K, Rb-rich phases (feldspars), which accumulate ^{87}Sr, are resistant to chemical weathering and that large amounts of ^{87}Sr are only released following Himalayan-type metamorphism, during which the ^{87}Sr migrates into more soluble phases. Recent studies in the Himalayan foreland attribute increases in riverine $^{87}Sr/^{86}Sr$ ratios to exposure of radiogenic metacarbonates (Quade et al., 1997). Studies in the Baltic Shield region (Ravizza and Peucker-Ehrenbrink, 1996) show that sialic crust also hosts very radiogenic Os (possibly in a sulfide such as molybdenite).

1.0 Turn left (south) on NY 9N/22.

1.7 Center of Port Henry.

South of Port Henry, the highway follows one of the Adirondack normal faults. Note oxidized outcrops of basement rocks. **The oxidation of sulfides is an important sink for atmospheric oxygen.** ➡ *Given that the insoluble iron oxide coating inhibits further weathering, does this process act as a negative (stabilizing) or positive (accelerating) feedback with respect to atmospheric pO_2?*

5.4 Turn left on NY 8 towards Crown Point and bridge to Vermont.

8.8 Turn left at Crown Point State Historic Site. A fee may be charged, $4.00 for vehicles with under 10 passengers and $25.00 for vehicles with 10 or more passengers.

9.4 **STOP 2: TOP OF CARBONATE PLATFORM, Crown Point, NY**

The sequence at the top of the carbonate platform records the onset of collision between the Laurentian passive margin and an outboard trench-arc system. The plate boundary may subsequently have converted from a site of net CO_2 production to a site of net CO_2 consumption, triggering the Late Ordovician ice age. The distinctive facies sequence, including the transition to flysch deposits (Stop 3), is a clear indicator of passive margin – arc collision; the abundance of fossils facilitates dating of the collision.

For a detailed account of the classic exposures at this stop, see Mehrtens and Selleck (this volume, Trip A3). We will focus on just three features pertinent to the theme of this trip. A disconformity (Station D of Baldwin, 1987) occurs in shallow water facies of the Valcour Formation; subsequent facies of the Valcour and Orwell Formations (Station G) record progressively deeper water; and the uppermost limestones of the Glens Falls Formation (Station I) contain argillaceous partings foreshadowing the arrival of easterly derived orogenic sediment.

A disconformity exposed at Station D and in the nearby moat is subtle but, in view of its possible regional extent and tectonic significance, provocative. Following deposition of Lower Ordovician carbonates on a slowly subsiding shelf, the Laurentian platform was locally uplifted as much as 500 meters (Jacobi, 1981) before subsequent drowning. In Newfoundland, karst features developed, and Mississippi Valley-type Zn-Pb mineralization formed within the basement. ➡ *Does the disconformity at Station D reflect the pre-trench forebulge?* The $^{87}Sr/^{86}Sr$ composition of unmetamorphosed marine carbonates is intermediate between low mantle values and very high sialic (continental crust) values. ➡ *What effect would the dissolution of these highly soluble carbonates, during the sculpting of this unconformity, have had on the global $^{87}Sr/^{86}Sr$ signal?*

Above the disconformity, the first beds are shallow water facies, but the overlying beds in the remainder of the Valcour Formation, Orwell Limestone, and Glens Falls Formation record deepening environments (Mehrtens and Baldwin, 1976; Selleck and Baldwin, 1985; Baldwin, 1987). In the Orwell, black chert is sparse near the base but forms a prominent discontinuous layer near the top. In the silica geochemical cycle, silica enters the marine system via terrestrial weathering, hydrothermal solutions, and dissolution of marine siliceous deposits. ➡ *Is it possible that the crude trend of increasing silica in the Orwell Formation reflects the influence of the approaching Taconic arc?*

The Glens Falls Limestone is fossil-poor, was deposited below the photic zone, and contains argillaceous partings foreshadowing the first easterly derived detritus. ➡ *Why do these partings weather recessively?*

Bradley (1989) compiled ages for the time of "platform drowning" both along strike and perpendicular to the strike of the Appalachian Orogen. The integrated time of "platform drowning" pre-dates by some ten or so million years the time of maximum extent of the Queenston clastic wedge (Rodgers, 1971), which was contemporaneous with the Late Ordovician ice age. ➡ *How much delay should be expected between the entry of a passive margin into a trench and maximal uplift and erosion of the resulting orogen?*

10.0 Turn left on NY 8 and proceed over bridge across Lake Champlain to Vermont.

11.1 Turn right on VT 125.

The erosional flux from the Taconic orogen, controlled in part by contemporaneous rainfall, competed with tectonic deformation in determining the architecture of the foreland fold and thrust belt. The low mountain to the east is Snake Mountain, held up by resistant quartzites of the Middle Cambrian Monkton Formation in the upper plate of the Champlain thrust. This thrust marks the approximate western limit of the Appalachian Orogen. The Taconic fold and thrust belt here is barely 10 miles wide, much narrower than the equivalent Alleghanian belt of the southern Appalachian Orogen. ➡*What was the original width of the foreland fold and thrust belt?* ➡ *Can the relationship noted by Hoffman and Grotzinger (1993) between aridity and tectonic style be applied to the Taconic foreland?*

18.2 Intersection with VT 22A. Proceed across highway and continue along unpaved road (Market Street).

19.3 Intersection with East Street. Continue east on Market Street.

20.05 Outcrops of Middle Ordovician shales on right. These shales lie stratigraphically above the carbonates at Crown Point. They display a prominent pencil cleavage, forming splintery rubble.

20.3 Park along road and walk to quarry.

STOP 3: ORDOVICIAN FLYSCH, Snake Mountain, VT

Clastic wedges that flank orogenic belts imply enhanced silicate weathering and organic C burial, the two principal CO_2 sinks.

"Flysch", a useful term for orogenic sediment of marine facies and early timing in the development of an orogen, overlies the carbonate platform. The Taconic flysch is composed of detritus from the collided passive margin and arc system. Heavy mineral assemblages in sandstones indicate the presence of eroding ophiolites within the Taconic orogen (Hiscott, 1978; Nelson and Casey, 1979; Rowley and Kidd, 1981). Chromite appears in the oldest flysch, considerably before the time of "platform drowning" and subsequent Queenston clastic wedge. Hiscott (1984) and Garver (1996) have shown how the trace elements Cr and Ni measured in shales can be used to infer the extent of exposed mantle through space and time.

The quarry outcrops expose thinly bedded dark gray shale beds, light gray calcareous beds, and a few tan-weathering dolomitic beds. ➡ *Given the arguments for an eastern source, how might these beds fit into the larger depositional framework?*

20.5 Turn left on VT 125. Proceed east to Middlebury. Classic karst features occur in the Champlain Valley. In Weybridge, a few miles to the north, an extensive cave system in Ordovician limestones illustrates the high solubility and consequent large, erratic porosity of these carbonates.

26.6 Turn right on VT 7. The mountain front to the east marks the boundary between the Cambro-Ordovician carbonate platform of eastern Laurentia and distal parts of the Laurentian margin, which have suffered considerably more deformation and metamorphism.

30.3 Turn left on VT 125 towards the mountain front.

30.9 Cross VT 116 near East Middlebury.

32.0 Cross bridge over Ripton Gorge and park on far side. Exercise caution returning by foot across narrow bridge. Descend to outcrops in stream below bridge.

STOP 4: BASE OF PASSIVE MARGIN, Ripton Gorge, VT.

Passive margins respond to sea level variations and host the principal records of pre-Mesozoic global change. They are sites of extreme fractionation, and major reservoirs of C (inorganic and organic), Ca, Mg, Si, and Al.

Thickly bedded, white orthoquartzites of the basal Cambrian Cheshire Formation crop out along the Middlebury River. Upstream, these beds rest unconformably on arkosic beds of the Pinnacle Formation, which in turn rest with profound nonconformity on gneisses of the Precambrian Mount Holly complex. The arkosic beds are interpreted as rift facies. The gneissic basement forms part of the Green Mountains Grenville Inlier.

Recent work on the New Jersey passive margin has shown an excellent agreement between sedimentary sequences and glacioeustatic cycles (Miller et al., 1996). The sequence boundaries are unconformities sculpted during glacioeustatic sea level fall. Cambro-Ordovician strata of the Champlain Valley show a number of transgressive-regressive cycles. ➡ *Given that most of the Early Paleozoic is thought to have been ice-free, what caused these cycles?* The major transgression coincident with Cheshire deposition was probably global in extent (e.g., Random Formation on the Avalon Platform in Newfoundland). Furthermore, orthoquartzite deposition was also widespread. Nicholas (1996) correlates a decrease in marine $^{87}Sr/^{86}Sr$ ratios with a surge of seafloor-spreading at this time.

The Laurentian passive margin is composed of materials that were produced by extensive weathering. Three orthoquartzite wedges in the stratigraphic section east of the Champlain thrust merge westward into the Potsdam Formation that transgressed Laurentia through the Cambrian. When clastic sediment was absent, the warm shallow waters were ideal for carbonate deposition. ➡ *What is the relative importance of tectonic setting, climate, and PCO_2 (which is thought to have been high through this interval) in producing compositionally mature sediment?* ➡ *Does supermature sediment imply, in addition to long residence time in a hot and wet weathering realm, high pCO_2 in order to completely weather the feldspars?*

The distribution of dolomite in the Champlain Valley sequence prompts questions about the Mg cycle. Cambrian carbonates are essentially all dolostones. Lower Ordovician carbonates include a mix of limestone through dolostone. The top of the sequence is nearly pure limestone. Dolomite clasts in limestone indicate penecontemporaneous formation of dolomite, but the age of the massive dolostone units is difficult to constrain (Reusch, 1976). Ultimately, most Mg is derived from weathering of mafic minerals such as olivine and pyroxene. ➡ *Why are the oldest but not the youngest carbonates in the Champlain Valley dolomitized?* ➡ *Is dolomite distribution simply facies controlled? Or does the decreasing abundance of Mg carbonate reflect diminishing Mg sources?*

32.0 Continue east on VT 125. We cross over an inlier of Grenville basement in the vicinity of the town of Ripton.

37.6 Breadloaf Campus of Middlebury College.

40.2 Middlebury Gap. On both sides of the Grenville basement, the mantling metasedimentary rocks, here assigned to the Hoosac Formation, are interpreted as rift facies. Along Interstate 89 (on strike to the north), these rift units are composed of metamorphosed coarse sediment locally riddled with black amphibolites (dikes indicating extension).

43.4 Turn left on road to Texas Falls.

43.9 Park at Texas Falls.

STOP 5: CORE OF TACONIC OROGEN, Texas Falls, VT.

The metamorphic core of the Taconic orogen contains high pressure mineral assemblages (Laird et al., 1984). The exhumation of these once deeply buried materials was linked with orographic precipitation, clastic sedimentation in the foreland (and overall style of the foreland), and C fluxes (degassing, organic C weathering, silicate weathering, and organic C burial).

The rocks at Texas Falls are green chlorite-muscovite-quartz schists. Magnetite is locally present. Dismembered and folded quartz veins are common, and indicate, along with rare isoclinally folded beds, extreme deformation. Schistocity strikes just east of north and dips approximately 50 degrees to the east, consistent with a position on the east flank of the Green Mountain Anticlinorium (GMA).

Rocks on the east flank of the GMA are generally interpreted as continental margin deposits later incorporated into the Taconic accretionary prism (e.g., "polydeformed and metamorphosed mixed assemblage at the precarious east edge of ancient North America" of Williams, 1978). Sedimentary C would have been released during growth of this accretionary prism as wet sediments progressively lost their volatile component. Modern analogs, e.g., the accretionary prism exposed in Barbados, host CO_2 springs. Subsequently, the exhumation of the deeply buried metamorphic rocks implies CO_2 consumption, via silicate weathering, and accelerated pace of the organic C cycle.

➡ *What is the relationship between the quartz veins and dehydration and release of C from the precursor sediments?* ➡ *How does the timing of quartz veining fit into the sequence of deformation?* ➡ *Considering that inorganic and organic C are affected by different reactions as they move through a subduction system (i.e., organic C may be oxidized and returned to the surface as CO_2 at a higher level and earlier than inorganic C), what might be the consequences for large-scale fractionation of C isotopes?*

The locus of high pressure Taconic metamorphism occurs on the east flank of the GMA (e.g., blueschists described by Laird et al., 1984). The occurrence of high pressure metamorphic rocks at the surface of the Earth implies considerable exhumation. If exhumation was achieved largely by erosion (as opposed to tectonic denudation via detachments), then the occurrence of these rocks also implies considerable silicate weathering and CO_2 consumption, especially where arc-ophiolite assemblages capped the structural pile. Significantly, most of the world's orogens preserve only small erosional remnants of ophiolites that are inferred to have been much more extensive in the past. Contemporaneous rainfall would have been a significant factor in determining the relative amounts of erosional and tectonic denudation. ➡ *What evidence can be used to infer paleo-rainfall over the Taconic orogen?* ➡ *Were sources of moisture available before the system evolved into an intracontinental suture?* ➡ *Do estimates of paleolatitudes place the orogen near equatorial rain belts?* ➡ *Could a positive feedback have developed between uplift and orographic precipitation soon after the young orogen began to rise?*

44.3 Turn left on VT 125.

47.3 Hancock. Turn right on VT 100 and proceed south.

48.2 View east of rugged terrane underlain by ophiolitic mélange. Knobby terrane typically characterizes mélanges made of resistant blocks enclosed in matrix of less resistant material. Generally, VT 100 follows the U.S. extension of the Baie Verte - Brompton Line interpreted as the fundamental suture between Laurentian rocks and accreted terranes (Williams and St. Julien, 1982). Of interest to historians of science, the terrane view of orogens put forth by Coney et al. (1980) has roots here in this part of Vermont: "...in the context of current plate tectonic theory, once one takes a single step east of the Green Mountain anticlinorium everything to the Bay of Maine is of suspect geo-political allegiance" (Coney et al., 1972; Williams and Hatcher, 1983).

49.8 Turn left (east) on Quarry Road.

50.3 Stay left on main road.

51. 3 Park above Vermont Verde Antique International quarry.

STOP 6: OPHIOLITE SUTURE, Rochester, VT.

A small serpentinite body is actively being quarried by Vermont Verde Antique International. PLEASE EXERCISE EXTREME CAUTION. THE OPERATOR HAS GRACIOUSLY OFFERED TO LET US EXAMINE BLOCKS FROM THE QUARRY. IF THE CRANE IS NOT IN USE, WE MAY BE ABLE TO EXAMINE THE SPECTACULAR EXPOSURE JUST EAST OF THE QUARRY. IF YOU HAVE A HARD HAT, ADDITIONAL ACCESS MAY BE PERMITTED.

Ultramafic and mafic rocks contain abundant Ca and Mg silicates. High temperatures associated with their genesis promote reactions that drive CO_2 into the atmosphere. Following generally tortuous histories, in the weathering environment these materials supply Ca and Mg ions which combine with atmospheric CO_2 to make seafloor carbonates.

Blocks in the quarry area display the variety of lithologies found in the ultramafic body. Some of the blocks are run through with a network of carbonate-talc veins, which weather light tan. Massive chromitite occurs in one block.

The contact between the serpentinite and overlying chlorite-epidote greenstone is well exposed on the east side of the quarry (Sanford, 1982). A 2 m thick zone of white-weathering rock dips moderately east and separates the ultramafic from the mafic rocks. The bottom meter is dominantly veined carbonate-talc-actinolite material and the top meter is a massive actinolite schist. Beneath this white layer, the top of the serpentinite body is sheared (platy, fish-scale serpentinite). The top contact of the white layer is also sheared, forming a talcy schist.

In Berner-type C cycle models, global degassing is estimated according to seafloor-spreading rates, assuming that most thermal release of C takes place at spreading centers and subduction zones. Appalachian ophiolites may not be representative of Iapetan crust and mantle, but they do illustrate the processes associated with generation of lithosphere. Direct release of mantle C may take place at spreading centers, although the abundance and distribution of C in the mantle are not well known. Additionally, spreading centers affect the C cycle indirectly via hydrothermal circulation. Seawater acidifies when it

passes through the crust due to the oxidation of reduced material, e.g. iron sulfide (pyrite). Upon release to the ocean, the acid waters cause the carbonate equilibria to shift by the conversion of carbonate to bicarbonate and of bicarbonate to carbonic acid, which exsolves CO_2 to the atmosphere.

The sedimentary cover of Paleozoic ophiolites is either siliciclastic material or chert. The calcareous nannoplankton deposits that cover Cenozoic oceanic crust are absent. Martin (1995) implies that higher pCO_2 during the Paleozoic did not permit growth or preservation of calcite plankton in the open ocean. This relationship is a corollary of the generally accepted solution to the "faint sun paradox" (i.e., to maintain a generally uniform mean temperature through time, greenhouse warming must have been stronger in the past; during the Early Paleozoic, the sun was 4.5% fainter than it is now).

Ultramafic rocks, due to their relatively low silica contents, are highly reactive when emplaced into crustal settings. Steep chemical gradients exist between peridotites and silica-rich surroundings, which cause extensive alteration zones to develop at the margins of these rock bodies. One style of reaction involves the production of talc-carbonate assemblages, making this process a sink for CO_2. ➡ *What is the likely source of the C trapped in these rocks?* Sanford (1982) has shown that much of the mineral and element distribution is explained by diffusion across the steep chemical gradients, with Mg migrating outward from the ultramafic body and Si and C migrating inward. He thought that this metasomatic redistribution of elements took place during Devonian metamorphism because it overprints the serpentine assemblage.

In the weathering environment, fresh ultramafic rocks are highly soluble. Olivine and pyroxene, which dissolve congruently, are efficient C sinks, consuming one mole of CO_2 per mole of Mg (or Ca). These materials are so reactive that locally groundwaters have been entirely stripped of C species, with carbonate deposited within the rock (as opposed to the more typical fate of carbonate, which is to be incorporated into marine tests). The resulting spring waters are dilute Ca-OH solutions with very high pH (Barnes et al., 1978; Drever, 1988). Elsewhere, waters derived by interaction with serpentinites have been reduced, locally giving off hydrogen gas. Reactions between CO_2 and these minerals are exothermic, and Klaus S. Lackner of Los Alamos has suggested that CO_2 might actually be used as an energy source (Beardsley, 1995).

51.3 Backtrack on Quarry Road.

52.8 Turn left (south) on VT 100.

53.8 Outcrops of greenstone on left.

54.7 Rochester. Turn left on Bethel Mountain Road (Rochester Mt. Rd. coming from Bethel).

56.0 Turn left (stay on Bethel Mountain Road).

58.8 Park on right side of road. Cross road to outcrops.

STOP 7: PRE-SILURIAN BASEMENT, Rochester Mountain.

The pre-Silurian basement beneath the Taconic unconformity includes marine sediments (C reservoirs) and volcanic materials, whose genesis was associated with CO_2 degassing.

These outcrops consist of green chloritic schists and rusty-weathering pyritic-graphitic schists, formerly black shales. The ages of these units are difficult to infer. The age, correlation, and origin of the graphitic units will be topics for discussion.

Volcanic rocks of the Lower Ordovician Moretown Formation lie a few miles to the east. The Moretown rocks are interpreted as an arc terrane. At the time of their eruption, the arc system was a source of CO_2. ➡ *What igneous and deformational processes within an arc system enable CO_2 to escape from rock reservoirs at depth to the atmosphere?* ➡*Prior to Mesozoic time, how would the absence of carbonates on the seafloor have affected the C cycle?*

Of fundamental importance to the C cycle is the fate of arcs following collision with passive margins. ➡ *Are they preserved somehow or are they destroyed by erosion when the upper plate of the arc system is elevated out of the ocean by underthrusting of buoyant sialic crust?* Uplift of arcs following collisions seems almost unavoidable. ➡ *Do extensional detachments develop, translating the volcanic carapaces to sites of preservation and exposing the plutonic levels of the arcs?* At any rate, the Taconic unconformity (Stop 8), sculpted following aborted subduction of the Laurentian passive margin, extends the full length and, arguably, nearly the full width of the Appalachian orogen.

59.3 Long roadcut, similar to previous stop (note 50 foot layer of graphitic schist; also minor hematitic schist). Note the generally lower and flatter topography to the east that characterizes the Siluro-Devonian metasedimentary terrane of the Connecticut Valley - Gaspe Synclinorium.

64.8 Turn left (north) on VT 12.

66.0 Turn right just before bridge across Gilead Brook onto dirt road. Park and walk north to the brook.

STOP 8A: SILURIAN COVER, Bethel, VT.

The Silurian cover sequence contrasts sharply with the Ordovician basement. The development of an extensive erosional surface carved deeply into arc terranes (formerly sites of degassing) implies silicate weathering and CO_2 consumption, which may have contributed to Late Ordovician global cooling.

Silurian cover and Ordovician basement rocks are well represented in stream boulders. Brown weathering calcareous quartzites (or silty limestones) are from Silurian strata. Light-weathering amphibolites with a strong fabric are from the Ordovician basement. The contact between these units is shown to lie near the VT 12 bridge just upstream (Doll et al., 1961). However, it is not easily located and neither is it easy to assess the extent to which the unconformity has been modified by Acadian events. (Note: The first exposures in the stream are metamorphosed Devonian turbidites, related to the Acadian orogeny.)

➡ *What implications does the strong contrast in composition between Silurian cover and Ordovician basement (e.g. quartzite on basalt) have for the C cycle?*

The extent, both along and across strike, and "depth" (i.e. amount of basement removed) of the Taconic unconformity bear on estimates of C fluxes that may have affected Late Ordovician climate. Silurian unconformities have been documented across nearly the full width of the Appalachian Orogen (Pavlides, 1968; Rodgers, 1971; Williams, 1978). ➡*How many of these are genetically related to the specific mechanism of Taconic collision with Laurentia?* In Maine, a Silurian unconformity is widespread, from the Jackman area (Silurian on Ordovician Attean Pluton) to coastal Maine (Silurian on basalts at Ames Knob, North Haven Island). However, in the Aroostook-Matapedia belt, sedimentation was apparently continuous and of a carbonate-rich nature through the Late Ordovician, although the margins of this basin received detritus. ➡ *The Ames Knob site is probably not a part of the Taconic orogen, based on faunal arguments, but if it is, what implications does this have for later strike-slip movements and placement of inboard elements outboard?*

The Silurian locally contains very red shales and sandstones (red beds). A positive carbon isotopic shift in the latest Ordovician suggests anomalous organic C burial and production of oxygen (O_2) and, perhaps, stratospheric ozone (O_3). ➡ *Is the implied decreased ultraviolet radiation, due to O_3 buildup, a viable explanation for the subsequent colonization of land by plants during the Silurian?*

66.0 Turn left (south) on VT 12.

70.4 Bethel. Turn left (east) on VT 107.

71.3 Turn left (north) on Christian Hill Road.

72.1 Turn left (northwest) on Davis Hill Road.

72.6 Park by outcrops on right.

STOP 8B: SILURIAN COVER, Bethel, VT.

Outcrops of shale and calcareous quartzite as in brook boulders.

72.6 Return to Killington-Pico region. Take VT 107 west to VT 100, turn left (south) to U.S.4.

APPENDIX A

ISOTOPE GEOCHEMISTRY AND THE CARBON CYCLE

The major isotopic fractionation associated with photosynthesis causes organic matter to be depleted in ^{13}C relative to ^{12}C. Variations in the C isotopic composition of seawater, inferred from calcium carbonate tests, are due to either 1) imbalances in the organic and inorganic fluxes (e.g., if the input flux is constant, seawater becomes enriched in ^{13}C when the organic C burial flux is increasing relative to the carbonate burial flux) or 2) variations in net fractionation (e.g., seawater becomes enriched in ^{13}C with more overall fractionation). Photosynthetic fractionation tends to be greater at high pCO_2 because plants can afford to be more selective.

Marine $^{87}Sr/^{86}Sr$ and $^{187}Os/^{188}Os$ ratios also bear on the geochemical C cycle because Sr and Os participate in reactions affecting C (weathering, hydrothermal exchange, dissolution of sediments). For these heavy isotope systems, the isotopic composition of the water that has interacted with a particular mineral assemblage depends largely on the parent/daughter ratios and ages of the minerals and the relative solubilities of the minerals. Table 5 summarizes effects of different C fluxes on pCO_2 and isotopic compositions. Table 5a contrasts the effects on pCO_2 and marine $^{87}Sr/^{86}Sr$ ratios of weathering soluble Ca and Mg-rich materials found in arc-ophiolite assemblages and of weathering less soluble sialic materials making up continental crust. Table 5b is an expanded version of Table 5a.

Carbon flux	Effect on pCO_2	Effect on $^{87}Sr/^{86}Sr$
weathering of sialic crust (J_{wl})	weak CO_2 sink (decrease)	increase
weathering of arc/ophiolite (J_{wa})	strong CO_2 sink (decrease)	decrease/flatten

Table 5a. Effects of silicate weathering on pCO_2 and marine $^{87}Sr/^{86}Sr$ ratios. Note J_{ws} is split into two component fluxes, J_{wa} (a for sima) and J_{wl} (l for sial).

C flux	Effect on CO_2	on $\delta^{13}C$	on $^{87}Sr/^{86}Sr$	on $^{187}Os/^{188}Os$
J_{mg}	increase	?	none	?
J_{mc}	increase	?	?	none
J_{bg}	decrease	increase	none	none
J_{bc}	none (long term)		none	none
J_{wg}	increase	decrease	none	increase if old
J_{wc}	none (long term)	weak increase	flatten	none
J_{wl}	decrease	none	increase	increase
J_{wa}	decrease	none	decrease/flatten	?

Table 5b. Carbon fluxes vs. effects (expanded version of Table 5a).

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GLACIAL GEOLOGY OF THE CENTRAL CHAMPLAIN VALLEY, VERMONT

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INTRODUCTION

This fieldtrip visits sites that illustrate the glacial and post-glacial deposits and history of the central Champlain Valley on the Vermont side. It is based on existing studies and information, generally conducted by numerous researchers in the 1930s (Chapman), 1970s (Wagner, Connally, others), and various graduate students throughout the decades. See the 1972 NEIGC guidebook for several pertinent articles and road logs.

The Champlain Valley was completely covered by Laurentian ice, which produced till, a few short sections of eskers, and striations on bedrock surfaces. During the receding stages of the ice, glacio-fluvial features such as kame terraces and kames were deposited. An extensive kame terrace is found along the east side of Rte. 116, from south and east of Hinesburg village southward to Starksboro village (and beyond). Kames are found at the north end of this kame terrace. As the ice was receding, meltwater was prevented from flowing northward due to the presence of the ice lobe in the northern end of the Champlain Valley, so a large pro-glacial lake was formed in the Champlain Valley, dammed by glacial deposits in the Glens Falls, N.Y. area (see 1972 NEIGC guidebook, and others, for details). Deltas formed at river mouths into the several stages of this lake (called Lake Vermont), which was much larger than present-day Lake Champlain. Bottom sediments were deposited, to depths of many tens of feet. Spillways formed in saddles, from smaller pro-glacial lakes ice-dammed in the hills to the east flowing into Lake Vermont or other pro-glacial lakes. When the ice finally receded north of the St. Lawrence River, salt water from the Atlantic Ocean entered the valley because the north end of the valley had been depressed below sea level by the weight of the ice. This bay of the Atlantic, called the Champlain Sea by geologists, also produced deltas and bottom sediments, and contained abundant marine life (one example of which is the whale skeleton seen in the Perkins Museum of Geology at UVM). Isostatic rebound eventually caused the north end of the Champlain Valley to rise above sea level, isolating the basin. Freshwater inputs rapidly produced a freshwater lake, the current Lake Champlain, which drains northward.

ROAD LOG

Stop #1: Mt. Philo State Park (if possible), for view of entire southern Champlain Valley;
Location: 0.6 miles east of Rte. 7 between North Ferrisburgh and Charlotte, at State Park Rd. and Mt. Philo Rd..

Stop #2: Shelburne "Two-till locality", to see:
* lodgement till;
* striations on bedrock.

Location: from Stop #1, return to Rte. 7 and go north for 5.6 miles, through Charlotte, to the Countryside Motel (east side of Rte. 7, 1 mile south of Shelburne village). Ask permission to park at motel, or park in pulloff on west side of Rte. 7 located 0.1 miles north of motel. "Two-till locality" is the steep eroded face of till on the northeast flank of McCabes Brook immediately west of the motel. At least two cross-cutting directions

of striations are seen on the bedrock surface exposed along the southwest edge of the brook, 100 feet downstream from the till face.

Stop #3: Lake or marine bottom sediments.

Location: Palmer or LaBerge farm fields, on either side of Thompson's Point Road, Charlotte. From Stop #2, go south on Rte. 7 for 5.5 miles, almost back to State Park Rd. (which goes east to Mt. Philo State Park). Turn west (right) on Thompson's Point Road. Ask permission at the Palmer farm (large white farmhouse on Rte. 7 located 2 houses south of Thompson's Point Rd., or at the LaBerge farm -- the first farmhouse on Thompson's Point Rd. west of Rte. 7 on the north side, at the corner with Greenbush Rd. Dig holes in any of the farm fields. Set topsoil plug aside; after re-filling hole with deeper soil, replace the topsoil plug.

Stop #4: Lake Vermont delta, built out over pre-existing kame terrace.

Location: Hinesburg Sand and Gravel, Inc., gravel pit, Rte. 116, in South Hinesburg. From Stop #3, go north to Charlotte, turn east (right) on the Charlotte-Hinesburg Rd. to Hinesburg village, turn south (right) on Rte. 116, go 3.3 miles south on Rte. 116 to South Hinesburg. Gravel pit is 0.1 miles south of South Hinesburg, on east side. Ask permission at the scale house to drive in and look. Go only where directed, and do not get out of vehicles -- federal mining regulations are ***very*** strict about unaccompanied visitors -- infractions could subject the owner to heavy fines, and would threaten his decades-long cooperation with visiting geologists.

On the way out, pull in to the usually-unmanned smaller state gravel pit located at the extreme western edge of the delta, behind the wooden equipment shed. If state transportation workers are present, ask permission. If not, be cautious, inspect the deposit only from the floor of the pit, and do not climb the pit face (dangerous overhangs).

This huge delta was created by the Winooski River (diverted by an ice dam from its "preferred" channel further north) flowing into Lake Vermont.

Stop #5: Kame terrace.

Location: east of Rte. 116, located 0.4 miles north of South Hinesburg, or 2.3 miles south of Hinesburg village. From Stop #4, go north on Rte. 116 for 0.4 miles, to a small private dead-end dirt road that heads east (right) up a steep hill from Rte. 116 just beyond a long wooden equipment shed and barn. Looks like a farm road or driveway. Drive up to the open field, about 0.2 miles east of Rte. 116. Ask permission of the farmer on Rte. 116, or the homeowners beyond the field. The fields are the top surface of a very large kame terrace that extends, with several large gaps, about 3 miles to the north (almost to Rhode Island Corners), and about 5 miles to the south (almost to Starksboro village). If permission is granted, walk east to the contact between the eastern edge of the kame terrace and the till-veneered hillside.

Stop #6: Kames.

Location: northeast of corner of Hinesburg-Richmond Rd. and Texas Hill Rd. From Stop #5, go north on Rte. 116. In 0.3 miles, coninue straight north on North Rd. where Rte. 116 curves sharply to the west. Continue north on North Rd. At about 1.7 miles north of Rte. 116, look for small gravel hills on either side of road. At intersection with Texas Hill Rd. and Hinesburg-Richmond Rd., observe the small hills northwest of this intersection. All of these sandy-gravelly hills previously mentioned are kames.

Stop #7: Tiny Lake Vermont delta.

Location: south side of Butternut Lane, 0.7 miles east of Rte. 2A in St. George. From Stop #6, go west on the Hinesburg-Richmond Rd. through Mechanicsville to Rte. 116, turn north (right) on Rte. 116, go north to Rte. 2A, turn right (north) on Rte. 2A, go about 2.2 miles north to Butternut Lane, turn east (right) on Butternut Lane, go about 0.7 miles on Butternut Lane to a narrow pullover on the south (right) side opposite a driveway. Ask permission at the house up the driveway. The deltaic surface is at the top of the streambank to the south; scramble up the steep hill to the east of the overhead utility lines. Back down at the bottom, find the lowest sands of the delta in the vertical streambanks just above the water surface, and the lake bottom sediments in the streambed a few tens of feet downstream from the pullover.

Stop #8: Spillway.

Location: South of West Oak Hill Rd., Williston. From Stop #7, go west on Butternut Lane back to Rte. 2A, turn north (right) on Rte. 2A, go north 1.6 miles to West Oak Hill Rd. Turn east (right) on West Oak Hill Rd., go about 1 mile (past Sunset Hill Rd. on the right) to a small gravel pit and construction yard on the right. Park, walk south over a small rise to the north edge of the spillway (flow was from east to west). Return to vehicles, continue east on West Oak Hill Rd., observing the spillway through the trees to the south (right). Its mouth is at the intersection of West Oak Hill Rd. and Oak Hill Rd. Notice the lag boulder pavement here.

Stop #9: Huge Champlain Sea delta (Winooski River).

Location: Sand pit east of Redmond Rd., Williston (owned by Hinesburg Sand and Gravel Co.; may be the site of a lined landfill in future years). From Stop #8, return west to Rte. 2A, and turn north (right). Go north past I-89 and through Tafts Corner (*see Windshield Site #9A below, beginning at Tafts Corner)* to the stoplight at Mountain View Rd., turn east (right) on Mountain View Rd, go east for 0.9 miles to Redmond Rd., turn north (left) on Redmond Rd., go north for about 1.1 miles. Just before the landfill gate is a dirt entrance road to a huge sandpit, often blocked with sandpiles to exclude vehicles. In 1997, the sandpit is not operating, and people on foot or bicycles enter at their own risk. Observe the topset and foreset beds, numerous ripple marks, and crossbedding. Stay away from banks with steep overhangs or large vertical relief.

Windshield Site #9A: Sand Dunes.

Location: Along both sides of Rte. 2A between Tafts Corner and Mountain View Rd., Williston. As you are driving to Stop #9, notice the small hills in the lawns of the commercial establishments and residences, and the cemetery (possibly?) along this section of Rte. 2A. These are sand dunes, blown from the exposed deltaic surfaces nearby.

THE FOLLOWING STOPS MIGHT HAVE TO BE DELETED, DUE TO TIME CONSTRAINTS:

Stop #10: More of the same huge delta (Winooski River into Champlain Sea).

Location: Airport Parkway, at north end of Burlington Airport runways. From Stop #9, return to Rte. 2A using Redmond Rd. and Mountain View Rd. Turn north (right) on Rte. 2A, go north 0.5 miles to River Cove Rd., turn west (left) on River Cove Rd., which becomes Poor Farm Rd. at some point. At 2.5 miles from Rte. 2A, Airport Parkway turns sharply west (right). Follow it 0.3 miles to a sandy pullover on the south (west) at the end of the runways. Observe the enormous flat surface to your south and west, and across the Winooski River to the north (where is found St.

Michael's College, the Fanny Allen Hospital, and Fort Ethan Allen), all of which is the deltaic surface formed by the Winooski River flowing into Lake Vermont.

Stop #11: Armored mudballs.

Location: In sandy face behind commercial buildings on north side of Rte. 2 between Industrial Avenue and Brownell Rd (may be gone in the future, as development continues). From Stop #10, work your way back south to Rte. 2 however possible, turn east, go 0.2 miles east of the intersection of Industrial Ave. and Rte. 2; look for sand banks behind commercial buildings.

Stop #12: Esker.

Location: North of Poker Hill Rd., Underhill Flats. From Stop #11, go east on Rte. 2 to Tafts Corner, then north on Rte. 2A to Essex Junction, then east on Rte. 15 to Underhill Flats. Beyond the village of Underhill Flats, turn northeast (right) on Poker Hill Rd., go about 0.2 miles. The esker is on your left about 200 yards from the road; covered with conifers; quite short (a few hundred yards); has a split north end. Ask permission of landowner.

Stop #13: Spillway.

Location: East of Rte. 15 from Underhill Flats, northward for about 3 miles to the height of land in the saddle. From Stop #12, return to Rte. 15, turn north (east; right). The spillway, which drained a major pro-glacial lake in the Lamoille River Valley southward into a lake in the Underhill-Jericho area, is easily seen just east of Rte. 15 for the next 2 to 3 miles. The mouth of the spillway was at the saddle (height of land of valley floor). The lag boulder pavement is extensively obscured by flooding from beaver ponds.

Stop #14: Charlotte the Whale.

Location: Perkins Geology Museum, University of Vermont. From Stop #13, return to Burlington by a variety of routes. Perkins Museum is in the Geology Dept. building on the north side of the UVM campus (off of Pearl Street, behind Ira Allen Chapel). Call for open times. Charlotte was a beluga whale excavated by railroad workers in the 1800s from a clay bank near Stop #3. Named after the town her skeleton was found in, she lived in the Champlain Sea, which was a bay of the North Atlantic Ocean that occupied the Champlain Valley after the ice receded north of the St. Lawrence River.

A TRANSECT ACROSS THE CONNECTICUT VALLEY SEQUENCE IN EAST-CENTRAL VERMONT

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Introduction

The origin, stratigraphic succession and subsequent style of collapse of the Silurian-Devonian Connecticut Valley sequence (CVS; also referred to here as the "Vermont sequence") in eastern Vermont has been debated for nearly one hundred years. Although many different studies have focused upon either local or regional structural and stratigraphic relationships, all of these can basically be grouped into two different models regarding the origin of the CVS and the succession of the different mapped formations. The first model proposes the Connecticut Valley - Gaspé synclinorium (Doll and others, 1961) in which Silurian-Devonian rocks on the east flank represented the eastern equivalent of the oldest strata that were unconformably on top of pre-Silurian rocks on the west flank, along the Richardson Memorial Contact (RMC). The pre-Silurian rocks on the west side of the synclinorium included the Moretown, Cram Hill and Barnard Members of the Missisquoi Formation which were correlated with eastern equivalents mapped as the Ordovician Albee, Ammonoosuc and Partridge Formations of the Bronson Hill anticlinorium (Figure 1; Billings, 1937, 1956). Because of the striking lithologic and stratigraphic similarities between the two different sequences (herein referred to as the Vermont and New Hampshire sequences), most workers agreed with the correlation. The nature of the boundary between the Silurian-Devonian part of the Vermont sequence (rocks within the core of the CVS) and the western limit of rock-types mapped as the New Hampshire sequence was, and still is, an item of intense debate. Some workers believed the contact was an erosional unconformity marked by the presence of the late Lower Devonian Putney Volcanics at the eastern extent of the Vermont sequence, and discontinuous quartz-pebble conglomerate at the base of the latest Lower Devonian Littleton Formation of the New Hampshire sequence (Hepburn and others, 1984, Trzcienski and others, 1992). Where mapped, this contact has become known as the "chicken yard line", after an exposure of the contact within a now-abandoned chicken yard in the town of Dummerston, Vermont (Hepburn and others, 1984). The other school of thought considers this contact to be either a pre-metamorphic thrust fault, which places older Littleton Formation of the New Hampshire sequence westward over younger Gile

Mountain and Waits River Formations of the CVS (the Whately Thrust of Robinson and others, 1984), or a syn- to post-metamorphic fault between different stratigraphic units in both the Vermont and New Hampshire sequences (the Monroe Line of Hall, 1959; Hatch, 1988a).

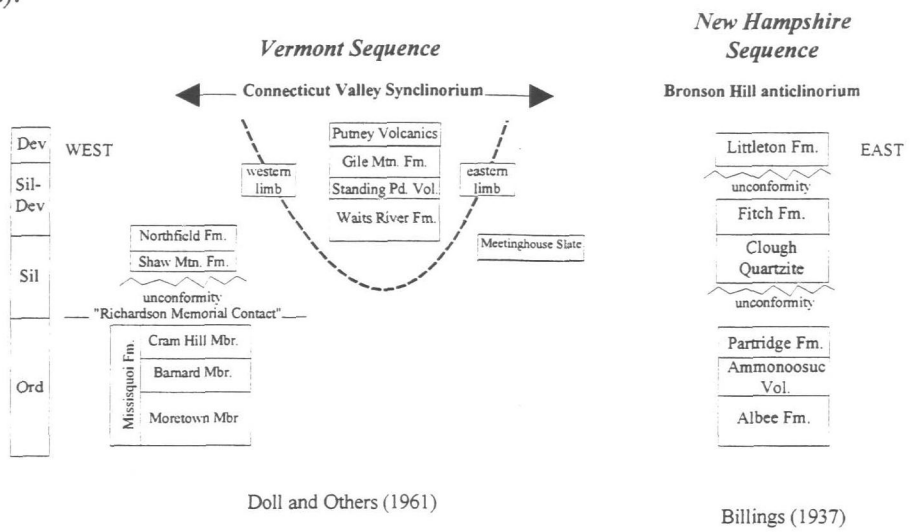

Figure 1. Stratigraphic correlation diagram for the Vermont and New Hampshire sequences. Vermont sections from Doll and others (1961); New Hampshire section from Billings (1937 and 1956).

The second model regarding the origin of the CVS, presented by Hatch (1987, 1988b), suggested that the CVS (or Connecticut Valley trough, CVT, as used by Hatch in these reports) displays depositional contacts with graded beds indicative of a stratigraphic succession younging towards the flanks of the sequence, and not towards the center, as proposed in the synclinorium model (Figure 2). The eastern and western flanks, represented by the Meetinghouse Slate Member of the Gile Mountain Formation, and the Northfield Formation, respectively, are actually younger than the Waits River Formation and parts of the Gile Mountain Formation that constitute the core of the of the CVS, or Connecticut Valley trough. Hatch thus interpreted the Connecticut Valley Trough as an anticlinorium rather than a synclinorium. This required fault contacts on both flanks, including a thrust fault along the RMC (the Dog River Fault Zone of Westerman, 1987) and a fault with protracted history along the eastern contact with the New Hampshire sequence (the Monroe Line of Hall, 1959, and Hatch, 1988b).

The purpose of this trip is to introduce new data regarding the structural, stratigraphic and metamorphic evolution of the CVS and adjacent New Hampshire sequence units that can be used to evaluate the previous models for the origin of the CVT. Since the western boundary of the trough (the RMC) will have been covered on Saturday's field trip by Ratcliffe and Armstrong on the Chester and Athens domes, this trip will focus primarily on the trough rocks and the eastern boundary with the New Hampshire sequence units.

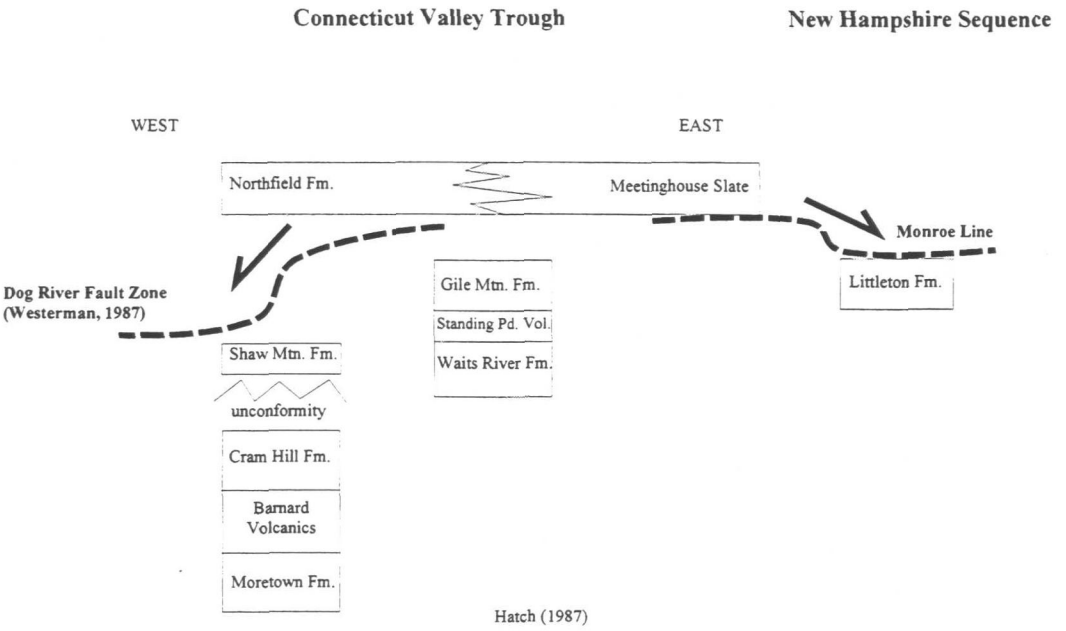

Figure 2. Stratigraphic correlation chart developed by Hatch (1987, 1988a), showing the Northfield Formation as a western facies of the Meetinghouse Slate and both stratigraphically above the Waits River and Gile Mountain Formations. Note the tectonic (faulted) nature of the boundary between the Northfield and Shaw Mountain units (the Dog River fault zone) and the boundary between the Meetinghouse Slate and Littleton Formation (the Monroe Line).

We shall discuss new 1:24,000-scale bedrock mapping and structural analysis conducted by Armstrong in the Bellows Falls, Walpole, Springfield and Saxtons River 7.5 x 15 minute quadrangles and Walsh in the northern part of the Springfield quadrangle, and the Mt. Ascutney 7.5 x 15 minute and Hartland 7.5 minute quadrangles (Figure 3), and ongoing petrologic studies by Spear in the same areas. Integration of these different disciplines will be used to focus on three interrelated thematic issues that will address the origin and subsequent destruction of the Connecticut Valley sequence, including:

1. The temporal and spatial relationships between attainment of peak metamorphic conditions during the Acadian Orogeny (illustrated by porphyroblast growth of index minerals such as garnet and staurolite) and Acadian deformational fabric development.

2. The distribution of Acadian pseudomorphic mineral textures in different lithologic units and structural levels, and their metamorphic and tectonic implications.

3. The palinspastic reconstruction of the CVS, including its sedimentary and volcanic history and its constraints on the present eastern extent of the Vermont sequence into areas previously believed to contain only New Hampshire sequence rocks.

Integration of results related to the three thematic issues will be discussed at most Stops and will expand as more information is gathered at the different localities during the course of the day. The goal of this methodology is to tie together the structural, lithologic and petrologic information into an internally consistent model that postulates the tectonic evolution of the Connecticut Valley sequence during the Acadian Orogeny. This, in turn, will help to define what we believe is the boundary between two very different tectonothermal regimes that were juxtaposed along a composite fault zone during, and possibly after, the Acadian Orogeny. This is the contact between rocks of what one could call "Vermont" and "New Hampshire" sequences.

Regional Geology

Bedrock in the Connecticut Valley sequence (CVS; Figure 3) and surrounding belts consists largely of, from west to east, Middle Proterozoic gneisses in the core of the Chester Dome, pre-Silurian metasedimentary, metavolcanic, and metaigneous rocks as a cover sequence immediately above the dome, Silurian and Devonian metasedimentary and metavolcanic rocks of the CVS, and Ordovician to Silurian and Devonian metasedimentary rocks informally referred to as the New Hampshire sequence (Figure 3; Billings, 1937, 1956; White and Jahns, 1950). In addition, the rocks are intruded by granitic dikes of the Devonian New Hampshire Plutonic Suite and, at Mount Ascutney, the Cretaceous White Mountain Plutonic - Volcanic Suite.

According to Doll and others (1961) the section above includes: 1) basement rocks -- Middle Proterozoic Mount Holly Complex; 2) eastern cover sequence -- Cambrian Cavendish and Hoosac Formations, undifferentiated Cambrian and Ordovician Pinney Hollow, Ottauquechee, and Stowe Formations, and Ordovician Missisquoi Formation; 3) Connecticut Valley sequence -- Silurian Shaw Mountain Formation, Devonian and Silurian Northfield Formation, Devonian Waits River and Gile Mountain Formations; and 4) New Hampshire sequence -- Ordovician Ammonoosuc Volcanics and Partridge Formation, Silurian Clough Quartzite and Fitch Formation, and Devonian Littleton Formation. The previous interpretation (Doll and others, 1961) maintains that the section of rocks from the Middle Proterozoic basement to the New Hampshire sequence is a non-faulted stratigraphic succession marked by three unconformities separating: 1) the Middle Proterozoic basement from the pre-Silurian cover sequence, 2) the pre-Silurian cover sequence from the Connecticut Valley sequence and 3) the Connecticut Valley sequence from the New Hampshire sequence.

Recent work to the west and southwest in the Cavendish, Chester, and Saxtons River quadrangles has redefined the relationships within the Middle Proterozoic rocks of the Chester and Athens Dome, the basement-cover relationships, and the internal structure

and stratigraphy in the pre-Silurian sequence (Ratcliffe, 1995a, b, c, in press, and Ratcliffe and Armstrong, 1995b, 1996, trip B-6 this report). The information shown on the State map by Doll and others (1961) represents the most complete work on the bedrock geology of the CVS prior to the recently published mapping (e.g., Ratcliffe and Armstrong, 1996; Walsh and others, 1996a and 1996b; Armstrong, 1997b) discussed in this report.

Rocks of the Connecticut Valley Sequence

The Connecticut Valley sequence consists of metasedimentary rocks of the Northfield and Waits River Formations, and metavolcanic rocks of the Waits River Formation. Also included in this sequence is a thin (0- to 10-m-thick) unnamed unit that is discontinuously exposed along the western side of the Northfield Formation, consisting of amphibolites, quartzite, granofels and schists. These rocks are very similar to rocks mapped in the Cavendish quadrangle (Ratcliffe, 1995a and in press) and are interpreted as a section of reworked volcaniclastic and metasedimentary rocks nonconformably overlying the Cram Hill Formation. In the Cavendish quadrangle this unit is interbedded at its top with calc-schists of the Waits River Formation (Ratcliffe, in press).

Metasedimentary Rocks

The majority of the metasedimentary rocks in the CVS are included within two lithologic groups: 1) gray carbonaceous schist and phyllite, and 2) gray carbonaceous schist and phyllite with interbedded impure siliceous limestone. A significant result of our recent 1:24,000-scale mapping is the separation of the carbonate-bearing from non-carbonate-bearing rocks in the CVS. The major criterion by which the rocks were separated is the presence or absence of the brown-weathering impure limestone beds, locally referred to as "punky" limestones. In addition to limestone beds, these rocks locally contain patchy-rusty-weathering, carbonaceous, calcite-bearing schists. This study groups rocks previously mapped as four formations (Northfield, Waits River, Gile Mountain, and Littleton of Doll and others, 1961) into two formations (Northfield and Waits River) based largely on the distribution of the limestone-bearing rocks.

On the Vermont State map (Doll and others, 1961) the Northfield Formation is shown as a continuous unit at the base of the Connecticut Valley sequence. Our mapping shows a discontinuous belt of gray carbonaceous schist and phyllite without limestone that varies from 0- to 260-m-thick along the western side of the CVS; the name Northfield Formation is applied to these rocks. The lower contact of the Northfield is interpreted as a nonconformity or disconformity based upon the discontinuous nature of Northfield units along the RMC and the presence of quartzite and quartz-pebble conglomerates and grits along the contact with the underlying Cram Hill Formation. This suggests that the Northfield is right side up and tops to the east. The quartzite and quartz-pebble conglomerates and grits crop out in three small separate bodies in the west-central part of the Springfield quadrangle. The units correlate in part with rocks mapped previously as Shaw Mountain Formation of Doll and others (1961), but we hesitate to use the name

Shaw Mountain because of the discontinuous distribution of the units and their uncertain correlation with similar rocks at the type locality in Northfield (Currier and Jahns, 1941). Evidence for the existence of a major fault at the base of the Northfield was not observed. The presence of a nonconformity, or a disconformity, and not a major fault, generally agrees with interpretations by Doll and others (1961) and Ratcliffe and Armstrong (1995b, 1996). This interpretation is inconsistent with recent findings by Hatch (1988b, 1991) and Westerman (1987) that place a major fault at the Northfield -- Cram Hill contact north of the Chester Dome.

The major metasedimentary units mapped as Waits River Formation include the carbonate- and non-carbonate-bearing gray schists and phyllites. Previous interpretations by Doll and others (1961) separated what we map as one formation (Waits River) into three formations (Waits River, Gile Mountain, and Littleton). The assignment of rocks to the Gile Mountain Formation was based on the assumption that rocks mapped as the Standing Pond Volcanic Member of Doll and others (1961) occur as a time-stratigraphic unit that separates older rocks with abundant limestone (Waits River) from younger rocks with little or no limestone (Gile Mountain). Our mapping indicates that limestones occur in roughly equal abundance on either side of the metavolcanic rocks and that it is not possible to separate two distinct formations based on the abundance of limestone-bearing rocks. Rhythmically bedded and graded sequences of gray phyllite and micaceous quartzite assigned to the Gile Mountain Formation, and reported by Fisher and Karabinos (1980) and Hatch (1988b) from areas north of this study, are absent in the quadrangles studied in this area by Armstrong and Walsh. For these reasons, we do not use the name Gile Mountain.

Doll and others (1961) show the gray schists and phyllites east of the easternmost belt of volcanic rocks (Figure 3) as Littleton Formation or undifferentiated New Hampshire sequence rocks. The contact between the volcanic rocks and the gray schists and phyllites to the east has been interpreted as an unconformity that separates the CVT (or Vermont sequence) from the New Hampshire sequence (Doll and others, 1961; Thompson and others, 1990; Thompson and others, 1993), and is informally referred to as the "chicken yard line" (Hepburn and others, 1984). Recent field trip articles (Boxwell and Laird, 1987; Trzcienski and others, 1992) refer to the location of the chicken yard line (CYL) unconformity at a roadcut 0.5 km west of the junction of I-91 and Route 11. At this CYL contact (our Stop 7), a conglomeratic quartzite and gray phyllite unit is exposed east of a greenstone. Other conglomeratic quartzite and gray phyllite and polymict conglomerate units, however, crop out both east and west of the CYL of Trzcienski and others (1992) at different places in the section. The gray phyllite and schist member is also lithologically similar on both sides of the contact. Our mapping shows that volcanic rocks, limestone-bearing schist and phyllite, and conglomeratic quartzite are present in a wide zone on both sides of the contact, suggesting that either there is no unconformity or that there may be several local disconformities. Generally, there is a decrease in the amount of metavolcanic, metavolcaniclastic and limestone-bearing rocks, and a corresponding increase in the amount of quartzite and conglomerate east

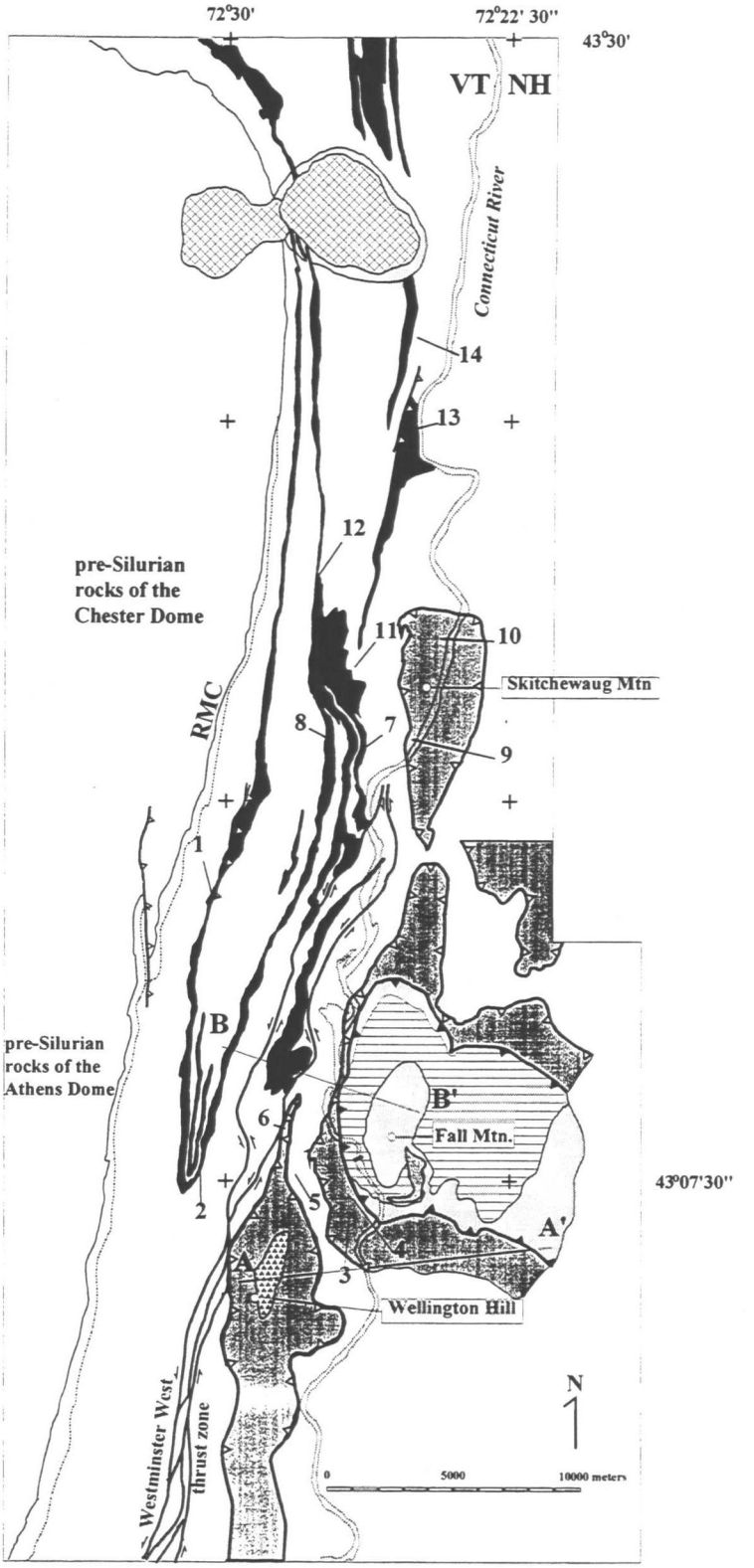

Figure 3. Generalized geologic map of the Connecticut Valley sequence and adjacent New Hampshire sequence. See explanation on next page.

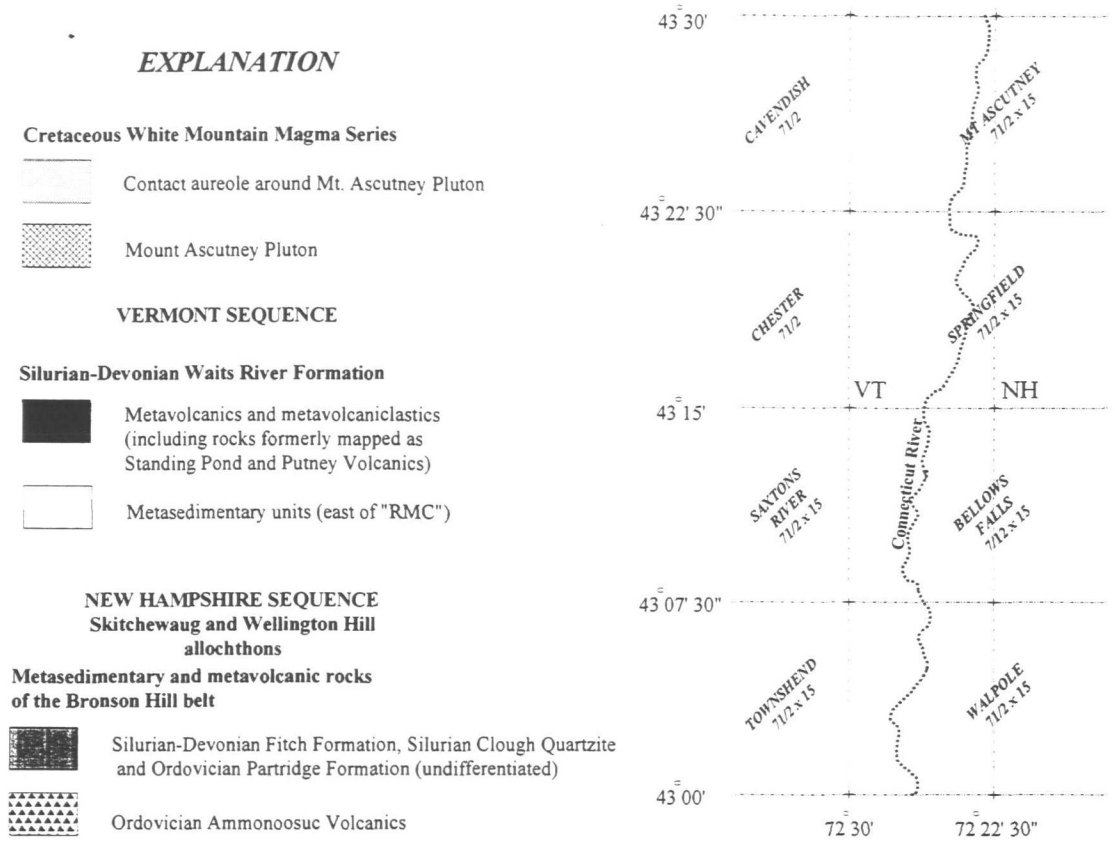

Figure 3 (continued). Explanation for geologic map

of where the CYL contact was previously defined but the transition cannot be assigned to a single horizon; this makes it difficult to assign regional significance to any particular contact. For these reasons, we interpret the rocks to form a continuous sequence devoid any of recognizable unconformities or evidence of hiatus in the depositional record. Thus, all of the units are included in the Waits River Formation. The use of the name Littleton for any part of this sequence implies that the Waits River Formation is in depositional contact with the Littleton Formation and would exclude the traditional New Hampshire sequence units, the Clough Quartzite and the Partridge Formation, from immediately below the Littleton. We have, therefore, excluded the usage of Littleton Formation for the easternmost carbonaceous schists in this area.

Metavolcanic Rocks

The metavolcanic rocks in the Connecticut Valley sequence consist of a heterogeneous assemblage of mafic and felsic volcanic rocks interbedded with volcaniclastic sedimentary rocks. The rocks were previously mapped as the Standing Pond Volcanic Member of the Waits River Formation (Doll and others, 1961), Putney Volcanics (Thompson and others, 1993), and even Ammonoosuc Volcanics of the New Hampshire sequence (Doll and others, 1961). The name Standing Pond amphibolite was originally applied by Doll (1944) to dark-green, fine- to medium-grained needle amphibolite at Standing Pond in the Strafford (VT) 15-minute quadrangle. Later, Doll and others (1961) used the Standing Pond Volcanic Member of the Waits River Formation to refer to all the volcanic and volcaniclastic rocks deposited, in their interpretation, between the Waits River and Gile Mountain Formations.

Our results indicate that all of the metavolcanic and metavolcaniclastic rocks are interbedded with the same pelitic metasedimentary rocks of the Waits River Formation, suggesting that they cannot be assigned to separate formations. For this reason, the metavolcanic and metavolcaniclastic rocks are mapped as unnamed members of the Waits River Formation. In accordance with Hatch (1991), therefore, we prefer not to use the name Standing Pond and instead refer to the rocks as unnamed volcanic rocks within the Waits River Formation until a more thorough evaluation can be made of the correlation with the rocks at the type locality of the Standing Pond (Doll, 1944).

The name Putney Volcanics (Hepburn, 1972; Trask, 1980) was introduced to separate the easternmost belt of Standing Pond as mapped by Doll and others (1961) in southeastern Vermont from other Standing Pond volcanic rocks to the west because it was considered "less mafic than the typical Standing Pond" (Trask, 1980, p. 133), and it could not be traced to the Standing Pond type locality. Our mapping shows that the felsic volcanic rocks, some of which would correspond to the Putney Volcanics as shown by Thompson and others (1993), do not occupy a single stratigraphic position. In addition, they are gradational with other metavolcanic rocks and metasedimentary rocks, and cannot be traced continuously to either of the type localities of the Standing Pond or the Putney. For these reasons the name Putney Volcanics is not used in this area.

On the State map, Doll and others (1961) show a small area of Ammonoosuc Volcanics northwest of Weathersfield Bow in the Springfield quadrangle. Our mapping shows that the rocks in that area belong to the felsic volcanic member of the Waits River Formation and are not part of the New Hampshire sequence. Recent work by Thompson and others (1993) also concluded that these rocks were not part of the Ammonoosuc Volcanics and called them Putney Volcanics instead.

Our mapping indicates that the metavolcanic rocks in the Connecticut Valley sequence can be subdivided into at least seven different units: laminated schist and granofels, large garnet and hornblende garbenschiefer schist, amphibolite, greenstone, hornblende-plagioclase gneiss, and felsic gneiss and quartzose granofels. The considerable across-strike and along-strike variation in the units is consistent with a volcanic and volcaniclastic origin for these rocks.

Rocks of the New Hampshire Sequence

A sequence of rocks, herein referred to informally as the New Hampshire sequence, originally defined by Billings (1937) in the Littleton, New Hampshire area to the north and subsequently modified in the Fall Mountain area by Kruger (1946) consists of several formations, including the Ordovician Ammonoosuc Volcanics and the Partridge Formation, the Silurian Clough Quartzite, the Silurian and Devonian Fitch Formation and the Devonian Littleton Formation. Our mapping shows that the New Hampshire sequence in this part of Vermont is restricted to the south-central part of the Springfield quadrangle and the west-central parts of the Bellows Falls and Walpole quadrangles, in areas immediately west of the Connecticut River (Bellows Falls and Walpole), and within the outlier of the Wellington Hill allochthon (Walpole quadrangle; Figure 3). In addition, the Ammonoosuc Volcanics, which occur in the core of an F1/F2 doubly-plunging anticline within the Wellington Hill allochthon, are not present within the rest of the Vermont parts of the Bellows Falls, Springfield or Mount Ascutney quadrangles.

Within the Wellington Hill allochthon (Figure 3), the Ammonoosuc Volcanics consists of very well bedded to massive felsic gneisses with distinctive hornblende- and pale-green to clear amphibole bearing (gedrite, cummingtonite?) compositional layers. The felsic volcanics are interlayered with subordinate hornblende-plagioclase (amphibolite) layers. This bimodal layering occurs on centimeter to meter scales. The bimodal sequence is itself interlayered with two other rock-types on a map scale: 1) massive, gray, white-weathering felsic, plagioclase-quartz gneiss without amphibole that may be either metamorphosed felsic volcanics or sill-like intrusives (contacts are generally parallel to the trend of compositional layering in the bimodal sequence); 2) Massive, hornblende-plagioclase gneiss that in many places crosscuts metasedimentary unit contacts and thus appears to be, at least in part, intrusive in origin.

The Partridge Formation consists of rusty-weathering sulfidic schists, gray, garnetiferous schist; massive to well-bedded quartz-plagioclase-biotite and quartz-plagioclase-

muscovite granofels with occasional carbonate and calc-silicate layers or lenses; sulfidic, biotite-bearing granofels; quartz-rich schist with coticule horizons, and thin, discontinuous quartz-pebble conglomerates that are lithologically identical to those found within the Vermont sequence rocks and the stratigraphically overlying Clough Quartzite. Several 1- to 2-m-thick, coarse-grained amphibolites are present within the schist and have contacts parallel to the S1 foliation; these may be dikes or sills. One 5- to 25-m-thick, massive felsic gneiss horizon that is similar to those found within the Ammonoosuc Volcanics in the Wellington Hill allochthon is also present and crosscuts metasedimentary units within the Partridge Formation.

The sulfidic schist and granofels units of the Partridge Formation are readily distinguished from the voluminous gray schists and phyllite of the Waits River Formation by their rusty orange weathering and granofelsic texture. The gray schist unit, however, is very similar to the Vermont sequence Whetstone Hill Member of the Ordovician Moretown Formation, west of the RMC. Both of these units are also very similar in appearance to the gray schists found within the Northfield Formation (the stratigraphically lowermost unit in the CVS) and the quartz-rich schists within the eastern part of what we map as Waits River Formation (previously mapped, in part, as Devonian Littleton Formation of the New Hampshire sequence). The interlayered nature of these gray schist with the other Partridge metasedimentary rocks mentioned above, allows distinction from the younger Silurian-Devonian strata.

The Partridge Formation is in sharp contact with poorly bedded, massive white to light-gray quartz-pebble to boulder conglomerate, which we map as the lower member of the Silurian Clough Quartzite. This contact has classically been interpreted as a significant erosional unconformity (Thompson, 1954; Doll and others, 1961; Thompson and others, 1993). The lower member of the Clough Quartzite is dominated by the poorly bedded conglomerate, having rounded clasts of vitreous, white vein quartz and rare clear clasts of quartzite. One locality, along the Connecticut River in Walpole, New Hampshire, contains several cobbles of granitic origin. Local discontinuous lenses, 1- to 3-m-thick, of chlorite-biotite-muscovite-garnet-quartz schist and chlorite-muscovite-plagioclase-quartz granofels are found locally within the upper part of the lower member.

The contact between the lower member and the upper member of the Clough Quartzite is defined by the first bed of vitreous white quartzite. Above the contact the upper member consists of a 25- to 100-meter thick section of well-bedded, vitreous, white and gray quartzite with only minor discontinuous lenses of conglomerate. At several localities, the lenses of conglomerate form channels that indicate tops are toward the upper member. The upper part of the upper member is bluish-gray quartzite and granofels interlayered with staurolite-garnet-biotite-chlorite-muscovite-plagioclase-quartz schist. Rare lenses of deeply-weathered, brown, quartz-calcite and calc-silicate rock, 0.5- to 3-m-thick, contains fossils. Tetracoral, brachiopods, pelycypods and a possible trilobite, described by Boucot and others (1958) and Boucot and Thompson (1963), support a Llandovery (Early Silurian) age. These lenses may represent carbonate horizons and relict coquina beds (Thompson and others, 1993).

In the Springfield quadrangle, quartzite and schist of the upper member of the Clough are in sharp contact with black to dark-gray biotite porphyroblastic, carbonaceous calcite schists and brownish-gray-weathering, well-bedded calc-silicates, granofels and marbles of the Silurian-Devonian Fitch Formation. The Fitch Formation is limited to the northern summit and north slopes of Skitchewaug Mountain. Excellent exposures occur in roadcuts along I-91 immediately south of the Route 143 overpass and within an abandoned quarry immediately east of the roadcuts. Well-preserved bedding is typically parallel to S1 foliation within the Fitch Formation.

The upper member of the Clough and the Fitch Formation are present only within the mapped part of the Springfield quadrangle. Here, these units are in sharp contact with gray phyllite and schist of the Waits River Formation. The contact is interpreted as a fault and is nowhere thought to be depositional. Elsewhere, including the Bellows Falls and Walpole quadrangles, the Fitch and upper Clough are absent, and conglomerate previously mapped as lower Clough is in contact with Waits River units. Thus, rocks previously mapped (Doll and others, 1961) as the Devonian Littleton Formation overlying the Fitch or the Clough are now mapped entirely as the Waits River Formation of the Connecticut Valley sequence and are interpreted to underlie rocks of the New Hampshire sequence.

Structural Geology

The oldest foliation in the Silurian and Devonian rocks is a bed-parallel schistosity (Acadian S1) containing rarely observed isoclinal folds with generally north- or south-plunging fold hinges (Acadian F1). Only in the hinge regions of these early F1 folds is it possible to see bedding that is not parallel to a foliation. Both the Vermont and New Hampshire sequence rocks possess a first generation (Acadian S1) schistosity, but they do not appear to have developed under the same metamorphic conditions (see section below on metamorphism). The S1 foliation in the New Hampshire sequence appears to have developed prior to staurolite-kyanite-grade metamorphism, but the S1 foliation in the CVS developed prior to the peak of garnet- to sub-garnet-grade metamorphic conditions. The second generation planar fabric in all of the Silurian and Devonian rocks of both Vermont and New Hampshire sequences (Acadian S2) varies from a non-penetrative cleavage to a penetrative schistosity. Folds associated with the second-generation planar fabric (Acadian F2) vary from open to isoclinal with generally consistent shallow plunges to both the north and south, but locally the plunges are quite steep. S1 and S2 are the most dominant, or visibly conspicuous, planar fabrics in the Silurian and Devonian rocks. Locally these two planar fabrics are parallel, and it is difficult to discern one from the other. S1 and S2 are deformed by a minimum of two younger cleavages.

The next youngest generation(s) of planar fabrics are broad to open folds with both shallow and steep fold hinges (F3) and associated mm to cm spaced cleavage (S3) from the southern part of the Springfield quadrangle northwards. South, into the Bellows Falls

and Walpole quadrangles the D3 structures are more intense and are related to significant faults and shear zones (see discussion below on the Westminster West fault zone). These structures have many different orientations, although they most commonly strike northeast and dip vertically to steeply northwest and southeast.

The youngest generation cleavage in the area is a 1 to 30 cm spaced cleavage that locally occurs as parallel kink bands or low-amplitude, high-wavelength folds with variable fold hinge orientations. Secondary minerals, largely quartz, calcite, and dolomite, occur as vein-filling material in the cleavage planes. This latest generation of cleavage generally strikes east west and dips sub-vertically, and is largely restricted to the eastern and southern parts of the map area. This cleavage, and the outcrop-scale and map-scale brittle faults in the area, may be related to Mesozoic extension (Hatch, 1988a).

The Skitchewaug Allochthon

Based upon mapping within the Skitchewaug Mountain area (present-day Springfield quadrangle), Thompson (1954) interpreted a regional structure, defined by the map pattern of contacts between rocks of the Fitch Formation, the Clough Quartzite and the Partridge Formation, as a refolded nappe fold, with an eastward opening sub-horizontal hingeline closing to the west, and a related axial surface folded into a synformal geometry by subsequent north-south oriented dome-stage folds. The resultant map pattern depicted a boomerang, or an arrowhead, with the point of the arrow to the north (on the north slope of Skitchewaug Mountain) and the flared sides extending southwest, with a southwestern limb of Clough Quartzite only tens of meters thick, and a southeastern limb of Clough that was hundreds of meters thick (Robinson and others, 1991). In addition, structural measurements in Thompson's reports showed bedding measurements and sedimentary topping directions in which beds on the north side of the structure were overturned, whereas beds on the south side were upright. This pattern suggests that the axis of the nappe actually lies within the arrowhead pattern, is parallel to the outer edges of the pattern, and the axial trace of the nappe fold is actually east-west oriented rather than north-south.

Armstrong's mapping in this area (Armstrong, 1997a, 1997b; Walsh and others, 1996a, 1996b) has revealed the following information about this structure: 1) the oldest foliation in the New Hampshire sequence rocks at Skitchewaug Mountain, S1, is primarily bed-parallel and is generally sub-horizontal in the core of the fold with a sharp change in orientation occurring on the extreme flanks of the structure, where the foliation progresses through the vertical into steep-eastward overturned orientations on the western flank of the structure, southward overturned on the north side, and vertical to westward overturned on the east side. This change in orientation is produced in part by interference folding related to subsequent F2 and F3 folding, and by rotation of S1 during subsequent D2 rotational strain; both mechanisms folded S1 and bedding into a domal structure. 2) The contact between the New Hampshire Sequence rocks within the Skitchewaug structure, and the surrounding carbonaceous schist of the Vermont sequence (Waits River Formation) is an S2 mylonitic fault contact, with local and map-scale lithologic truncations, truncation of S1 fabric within the New Hampshire sequence rocks, and a

surface of metamorphic disparity between garnet- to staurolite-grade New Hampshire sequence and the sub-garnet-grade rocks of the Vermont sequence (Figure 4). Lithologic truncations include a near complete loss of the Clough Quartzite along the southwestern part of the western limb of the structure. This area, which coincides with the thin (10-meter-thick) southwestern limb of Thompson's nappe structure, occurs in an area of pervasive outcrop, in which the highly strained Clough can be traced along for over 500 meters to a location where sulfidic schist of the Partridge and carbonaceous, non-sulfidic schist that we map as Waits River Formation, are separated by mylonitized quartz-pebble conglomerate, caught within the fault zone, that is less than 3 meters thick. This contact post-dates the F2 phase of domal folding and garnet- to staurolite-grade metamorphism in the New Hampshire sequence rocks. Subsequent north-south oriented F3 folds deform this S2 fault into a basinal geometry such that the higher grade upper plate New Hampshire sequence is a klippe or outlier of previously deformed and metamorphosed rocks above lower grade rocks of the Vermont sequence.

We agree with previous interpretations that rocks included within the Fitch Formation, Clough Quartzite, and Partridge Formations are structurally above carbonaceous schist. We differ in the interpretation that the sections are coherent (i.e. the carbonaceous schists are Vermont Sequence and not New Hampshire Sequence). Our mapping and structural analysis of S1 orientations also indicate that the Partridge is structurally beneath the Clough, and the Clough is structurally beneath the Fitch everywhere within this structure, and that there is no nappe axial surface. In addition, we have not been able to identify bedding or related sedimentary topping criteria in the areas from which the overturned beds were reported by Thompson (1954). Although speculative, we believe that the overturned bedding data was probably derived from the notion that the fabric measured by Thompson was bedding in carbonaceous rocks he ascribed to the Littleton Formation; since the Littleton appeared to be structurally beneath the Clough, it would lead him to the conclusion that the section, and therefore, the measured beds, were overturned. Thus, the nappe structural model appears to have been generated primarily from stratigraphic arguments that are predicated on a coherent section of Clough, Fitch (where present), and most importantly, stratigraphically uppermost Littleton Formation. In our interpretation, the presence of upper and lower plate truncations, a sharp metamorphic contrast, and coherent stratigraphy within, but not between, the two sequences favors a thrust model for the Skitchewaug Mountain area.

Wellington Hill Allochthon

Rocks within the vicinity of Wellington Hill were interpreted as occurring within an antiformal structure that comprised either autochthonous New Hampshire sequence rocks (Thompson and others, 1968) or the upright limb of the Bernardston nappe (Thompson and Rosenfeld, 1979; Robinson and others, 1991). Recent mapping by Armstrong in this area indicates that rock-types associated with the Ammonoosuc Volcanics and Partridge Formation do occur within a doubly-plunging (north-south) antiformal structure, as described previously by Thompson and others (1968) and Thompson and Rosenfeld (1979). The antiform is defined by the folding of bed-parallel S1 foliation by a syn-S2 fold phase with an upright, nearly vertical, north-south-trending axial surface.

Furthermore, the New Hampshire sequence rocks reside above a syn-S2, sub-horizontal thrust fault which emplaced these rocks over the structurally lower rocks of the Vermont sequence Waits River Formation (Figure 4). The F2 antiformal structure within the overlying New Hampshire sequence rocks may be a fault-related fold that formed during D2 westward transport over the Vermont sequence rocks. The Wellington Hill allochthon appears to be an outlier of the Skitchewaug allochthon, located approximately 1 km to the east in the northern part of the Walpole quadrangle and the southern part of the Bellows Falls quadrangle. Thus, both structures would reside above the same S2 thrust surface, informally called the Skitchewaug Mountain thrust.

Westminster West Fault Zone

Lithologic units and S2-age fold and thrust fault structures in both the Vermont and New Hampshire sequences are truncated by at least two discrete zones of mylonitic fabric that compose what has been called the Westminster West fault zone (Figures 3 and 4). Armstrong (1995) first recognized this structure within the Townshend quadrangle, immediately southwest of the Walpole quadrangle; the southernmost quadrangle visited during this field trip. In Townshend, Armstrong (1995) observed that S2 fabric in the Vermont and New Hampshire sequences was highly crenulated, sheared and transposed into a secondary penetrative foliation, S3. Kinematics associated with the S3 shear zones, including asymmetric mica fish and quartz pressure shadows on porphyroblasts, rotated foliation, and offset planar fabric all indicated left-lateral strike-slip motion. These same features also display consistent east-over-west motion for dip-slip component offset. Associated elongation lineations indicate that the shear zone may have had multiple periods of dip-slip, strike-slip and oblique-slip motion over a protracted movement history. Relative-age relationships between porphyroblast growth and fabric development indicate that the latest phases of motion were post-peak metamorphic, and post-date the regional metamorphic field gradient. This is substantiated by the juxtaposition of chlorite-grade to biotite-grade rocks of the Vermont sequence (to the west) with garnet- to staurolite/kyanite-grade rocks of the New Hampshire sequence (to the east) along the eastern splay of the shear zone.

The Westminster West fault zone can now be traced continuously from the Townshend quadrangle northeastward into the Walpole and Bellows Falls quadrangles where the eastern splay is eventually buried underneath recent alluvial deposits of the Connecticut River. The western splay is unmapped north of the Bellows Falls quadrangle, and may terminate in the Springfield quadrangle.

Metamorphism

Acadian metamorphic grade within the CVT ranges from the chlorite and biotite zones to the garnet zone, and ultimately into an area of staurolite-kyanite zone in the Bellows Falls vicinity (Figure 5). Rocks in the western part of the map area are at garnet-grade as evidenced by abundant garnet porphyroblasts in pelitic and semi-pelitic rock types of the Waits River Formation, and abundant garnet in the Northfield Formation.

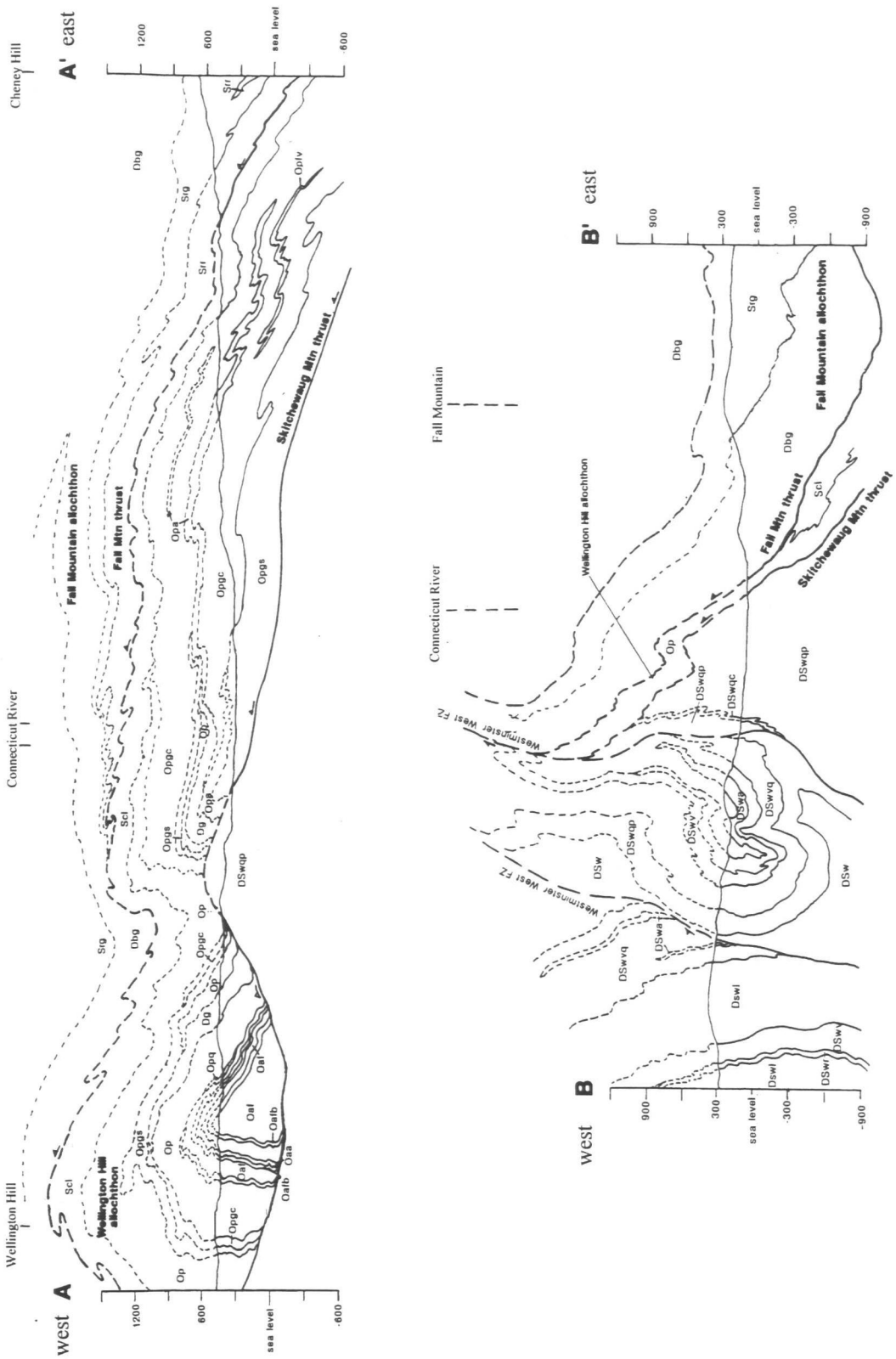

Figure 4a. Geologic cross sections A-A' (Walpole quadrangle; see registration locators in Figure 3) and B-B' (Bellows Falls quadrangle). Individual units shown in correlation of map units (Figure 4b) are described in Walsh and others (1986a, 1986b) and Armstrong (1997b). Vertical scale in meters.

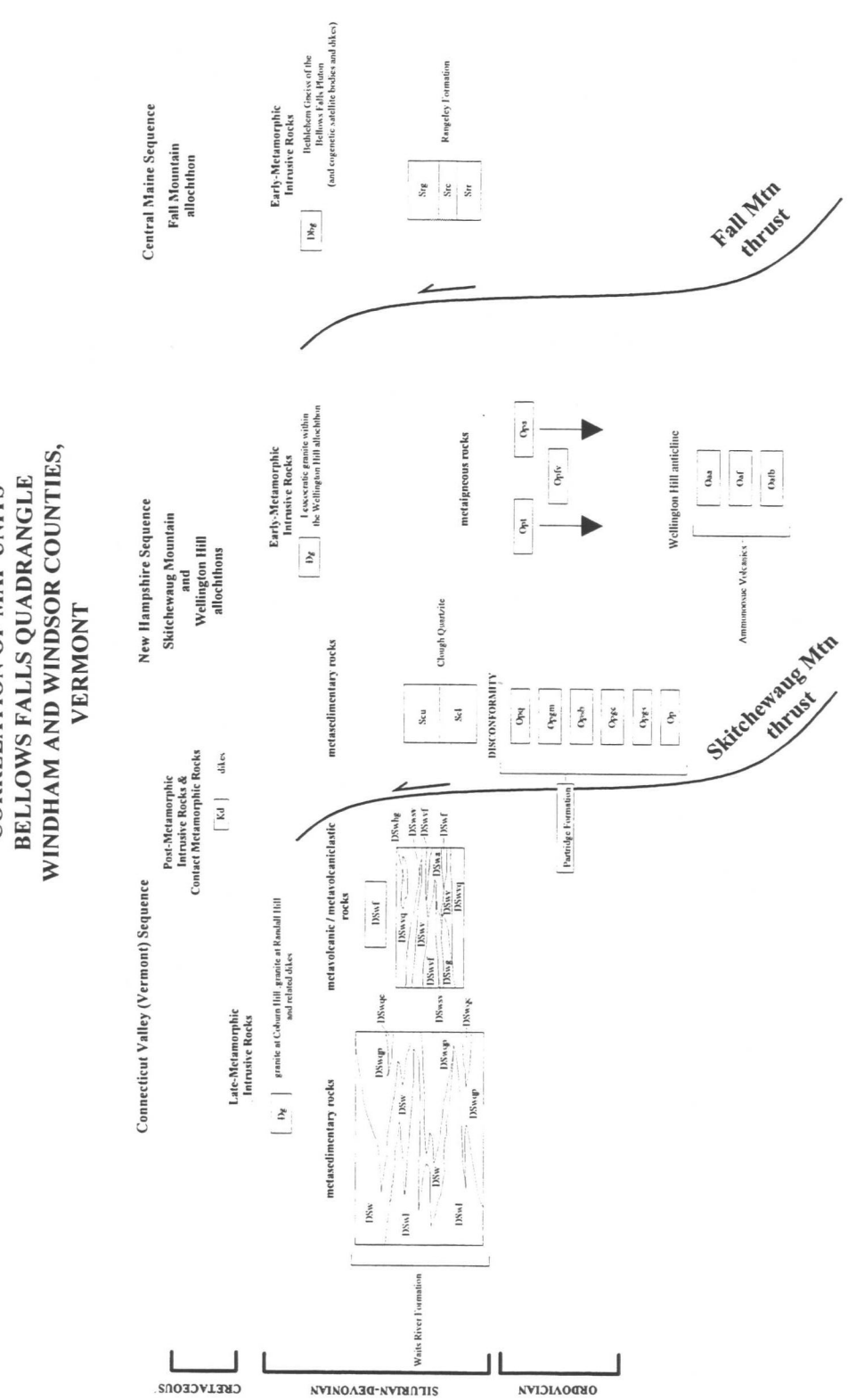

Figure 4b. Correlation of map units for cross sections in Figure 4a.

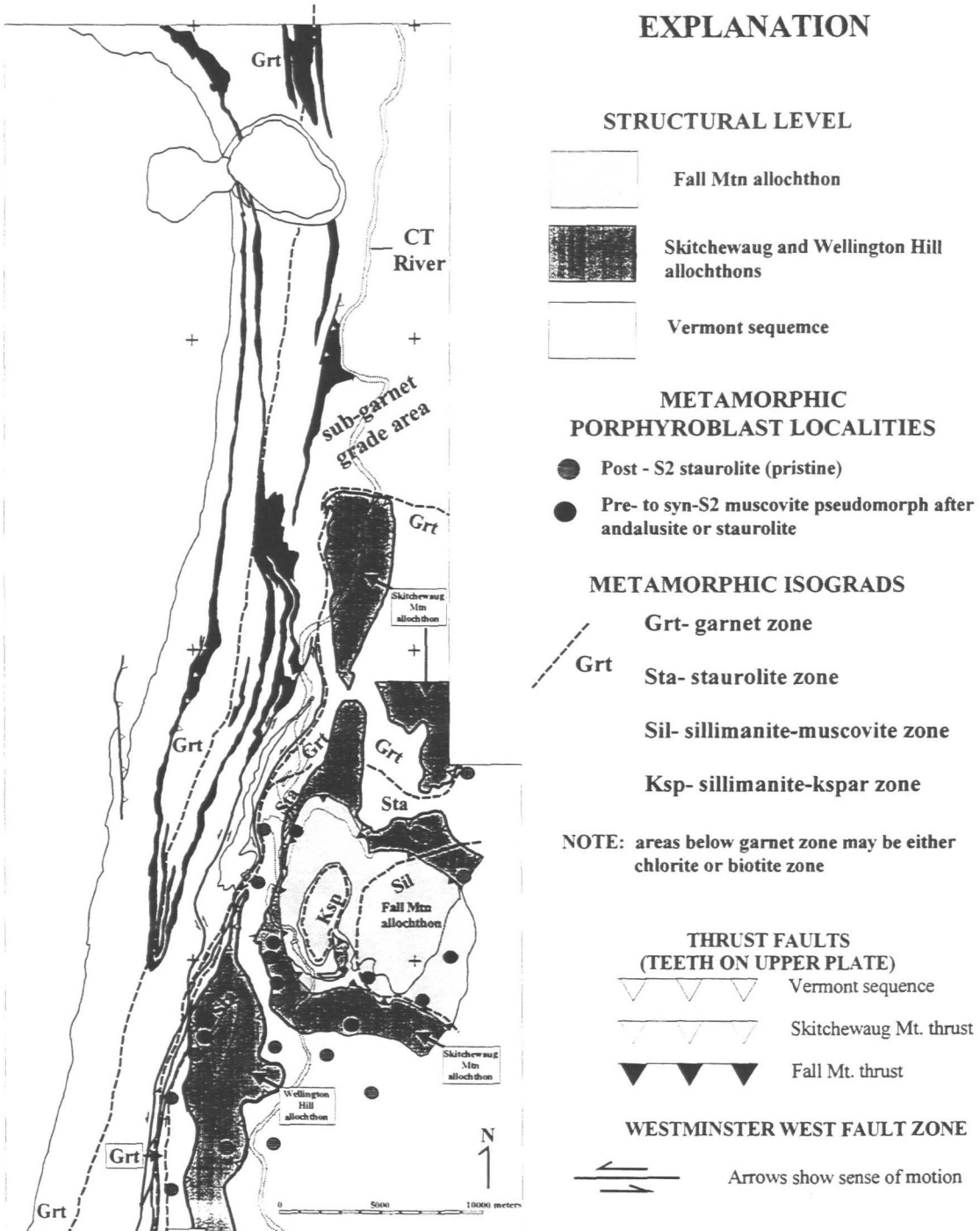

Figure 5. Metamorphic map of the study area, showing isograd boundaries, pseudomorph localities (discussed in text), and patterns for the three different structural levels. Also note that the volcanics within the Waits River Formation are patterned

Large garnets are found sporadically within the gray schists in the western part of the Waits River Formation. Large garnet and hornblende porphyroblasts are common within the western two belts of volcanic lithologies in the Waits River Formation. The eastern part of the Waits River Formation is devoid of garnets, and only the black schists and specific parts of the laminated schist and granofels contain biotite. The easternmost belt of felsic volcanics contains no biotite but does contain actinolite as pseudomorphs after pyroxene.

The garnet isograd associated with the western belt of garnet-grade rocks cuts across both D1 and D2 structures, including the axis of the F1 syncline that defines the CVS in this region (Figure 5). In the southern part of the Springfield quadrangle, the garnet isograd (garnet-in to the west) appears to be sharp and is approximately located along the syn-Acadian S2 shear zone near Hartness Park, east of the town of Springfield. In the Mount Ascutney quadrangle, the garnet isograd trends roughly north-northeast and cuts across several lithologies whose contacts are parallel to either S1 or S2; therefore, the isograd is not compositionally controlled and indicates increased metamorphic grade to the west. At Stop 1 we will show evidence that the growth of garnet porphyroblasts and attainment of peak temperature conditions occurred *after* the development of both S2 *and* subsequent northeast- and northwest-trending crenulation cleavage and penetrative foliation related to Acadian dome development in this area (locally, Acadian S3 and S4; dome-stage deformation).

In the Skitchewaug Mountain area, staurolite-grade rocks (possibly with anomalous high-alumina bulk compositions) occur within pelitic schists in the upper Clough Quartzite unit. No other rocks within the New Hampshire sequence here contain staurolite. The Partridge Formation contains only garnet, although some chlorite clots in the Partridge may be pseudomorphs after staurolite. The appearance of key metamorphic minerals in the New Hampshire sequence is strongly controlled by bulk compositional variation and, therefore, variation in metamorphic intensity (pressure, temperature) may not be as significant a factor within these structurally isolated rocks. Metamorphism within the New Hampshire sequence rocks appears to either postdate S1 or is synchronous with its development, but it certainly predates S2, and is likely related to a metamorphic event that occurred either prior to or during transport of the New Hampshire sequence rocks onto the Vermont sequence.

Within the Bellows Falls and Walpole quadrangles, some interesting results regarding the tectonothermal evolution of New Hampshire and Vermont sequence rocks are being acquired through the integration of mapping, structural analysis and metamorphic petrology. In several recent publications (Spear and others, 1990; Spear, 1992) and during ongoing field and analytical studies, Spear has been able to show the general distribution of at least three different metamorphic styles within rocks of both the Vermont and New Hampshire sequences, and a third sequence (the Central Maine sequence; CMS) which appears to structurally overlie the other two along another west-directed thrust fault (the Fall Mt. thrust; Figures 3 and 4). Rocks of the CMS include gray

Figures 6 and 7. Photomicrographs of porphyroblast textures from the New Hampshire sequence.

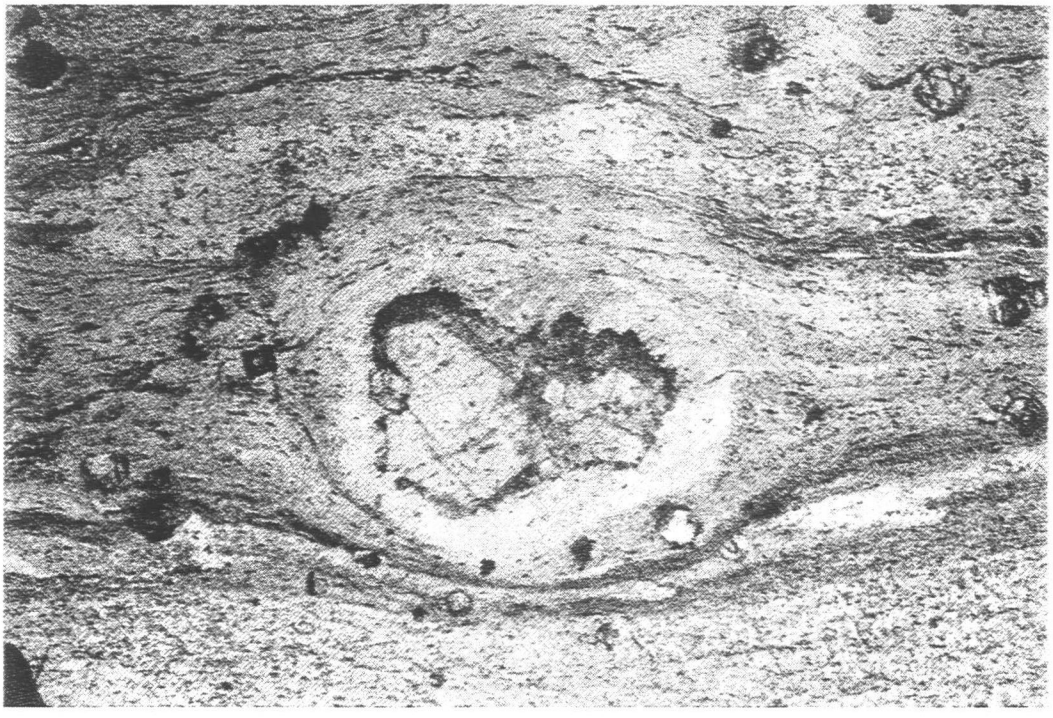

Figure 6a. Pseudomorph after early staurolite from the middle part of the Skitchewaug Mt. allochthon (contact metamorphic in origin). Note small garnet inclusions in pseudomorph and in matrix. Fabric (S2) postdates staurolite and predates pseudomorph reaction. Width of view = 0.5 cm.

Figure 6b. Pseudomorphs after andalusite (contact metamorphic in origin) from the upper part of the Skitchewaug Mt. allochthon, near contact with Fall Mt. thrust. Note small garnets in matrix. (Width of view = 1 cm).

Figure 7a. Porphyroblasts of late staurolite (regional metamorphic; outlined in white) from the Skitchewaug allochthon, plus early garnet (contact metamorphic; black outlines). Late staurolite overgrows both the garnet and S2 fabric. Width of view = 1 cm.

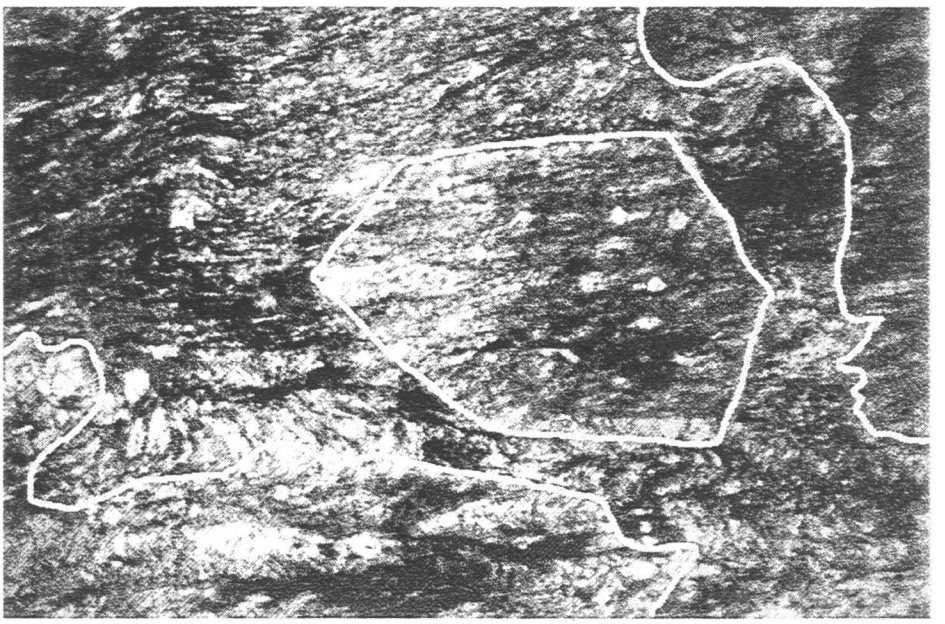

Figure 7b. Porphyroblast of late (post-S2) staurolite (regional metamorphic; white outline) from the lower part of the Skitchewaug allochthon. Small garnet inclusions are contact metamorphic in origin. Width of view = 0.5 cm.

and sulfidic schist of the Silurian Rangeley Formation, which are intruded by at least one large, sill-like body of peraluminous to metaluminous granite, called Bethlehem Gneiss. This sill is referred to as the Bellows Falls Pluton (Kruger, 1946).

Rangeley Formation rocks that are enveloped by the pluton were contact metamorphosed (the first metamorphic style) during intrusion of the pluton to Sillimanite-Kspar grade. Sillimanite typically occurs within the matrix and as pseudomorphs after andalusite. Migmatization of the schist is prevalent. Rocks of the Rangeley Formation below the pluton (but still in intrusive contact with it) were contact metamorphosed to andalusite-biotite- and to staurolite-biotite-grade conditions. The early staurolite and andalusite were subsequently pseudomorphed by muscovite, possibly during the early stages of D2 transport of these rocks along the Fall Mountain thrust (Spear, 1992). Intrusion of the Bethlehem Gneiss, therefore, predated westward transport of the Fall Mountain allochthon.

Rocks of the New Hampshire sequence, structurally underneath the Fall Mountain thrust, also contain muscovite pseudomorphs after staurolite (and in a few places, possibly after andalusite) which may have a similar contact and subsequent replacement origin to the rocks described in the Rangeley beneath the pluton. A map showing the distribution of the pseudomorphic texture shows that New Hampshire sequence rocks in the Bellows Falls area, and within the Wellington Hill allochthon, contain the pseudomorphs (Figure 6). Rocks of the Vermont sequence do not contain the pseudomorphs. The pseudomorph distribution can be related to two major points that we will refer to continually on this field trip:

1. The pseudomorphs are associated with a relatively high-T, low-P metamorphism that can be spatially related to the Bethlehem Gneiss of the Bellows Falls Pluton. Thus, the first metamorphism recognized is interpreted to be related to contact metamorphism. A U-Pb age on the Bethlehem Gneiss (reported in Kohn and others, 1992), interpreted to be a crystallization age, is 407 Ma. This would also be a general age for the contact metamorphism.

2. The distribution of contact metamorphism-related pseudomorphs indicates that the CMS rocks must have overridden the rocks of the New Hampshire sequence while the latter were still hot. The absence of pseudomorphs in the Vermont sequence rocks (including the gray schist that immediately underlies the New Hampshire sequence rocks along the Skitchewaug Mountain thrust) indicates that they were not in close proximity to the CMS-New Hampshire sequence rocks during the contact metamorphism. Furthermore, pseudomorphs within the Wellington Hill allochthon are consistent with the structure being an outlier of the Skitchewaug Mt. allochthon and not an antiformal exposure of rocks in an autochthonous position from below the Vermont sequence. If the Wellington Hill rocks were structurally beneath the Vermont rocks (and autochthonous with respect to the Vermont sequence), they would not contain contact metamorphic pseudomorphs, or the Vermont sequence rocks would contain them as well. We believe that this relationship gives strong

support to the structural model put forth by Armstrong (1997a, 1997b) that the New Hampshire sequence rocks, in this vicinity, structurally overlie the Vermont sequence rocks.

Both the New Hampshire and Vermont sequences contain staurolite-kyanite-grade rocks in close proximity to the Skitchewaug Mountain thrust. Staurolite growth in the New Hampshire sequence rocks is neoblastic, and is demonstrably younger than the larger staurolite and pseudomorphic texture related to contact metamorphism; the older staurolite is replaced by muscovite that grew during the development of the regional syn-thrusting, S2 fabric. The younger staurolite neoblasts typically overgrow the S2 fabric (Figure 7). Pristine staurolite within the Vermont sequence rocks appears to be either late- or post-S2. Spear (1992) attributed this phase of staurolite growth to loading of thrust sheets during the "nappe-stage" deformation (the second metamorphic style). Integration of the petrology with the structural/fabric analysis conducted by Armstrong and Walsh is very consistent with attainment of staurolite-grade conditions as a result of loading of the thrust sheets associated with the Skitchewaug Mountain and Fall Mountain thrusts (Armstrong, 1997b; Walsh and others, 1996a, 1996b). Retrogression of the staurolite-kyanite-grade assemblages (and possibly garnet-grade assemblages to the west within the Vermont sequence), with replacement of staurolite and garnet by chlorite (the third metamorphic style), is related to the development of the S3 mylonitic fabric of the Westminster West fault zone.

Geochronology

Figure 8 shows a selection of published ages that constrain the timing and cooling history of metamorphism in central New England. In Vermont, growth of garnet at Gassetts occurred ca 378 Ma (Vance and Holland, 1993), which also is the lower limit of deformation (380 Ma, Naylor, 1971), and tightly constrains the age of Acadian metamorphism. In New Hampshire, metamorphism is more complex and prolonged. The ages of the New Hampshire magma series (Kinsman Quartz Monzonite and Bethlehem gneiss, including the Bellows Falls and Indian Pond plutons) is known to be 404 - 413 Ma. Ages of monazites, mostly from the high-grade part of central New Hampshire, are 392 - 402 Ma (Eusden and Barreiro, 1988), but the significance of these ages relative to the history of metamorphic recrystallization is not known. There is an indication of Late Devonian (ca. 360 Ma) recrystallization in central Massachusetts (south of the map limits), and Permian/Carboniferous ages (ca. 285-295 Ma) in the Pelham dome have been interpreted as indicative of Alleghenian deformation and metamorphism affecting the basement rocks (e.g. Gromet and Robinson, 1990; Tucker and Robinson, 1993).

The post-peak metamorphic cooling histories of New Hampshire and Vermont are also disparate. Published $^{40}Ar/^{39}Ar$ dates (Harrison and others, 1989; Spear and Harrison, 1989) have revealed considerable details regarding New England Paleozoic cooling history.

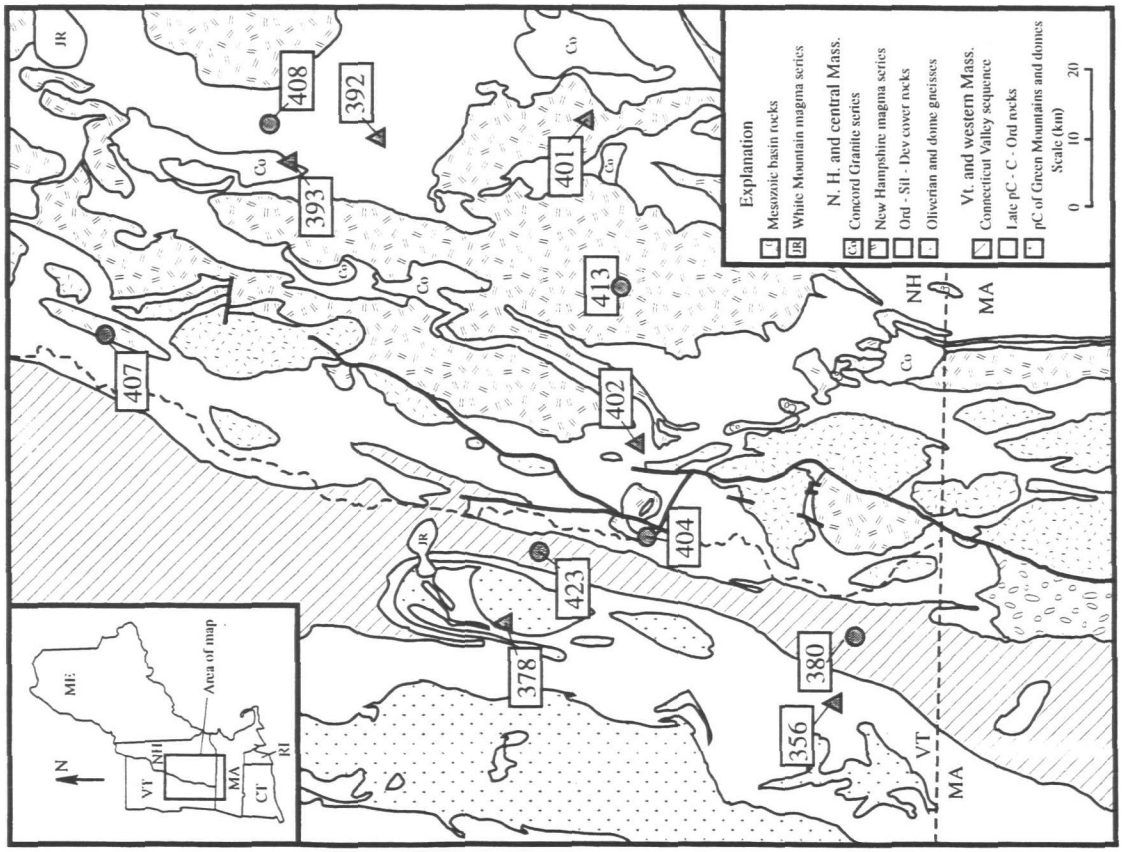

Figure 8. Selected metamorphic ages from western and central New England. Symbols on map correspond to authors shown to the right.

One of the striking features of these data is the difference in cooling ages of hornblende, muscovite and biotite in Vermont compared with New Hampshire. Age differences across the Vermont-New Hampshire border are approximately 375 versus 350 (hornblende, in Vermont and New Hampshire, respectively), 345 versus 250 (muscovite) and 330 versus 240 (biotite), although there are some anomalous, as yet unexplained older ages in both states. Harrison and others (1989) interpreted the age differences as being due to hinged differential uplift between two terranes, perhaps as the result of normal faulting, and suggested a correlation with the Mesozoic border fault, although they realized that the density of data were insufficient to clearly define the boundary. They also recognized that differential movement must have commenced nearly 100 Ma prior to the initiation of documented graben formation along the Mesozoic border fault, and did not rule out the possibility of a previously unrecognized structure as being responsible. The age discontinuity also lies in approximately the same area as the observed metamorphic discontinuity between garnet and staurolite grade rocks. It is suggested here, therefore, that composite and long-lived movement along the Westminster West fault zone (and reactivation of the zone) is responsible, at least in part, for the disparate metamorphic and cooling histories in this region.

Summary and Discussion

The results of our mapping, structural analysis, and petrologic studies show details not previously reported for the rocks in the Connecticut Valley sequence, including:

- Rocks associated with the Vermont sequence, including volcanics and gray schist of the Waits River Formation, extend much farther into New Hampshire than previously thought.

- Rocks previously mapped as the Standing Pond volcanics do not occur as a single time-stratigraphic unit that separates different metapelitic sequences, previously mapped as the Waits River (below) and Gile Mountain (above) Formations. Rather, volcanic rocks previously mapped as Standing Pond (and Putney) volcanics occur within at least five different belts internal to the Waits River Formation.

- Rocks of the Connecticut Valley sequence appear to be distributed within a large F1 synformal structure, to the south, within the Townshend, Saxtons River, Walpole and Bellows Falls quadrangles. This structure, however is not well defined to the north within the Springfield and Mt. Ascutney quadrangles, where the Waits River lithologies, at least in part, may compose a homoclinal, east-facing sequence.

- The regional Barrovian metamorphism, interpreted here to be related to loading of the Vermont sequence by the Skitchewaug Mountain and Fall Mountain allochthons, postdates the F1, F2, *and* F3 structures in the western part of the CVS that are related to the formation of Acadian domes (the Chester and Athens domes, in this area;

Figure 3). Barrovian metamorphism postdates the regional S2 fabric (associated with thrusting and loading of the allochthons) in the east near the Vermont-New Hampshire sequence contact. The transition in relative-age relationships appears to change across the post-peak-metamorphic Westminster West fault zone. The cause of change in relative-age relationships is not yet clearly understood, but probably is related to diachroneity in the evolution of structures during thrusting and associated loading of the different lithotectonic terranes.

- New Hampshire and Central Maine sequence rocks, everywhere mapped, are allochthonous with respect to the structurally underlying Vermont sequence. Metamorphic (pseudomorph) textural relationships strongly suggest that the New Hampshire and Central Maine rocks experienced and early contact metamorphism that is absent from the Vermont sequence, and predated the regional Barrovian metamorphism that affected all sequences. This is strong, independent evidence for the structural-based model for the Skitchewaug Mountain and Fall Mountain allochthons.

Previous workers interpreted key belts of volcanic rocks as time-stratigraphic units between different formations (Lyons, 1955; Doll and others, 1961). The Putney Volcanics (Hepburn, 1972; Trask, 1980) are shown in most places between the Gile Mountain and Littleton Formations and the Standing Pond separates the Waits River and Gile Mountain Formations. The map distribution we find indicates that volcanism occurred throughout fluctuations in carbonate sand deposition because the volcanic rocks do not separate carbonate-bearing from non-carbonate-bearing sequences. The map pattern suggests that the volcanic rocks transgress unit boundaries within the sequence; an observation also noted by Lyons (1955) immediately north of the map area, in the 15-minute Hanover quadrangle. The presence of limestones on both sides of volcanic units throughout the sequence and the probability of multiple volcanic horizons complicate criteria for separating the Waits River Formation from the Gile Mountain Formation on a stratigraphic basis. The use of the name Gile Mountain Formation is further complicated because non-limestone-bearing lithologies including gray slate to phyllite and quartzite, feldspathic quartzite and mica schist, and rhythmically graded schist and micaceous quartzite typical of the Gile Mountain Formation from east-central Vermont (Hatch, 1988b) are absent in this area. The mapping illustrates the need to re-evaluate the regional distribution of limestone-bearing and volcanic rocks within the Connecticut Valley sequence to assess the validity of previously mapped belts of the Gile Mountain Formation.

Refolded F1 folds in the area are not useful for determination of regional folds because the folds are commonly rootless, show no asymmetry, and are too widely distributed. Recent analyses of sedimentary topping criteria to the north (Fisher and Karabinos, 1980; Hatch, 1988b) provide evidence for map-scale F1 folds such as the synform seen on Stop 2 of this trip (the CVS synform). The lack of unequivocal fold and topping criteria means

that interpretations as to the location of such folds in this area are conjectural. Several possibilities exist, however, that could explain the overall distribution of units.

The classic works by Rosenfeld (1968, 1970) present the apparent rotation of inclusion trails in large garnets from the volcanic rocks as evidence for changes in vergence across the limbs of the nappes. Recent work by Thompson and others (1993) use this evidence to support the nappe model. Our mapping shows that the garnet isograd cuts across the distribution of the units and that garnet growth is syn- to post S2 development, making it difficult to correlate garnet growth with early F1 nappe-stage folds Stops 1 and 2 of this trip). The current controversy over whether garnets actually rotate (Bell and Johnson, 1989; Ratcliffe and others, 1992; Ratcliffe and Armstrong, 1995a) also raises considerable questions regarding the facing direction of major fold limbs in the area.

Road Log

Assemble in Springfield, Vermont at the Springfield Plaza parking lot in front of the Grand Union (across the street from the McDonald's) at the junction of Vermont Route 11 and 106. **Departure time is 7:30 am. Bring a lunch.**

Mileage	
0.0	Turn right out of the parking lot onto Route 11 East.
0.7	Downtown Springfield.
1.0	Blinking yellow light. Route 11 bears left, go straight on South Street (road between Sunoco & Dubanevich's Market).
1.3	Proceed straight.
1.9	Springfield High School on right.
3.4	Hardscrabble Corner at Stop sign, turn right onto Brockways Mills Road.
4.4	Road forks, bear left.
4.7	Outcrop of Waits River Formation schist on right.
5.7	Cobble Hill to left (east) is underlain by large-garnet and hornblende schist (garbenschiefer), a volcaniclastic unit in the Waits River Formation with baseball-size garnet porphyroblasts.
6.3	Waits River Formation limey schist, volcaniclastics and garbenschiefer outcrops on left in pasture.
8.2	Pull off on left into dirt road for Stop 1.

Stop 1:
Brockways Mills - Volcanics and volcaniclastics of the Waits River Formation (Saxtons River 7.5 x 15-minute quadrangle). Proceed across railroad tracks along east side of hay field down to Williams River exposures.

Rocks within the bed of the river consist of silvery-gray calcitic pelite with rare, thin, brown-weathering siliceous limestones, and various metavolcanic and volcaniclastic rocks. By walking upstream, you will go through the metapelite and limestones. Notice that the dominant planar fabric, the S2 schistosity, is defined by strongly aligned white mica and chlorite, and by numerous sheared out and transposed quartz veins. Contacts between the lenses of limestone and the surrounding metapelite are parallel to both S1 and S2. In many places, you will also be able to observe F2 folds of quartz veins, and occasional folds of an earlier foliation, S1. The abundant garnet porphyroblasts within the metapelite overgrew both S1 and S2. S2 within this outcrop generally strikes north south and dips very steeply either west or east. In many places, you will also notice that the dominant fabric is folded by a set of younger crenulations, which trend north-northwest, and dip steeply west. This fabric, S3, is locally penetrative and defined by aligned white mica becoming a secondary shear-band foliation. The angular relationship between S2 and the superposed S3 typically produces a series of asymmetric folds, F3, that have a clockwise sense of rotation (Figure 9). In his study of snowball garnets, Rosenfeld (1968) attributed the clockwise inclusion patterns within garnet porphyroblasts at this locality to active flow of material off the eastern flank of the Athens and Chester domes. This flow, related to diapiric upwelling of deeper crustal material was thought to be the cause for the development of "spruce-tree folds" which yield opposite senses of rotation to parasitic folds that form on traditional buckle, or passive-flow folds (Figure 9).

Downstream from the metapelite exposures, you will walk across the contact with a sequence of large-garnet and hornblende rocks (called "garbenschiefer") that we think represent admixed mafic and pelitic sediments at the interface of volcanic and pelitic deposition in this ancient basin. Look closely at the large garnets and the included fabrics. Notice that the garnets include fabric that displays either:

1. Planar inclusions that define a foliation that is continuous with the matrix foliation, S2.

2. Crenulated schistosity that has an axial plane in a coplanar orientation with either F2 or, most commonly, F3 axial planes and related schistosity in the surrounding matrix. These folds typically have clockwise asymmetry, similar to the asymmetry of the F3 folds that were observed in the metapelite exposures and within the matrix folds in close proximity to the garbenschiefer garnets. Also, notice that the orientations of these "axial planes" within the garnets are very consistently oriented.

We believe that these observations, regarding the crossing-angles of S2 and S3, the continuity of planar fabric within and immediately outside of garnets, the similar geometry of crenulate folds within and outside of garnets, the consistency of orientations of axial-plane orientations within many of the garnets, and the lack of any rotation of fabric within

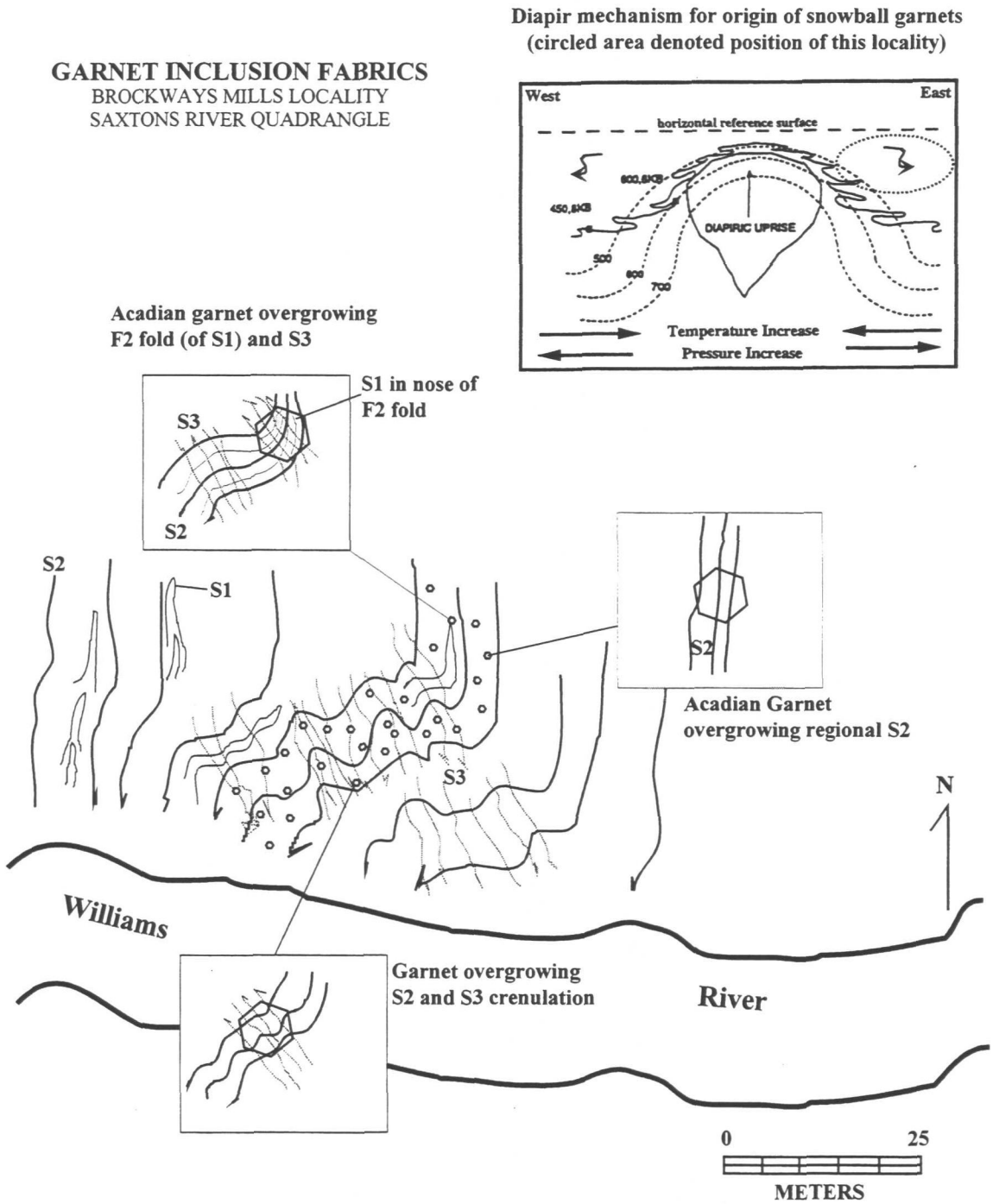

Figure 9. Schematic map of deformational fabric and garnet porphyroblast textural relationships. Inset map shows diapir model for the evolution of snowball garnets, as depicted by Rosenfeld (1970).

garnets (S2) that is continuous with S2 matrix foliation, are consistent with garnet growth after or during the development of S3 crenulation cleavage, penetrative foliation and F3 folding (Figure 9). The striking similarity and orientations of structures within the matrix of both the metapelite and volcanic/volcaniclastic rocks, and the lack of heterogeneous rotational geometries (garnets that show a broad range of angular rotation) argue against a syn-growth rotational origin for the garnets. The implications for this are great, in that the non-rotational model requires relatively static garnet growth, and attainment of peak-temperature metamorphic conditions after the development of S1, S2 and S3 deformations and thus, after at least the initial development of the Acadian dome structures (the Chester and Athens domes) further to the west. In addition, the F3 fold asymmetry would be related purely to the angular relationships between the orientation of pre-existing S2 and the developing S3; this would not necessitate a diapiric origin for the formation of either S2, S3 or the domes themselves.

Further downstream, the garbenschiefer is in sharp contact with thin (less than 1 meter-thick) layers of metapelite that are interbedded with more garbenschiefer that contains substantially smaller garnets (but with big fascicles of hornblende). This interbedded zone is approximately 20-meters-thick, and on the downstream (west) side, is in contact with a 1- to 5-meter-thick zone of chlorite-muscovite-plagioclase (felsic) volcanics that contain angular fragments of ilmenite-muscovite-plagioclase granofels, 1 mm to 1 cm in diameter, that may be small volcanic bombs. This rock is mappable to the south for more than 3 km, and is similar in composition to the rocks at Stop 2. Going further downstream, these rocks pass into more garbenschiefer and finally, into a 70- to 100-meter-thick section of amphibolite, which along with the garbenschiefer, makes up the majority of volcanic rock within the westernmost belt of Waits River volcanics; formerly mapped as the Standing Pond Volcanics (Doll and others, 1961).

Return to vehicles and continue south along Brockways Mills Road.

Mileage	
8.6	Brockways Mills, bear left and cross bridge and railroad tracks.
9.0	Stop sign at junction with Route 103. Turn left (south).
9.2	Turn right on Pleasant Valley Road towards Saxtons River.
11.0	Outcrops of hornblende-plagioclase gneiss and schist with interbedded limestone of the Waits River Formation in pasture on left (east).
13.8	Bear left, continue downhill to Stop sign.
13.9	Stop sign at junction with Route 121. Continue straight on 121 East.
14.2	Village of Saxtons River. At Stop sign, bear left.
14.25	Turn right onto River Road.
14.3	Make right into parking area on Saxtons River for Stop 2.

Stop 2:
Saxtons River - Volcanic and volcaniclastic rocks of the Waits River Formation showing F1 and F2 folds (Saxtons River 7.5 x 15-minute quadrangle).

Rocks at this exposure consist of quartz-chlorite-muscovite-plagioclase with minor amounts of ilmenite and calcite, which is typically weathered-out forming the distinctive pitting parallel to the dominant foliation, S1. Notice that the rocks also contain strong, S1-parallel compositional layering. 1 to 5 mm plagioclase-ilmenite-muscovite-bearing fragments, similar to the felsic bombs at Stop 1, are locally abundant. Locally, 1- to 20-cm-thick layers of tan-weathering, plagioclase-rich beds of fine-grained felsic tuff are also parallel to S1. The S1 fabric is pervasively folded by upright F2 folds, which have a weak to moderately developed S2 foliation within the F2 axial surfaces. F2 folds generally have very shallow southward plunges, but some areas have either horizontal or shallow northward plunges. In several locations at this exposure, S2 fabric is folded by a set of shallow-dipping kink bands with either reverse or normal offset of S2 fabric. On the north side of the falls, approximately 50 feet east (downstream) of the bridge, one can find zones where S1 fabric is axial-planar to F1 folds that have closures defined by folded layers. Several of these closures may actually contain a pre-S1 cleavage that is parallel to the beds. Rosenfeld (1954) and Thompson and Rosenfeld (1979) previously interpreted the dominant set of folds as "nappe-stage"; folds related to the development of large recumbent structures during the early part of the Acadian Orogeny. It was suggested that the clockwise sense of rotation of these south-plunging folds was indicative of position on the eastern limb of an antiformal structure. Thompson and Rosenfeld (1979) and Doll and others (1961), however, mapped the gray metapelite and limestones immediately west of these rocks as Gile Mountain Formation, and thus stratigraphically above the volcanic rocks at this locality, which were called Standing Pond Volcanics. Since these workers interpreted the Standing Pond as a time-stratigraphic unit that separated stratigraphically underlying Waits River Formation from the overlying Gile Mountain Formation, presence of the Gile Mountain rocks within the core of this antiform required the structure to be a synclinal nappe that was subsequently overturned into an antiform during subsequent dome-stage deformation.

Our mapping and structural analysis within this area has led to the following observations:

1. The dominant set of folds (our F2) at this locality fold an older foliation (our S1), which locally can be seen associated with folded beds (beds folded by F1). Immediately south of this locality, this belt of volcanics joins with the westernmost belt of volcanics within the Waits River Formation (seen at Stop 1), forming the southern closure of a large fold structure. At this closure, F2 folds on both limbs show the same rotation sense, and plunge south. Older F1 folds have very steep to moderate northward plunges and show opposite sense of rotation, consistent with the closure being related to an early, north-plunging synform. We believe that the structure, geometrically speaking, is an F1 synform with subsequent F2 structures superposed upon it. F2 would therefore produce further shortening on a preexisting F1 synform.

2. The distribution of volcanic rocks within metapelite lithologies is not as simple as previously portrayed. Regardless of the causes, the volcanics are distributed in at least five different belts, from west to east, across the Connecticut Valley Trough. Only the westernmost two belts (one seen here at Stop 2, the westernmost one seen at Stop 1) come together to form a demonstrable fold closure. The other three reside in different structural and stratigraphic positions further east. Although all three of these easternmost belts have very heterogeneous distributions of volcanic and volcaniclastic rocks, and occur within somewhat different metasedimentary hosts, they do contain similar lithologies. These relationships are not consistent with the volcanic rocks being distributed within a single time-stratigraphic unit. Furthermore, there is no sedimentary evidence for topping reversals across the individual belts that would indicate that they mark the positions of nappe-fold axial traces related to the folding across the belt of a single stratigraphic horizon, as portrayed in the cross-sections shown by Doll and others (1961) and subsequent Figures by Thompson and Rosenfeld (1979) and Thompson and others (1993).

3. Metapelitic rocks on both sides of this limb of volcanic/volcaniclastic rocks contain beds of brown-weathering siliceous limestone that is diagnostic of the Waits River Formation. Our mapping indicates that there is no lithological evidence for separation of the metapelitic rocks into two different formations as done previously; separation was based exclusively upon the premise that the volcanics at this Stop represent a unique time-stratigraphic boundary.

The previous three points represent our case for combining all of the rocks within the Connecticut Valley trough, at this latitude, into the Waits River Formation. Furthermore, the distribution of lithologies, coupled with the structural arguments presented above, lead us to conclude that there is no unequivocal evidence for a regional nappe structure in this belt. A simple synclinal structure is the simplest and most consistent model for the geometry of this part of the trough.

In regard to the Acadian metamorphism in this area, the rocks at this locality are sub-garnet (biotite) grade. Garnet is present within the metapelite immediately west of this locality (approximately 0.5 km west of the bridge). Distribution of garnet-bearing rocks shows that the related garnet isograd cuts across both this F1 synclinal structure and younger F2 structures, to the north within the northeastern part of the Saxtons River quadrangle and the adjacent Bellows Falls quadrangle to the east. Similar relationships are found within both the Springfield and Mt. Ascutney quadrangles as well. These cross-cutting relationships are consistent with the fabric-mineral growth age relationships seen at Stop 1 which indicate a post-S2 and locally, a post-S3 growth history for garnets and related metamorphic peak-T attainment. These relationships are very consistent in this region. This relationship will be compared to the timing of porphyroblast growth and S2/S3 fabric development further east during subsequent Stops.

Return to vehicles and go back to Route 121. Turn right onto 121 East.

Mileage	
14.4	Vermont Pretzel Company on right.
15.5	Crossing Westminster West fault zone.
17.3	Cross bridge over Saxtons River. Outcrops on the north side of the bridge are staurolite grade rocks of the Waits River Formation.
17.8	Turn right at monument green onto Gage Street.
18.0	Covered Bridge Road on right, continue straight onto dirt road.
18.2	Pull over to right side of road for Stop 3. Head down path to river exposure.

Stop 3:
Saxtons River - Graded beds with staurolite replaced by muscovite. Walpole 7.5 x 15-minute quadrangle. Spear and others (1990) sample location 79-66).

The purpose of this Stop is to examine rocks of the staurolite zone and to discuss their structural position, reaction history and metamorphic P-T path. (Also see Spear, 1992; Stop 3, Spear, 1993, and Thompson and Rosenfeld, 1979, Stop 1).

Recent mapping by Armstrong has shown that rocks at Stop 3 are part of the same package as seen at Stop 4, and constitute part of the New Hampshire sequence. These rocks are in tectonic contact with rocks like those we shall see at Stop 5 along an early D2 thrust fault, herein referred to as the Skitchewaug Mountain thrust. Rocks at this outcrop are now mapped as a unit within the Partridge Formation. Note that they display excellent graded beds. Tops can readily be determined by examining direction of fining (note that the large crystals have grown in the clay layers at the tops of the beds). The rocks here are isoclinally folded, which reverses the direction of tops. A particularly good example of an isoclinal fold hinge can be seen in the rocks immediately to the right (upstream) of the upper pool. Stand facing the pool and note the topping direction, which is inverted to the west. Now note the topping direction of rocks to the right of the pool at head level (2 meters above the water). Here the rocks top up to the east. The fold hinge is exposed at the pool's edge and plunges into the pool.

The assemblage here is staurolite + garnet + biotite + muscovite + quartz. Retrograde chlorite is present in the matrix. Most of the staurolite is gone in these rocks, being replaced by pseudomorphs of muscovite (Figure 6a). Locally, one can find relict staurolite cores in the pseudomorphs.

The early staurolite-grade metamorphism seen in these rocks predates the thrusting, and has been interpreted by Spear to be related to contact metamorphism, possibly with granitoids such as the Bethlehem Gneiss of the Bellows Falls Pluton. Subsequent replacement of the staurolite by muscovite may be related to the retrograde reaction:

$$\text{staurolite} + \text{biotite} + H_2O = \text{chlorite} + \text{muscovite} \pm \text{garnet}$$

which is a discontinuous reaction in the KFMASH system. The replacement of staurolite by a potassic phase such as muscovite is common in replacement textures (e.g. Carmichael, 1969; Foster, 1981, 1983) and attests to the higher mobility of K relative to Al. The alteration of staurolite to muscovite seen here is completely different from that observed lower in the section and to the west, where staurolite is replaced by chlorite.

Mapping on Fall Mountain by Kruger (1946) and by Allen (1984, 1985) and Chamberlain and others (1988), just east of Stop 5, on the New Hampshire side of the Connecticut River, has shown that the Bethlehem Gneiss of the Bellows Falls Pluton intrudes rocks on both the upper and lower contacts. We agree with Allen (1984, 1985) that the rocks above the pluton, on Fall Mountain, are most likely lithologies associated with the Silurian Rangeley Formation of the Central Maine sequence, and were previously transported from the east from a deep-water basin on the east side of the Bronson Hill. These rocks would likely be the eastern (distal) correlative of the Clough Quartzite. Westward transport of the Rangeley over the Partridge Formation and Clough Quartzite of the New Hampshire sequence would have occurred early during the Acadian Orogeny. This fault has been mapped by Allen (1985) and called the Fall Mountain thrust.

Ongoing mapping by Armstrong in the Fall Mountain area indicates that rocks below the pluton, previously mapped as Partridge Formation, may actually be more Rangeley, such that the Bellows Falls Pluton is entirely within the Rangeley. This is similar to the distribution of other New Hampshire series granites farther south in the Keene area, such as the Ashuelot Pluton that is in intrusive contact with Rangeley, both above and below (Armstrong, 1986; Elbert, 1988). Although there may not be an intrusive contact between the Bellows Falls pluton and rocks of the New Hampshire sequence in this area, Spear has identified numerous localities where early muscovite pseudomorphs after staurolite, and possibly, after andalusite occur within rocks now mapped by Armstrong as within the Partridge Formation. A map made by Spear of the distribution of these early muscovite pseudomorphs correlates extremely well with Armstrong's geologic map showing the distribution of rocks within the structural levels occupied by the New Hampshire and Central Maine sequences (black dots in Figure 5). No early pseudomorphs have been found within rocks mapped as Vermont sequence.

The presence of early pseudomorphs in the New Hampshire sequence rocks suggests that these rocks were heated prior to the Barrovian metamorphic event that produced the younger idioblastic staurolite during or after S2 deformation. The early heating event may herald the arrival of the Fall Mountain thrust sheet; heating of the New Hampshire sequence at low pressures, as discussed by Spear and others (1990) and Spear (1992, 1993) would imply that the Bellows Falls pluton was still hot and that the Fall Mountain thrust sheet had not yet achieved thermal equilibrium following the intrusion of the Bellows Falls Pluton. In other words, hot rocks were being structurally emplaced over colder rocks. Furthermore, this requires that the rate of cooling and attainment of thermal equilibrium is slower (in this case) than the rate of tectonic transport. Subsequent regional staurolite-grade

metamorphism followed this early retrograde event, with staurolite growth related to loading of rocks within this structural level (the Skitchewaug Mountain allochthon) by rocks at higher crustal levels.

Return to vehicles and continue on Gage Street.

Mileage	
18.6	Turn right into "The Basin Farm."
18.7	Pull off to right side of road for Stop 4.

Stop 4:
New Hampshire sequence - Graded beds of biotite granofels of the Partridge Formation in contact with conglomerate of the Clough Quartzite (Walpole 7.5 x 15-minute quadrangle).

Exposure on the road is quartzite- and quartz-pebble conglomerate of the Silurian Clough Quartzite; part of the New Hampshire sequence. Scramble down into the riverbed by the old bridge abutment. The first exposures here are of the Clough. The dominant fabric is composite S1/S2. Note that pebbles on foliation surfaces show little stretching and weak to moderate flattening. Strain is not high. Where present, bedding, defined by pebble sorting in lensoidal channels, dips 30 to 40 degrees to the east (strike of bedding is parallel to S1/S2 which is northwesterly).

The sharp contact between the conglomerate and well-bedded and mostly well-graded biotite-bearing granofels, muscovite-rich granofels, feldspathic quartzite and wacke, and calcite-bearing biotite granofels, is approximately 50 meters upstream. Notice that sedimentary topping directions from graded beds yield both east- and west-facing beds across 1 meter to tens of meters of distance; similar to the reversals in tops seen at Stop 3 due to isoclinal fold repetition. F2 fold closures, however, are difficult to find here. Reversals in tops are found throughout this entire exposure, which extends around the left-hand bend (looking upstream) for more than 300 meters, to within 250 meters of the rocks at Stop 3.

These rocks were previously interpreted to be within the Devonian Fitch Formation, and thus, stratigraphically above the Clough Quartzite, to the east (Thompson and Rosenfeld, 1979). These very distinctive rocks are traceable both immediately south (onto Bald Hill, visible across the river), and north onto Oak Hill. In both areas, these well-bedded rocks:

1. Contain massive, hornblende-plagioclase±garnet±cummingtonite amphibolite, 1- to 20-meters-thick and several hundred meters in length. Detailed mapping in areas of good control shows that the amphibolites are in sharp contact with the granofels, as well as rusty-weathering, sulfidic schist, gray garnetiferous schist, and sulfidic, pyrrhotite- and pyrite-bearing, fine-grained quartzite. These lithologies occur on both sides of the amphibolite such that the amphibolite crosscuts lithologic contacts and appear to be

intrusive. Texturally, they are very similar to other mafic horizons found within the Partridge Formation in adjacent New Hampshire, and to amphibolites found in the Ordovician Cram Hill Formation of the pre-Silurian sequence around the Chester and Athens domes. They are very different in mineralogy and texture to mafic phases of the Bethlehem Gneiss; part of the Devonian New Hampshire Series Plutonic Suite.

2. Contain a mappable, 50- to 100-meter-thick by 1 km long body of coarse-grained, massive, chlorite-muscovite-quartz-plagioclase-garnet sill of tonalite. In addition, several bodies of massive felsic volcanics with quartz and plagioclase fragments have been mapped within the granofels to the east in New Hampshire, along the same structural position as the amphibolite and tonalite. Both the tonalite and the volcanics are very similar to igneous rocks mapped by Armstrong in the Ordovician Cram Hill Formation, in the Brattleboro, Newfane, Townshend, and Saxtons River quadrangles, to the south and west of this area (Armstrong, 1994; Ratcliffe and Armstrong, 1996). It is important to note that the Cram Hill has been tentatively correlated with the Partridge Formation in previous studies (Doll and others, 1961).

Rusty-weathering, sulfidic schist and quartzite, along with interbedded gray schist, present on Oak Hill, immediately north of this exposure, have been previously mapped as Partridge Formation (Thompson and Rosenfeld, 1979; Doll and others, 1961). In both of these studies, the Partridge was shown in direct contact with the Fitch, to the west, along a complex appendage, with the Clough terminating along a triple-junction contact with the Fitch and the Partridge. The Partridge rocks were shown terminating in a fold closure immediately south of Oak Hill. Based upon our mapping, we interpret the bedded granofels to be part of the Ordovician Partridge Formation and not part of the Fitch Formation, as in previous studies. The interbedded nature of the granofels with Partridge Formation-like sulfidic schist and quartzite, and the presence of tonalite, felsic volcanics and amphibolite are consistent with this reinterpretation. Our mapping implies that a significant amount of what had been mapped as granofels within the Fitch, west of, and thus, underneath the east-dipping Clough Quartzite (the overturned limb of the Skitchewaug nappe; Thompson and Rosenfeld, 1979), is actually part of the Ordovician section below (and thus, right-side-up) the Clough Quartzite. Since this belt of granofels can be traced across the river into New Hampshire, a large area of what had been interpreted to be Fitch may be Partridge. This raises an additional question: If rocks previously mapped as Fitch (rocks above the Clough) are actually within the Partridge Formation (below the Clough), and the Acadian nappe geometries are essentially based upon the stratigraphic succession, how does this modification of stratigraphy affect the nappe model?

Return to vehicles and make a U-turn. Return to Gage Street, turn right.

Mileage	
19.0	Turn right onto 121 East.
19.2	Junction with Pleasant Street, bear right into village of Bellows Falls.
19.4	Junction with U.S. Route 5, turn right onto 5 South. View of Connecticut River and Fall Mountain to left (east).

19.8	Cross bridge over Saxtons River. Exposures of Bethlehem Gneiss below.
22.0	Bear right off of U.S. 5 and follow sign for I-91.
22.7	Turn right onto I-91 North.
24.2	Cross bridge over Saxtons River.
25.1	Roadcuts of staurolite-bearing schists mapped as Waits River Formation.
25.2	Pull off on right at large roadcuts for Stop 5.

Stop 5:
Roadcuts of large staurolite-bearing schist of the Waits River Formation (Vermont sequence; Bellows Falls 7.5 x 15-minute quadrangle).

The purpose of this Stop is to examine staurolite zone rocks that are structurally below the Skitchewaug Mountain thrust within rocks of the Vermont sequence.

The roadcut consists of ilmenite-chlorite-plagioclase-garnet-staurolite-muscovite-quartz metapelite with some millimeter to centimeter-scale compositional layering of metapelite and quartzite or quartz-rich pelitic horizons. The rocks here were originally mapped as the Littleton Formation in the Skitchewaug nappe and are now reassigned to the Vermont sequence. The most obvious metamorphic feature is the large staurolite (Figure 6b; Figures 7a and 7b). The staurolite is late in the paragenetic sequence and overgrew early foliations as well as included garnets. The staurolite producing reaction is garnet + chlorite + muscovite = staurolite + biotite + H_2O. An interesting note about this particular paragenesis (late, large staurolite) is that it is observed along strike to the south into Massachusetts and to the north to the Littleton area.

No P-T paths have been calculated from rocks at this locality, but a staurolite zone rock from along strike to the north (location BF-18 of Spear and others, 1990, and Spear, 1992) has yielded the P-T path shown in Figure 10. The path shows simultaneous increases in temperature and pressure as would be expected from a rock that was being loaded by the emplacement of higher level allochthons.

Note that the staurolite in this outcrop is rather large (5 mm to 2 cm), relatively pristine with only minor replacement by chlorite. Staurolite porphyroblastic textures at this outcrop differ from those at Stop 3 in that:

1. The staurolite here overgrows the S2 fabric, whereas staurolite predates S2 at Stop 3.

2. Staurolite replacement by chlorite here is minimal. Staurolite at Stop 3 is almost completely replaced by muscovite – not chlorite.

3. Muscovite pseudomorphs after staurolite at Stop 3 grow within the S1/S2 intersection, and muscovite is aligned within S2 in the matrix. Thus, staurolite replacement at Stop 3 occurred prior to staurolite growth at Stop 5. Small, idioblastic, matrix staurolite in

rocks near Stop 3 appear to overgrow S2 and represent a second staurolite growth phase related to the same metamorphism responsible for staurolite growth at Stop 5.

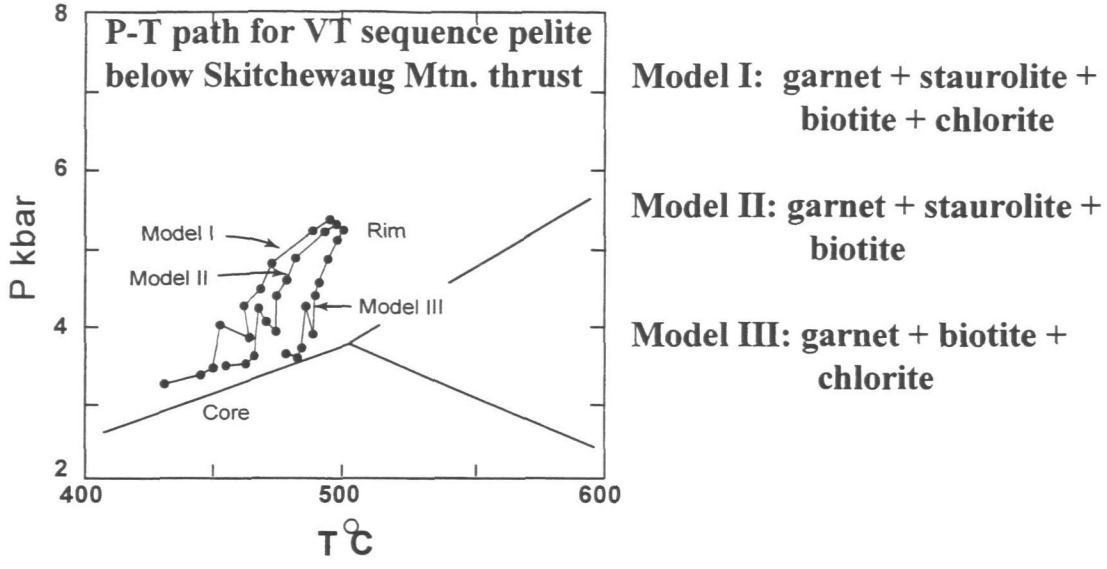

Figure 10. P-T path computed for 3 model assemblages for pelitic sample from lower plate of the Fall Mountain thrust (within the Skitchewaug Mountain allochthon). The three different models correspond to three different and theoretically possible, equilibrium assemblages for this sample. Regardless of the model assemblage, all three paths yield heating and loading paths during prograde regional metamorphism. Points on the individual paths represent individual microprobe analyses on a single garnet.

These relationships suggest that the regional Barrovian metamorphism in this area was at staurolite-grade (or staurolite-kyanite-grade, as we shall see at the next Stop), was the result of D2 thrust transport of the combined Fall Mountain/New Hampshire sequence rocks, and related loading, and that attainment of peak-T conditions occurred immediately after D2 deformation. The abundance of early pseudomorphs in the New Hampshire sequence and the total absence of these pseudomorphs within rocks of the Vermont sequence yields the following chronology of events;

1. The New Hampshire sequence was overridden by the Fall Mountain thrust sheet.

2. The juxtaposed package cooled following continued heating of the structurally underlying, colder rocks.

3. The lower plate rocks were retrograded, possibly due to dewatering of the pluton after crystallization.

4. This package was then transported west along a structurally deeper fault, the Skitchewaug Mountain thrust, which emplaced the Fall Mountain/New Hampshire sequence package on top of the Vermont sequence rocks – resulting in the regional Barrovian metamorphism.

Return to vehicles and continue on I-91 North.

Mileage:

26.2	Roadcuts of sulfidic schist of the Partridge Formation
26.6	Pull off to right for Stop 6.

Stop 6:
Roadcuts of sulfidic schist and granofels of the Partridge Formation (Bellows Falls 7.5 x 15-minute quadrangle).

Rocks at this Stop consist of very rusty-weathered, gray and tan sulfidic schist, biotite granofels, aluminous schist, graphitic, garnet-staurolite schist and rare coticule and calc-silicate beds. These rocks have been mapped by Armstrong (1997b) and by Thompson and Rosenfeld (1979) as Partridge Formation within the northern end of the Wellington Hill structure. Immediately east of this roadcut, at Minards Pond, the Partridge also includes several 5- to 25-meter-thick, massive amphibolite horizons, similar to those described in the text, which probably represent metamorphosed mafic dikes or flows. Many of the quartz veins in this and the outcrop immediately to the south contain large muscovite and kyanite porphyroblasts that occur both within the veins and along their margins with surrounding metapelite and granofels.

The dominant foliation in these rocks is S2, which is typically mylonitic near the contact with the underlying Vermont sequence rocks. Many of the folds seen in this outcrop, however, are related to F3 folding of the S2 foliation, which produces the doubly-plunging synformal structure that is cored by the rocks of the Wellington Hill allochthon. Notice that the F3 folds porpoise through the horizontal at this locality, with some folds plunging south and some north. At the northern closure of this structure (approximately 1.5 km northeast of this locality), the structure plunges south at approximately 15° to 20°, with rocks like those seen here lying on top of gray schist like those seen at Stop 5.

Stand on top of the outcrop and look west across the highway, you will see sub-garnet-grade rocks of the Waits River Formation that show substantial evidence for retrogression. The mylonitic fabric in those rocks is related to the Westminster West fault zone (Figure 3), which cuts the Skitchewaug Mountain thrust and places structurally lower rocks from the east upon structurally higher rocks to the west. Although movement was primarily strike-slip (sinistral), observed kinematic indicators are associated with a transport direction that had a substantial dip-slip (thrust) component as well. This fault is responsible for localized post-peak-T retrogression (replacement of the Barrovian staurolite and garnet by late chlorite) in this area.

Note that we are only about 1 km to the east of chlorite zone rocks. If the chlorite zone represents temperatures no greater than 450° C and the staurolite isograd appears at 550° C, this requires a metamorphic field gradient of ca. 100°/km, which is too high to represent a geothermal gradient for Barrovian metamorphism. This implies a major structural discontinuity between the chlorite and staurolite zone. There is evidence for extensive alteration near the western part of the staurolite zone where staurolite is replaced by chlorite. This chlorite alteration may well have been produced during emplacement of the staurolite grade rocks onto low-grade rocks along the Westminster West fault zone. Fluids required for the alteration may have been derived from the dewatering of the lower grade rocks during emplacement. It is also possible that this structure explains, at least to some extent, the discrepancy in $^{40}Ar/^{39}Ar$ ages between Vermont and New Hampshire, mentioned in the geochronology section of the text, above.

Return to vehicles and continue on I-91 North.

Mileage:

27.0	Roadcuts of sub-garnet grade rocks of the Waits River Formation.
27.6	Intermediate to mafic volcanics and volcaniclastic rocks of the Waits River Formation in median roadcuts.
29.0	Roadcuts of Waits River Formation.
29.7	Cross bridge over Williams River.
31.5	Peraluminous New Hampshire Series granite dike intrudes Waits River Formation in median roadcut.
33.3	Interlayered schist and limestone with volcanic and volcaniclastic rocks of the Waits River Formation at rest area.
34.7	View of Skitchewaug Mountain at 1 o'clock.
35.6	Cross bridge over Black River. View of the Connecticut River to right (east).
36.0	Take Exit 7 off I-91. Go west on Route 11.
36.8	Turn right into Holiday Inn Express hotel and Howard Johnson's restaurant parking lot for Stop 7.

Stop 7:
Contact between volcanic and volcaniclastic rocks and gray schist with conglomerate horizons of the Waits River Formation (Springfield 7.5 x 15-minute quadrangle).

Walk over to the roadcut on the north side of Route 11, just west of the parking lot. Rocks on the east side of the cut are gray, carbonaceous, quartz-rich phyllite and schist with interbedded, thin gray quartzite, and matrix supported conglomerate previously mapped as Littleton Formation. Greenstone and green chlorite-muscovite-quartz-plagioclase schist on the west side of the cut are volcanics previously mapped as Putney Volcanics. This outcrop has been visited by many different geologists over at least the past thirty years. Thompson and others (1993) interpreted the contact between the gray schist and phyllite and the green volcanics as the boundary between the Vermont and New Hampshire sequences. The presence of conglomerate was interpreted as evidence that the contact was an erosional unconformity with younger rocks to the east. This outcrop is considered by many workers to be one of the best exposures of the contact (the "chicken yard line", or CYL, of Hepburn and others, 1984) between the Vermont and New Hampshire sequences.

Our mapping shows that conglomeratic quartzite with gray phyllite and polymict conglomerate units crop out both east and west of the proposed CYL at different places in the section. The gray phyllite and schist is also lithologically similar on both sides of the contact. Our mapping shows that volcanic rocks, limestone-bearing schist and phyllite, and conglomeratic quartzite are present in a wide zone on both sides of the contact, suggesting that either there is no unconformity or that there may be several local disconformities. Generally, there is a decrease in the amount of volcanic and limestone-bearing rock and a corresponding increase in the amount of quartzite and conglomerate east of where the CYL contact was previously defined but the transition cannot be assigned to a single horizon; this makes it difficult to assign regional significance to any particular contact. For these reasons, we interpret the rocks as a continuous sequence devoid of any recognizable unconformities or evidence of hiatus in the depositional record. Thus, all of the units are included in the Waits River Formation. The use of the name Littleton for any part of this sequence would imply a coherent stratigraphic column that would join the Vermont sequence Waits River Formation with the New Hampshire sequence Littleton Formation, and exclude the traditional New Hampshire sequence units, the Clough Quartzite and the Partridge Formation, from immediately below the Littleton. We have, therefore, excluded the usage of Littleton Formation for the easternmost carbonaceous schists in this area.

Start by looking along the eastern margin of the outcrop, and progressively work your way west. Note that bedding, defined by interlayered quartzite and schist, is parallel to the regional S1 foliation. Locally, tops can be found that reverse across the outcrop, and can be explained by observable F2 isoclinal folds that fold bedding and S1 foliation. Similar F2 folds are common throughout the area and complicate the use of limited topping information. Post S2 shearbands are also present here and have shallow-dipping axial surfaces. At the gray schist-greenstone contact there is a locally well-developed fault fabric parallel to the dominant schistosity in the rock. Just to the right of the contact, you should

see several medium-gray lenses of sheared vein-quartz-pebble conglomerate. Some argument has arisen in past Stops here as to whether these pods are indeed sheared conglomerate or just mylonitized vein quartz. If you look up about ten feet, however, you will see a larger pod that has distinctive stretched pebbles, with aspect ratios (in the two dimensions you are looking at) of approximately 10:1. The texture associated with these pods (the ribbony or striated nature of the quartz within them) is typical of highly deformed pebble conglomerate in this area. The shear fabric does not affect the map pattern, and cannot be traced to any mapped faults in the area. The continuous stratigraphy and lack of any significant faults suggests that this contact is simply a sheared contact within Vermont sequence rocks.

Our mapping of the distribution of the volcanic rocks in this area indicates that at least five different belts occur from west to east, at this latitude. Although the westernmost two belts are the same belt (see Stop 2 discussion), the three easternmost belts may be distinct volcanic sequences. Another belt of felsic metavolcanics, lithologically similar to the one at the next Stop (Stop 8), is present across-strike, to the east of the belt mapped here. Our mapping shows that it does not connect with this belt, but it was shown as the continuation of the Putney Volcanics on a map presented in Thompson and others (1993).

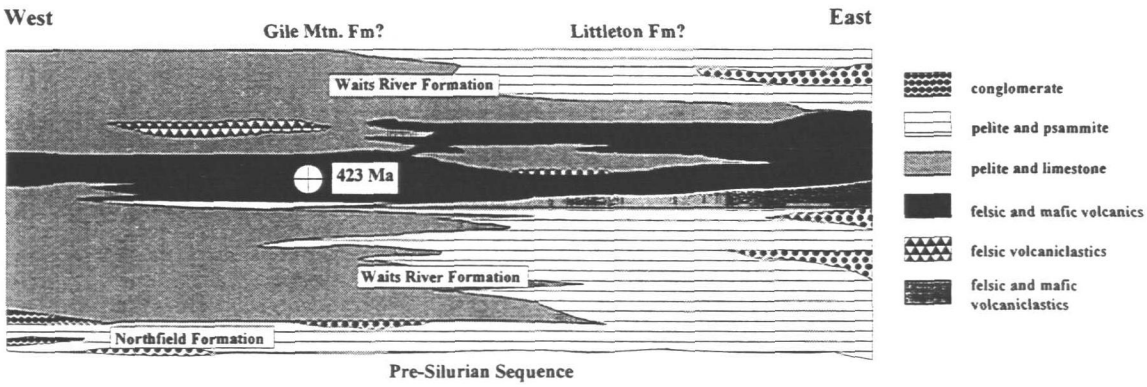

Figure 11. Diagram showing the distribution of units within the Connecticut Valley trough.

If the stratigraphic relationships presented here are correct, then the volcanic rocks at this Stop do not occur at the top of the CVS, but rather somewhere in the middle or even near the stratigraphic bottom of the CVS. Figure 11 shows a schematic model for the distribution of metasedimentary rocks and metavolcanic rocks in the Vermont sequence. The regional stratigraphic and tectonic implications of this interpretation and model will be discussed at many of the upcoming Stops.

Return to vehicles and turn right out of Holiday Inn onto Route 11 West.

Mileage
37.8 Pull off to right for Stop 8.

Stop 8:
Felsic volcanic rocks of the Waits River Formation including the dated rock (423±4 Ma) of Aleinikoff and Karabinos (1990, sample VT/Sp 1-85) and Hueber and others (1990) (Springfield 7.5 x 15-minute quadrangle).

These rocks consist of well-layered, medium- to coarse-grained, chlorite-muscovite-quartz-plagioclase metavolcanics with some horizons containing quartz fragments or lapilli. Variation in abundance of these fragments helps to define bedding. Notice that there are numerous, 1- to 10-cm-thick layers of finer-grained felsic material. Near the right-hand (west) side of the roadcut (across from the junction with Goulds Mill Road), there is a 50-cm-thick fine-grained layer, dated by Aleinikoff and Karabinos (1990), that is in sharp contact with the coarse-grained volcanics. The fine-grained layer is folded, but the contact is always parallel to the dominant foliation, S1. In thin section , and with a hand lens, S1 is visible within the fine-grained layer. Also present, especially within the coarse-grained material, is a second foliation, S2, which is axial planar to the folds here (F2). F2 plunges steeply down dip. If you look closely on horizontal surfaces, you can see folded S1 in F2 hinges.

Aleinikoff and Karabinos (1990) analyzed zircons from the fine-grained layer (also reported in Hueber and others, 1990). They acquired a nearly concordant U-Pb 207/206 age of 423±4 Ma. They interpreted the layer as a fine-grained dike that cut the dominant foliation within the coarser-grained host. Our analysis of this outcrop suggests that there is no evidence of a fabric truncation between the fine- and coarse-grained material, and that the angular discordance between fabrics can be explained by the obliquity between S1 and S2, and the lack of penetrative S2 within the fine-grained material which gives an apparent, cross-cutting relationship; S2 refracts across the fine-grained layer, and where folded, the contact (and layer-parallel S1) appears to truncate S2 within the coarse-grained material. A careful examination of bedding orientations within the coarse-grained material, however, shows that it is completely concordant with the fine-grained layer's contacts.

If the dated layer were a dike, and it did cut S1, then S1 in the coarse-grained rock would have to be pre-423 Ma. If this were true, then the deformation, and hence, the rock itself, would have to been deformed in either the Taconian Orogeny (making the rocks here Pre-Silurian), or some previously unobserved pre-423 Ma, but post-Taconian orogenic event. Based upon Silurian fossil ages from the Shaw Mountain Formation, and other regional considerations, both of these hypotheses seem very unlikely. We interpret the layer as a bed because of the lack of unequivocal cross-cutting relationships and the presence of

many similar, yet thinner, layers within the exposure. We therefore interpret the U-Pb data as the Silurian age for the deposition of the felsic volcanic unit at this locality.

Return to vehicles and turn right onto Goulds Mill Road. Cross bridge over Black River.

Mileage	
38.6	Junction with Route 11, turn left (east).
39.2	Pass under I-91.
39.8	Turn left onto U.S. 5 North, before Cheshire Toll Bridge.
40.3	Sulfidic schist of the Partridge Formation on left (west).
40.5	Pull off to right side of U.S. 5 for Stop 9. Proceed south along road to roadcut on the west side of Route 5. **BEWARE OF TRAFFIC!**

Stop 9:
Partridge sulfidic schist and amphibolite dikes (Springfield 7.5 x 15-minute quadrangle).

The purpose of this Stop is to show the two main lithologies that occur within the Partridge Formation here at Skitchewaug Mountain. These include very rusty-weathering, sulfidic, pyrrhotite-chlorite-plagioclase-quartz-muscovite schist with rare garnet. The second lithology is hornblende-plagioclase amphibolite. The amphibolite is very similar to the amphibolite in the Partridge in the Bellows Falls and Walpole quadrangles. It also has similar sharp contacts with the schist, and in some places, fine-grained margins that may be relict chilled margins suggesting that some amphibolites may be dikes. The lack of continuous bedding within the schist in these outcrops makes it difficult to evaluate the possibility of truncations of intrusive origin.

Structurally, the dike-schist contacts and relict compositional layering within the schist are generally parallel to the dominant foliation in the New Hampshire sequence rocks at Skitchewaug Mountain. Notice that S1 strikes northeast and dips to the northwest. S1 formline maps indicate that S1 is deformed into a doubly-plunging, or domal pattern, with Partridge in the core, Clough in the middle, and the stratigraphically younger Fitch Formation on top. The contact of these New Hampshire sequence units with the surrounding gray schists of the Waits River Formation (Vermont sequence), however, is a S2 fault contact. S2 formlines indicate that the S2 fault surface and associated F2 axial surfaces define a doubly plunging synformal, or basinal, geometry (Figure 4, and cross sections in Walsh and others, 1996a). The antiformal New Hampshire sequence therefore tectonically overlies a fault surface that is synformal in shape. This is very similar to the geometry described for the Wellington Hill allochthon (Figure 4); the western continuation of the Skitchewaug Mountain thrust sheet. This interpretation differs from the classic one which depicts the rocks of the New Hampshire sequence involved in a west-verging, recumbent nappe-fold, subsequently refolded about a north-south synformal axis (Thompson, 1954; Robinson and others, 1991).

If our model is correct, then the gray carbonaceous rocks immediately west and east of Skitchewaug Mountain are part of the Vermont sequence and not the Littleton Formation of the New Hampshire sequence. We believe that the entire length of what has been previously mapped as the overturned limb of the Skitchewaug nappe, from Springfield to the Walpole quadrangle, is the fault contact between structurally overlying Clough Quartzite and underlying Waits River Formation. It may be that in places where the Clough and Fitch are absent, the fault has cut down into the Partridge within areas of complex F1 or F2 folding, such as the domal structure at Skitchewaug Mountain. This hypothesis directly questions the premise that all gray schist within this area of the Bronson Hill Belt belongs to the Littleton Formation, and that it is stratigraphically above the Clough-Fitch sequence. Careful remapping and re-evaluation of the New Hampshire sequence, especially in western New Hampshire, appears to be necessary.

Return to vehicles and continue on U.S. 5 North.

Mileage	
41.1	Outcrops of Clough conglomerate on left (west). View of cliffs of Clough at 12 o'clock dipping gently to north.
42.9	Pull off to right before junction with Old Connecticut River Road for Stop 10. Cross over to roadcut on west side of Route 5.

Stop 10:
Fossils in the Upper Clough Quartzite- Skitchewaug Mountain (Springfield 7.5 x 15-minute quadrangle).

This locality was discussed in a papers presented by Boucot and others (1958), and Boucot and Thompson (1963) that focused on Llandovery (Early Silurian) fossils within the Clough in several different locations in New Hampshire and eastern Vermont. The upper generally consists of tan to gray, vitreous quartzite, but most of the exposure here is within the lowermost part of the upper Clough, and consists of massive, poorly-bedded quartz pebble- to boulder-conglomerate with distinctive horizons of weathered, calcite-bearing, siliceous limestone. The weathering of the limestone has produced the distinctive pitted texture of the rock. The presence of abundant fossilized shell fragments (tetra coral, crinoids, brachiopods, and possibly some trilobites, appears to be a death assemblage; these horizons may be relict coquinas in a near-shore to shore environment (Thompson and others, 1993). Ovate fossil remnants, now calcified, can be seen on the surfaces of many of the coquina horizons. Calc-silicate and marble of the Silurian and Devonian Fitch Formation can be seen a few tens of meters north of this locality on the west side of Route 5.

Notice that S1 is nearly absent in the massive conglomerate phase, with S2 only moderately developed, and dipping southwards at about 10 to 20 degrees. S1 is defined by the coquina layers and dips steeply to the north. Steeply north dipping S1 cut by shallow-south-dipping

S2 is consistent with the domal geometry for the folded S1 in the New Hampshire sequence rocks.

Return to vehicles and continue on U.S. 5 North.

Mileage	
43.4	Roadcuts of Fitch Formation.
45.3	Junction with Route 143, turn left onto 143 West.
45.9	Glacial erratics from Mount Ascutney on right. Part of the Ascutney boulder train.
46.6	View of Mount Ascutney at 4 o'clock.
46.6	Pass under I-91. Late shear zone in Waits River Formation rocks.
47.9	Bear right, follow 143 West.
48.2	Outcrops of quartz pebble conglomerate within Waits River Formation on right (north) in pasture.
48.6	Turn left at monument onto Crown Point Road.
48.9	Pull off to right at 117 Crown Point Road for Stop 11.

Stop 11:
Chestnut Hill - matrix supported conglomerate in the Waits River Formation. Outcrops of conglomerate on hill behind house (Springfield 7.5 x 15-minute quadrangle).

This matrix-supported conglomerate occurs as a discontinuous lens (approximately 300 meters long by 50 meters thick) within a gray to black, biotite-bearing schist unit within the Waits River Formation. The large whaleback outcrop in the woods contains numerous rounded to angular clasts of gray to bluish-gray, vitreous quartzite, clear to bluish-gray, coarse-grained vein quartz, greenish-gray grit (which may be volcanic in origin), and rectangular to angular autoclastic fragments of the matrix. Although most clasts are of pebble-size (mm - to several cm), many clasts are several to tens of centimeters in diameter, with one quartzite clast (a boulder) over 25 cm in diameter. The clasts seen here do not appear to have come from a local source (based upon the clast types), but the angular nature of many of the clasts suggests that they may not have been transported very far. Are they remnants of a subsequently eroded higher structural or stratigraphic level within the Connecticut Valley sequence? Several smaller outcrops to the north and west contain clasts that have a preexisting foliation not present within the matrix. This foliation therefore predates the deposition of the conglomerate. If this observation is correct, then the clast source(s) may actually be a lower structural level that was undergoing erosion during deposition of this part of the trough. Conversely, they may have come from a higher structural level that underwent deformation during a previous tectonism. Since no such rocks have been recognized within the cover of the Bronson Hill belt (the New Hampshire sequence), these rocks do not appear to have a

local eastern source (if the foliation is indeed the result of a pre-Acadian event). Several of these outcrops also contain graded sandy horizons which face east towards the contact with the New Hampshire sequence at Skitchewaug Mountain. A probable source may be pre-Silurian rocks of the Taconide zone to the west.

These rocks were previously mapped as Clough Quartzite within the New Hampshire sequence (Doll and others, 1961). The association of this conglomerate here at Chestnut Hill with gray schists, interbedded siliceous limestones, mafic volcanics, and heterogeneous volcaniclastic rocks indicates to us that this unit is part of the Vermont sequence, and specifically part of the Waits River Formation. Although the Waits River Formation does contain numerous quartz-pebble conglomerate horizons that appear to be isolated channels within an otherwise relatively quiescent basin, this is the only locality where we have seen a matrix-supported conglomerate with clasts other than of fine-grained quartzite and milky-white vein quartz. It should be pointed out that a similar matrix-supported conglomerate with schist and quartzite clasts is present within the Clough Quartzite in the Walpole quadrangle, to the south. At that locality, however, the matrix-supported conglomerate is interlayered with clast-supported quartz-pebble and cobble conglomerate that is stratigraphically overlain by calc-silicate and marble mapped as Fitch Formation.

Return to vehicles, make U-turn, and return to Route 143.

Mileage	
49.3	Turn left at Stop sign onto 143 West.
49.5	Turn right onto Eureka Road.
49.7	Outcrops in pasture to left were previously mapped as Putney volcanics. Outcrops on road were previously mapped as Littleton Formation, but contain impure limestones. Rocks now mapped as Waits River Formation.
50.5	Junction with Woodbury Road, continue straight.
51.3	Turn left onto Barlow Road.
51.9	Junction of Barlow Road, Town Farm Road, Mile Hill Road, and Highland Road. Pull off to right for Stop 12.

Stop 12:
Volcanic units in the Waits River Formation (Springfield 7.5 x 15-minute quadrangle).

Outcrops are located in a meadow south of the four road junction .

This Stop illustrates four typical sub-garnet grade volcanic and volcaniclastic map units in the Waits River Formation. Exposures on the eastern side of the meadow include well-layered quartz-plagioclase granofels mapped as the felsic volcanic unit. Some layers within the felsic volcanic unit are mineralogically and texturally similar to the dated felsic rock at Stop 8. The felsic rocks are interlayered with mafic schist, greenstone, and chlorite-biotite-muscovite-quartz-plagioclase schist and granofels towards the west in these meadow exposures. Going west, the next outcrops are of a medium-grained hornblende-plagioclase gneiss or amphibolite, and finally a fine-grained greenstone. Both units are interpreted as mafic volcanic rocks. This sequence of four map units is continuous northwards for approximately 2 km. The felsic volcanic unit and the hornblende-plagioclase gneiss become discontinuous or absent to the north. At Mount Ascutney, 12 km to the north, the belt consists entirely of greenstone and amphibolite. The considerable along and across strike variation in the units is consistent with a volcanic and volcaniclastic origin for these rocks.

Return to vehicles and make a U-turn.

Mileage	
52.7	Turn left onto Eureka Road.
53.9	Road turns to dirt.
54.1	Bear right at fork onto Bowen Hill Road.
54.3	Junction with Old Bow Road, continue straight.
55.2	Stop sign at junction with Downer Hill Road, turn right.
55.7	Cross I-91.
56.3	Roberts Road on left, continue straight.
56.6	Junction U.S. 5, turn left.
58.2	Outcrop on left of felsic volcanics of the Waits River Formation.
58.4	Pull off to right for Stop 13.

Stop 13:
Felsic volcanics of the Waits River Formation (Mt. Ascutney 7.5 x 15-minute quadrangle).

Roadcut located on the west side of U.S. Route 5, Weathersfield, Vermont

This exposure is typical of the easternmost belt of felsic volcanic rocks mapped in the Mt. Ascutney and Springfield quadrangles. These sub-biotite grade rocks consist largely of chlorite-muscovite-quartz-plagioclase schist or phyllite with beds of quartz-plagioclase granofels and lesser greenstone. Locally, the granofelsic layers contain 1-2 mm porphyroclasts that, in thin section, resemble phenocrysts in a cryptocrystalline matrix of quartz and feldspar with accessory epidote, chlorite, and muscovite. These rocks are interpreted as felsic crystal tuffs, and mafic and felsic volcaniclastic rocks. These rocks are similar to the felsic volcanic rocks at Stops 8 and 12, but are not the same map unit and are interpreted as a separate and stratigraphically higher collection of felsic volcanic rocks in the Waits River Formation.

Return to vehicles and continue north on U.S. 5.

Mileage	
60.5	Junction U.S. 5 and Route 131 at light, turn left (west).
61.0	Pass under I-91.
61.1	Turn left into Park and Ride for Stop 14:

Stop 14:
Interbedded schist and impure limestone of the Waits River Formation showing F1, F2, and F3 folds Formation at I-91 Exit 8 roadcut (Mt. Ascutney 7.5 x 15-minute quadrangle).

This roadcut consists of dark- to light-gray, carbonaceous schist and phyllite with interbedded dark blue-gray, rusty brown weathering, impure siliceous limestone, and gray micaceous quartzite. Limestone beds occupy 40 to 50 percent of the outcrop and range in thickness from several cm to 30-cm-thick. These rocks were mapped previously as the Gile Mountain Formation on the State map (Doll and others, 1961) because they crop out east of (and structurally above) a belt of volcanic and volcaniclastic rocks exposed approximately 300 m to the west. These rocks with interlayered limestone are widespread and occur on both sides of the volcanic units, making it difficult to define the Waits River and Gile Mountain Formations on a lithologic basis in this area.

Three fold generations can be seen in the rocks. The earliest folds (F1) are rootless and isoclinal. The second generation folds (F2) are tight to isoclinal and are associated with the dominant foliation in the rock. A third set of folds (F3) with an associated crenulation cleavage weakly deforms the two earlier fold generations. A cretaceous lamprophyre

dike intrudes the deformed rocks parallel to a prominent joint set that trends 290°, and is exposed west of the I-91 south off ramp near the junction with route 131.

END OF TRIP

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Trip B1: THE APPALACHIAN FORELAND AS SEEN IN NORTHERN VERMONT

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INTRODUCTION

Central Vermont provides and excellent field laboratory for studying the tectonic transition between the foreland and the pre-Silurian hinterland of western New England because many of the major lithotectonic units are exposed in a 40 km wide belt (fig. 1) and the underlying Middle Proterozoic granitic basement is present in the small, but tectonically significant, Lincoln massif (fig. 2). This one-day field trip will concentrate on the foreland and the western part of the hinterland.

The rocks of the foreland consist of a basal rift-clastic sequence overlain by carbonate and siliciclastic rocks of the platform whereas the rocks of the hinterland consist of metamorphosed pelitic, psammitic, and mafic volcanic rocks of the slope-rise and outer rise sequence (fig. 3). Major west-directed thrust faults cut the section and become more numerous to the east. The metamorphic grade ranges from essentially unmetamorphosed, although well lithified and cleaved, sedimentary platform rocks of the foreland to kyanite-chloritoid grade in the hinterland. Temperatures estimates range from 200^0 C or lower along the Champlain thrust to nearly 500^0 C in the central part of the hinterland (fig. 2). Using these temperatures and the preserved fabrics along many of the faults, deformation is estimated to have occurred at near-surfaces conditions in the western foreland (2.5 km or less) and at much greater depths in the hinterland (17-20 km) as a result of stacked and highly deformed thrust sheets (Strehle and Stanley, 1986). Based on regional considerations (Stanley and Ratcliffe, 1985) and available isotopic-age analysis, deformation and metamorphism throughout the western (foreland) and central (accretionary complex) part of the belt largely occurred during the Taconian orogeny (Middle Ordovician). Younger Acadian deformation (Middle Devonian) has severely

deformed the Ordovican Moretown and Cram Hill Formations (fore-arc sequence) as well as the Silurian and Devonian section. The Acadian deformation decreases in intensity westward into the Taconian accretionary complex where its influence is thought to be minimal. Future isotopic work must clarify the extent and relative importance of these two orogenic events. During the Early Cretaceous alkalic dikes and normal faults cut the existing geology and represent abortive rifting during the opening of the North Atlantic (McHone, 1987; Stanley, 1980).

WESTERN VERMONT - The structure of the Vermont foreland is dominated by major, north-trending folds and imbricate thrust faults (Doll and others, 1961 and the western part of fig. 2). The largest of these, the Champlain thrust, extends form southern Quebec to Albany, New York and places the older platform sequence over the younger foreland basin. Estimates of westward displacement range from 15 km. to 100 km (Stanley and others, 1987). Seismic traverses across western Vermont show that the Champlain thrust dips eastward at about 15 degrees beneath the hinterland of the Green Mountains and that many of the major folds of western Vermont are formed as duplexes and related structures on this and other faults in the foreland. Along the Green Mountain front several of the larger anticlines appear to be fault-propagation folds (Suppe, 1985; fig. 2). Seismic information suggests that some of the high angled-faults in the western foreland are older than the Champlain thrust whereas others are younger. The increase in deformation towards the Champlain thrust is well illustrated in selected outcrops (Lessor's Quarry and the "Beam") in the Ordovician rocks of thelower plate exposed on the Champlain Islands (figs. 1 and 4).

An eastern, but smaller Hinesburg thrust (figs.1 and 2), places transitional and rift clastic rocks over the platform sequence and, as such, forms the boundary between the foreland rocks of western Vermont and the hinterland. Dorsey and others (1983) have demonstrated that the Hinesburg thrust developed as a breakthrough thrust on the overturned limb of the Hinesburg nappe and therefore is quite different from the geometry of the Champlain thrust. Westward displacement is estimated to be 6 km.

HINTERLAND - In central Vermont the earlier mapping by Cady and others (1962) and Osberg (1952) have been very helpful because it has defined many of the formations of the pre-Silurian hinterland and, as such, has highlighted the areas that must be reexamined in light of plate tectonic theory and recent findings in other hinterlands of the world. Furthermore, the earlier petrologic and isotope work of Albee (1965, 1968), Garlick and Epstein (1967), Cady (1969), Laird and Albee (1981), Lamphere and others (1983), and Laird and others (1984) is particularly important in defining metamorphic facies and establishing the physical conditions and timing of metamorphism and deformation. Much isotopic-age analyses, however, must be done in order to clarify the relative importance of the Acadian orogeny and understanding the thermal history of this part of the orogen (Sutter and others, 1985).

Prior to the 1970's, the formations in the Vermont hinterland were considered to be a coherent, depositional sequence, although considerable internal flattening and shearing were recognized (for example, Doll and others, 1961). As a result of mapping in western Massachusetts (Zen and other 1983) and Vermont (for example, the numerous published work by Stanley and his students, Peter and Thelma Thompson, Doolan and his students, and extensive mapping in southern Vermont by Ratcliffe, Walsh and Armstrong during the last 6 years) ductile faults are now recognized as a fundamental feature of the hinterland. The work in central and northern Vermont have further shown that the faults occurred at different times during the evolution and metamorphism of the orogen (for example, Stanley and others, 1987, and Stanley, 1991. Perhaps the most important of these are the early or premetamorphic faults because they control the distribution of major rock types and have been transported the farthest. The older high-pressure metamorphic series recognized by Laird (1984, 1987; fig. 4) in the mafic schists of the hinterland (fig. 2) formed either during or after these early thrust faults. As a result, the only reliable criterion for these faults is the truncation of units on either the footwall or the hangingwall. Later synmetamorphic and late metamorphic faults offset these older thrusts. Displacement criteria for many of these faults indicate east-over-west motion.

ITINERARY

The first two stops are located on South Hero Island, the southern-most island of the larger islands in northwestern Vermont (fig.4).

Stop 1 LESSOR'S QUARRY (Directly south off of Sunset View Road) - This quarry is located in the fossiliferous Glens Falls Limestone of Middle Ordovician age (fig.5). The Glens Falls contains excellent example of graded beds of fragmented fossil material representing distinct pulses of carbonate sedimentation possibly caused by storms. Study the stratigraphy of the larger blocks.

The quarry contains a number of imbricate bedding plane thrusts which can be studied by mapping all the walls. The most conspicuous thrust, called the Lessor's Quarry thrust, is well exposed on the north wall. Conspicuous slickenlines, east-dipping fault-zone cleavage, and bent S1 cleavage indicate that the upper plate moved to the west-northwest over the lower plate. The amount of displacement is unknown. A conspicuous syncline forms the lower plate on the east side of the north wall. This syncline folds an older bedding-plane thrust (blind thrust) which dies out in small faults and pressure solution features on the western limb of the syncline. Multiple cleavages along the blind thrust indicate that it moved westward before it was folded by the syncline. The syncline and its eastern antcline probably developed over a deeper ramp. You will note that the Lessor's Quarry thrust is not folded by the syncline or its eastern anticline, but instead truncates the blind thrust. We believe that the Lessor's Quarry thrust is an excellent example of an out-of-sequence thrust in the foreland.

The S1 cleavage is a superb example of pressure solution. You will note that it is discontinuous, filled with foliated, black clay-like material. S1 off sets bedding where the bedding-cleavage angle is less than 90 degrees. Furthermore, fossils are truncated by the cleavage. The calcite from the pressure solution was largely deposited in fractures and along fault surfaces.

Discussions will focus on the structural history preserved in the quarry.

Stop 2 "THE BEAM" (fig. 6) (Outcops on Rt 2 1 mile east of Lessor's Quarry). This is a suberb outcrop that serves as a field laboratory for research and teaching of foreland deformation. Please study it. Use cameras but not hammers. This exposure has been thoroughly discussed in the following paper: **Stanley, R.S., 1990, The evolution of mesoscopic imbricate thrust faults - an example from the Vermont foreland: Journal of Structural Geology, v. 12, p. 227-242.**

The outcrop is located in the Cumberland Head Formation (Middle Ordovician) 5 miles west of the exposed front of the Champlain thrust fault or approximately 4600 feet (1400 meters) below the restored westward projection of the thrust surface (fig.4). The major questions that will be discussed are: 1. How do ramp faults form?, 2. Are there criteria to determine if imbricate thrust faults develop toward the foreland (west) or the hinterland (east)?, 3. What is the relation between faulting and cleavage development?, 4. What processes are involved in the formation of the fault zones?, 5. Are there criteria that indicate the relative importance and duration of motion along the fault zone?, 6. Is there evidence to suggest that abnormal pore pressure existed during faulting?, and finally 7. What is the structural evolution of the imbricate faults and cleavage formation? The first six questions will largely be addressed by evidence at the outcrop. The seventh question will be answered by palinspastically restoring the imbricated and cleaved sequence to its undeformed state.

ISOTOPE ANALYSIS, SYSTEMS DYNAMICS AND THE "BEAM" - AN INTERLUDE (Stanley and others 1996)

In order to determine if the calcite in the veins was derived locally from the surrounding shales, samples of the micrite and host-rock (shale) microlithons, cleaved shale, and of sparry calcite from the younger veins and floor and ramp thrusts were analyzed for stable isotopic composition. The mean carbon and oxygen isotopic composition of the younger calcite veins is similar to that of the host rock (1.3 +/- 0.02 ‰ and -7.2 +/- 0.1 ‰, respectively) which suggests local dissolution of local matrix carbonate and precipitation as vein calcite under fairly constant temperatures. Calcite veins related to different faulting events also show similar carbon and oxygen isotopic composition,

suggesting that deformation occurred at each fault under similar conditions. However, fault veins were found to be enriched in 13C relative to host rock by approximately 0.2 ‰, still indicating dominant matrix control on isotopic composition. This difference is not considered to be significant at this scale, although it may indicate a temperature difference of a degree or two between carbonate dissolution and precipitation

Based on the foregoing geological and isotopic evidence a dynamic simulation using STELLA is constructed based on the two difference deformation mechanisms for the beam - pressure solution in the calcareous shales and fluid-induced faulting in the micrite "beam". We suggest that the calcite-rich fluids expelled from the surrounding shales developed abnormal fluid pressures along the faults surrounding the "beam". To represent this dynamic system in STELLA four basic components of the system are represented by icons (fig. 7). The rectangular icons are stocks which represent integrals or accumulations of quantities over time (for example, fault zone fluid, uncleaved rock, displacement). The arrows with values are flows which represent quantities per unit time or differentials (for example, movement rate, fluid outflow, pressure solution). The circular icons or converters represent specific values, mathematical relations or graphs that feed data into the flows or record mathematical relations during the simulation (for example, dissolution fraction, hydraulic conductivity, decay constant). Most of the icons are hooked together by arrows that represent x to y (arrow head) dependency. These are the 4 basic tools of system thinking as originally described by Jay Forrester of MIT and subsequently amplified and popularized by Barry Richmond and associates at High Performance Systems Inc. (1990, 1992, for example). The values and description of each icon are documented in equations which will be available during the conference. Not all the parameters in the simulation have been measured for this outcrop. Such values as hydraulic conductivity are taken from standard hydology references. This parameter should change substantially from uncleaved rocks where it is probably similar irrespective of direction compared to cleaved rocks where the hydraulic conductivity is probably far smaller (slower) across the cleavage than it is parallel to the cleavage. Other such values as dissolution fraction are a guess since we are unaware of pressure solution reaction rates. Thus the model presented here is a valid representation of the geological system. It is, however, not verified since many the important inputs have not been measured.

Figure 8 illustrates a single run for 2000 time units which represent 200,000 years in this simulation. Note that the "time" between faulting events, represented by the "sawtooth" graph, increases as time approaches 130,000 years. Beyond this time faulting ceases since most of the uncleaved rock has been converted into cleaved rock (98%). Thus the rock is essentially dewatered and fluid is no longer an aid in faulting. In short, the outcrop has strain hardened and further deformation has to be in the form of other local mechanisms (cleavage-controlled faulting, for example) or deformation must move toward the craton where undeformed rocks are located.

Note how the graph of displacement changes during the simulation. During each faulting event the graphs segment is vertical and records "rapid" displacement. In this simulation this distance would occur within a 100 year period which represents a single time step in the calculation. During the intervals between fault events, the graphs are curved and represent the slow shortening provided by the pressure solution process. As time progresses during the simulation, these curve segments are more horizontal reflecting the decreasing rate of cleavage formation as less and less undeformed shale is available.

The time step or dt used in this particular simulation represents 100 years. This value was selected so that the behavior of the subsystems for fluid-controlled faulting and pressure solution could be demonstrated within a reasonable length of computational time. We believe that this time step is too large and should be reduced by at least a factor of 10 to represent more accurately the displacement during faulting. Furthermore, the vertical displacement on the graph that represents a "single fault event" is far too simple and must represent a whole series of small displacements or microfaulting events that result from the continual alternation between pressure solution and fault displacement. Thus the dynamic simulation is only a gross simplification of what was probably occuring during the shortening at the "beam". It does, however, illustrate quite nicely the "feedback" processes that can occur during deformation and the benefit of viewing geological processes from a systems dynamic perspective.

Continue southwardward on Rt 2 through South Hero Village to Interstate 89 for a distance of approximately 10 miles (16 kilometers). Travel south approximately 10 miles (16 km) on Interstate 89 to exit 14 where you turn west on Williston Road (Rt 2). Travel

west several miles until you reach the Burlington waterfront. Turn north on Rt 127 and travel 3 miles (4.8 km) to Burlington High School. Turn left onto the road at the light and proceed to the Episcopal Diocece of Vermont. **Read Stanley, R. S., 1987, The Champlain thrust, Lone Rock Point, Burlington, Vermont: Geological Society of America, Centennial Field Guide for the Northeast Section, p. 67-72** for further instructions and a discussion of the outcrop.

Stop 3 CHAMPLAIN THRUST FAULT AT LONE ROCK POINT. This is by far the best exposure of the Champlain thrust fault in the northern Appalachians. Furthermore, it is one of the best exposures of a foreland thrust fault in the United States. Wave erosion of the weaker shale of the lower plate has exposed the fault zone for a distance of over 1 mile (1.6 km). Copies of the aforementioned paper will be distributed to participants. Refer to figures 2, 12,13,14, and 15.

Return south along Rt 127 to the junction with Rt 2 in Burlington. Travel east several miles to South Burlington where you will find the junction of Rt 2 and Rt 116. Travel south on Rt 116 approximately 10 miles (16 km) to the junction of Rt 116 and the Hineburg-Shelburne Road. A 4-way stop light marks the junction. There is a small shopping center on the west side of the road. Turn east (left) onto Champlain Valley High School Road. Go pass the high school and take the first left (north) at the 4-way junction onto Richmond Pond Road. Travel a mile up the road an park near the barn on the west side. Walk west down the dirt road about 0.25 miles (0.4 km) and take the dirt road heading north to the outcrop of the Hinesburg thrust fault which is located at the base of the wooded hill.

Stop 4 HINESBURG THRUST FAULT (fig. 9). This is the type locality of the Hinesburg thrust fault which, for central and north Vermont, is the major structural boundary between the carbonate-siliciclastic platforms of the foreland to the west and the highly deformed, and metamorphosed rift - drift sequence of the Taconian hinterland to the east. Westward displacement is in the order of 4 miles (6.4 km). Although the Lower Ordovician carbonate rocks of the lower plate are poor exposed here, the upper plate of argillaceous quartzite of the Lower Cambrian Cheshire Formation and stratigraphic lower

Fairfield Pond Phyllite form the cliffs along the western front of the hill. These rocks contain many such fault related structures as deformed extension fractures, isoclinal folds, shear bands, mylonitic textures, slickenlines, and pressure fringes around pyrite. Many of these have been analyzed by Stanley and his students (Gillespie, 1975, Strehle, 1985; and Stehle and Stanley, 1986). Stanley, Martin and Smith (1993) studied the "z" shaped extension fractures. The results of this and earlier work will be discussed. We will concentrate on the following features:

1. Is the conspicuous layering in the basal part of the upper plate, bedding or foliation?
2. The presence of asymmetric fault-related folds and their relation to the Hinesburg recumbent fold.
3. The prominent mineral lineation consisting of elongate quartz and clusters of quartz grains.
4. "Z" shaped quartz veins, their correlation and history throughout the upper plate.
5. Rare, east-dipping shear bands.

Return to the junction of Rt 116 and Champlain Valley High School Road and turn south. Travel through the village of Hinesburg 16 miles (26 km) to the junction of the New Haven River and Rt 116 just north of the village of Bristol. Turn east onto the Lincoln Gap Road.

STOP 5- WESTERN CONTACT OF THE EASTERN LINCOLN MASSIF (fig. 10) (second bridge east of Lincoln on the Lincoln Gap Road). This outcrop shows the western contact between the Middle Proterozoic rocks of the Eastern Lincoln massif and the overlying basal conglomerate of the Pinnacle Formation. Generally this contact is an erosional unconformity which dips steeply west or is slightly overturned with gentle north-plunging parasitic folds (DelloRusso and Stanley, 1986), Here, however, the contact is offset across a west-directed thrust fault Cobb Hill thrust fault) and the associated folds in the cover have been rotated toward the transport direction. The Grenvillian foliation of the Middle Proterozoic rocks is progressively overprinted by the Taconian schistosity (K/Ar age of 410 Ma on biotite, Cady 1969) of the cover as the contact is approached from the basement. The basal conglomerate of the Pinnacle is generally a quartz cobble

conglomerate with minor pebbles and cobbles of granitic rocks. Here, however, large granitic cobbles and boulders form lensoid deposits separated by quartz-feldspar metawacke. The origin of this deposit is controversial, but Stanley, following Tauvers (1982), will argue that they represent ancient channels that have been subsequently deformed into nearly reclined folds.

Stop 6 EASTERN CONTACT OF THE LINCOLN MASSIF (fig. 11) - The South Lincoln thrust fault at the bridge at South Lincoln. This outcrop along the New Haven River contains a mylonitic sliver of Middle Proterozoic gneiss in thrust contact with biotite-chlorite schist of the Hoosac (Pinnacle) Formation. A thick conglomerate containing quartzite and granitic gneiss pebbles, cobbles, and boulders in a matrix of biotite-chlorite schist and metawacke overlies the gneiss. The clasts are identical to the Middle Proterozoic rocks in the Lincoln Massif. Late northeast-plunging folds deform the fault zone. These folds are in turn overprinted by late biotite that is randomly oriented across the dominant schistosity. Based on the fact that the feldspar in the mylonitic gneiss has been dynamically recrystallized and the surrounding rocks have been metamorphosed to the garnet grade the temperature is estimated to have been in the order of 435^0 C to 450^0 C or slightly higher (fig. 2). These temperatures would correspond to pressures in the order of 5 to 6 kbars (17 to 20 km) (Strehle and Stanley, 1986). This synmetamorphic fault zone with it sliver of basement is typical of the eastern margin of the Lincoln massif. In fact, about 10 km. to the south fault zones like this become so numerous that they essentially make up the eastern half of the Lincoln massif as it is shown on the Geological Map of Vermont (Doll and others, 1961).

Return on the dirt road directly west of the New Haven River to the junction of the Lincoln Gap Road. Stop by the bridge that leads back to Lincoln. Do not go up the Lincoln Gap Road.

TECTONIC SYNTHESIS (figs 12 - 15)

The tectonic evolution of the western part of the central Vermont transect largely involves the failure of the east margin of sialic crust of ancient North America. Based on

the geometry, regional continuity, and symetamorphic fault zone assemblages the sequence evolved from east to west with the Champlain thrust zone forming the regional decollement

In order to summarize the evolution of central Vermont 5 retrodeformed sections are shown in figures x to 10x. The sections begin with the present cross section for the western part of central Vermont (figure 2). This section shows structural detail, critical isotopic ages, and the temperature and pressure information inferred from mineral compositions or isotopic ratios. We begin with these sections and retrodeform the sequence using sedimentological relations, structural sequence, and metamorphic-isotopic information. By its very nature, any palinspastic analysis for rocks as highly deformed as those in western New England is largely qualitative since key markers and critical age information needed to measure displacement are lacking. The westward displacement of 70 km on the Middle Proterozoic slices, however, can be calculated from their relationship in cross section. The important relations shown in each section are discussed in the captions for figures 12 through 15 and are not repeated here.

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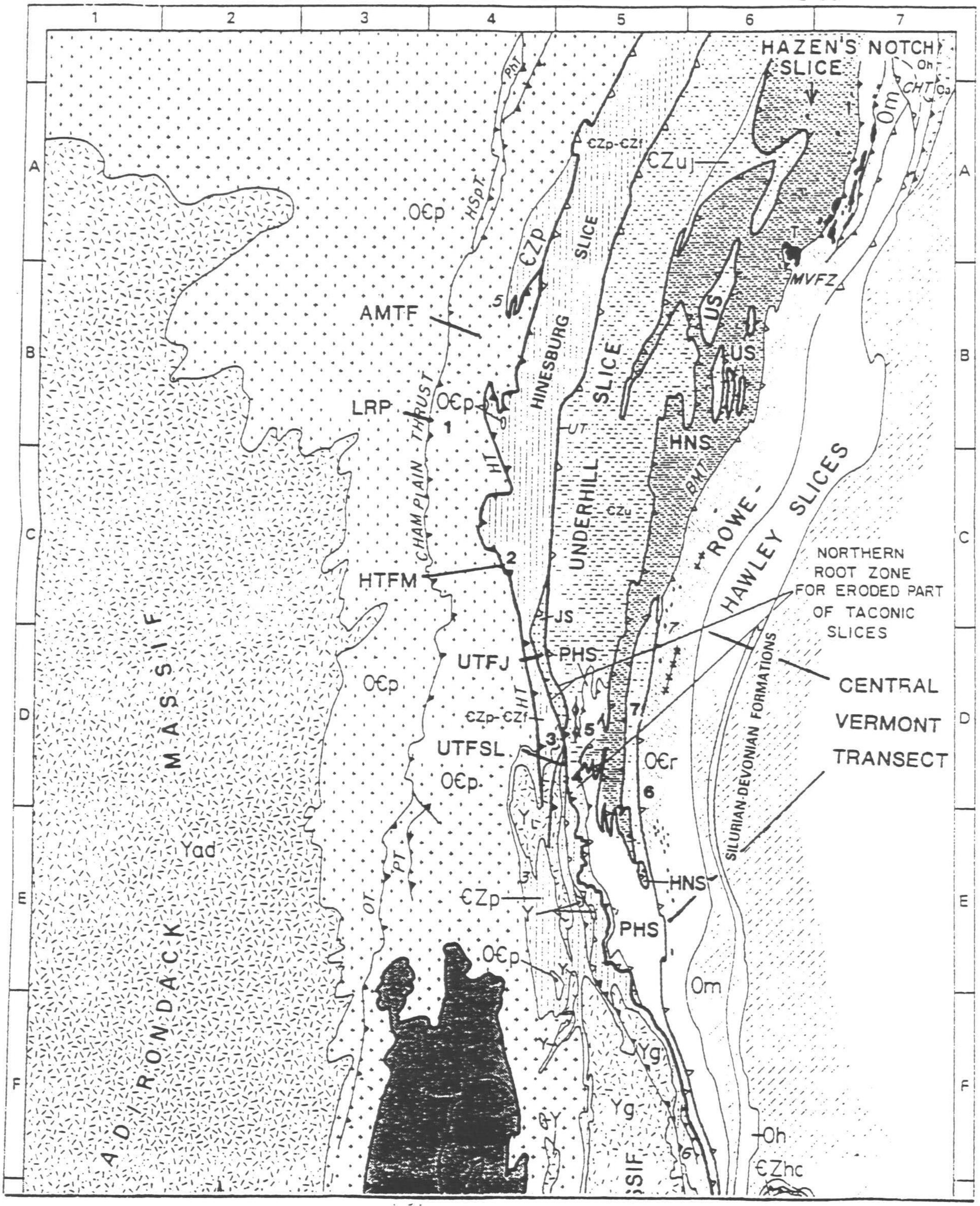
73°30'
73°00'
72°30'
45°00'
44°30'
44°00'
43°30'
HAZEN'S NOTCH SLICE
HINESBURG SLICE
UNDERHILL SLICE
ROWE - HAWLEY SLICES
NORTHERN ROOT ZONE FOR ERODED PART OF TACONIC SLICES
CENTRAL VERMONT TRANSECT
SILURIAN-DEVONIAN FORMATIONS
ADIRONDACK MASSIF
CHAMPLAIN THRUST
AMTF
LRP
HTFM
UTFJ
UTFSL
HNS
PHS
JS
US
MVFZ
BMT
HSpT
HT
UT
PT
OT
CHT
O€p
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€Zu
O€r
Om
Oh
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Yad
Yg
Y

FIGURE 1

FIGURE 1 Interpretative Geologic Map of Vermont and eastern New York modified from Stanley and Ratcliffe (1985, pl. 1, fig. 2A). The following symbols are generally listed from west to east. Yad, Middle Proterozoic of the Adirondack massif; Yg, Middle Proterozoic of the Green Mountain massif; YL, Middle Proterozoic of the Lincoln massif; Y, Middle Proterozoic between the Green Mountain massif (Yg), and the Taconic allochthons (medium grey); OCp, Cambrian and Ordovician rocks of the carbonate-siliciclastic platform (DRIFT STAGE); rift-clastic sequence of the Pinnacle (CZp) and Fairfield Pond Formations (CZf) and their equivalents on the east side of the Lincoln and Green Mountain massifs. CZu, Underhill Formation; CZuj, Jay Peak Member of the Underhill Formation (RIFT STAGE IN HINTERLAND); OCr, include RIFT STAGE rocks of the Pinney Hollow and Stowe Formations; FOREARC SEQUENCE (Missisquoi Group) include Om, Moretown Formation and Oh includes Harlow Bridge and Cram Hill Formations in Vermont and Hawley Formation in Massachusetts. Silurian-Devonian sequence include Shaw Mountain, Northfield and Waits River Formations in the area labelled "Silurian-Devonian Formations". Symbol T in A6 is the glaucophane locality at Tilliston Peak. Short line with X's (Worcester Mountains and Mt Elmore) and line with rhombs (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones if Albee (1968). Irregular black marks are ultramafic bodies (U). , PhT, Philipsburg thrust; HSpt, Highgate Springs thrust; PT, Pinnacle thrust; OT, Orwell thrust; AMTF, Arrowhead Mountain thrust fault; HT; Hinesburg thrust; HTFM, the Hinesburg thrust fault at Mechanicsville; UT, Underhill thrust; UTFJ, the Underhill thrust fault at Jerusalem; UTSL, Underhill thrust fault at South Lincoln; ; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens North slice; MVFZ, Missisquoi Valley fault zone; PHS, Pinney Hollow slice; BMT, Belvidere Mountain thrust; CHT, Coburn Hill thrust; Oa, Ascot-Weedon sequence in grid location 7A.

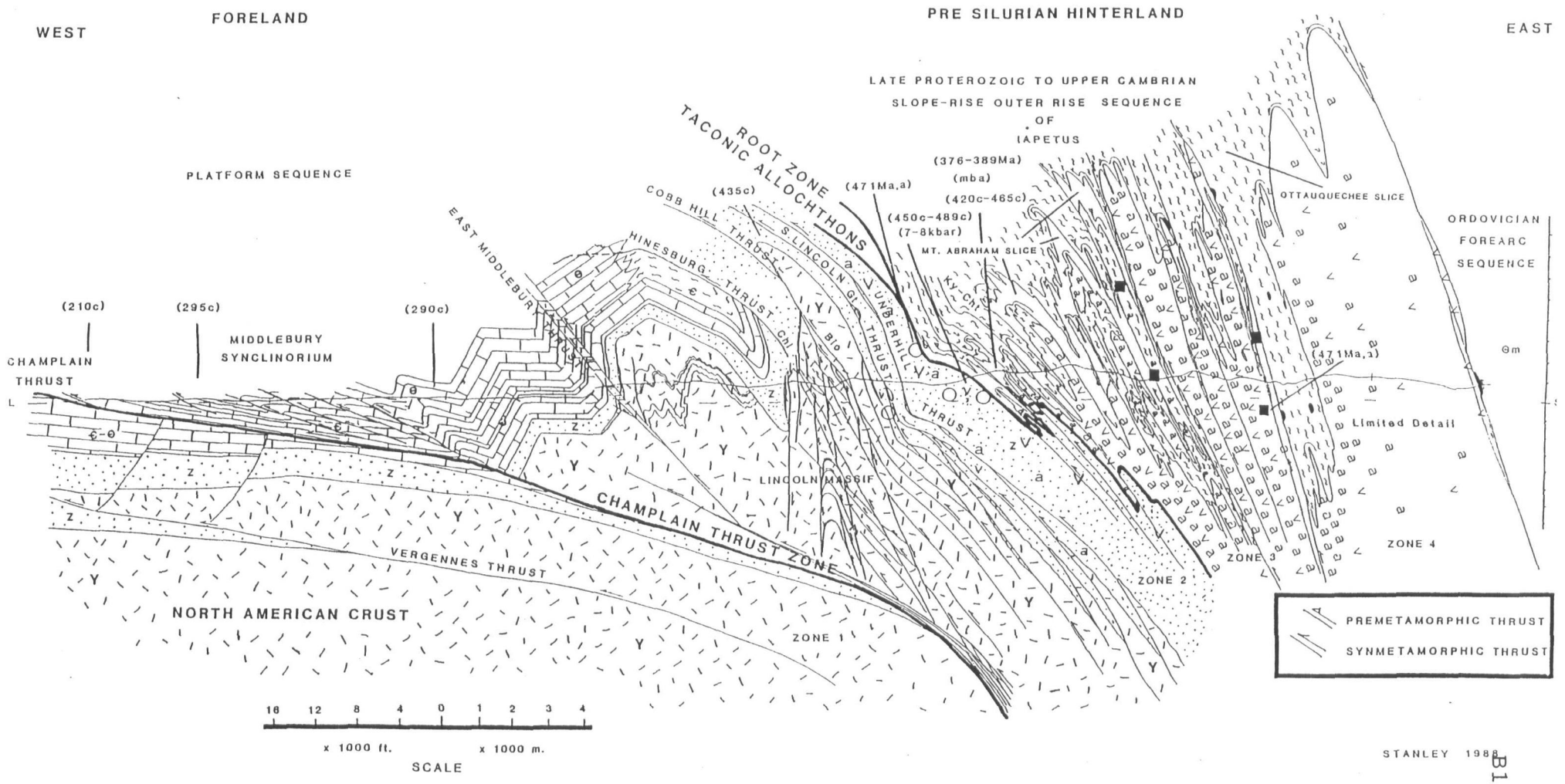

WEST
FORELAND
PRE SILURIAN HINTERLAND
EAST
PLATFORM SEQUENCE
LATE PROTEROZOIC TO UPPER CAMBRIAN
SLOPE-RISE OUTER RISE SEQUENCE
OF
IAPETUS
ROOT ZONE
TACONIC ALLOCHTHONS
(376-389Ma)
(mba)
(420c-465c)
(435c)
(471Ma,a)
(450c-489c)
(7-8kbar)
COBB HILL THRUST
S.LINCOLN GL.
EAST MIDDLEBURY FAULT
HINESBURG THRUST
UNDERHILL THRUST
MT. ABRAHAM SLICE
OTTAUQUECHEE SLICE
ORDOVICIAN
FOREARC
SEQUENCE
(210c)
(295c)
(290c)
MIDDLEBURY
SYNCLINORIUM
CHAMPLAIN
THRUST
LINCOLN MASSIF
CHAMPLAIN THRUST ZONE
VERGENNES THRUST
NORTH AMERICAN CRUST
ZONE 1
ZONE 2
ZONE 3
ZONE 4
Limited Detail
PREMETAMORPHIC THRUST
SYNMETAMORPHIC THRUST
16 12 8 4 0 1 2 3 4
x 1000 ft.
x 1000 m.
SCALE
STANLEY 1988

FIGURE 2

FIGURE 2. Geological cross section through the pre-Silurian foreland and hinterland along a latitude of 44^0 N. The section represents the Taconian Accretionary Wedge. It is based on mapping by Cady (1945), Washington (1987), and Harding and Hartz (1987) in the foreland, by Tauvers (1982) and DelloRusso and Stanley (1986) in the Lincoln massif, and by Lapp and Stanley (1986), Prewitt (1986), Stanley (1986, 1987), Kraus (1987), Haydock (1988), Walsh (1989), Kimball-Falta (1991), and Armstrong (1993) in the Taconian part of the hinterland. North American Middle Proterozoic (Y) crust is shown by the random dash pattern. The overlying foreland rocks consist of Late Proterozoic (Z) rift clastic rocks (stippled pattern) and Cambrian and Ordovician platform rocks (carbonate and siliciclastic rocks overlain by shales) shown by the stacked rectangles. Symbols for the deformed and metamorphosed slope-rise and outer rise sequence correspond to those in figure 1. The sequence designations are those used in Stanley and Ratcliffe (1985) and figure 1 The black lenses in the hinterland sequence represent serpentinites. Zones 1 through 4 are based on mafic rock geochemistry (Coish, 1987, 1988). Temperatures of coexisting mineral assemblages are given as (i.e.290^{0}c) and are based on carbon and oxygen isotope analyses by Sheppard and Schwarcz (1970) for the foreland, oxygen isotope analyses by Garlick and Epstein (1967) and calcite-dolomite and amphibole-plagioclase temperatures (Laird and others, 1984) for the western part of the hinterland. The 471 Ma ages are based on 40Ar/39Ar analyses of hornblende by Laird and others (1984). The 376-389 Ma ages are based on 40Ar/39Ar total fusion ages of muscovite and biotite (Laird and others, 1984; Lamphere and others, 1983). The symbol a refers to amphibole, b refers to biotite, and m refers to muscovite. The metamorphic series assignment is based on amphibole analyses by Laird (1987). The solid squares represent hornblende from the high-pressure metamorphic series and the open circles represent hornblende from the medium-pressure metamorphic series. Based on stratigraphic and sedimentologic information, Stanley and others (1989) suggested that the root zone for the northern extension of the Taconic allochthons is located along a complex zone of pre- to late metamorphic faults that forms the western boundary of the albite-rich and aluminous rocks of the Hazens Notch, eastern Underhill and Mt. Abraham formations. This location is supported by the fact that the metabasalts in the Taconic allochthons are chemically more similar to the mafic rocks of the western Underhill and Pinnacle formations (Zone 2) than they are to the mafic rocks to the east in Zones 3 and 4. The exposed Middle Proterozoic rocks of the Lincoln massif consist of two major anticlines in which the
eastern one has been severely flattened by numerous thrust faults which contain Paleozoic mylonites with east-over-west fabrics. A third anticline is buried beneath the platform sequence to the west of the East Middlebury thrust (Harding and Hartz, 1987). The breakthrough faults (East Middlebury thrust) and geometry of the western anticline suggests that the three anticlines of Middle Proterozoic rocks began their development as fault-propagation folds (Suppe, 1985). The eastern anticline was then flattened by the development of axial-surface cleavage and extensive high-angle faults. The available PT information suggests that this deformation in the massif occurred over temperatures that ranged from 290^{0}C to the west to 435^0 C along the eastern border of the massif. These temperatures would correspond to pressures in the range of 3 kbars (10 km) to the west and 4.5 to 5.0 kbars (16 to 18 km) to the east assuming a standard geothermal gradient of 20^{0}C to 30^{0}C per kilometer (Strehle and Stanley, 1986, fig. 5). This information again suggests that there was a significant tectonic load over the Lincoln massif when these structures formed. The minor imbricate faults of the Middlebury synclinorium are based on work by Washington (1987). The slice of North American crust floored by the Vergennes thrust is an interpretation based on some unpublished seismic reflection work. Shortening across this section is estimated to be between 400-500 km. of which approximately 70 km involved the folding and faulting of North American crust.

LITHOTECTONIC CORRELATION CHART FOR CENTRAL VERMONT

Figure 3

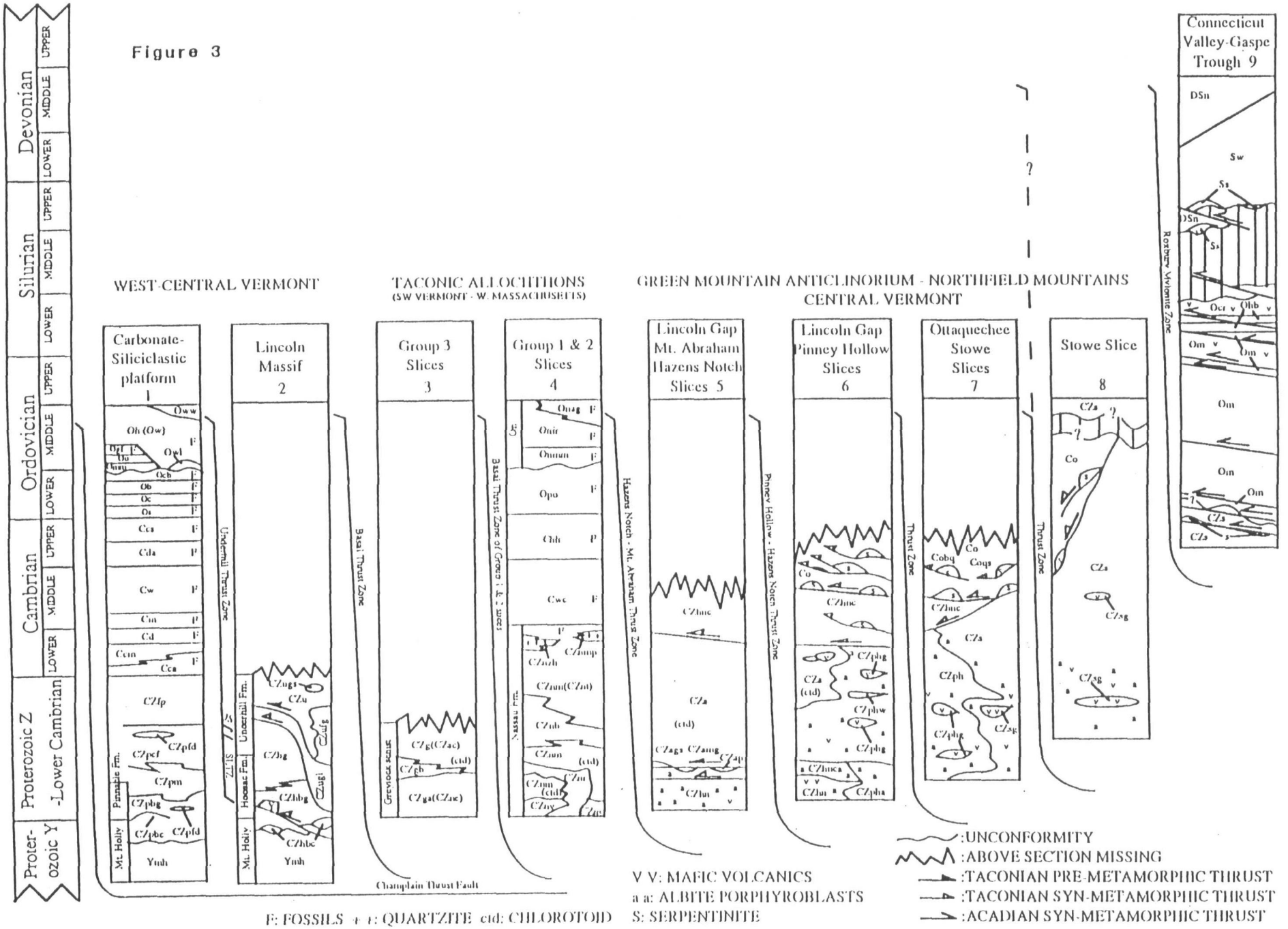


FIGURE 3 Lithotectonic correlation chart for central Vermont. Symbols are explained in Doll and others (1969) and Stanley and others (1988, 1993). The units covered in this field trip are listed in columns 1 and 2. The exposed Proterozoic Y basement of the Mount Holly Complex (Ymh) consists of well foliated granitic gneiss with minor mafic rocks, augen gneiss, quartzite, and aluminous schist (DelloRusso and Stanley, 1986). These rocks are overlain by rift clastic rocks consisting of metawacke, metaarkose, finer grained metaclastic/ metapelitic rocks, and salmon-colored dolostone and associated chloritc metawacke of the Pinnacle Formation (CZp) in column 1. Polymictic pebble to boulder conglomerate (both matrix and clast supported), quartz pebble conglomerate are found along the profound unconformity separating the basement from its overlying rift cover. Correlative rocks to the east (CZu, CZh, column 2) are more pelitic, contain mafic schist and thin quartzite, but still contain metawacke although it is far less abundant than it is to the east across the South Lincoln-Underhill thrust (fig. 1). Polymictic matrix supported conglomerates are found between slivers of Mount Holy basement and the eastern cover (DelloRusso and Stanley, 1986). The classical fossiliferous carbonate-siliciclastic platform sequence first described by Wallace Cady (1945) (from Cc to Ogl of column 1) overlies the rift clastic sequence and represents the drift stage for the western part of the Appalachians. These rocks form much of the geology of the upper plate or slice of the Champlain thrust fault. These rocks pass upwards into calcareous and non calcareous Middle Ordovician shales (Osp, Oi, Oh, Ow) that represent clastic debris eroded off the Taconian accretionary prism emerging to the east and deposited in a foreland basin during Middle Ordovician time. The Champlain thrust fault places the drift sequence on top of this shale - rich foreland section.

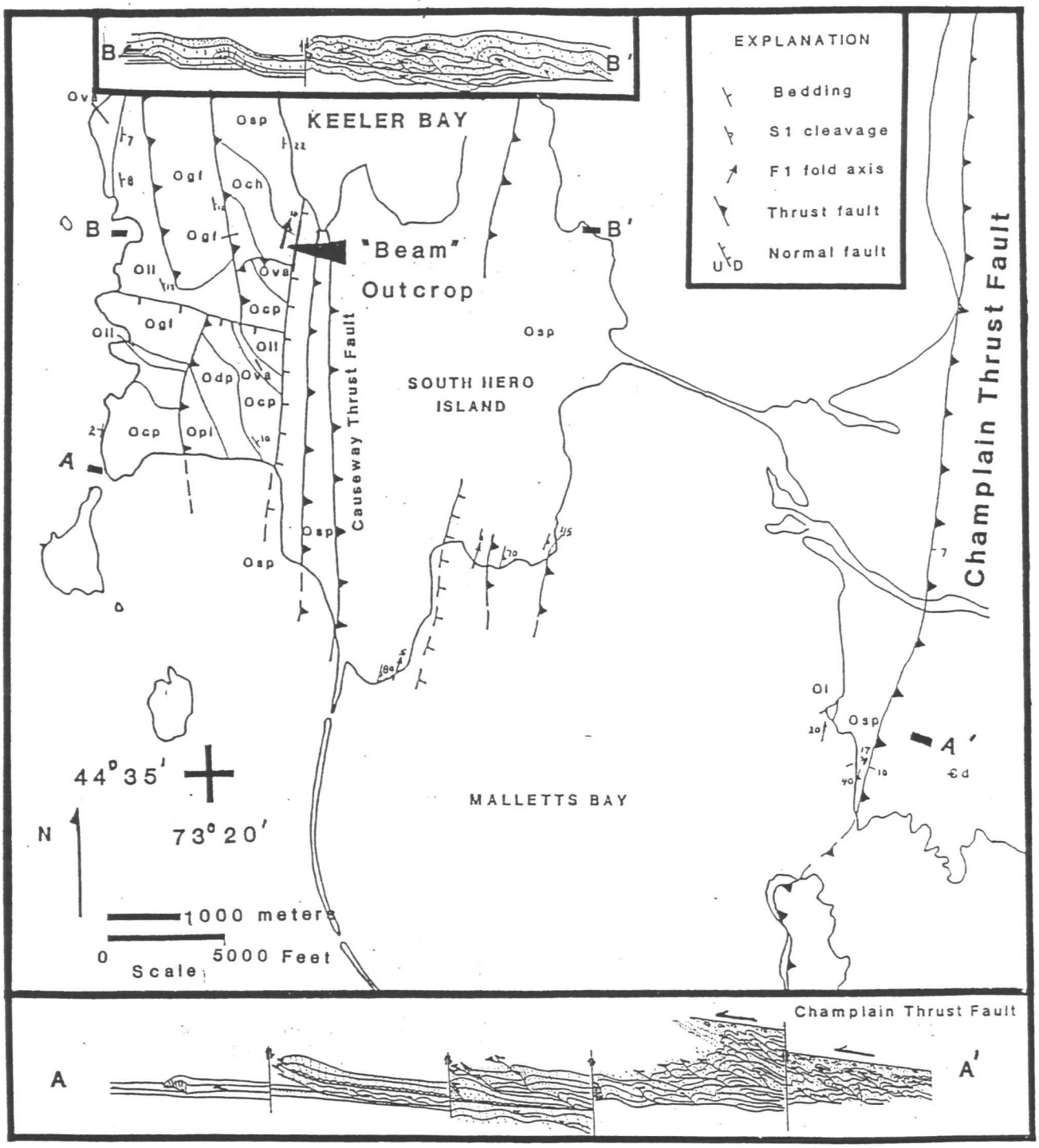
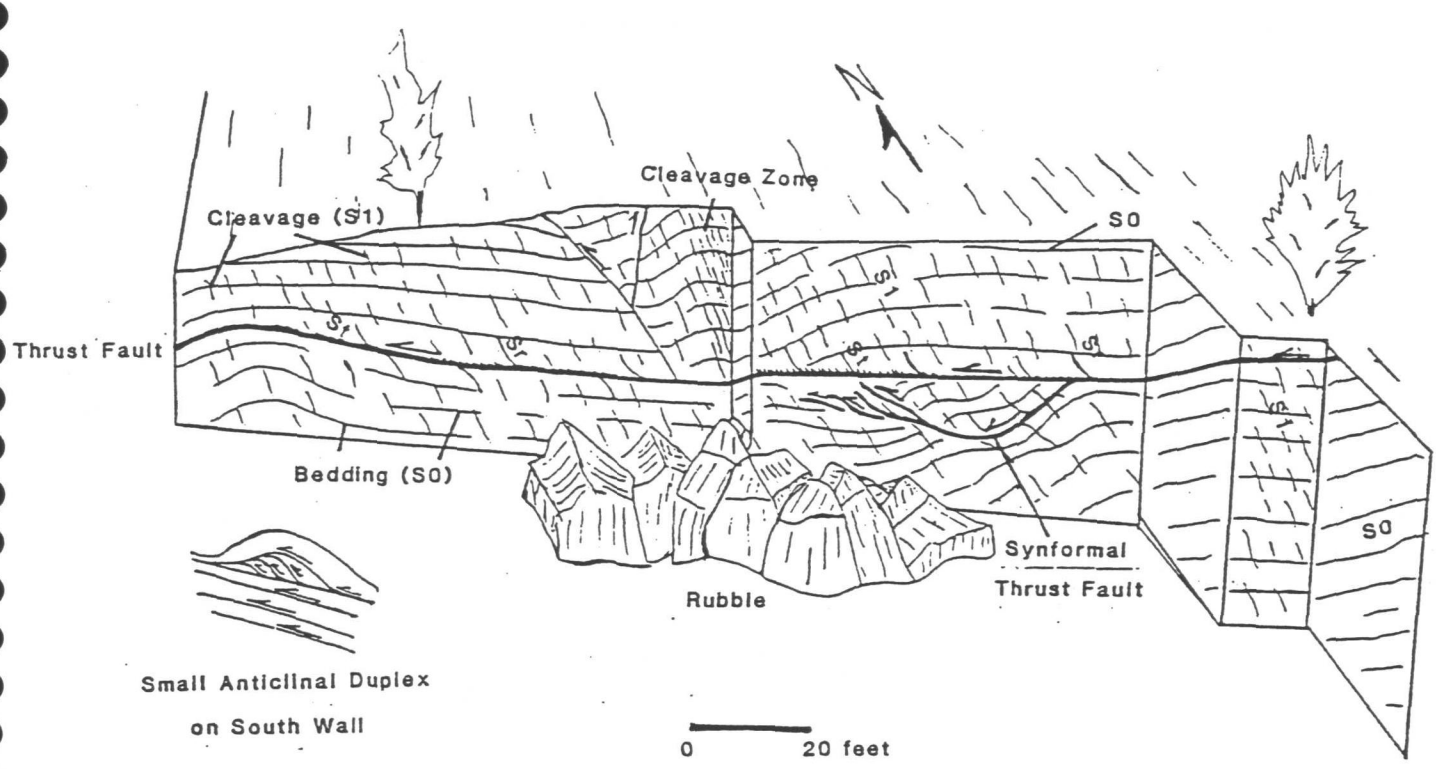


FIGURE 4 Geologic map and cross sections of the southern part of South Hero Island and eastern shore of Lake Champlain in the vicinity of Malletts Bay, Vermont. Geology by K. Leonard, 1985. Stops 1 and 2 ("The Beam") are along cross section B-B'



Stanley (1987)

THRUST FAULTS AND RELATED STRUCTURES

AT

LESSOR'S QUARRY, SOUTH HERO ISLAND, VERMONT

FIGURE 5

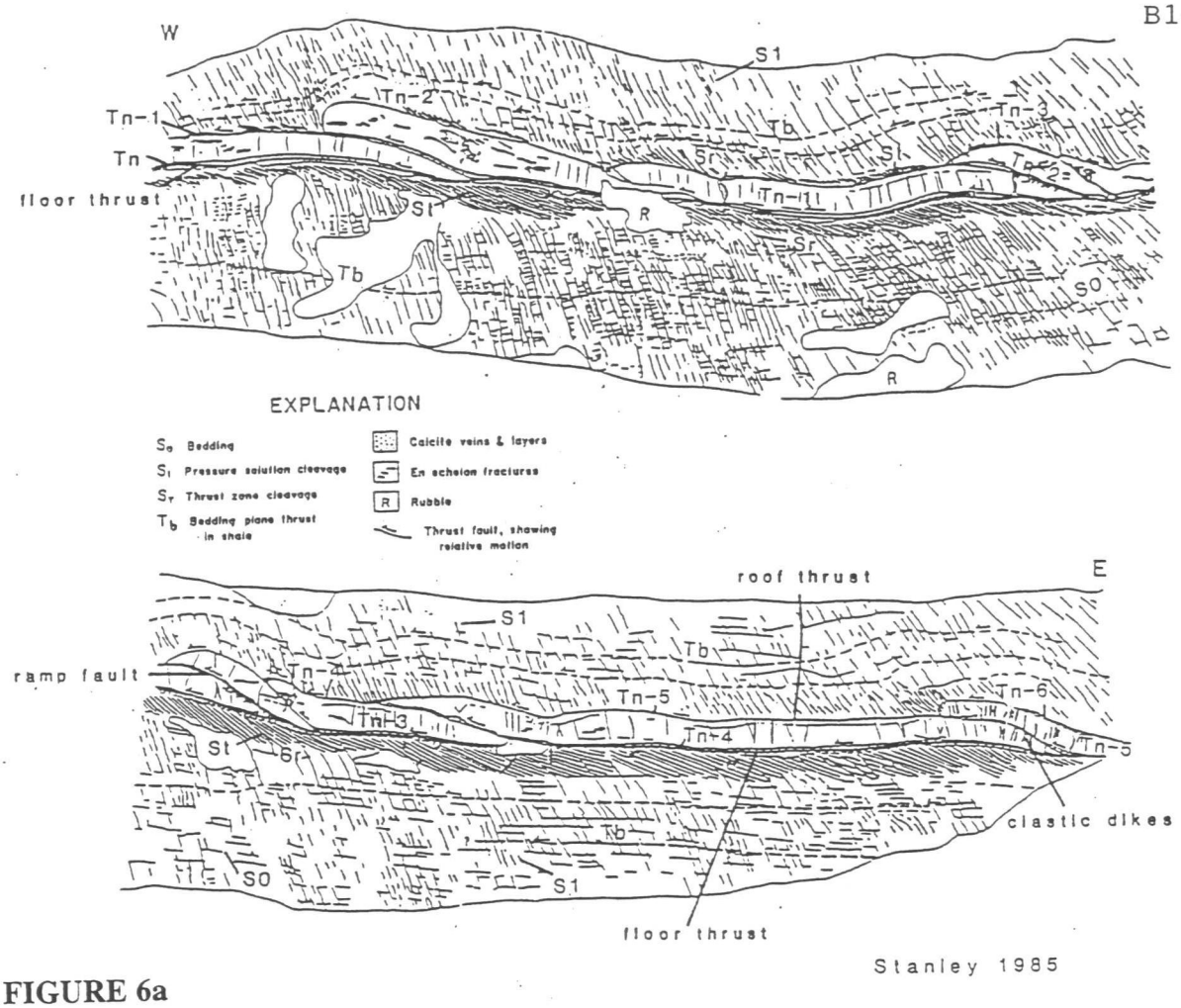


FIGURE 6a

DEFORMATION OF THE CUMBERLAND HEAD FORMATION

SOUTH HERO, VERMONT

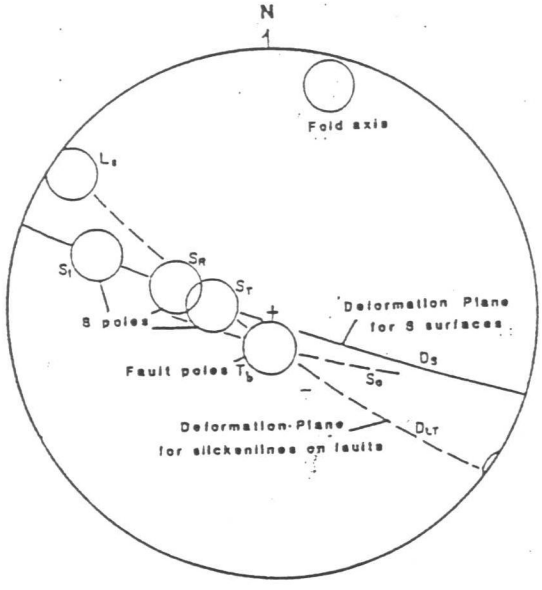


FIGURE 6b Lower hemisphere equal area projection showing the dominant position of the major structural elements in the outcrop of the "beam". The structural elements are identified in figure 7a. **Ls** refers to the dominant orientation of slickenlines on the thrust faults. The slickenlines on the older thrust faults are rotated along the deformation plane for slickenlines

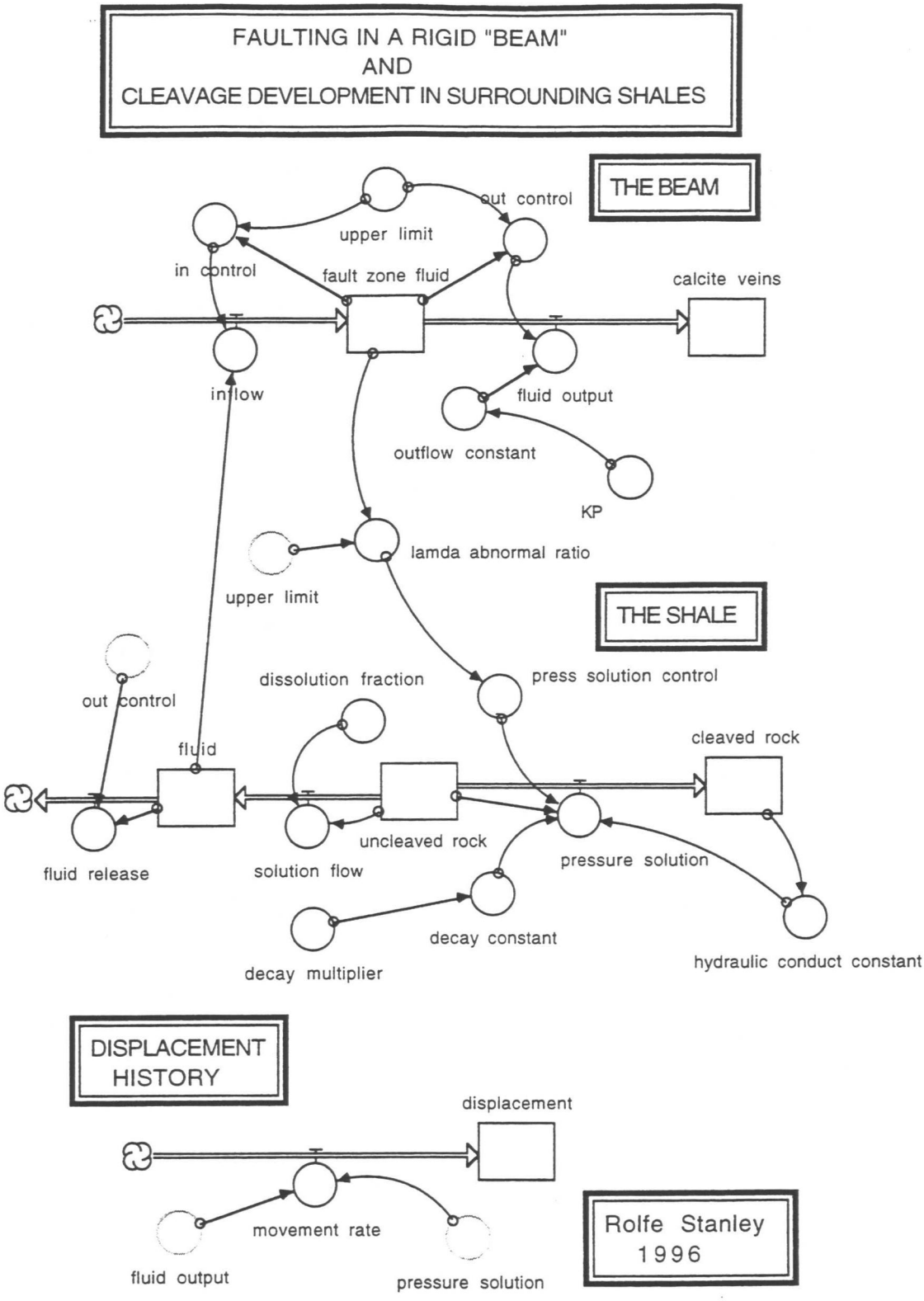


FIGURE 7 STELLA simulation model for faulting and cleavage development in the "Beam". Rectangles are **stocks**, arrows with attached circles are **flows**, isolated circles are **converters**, and arrows connecting the above elements are **connectors.** The function of the icons are explained in the text. The system is divided into 3 parts, the upper subsystem represents fluid-controlled faulting the "beam". The middle subsystem represents the development of cleavage by pressure solution in the shale. The displacement history of movement is recorded in the lowest subsystem. Labels identify the geological function of the components of the system. Equations are available upon request.

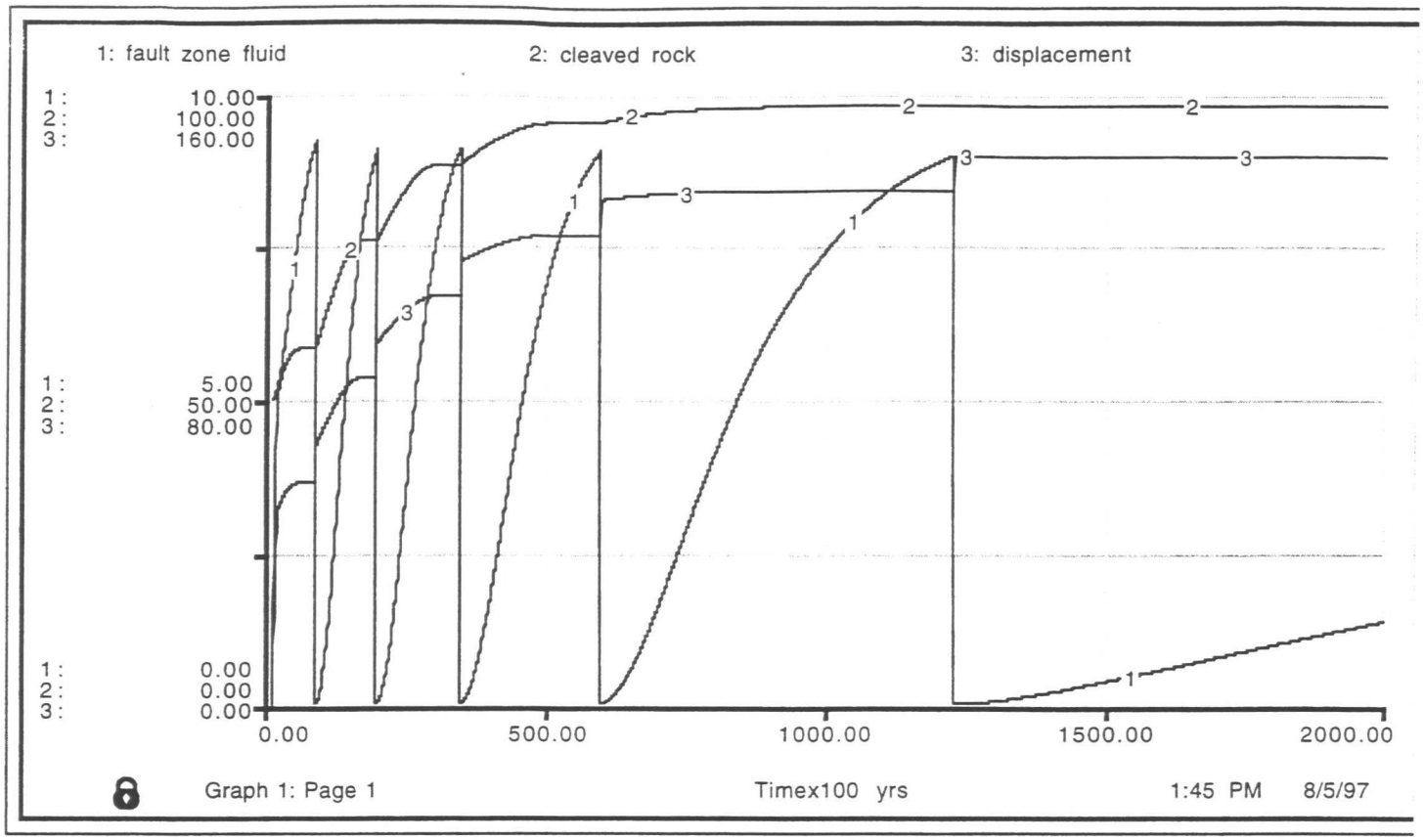


FIGURE 8 Graphical output of the STELLA simulation showing values for the percentage of fluid in the fault zone (**fault zone fluid, graph 1**) where 5 represents a standard hydrostatic condition or a Hubbert-Rubey ratio of 0.5. **Graph 2** represent the percentage of cleaved rock in the outcrop where the stock for uncleaved rock was originally 100%. **Graph 3** represents the displacement of total shortening of the outcrop. The simulation is run for 2000 time step with each time step representing 100 years. This time step is dictated by 1) the value of hydraulic conductivity in the shales and 2) computational time and memory.

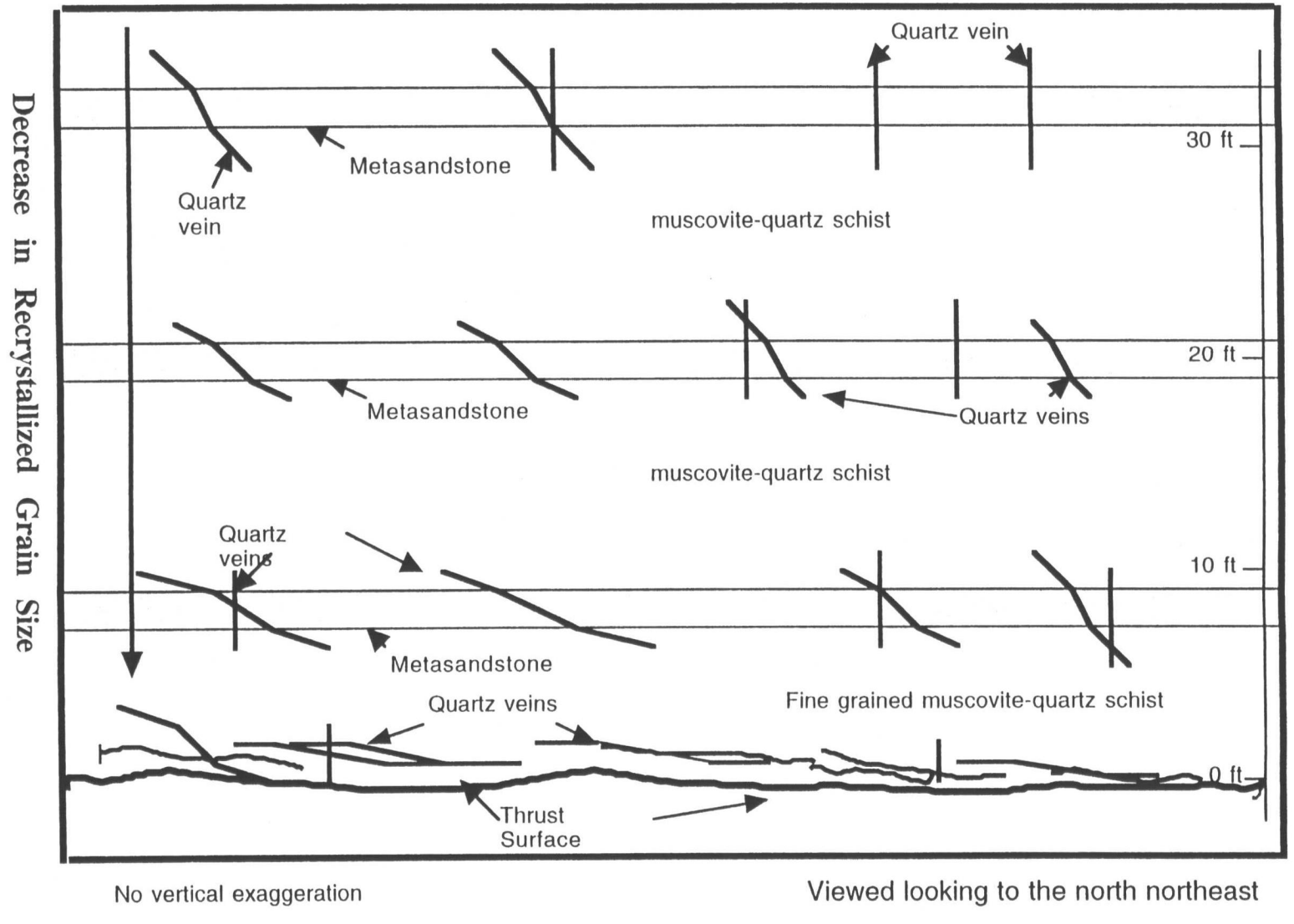
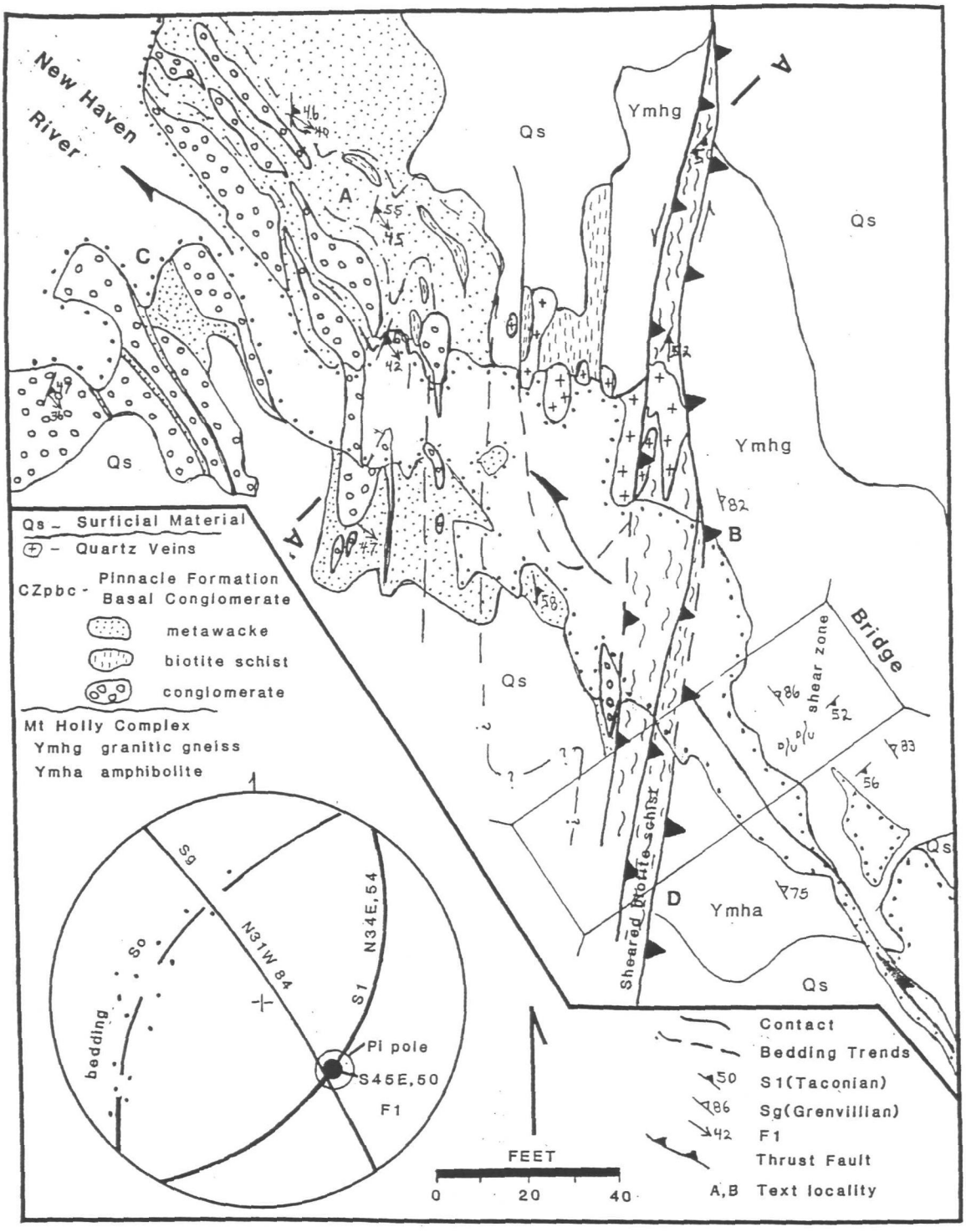


FIGURE 9

SCHEMATIC CROSS SECTION OF THE HINESBURG THRUST FAULT AT MECHANICSVILLE, VERMONT



BASAL LATE PROTEROZOIC

UNCONFORMITY AT LINCOLN, VERMONT

FIGURE 10

Tauvers, 1982 Stanley and DelloRusso 1985

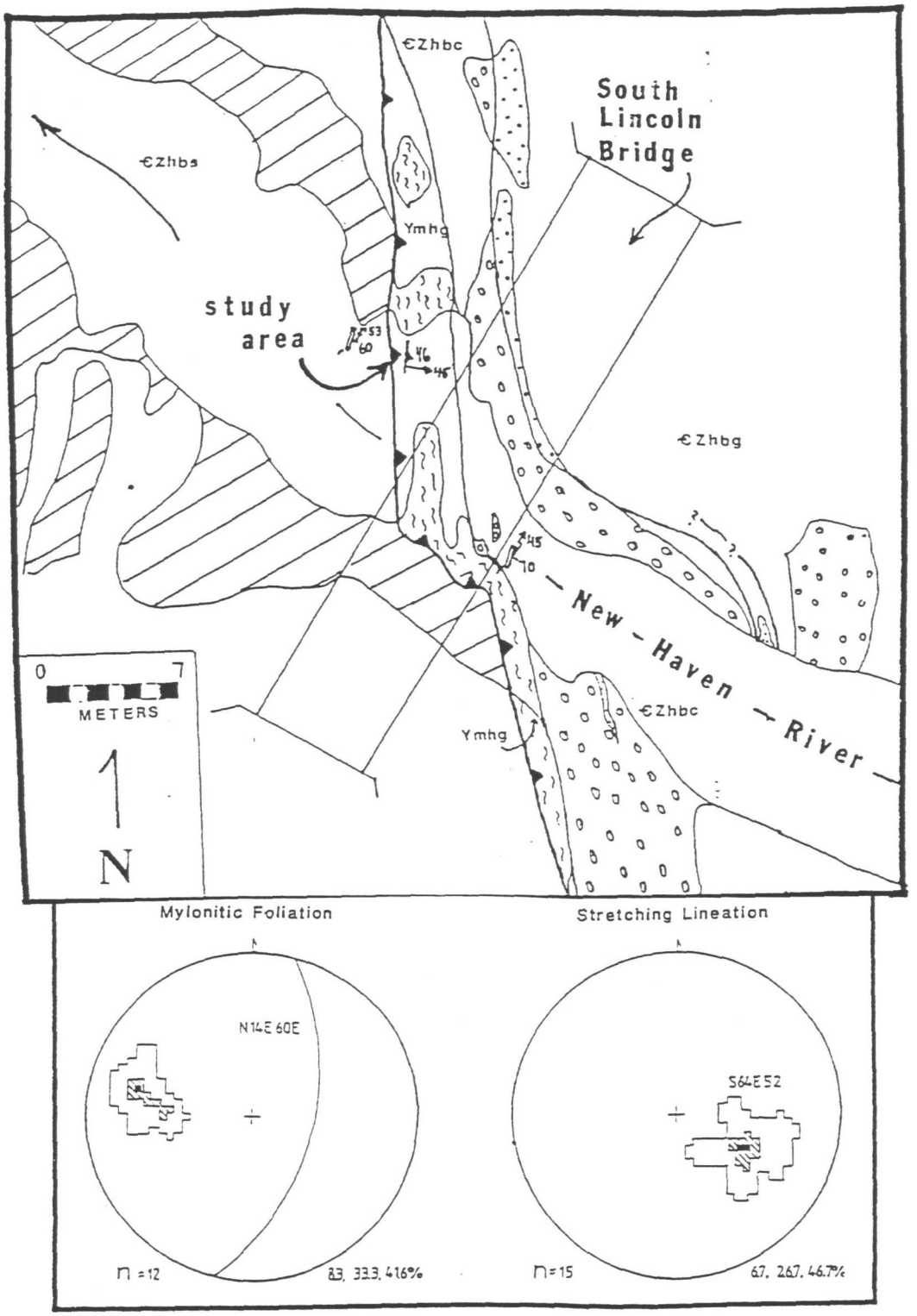


FIGURE 11 Geology at the bridge at South Lincoln, Vermont. Here the mylonitic granitic gneiss of the Mount Holly Complex (Ymhg) is in thrust contact with the Hoosac Formation (CZhbs). A pervasive, platey mylonitic foliation (Sm) and a penetrative mineral lineation shown in equal area projection dominate the granitic gneiss. A matrix-supported, cobble to boulder, polymictic conglomerate with thin metawacke beds (CZhbc) overlies the granitic gneiss. The thrust contact and all adjacent units are folded by F n+1 folds with counterclockwise sense. Hinges of isoclinal Fn folds are nearly parallel to the mineral lineation in the gneiss. Geology by Tauvers (1982) and subsequently modified by Stanley. Tick marks indicate the north direction. Percentages are points per 1% area.

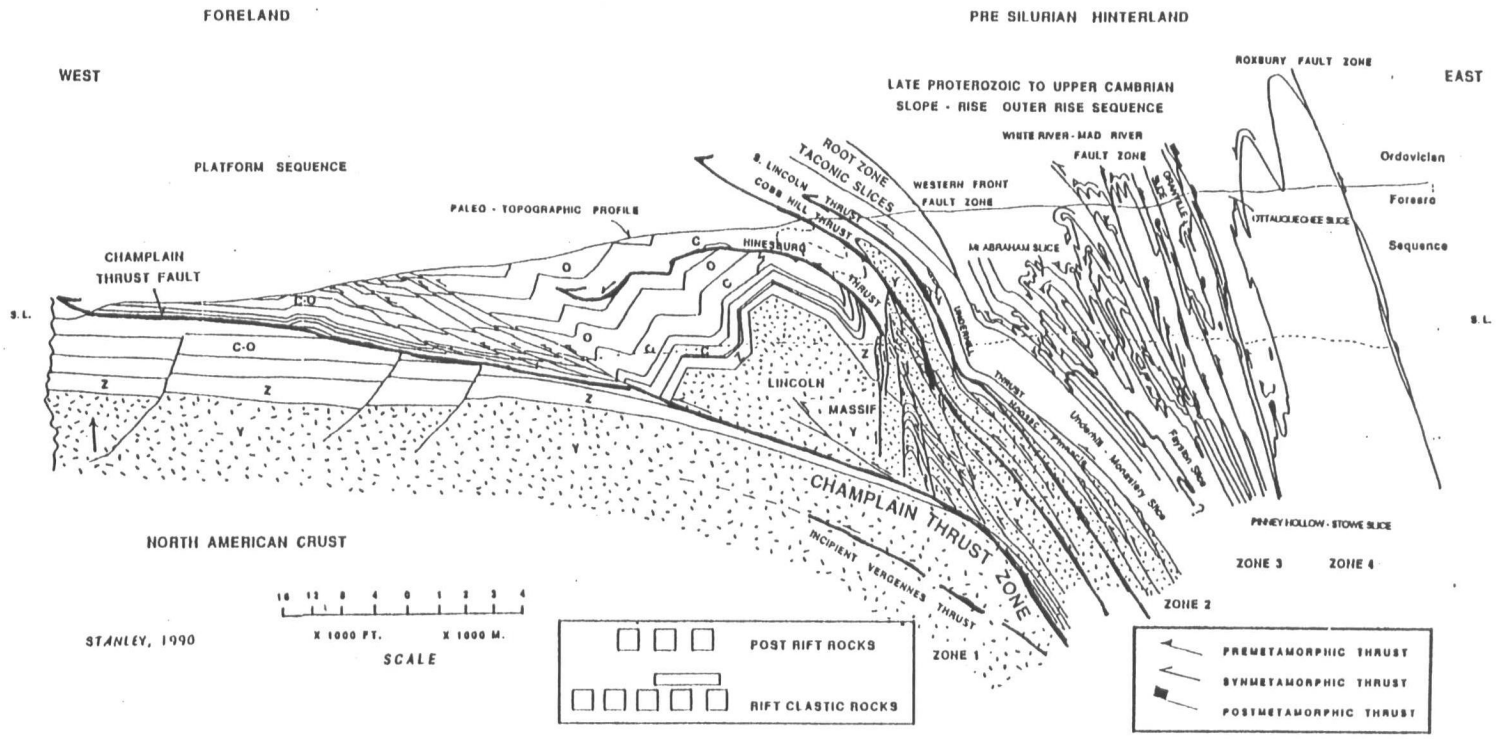


FIGURE 12 Enlarged cross section for the western part of the central Vermont transect showing the main aspects of the Taconian Accretionary Wedge. Here the Vergennes thrust fault of figure 2 is retrodeformed by 1.6 km to its eastern location beneath the Champlain thrust fault. The Lincoln Massif on the upper plate of the Champlain thrust is deformed by a fault propagation fold associated with the East Middlebury thrust fault. The eastern part of the massif is cut by numerous ductile faults (Cobb Hill thrust fault, CHT) and shear zones that were flattened as the eastern part of the section (South Lincoln thrust fault, for example) was rotated clockwise into a steeper position compared to figure 11. At this time the Hinesburg thrust fault, which nucleates from between the two anticlines of the Lincoln Massif, was locked as structures beneath it formed in the foreland. High angle faults below the Champlain thrust cut the base of the Middle Ordovician sequence and may represented structures formed on the peripheral bulge as the lower plate was bent toward the east-dipping subduction zone. Remaining symbols are similar to those in Figure 2. Compare this figure with figure 2.

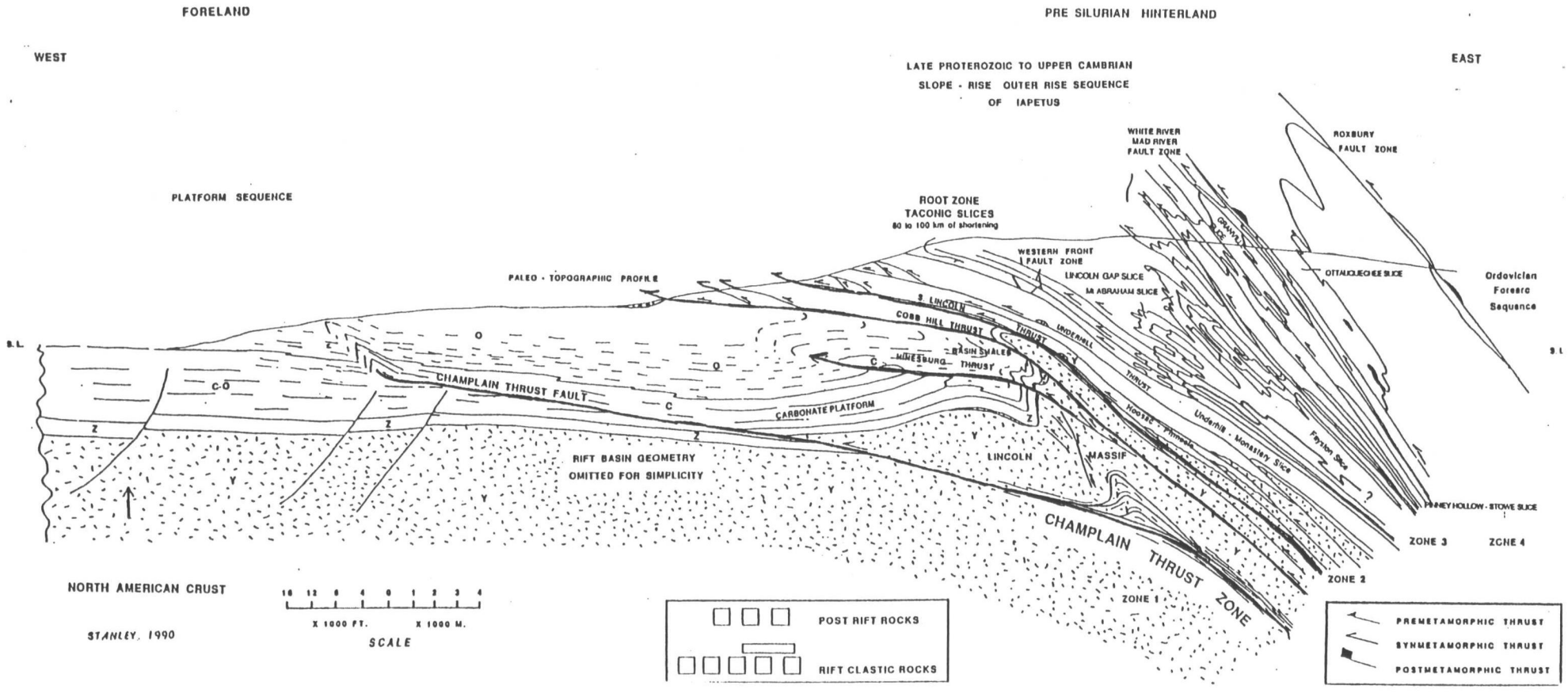


FIGURE 13. This section shows unfolding of the Lincoln Massif as the East Middlebury thrust fault is retrodeformed. These events are associated with the gradual retrodeformation of the Champlain thrust zone whose western tip is shown here as a fault propagation structure. The Cobb Hill and Hinesburg thrust faults were active at this time. Synmetamorphic faults were probably also active in the hinterland as the mountain system adjusted to westward displacement. Remaining symbols are explained in the previous figure.

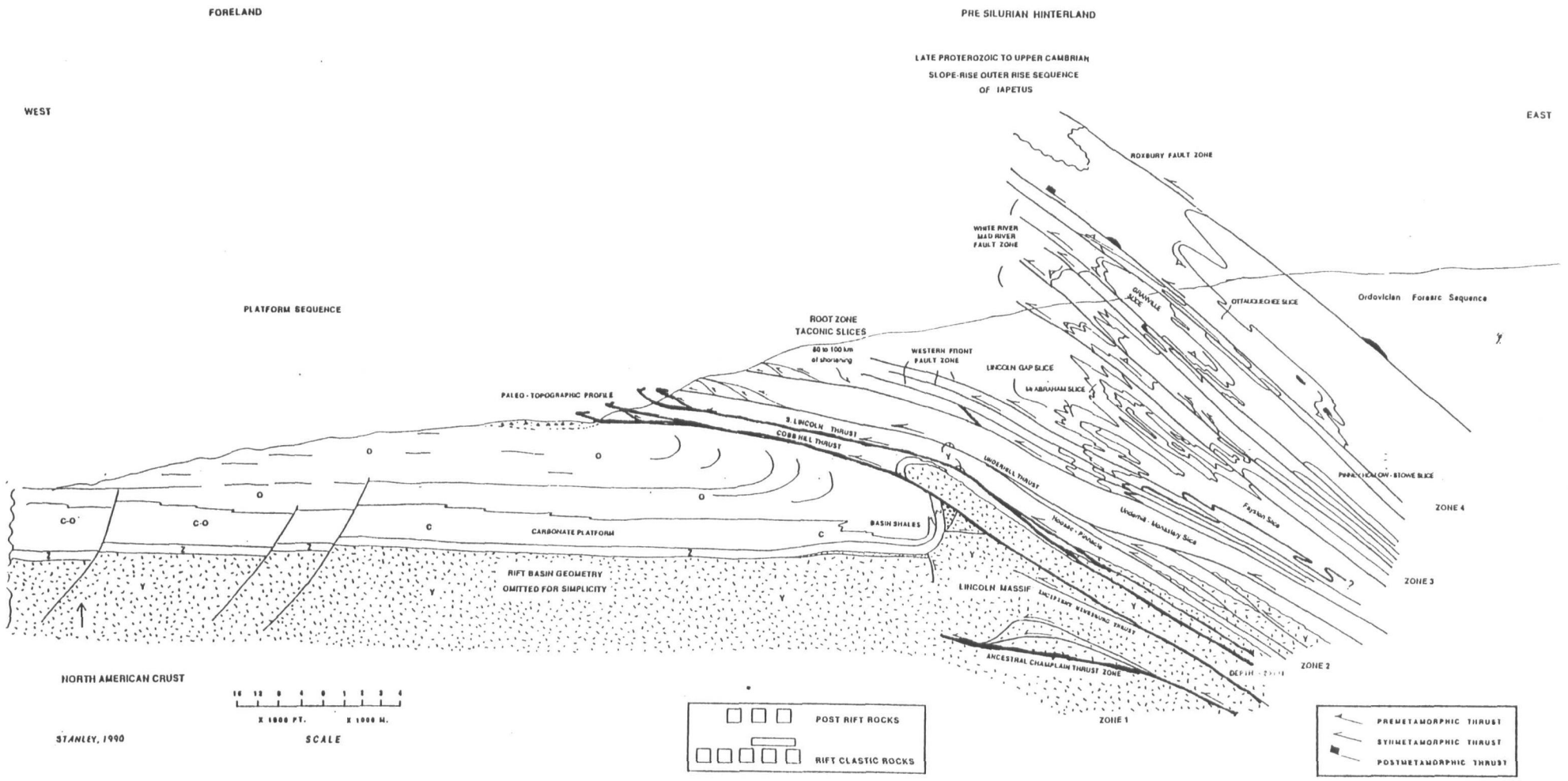


FIGURE 14. In this section the tip of the Champlain thrust has been retrodeformed to the eastern part of the Lincoln Massif. During this time the Cobb Hill thrust (CHT) and associated faults to the east (the South Lincoln-Underhill thrust faults), clipped off slices of Middle Proterozoic crust as the accreted rift-drift section to the east was driven across the continental margin. Synmetamorphic thrust faults to the east were active during this time with kyanite-chloritoid grade metamorphism occurring lower in the accreted pile.

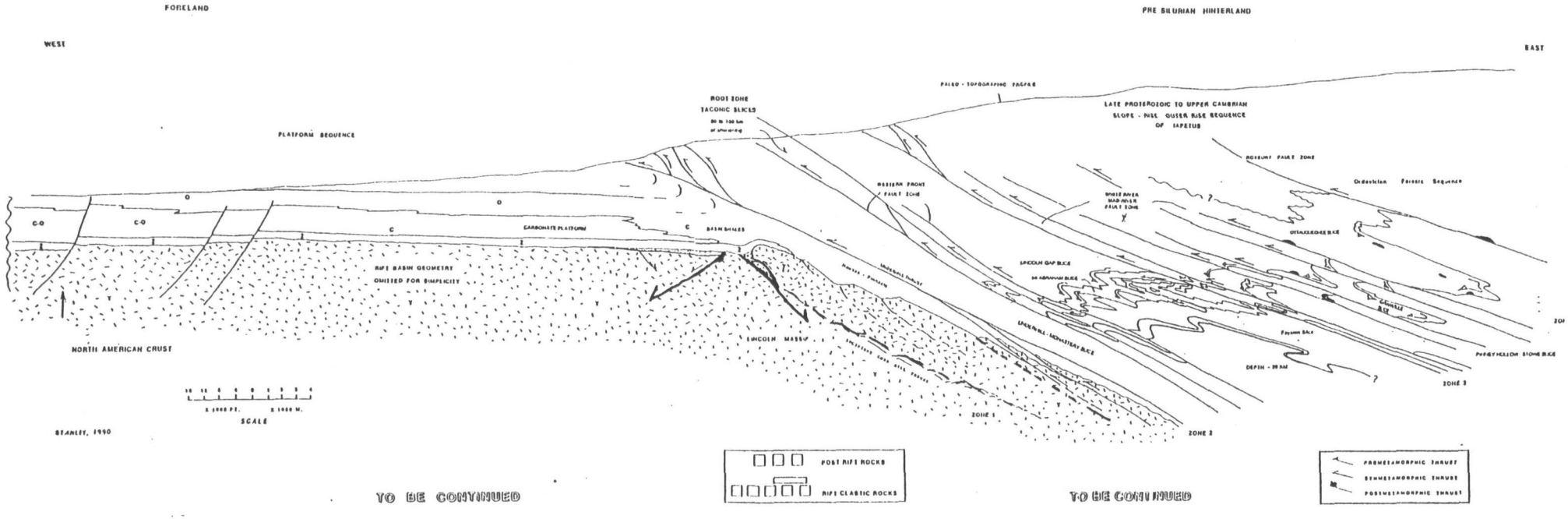


FIGURE 15. In this section the Cobb Hill thrust fault and thrusts along the eastern margin of figures 1 and 2 have been retrodeformed to an earlier stage. Here the eastern margin of the massif is shown as a series of anastomosing shear zones that may have developed as normal faults during early rifting (Warren, 1990). These faults were reactivated as thrust faults during compression. Synmetamorphic thrust faults were active to the east during this time with kyanite-chloritoid grade metamorphism occurring lower in the accreted pile. These faults cut the older pre/early metamorphic faults shown by curved arrows. The medium high-pressure metamorphism occurred either before or during this early event since barrositic hornblende is found in different slices.

PLUTONIC AND HYPABYSSAL INTRUSIONS OF THE EARLY CRETACEOUS CUTTINGSVILLE COMPLEX, VERMONT

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INTRODUCTION

The Monteregian Hills (MH) and White Mountain (WM) igneous provinces represent a major period of continental anorogenic igneous activity that occurred between 240 and 100 Ma (Foland and Faul, 1977). The continental magmatism can be temporally divided into two major episodes, Jurassic (170-200 Ma) and Cretaceous (*ca.* 125 Ma). The older period of igneous activity is confined to the WM province (Fig. 1), is largely felsic in character, and, with the notable exceptions of Red Hill and Rattlesnake Mountain, the rocks are silica-saturated to silica-oversaturated. The largest intrusive complex is the White Mountain batholith which consists of a number of overlapping caldera complexes and associated granitic plutons (Eby et al., 1992). The younger period of igneous activity encompasses both the MH and WM provinces and is the focus of this field trip.

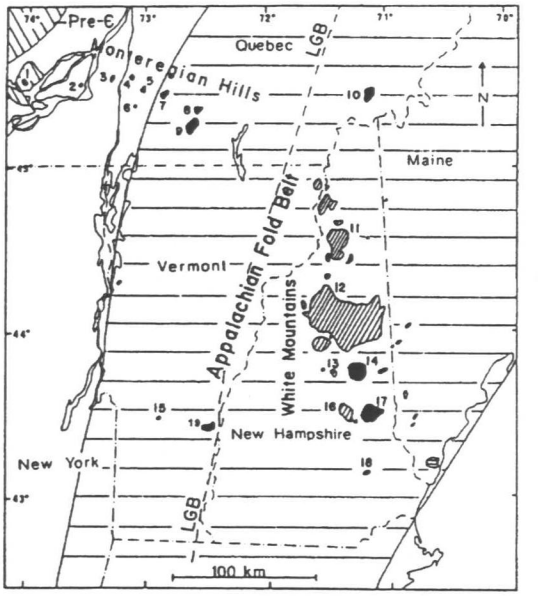


Fig. 1. Geological setting and locations of the major Monteregian Hills and White Mountain plutons. Horizontally shaded areas, plutons emplaced between 240 and 200 Ma. Diagonally shaded areas, plutons emplaced between 200 and 165 Ma. Filled areas, plutons emplaced between 140 and 100 Ma. Plutons indicated by number: 1, Oka; 2, Mont Royal; 3, Mont St. Bruno; 4, Mont St. Hilaire; 5, Mont Rougemont; 6, Mont Johnson; 7, Mont Yamaska; 8, Mont Shefford; 9, Mont Brome; 10, Mont Megantic; 11, Pliny Range; 12, White Mountain batholith; 13, Red Hill; 14, Ossipee; 15, Cuttingsville; 16, Belknap Mountain; 17, Merrymeeting Lake; 18; Mount Pawtuckaway; 19, Ascutney. LGB, limit of Grenville basement. After Eby (1987).

In several stages from 130 to 100 million years ago, groups of plutons, dike swarms, and individual intrusive complexes combined to form an igneous province that stretches 400 km from the central Adirondack Highlands of New York eastward through southern Maine, and 350 km from the Monteregian Hills (MH) of southern Quebec southeastward through New Hampshire (Fig. 1). Early Cretaceous alkalic plutons in New England are called the younger White Mountain (WM) igneous province by Eby (1987), and all of the Early Cretaceous intrusions have been labeled the New England-Quebec (NEQ) igneous province by McHone and Butler (1984). The MH and WM igneous provinces appear to be products of anorogenic, intraplate rifting, perhaps forming in concert with stages of North Atlantic oceanic rifting events (McHone and Butler, 1984), although tholeiitic Mesozoic magmas associated with Triassic-Jurassic basins around the entire North Atlantic ocean are much more widespread (Manspeizer, 1988).

Similar but somewhat younger intrusions (and volcanic products) are found in the continental shelf (Hurtubise and others, 1987), and other intraplate alkaline intrusions/volcanoes extend into the western North Atlantic as the New England seamount chain (Duncan, 1984). Morgan (1983), Duncan (1984) and

others have linked the Cretaceous MH and WM intrusions to the New England seamounts as expressions of an age-progressive *hotspot track*, analogous with the Hawaiian model. In the New England model, the northwestern MH plutons are at the older (*ca.* 124 Ma) end, progressing through 122 Ma WM plutons in New Hampshire, and through the seamounts from about 110 to 70 Ma far out into the North Atlantic Ocean.

The hotspot track model has immediate appeal but also immediate problems, some of which have been discussed by McHone (1996). The Cuttingsville complex is petrologically similar to the MH plutons, but it is located at least 120 km southwest of the MH-WM section of the *hotspot track*. More importantly, the Cuttingsville magmas formed about the same time as New England seamounts far to the southeast, which is at least 20 m.y. too young for the New England portion of the track.

The younger WM and MH plutons are intruded into several terranes: the Precambrian Grenville province of the Canadian shield, the flat lying Cambro-Ordovican sediments of the St. Lawrence Lowland, and the deformed Lower Paleozoic section of the Appalachian orogene. A notable feature is that silica-undersaturated rocks are confined to plutons emplaced to the west of Logan's line (i.e. into thinner continental crust). This is well illustrated by the MH plutons which cross all three terranes. Plutons emplaced into the Grenville province and St. Lawrence Lowlands are characterized by the near absence of silica-saturated rocks and their largely mafic character. Eastward the plutons become larger, felsic rocks become a significant component, and silica-saturated to oversaturated lithologies are found. Only silica-saturated rocks are found At Mont Megantic (Fig. 1). Unlike the older WM, mafic rocks are important components of most of the younger WM plutons. Basalt outcrops at Ossipee, diorite at Merrymeeting Lake, Mount Pawtuckaway, and Ascutney, and essexite at Cuttingsville. Eby (1987) suggested that these variations could be explained by differences in the thickness of the continental crust and the depth of melting. In the thicker Appalachian section larger volumes of melt were produced in the mantle at shallower depths, and there was more opportunity for interaction between these melts and the crust. Eby (1985) noted that in terms of Sr and Pb mafic silica-undersaturated rocks in the MH and mafic silica-saturated rocks in the WM were isotopically indistinguishable, suggesting a similar source (isotopically depleted mantle) but different degrees of melting.

At Mont Shefford, Mont Brome, and Cuttingsville quartz-bearing and nepheline-bearing rocks exist in spatial and temporal proximity. These three plutons occur in the transition zone between the Pre-Cambrian Grenville and the folded Appalachians. Landoll and Foland (1996) have argued that the undersaturated and oversaturated rocks are derived from the same magma with production of the oversaturated rocks by AFC processes occurring at high level. On this field trip we will examine the Cuttingsville complex where both mafic and felsic silica-undersaturated and oversaturated rocks occur in the same pluton.

Alkalic dikes across the region display a bimodal range of mafic and felsic types in overlapping swarms, each group having somewhat distinctive ages and physical characteristics. Given that they cooled only a few kilometers below today's surface, the dikes are intermediate in crystal character between phaneritic plutons and aphanitic volcanic rocks. Good eyesight and a hand lens are required to distinguish the minerals and textures of the dikes, but with care and experience, most can be classified in the field. Unlike the great quartz tholeiite intrusions of southern and eastern New England, most of the dikes are too small to have produced flood basalts or large volcanic edifices. Yet, as shown by xenoliths of spinel peridotite and other mantle rocks (McHone and Williams, 1985), these magmas, in dikes only a few meters wide, ascended from mantle depths of 100 km or more.

This field excursion presents ideas and data that have been slowly gathered for nearly twenty years. As in other regions, relatively few of the geologists who produced the quadrangle bedrock maps for the area paid much attention to the intrusions. Some older papers contain very useful information, such as those by Marsters (1889), Eggleston (1918) and Brace (1953). More recent work on the pluton at Cuttingsville includes papers by Laurent and Pierson (1973) and Robinson (1990), and studies by Wood (1984) and Eby (1992).

Dike locations in Vermont were studied by McHone during his thesis work at the University of Vermont in the early 1970's. Other field visits were made with J. Robert Butler during and after McHone's Ph.D. work at the University of North Carolina at Chapel Hill (1974-1978), and some field work was conducted with Chiasma Consultants, Inc. for the National Uranium Resource Evaluation (1978-1980). Eby investigated the Cuttingsville plutonic complex in the mid-1980s (NSF EAR-8600058). We made a field tour in June of 1997 in preparation for this field trip.

The western margin of the Cretaceous New England - Quebec igneous province is formed with three lobes, or subprovinces, that extend westward from northern New England (Fig. 2). On the northern side, the Monteregian Hills subprovince of southern Quebec is known for its carbonatites and ultramafic stocks, as well as for alkali lamprophyre dikes. Older K-Ar radiometric dates in Quebec are mainly between 110 and 130 Ma (Eby, 1984), although subsequent work by Foland *et al.* (1986) suggests that many of the Monteregian Hills intrusions are close to 124 Ma in age.

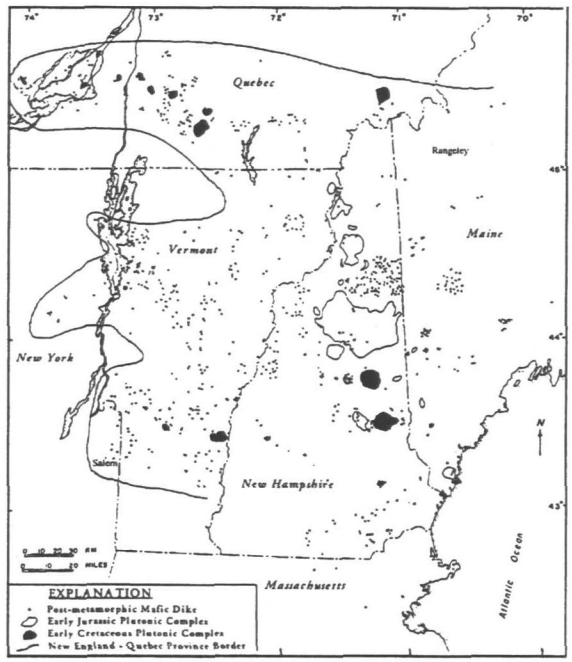


Fig. 2. Distribution of post-metamorphic dikes in northern New England.

Igneous rocks are unknown in the northernmost Lake Champlain Valley, but north and south of Burlington, Vermont there are several hundred lamprophyre and trachyte dikes exposed along shorelines, roadcuts, streams, and hillsides. Lamprophyre dikes of this subprovince are distributed westward into the central Adirondack Highlands of New York, and eastward into north-central Vermont (McHone and Corneille, 1980). Champlain dikes are identical to Monteregian dikes, including carbonate-rich types, but associated plutonic complexes are fewer and smaller in the Lake Champlain region than in Quebec. Radiometric dates indicate ages near 135 Ma for monchiquites, 125 Ma for trachytes and syenites, and 115 Ma for camptonites (McHone, 1987), but this very neat division needs better confirmation. Because the Champlain Valley intrusions are close analogs to the Monteregian intrusions, their ages might actually all be nearer to 122 to 124 Ma.

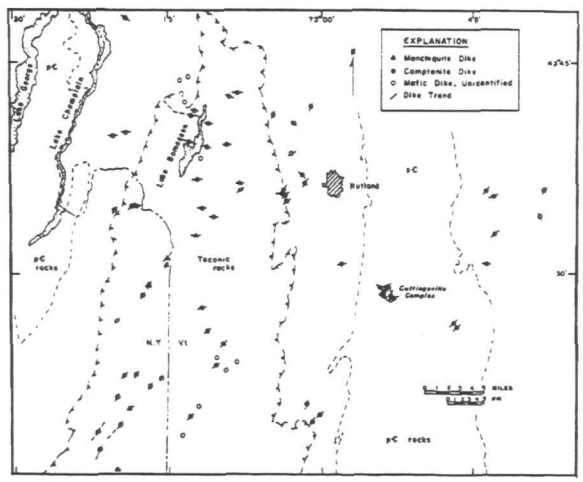


Fig. 3. Dike locations in the northern Taconics region, Vermont and New York.

In the southern, upper Champlain Valley (the lake flows northward), there is a "virtual" gap in igneous rocks, with only a few stray dikes known at Vergennes, Middlebury, Westport (New York), and Orwell. The third lobe (herein labeled "NT", for Northern Taconics) has around 70 known dike localities, some of which are probably exposures of the same dike, distributed along the northern Taconic region between Proctor and Dorset, westward a few kilometers into eastern New York, and

eastward across the Vermont Valley into the Green Mountains southeast of Rutland (Fig. 3). Many of the intrusions have petrologic characteristics that are distinct from the northern NEQ dikes, but there are also some very similar examples. Except for a few trachytes near Rutland, all of the dikes so far studied are lamprophyres. Dates are mostly 100-110 Ma (Table 1).

Little detailed work has been done on the dike rocks, but the Cuttingsville plutonic complex southwest of Rutland has received attention since the 1970's from mineral companies as well as by research geologists. We consider the Cuttingsville intrusion to be connected with the Taconic subprovince on the basis of age and petrology, although the distribution of dikes does show fewer examples towards Cuttingsville (Fig. 3). Eastward from Cuttingsville, dikes remain fairly common (10 to 20 examples have been mapped in some of the 7.5' quadrangles), and they merge into the regional camptonite swarms of eastern Vermont, New Hampshire, and Maine (McHone, 1984).

INTRUSIVE TRENDS AND STRUCTURES

Dikes around Cuttingsville, Vermont have a NE-SW preference, unlike dike groups farther north (Fig. 4). In southern Quebec, the Monteregian Hills dike group has a WNW-ESE maxima, examples for which are found the entire distance from Montreal to northwestern Maine (McHone, 1978a). Dikes of the Lake Champlain Valley region have a very distinct E-W preference (McHone and Corneille, 1980). The NE-SW dike trend is common not only across southern Vermont, but also prevails across northern New England east of the Green Mountain anticlinorium (Fig. 4).

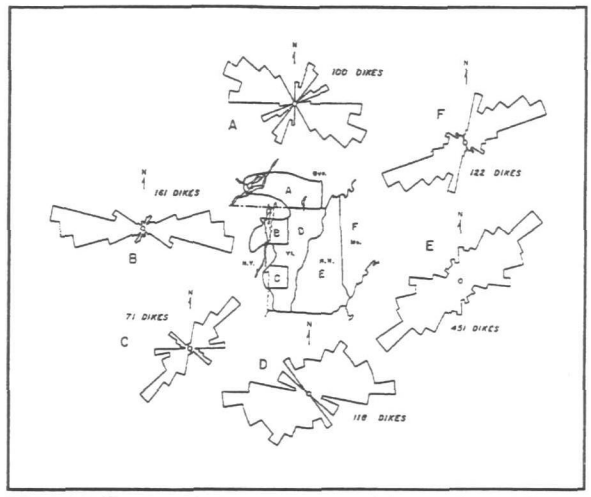


Fig. 4. Dike trends (rose diagrams) in northern New England. Rose diagram labels refer to areas shown in the central figure, with rose C representing dikes around Cuttingsville and the northern Taconic Mountains.

ALKALIC DIKE TYPES

Regional dike types around Cuttingsville include monchiquite (nephelinite), camptonite (basanite), bostonite (trachyte), and spessartite (andesite). All of these igneous types are presumed to be related through mantle/crustal processes of fractional melting, differentiation, and crystallization, but it may be unlikely that they were at one time all co-magmatic.

Monchiquite is a very mafic, granular, analcite-bearing, olivine-bearing, augite-rich alkali basalt, often with appreciable calcite (in spheroidal bodies), phlogopitic mica, and kaersutitic amphibole. Feldspar (Ca-plagioclase) is poorly developed or lacking. Monchiquite is commonly dark gray in color.

Camptonite can look much like monchiquite, except that olivine is rare or absent, kaersutite is common to abundant, and plagioclase is more abundant than analcite. Phenocrysts are only mafic (augite and/or kaersutite), rather than felsic, although feldspar in fact does occur as phenocrysts in some dikes that otherwise are "good" camptonites. Camptonite dikes usually have a brownish to rusty gray range of colors.

Spessartite dikes lack olivine and analcite, but plagioclase (intermediate Ca) is well developed and present as phenocrysts as well as intergrown with augite in the groundmass. Phenocrysts (or megacrysts) of kaersutite serve to distinguish spessartite from tholeiitic dolerite (diabase) dikes that are common in other parts of New England. Spessartite often shows a distinctly greenish or purplish cast as well as gray colors.

The intrusive breccias north of Cuttingsville (stops 3 and 4 of this trip) have a spessartitic to "andesitic" matrix.

Bostonite is a name that in a strict sense applies only to felsic (anorthoclase-rich) dikes that have a "felty" clumped-grain texture, which is not always present. Trachyte, although used for volcanic rocks as well, is a better general term. Minor minerals include oxidized biotite, quartz, and clay products. Some examples show well-formed alkali feldspar and/or quartz phenocrysts. Trachyte dikes may be iron-stained, but they are generally light brown to cream-colored on fresh surfaces. The quartz syenite of the Cuttingsville complex is chemically similar to trachyte, but at stops 5 and 7, the syenite has been altered and enriched by sulfides.

TABLE 1. RADIOMETRIC DATES, NORTHERN TACONIC IGNEOUS ROCKS

Sample	Description	Date (Ma)	Reference
PO-1	Hbl spessartite, Route 4 roadcut Poultney quadrangle lat 43°32'05"N lon 73°10'34"W	113 ± 4	(1)
WR-3	Hbl spessartite, Route 4 roadcut West Rutland quadrangle lat 43°30'49"N lon 73°03'20"W	108 ± 4	(2)
RT-7	Andesitic breccia, Shrewsbury Rutland quadrangle lat 43°30'57"N lon 72°53'56"W	101 ± 2	(1)
Cuttingsville complex	Sodalite syenite	100 ± 2	(3)
	Biotite foyaite	103 ± 2	(4)
	Essexite	99 ± 2	(4)
	Essexite	103 ± 4	(5)
	Bio-hbl syenite	98 ± 11	(6)
	Bio-hbl syenite	100 ± 8	(6)
	Hornblende syenite	97 ± 8	(6)
	Essexite	100 ± 8	(6)

Note: K-Ar whole-rock or biotite ages, except for Cuttingsville (Ref. 5) which are titanite fission track ages. Ages have been corrected, where necessary, using the new IUGS decay constants (Steiger and Jäger, 1977). References: (1) McHone and McHone (1993); (2) Zen, 1972; (3) personal communication, H. Kreuger, Geochron; (4) Armstrong and Stump, 1971; (5) Stone & Webster, unpublished date referenced in Kanteng (1976); (6) Eby, unpublished titanite fission track ages.

CUTTINGSVILLE

General Geology

The Cuttingsville complex is an elliptical shaped, northwest trending intrusion, 2.3 km long and 1.7 km wide. The various plutonic and hypabyssal units were intruded into Precambrian schists and gneisses. The pluton was originally mapped by Eggleston (1918) and subsequently re-investigated by Kanteng (1972), Laurent and Pierson (1973) and Wood (1984). The geological map (Fig. 5) is modified from Eggleston (1918) on the basis of additional field work in the summer of 1986. Geophysical studies indicate that the intrusion is pipe-shaped with near-vertical sides which extend to a depth of at least 2 km (Griscom and Bromery, 1968). Radiometric dating (Table 1) indicates an emplacement age of *ca.* 100 Ma, significantly younger than the ages of the other younger White Mountain intrusions.

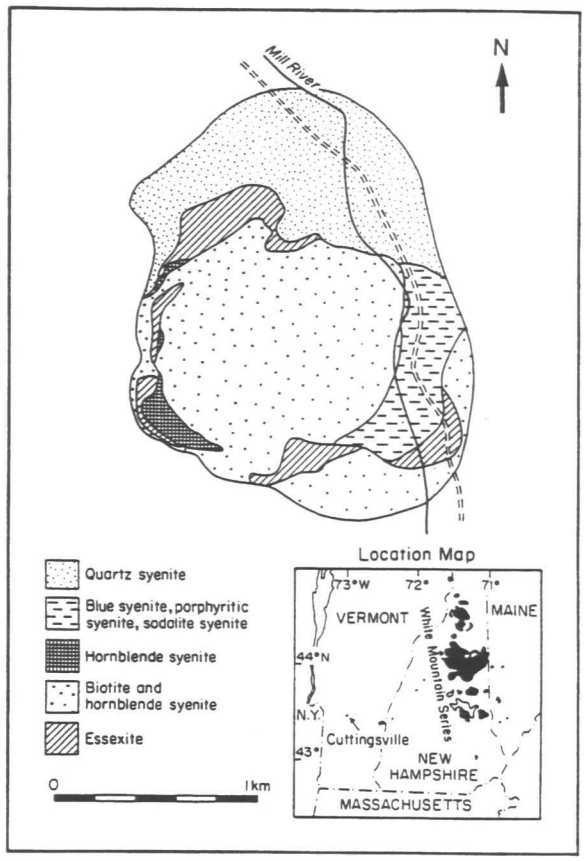


Fig. 5. Geological map of the Cuttingsville complex. Modified from Eggleston, 1918.

The intrusive contacts with the country rock are discordant and have near vertical dips. The contact zone is 5 to 10 meters wide and contains country rock xenoliths that are rounded due to abrasion (Pierson, 1970). The contacts between the intrusive phases are both sharp and gradational indicating that the various phases were emplaced in a relatively short time. The contacts between the late stage dikes and the rocks of the pluton are sharp, but the absence of chilled margins suggests that the dikes were emplaced before the pluton had cooled to near ambient temperatures.

Laurent and Pierson (1973) distinguished, from core to margin, four main intrusive phases. (1) - hastingsite foyaite and biotite foyaite (pulaskites), (2) - dioritie and porphyritic diorite, (3) - essexite, porphyritic essexite, plagifoyaite and sodalite foyaite, (4) - alkaline quartz syenite (larvikite) and syenite. The nomenclature of Laurent and Pierson (1973) has been modified (as shown on the geological map) to conform to that recommended by the IUGS Subcommission on the Systematics of Igneous Rocks, the term essexite has been retained for the nepheline-bearing gabbro (diorite). Except for very local contamination the diorites are all nepheline normative (and usually nepheline bearing) and hence mapped as essexites. In addition to the intrusive phases distinguished by Laurent and Pierson (1973) there is a porphyritic syenite, which varies from slightly nepheline normative to slightly quartz normative, that contains inclusions of all the other intrusive units. With the exception of late stage felsic and mafic dikes, the porphyritic syenite represents the last phase of intrusive activity.

Petrography

Essexite. In hand specimen the rock is light to dark gray, medium- to coarse-grained, and occasionally porphyritic. Mafic minerals often constitute more than 50% of the rock. The clinopyroxene occurs as pink titanaugite rimmed by light green augite and as separate light green augite grains. The amphibole is kaersutite, in some specimens amphibole is much more abundant than clinopyroxene. Plagioclase is usually strongly zoned. Accessory minerals are Fe-Ti oxides, titanite, and red-brown to straw-brown biotite. Anhderal nepheline and alkali feldspar occur as interstitial grains (Fig. 6).

Biotite-hornblende syenite. In hand specimen the rock is white to light gray and medium- to very-coarse-grained. The rock is dominantly composed of perthitic alkali feldspar, the feldspar occasionally has a plagioclase core. Plagioclase is found as a minor phase in some specimens. Mafic minerals are a light green augite and a red brown biotite. Light green amphibole is occasionally found. Accessory minerals are Fe-Ti oxides, apatite, and titanite.

Hornblende syenite. In hand specimen the rock is white to light gray and medium- to coarse-grained. The rock is dominantly composed of perthitic alkali feldspar with minor plagioclase, both showing a lath-like habit. The mafic minerals are an olive green to reddish amphibole and a straw brown to dark brown biotite. Titanite is an abundant accessory mineral, apatite and Fe-Ti oxides occur in minor amounts (Fig. 7).

Porphyritic syenite. In hand specimen the rock is light gray and porphyritic. Perthitic alkali feldspar, plagioclase, and occasionally a light green augite occur as phenocryst phases in a fine- to medium-grained groundmass of alkali feldspar, plagioclase, light green augite, brown mica and an Fe-Ti oxide. Accessory minerals are titanite, titanomagnetite, and apatite. Kaersutite and titanaugite occur locally as xenocrysts partly to totally altered to red brown biotite (Fig. 8).

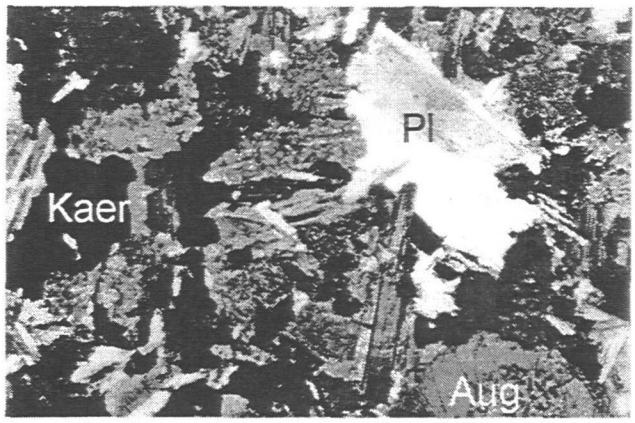


Fig. 6. Medium grained essexite (CV2). Width of field of view, 5.4 mm. Crossed nicols. Aug - augite, Kaer - kaersutite, Pl - plagioclase.

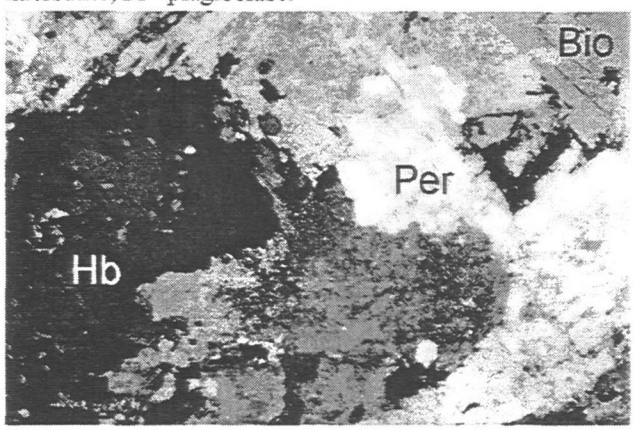


Fig. 7. Coarse-grained hornblende syenite (CV24). Width of filed of view, 5.4 mm. Crossed nicols. Bio - biotite, Hb - hornblende, Per - perthite.

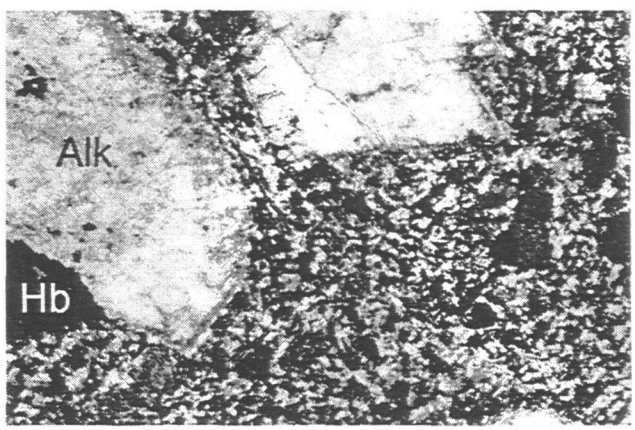


Fig. 8. Porphyritic syenite (CV8). Width of field of view, 5.4 mm. Crossed nicols. Alk - alkali feldspar, Hb - hornblende.

Sodalite syenite. In hand specimen the rock is medium gray to blue in color and medium- to coarse-grained. Blocky to lath-shaped perthitic alkali feldspar is the major mineral. Sodalite occurs interstially in variable amount. Randomly oriented clots of mafic silicates consist of green to greenish-brown aegirine augite, brown to straw brown biotite, and Fe-Ti oxides. Zircon and apatite occur in accessory amounts (Fig. 9).

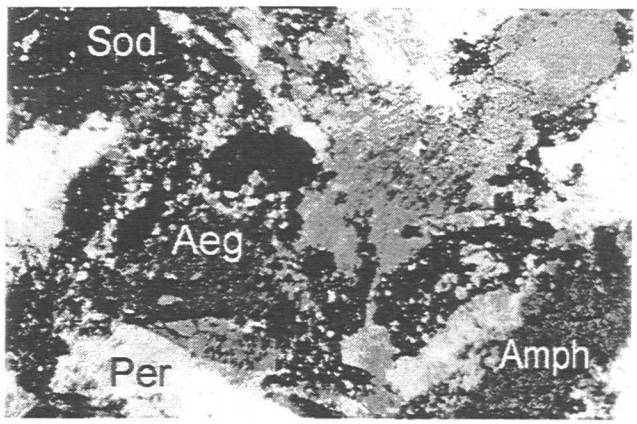


Fig. 9. Medium-grained sodalite syenite (CV23). Width of field of view, 5.4 mm. Crossed nicols. Aeg - aegirine, Amph - amphibole, Per - perthite, Sod - sodalite.

Quartz syenite. In hand specimen the rock is light gray, greenish on fresh surfaces, medium-grained and equigranular. Perthitic alkali feldspar and interstital quartz comprise 90% to 95% of the rock. Mafic minerals are green pleochroic aegirine, sparse reddish amphibole, and red-brown biotite. Grunerite has been found associated with biotite. An as yet unidentified mineral (provisionally named "honeyite") is found in several specimens. The mineral is light yellow in color, weakly pleochroic, and has an equant to slightly elongate habit. The elongate grains show inclined extinction. Accessory minerals are Fe-Ti oxides, apatite and zircon (Fig. 10).

Dike rocks. Both felsic and mafic dikes cut the Cuttingsville complex and the country rock. The *phonolites* are very fine-grained, dark green in hand specimen, and in thin section show a felted mass of alkali feldspar and pleochroic green aegirine. The mafic dikes are either fine-grained *diorite* (essexite) or *"camptonites"*. The so-called camptonites are *sensu stricto* not lamprophyres since they have, in addition to titanaugite and kaersutite, plagioclase phenocrysts. Fe-Ti oxides occasionally occur as a phenocryst mineral. Some of the mafic silicate phenocrysts have been extensively replaced by chlorite and most have reaction rims. The fine-grained matrix consists of plagioclase (and alkali feldspar?), red-brown pleochroic amphibole, Fe-Ti oxides, and accessory red-brown to brown biotite (Fig. 11).

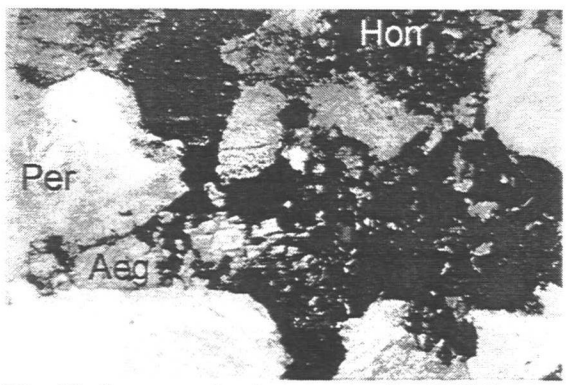


Fig. 10. Coarse-grained quartz syenite (CV38). Width of field of view, 5.4 mm. Crossed nicols. Hon - "honeyite", Per - perthite.

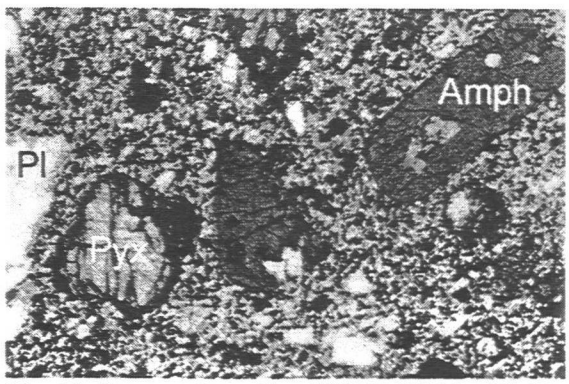


Fig. 11. *Lamprophyre* dike (CV32). Width of field of view, 5.4 mm. Crossed nicols. Amph - amphibole, Pl - plagioclase, Pyx - pyroxene.

Mineral Chemistry

Cuttingsville minerals have been analyzed by electron microprobe. The data are not yet published, but for some of the major minerals the chemistry is summarized in graphical form and briefly discussed below.

Feldspar. Plagioclase compositions vary from An_{62} to almost pure end member albite (Fig. 12). Some of the most Ca-rich plagioclases are found in the porphyritic syenite, but thin section observations suggest that these are plagioclase xenocrysts. Alkali feldspar compositions vary from Or_{90} to essentially pure end member albite. Feldspars from the mafic dikes show the highest equilibration temperatures, but the range in feldspar compositions indicates that there has been significant subsolidus unmixing. The two feldspar geothermometer of Fuhrman and Lindsley (1988) yields equilibrium temperatures of 680°C for a porphyritic syenite (CV15), 700°C for a quartz syenite (CV21) and 480°C for a biotite-hornblende syenite. While all the temperatures are subolidus, the relatively high equilibrium temperatures for the porphyritic syenite and quartz syenite indicate relatively rapid cooling, while the lower equilibrium temperature for the biotite-hornblende syenite indicates that this unit cooled relatively slowly.

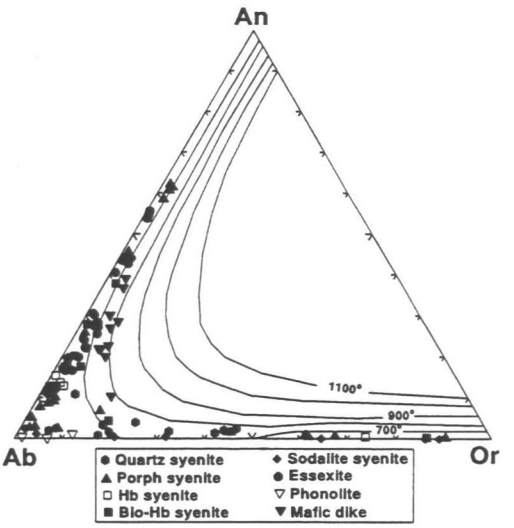


Fig. 12. Feldspar compositions projected into the Ab-An-Or ternary diagram. One bar feldspar solvi (°C) are from Nekvasil (1992).

Pyroxene. Chemically the pyroxenes are diopside and aegirine-augite (Fig. 13). Compared to many alkaline complexes, the Cuttingsville pyroxenes are not particularly Fe- or Na-rich, with only the pyroxenes from the sodalite syenite showing the typical degree of Na-enrichment (but not Fe). Hence the magmas from which these pyroxenes crystallized were not strongly alkaline. This projection does not differentiate between Ti-rich and Ti-poor pyroxenes, but optically the pyroxenes found in the essexite are pink pleochroic titanaugites and light green pleochroic augites. Pink titanaugites are also found in some of the syenites as xenocrysts.

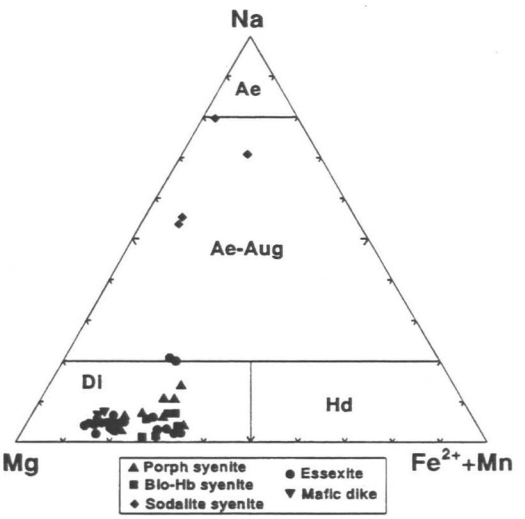
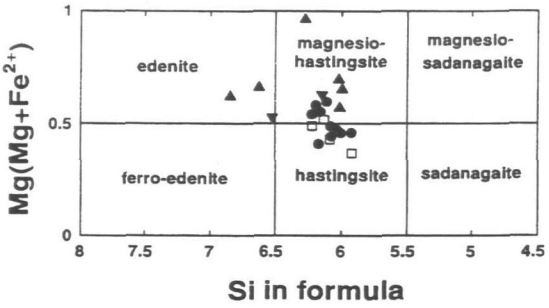


Fig. 13. Pyroxenes compositions plotted on the classification diagram of Flohr and Ross (1990).

Amphibole. Amphiboles in the essexite vary in composition from kaersutite to hastingsite (Fig. 14). Kaersutite is the characteristic amphibole of the essexite and the kaersutite found in the syenites is xenocrystic. Magnesio-hastingsite, hastingsite, and edenite are found in the syenites. These are all calcic amphiboles, and hence the amphiboles are also not alkali-rich, in agreement with the pyroxene chemistry.

Biotite. Biotite compositions are projected into a portion of Al-Mg-Fe^{2+} on Figure 15. Of note is that the biotites show restricted compositional variation, are not particularly Fe-rich, and are more aluminous then biotites from many other alkaline complexes. The most strongly Fe-enriched biotite is from the phonolite, and has a composition similar to that observed for biotites from some strongly alkaline nepheline syenite complexes in Malawi. While the analyses are not considered reliable due to the large amount of



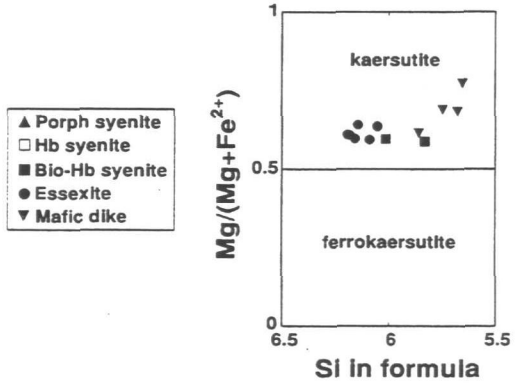


Fig. 14. Amphibole compositions plotted on the classification diagrams of Leake et al. (1997). Amphibole formulae calculated on the basis of 23 (O) with 2 (OH, F, Cl).

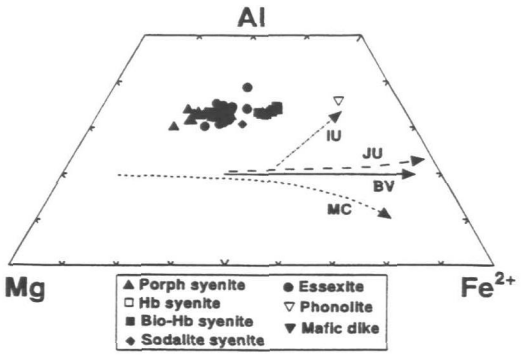


Fig. 15. Biotite compositions plotted in a portion of the Mg-Al-Fe^{2+} ternary diagram. Biotite formulas calculated on the basis of 24 (O+OH+F+Cl). IU = Ilomba and Ulindi (unpublished data, Woolley), JU = Junguni (Woolley and Platt, 1988), BV= Beemerville (unpublished data, Eby) and MC = Magnet Cove (Flohr and Ross, 1990).

alteration, and are not plotted on Figure 15, biotites from the quartz syenite are Fe-rich. Also plotted on Fig. 15 are trends for biotites from other nepheline syenite complexes: Ilomba and Ulindi, North Nyasa alkaline province, Malawi; Junguni, a sodalite-rich syenite complex, Chilwa alkaline province, Malawi; nepheline syenites of the Beemerville complex, New Jersey; and essexites through nepheline syenites of the Magnet Cove complex, Arkansas. The Cuttingsville biotites are characterized by their relatively high Al content compared to the other alkaline complexes.

Rock Chemistry

An extensive major, trace element, and radiogenic and stable isotope data base now exists for the Cuttingsville complex. A detailed discussion of these data is beyond the scope of this guidebook, but there will be an opportunity to discuss the data in the field. A manuscript is currently in preparation which will describe the results in detail. Laurent and Pierson (1973) distinguished two major petrologic series at Cuttingsville, both starting with essexite, one of which differentiated towards more silica-undersaturated magmas, and the other towards silica-saturated to over-saturated magmas culminating in the quartz syenites. Neither the unpublished data of Eby or Wood (1984) show this sharp two series distinction (Fig. 16). The essexites are all broadly similar in compositon irrespective of grain size, usually contain minor nepheline, and are Ne-normative largely because of the presence of kaersutitic amphibole rather than an abundance of modal nepehline. On the basis of petrography and chemistry, many of the syenites are cumulate rocks, and the late-stage porphyritic syenites contain enough xenocrystic material plus phenocrysts so that they do not, in most cases, represent liquid compositions. The quartz syenites do show what could be interpreted as a liquidus trend and the phonolite and sodalite apparently approximate liquid compositions (Fig. 17). The sodalite syenite, phonolite, and some of the porphyritic syenites plot in the thermal trough leading towards

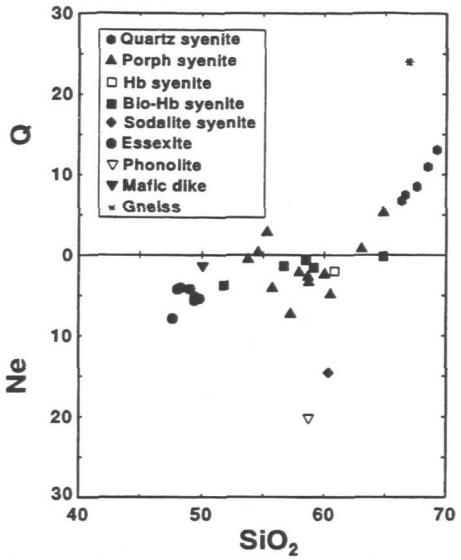


Fig. 16. Normative nepheline and quartz versus SiO_2 (wt. %).

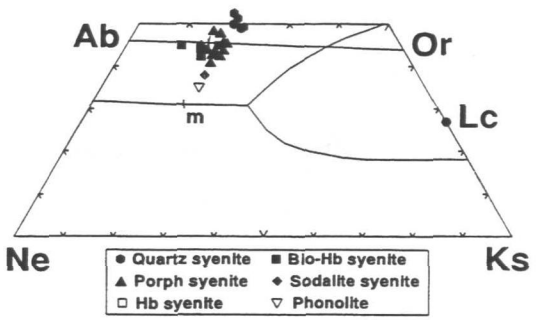


Fig. 17. Syenite compositions projected into the 1 kbar Ne-Qtz-Ks ternary phase diagram (Henderson, 1984). The phonolite and sodalite syenite plot close to the thermal minimum while some of the porphyritic syenites plot in the trough leading to the thermal minimum.

the low temperature minimum in the silica-undersaturated portion of the Ne-Qtz-Ks 1 kbar phase diagram (Fig. 17). The quartz syenites fall in a cluster just in the silica-oversaturated side of the phase diagram. Trace element and isotopic data suggest that these syenites have undergone evolution by AFC processes, however the amount of contaminant is apparently extremely small, probably less than 5%.

Petrogenetic Considerations

This topic will be more fully considered on the field trip. In brief, the sequence of events indicates that the essexites were the first phase emplaced in the conduit. Subsequent intrusions of syenitic magmas removed much of the early essexite, but remants are preserved both as continuous outcrop and as xenoliths, and xenocrystic minerals, in the syenites. Because of exposure and accessibility, much of the syenitic material you will see on the field trip is porphyritic. However, the bulk of the intrusion is composed of coarse-grained syenite and the petrography and chemistry of the coarse-grained syenites indicates that they are cumulates from a syenitic magma. While the essexites and some of the syenites are normatively silica-undersaturated, nepheline is found only in minor amounts and sodalite is confined to the sodalite syenite where may have been produced by crystallization from late stage fluids. The quartz syenites, the most silica-oversaturated rocks in the complex, never contain more than 5% modal quartz. Hence the Cuttingsville magmas were slightly silica undersaturated to slightly silica oversaturated.

Sr, Pb and O isotopic data indicate that the bulk of the Cuttingsville rocks were crystallized from magmas that had undergone little interaction with crustal rocks. For example, for the essexites and most of the syenites $(^{87}Sr/^{86}Sr)_i$ varies from 0.7034 to 0.7038. Only in the case of the quartz syenites and the quartz-bearing porphyritic syenites are higher $(^{87}Sr/^{86}Sr)_i$ ratios found (up to 0.7083). However, in the case of the quartz syenites absolute Sr abundances are low (<50 ppm) so these elevated ratios can be produced through minor interaction with the surrounding gneisses. Both the quartz syenite and the sodalite syenite show significant negative Eu anomalies, none of the other units have negative Eu anomalies and the coarse-grained syenites typically have small positive Eu anomalies. There is abundant evidence that all of the syenitic magmas were feldspar saturated, so these magmas were not above the liquidus. The constraints of phase petrology therefore require that the silica-saturated or undersaturated character of the magmas was determined during the early stages of magmatic evolution. While there is evidence that AFC processes

played a role in the evolution of the quartz syenite magma, this magma was already silica saturated and had undergone a significant degree of differentiation, before it was emplaced at high levels. Hence if crustal contamination played a role in determining the silica-saturated or undersaturated character of these magmas, the crustal contamination must have occurred early in the evolution of the original mantle-derived melts while they were still at relatively great depth in the crust.

SITE LOCATIONS

The region is generally rural, and famous for its scenery. Motels and other amenities are most abundant in Rutland, but that city also has the most unpleasant traffic flow of the area. U.S.G.S. topographic maps at 1:24,000 scale for the area include the Rutland (1980) and Wallingford (1986) sheets. We have found The Vermont Atlas and Gazeteer (DeLorme Mapping Co.) to be generally useful and widely available.

Along the way, a lunch break will be made in the village of Cuttingsville, probably after Stop 5. Gasoline is not available in Cuttingsville, but restaurants and gas stations are abundant along Rte. 7 in Rutland, between the meeting headquarters and the first stop of this trip. The total travel distance from start to finish is about 11 miles.

The first three stops are near roads, while the remaining sites are a few minutes walk from parking areas. Please be sensitive to private property at and near these sites, and watch for traffic, which can be passing by at high speeds.

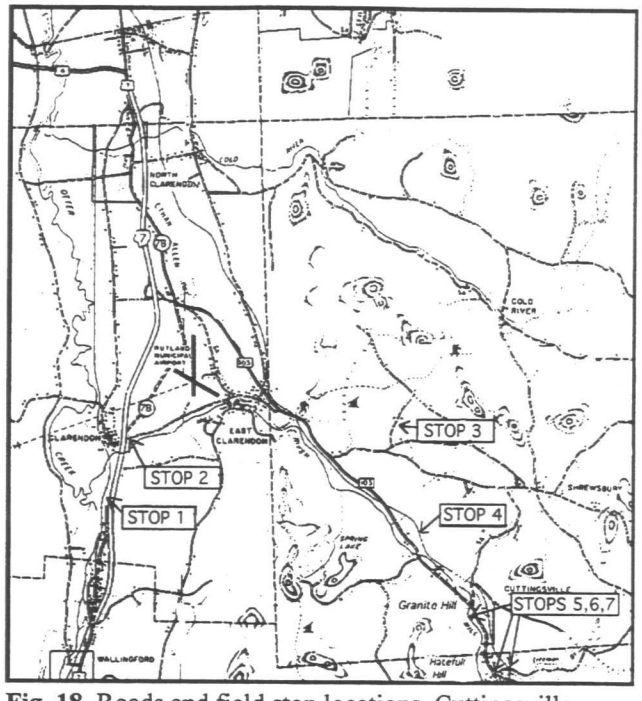


Fig. 18. Roads and field stop locations, Cuttingsville region.

ROAD LOG AND SITE DESCRIPTIONS

0.0 miles. START, Rte. 7 rest area.

We will assemble and start at 9:00 AM sharp from the highway rest area on the west (right) side of Rte.7, which is about 0.4 miles south of the intersection with Rte 103. The most direct route to this starting point from the Long Trail Lodge is to travel west on Rte. 4 to the intersection with Rte. 7 at Rutland, and then south (left) on Rte. 7. The intersection with Rte. 103 is several miles south of the city. Please arrive at least five minutes before our 9:00 AM start.

From the rest area, continue southbound on Rte. 7.

2.1 miles. Roadcut on west (right) side of Rte. 7. Pull off near southern end of cut.

STOP 1. SOUTH CLARENDON TRACHYTE DIKES

This very fractured dike (RT-4; AZ300,86; 230 cm) is hardly recognizable as an igneous intrusion until you examine the rock fragments. There is a very "shaley cleavage" developed parallel to the dike walls,

which must have a tectonic cause. The Ordovician dolostone country rock shows both faults and hydrothermal alteration at other outcrops in the area.

On the northern end of the cut there is a smaller, dark-weathering trachyte (RT-1; AZ 060,75; 14 cm) exposed. Small fingers of this dike have a green color, similar to that seen in other very thin trachytes that intrude dolostone (McHone, 1987). Perhaps a chemical reaction with the Mg-rich country rock has produced a fine-grained green mineral that is diluted in thicker dikes.

Compositionally, the trachytes are quartz syenite. In contrast to the abundant trachyte/bostonite dikes of the Lake Champlain Valley south of Burlington, there are only a handful of such felsic dikes around Rutland. As an end-member differentiate, the trachyte dikes are thought to be offshoots of shallow-level plutons. McHone and Wagener (1982) report 6 ppm U3O8 and 44 ppm eTh for dike RT-4, in agreement with other trachytes in Vermont.

Continue 0.2 miles farther south to turn around. Turn around by pulling to the right into a side road, then left across Rte. 7 to head back to the north. BEWARE of high-speed traffic!

3.1 miles. Roadcut with dikes on the east (right) side. Turn right onto old Rte. 7, then right uphill behind the roadcut, park (dead end road).

STOP 2. SOUTH CLARENDON DIABASE DIKES

Diabase, as used here, is essentially a basaltic rock that is appreciably altered, generally by hydrothermal solutions or weathering rather than by burial metamorphism (although the term has been used for low-grade metamorphosed rocks as well, for which we prefer "meta-diabase"). As at this site, it can be difficult to see much original texture and primary mineralogy. This dike (RT-2) and its small neighbor (RT-3) on the southern end of the cut may have originally been camptonite, but the mafic minerals are so changed that it is difficult to classify.

This E-W dike (AZ 084,86; 124 cm wide) has a small but distinct positive magnetic anomaly, as measured with a portable proton precession magnetometer (Fig. 19). The magnetic expression can be traced to the east for several hundred meters, as far as the Mill River. The dike is exposed in the river gorge on this magnetic line, and could possibly be traced much farther.

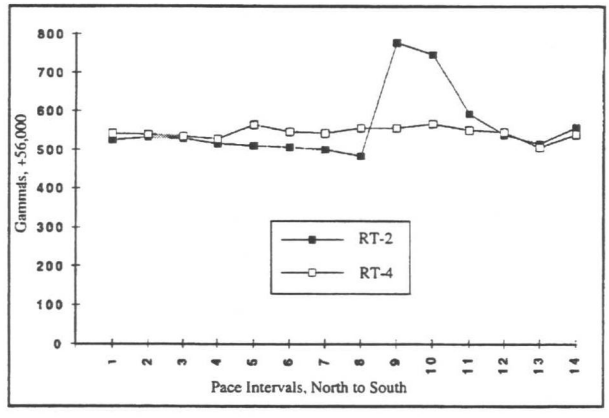


Fig. 19. Magnetic anomaly across diabase dike RT-2, traversed along the road ditch. A similar traverse across trachyte dike RT-4 reveals no magnetic expression, as expected for trachyte.

Turn right, away from Rte.7 and follow gravel road along the Mill River, to East Clarendon.

2.6 miles. Turn right at stop sign, in tiny village of East Clarendon.

3.0 miles. Turn right (southeast) onto Rte. 103.

4.2 miles. Turn left onto Maplecrest Farm Road. Go uphill, eastward.

5.0 miles. Bear left at intersection.

5.1 miles. Bear right at intersection.

5.2 miles. Stop along road near edge of woods to your right (east). Outcrops are low rock mounds in the woods.

STOP 3. SHREWSBURY INTRUSIVE BRECCIA

This site is on private property of Mr. Arthur Pierce, whose residence is at the last intersection. Please do not damage the fence or other property.

The map by Brace (1953) shows this breccia and a few others to the southeast in vague ovals, because exposure is poor. It was certainly a violent intrusion, full of clasts of local metamorphic rocks of the Grenvillian Mt. Holly complex, and there may be several "pipes" as the true forms of the intrusions. This site provided samples for Paul Doss (1986), who cataloged many of the lithologies within the breccia. Doss (1986) looked especially for sedimentary clasts of the Champlain Valley sequence, which would prove an overthrust relationship of the western Green Mountains, but none were identified.

The dike matrix is fairly fresh in a few places between xenoliths, and has a very volcanic, andesitic look in thin section. The date of 101 Ma (Table 1) is reasonable and indicates little contamination by K or Ar from the country rocks.

Turn around, head back to Rte. 103.

6.2 miles. Turn left (southeast) onto Rte. 103.

7.1 miles. RR crossing. Pull off highway onto right shoulder

STOP 4. HYPABYSSAL INTRUSIONS, NORTHERN RR CUT

Walk south approximately 600 m to the first of two railroad cuts. Eggleston (1918, p. 384) describes the set of outcrops along the railroad as follows: "The northern eruptives, where exposed, are much more involved with the country rocks than is the case on Granite Hill [the main intrusion]. Contorted and brecciated gneisses frequently alternate with eruptives along the railroad, and flank them on the northwest. The geological map [Fig. 1, Eggleston, 1918] owing to the limitations of its scale, gives a quite inadequate impression of the intricate relations between eruptives and country rock, especially in the case of the more northern area." At this stop you will have an opportunity to examine these complex relationships. In the first of the railroad cuts several varieties of porphyritic syenite are found cutting the gneisses. At the southern end of the first set of outcrops is a 6 m wide dike that weathers to in a knobly pattern. This dike is crowded with xenocrysts (photomicrograph, Fig. 11). Proceed south another 400 m to the next railroad cut. A porphyritic fine-grained syenite dike, approximately 20 m wide, cuts the gneisses toward the northern end of the railroad cut. Return to vehicles.

Continue SE on Rte. 103.

8.8 miles. If we have only a few vehicles, turn into the Stewart Ford dealer lot on right, and park in the back away from dealer stock (this is one of the oldest car dealers around, having been in business here since 1915). Several cars can also park in front of the closed storefront across and west of the car dealer. Otherwise, if we have more than 4 or 5 vehicles, we may park a little farther down the road in the hardware store/restaurant lot, and walk back.

STOP 5. NORTHERN GRANITE HILL AND MILL RIVER SYENITE FACIES

Walk up an old quarry/logging road into the woods above the car lot. Approximately 380 m up the road is an outcrop of quartz syenite cut by porphyritic syenite with pegmatitic zones. Continue for another 210 m to an outcrop of typical coarse-grained essexite. A phonolite dike cuts the essexite at the SW edge of the outcrop.

Return to the parking lot and walk south about 160 meters along the Mill River, if conditions permit (low water is helpful). Outcrops of quarz syenite, cut by porphyritic syenite, are exposed in the river. The contact between the porphyritic syenite and the quartz syenite is irregular indicating that the quartz syenite was still soft at the time of intrusion of the porphyritic syenite. Blocks of quartz syenite and essexite are found in the porphyritic syenite. Continue along the river another 170 m to a lamprophyre dike.

Continue SE across the bridge.

9.0 miles. Turn right into the entry road and parking lot between the restaurant and hardware store.

LUNCH STOP. The restaurant has been warned that we may appear. Soft drinks are also available at the hardware store. Please try to limit your lunch break to 45 minutes.

10.0 miles. Turn left into the left branching road just past the small stream bridge.

10.1 miles. Park along the road near the railroad crossing. Do not block the road, and do not park on the tracks!

STOP 6. SOUTHERN RR CUT

Walk approximately 400 m to the north along the railroad tracks. Medium-grained sodalite syenite is exposed in a 15 m long outrcrop on the east side of the railroad tracks. This is the location of the sodalite syenite (CV23) shown in Figure 9.

Return back down the road to Rte. 103.

10.3 miles. Park on the wide areas at the side road, but not on Rte 103.

STOP 7. SOUTHERN MILL RIVER SYENITE-DIORITE FACIES

This site requires participants to wade across the Mill River, which is our reason for making this the last stop! Although we do not expect the river to be particularly deep or swift, there is some risk of falling. If you are up for it, this site has some of the most interesting and varied exposures of the plutonic sequence to be found anywhere in the complex.

An almost continous sequence of outcrops occurs in a 350 m zone along the west side of the Mill River. Going north, the sequence begins with medium-grained, locally porphyritic, essexite. The essexite is cut by both mafic and felsic dikes. The felsic dikes often show a very anastomosing relationship with the essexite, and in places the dikes are disrupted suggesting that the essexite was plastic when the dikes were intruded. Continuing northward larger syenitic zones are encountered, often containing numbers mafic inclusions and abundant sulfides. At approximately 150 m medium-grained syenite with moderately abundant sulfides becomes the dominant unit. Continuing northward fine-grained syenite, often bluish in color, becomes the dominant lithology. Porphyritic syenite is associated with this unit, apparently as a facies of the blue syenite. Locally mafic inclusions are common, and the syenite outcrops are cut by numerous mafic dikes.

At the end of the syenite section, return to the Mill River crossing point. Rewade the Mill River, and return to cars for R&R.

END OF TRIP

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Basement-Cover Relationships in Southern Vermont

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INTRODUCTION

Soon after the development of plate tectonic theory, a variety of models invoking subduction near the continental margin of ancient North America (Laurentia) were developed to explain the Ordovician Taconian Orogeny in the Northern Appalachians (Dewey and Bird, 1970; Chapple, 1973; St. Julien and Hubert, 1975). Later, Rowley and Kidd (1981) and Stanley and Ratcliffe (1985) presented a refined plate tectonic reconstruction in which the Bronson Hill arc formed above an east-dipping subduction zone and collided with the passive continental margin of Laurentia as it was drawn into the subduction zone. For the past fifteen years this model has been the starting point for most studies of the igneous, metamorphic, and deformational history of western New England. The model of an arc colliding with a passive continental margin still seems like the best explanation for the current juxtaposition of tectonic units in the Taconian thrust belt. However, radiometric dating and geochemical studies (Karabinos and Aleinikoff, 1991; Karabinos and Tucker, 1992; Karabinos et al., 1993; Karabinos and Williamson, 1994; Karabinos et al., 1996) indicate that an arc considerably older than the Bronson Hill arc is preserved in western Massachusetts and eastern Vermont (Figs. 1 & 2), as suggested by Thompson (1990).

This older arc was first called the Shelburne Falls arc by Karabinos and Tucker (1992) for excellent dated exposures of the Collinsville Formation in the Shelburne Falls dome in northwestern Massachusetts. It includes the Barnard Volcanic Member of the Missisquoi Formation in Vermont. Karabinos et al. (1993) suggested that the Shelburne Falls arc, which formed between 485 to 470 Ma, collided with Laurentia during the Taconian orogeny and that the Bronson Hill arc formed after the Taconian orogeny above a newly developed west-dipping subduction zone east of the accreted Shelburne Falls arc. This suggestion is consistent with the data of Tucker and Robinson (1990) who showed that the Bronson Hill arc in central Massachusetts and southern New Hampshire formed between 454 to 442 Ma and is, therefore, too young to be the arc that collided with Laurentia during the Taconian orogeny.

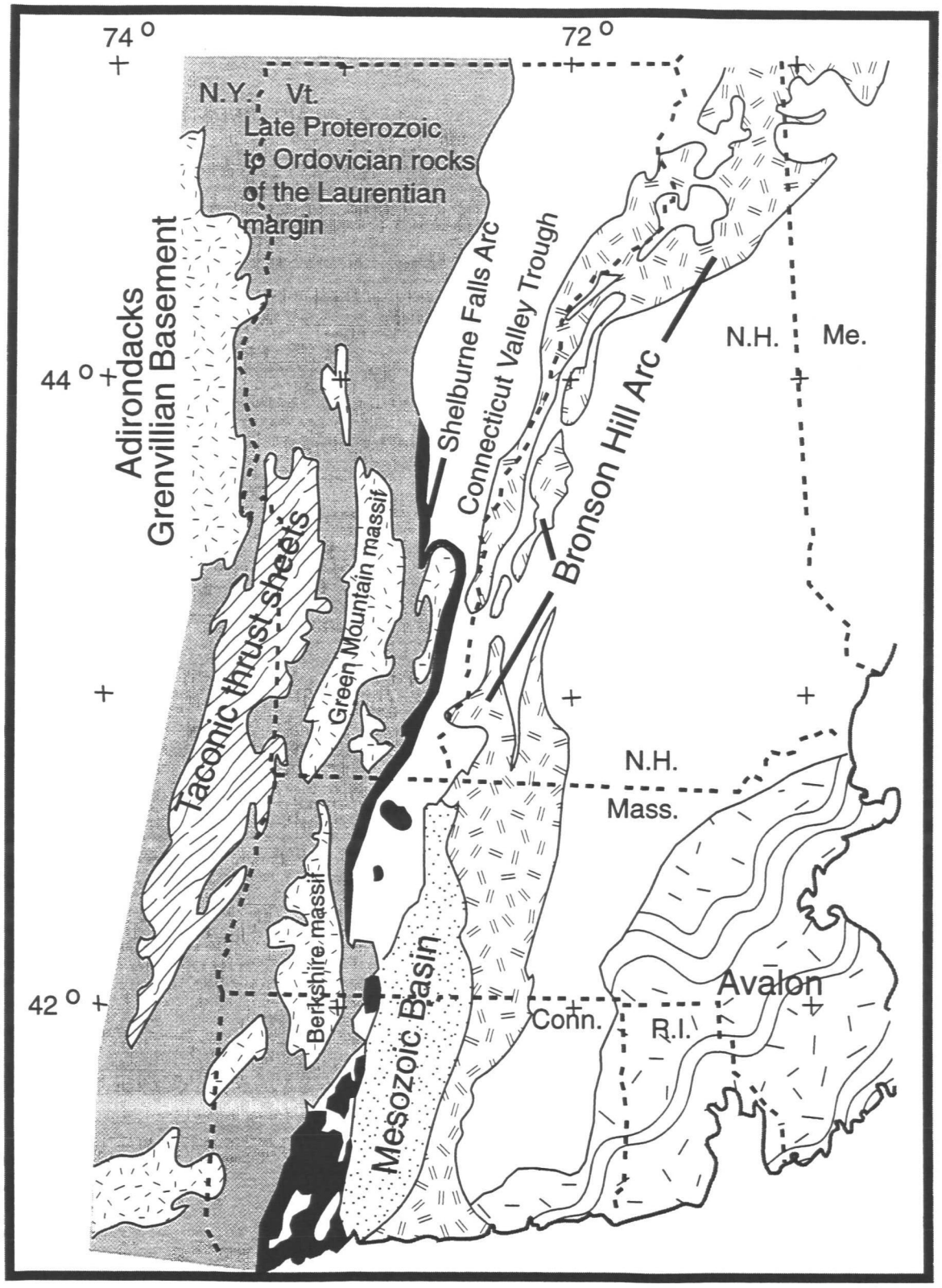


Figure 1. Simplified tectonic map of New England.

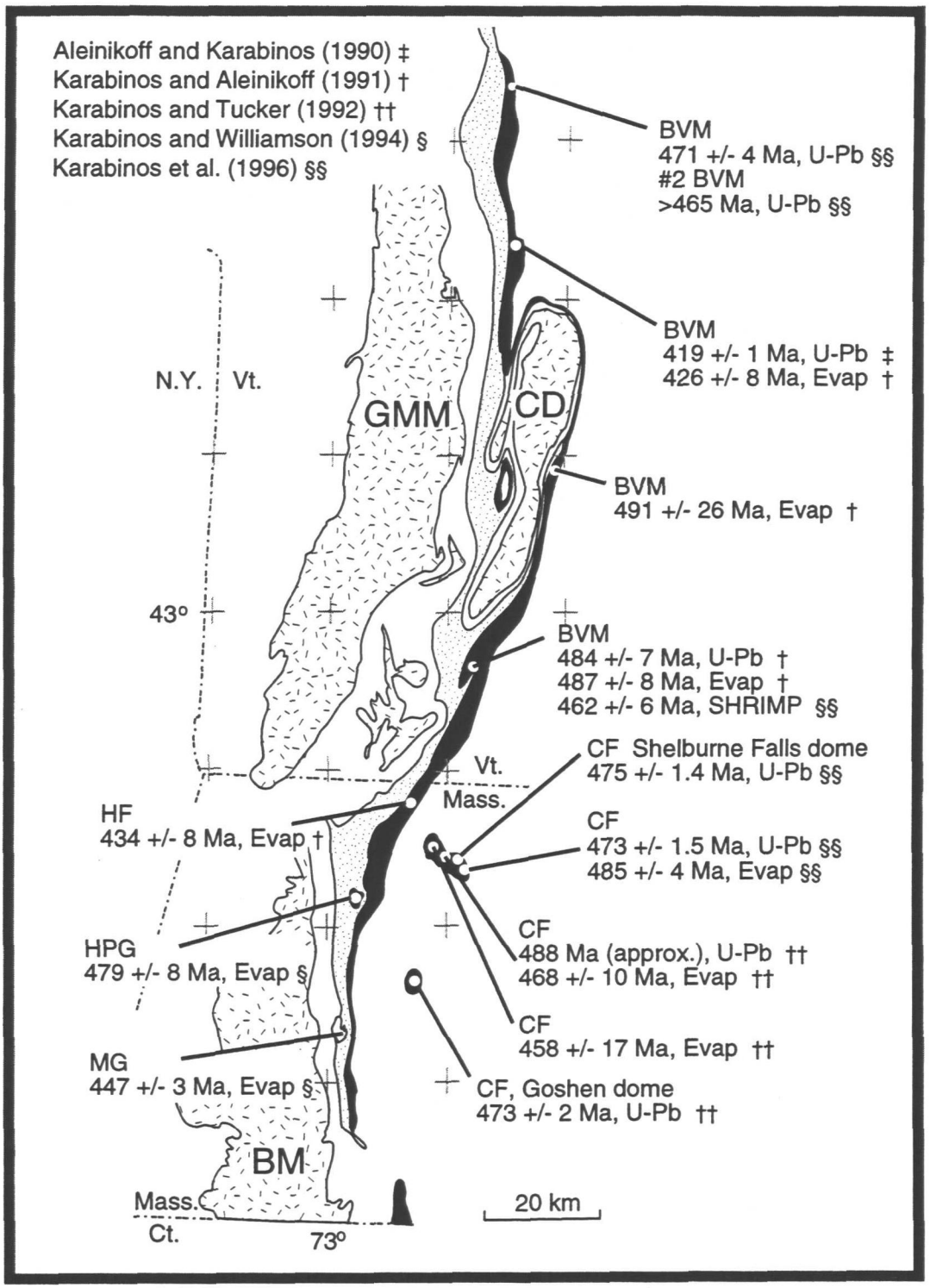


Figure 2. Summary of age data from the Shelburne Falls arc, shown in black. BVM- Barnard Volcanic Member, CF- Collinsville Formation, HPG- Hallockville Pond Gneiss, HF- Sill in the Hawley Formation, MG- post-Taconian Middlefield Granite. U-Pb- isotope dilution age, EVAP- evaporation age, SHRIMP- ion-microprobe age. BM- Berkshire massif, CD- Chester dome, GMM- Green Mountain massif.

If the plate tectonic model of an arc colliding with a passive continental margin during the Taconian orogeny is correct, it should be possible to identify the following features in the thrust belt:

1) The passive margin composed of Middle Proterozoic Grenvillian basement rocks and unconformably overlying Late Proterozoic to Ordovician cover rocks. The basal units of the cover sequence presumably record rifting whereas the upper units should reflect deposition in either a shallow shelf environment to the west or a deep slope-rise environment to the east.
2) The accretionary wedge or remnants of oceanic crust and sediments scraped off of the oceanic realm of the Laurentian plate as it was subducted under the Shelburne Falls arc.
3) Vestiges of the Shelburne Falls arc that collided with Laurentia.

It is important to identify these tectonic elements not just to confirm the plate tectonic model for the Taconian orogeny, but also to reconstruct the details of deformation and metamorphism during collision. Ongoing debates concerning the geology of this region center on which formations to include in the tectonic elements described above. Toward this end the problems we wish to address in this field trip are:

- How to distinguish between Grenvillian basement and younger cover rocks from the passive margin.
- How to distinguish between cover sequences that formed in the shallow shelf vs. the deep slope-rise of the passive margin.
- The boundary between rocks deposited on the continental margin vs. oceanic crust.

As debate about the history of western New England flourishes, it is easy to overlook the fact that tremendous progress has been made in understanding the geology of this complex region. For the most part, workers active in western New England agree about what tectonic elements exist but are hammering out the details of where to draw the boundaries and the kinematic details of how the units were juxtaposed.

GEOLOGIC SETTING

Easton (1986) described the Grenville Province in North America as a complex mosaic of rocks of widely ranging ages repeatedly deformed and metamorphosed during the Middle Proterozoic Grenville Orogenic cycle. By the Late Proterozoic, erosion had beveled this ancient mountain belt and continental rifting had formed basins which were filled by coarse clastic sediments and interlayered bimodal volcanics. Rifting gave way to production of new oceanic crust in the Late Proterozoic to Early Cambrian. As the proto-Atlantic (Iapetus) developed, a passive margin was established on ancient North America (Laurentia). The Grenvillian rocks found in the Green Mountain massif and Chester and Athens domes formed the "basement" on which Late Proterozoic to Middle Ordovician "cover" sediments were deposited.

The continental margin rocks can be divided into two distinct zones based on the nature of the Cambrian and Ordovician sediments. Both zones contain coarse basal clastic sediments of Late

Proterozoic age (the Dalton and the lowest part of the Tyson Formations in Vermont). West of the Green Mountain massif (in present coordinates), the basal clastics grade into Cambrian and Ordovician quartzites and carbonate sediments which were deposited in shallow water on the continental shelf (the Cheshire Quartzite, Dunham Dolomite, Monkton Quartzite, Winooski Dolomite, Danby Formation, Clarendon Springs Formation, and Shelburne Marble in Vermont). East of the Green Mountain massif fine grained shales and siltstones of Cambrian to Ordovician age were deposited in deeper water of the continental slope and rise (the Late Proterozoic to Cambrian Hoosac and Pinney Hollow Formations in Vermont). The divergence in sedimentary regime between the shelf and slope-rise sequences may have begun during the Late Proterozoic rather than the Cambrian, but precise dating of these rocks is impossible at present.

Regionally, dark grey to black carbonaceous flysch is found on top of both the shelf and slope-rise sequences (Hortonville Formation in Vermont, Walloomsac Formation in Massachusetts). This unit was eroded from active thrust sheets during the Taconian orogeny and was commonly overridden by the thrusts. Fossil dating of the syn-orogenic flysch provides the best control for age of Taconian thrusting available (Zen, 1967). Deposition was time transgressive and became younger to the west.

East of the Hoosac Formation (and Pinney Hollow Formation in Vermont) are rocks interpreted as the remains of an ancient ocean basin (Ottauquechee and Stowe Formations in Vermont, the Rowe Schist in Massachusetts). These rocks are composed of pelitic schist, thin layers of quartzite, amphibolite, and ultramafic rocks which probably formed as shale, chert, and oceanic crust/lithosphere, respectively.

Quartz and feldspar rich schists and gneisses commonly containing mafic layers and some pelitic schist layers are usually interpreted as the remains of a forearc basin (the Moretown Member of the Missisquoi Formation in Vermont, Mortetown Formation in Vermont).

Bimodal volcanics and intrusives with or without interlayered pelitic schists (the Barnard Volcanic Member of the Missisquoi Formation in Vermont, the Hawley and Collinsville Formations and the Hallockville Pond Gneiss in Massachusetts) are the remnants of the Shelburne Falls arc.

The Silurian and Devonian meta-sediments (the Shaw Mountain, Northfield, Gile Mountain, and Waits River Formations in Vermont) are calcareous pelitic schists and quartzites deposited in a large basin, the Connecticut Valley-Gaspe trough that extended at least from Connecticut through Quebec. These rocks were deposited after the Taconian orogeny, but they were deformed during the Devonian Acadian orogeny. They are important in understanding the Taconian orogeny in two ways. First, the effects of Acadian deformation and metamorphism have to be taken into account when examining the Taconian orogeny in regions of overlap (Rosenfeld, 1968). Second, the Collinsville Formation in Massachusetts, part of the Shelburne Falls arc, is exposed in domes surrounded by the Silurian and Devonian metasediments.

STRATIGRAPHIC SUMMARY

Middle Proterozoic Basement Rocks

Mount Holly Complex

Yma, Augen Gneiss: Microcline, plagioclase, quartz, biotite, muscovite, epidote gneiss. Also occurs as 1-3 m thick layers within other rock types in the Mount Holly Complex. U-Pb zircon ages of approximately 1125 Ma (Karabinos and Aleinikoff, 1990).

Ymq, Quartzite and feldspathic quartzite: Clean, massive, vitreous, blue quarzite. Feldspathic and micaceous quartzite commonly containing chlorite and garnet. Rare beds of quartz, muscovite, paragonite, chloritoid schist.

Ymm, Marble and calc-silicate rock: Coarse-grained calcite marble. Tremolite, talc, calcite, epidote, phlogopite schist; may contain chlorite, plagioclase, and microcline.

Ymf, Felsic Gneisses: Heterogeneous plagioclase, quartz, ±K-feldspar, biotite, chlorite, epidote, muscovite gneiss; may contain altered garnet.

Eastern Cover Sequence Rocks

Tyson Formation

CZtc, Basal Conglomerate: Sand-sized detrital grains to boulders of blue quartz or quartzite, feldspar, and less commnly lithic fragments of gneiss in a matrix of quartz, albite, muscovite, biotite, chlorite schist.

CZts, Schist: Quartz, albite, muscovite, chlorite, biotite schist; commonly contains weathered pits and nodules of carbonate minerals. Light gray carbonate-rich quartz, muscovite schist.

CZtq, Quartzite: Light gray, fine-grained quartzite containing minor amounts of muscovite and feldspar.

CZtd, Dolomite: Fine-grained, buff-weathering, massive dolomite; many beds contain detrital quartz grains. Magnetite-rich near upper contact with Hoosac Formation.

Hoosac Formation

CZh, Albite porphyroblast, quartz, biotite, muscovite, chlorite schist near base. Paragonite, chloritoid, large garnet schist ± kyanite, ± staurolite. Commonly, garnet displays distinctive unconformity texture. Dark gray graphitic quartz, albite, muscovite, biotite, chlorite phyllite; commonly contains beds of dark quartzite, rarely contains thin beds of dolomite.

Pinney Hollow Formation

CZph, Quartz, albite, muscovite, chlorite phyllite or fine-grained schist. Quartz, muscovite, paragonite, chloritoid, chlorite phyllite. Less commonly albite-porphyroblast, quartz, muscovite, biotite, chlorite schist. Includes epidote, chlorite, quartz, plagioclase, muscovite greenstone.

Rocks of Uncertain Affinity

Pinnacle Formation

CZp: Medium to dark gray quartz, feldspar, biotite, muscovite, chlorite metagraywacke. Detrital grains of quartz, feldspar, and mica typical. Well bedded. Contains beds of quartz, albite, muscovite, biotite, chlorite schist.

Western Cover Sequence Rocks

Dalton Formation

CZd: Brown to grey conglomerate in a quartz, feldspar, muscovite matrix.

Cheshire Quartzite

Cc: White, vitreous, massive quartzite. Interbeds of dark gray quartz, muscovite, biotite, chlorite phyllite and feldspathic, micaceous quartzite common near base.

Dunham Dolomite

Cd: Buff-weathering dolomite containing siliceous partings and detrital quartz grains.

Monkton Quartzite and Winooski Dolomite

Cmw: Interbedded impure quartzite and buff, orange, yellow, or dark gray dolomite. Green and dark gray beds of phyllite also present. Winooski Dolomite contains somewhat less quartzite than the Monkton Quartzite. Difficult to separate these units in the Rutland area.

Danby Formation

Cda: Vitreous quartzite interbedded with gray calcitic dolomite. Dolomitic quartzite and quartzose dolomite. Cross-bedding common.

Clarendon Springs Formation

Ccs: Gray calcitic dolomite.

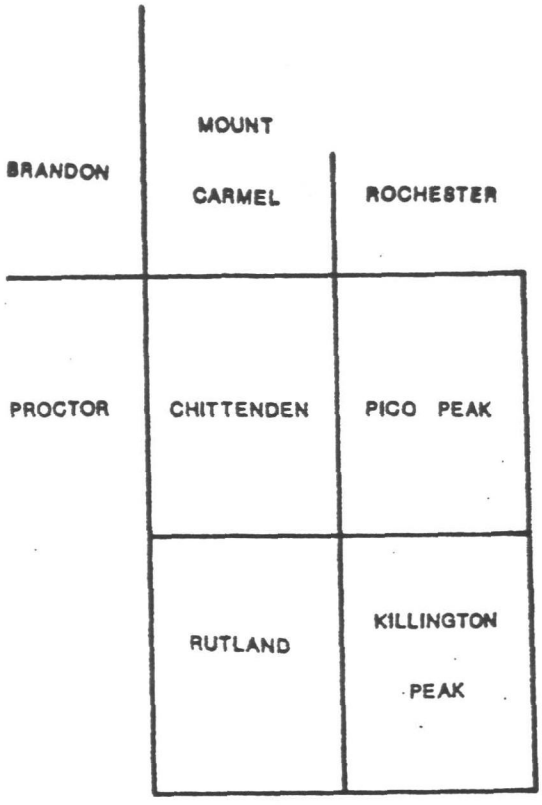
Shelburne Marble

Os: White calcite marble. Intermediate gray dolomite unit.

DISTINGUISHING BASEMENT FROM COVER ROCKS

Rocks belonging to the Grenvillian basement (Fig. 3) experienced sillimanite grade metamorphism and higher during the Middle Proterozoic. Cover rocks went through biotite to garnet grade metamorphism in the vicinity of the Green Mountain massif and kyanite grade metamorphism near the Chester dome. Thus, it is usually possible to distinguish basement from cover rocks on the basis of metamorphic mineral assemblages. This approach can be less than straightforward in the Green Mountain massif where basement rocks have been retrograded to assemblages similar to the biotite grade cover rocks near fault zones.

Partial melting of basement rocks during Middle Proterozoic orogenesis led to widespread development of pegmatites which are commonly highly deformed. Such pegmatites are absent from cover rocks in the Green Mountain massif. Pegmatites are locally present in cover rocks



EXPLANATION

WESTERN COVER SEQUENCE

Os: Shelburne Marble
€cs: Clarendon Springs Formation
€da: Danby Formation
€mw: Undifferentiated Monkton Quartzite and Winooski Dolomite
€w: Winooski Dolomite
€m: Monkton Quartzite
€d: Dunham Dolomite
€c: Cheshire Quartzite

€Zp: Pinnacle Formation (uncertain affinity)

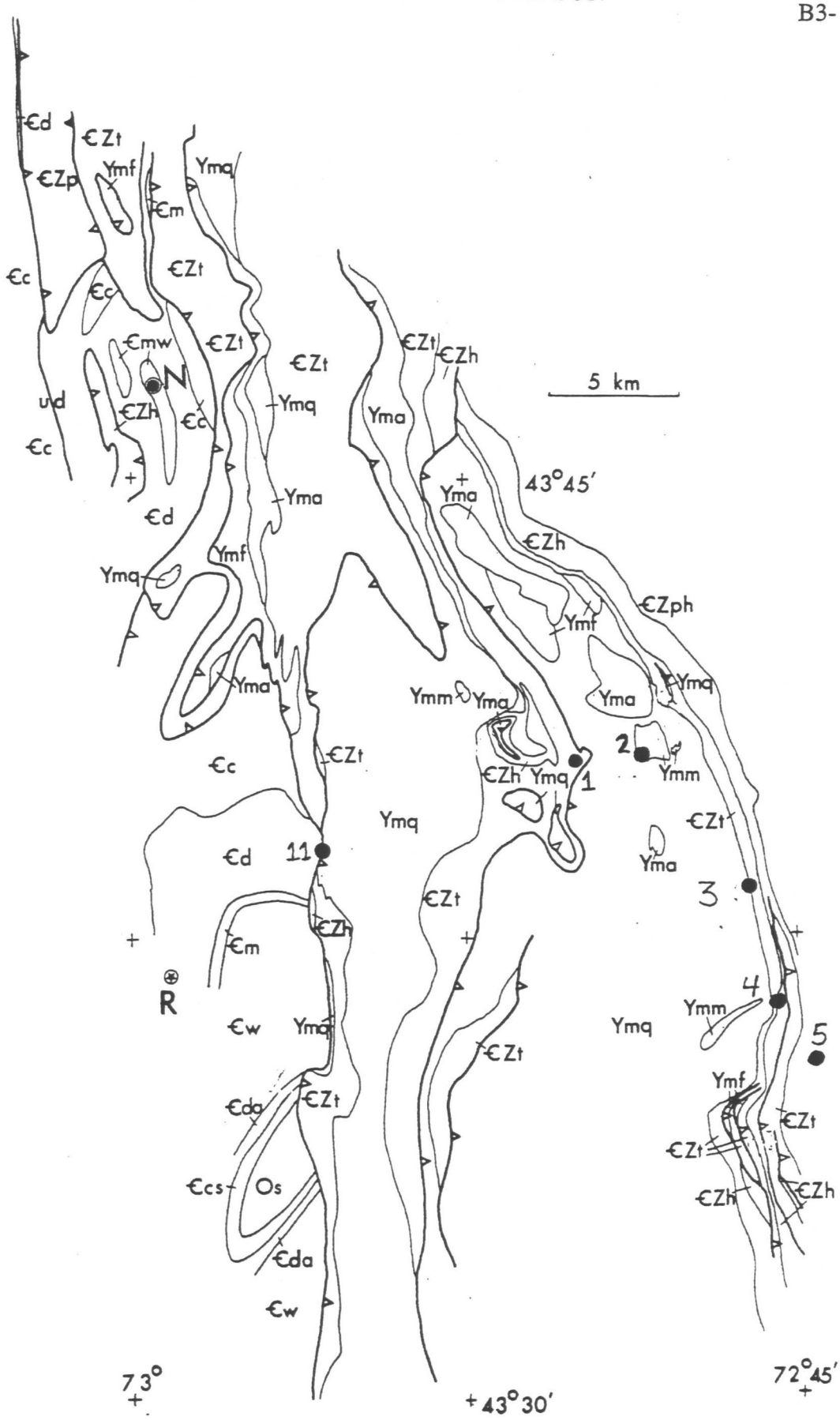
EASTERN COVER SEQUENCE

€Zph: Pinney Hollow Formation
€Zh: Hoosac Formation
€Zt: Tyson Formation

-------- unconformity --------

Mount Holly Complex
(Relative ages uncertain)
Ymf: Felsic gneiss
Yma: Microcline augen gneiss
Ymm: Marbles and calcsilicates
Ymq: Quartzites and feldspathic quartzites

Figure 3. Geologic map of the northern end of the Green Mountain massif. Structure in the western cover sequence south of Rutland, Vt., (R) based on Brace (1953) and Doll et al. (1961). Heavy lines- thrust faults, teeth point to upper plate.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around the Chester dome, but they are post- or late-kinematic and are weakly deformed. Cover rocks throughout the region may contain feldspar veins, which we suppose could be confused by some as pegmatites. Such veins, however, do not possess typical pegmatite mineral assemblages of quartz, feldspar, and mica; they are usually made up exclusively of plagioclase.

Another distinctive feature of the basement rocks is the presence of blue quartz, particularly as veins. The blue color is produced by tiny rutile needles exsolved from quartz and is characteristic of very high grade metamorphic rocks that experienced a later episode of upper greenschist to lower amphibolite facies metamorphism (Niggli and Thompson, 1979). The early high-temperature metamorphism is necessary for TiO_2 to be soluble in quartz and the later lower-temperature heating episode is necessary for the TiO_2 to exsolve and form rutile needles. Some care is needed when using this criterion for distinguishing between basement and cover, however, because detrital quartz grains in cover rocks which were derived from basement rocks may be blue. Blue quartz veins are not found in cover rocks, however, so their presence provides a useful diagnostic feature exclusive to basement rocks.

One problem in distinguishing basement from cover rocks is that similar meta-sedimentary lithologies are found in both packages. For example, albite gneisses are found in both the Hoosac Formation and the Mount Holly Complex. These rocks frequently do not contain mineral assemblages that tightly constrain peak metamorphic conditions. The best methods for assigning such rocks to the correct unit rely on the presence or absence of deformed pegmatites and the proximity of other rock types.

An interesting basement-cover problem surfaced recently. Ratcliffe (1994) reassigned calc-silicates, quartzites, and pelitic schists in the Chester dome and elsewhere shown as cover rocks belonging to the Cavendish Formation by Doll et al. (1961) to the Mount Holly Complex based on inferred structural relationships. Ratcliffe et al. (1996) further concluded that these rocks, formerly thought to be Late Proterozoic to Early Cambrian in age must be older than 1.4 Ga based on $^{207}Pb/^{206}Pb$ SHRIMP ages on a tonalite which they interpreted as intruding the Cavendish Formation. This new age assignment of the Cavendish Formation demanded a completely new interpretation of the structure of the Chester dome area, which has appeared in recently published quadrangle maps (e.g. Ratcliffe, 1995).

To test this new age assignment, Karabinos collected a sample of quartzite from the Cavendish Formation in Cavendish Gorge very close to the inferred intrusive contact. Zircons separated from the quartzite form a heterogeneous population of detrital grains. Grain shapes range from nearly spherical to nearly euhedral. Virtually all the grains have pitted surfaces and show at least some rounding of edges and terminations. Single zircon grains were analyzed at Brown University using the evaporation method originally described by Kober (1986, 1987) and modified by Karabinos and Gromet (1993).

The age data are summarized in Figure 4 and it is clear from the young ages of the detrital zircon grains that the quartzite cannot be older than 1.4 Ga, as suggested by Ratcliffe et al. (1996). The youngest zircons are approximately 940 to 980 Ma and were probably derived from the Stamford Granite Gneiss or the Bull Hill Augen Gneiss (Karabinos and Aleinikoff, 1990). These

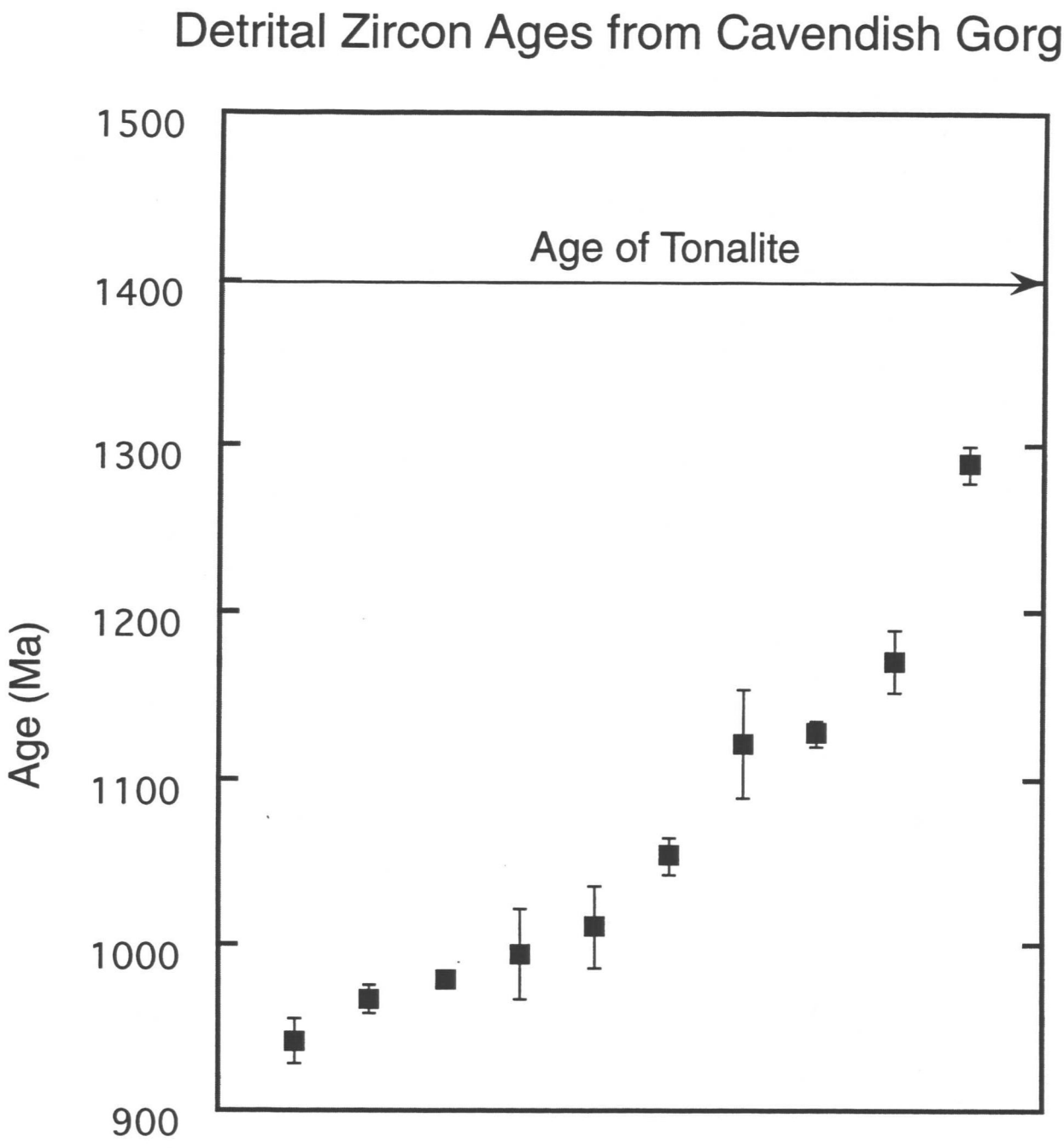

Figure 4. Ages of detrital zircon grains from quartzite from Cavendish Gorge. Reference line gives age of tonalite reported by Ratcliffe et al. (1996). Error bars given at 2-sigma level.

young ages are significant because the Stamford Granite Gneiss and the Bull Hill Augen Gneiss formed after the Grenville orogenic cycle; therefore the metasedimentary units of the Cavendish Formation must also post-date Grenville deformation and metamorphism, in accord with earlier interpretations.

We believe that some rocks shown as Cavendish Formation by Doll et al. (1961) may be part of the Mount Holly Complex but that most of the formation is part of the cover sequence. The metasedimentary package in the Cavendish Formation that occurs throughout southern Vermont and includes calcsilicate or marble, quartzite, albite schist, and high-aluminum schist is best correlated with the upper part of the Tyson Formation and the Hoosac Formation (Thompson et al., 1993a,b).

DISTINGUISHING SHELF FROM SLOPE-RISE SEQUENCES

It would be very useful to be able to reconstruct unambiguously the depositional environments of cover sequence rocks. We would be able to restore thrust sheets to their original relative order and constrain the kinematics of thrusting. At a minimum, we should strive to determine which cover rocks were deposited on the shelf and which on the slope-rise of Laurentia. One difficulty is that the basal units of both the western and eastern sequences formed during early stages of rifting and are hard to distinguish. Another problem is that there may well have been local ranges (horsts) surrounded by basins (grabens) that persisted for some time after rifting of the continental margin. There may even have been large plateaus far off the coast where sediments would resemble shelf sequence rocks more than adjacent deep-water deposits.

On the west flank of the Green Mountain massif between Wallingford and Bennington, Vermont, the basement rocks of the Mount Holly Complex are unconformably overlain by basal clastics of the Dalton Formation. In this area the Dalton Formation is mostly conglomeratic and thin (< 100 m). Above it is quartzite and marble of the western shelf sequence.

East of the Green Mountain massif basement rocks are unconformably overlain by the Tyson Formation which includes a basal conglomerate, pelitic schist, quartzite, and dolomitic marble. The Hoosac Formation overlies the dolomitic marble of the Tyson Formation. The Hoosac Formation is dominated by albite schist but includes highly aluminous chloritoid-paragonite schist and minor quartzite and dolomitic marble (Plymouth Member). The Pinney Hollow Formation overlies the Hoosac Formation and is dominated by highly aluminous schist but also contains many layers of albite schist. We interpret this sequence of rocks as being part of an eastern cover sequence probably deposited on the slope of Laurentia. It is difficult to directly correlate these units with the western shelf sequence because of the lack of fossils east of the Green Mountain massif. The dolomitic marbles found in the eastern sequence may record periods of shallow water deposition. Indeed, the iron rich horizon at the top of the dolomitic marble of the Tyson Formation, which is also present at the top of the marble unit of the Cavendish Formation, has been interpreted as a terra rosa by Thompson (1972). An alternative interpretation is that at least some of the discontinuous quartzite and marble layers originated as sedimentary lenses and

were derived from shelf sequence rocks to the west. Keith and Friedman (1977) proposed that similar quartzite and marble lenses in Cambrian rocks of the Taconic sequence formed by fluidized sediment flows and debris flows, which carried shelf derived material into deeper water. The Tyson, Hoosac, and Pinney Hollow Formations contain several lithologies in common (pelitic schist and discontinuous quartzite and marble lenses) that suggest that their depositional environment varied over time but was not dramatically different. There are faults within this sequence of rocks, but we do not believe that they are major tectonic boundaries.

Karabinos (1988) used the characteristics of cover sequence rocks to divide the Green Mountain massif into a tectonically higher, northeastern part overlain by eastern cover rocks and a lower, southwestern part overlain by western cover rocks.

DISTINGUISHING SLOPE-RISE AND OCEANIC SEQUENCES

This is arguably one of the least important problems in reconstructing the tectonic history of the Taconian orogeny. The difference between sediments deposited on the continental rise and on oceanic crust must be minor. Furthermore, sediments from both environments were deposited on the Laurentian plate and later incorporated in an accretionary wedge, so faults separating oceanic from continental rise rocks are not plate boundaries or sutures; they are just faults. The plate boundary should be preserved between the arc complex and the accretionary wedge. Nonetheless, it would be useful to be able to recognize sediments deposited on oceanic crust to help confirm the plate tectonic model for the Taconian orogeny and to estimate crustal shortening.

In Massachusetts this boundary is identified as the contact between the Hoosac Formation and the Rowe Schist (e.g. Stanley and Ratcliffe, 1985). We suggest that in Vermont the boundary roughly coincides with the contact between the Pinney Hollow and Ottauquechee Formations. Mafic units in the upper part of the Pinney Hollow Formation show geochemical trends that can be interpreted as reflecting volcanism on highly attenuated continental crust (Coish, 1993). Furthermore, we do not find evidence for a major tectonic boundary or important lithologic variations between the Hoosac and Pinney Hollow Formations as suggested by Stanley and Ratcliffe (1985). However, the Ottauquechee Formation contains rock types different from those in the Hoosac and Pinney Hollow Formations. It is composed of pelitic schist, thin layers of quartzite, amphibolite, and ultramafic rocks which may have formed as shale, chert, and oceanic crust/lithosphere, respectively.

Perhaps the most important tectonic boundary in eastern Vermont is the contact between the Stowe and Missisquoi Formations. The Moretown Member of the Missisquoi Formation is a likely candidate for the forearc of the Shelburne Falls arc, although it contains detrital zircons of Grenvillian provenance (Karabinos and Gromet, 1993), so it was probably deposited near the Laurentian margin. The Barnard Volcanic Member of the Missisquoi Formation, as noted in the introduction, is part of the Shelburne Falls arc and, therefore, is a relict of a separate plate from Laurentia.

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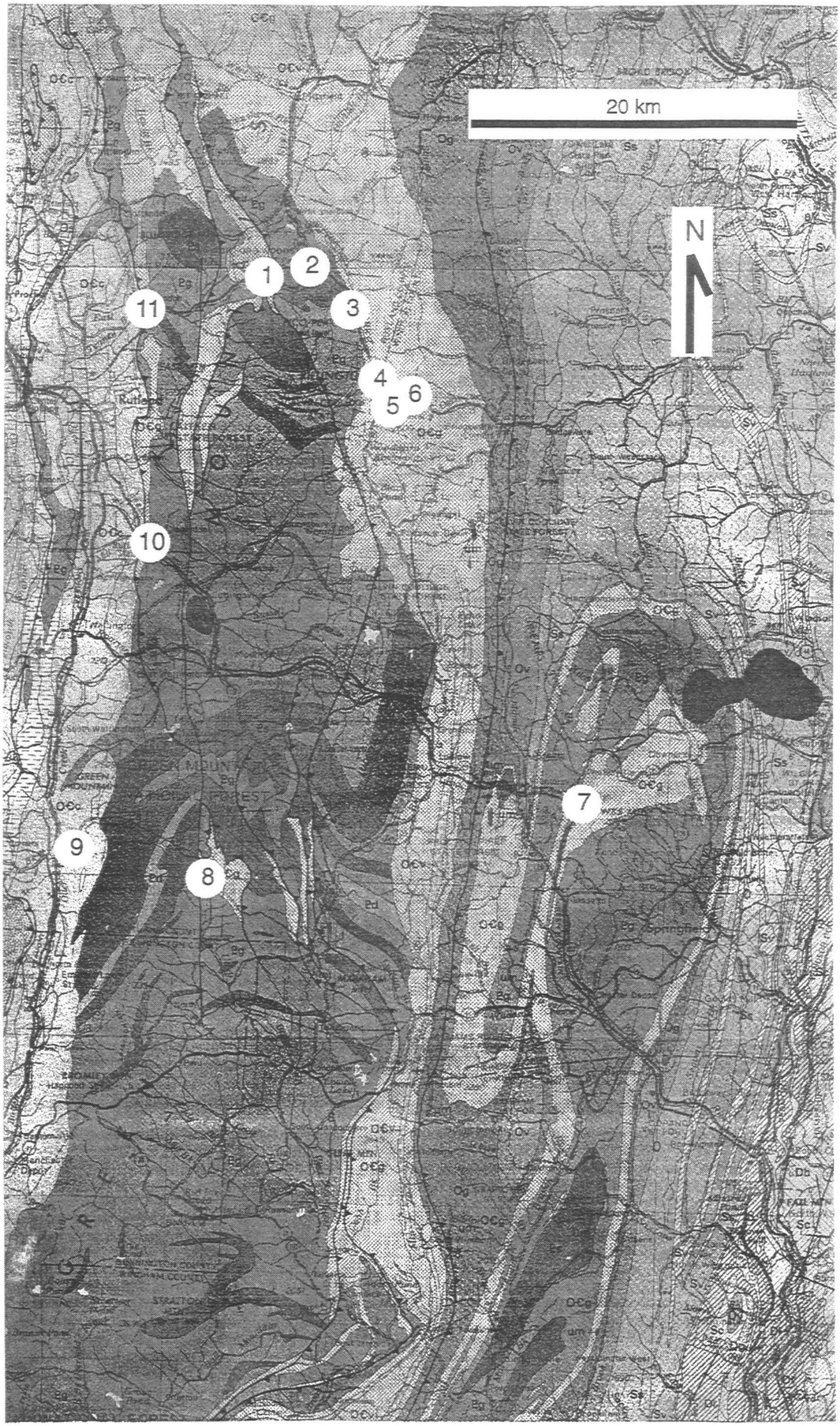

Figure 5. Geologic Map from Thompson et al. (1990) showing location of stops.

ROAD LOG

0.0	0.0	Start at Pico Peak ski area. Right on Rt. 4 E. Figure 5 shows the stops on this trip. Stops in the northern part of the Green Mountain massif are also shown in Fig. 3.
0.6	0.6	**STOP 1.** Right into parking area across from Deer Leap. Cross road and follow Long Trail to outcrops of quartz, plagioclase, muscovite, chlorite schist with abundant quartz veins. We believe this unit is part of the cover sequence and correlate it with the Tyson Fm. Higher up the trail are exposures of greywacke and conglomerate. Fine-grained dolomitic marble is exposed at the western base of the slope. Continue east on Rt 4.
2.1	1.5	Left on Rt 100 N.
2.6	0.5	**STOP 2.** Right into Kent Pond parking area. Follow Appalachian Trail N to bridge. From bridge downstream to Kent Pond are exposures of interbedded quartzite, marble, and calc-silicate rocks of the Mount Holly Complex. This package is common on the broad terrace east of Pico Peak and west of the Rt. 100 valley. The coarse grained marble and calc-silicate rock of the basement is very different from the fine grained dolomitic marble of the cover sequence. Return south on Rt. 100.
3.2	0.6	Left on Rt. 4 E & Rt. 100 S.
5.2	3.0	**STOP 3.** Pull over to right. Outcrops of deformed conglomerate of the Tyson Fm close to the basement-cover contact which is highly sheared. Continue on Rt. 4 E & Rt. 100 S.
7.9	2.7	Turn left on Mission Farm Rd.
8.5	0.6	**STOP 4.** Pull over to right. Outcrops of Hoosac Fm. above dolomitic marble of the Tyson Fm. Albite schist of Hoosac Fm. contains abundant magnetite. Dolomitic marble of Tyson Fm. contains iron oxide layers that were once mined for ore. The iron-rich zone may reflect subaerial weathering. The quartzite and marble of the Tyson Fm. and albite schist of the Hoosac Fm. form a distinctive package found throughout the Chester, Athens, and Wilmington domes and the Green Mountain massif which we will see at Stops 7 & 8. Continue south on road.
8.7	0.2	Left on Rt. 4 E & Rt. 100 S.
9.8	1.1	Continue on Rt. 4 E.
10.1	0.3	**STOP 5.** Pull over to right. Outcrops on left of Pinney Hollow Fm. Green quartz, muscovite, paragonite, +/- chloritoid phyllite and schist. Some layers are albitic and indistinguishable from albite schist of the Hoosac Fm. We find it hard to believe that there is a profound tectonic

boundary between the Hoosac and Pinney Hollow Fms. because they are so similar lithologically.
Continue on Rt. 4 E.

11.3	1.2	**STOP 6.** Pull over to right. Outcrops on right of Ottauquechee Fm. Carbonaceous, sulfidic phyllite interbedded with with carbonaceous quartzite. Mafic layers are not present here but are common in the Ottauquechee Fm. The boundary between the Pinney Hollow and Ottauquechee Fms. seems to be a likely candidate for the transition from continental margin to oceanic sediments deposited on the Laurentian plate in the Cambrian and Early Ordovician. The strain displayed by this unit is enormous; it appears to have acted like rollers or ball bearings allowing overlying units to be transported westward. Continue on Rt. 4 E.
11.4	0.1	Reverse direction at small road on right. Head west on Rt. 4.
13.0	1.6	Left on Rt. 100 S.
16.2	3.2	**OPTIONAL STOP.** Large outcrop on left of dolomitic marble. Outcrop on right of conglomerate. Both are part of the Tyson Fm.
27.7	11.5	Left on Rt. 100 S & Rt. 103 S.
29.7	2.0	In Ludlow continue on Rt. 103 S where it diverges from Rt. 100 S.
32.7	3.0	Outcrop on left of ultramafic rocks altered to serpentine.
33.0	0.3	Left on Rt. 131 E.
35.6	2.6	Right on dirt road just before water treatment ponds.
35.8	0.2	**STOP 7.** Left into parking area. Cavendish Gorge in the Chester dome. We will walk through the gate to the power plant. Near concrete footings is the site of a sample collected by Ratcliffe and Aleinikoff of a tonalite that gave a SHRIMP zircon age of approximately 1.4 Ga. They interpreted the contact between the tonalite and the metasedimentary units as intrusive and concluded that the Cavendish Formation is older than 1.4 Ga. However, the quartzite, marble, calc-silicate, and albitic schist of the Cavendish Fm. are really very similar to the rocks seen at the top of the Tyson Fm. and base of the Hoosac Fm. on the east side of the Green Mounain massif and we would correlate them. To test this new age assignment, Karabinos collected a sample of the quartzite and extracted detrital zircons. The evaporation method gave ages as young as 940 Ma (see Fig. 4). Therefore, the Cavendish Formation cannot be as old as postulated by Ratcliffe and Aleinikoff. We still maintain that these rocks are part of the cover sequence based not only on the zircon ages but also on the metamorphic history preserved in the rocks. Return to Rt. 131.
36.0	0.2	Left on Rt 131 W.
38.6	2.6	Right on Rt. 103 W.

41.9	3.3	Left on Rt. 100 S.
49.3	7.4	Intersection with Rt. 151. Left, on Rt. 100 S.
52.9	3.6	Weston. Right at village green. Go over bridge.
53.1	0.2	Left on Landgrove Road.
57.4	4.3	Hard right on Little Michigan Road.
57.7	0.3	Right on U.S.F.S. 10.
63.3	5.6	**STOP 8.** Pull over. Devils Den. Outcrop on right of fine grained dolomitic marble and quartzite. Upslope to right is albite schist. There are also mylonitic rocks and basement lithologies here. This was shown as Cavendish Formation by Doll et al. (1961), and we would correlate it with the upper part of the Tyson Fm. and lower part of the Hoosac Fm. and with rocks at Cavendish Gorge. We also believe that these rocks unconformably overlie basement and that they are part of a deep water cover sequence. If these interpretations are correct, a major fault must separate these rocks from the western shallow water cover sequence and its unconformably underlying basement which we will see at the next stop. Continue on U.S.F.S. 10.
70.8	7.5	**STOP 9.** Pull over to left in parking area on far side of bridge. Outcrops in stream of clean white vitreous Cheshire Quartzite. Cross beds visible in some exposures. This is typical of Cheshire Quartzite on the west flank of the Green Mountain massif from Bennington north to Wallingford. The Dalton Fm. underlying the Cheshire Quartzite is typically an arkosic conglomerate not more than approximately 100 m thick. Above the Cheshire Quartzite is the Dunham Dolomite and the rest of the western shelf sequence. North of Wallingford the cover rocks on the west side of the massif are very different and include minor conglomerate and abundant greywacke, schist, and dolomitic marble that resemble the Tyson and Hoosac Fms. from the east side of the massif. This cover sequence and underlying basement rocks are separated from rocks clearly belonging to the western shelf sequence by thrust faults. Coninue down road.
71.8	1.0	Right on Rt. 7 N.
84.7	12.9	Right at sign for Rt. 7B but continue straight at 20 m instead of turning left on Rt. 7B
86.3	1.6	Right on Airport Road.
86.7	0.4	Right on Rt 103 S.
87.2	0.5	**STOP 10.** Right into parking area just before railroad tracks. Follow Appalachian trail S to suspension bridge. Outcrops below bridge of quartz-rich schist and greywacke. These rocks are typical of the basal cover sequence found on the west side of the Green Mountain massif north of Wallingford. Some layers of albite schist and chloritoid bearing

schist downstream. Rare pebbly layers also present. If time permits we will cross Rt. 103 and follow the Appalachian trail N to look at the basement-cover contact in the power line west of the trail.
Return on Rt. 103 W.

87.7	0.5	Right on small road and head north.
90.5	2.8	Good view to northeast of Bald Mountain.
90.9	0.4	Stop sign. Straight then bear right.
91.1	0.2	Stop sign. Left.
93.0	1.9	Stop sign. Straight.
93.4	0.4	Stop sign. Right on Stratton Road.
94.0	0.6	Stop sign. Right on Stratton Road
95.0	1.0	Stop sign. Left on Town Line Road.
96.9	1.9	Stop sign. Right on Rt. 4 E.
98.1	1.2	**STOP 11.** Pull over to right. Outcrops on right of quartzite and schist of the Mount Holly Complex. Blue quartz veins and pegmatites typical of basement lithologies. Rutland City contracted for a water well to be dug on the far side of the pasture across Rt. 4 in the early 1980's. The well went down about 500 feet without hitting bedrock- not exactly typical of Cheshire Quartzite or Dalton Fm. which is what should be there if the basement-cover contact were a simple unconformity. There is strong evidence for a thrust along the western contact of the massif. It was mapped by Brace (1953) and Karabinos (1988). Continue on Rt. 4 E.
103.3	5.2	Pull into parking lot of Pico Peak ski area.

ECONOMIC GEOLOGY OF THE DIMENSION STONE INDUSTRY EASTERN VERMONT GRANITE BELT

by

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INTRODUCTION

Several Devonian granitoids intrude the Siluro-Devonian metasedimentary rocks of the Connecticut Valley-Gaspé Trough in eastern Vermont between the Massachusetts and Quebec borders, and many of the smaller, finer grained bodies have been quarried for dimension stone at one time or another. During this field trip, we will examine active dimension stone quarry operations in three small intrusions — the Bethel, Barre, and Woodbury bodies. Each of these quarry areas produces a distinctive grey to white granite. Each area also has a proud history of commercial activity. The Bethel area opened about 1902, and the Barre and Woodbury areas have been active for over 100 years. And, interestingly, each district produced the granite for a state capitol building — Bethel White Granite was used for the Wisconsin state capitol in Madison; Barre Granite quarries produced the stone for the Vermont State House in Montpelier, and Woodbury Granite was used to construct the Pennsylvania capitol building in Harrisburg. (Woodbury Granite is also used for part of the state capitol in Boise, Idaho.)

Figure 1 Historic distribution of granite quarry districts in Vermont. 1 Bethel, 2. Barre, 3. Woodbury. Modified from Dale (1923).

Although only one of the quarries (Woodbury) that we will visit today produces stone that is geologically classified as granite, the term granite in the dimension stone industry means *any* igneous rock (yes, including so-called black granites — diabase, gabbro and anorthosite). We apply the industry use of the term granite herein.

Today, these three granite quarry areas represent the remnants of what was once an industry that stretched from Dummerston, near Brattleboro in southern Vermont, to Derby at the Quebec border (Figure 1). Nonetheless, Vermont is one of the top two granite producing states in the US.

NEIGC last officially visited the Barre Granite

district in 1961, with a field trip led by Brewster Baldwin of Middlebury College (Baldwin, 1961). To our knowledge, the Bethel and Woodbury quarries have not been included in NEIGC field trips in the past.

The quarries we visit on this field trip are all owned and operated by Rock of Ages Corporation of Barre, Vermont (Bethel and Barre) and its parent company, the John Swenson Granite Company of Concord, New Hampshire (Woodbury). Rock of Ages operates a Visitor Center in Barre that welcomes visitors to viewing platforms and tours. The Bethel and Woodbury quarries are not open to the public, however, and should not be visited without first obtaining permission from the owners.

REGIONAL SETTING

The Connecticut Valley-Gaspé Trough in eastern Vermont is generally characterized by tightly folded phyllitic and calc-silicate metasediments of the Siluro-Devonian Gile Mountain and Waits River Formations. The metasediments are interpreted to have been deposited in the remnants of the Iapetus ocean. Dozens of granitoids of the New Hampshire Plutonic Series intrude the Gile Mountain - Waits River metasediments (Doll and others, 1961). A few New Hampshire Plutonic Series plutons also intrude the Ordovician Missisquoi Formation on the west, and the Maidstone and Victory plutons in the Northeast Kingdom extend into the Ordovician metasediments on the east.

In general, the intrusions are two-feldspar, two-mica minimum-melt type granitoids. There likely was no complex fractionation history. In general, the intrusions are unzoned, relatively uniform in composition, and not associated with mafic phases. The intrusions range in size from small exposures a few hundred feet long to plutons about 40 miles in diameter.

Contacts with the surrounding metasediments are mildly to strongly discordant. Metamorphic aureoles are associated with most of the larger intrusions, indicating that the plutons were intruded after the peak of Acadian regional metamorphism. Naylor (1971) obtained K-Ar dates on primary micas in several of the Vermont granitoids that clustered around 380 Ma. He concluded that the Acadian orogenic event was brief, but intense, since the Devonian plutons crosscut Siluro-Devonian sediments that had been folded and regionally metamorphosed before intrusion.

DIMENSION STONE QUARRYING

The geologic requirements for dimension stone quarrying are relatively straightforward and simple — a desirable deposit contains a marketable stone that is, first of all, pretty, uniform in color, composition, and texture, free of minerals that weather badly, relatively massive (i.e., unfractured), and located with easy access to transportation. Finding such bodies is more difficult than one might think.

Dimension stone quarrying in the granites of eastern Vermont has been concentrated in the

smaller bodies, which are typically fine grained and grey in color. The larger plutons generally are coarser grained, less homogeneous in color and texture, and commonly contain pegmatite veins. Dale (1923) provides detailed descriptions of the quarries in existence in the early part of the century, and his work remains as the most comprehensive reference on granite dimension stone resources in New England. Few of the active granite quarries in New England today are not described in Dale's work.

Quarrymen have an intuitive understanding of rock mechanics and have always used the natural fracture properties of granite to extract rectilinear blocks. They are, of course, constrained by large fractures and jointing, but good quarrymen also use the micro-structure of granite to split out blocks. In the dimension stone industry, the three directions in which a block of granite will split are called rift, grain, and hardway (or headgrain), in order of decreasing ease of splitting.

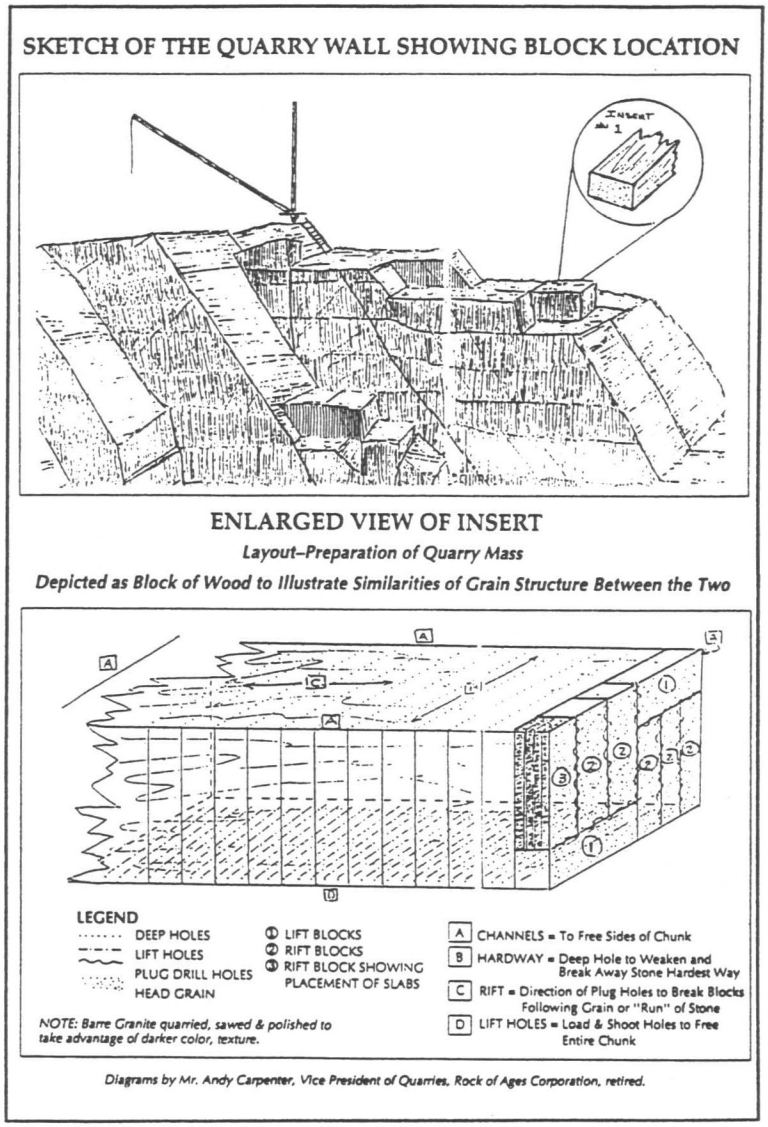

Figure 2 Sketch of quarry process taken from Rock of Ages (1992).

The rift is a plane containing the densest population of subparallel microcracks, and it is developed to varying degrees in different granites. The rift is relatively well developed in most of the Vermont granites, but granites in some areas have essentially no rift. Rift is parallel to sheeting in most granites and, in such granites, it is probably genetically related to unloading of granite as it reaches the earth's surface. In the Barre and Woodbury granites, the rift is a near vertical direction, however, and we infer that it was imposed tectonically. Rift can be recognized in thin section and hand specimens as a set of subparallel cracks that cross quartz grains. A well developed rift makes granite easier to split by hammering small wedges (plugs and feathers) inserted in small diameter shallow holes. The other directions generally must be drilled more deeply.

In some ways, quarry technology has advanced enormously, and in other ways, the same

processes are still in use that were used by the ancient Egyptians. Current quarry technology has moved away from deep quarries with stationary lifting derricks toward quarries with drive-in access and that use mobile lifting equipment wherever possible. Noise concerns have reduced the use of oxygen and other torches to burn channels to free large chunks of granite from the quarry walls prior to drilling out blocks. Torches are being replaced by high speed slot drills for cutting channels. Diamond wire saws have come into more common use in granite quarries. But there is still a lot of drilling and hard work involved. Figure 2 is a sketch that depicts the process of quarrying a block of Barre Granite.

BETHEL WHITE GRANITE

The Bethel White Granite is a light colored, fine to medium grained quartz monzonite that occurs in a small, northerly striking semi-concordant intrusion on the eastern flank of Christian Hill. As shown on Ern's (1963) geologic map of the Randolph 15' quadrangle, it is about 5000 feet long and 600 feet wide. The contacts are not well exposed but are approximately parallel to the strike of the surrounding Waits River Formation metasediments. A much smaller body occurs a few hundred feet to the east, adjacent to the Interstate.

One of the most interesting aspects of the Bethel White granite to geologists is its orbicular texture. Rather than occurring as disseminated crystals as in most granites, the biotite occurs in concentric clumps elongated parallel to flow. The orbicules commonly include feldspar and quartz grains, but it's not clear that there is a consistent nucleus. In places, orbicules a few inches long occur, but most are smaller. Swarms of orbicules occur, especially in the southern exposures, and partially resorbed xenoliths and schlieren are also common. In the orbicules of the eastern limb of the Bethel Granite, garnet occurs near the center of some of the orbicules. Similar orbicular granites occur in other small bodies in Vermont, such as a well known occurrence in Craftsbury, but none of them has been quarried to any extent.

Eastward dipping sheeting fractures are well exposed in the quarries, and the vertical spacing of the sheets range from a few inches near the surface to over 15 feet at depth. Other joint sets generally strike about N85E. The quarrier reports that the Bethel White Granite does not possess a prominent rift and that the granite quarries with about the same ease in three directions. Dale (1923) states that the rift is nearly horizontal to about 15°, and that the grain is nearly vertical and oriented N17E.

From a commercial standpoint, a pure white granite is desirable, and the orbicules are considered undesirable. Thus, quarry activities have been concentrated where the granite is more uniformly white in color. The Bethel White quarries opened about 1902, and as noted in the introduction, provided granite for the state capitol in Madison, Wisconsin, as well as portions of the Smithsonian Institution and Union Station in Washington.

In recent decades, Bethel White granite has been far more popular abroad than domestically, and it has always seen more use as a building stone than as a monumental stone. A limited amount of Bethel White granite is used as skid caps and tank liners for pickling lines

in steel plants. But almost all production from this quarry is now shipped as unprocessed saw blocks to Europe and Asia for use as building stone.

BARRE GRANITE

The granite quarried at Barre is actually a granodiorite that occurs in a sinuous body elongated parallel to the strike of the enclosing metasediments of the Gile Mountain Formation on the west and Waits River Formation on the east. Richter (1987) provides a general description of the granite and quarries. The pluton is shown on the state bedrock geological map (Doll and others, 1961) and in Murthy (1957) as being about 5 miles long and almost 2 miles wide. A large septum of Gile Mountain Formation roughly follows the road between Graniteville and Websterville. Primary flow fabrics are faintly visible in most quarry faces, and xenoliths in various stages of digestion by the magma are abundant near the contacts.

Two styles of jointing are evident in the Barre quarries — high angle joints with a northeast strike, and sheeting joints that follow the topography. In the Adam Quarry (Stop 2A), high angle joints predominate. They dip toward easterly at about a 45° angle, are commonly coated by greenish chloritized alteration products, and some have slickensides. In contrast, fracturing in the Smith Quarry (Stop 2B) is characterized by sheeting joints. The vertical spacing of the sheets increases from a few inches near the original ground surface to greater than 30 feet at depth. The rift is nearly vertical in all of the Barre quarries and generally strikes about N50E. The grain is approximately horizontal and is called the lift in the Barre quarries.

Quarrying activity began in the 1830's with the extraction of granite on Cobble Hill for the Vermont State House in Montpelier. The quarry industry became a major force in the Barre area after the Civil War, when the railroad was constructed and steam-powered tools made quarrying a bit easier.

Quarrying has been concentrated in the southern part of the pluton, where the granite is uniformly fine grained and darker grey in color. The finer grained, darker granite is favored for cemetery monuments, and lighter grey, coarser Barre Granite is used as building stone and for fabricating industrial products. The Barre Granite district is one of the major granite producers in the US, quarrying almost 1 million cubic feet of granite annually.

Cemetery monuments and building stone are familiar products of the dimension stone industry, but the industrial use of granite products is one aspect of the granite industry unfamiliar to most. Rock of Ages is a major producer of granite industrial products, and its market is worldwide. Barre Granite is ground into precision tools such as toolmaker's flats, straightedges, master squares, parallels, angle plates, and air-ride riser blocks. Surface plates of various sizes and configurations are used as stable inspection surfaces and machine bases in low and high tech industries. Granite press rolls are large cylinders of post-tensioned granite up to about 30 feet long and 5 feet in diameter. They are used in paper mills, turning at high speed in large machines to de-water the pulp as part of the paper manufacturing process. Rock of Ages is one of only two companies in the world that produces granite press rolls.

Most of the granite quarried in Barre is used in cemetery monuments, however. Barre Granite is used as building stone to a lesser extent. Modern-day building projects include the 70-story Texas Commerce Bank Tower and the ARCO corporate headquarters in Texas.

WOODBURY GRANITE

The Woodbury Granite is a medium grained medium to light grey granite that occurs as numerous small discrete knobby exposures poking through the Waits River Formation in the vicinity of Woodbury and Hardwick (Doll and others, 1961). Konig (1961) shows the locations of most exposures in his bedrock map of the Plainfield 15" Quadrangle. They presumably are connected at depth, and may represent the roof of a larger body. Several of the larger exposures have been quarried in the past, and some toward the northeast are described as porphyritic. The only active quarry in the Woodbury granite is located in the largest knob about a mile east of the town center.

The Woodbury Granite is lighter colored and coarser grained than Barre Granite. Southeast dipping flow textures are readily visible, but xenoliths are rare in the active quarry.

Fracturing in the Woodbury Granite quarry is dominated by sheeting joints that follow the surface topography. High angle jointing is widely spaced, and large quarry blocks can be extracted. The rift is vertical and its strike is reported by Dale (1923) to be N35E. The stone splits well in three directions, and this quarry has the least waste of any included in this field trip.

The Woodbury Granite quarries opened in the 1880's. Because it is relatively coarse grained, light grey in color, and does not exhibit strong contrast between polished and rough finishes, Woodbury Granite has not been widely used for monuments. Woodbury Granite has largely been used for building stone. The Pennsylvania capitol in Harrisburg is the largest building to have been constructed of Woodbury Granite, and construction of major additions to the capitol complex in the late 1980's spurred reopening and development of the modern Woodbury Granite quarry. Currently, most production from this quarry is used for cut landscape material (benches, walls, posts, etc.).

ACKNOWLEDGMENTS

We thank Helen Mango of Castleton State College for inviting us to participate in this year's NEIGC. We also thank Rock of Ages Corporation and the John Swenson Granite Company for permitting us to visit their active quarries and fabrication facilities.

ROAD LOG

We assemble in the parking lot at McCullough's Quik Stop and Citgo Station on the south side of Route 107 (Figure 3). The meeting place is located 1.4 miles east of the junction of Route 12 and Route 107 in downtown Bethel if you are coming from the Pico-Killington area, and 0.7 miles west of exit 3 of I-89 if arriving from the interstate. Tank up on gas, if you haven't already.

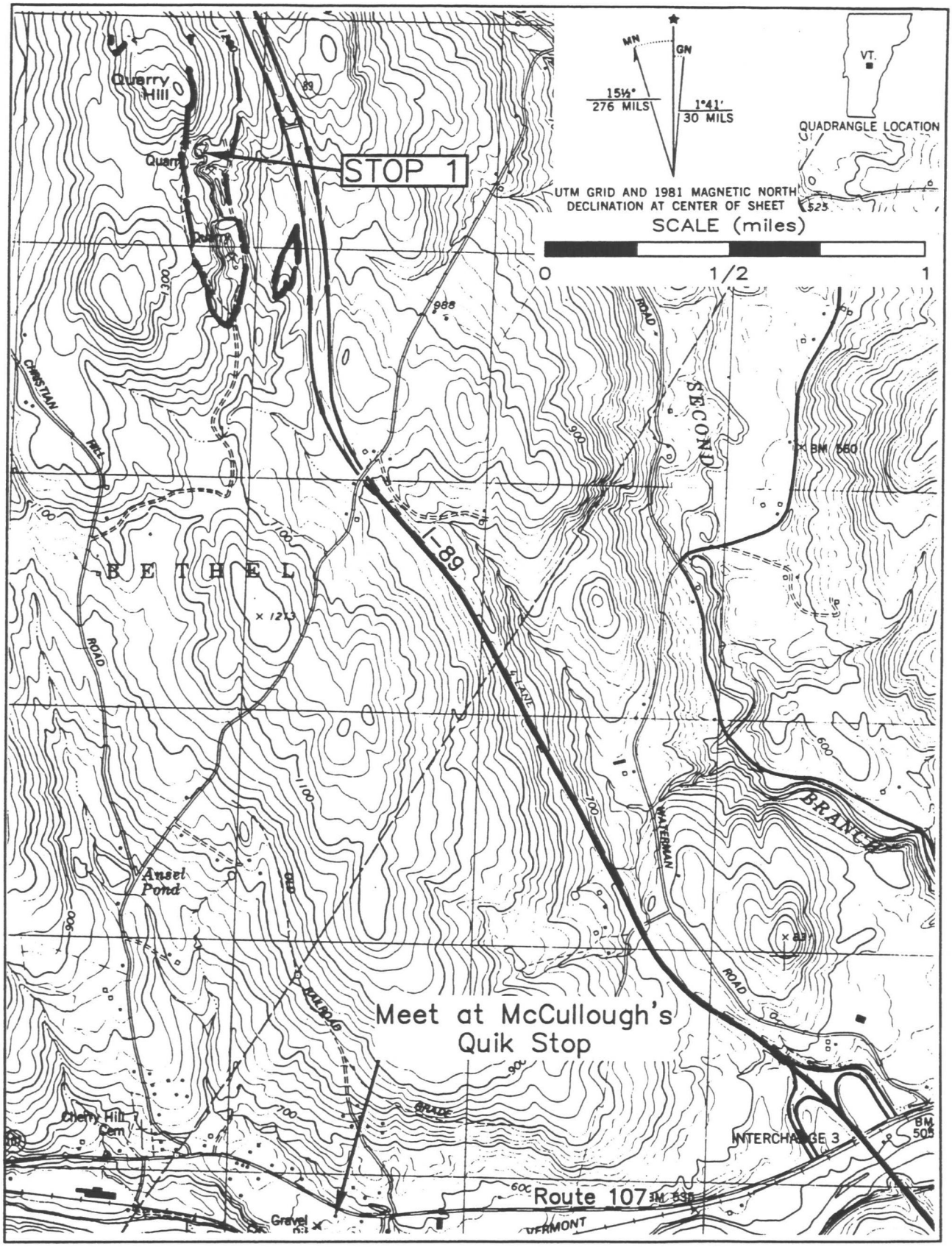

Figure 3 Location of Stop 1, Bethel White Granite quarry. Approximate granite contacts shown as dashed lines. Base map South Royalton 7.5" USGS topographic quadrangle.

Incr.	Cum.	
0.0	0.0	Exit the parking lot at McCullough's Quik Stop, and turn left (west) on Route 107.
0.4	0.4	Turn right (north) on Christian Hill Road. The road climbs a hill, passing a cemetery on the left.
0.8	1.2	Fork in the road. Take the left (north) fork and pass Ansel Pond.
0.9	2.1	Turn right (east) on gravel road.
1.7	3.8	Park in storage yard at north end of roadway.

STOP 1. BETHEL WHITE GRANITE QUARRY. (45 minutes, location shown in Figure 3) The Bethel White Granite quarry currently is a drive-in operation, and we will walk down to the working area. The north and west faces of the quarry rise to the natural ground surface of the hill and provide beautiful views of the sheeting joints. Primary igneous flow and Bethel's famous orbicules are easily recognized in the faces as well. In the southwestern corner of the active quarry, an east-west lamprophyre dike is exposed. The dike crosses the granite and what is either a large xenolith or the contact with phyllites of the Waits River Formation.

South of the dimension stone quarry are large exposures where crushed stone was quarried for construction of I-89. When the exposures were fresh, prominent bands of schlieren and swarms of orbicules were visible. Currently, some large irregular blocks in that area are partially covered with bright blue silicone. Reportedly, the silicone was applied by a company that made simulated rocks for the movie industry.

Reverse direction and return down gravel road. It's about a ½-hour drive to the next stop.

1.7	5.5	Turn left (south) on Christian Hill Road.
0.9	6.4	Pass Ansel Pond and stay right when merging with fork from left.
1.2	7.6	Turn left (east) on Route 107.
2.4	10.0	Junction Route I-89. Turn left (north) onto I-89 at exit 3.
4.8	14.8	Take exit 6, South Barre exit, and proceed on Route 63.
2.1	16.9	Good view of quarries ahead.
1.9	18.8	Junction Route 14 at a light. Cross Route 14 and proceed up the hill on Middle Road to Graniteville Road.
1.5	20.3	Rock of Ages Craftsman Center on left. Stay right.
0.3	20.6	Keep left at fork in road.
0.7	21.3	Turn right (south) on private gravel road. Pause to allow cars to re-assemble. Proceed south.

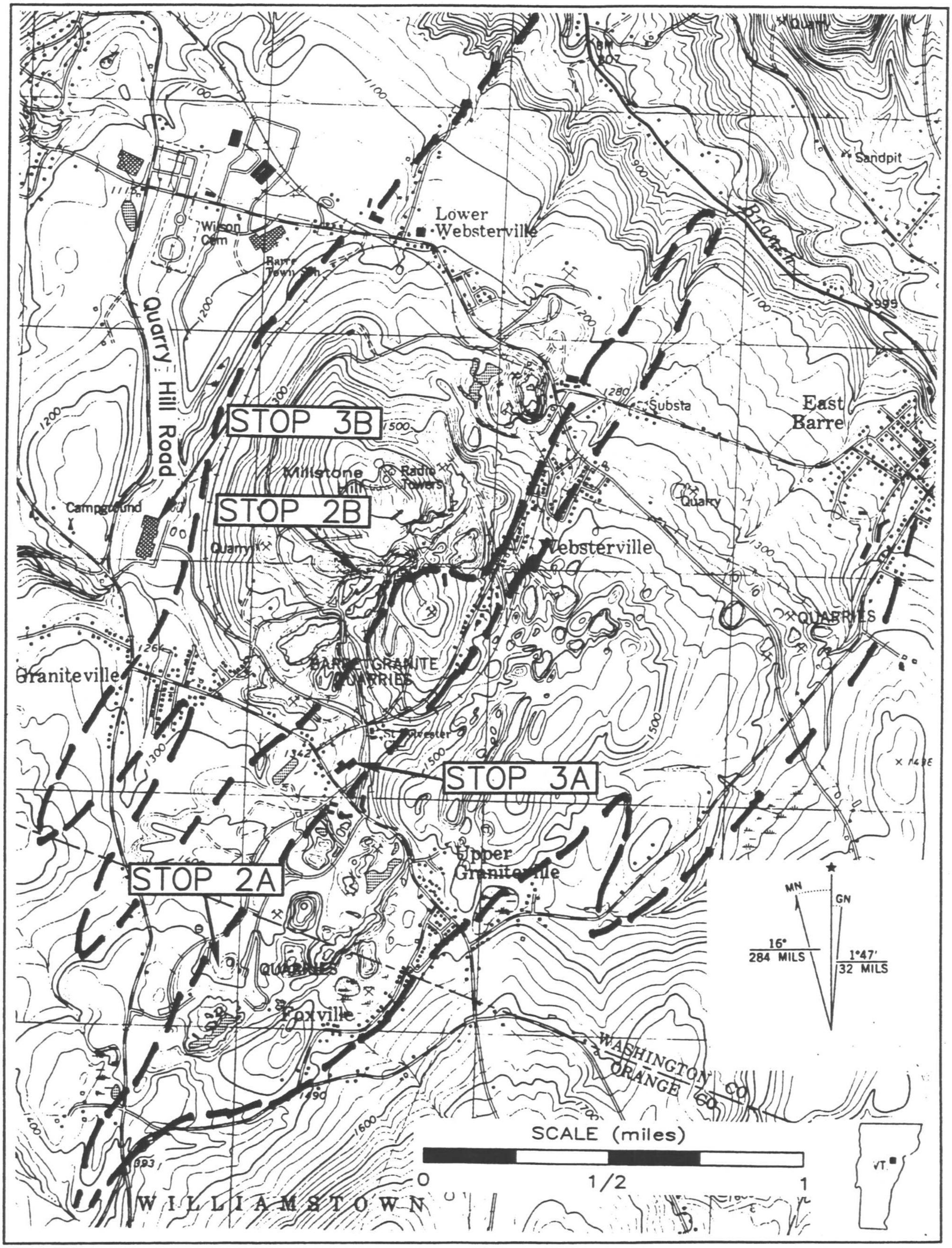

Figure 4 Locations of Stops 2 and 3 in the Barre area. Approximate granite contacts shown as dashed lines. Base map Barre East 7.5" USGS topographic quadrangle.

STOP 2. BARRE GRANITE QUARRIES. (1½ hours) We will visit two active quarries in Barre — the Adam Quarry and the Smith Quarry. Note the different styles of fracturing and the difference in grain size and color of the granite in the two quarries. Note also the different approaches to quarry development. See Figure 4 for the locations of the stops.

Stop 2A. Adam Quarry. The Adam Quarry is located in the southern end of the eastern limb of the Barre pluton. The quarry has been developed as a drive-in operation, and mobile loaders rather than derricks are used to move blocks. Pike Industries has an agreement with Rock of Ages to drill and blast areas for crushed stone in the upper edges of the quarry where the rock is too fractured to quarry as dimension stone. At the production levels, the structure of the granite is well displayed. There is a prominent set of NE striking joints that dip about 45° to the SE. Sheeting is not well developed in this quarry. The waste factor in this quarry is relatively high due to the fractures, but the granite is fine grained and dark blue grey in color, making it the premium grade material quarried in Barre. Most of the production from the Adam Quarry is used for monuments.

The Adam Quarry is a relatively new development, opened after an extensive diamond core drilling project in the early 1970's. It is adjacent to, and merges with, the older Pirie Quarry to the north.

Stop 2B. Smith Quarry. The Smith Quarry is located on the eastern flank of Millstone Hill just west of the septum of country rock that underlies the road between Graniteville and Websterville. We will visit the U-11 derrick location, located at the edge of the deepest part of the quarry. IF YOU ARE AFRAID OF HEIGHTS, STAY WELL BACK IN THE STORAGE YARD. There are a lot of interesting blocks in the storage yard, and the surface exposures around the edges, which had been quarried for crushed stone some years ago, exhibit thin sheeting, several large xenoliths, and a lamprophyre dike.

The U-11 derrick is the largest derrick in Barre and is rated for a 250-ton lifting capacity. All of the large blocks for surface plates, press rolls, mausoleum roofs, etc. are quarried at this location. The quarry is over 400 feet deep. Looking over the edge provides a mightily impressive view for those not afraid of heights.

Fracturing at the Smith Quarry is characterized by sheeting joints. High angle jointing is far less well developed here than at the Adam Quarry, making it possible to extract large blocks. The waste factor is lower in this quarry as well. But because the granite is slightly coarser grained, lighter colored, and more "wavy" (i.e., flow lines are more visible), it cannot command as high a price as that from the Adam Quarry.

LUNCH BREAK AND PIT STOP AT THE VISITOR CENTER. There are picnic tables and a "dining car" outside the Visitor Center. A food stand offers limited lunch materials and soft drinks. The Visitor Center

has some nice displays and granite souvenirs for sale.

0.0 0.0 Reset mileage after lunch. Turn left (northwest) out of the Visitor Center parking lot onto Graniteville Road.

0.1 0.1 Turn right into driveway at Saw Plant. Park near first door.

STOP 3. ROCK OF AGES FABRICATION PLANTS. (1½ hours) There are three plant buildings that we will visit — the saw plant, the Craftsman Center, and the Press Roll Fabrication Plant.

Stop 3A. Saw Plant. In the Saw Plant, quarry blocks are slabbed using computerized diamond saws with blades exceeding 10 feet in diameter and beds that can hold several blocks. The diamond saws operate 24-hours a day and cut about 15 square feet per hour. Older multi-strand wire saws are still used for certain work. The wire saw uses twisted wire traveling at high speed in a continuous loop. Silicon carbide abrasive drawn through the stone by the twisted wire does the actual cutting.

On leaving Saw Plant, turn right on Graniteville Road.

0.6 0.7 Bear right (north) at fork, continue on Graniteville Road.

0.3 1.0 At fork in road, take hard right into driveway of Craftsman Center. Park in lot at front of building.

Stop 3B. Craftsman Center and Press Roll Fabrication Plant. Rock of Ages' Craftsman Center produces cemetery monuments, mausoleums, sculptures, and industrial products. We will walk through the plant floor and see monuments and mausoleums in various stages production. Much of the work is automated and much is standardized, but there is still a large amount of handwork that goes on. In the Industrial Products area, we will see evidence of the precision work that goes into the fabrication of surface plates, machine bases, and other technical assemblies. Granite press rolls for the paper mill industry are fabricated in a separate facility to the rear of the Craftsman Center. A 200-ton overhead crane, giant lathes and related equipment are used to produce the press rolls.

On leaving Craftsman Center driveway, turn right (north) on Quarry Hill Road. It's about a ½-hour drive to the next stop.

1.1 2.1 Wilson Cemetery on right.

0.7 2.8 Railroad crossing.

0.6. 3.4 Junction Route 14. Turn right (north) on Route 14.

0.9 4.3 Junction Route 302. Continue on Route 14/Route 302 by turning left (west) on Main Street.

1.0 5.3 Routes 302 and 14 separate at a light, corner of Maple Avenue. Continue on Route 14 by turning right (north) on Maple Avenue.

0.8 6.1 Turn left into the main entrance of Hope Cemetery.

STOP 4. HOPE CEMETERY. (20 minutes) We will spend a few minutes visiting some of the expressive granite memorials in the Hope Cemetery. If time permits, we will stop to examine a few monuments. Otherwise, we will motorcade slowly through the cemetery, and you can revisit at a later time, if you wish. Enter the cemetery through its main gate and follow the entrance road to a T. Turn left, and wind around toward the south. Take a short loop road near a fountain, pause to admire the nearby Corti memorial, and loop back to the left toward the cemetery exit.

All of the cemeteries in the Barre-Montpelier area contain striking granite memorials, but the Hope Cemetery is considered the showcase. Most of the monuments in Hope Cemetery use Barre Granite, and steeled finishes are more popular than polished. The workmanship is simply outstanding.

There are too many interesting monuments to describe in detail. However, one of the most beautiful monuments is that of Elia Corti, a noted Barre sculptor and apparently an innocent bystander who died when someone fired into the crowd during a scuffle between Socialists and Anarchists in front of Barre's Socialist Block in 1903 (Clarke, 1989). His monument was carved by his brother and brother-in-law, and contains a full-size likeness with his stone working tools resting near his feet.

0.5± 6.6 Exit the Hope Cemetery through the main gate. Turn left (north) on Route 14.

4.2 10.8 Join US Route 2 west for a short distance. Take on Route 14 when US Route 2 splits off.

3.4 14.2 Junction Route 214 in North Montpelier. Stay on Route 14.

4.8 19.0 General Store in East Calais. Stay on Route 14.

Stay on Route 14 through South Woodbury. (Don't have mileage)

6.1 25.1 Woodbury. Turn right (east) on Cabot Road in Woodbury.

0.3 25.4 Pavement ends.

0.4. 25.8 Turn left (north) on unmarked gravel road.

0.8 26.6 Take sharp switchback left up the hill.

0.3. 26.9 Gate.

0.6 27.5 **STOP 5. WOODBURY GRANITE QUARRY.** (30 minutes, location shown in Figure 5) The Woodbury Quarry is another drive-in operation using mobile lifting equipment. Sheeting fractures and primary igneous flow are well displayed in the quarry faces. Very few high angle joints disrupt the quarry, and there is very little waste in this operation.

End of field trip.

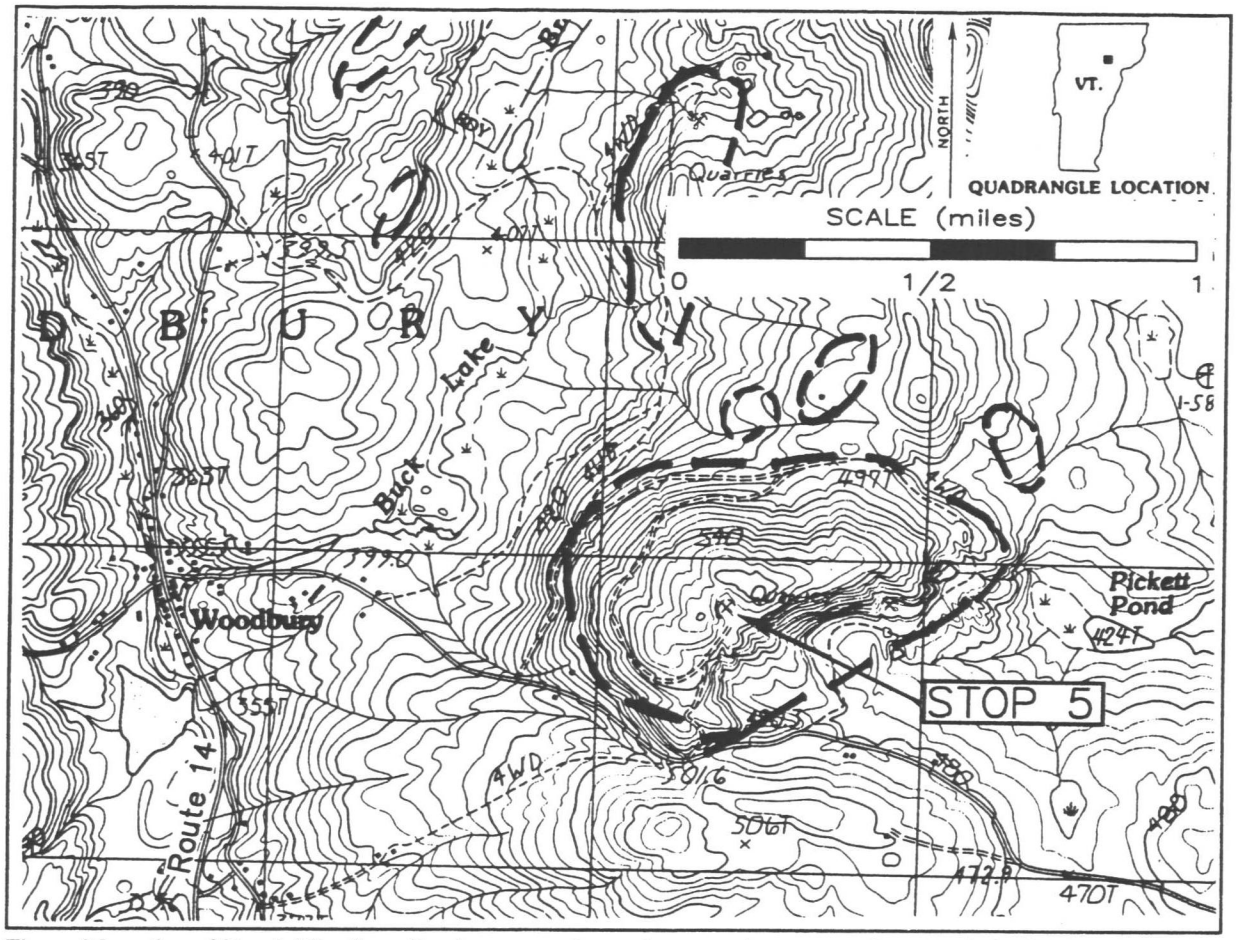

Figure 5 Location of Stop 5, Woodbury Granite quarry. Approximate granite contacts shown as dashed lines. Base map Woodbury 7.5" USGS topographic quadrangle.

To return to the Pico Peak Ski Area for the Banquet, follow the quarry road back out to Cabot Road, turn right (west) returning to Route 14, and turn left (south). Retrace your steps along Route 14 south to Barre. At the light at the south end of Maple Avenue in Barre (junction with US Route 302 and Route 62), go straight on Route 62 and follow it to I-89. Take I-89 south to Exit 3 (Bethel). Go right (west) on Route 107. Stay on Route 107 approximately 13 miles to Route 100. Take Route 100 left (south) approximately 11 miles to US Route 4. Go right (west) on US Route 4 a few miles to the Pico Peak Ski Area.

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Trip B5:
Environmental Hydrogeology: A Two-part Fieldtrip.
1) Disposal of Treated Sewage Effluent by Spray Irrigation on a Forested Hillside;
2) Air Sparging and Soil Vapor Extraction at a Gasoline Contamination Site.

Chris Aldrich, Heindel and Noyes, P.O. Box 64709, Burlington, VT 05402-4709

1) Disposal of Treated Sewage Effluent by Spray Irrigation on a Forested Hillside

Several of Vermont's ski areas make extensive use of the deep well-drained soils developed on glacial tills on moderately sloping forested hillsides to treat and dispose of hundreds of thousands of gallons per day of sewage. Secondary-treated disinfected sewage effluent is sprayed during most months of the year at low application rates on hillsides especially selected to have deep and permeable A and B horizons, moderate slopes, no bedrock outcrops, stable vegetation, and large setback distances to surface waters. These spray areas serve the dual purpose of adding a third degree of treatment to the effluent, and dispersing the wastewater in an environmentally benign manner. Personnel from Stratton Mountain Corporation will assist the trip leader in providing a tour of the entire wastewater treatment and disposal system at a major eastern ski area, from the secondary treatment plant to the operating spray irrigation area. The results of many years of monitoring of groundwater, surface water, and aquatic biota at and below the spray site will be presented.

2) Air Sparging and Soil Extraction at a Gasoline Contamination Site

The site of a subsurface leak of several thousand gallons of gasoline will be visited to inspect air sparging and soil vapor extraction systems installed and operated by the trip leader. The site will provide the backdrop for a discussion of how to lay out air sparging and vapor extraction systems, and what they can (and cannot) be expected to accomplish. Using this active remediation site, the trip leader will demonstrate the use of various field instruments (photo-ionizing detector (PID), and meters to measure oxygen, carbon dioxide, methane, air velocity, and vacuum). These real-time measurements allow remediation specialists to fine-tune the extraction systems in response to changing conditions, producing maximum remediation results while minimizing energy and labor costs. Monitoring data will be presented and evaluated.

TRIP B-6. STRATIGRAPHY, GEOCHRONOLOGY, AND TECTONIC EVOLUTION OF THE BASEMENT AND COVER ROCKS OF THE CHESTER AND ATHENS DOMES

by

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ABSTRACT

Laurentian basement rocks of the Mount Holly Complex (~1.4 to 1.1 Ga) and the post-Grenville Bull Hill Gneiss (955 Ga) in the Chester and Athens domes are overlain unconformably by Late Proterozoic and Cambrian rocks that are only locally preserved beneath a more extensive Ordovician tectonic cover that was inherited from Taconian thrust faulting and crustal accretion of an island-arc terrane. Calc-alkaline intrusive and volcanic rocks from the Moretown and Cram Hill Formations in this tectonic cover have U-Pb zircon ages of 496 Ma to 462 Ma. These ages and the internal structure of the cover rocks indicate that the Moretown Formation, ultramafic rocks, and the Cram Hill Formation were tectonically assembled outboard of Laurentia and then emplaced later on the Laurentian basement in the Taconic Orogeny. New 1:24,000 mapping of the core of the domes and the cover rocks shows that the Acadian deformation of this pre-assembled thrust stack involved multiple, non-coaxial shortening in three or four Acadian deformational events, at pressures of 7 to 10 Kbars and temperatures of 500° to 600°C. Detailed mapping of strain markers within the Mount Holly Complex in the core of the domes and the structures in the Silurian and Devonian rocks do not support the previous concepts of a regional nappe-stage Acadian folding event. In fact the map data rule out that interpretation.

This trip will focus primarily on the detailed internal structure of the domes and less on the stratigraphy of the Cambrian through Ordovician tectonic cover. The cover rocks are described more thoroughly in Trip C-6. This trip includes stops in the Ludlow, Cavendish, Chester, Andover, and Saxtons River quadrangles. The topics discussed overlap with Sunday Trip C-6, and to some extent with Friday Trip A-6.

INTRODUCTION

Since 1992 we have completed new 1:24,000 bedrock geologic maps of the rest of the Chester and Athens domes not covered in our 1992 NEIGC (Ratcliffe and others, 1992) field trip (see figure 1 and index map in explanation). That trip presented a detailed discussion of the stratigraphy and structure of the southern part of the Athens dome, and the Rayponda and Sadawga domes to the south. That information will not be repeated here. Preliminary geologic maps are now available for all of the Chester and Athens domes and much of the surrounding area. The new maps released since 1992 or not covered on the 1992 trip include Mount Holly (Ratcliffe, 1992), Mount Holly and Ludlow quadrangles (Walsh and others, 1994), Plymouth (Walsh and Ratcliffe, 1994a, b), Cavendish (Ratcliffe, 1995a), Chester (Ratcliffe, 1995b, c), Andover (Ratcliffe, 1996a, b), Saxtons River (Ratcliffe and Armstrong, 1995, 1996) and Newfane and Brattleboro (Armstrong, 1994). In addition preliminary maps are available for adjacent areas. These quadrangle maps are the Weston (Ratcliffe and Burton, 1996), Springfield and Mount Ascutney (Walsh and others, 1996a, b), Bellows Falls (Armstrong, 1997), and Rochester (Walsh and Falta, 1996a, b) quadrangles. The following standard USGS maps are published or are in press, Woodford (Burton, 1991), Mount Snow and Readsboro (Ratcliffe, 1993), Jamaica and Townshend (Ratcliffe, 1997), West Dover and Jacksonville (Ratcliffe and Armstrong, in press), Cavendish (Ratcliffe, in press) and Chester (Ratcliffe, in press).

Field mapping is completed or will be by the end of the 1997 field season of these additional quadrangles: Chittenden, Pico Peak, Hartland, Mount Carmel, Rutland, Wallingford, Danby, Peru, Londonderry, Walpole, and Killington Peak.

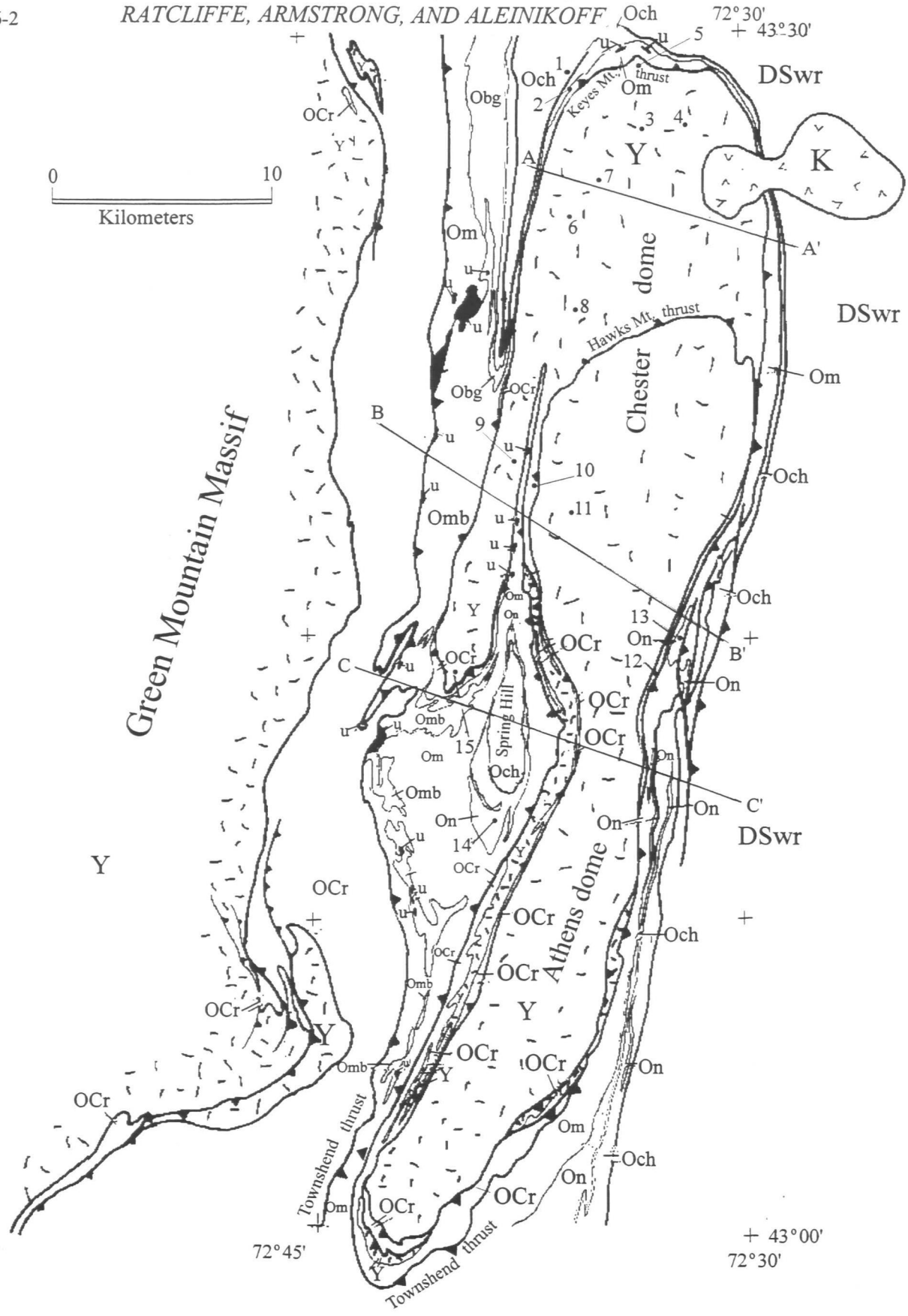


Figure 1. Generalized geologic map of the Chester and Athens domes showing major lithotectonic units; Laurentian basement of the Mount Holly Complex (Y), and tectonic cover of the domes including the Moretown and Cram Hill Formation and Rowe Schist above the Keyes Mountain and Townshend thrusts. For sources see index map and text. Section lines approximately located see figure 4 for more detailed map.

EXPLANATION B-6 -- Figure 1

Cretaceous

K

Alkaline intrusive rocks at Mount Ascutney

Devonian and Silurian

DSwr

Waits River and Northfield Formations including volcanogenic members of the Waits River Formation (Standing Pond volcanics of Doll and others, 1961)

Och
Obg
On

Cram Hill Formation and igneous rocks of the North River Igneous Suite (Ont) and Barnard Gneiss (Obg) of Richardson (1924)

Om
Omb

Ordovician and Cambrian

Moretown Formation including Whetstone Hill member of Missisquoi Formation of Doll and others (1961) Omb rusty schist member

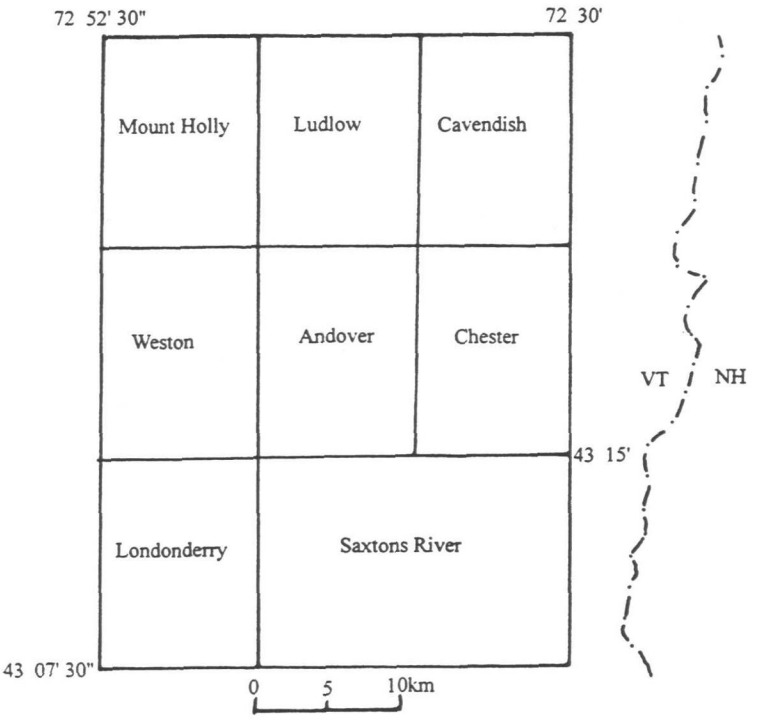
Oer

Hoosac, Pinney Hollow, Ottauquechee, Stowe and Rowe Formations undivided

Middle Proterozoic

Y

Core gneisses and schists of the Green Mountains massif and Chester and Athens domes (including units of the Mount Holly Complex, Cardinal Brook Intrusive Suite and Cavendish Formation



U

Ultramafic rocks

thrust fault teeth on upper plate

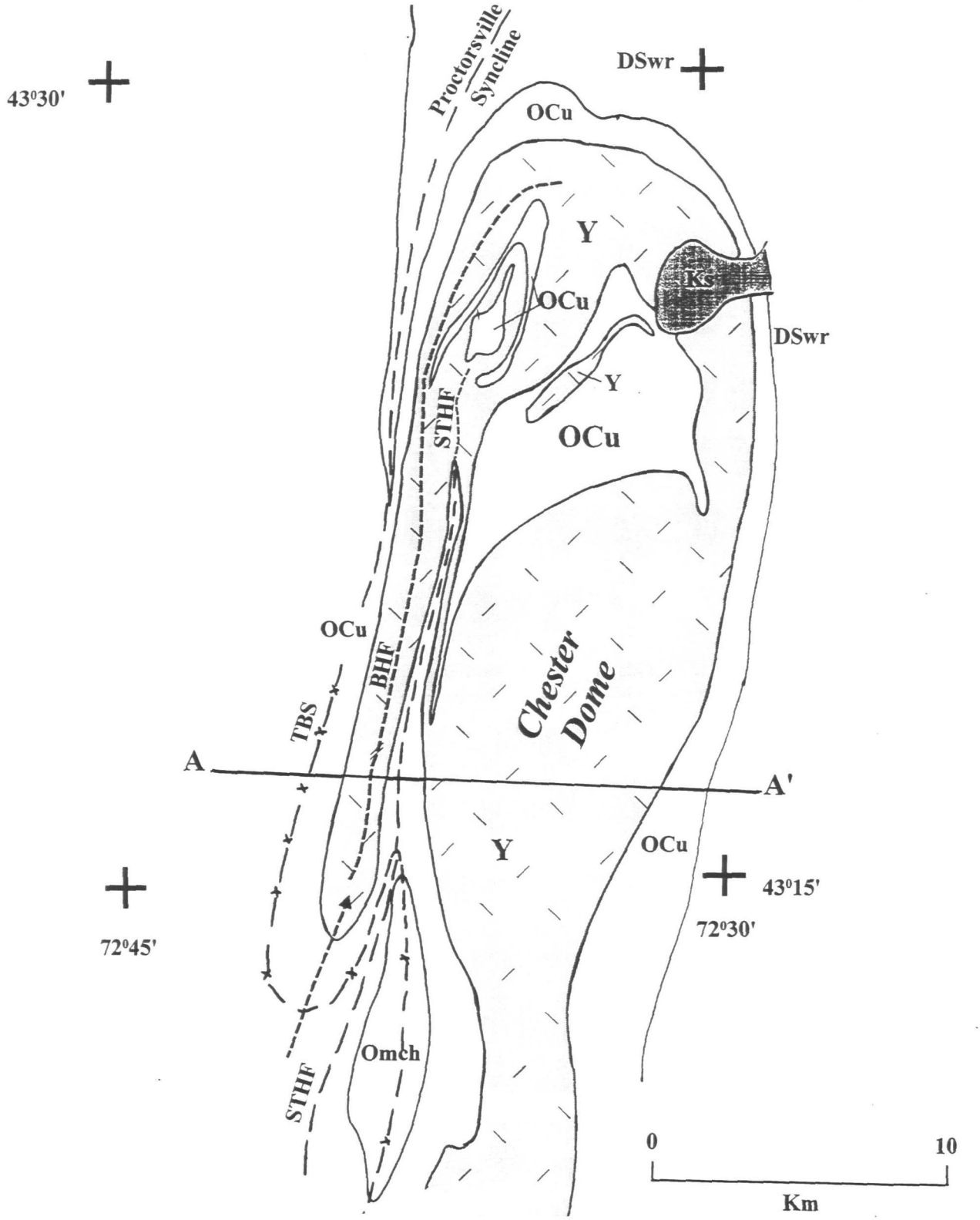


Figure 2. Schematic map of the Chester dome after Doll and others (1961) and Thompson and others (1993) showing axial traces of major Acadian folds in interpretation of Doll and others (1961) along section A-A' (figure 3). Rocks of Mount Holly Complex (Y), Ordovician and Cambrian cover rocks undifferentiated (O€u) (here including Cavendish Formation at Hawks Mountain and at Star Hill in cover sequence), Ordovician Cram Hill Formation (Omch) shown only at Spring Hill, Cretaceous intrusive at Little Ascutney Mountain (Ks). Fold axial traces in interpretation of Doll and others (1961) nappe stage Acadian folds: Townshend-Browington syncline (TBS), Butternut Hill fold (BHF) and Star Hill fold (STHF).

The preliminary and simplified digital maps cited above are available in diskette or paper format from the Office of the State Geologist, Department of Natural Resources, 103 South Main Street, Waterbury, VT 05671-0301. The nondigital USGS Open-File maps are available from U.S. Geological Survey, Information Services, Box 25286, Federal Center, Denver, CO 80225.

As a result of our mapping and that of our colleagues, new 1:24,000 geologic map coverage of virtually all of the Green Mountain massif, Eastern Cover Sequence and the Chester, Athens, Rayponda and Sadawaga domes has been completed. In addition one of us (JNA) has been conducting a program of conventional U-Pb and Ion Microprobe dating of zircon in support of the field mapping efforts. New geochronologic data from the basement gneisses of the Green Mountain massif and core rocks of the Chester and Athens domes, in conjunction with geochemical data, has recently established the age of the Laurentian basement rocks ranges from about 1400 Ma to 955 Ma. Our new U-Pb conventional and SHRIMP ages are given in tables 1 and 2. Additional emphasis has been placed on dating metavolcanic and metaintrusive rocks in the cover sequence. These data will be discussed on this trip and are presented in the text for, Trip C-6.

PREVIOUS WORK

The bedrock geology of the Chester and Athens domes was originally mapped by J.B. Thompson, Jr. (1950) and by J.L. Rosenfeld (1954). Their data were incorporated into the Centennial Geologic Map of Vermont (Doll and others, 1961) at a scale of 1:250,000. More recent work by them was used in compilation of the Glens Falls 1°x2° quadrangle (Thompson and others, 1990) at a scale of 1:250,000, and by Thompson and others (1993) at a scale of 1:125,000. The basic elements of the geology have not changed greatly since Thompson's and Rosenfeld's work in 1950 and 1954, until recently.

A recently published map of the Chester-Athens dome and surrounding areas by Thompson, Rosenfeld and Chamberlain (1993) shows considerable modification of the location of some contacts in the Saxtons River quadrangle from those drawn previously by Rosenfeld (1954) and by Doll and others (1961). The principal modifications include greater definition in the cover rocks northwest of Grafton and reassignment of a belt of dark biotite schist in the Round Mountain area (in the Andover quadrangle) from the Moretown or Cram Hill Formation of their previous versions to the Ottauquechee Formation. We do not agree with this assignment and believe that previous correlation of these rocks with the Cram Hill or Moretown Formation as shown on Doll and others (1961) is more correct. In addition, Thompson, Rosenfeld, and Chamberlain (1993) have mapped a loosely defined unit of slabby gneisses of uncertain age near the margin of the dome. They show these rocks as forming a belt approximately ½ km wide along the southwestern border of the dome at the southern boundary of the Saxtons River quadrangle. In the southern part of the Chester dome, at the northern border of Saxtons River quadrangle, and extending to a point approximately 4 km south, they show a south-closing belt of Tyson Dolomite (interpreted by them to be Cambrian and or Late Proterozoic) which forms the base of their slabby gneiss unit These slabby gneisses coincide with mylonitic varieties of the Mount Holly Complex and of the Bull Hill Gneiss as we map them, as well as local occurrences of rocks of the Hoosac Formation. Elsewhere their slabby gneiss-Mount Holly contact appears to coincide with what we map as the Bull Hill Gneiss-Mount Holly Complex contact, for example along the eastern margin of the Chester and Athens domes (fig. 4).

The simplified geology of the Chester and Athens domes as summarized from Thompson and others (1993) is shown in figure 2. A cross section across the Chester dome in figure 3 (after E-E' of Doll and others, 1961) shows that Middle Proterozoic core rocks and lower and middle Paleozoic cover rocks are folded in a large recumbent, west-verging nappe, the Butternut Hill fold, and an ancillary recumbent syncline, the Star Hill fold. According to models of Thompson (1950) and of Rosenfeld (1968) later Acadian deformation, accompanied by the buoyant uprising of lighter core rocks, produced upwarping of the dome and adjacent tightly pinched synclines.

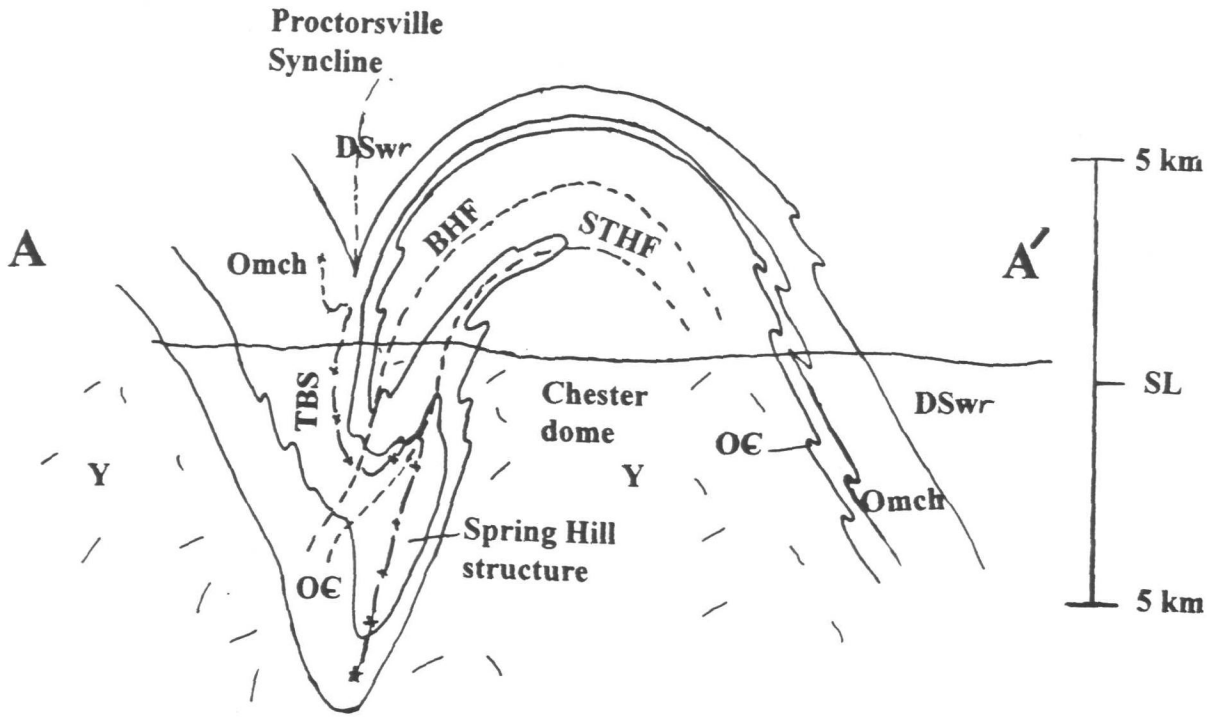


Figure 3. Schematic section along A-A' in figure 2 after Doll and others (1961) from east flank of Green Mountain massif on west across Chester dome showing nappe stage Butternut Hill, Star Hill folds and the Townshend-Browington syncline in interpretation of Doll and others (1961), and of Downie (1982). Units as in figure 2, except O€u of figure 2, is subdivided to show the combined Moretown and Cram Hill Formations (Omch) as a separate unit. Section extends farther west than A-A' line on figure 2. In interpretation of Thompson and others (1993) both the Cavendish (outer belt of cover rocks at Star Hill) and the Pinney Hollow Formation (inner belt of cover at Star Hill) are separated from basement in the Star Hill fold (STHF) by megaboudinage after Downie (1982), in an interpretation similar to that used for the isolation of rocks in the Spring Hill structure as proposed by Rosenfeld (1954). The interpretations shown here are not substantiated by our mapping see figures 4 and 5 for our interpretation.

Similar cross sections in Thompson (1950) and Thompson and others (1986) cite the northern plunge of the Butternut Hill and Star Hill folds and use the rotation senses and plunges of minor folds to support the synformal and antiformal shape of the Butternut Hill and Star Hill folds, respectively.

The Star Hill fold of Thompson (1950) originally extended from near Grafton in the Saxtons River quadrangle to Star Hill (Doll and others, 1961) in the Cavendish quadrangle. Subsequent versions of the geology by A.B. Thompson, Lyttle, and Thompson (1977), J.B. Thompson, Jr., and others (1990), and J.B. Thompson, Jr., and others (1993) terminate the fold at Cavendish south of Star Hill and show isolated areas of cover rock at Star Hill due, in part, to more detailed mapping of the Star Hill area by Downie (1982). This version appears on figure 2.

Nisbet (1976) mapped the Keyes Mountain area north of Star Hill in detail and did reconnaissance studies of the Butternut Hill fold. He concluded that Thompson's use of rotation sense of minor folds in confirming the form of the major folds was not consistent and that many of these minor folds were the result of late interference folds; consequently, they could not be used uniquely to determine stratigraphy or structure.

In Downie's interpretation (1982, p. 269), the Hoosac and Pinney Hollow Formations occur as an inner cover in the core of the Star Hill fold (the inner belt of O$\in$u on fig. 2) following the earlier usage of Thompson (1950) and Doll and others (1961). According to Downie, these rocks are thought to be the emergence of the mantle of the dome exposed in a major antiform. A continuous belt of Mount Holly Complex (Y on fig. 2) encircles the core of the antiform; this belt was originally mapped as Hoosac Formation by J.B. Thompson, Jr. (1950). Beyond this belt Downie mapped calc-silicate gneiss, granofels, and schist tentatively as Hoosac Formation; this belt corresponds to the outermost belt of the Cavendish Formation at Star Hill (fig. 2) as shown by Doll and others (1961), shown as O$\in$u on figure 2. Rocks of the Mount Holly Complex again encircle the area. See the text discussion of Downie's work for Stop 6.

Purpose of present mapping

Previous geologic maps by Thompson (1950), Doll and others (1961), and Thompson and others (1990, 1993) do not show subdivisions of the Mount Holly Complex within the core of the Chester dome in the Cavendish, Chester, or Saxtons River quadrangles. Because the relationship of cover rocks to the basement rocks is critical to determining the structure of the dome, an effort was made to separate rock types within the core of the dome. Separating the rock types is necessary in order to evaluate whether the Cavendish Formation is a cover rock or part of the Mount Holly Complex.

Key questions that we have addressed include:

1. Are the core rocks of the dome comparable to Mount Holly rocks of the Green Mountain massif?
2. Are the schists of the Cavendish Formation part of the basement rocks or are they cover rocks assignable to the Late Proterozoic and Lower Cambrian?, Hoosac Formation?
3. Is the internal structure of the dome compatible with isoclinal and recumbent folding of the basement and cover into nappes?
4. Are the cover rocks east and west of the dome in depositional contact with the core rocks and is the stratigraphy intact, or are these rocks bounded by faults?

Major Lithotectonic Units

In very general terms, the rocks of the Chester and Athens domes can be assigned to the following lithotectonic assemblages following Stanley and Ratcliffe (1985):

(1) Rocks of the Laurentian plate. These include the Middle Proterozoic Mount Holly Complex and its immediate unconformable cover of the Tyson, Hoosac, and Plymouth Formations that together formed the basement and cover affected by ancient Late Proterozoic rifting to form the Laurentian passive continental margin in the Cambrian and Ordovician.

(2) Rocks of the late-rift and oceanic-drifting stage. These include the Pinney Hollow, Ottauquechee, and Stowe Formations, and Rowe Schist of southern Vermont. Although these rocks were deposited in a deepening and widening Iapetan ocean basin, sediment polarity was largely from the Laurentian margin and these rocks are not therefore truly exotic rocks. Their deposition probably spanned the transition from rift subsidence to ocean basin formation and drifting.

(3) Rocks of the oceanic arc and fore-arc, and accretionary prism. These include the rocks of the Moretown Formation and the Cram Hill Formation, and may include felsic and mafic volcanic or intrusive rocks of the Barnard Gneiss (of Richardson, 1924) and the North River Igneous Suite of Armstrong (1994). These rocks and the contained ultramafic rocks are exotic to the Laurentian plate.

(4) Post-accretionary and late Taconian through post-Taconian metasedimentary and volcanic rocks. These may include some rocks in the uppermost part of the Cram Hill Formation and the Silurian and Devonian rocks of the Waits River Formation.

In addition minor Acadian granite dikes accompanied Acadian doming and were affected by tightening of folds on the margins of the domes late in the Acadian orogeny. Abundant post-tectonic Cretaceous lamprophyre, diabase and syenitic dikes and plutons are related to a major episode of intraplate alkaline plutonism.

STRATIGRAPHY

The description of rock units, nomenclature, and symbology below is taken largely from the text and description of map units for the Chester and Cavendish quadrangles and should be used in conjunction with the Open-File geologic maps of the Cavendish (Ratcliffe, 1995a), Chester (Ratcliffe, 1995b, c), Andover (Ratcliffe, 1996a, b), and Saxtons River quadrangles (Ratcliffe and Armstrong, 1995, 1996) as no detailed map or text containing all the data discussed accompanies this guidebook article.

Rocks of the Chester and Athens areas range in age from Middle Proterozoic to Devonian. The core rocks of the domes, including the Cavendish Formation or the Hoosac Formation of Thompson and others (1993), on Hawks Mountain and at Star Hill are here all regarded as Middle Proterozoic (figs. 1 and 4). The core rocks include intrusive and layered gneisses of the Mount Holly Complex and the younger intrusive Bull Hill Gneiss of the Cardinal Brook Intrusive Suite (fig. 4).

A 0.5- to 0.75-km-thick section of cover rocks (fig. 1) overlies the Mount Holly Complex and the Bull Hill Gneiss on the eastern and western sides of the domes. The ages of the rocks and formation assignments, in this belt, are highly uncertain but they are thought to range from possibly Late Proterozoic to Early Devonian. The lower part of the section consists of schists, amphibolites, granofels, and metavolcanic or intrusive rocks previously assigned to the Hoosac, Pinney Hollow, and Ottauquechee Formations (shown undivided as O$\in$r on figure 1); the upper part consists of the overlying Moretown Formation (Om) and associated Barnard Gneiss (of Richardson, 1924) or North River Igneous Suite (of Armstrong, 1994) shown as Obg and On, respectively, on figure 1.

Previous maps (Thompson and others, 1993) show these cover rocks as continuous parallel units. The present mapping does not support that interpretation because the mapped units are neither continuous nor are they in a regular succession. Internal faulting and folding, as well as a major fault detachment, along the base of the section are indicated. The major part of the section is assigned to the Moretown Formation and interpreted to be in fault contact with the basement rocks in many areas.

The top of the exposed section consists of a sequence of interlayered greenstones, quartzites, coticule, and schist assigned to the Ordovician Cram Hill Formation (Och on figure 1). Intrusive rocks of the North River Igneous Suite (On) (of Armstrong, 1994) and metavolcanics of the Barnard Gneiss (Obg) of

Richardson (1924) are present in the Moretown and Cram Hill Formations. These intrusive and volcanic rocks have U-Pb and SHRIMP ages between 496±8 and 462±6 Ma (Table 2). Overlying these are rocks of the Northfield and Waits River Formations of Silurian and Devonian age (DSwr).

Cover rocks west of the dome in the Spring Hill fold correlate with the lower parts of the cover sequence east of the dome and include, locally, rocks of the Hoosac Formation and Rowe Schist shown collectively as OCr on figure 1.

Of the numerous Taconian and Acadian regional thrust faults, we have shown only a few on figure 1. The Hawks Mountain thrust roots in the Chester dome and separates the overlying Cavendish Formation from the underlying Baileys Mills Tonalite Gneiss (see discussion below).

The Keyes Mountain thrust places rocks of the lower part of the cover sequence on the Mount Holly Complex or Hoosac Formation (where present). The Townshend thrust places the very complexly folded and internally faulted Moretown Formation and its included ultramafic rocks on the lower part of the cover sequence or on the Mount Holly Complex. The Moretown Formation and some intrusive rocks of the North River Igneous Suite (of Armstrong, 1994) or of the Barnard Gneiss (of Richardson, 1924) were tectonically assembled as an accretionary wedge and arc intrusives before final emplacement on the Townshend thrust (see Trip C-6).

STRATIGRAPHY OF THE MOUNT HOLLY COMPLEX

Rocks of the Mount Holly Complex crop out within the core of the domes and on both flanks of the Spring Hill fold along the western edge of the domes. Some units in the Mount Holly Complex that we have mapped are shown in figures 4 and 5. The Mount Holly Complex consists of both paragneisses and intrusive rocks. Not all of the units described below appear in figures 4 and 5, but are shown on the published or Open-File maps of the Andover, Cavendish, Chester and Saxtons River quadrangles. The dominant rock type of the paragneiss section consists of well-layered biotite-quartz-plagioclase gneiss (Ybg) that is the host for a variety of other distinctive rocks, including rusty-weathering schists or gneisses (Yrs, Yrg) that are closely associated with thin layers of quartzite, calc-silicate rocks, and with marble (Ycs, Ym) most of which are to thin to be shown in figures 4 and 5. Abundant layers of amphibolite (Ya) can be mapped locally. The calc-silicate rocks and marbles (Ycs, Ym) and schists (Yrs, Yrg) are generally discontinuous, but locally continuous belts that are associated with very garnet-rich biotite gneiss or schist that is as garnetiferous as the Gassetts Schist Member of the Cavendish Formation can be mapped. One belt of such rocks extends along the eastern side of Chester dome from the North Springfield Reservoir (NSR on fig. 4) south almost to State Route 11, where it disappears in a belt of granite gneiss (Ygg).

A parallel belt of similar rocks (Yrg, Yrs, Ym, Ycs) to the east extends from the Black River at the eastern margin of the Chester quadrangle southwestward through Whitney Hill (WH on fig. 4) to the southern border of that map. Bull Hill gneiss occurs to the east. This belt of rocks has been mapped previously (Thompson and others, 1993; Doll and others, 1961) as the equivalent of the Cambrian Tyson Formation which is normally found near the base of the cover sequence in the Green Mountains. They therefore interpreted this belt of schist and calc-silicate as the base of their cover sequence. The present mapping indicates that these calc-silicate rocks and associated schists are all well within the Mount Holly Complex, and west of the Bull Hill Gneiss.

South of State Route 11, a very persistent but thin belt of rusty muscovite-chlorite schist and associated garnetiferous biotite gneiss (Yrg) extends southwestward to the southern border of the Chester quadrangle enters the Saxtons River quadrangle extends back to the northwest for about 3 km to a point north of State Route 103, at the western side of the Chester dome (fig. 4).

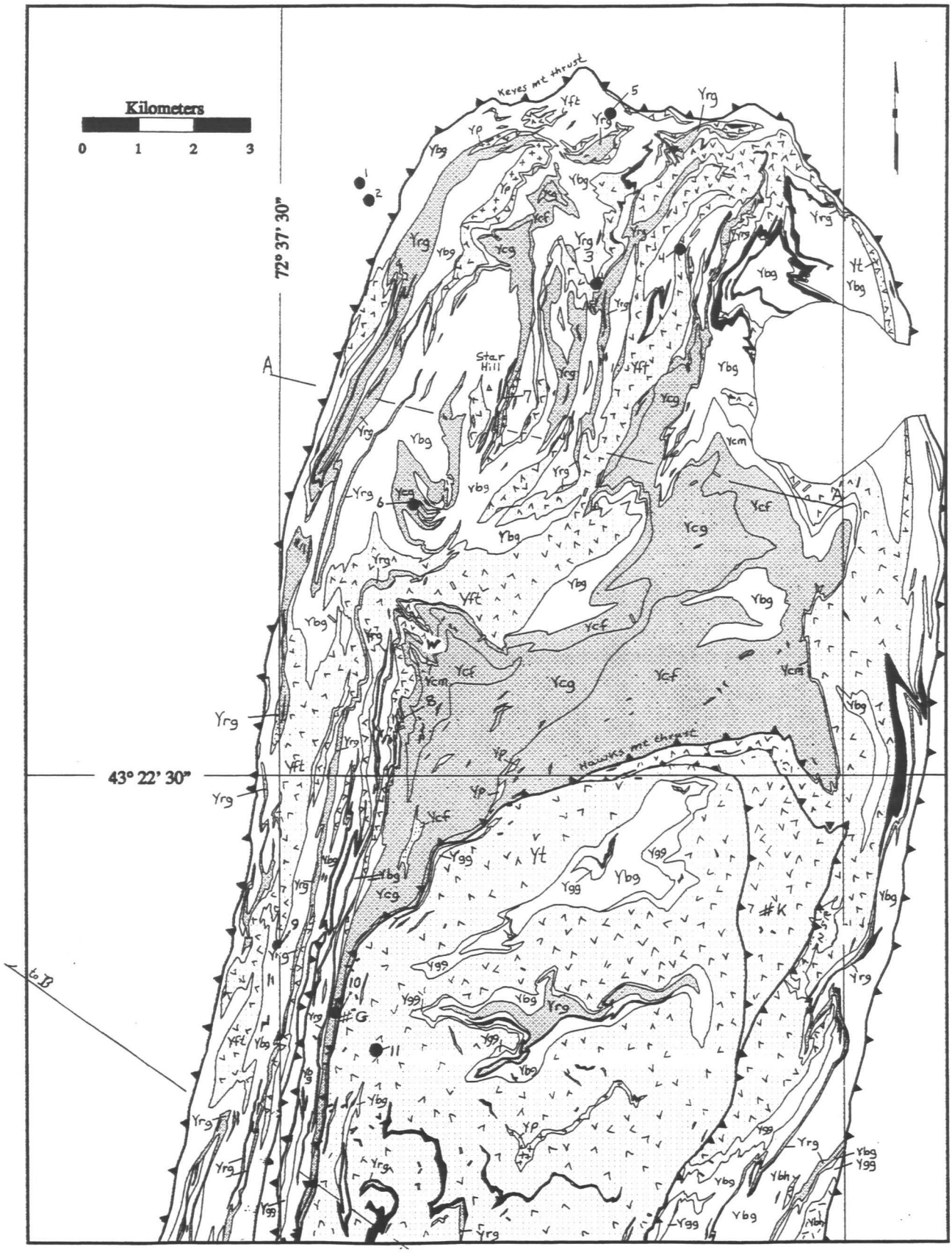


Figure 4. Generalized geologic maps of the core of the Chester and Athens domes showing subdivisions of the Mount Holly Complex (two maps overlap in coverage). For complete geologic maps and identification of all units see the Open File maps of the Chester, Cavendish, and Saxtons River quadrangles referred to in the introduction. Section lines identify approximate location of cross sections in figure 6. Abbreviations of place names: Chester (C), Grafton (GR), Whitney Hill (WH), Gassetts (G), Whitesville (W), Bull Hill (BH), Kendricks Corner (K), and North Springfield Reservoir (NSR).

EXPLANATION OF SYMBOLS ON CROSS SECTIONS B-6 -- Figure 6

DSw — Northfield and Waits River Formations

----------Unconformity----------

Och — Cram Hill Formation and associated volcanic rocks

Om — Moretown Formation and Intrusive rocks of the North River Igneous Suite, and Barnard Gneiss of Richardson (1929)

Omb — Black schist unit of Moretown Formation

OZu — Ultramafic Rocks

----------Undivided----------

OЄr — Pinney Hollow, Ottauquechee, Stowe Formations and Rowe Schist

ЄZh — Hoosac Formation

----------Unconformity----------

Ybh — Bull Hill Gneiss Member of Cardinal Brook Intrusive Suite

MOUNT HOLLY COMPLEX

Intrusive Rocks

Ygg — Granitic gneiss

Yt — Baileys Mills Tonalite Gneiss

Yft — Felchville Gneiss

Metasedimentary rocks of Mount Holly Complex

Ycg — Gassetts Schist Member

Ycf — Feldspathic granofels

Ycm — Calc-silicate rock and dolomite marble

Cavendish Formation

Ybg — Biotitic paragneiss, calc-silicate rock, quartzite and calcite or dolomite marble

Ya — Amphibolite (in part intrusive or extrusive)

Yrg — Rusty garnet schist and quartzite (contains garnet-rich schist like Gassetts Member of the Cavendish Formation)

Thrust fault teeth on upper plate (T) Taconian, (A) Acadian, (A?) composite age

B-6 -- Figure 4

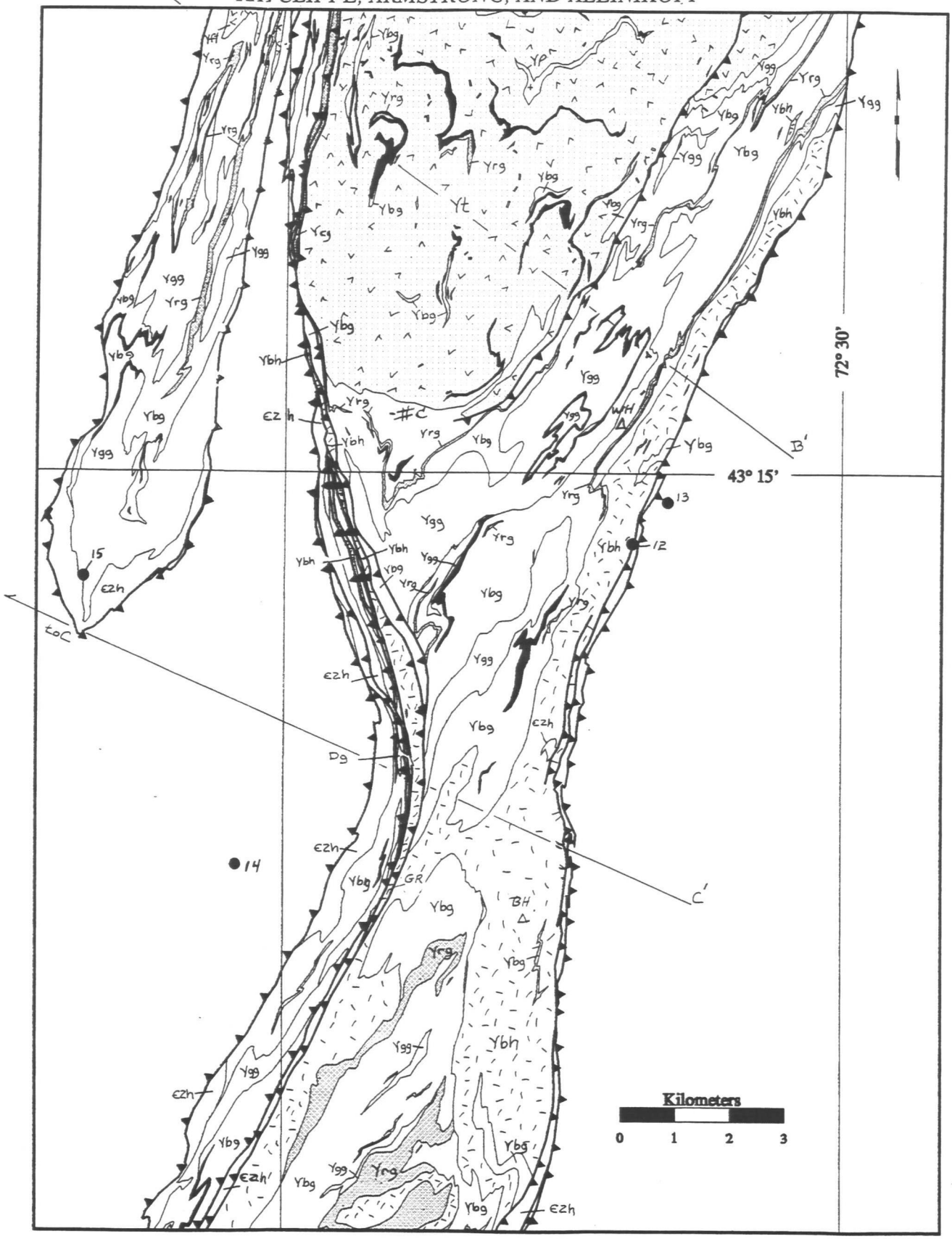


Figure 4. Generalized geologic maps of the core of the Chester and Athens domes showing subdivisions of the Mount Holly Complex (two maps overlap in coverage). For complete geologic maps and identification of all units see the Open File maps of the Chester, Cavendish, and Saxtons River quadrangles referred to in the introduction. Section lines identify approximate location of cross sections in figure 6. Abbreviations of place names: Chester (C), Grafton (GR), Whitney Hill (WH), Gassetts (G), Whitesville (W), Bull Hill (BH), Kendricks Corner (K), and North Springfield Reservoir (NSR).

In the northwestern part of the Chester quadrangle and adjacent Cavendish quadrangle many narrow belts of schist, calc-silicate rocks, and marble (Yrs, Ycs and Ym) are interlayered in biotitic gneiss (Ybg), and extend the length of these two maps parallel to other units in the Mount Holly Complex, including belts of the Cavendish Formation.

A large area of kyanite-biotite-muscovite-garnet-plagioclase quartz schist (Yrg on fig. 4), almost 4 km long, is present north of State Route 10 in the center of the Chester quadrangle. The schists within this belt, as elsewhere, that are mapped as (Yrg) in the Mount Holly Complex of our reports closely resemble schist within the Cavendish Formation (Ycg, Ycfs) (see section on Cavendish below).

Cavendish Formation of the Mount Holly Complex (Ratcliffe, in press)

Cavendish Formation at Hawks Mountain

The name Cavendish Schist was first applied by Richardson (1929) to the large mass of biotite-quartz-plagioclase schist and more lustrous garnet-kyanite-staurolite-biotite-muscovite-quartz schist exposed near the village of Gassetts and on Hawks Mountain in the towns of Chester and Baltimore. The more lustrous unit, well exposed in roadcuts on State Route 103, 0.5 km west of the village of Gassetts (G on fig. 4), was identified as Gassetts schist (Richardson, 1929). The term Gassetts schist has been informally used by many authors (for example, Thompson, 1975). A regional compilation by Doll and others (1961) did not show the Gassetts as a separate unit but used the name Cavendish Formation. The Cavendish as used by Doll and others (1961) included three discontinuous members; in ascending order they were an unnamed marble member, the Readsboro Schist Member (equivalent to the Gassetts Schist and Cavendish Schist of Richardson, 1929), and the Bull Hill Gneiss member.

In subsequent reports (A.B. Thompson, Lyttle, and Thompson, 1977; Thompson and others, 1990, 1993) the name Cavendish was abandoned, the name Hoosac Formation was used to refer to the schistose rocks, and the basal marble was retained as an unnamed member. The Bull Hill Gneiss was viewed as a metavolcanic member also of the Hoosac Formation.

The results of mapping of the Cavendish (Ratcliffe, 1995a, in press) and Chester (1995b, c, in press) quadrangles, and Saxtons River quadrangles (Ratcliffe and Armstrong, 1995), as well as regional studies (Ratcliffe, 1991) outside these areas have shown that the Bull Hill Gneiss is not interlayered with the Hoosac Formation but instead is intrusive into the Mount Holly Complex. Furthermore, rocks previously mapped as schists of the Cavendish Formation now appear to be gneiss and schist of the Mount Holly Complex. The Cavendish is now assigned to the Mount Holly Complex on the basis of 1) the intrusive contact of the Felchville Gneiss against the Cavendish and 2) the similarities between schistose units (Yrs, Yrg, Yrgt) and calc-silicate rocks or marbles (Ycs, Ym) of the Mount Holly Complex and comparable marble, calc-silicate rocks and schists of the Cavendish Formation as revised (Ratcliffe, in press). The Cavendish Formation includes three members (1) a discontinuous unnamed marble and calc-silicate member (Yccs or Ycm), (2) Gassetts Schist Member (Ycg), an aluminous garnet-rich schist, and (3) an unnamed feldspathic schist or granofels member (Ycfs).

These three members of the Cavendish Formation (Ratcliffe, in press) extend into the Chester quadrangle from the large mass on Hawks Mountain in the Cavendish quadrangle. The Gassetts Schist Member is locally bordered on the west and east by a belt of feldspathic schist and granofels (Ycfs) and locally by a narrow belt of marble and calc-silicate rock (Yccs). These marbles and calc-silicate rocks are discontinuous and the contact between biotite-quartz-plagioclase gneiss (Ybg) of the Mount Holly Complex, and the Gassetts Schist Member without the intervening schist and or granofels (Ycfs or Yccs), can be observed along the western slopes of Hawks Mountain about 1 km south of the northern border of the Chester quadrangle. In the large roadcuts west of Gassetts, a narrow zone of highly mylonitic phlogopite-calcite schist and calcite marble 0.5 m thick occurs at the contact between the Gassetts Schist Member (Ycg) and the Schist and Granofels Member (Ycfs).

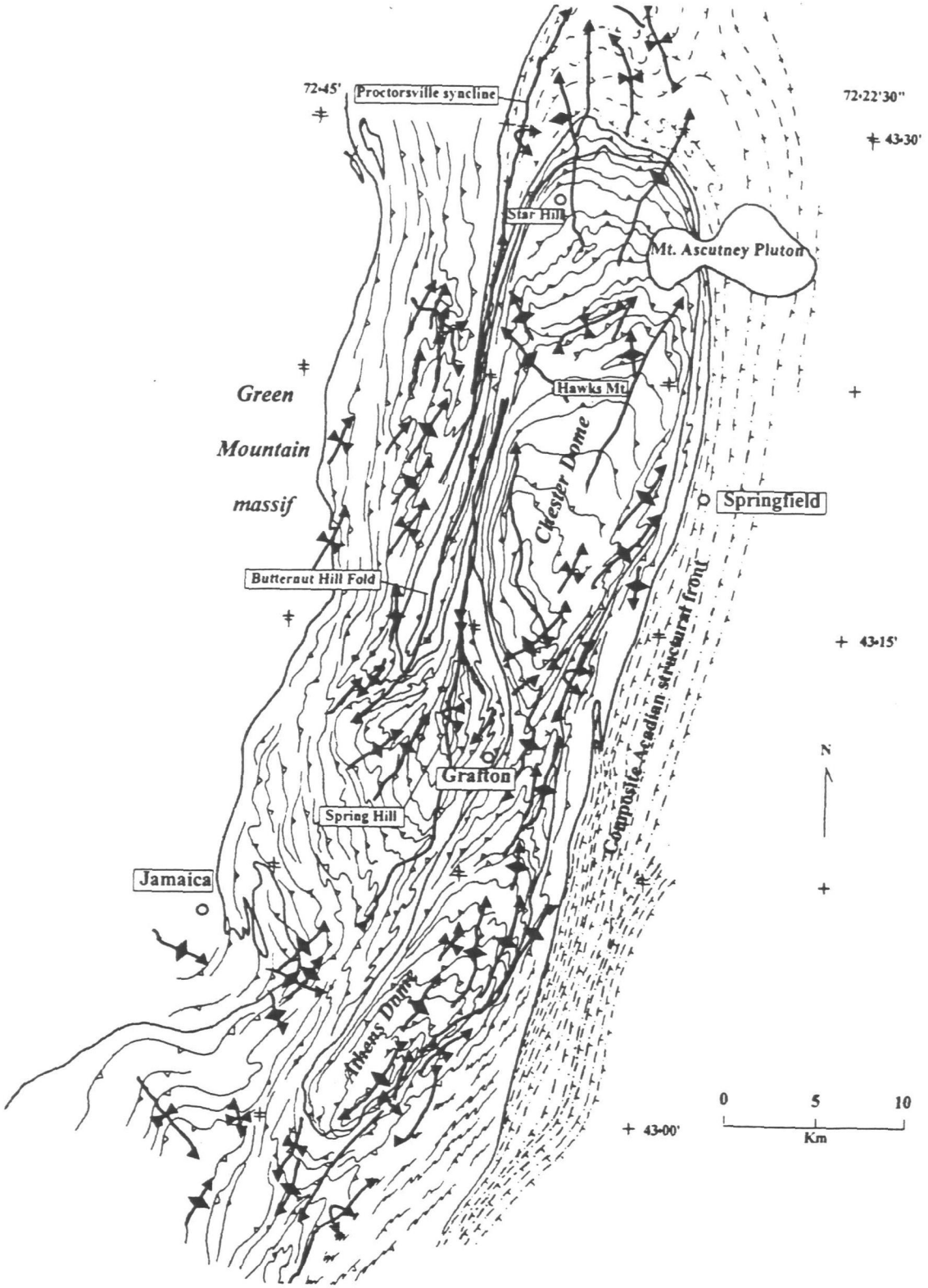


Figure 5. Schistosity form-line map of the Chester and Athens domes showing Taconian schistosity in pre-Silurian rocks and Acadian schistosity in Silurian and Devonian rocks. Small triangles or barbs show dip direction. Schistosity within the core of dome may be a composite of Acadian and Taconian structure. Bold axial traces show form and plunge of Acadian (F_3, F_4, folds responsible for formation of the dome. On the eastern side of the dome multiple phases of Acadian deformation overlap and result in accumulated strain in late Acadian folds that in general "young" to the east. Late stage flattening or tightening also occurred on the west flank of the dome in the Proctorsville syncline.

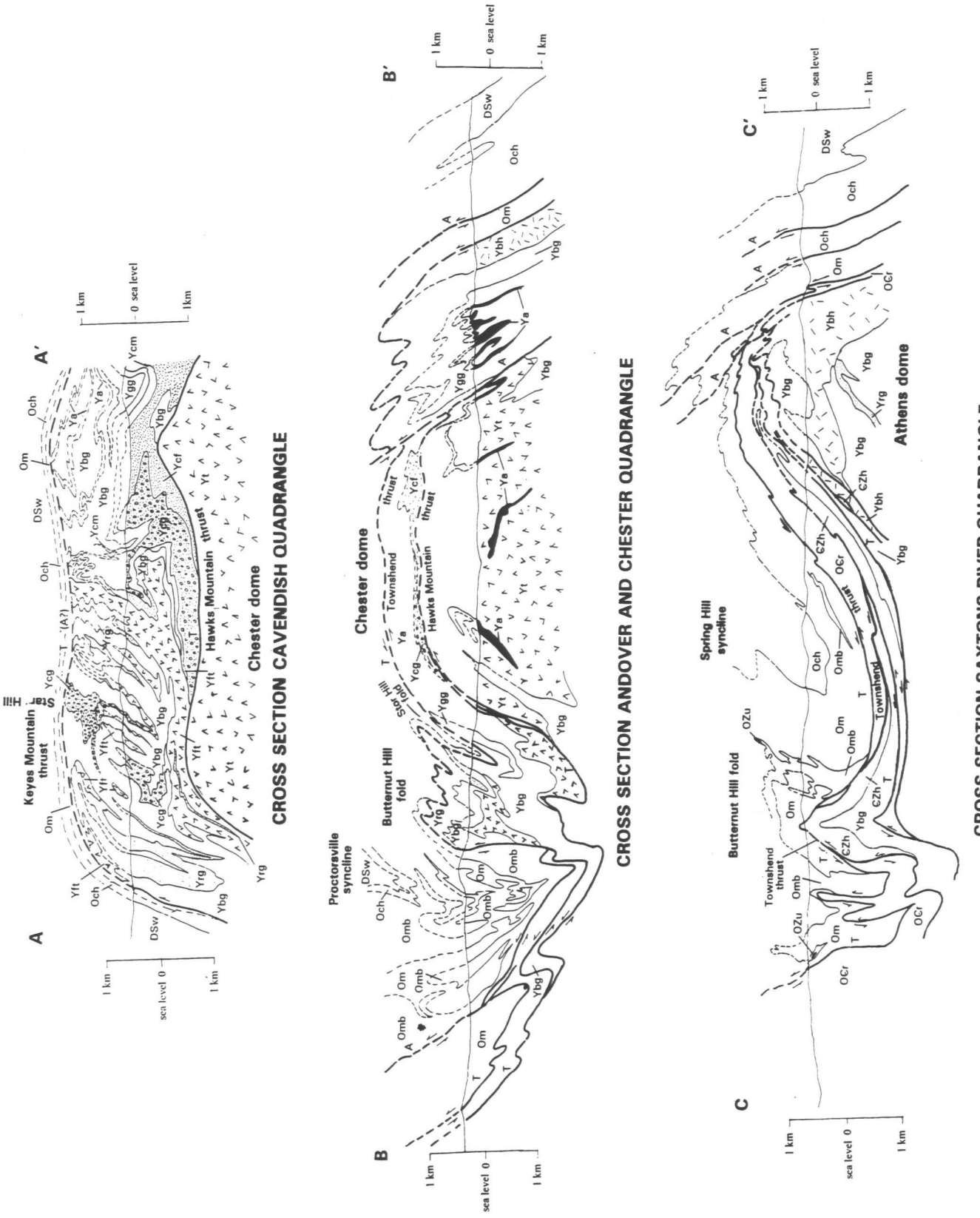


Figure 6. Cross sections of the Chester and Athens domes in the Cavendish, Chester and Saxtons River quadrangles (see text for sources of data). Approximate location of section lines shown in figure 1.

Contact relationships of Cavendish Formation in the Hawks Mountain Mass

Extensive exposures of contacts between feldspathic schist and marble in the Cavendish (Ycfs and Ycm) occur along the Black River in the Cavendish quadrangle northward from Cavendish Gorge to 60 m north of the electric powerplant (see Stop 8). Here, gray-to-white, fine-grained dolomite-calcite marble (±diopside, actinolite, and talc), rare vitreous white quartzite, and dark-gray albitic granofels are interbedded in a 3-m-thick band. The contacts are gradational, demonstrating clearly the interbedded nature of the feldspathic schist, granofels, marble, and quartzite. The calcite-dolomite-quartz mineral assemblage is dominant but occurs in conjunction with knotty appearing calc-silicate rocks containing masses of tremolite, chlorite, talc, and, less commonly, diopside and coarse phlogopite. Immediately above the dolomite and quartzite is a white aplitic gneiss unit (Yft), a local variant of the Felchville Gneiss. The dated sample C718 in table 1 comes from this locality. In the gorge area, the white aplitic gneiss, which is too thin to map at 1:24,000, is in contact with feldspathic schist, white dolomite, and more aluminous schist of the Cavendish (units Ycfs, Ycm, and Ycg). Along the west bank of the gorge, the aplite truncates and encloses several marble and calc-silicate gneiss units (Ycm, Yccs); the best exposures are near the vertical joint face of the west wall of the gorge from the dam and extending downstream about 50 m (Stop 8). On the east bank of the gorge, the same white aplitic gneiss is in contact with aluminous and feldspathic variants of the Cavendish. Farther downstream, a coarser grained variant of the Felchville (Yft) is in contact with the Cavendish Formation (Ycfs).

Along both banks downstream of the powerplant, a coarse diopside-hornblende scarn is present where the aplitic gneiss of the Felchville Gneiss is in contact with white dolomite marble. West of the Black River in the woods about 200 m south of the powerplant, calc-silicate gneiss merges into the aplite gneiss of the Felchville Gneiss. Elsewhere, this aplite contains distinct xenoliths of biotite gneiss (Ybg) of the Mount Holly Complex. The aplitic rock (Yft) is clearly an intrusive rock. Pegmatite pods occur throughout the Cavendish. The pegmatites (figs. 7A, C and D) are irregular, strongly foliated and retrograded rocks that contain chlorite-ilmenite replacement of once coarse biotite, hornblende or both. Acadian biotite-muscovite granites in contrast are straight-walled weakly deformed dikes of biotite-muscovite granite.

Similar relations can be seen near Whitesville, in the low hills northwest of the Black River, and in the Black River opposite the intersection of State Route 131 and Windy Hill Road in the Cavendish quadrangle. All along the contact between the Cavendish and Mount Holly Complex rocks, from the southern border of the Cavendish quadrangle northward to Hurricane Hill in the northeastern part of the Cavendish quadrangle, are tiny interlayered conformable lenses of calc-silicate gneiss, marble, and aluminous schist that closely resemble units of the Cavendish within the biotite-quartz-plagioclase gneiss (Ybg) and the aplitic gneiss (Yft) of the Felchville Gneiss of the Mount Holly Complex. The aplitic gneiss and trondhjemite gneiss (Yft) of the Felchville intrude all other paragneiss units of the Mount Holly Complex, as well as the Cavendish Formation. Because Cavendish units are found within the Mount Holly and vice versa and Felchville units are found in both, it seems logical that the entire mass of Cavendish should be considered part of the Mount Holly Complex.

A series of calc-silicate gneisses, marbles, and aluminous schist layers within the Mount Holly Complex south of Cavendish Gorge occur west of and parallel to the Cavendish contact. These units include lenses of aplitic to granodioritic gneiss (Yft) and extend southward into the Duttonsville Gulf area in the Chester quadrangle.

Previously, several layers of calc-silicate gneiss and calcite-dolomite marble exposed in Cavendish Gorge were mapped as a single unit at the base of the Cavendish (Doll and others, 1961); our work indicates this clearly is not the case. The calc-silicate gneiss exposures at Gassetts studied by A.B.

Thompson (1975), the marble sampled by Sheppard and Schwarz (1970) in Duttonsville Gulf, and the marble and calc-silicate gneiss exposures in Cavendish Gorge are probably not the same layer, but, instead are different layers within the Mount Holly Complex.

Hawks Mountain thrust and the Lower Contact of the Cavendish Formation

The eastern and lower contact of the Cavendish Formation is well exposed in a series of cliffs from a point west of Baileys Mills in the Chester quadrangle northward, on the east- and south-facing slopes of Hawks Mountain in the Chester and Cavendish Quadrangles. All along this contact, the Cavendish Formation has a transitional contact with the Baileys Mills Tonalitic Gneiss of the Mount Holly Complex (Yt) and with a biotitic-feldspathic plagioclase-augen mylonite gneiss (Yta and Ytm of the Chester quadrangle report) into highly sheared mylonitic schist of the Cavendish (Ycfs). The transition zone is 50 to 200 m thick (see figs. 6 and 7).

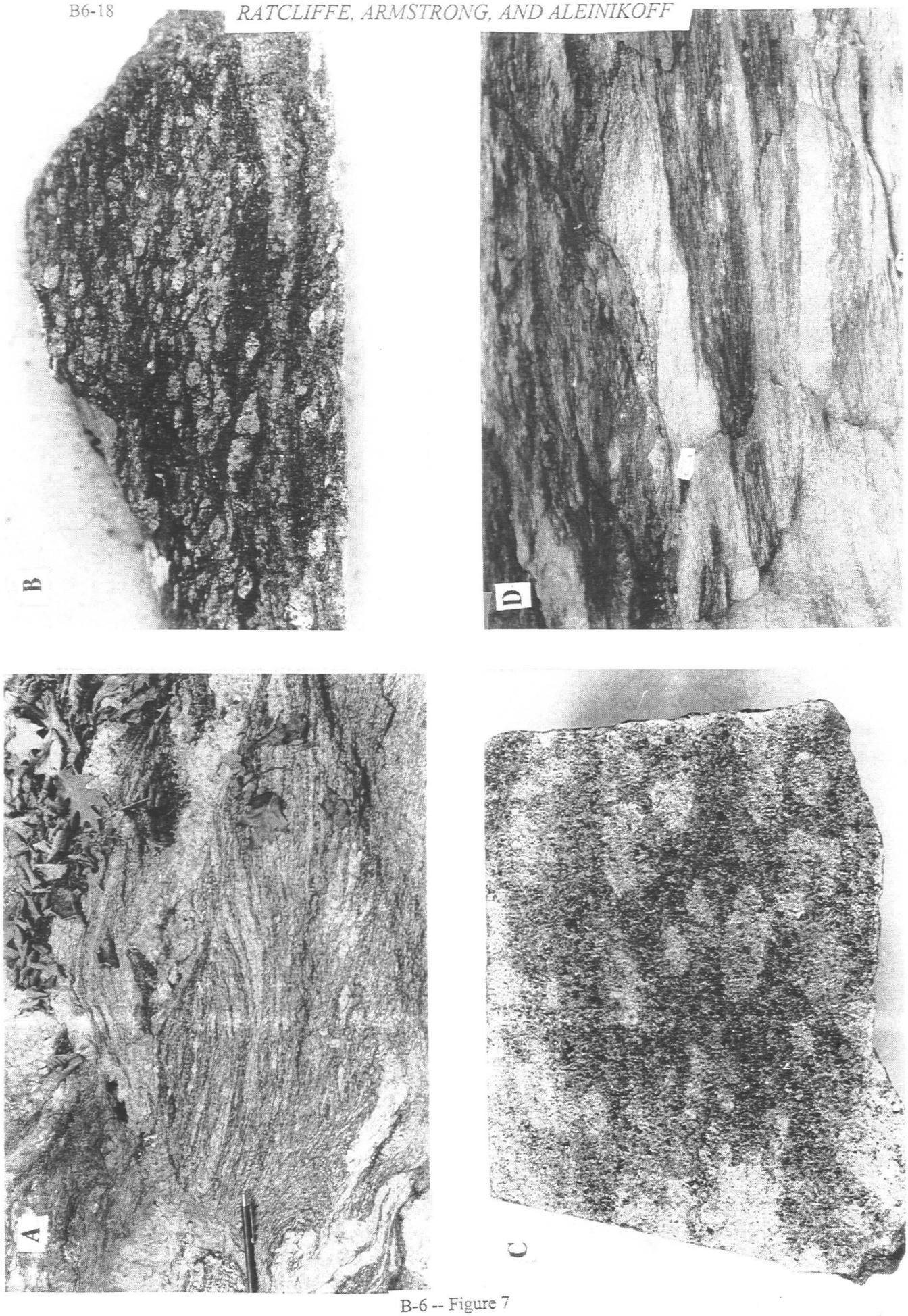
The lower transition into gneiss is exposed on the lower slopes of Hawks Mountain, below approximately 1,300 ft elevation in the west. In this area, coarse-grained, biotite-spotted trondhjemite and tonalite gneiss (Yt) is recognizable despite an intense, penetrative mylonitic foliation that dips approximately 15° to 10° to the north and northwest (fig. 7A). Zones of mylonite, spaced meters apart, transect the coarse gneiss and weave around elongate blocks, knots, and sheared-off fold noses of the gneiss, producing marked zones of disarticulation in the plane of the mylonitization. This zone gradually passes to lower elevations of 1,100 ft at the east end of Hawks Mountain in the Cavendish quadrangle.

All along the south-facing cliffs of the Hawks Mountain, there is a gradual transition upward from the above mentioned basal zone into increasingly sheared, darker, finer grained biotite mylonite schist. This zone of mylonite schist contains two distinctive rocks. The first is a phacoidally sheared plagioclase-"augen" mylonite (figs. 7B, C) in which 1- to 2-cm porphyroclasts of plagioclase account for 70 percent of the rock. The other is a dark, porphyroclastic, biotite mylonite schist, which contains smaller (1 to 0.5 cm) mylonitic and milled composite “grains” of plagioclase, biotite, and quartz. The "grains" have distinctive growth tails and asymmetric shapes typical of mylonitic porphyroclasts modified by recrystallization and dynamic regrowth.

Within about 20 m of the contact with the Cavendish, the rocks are fine grained, uniformly foliated schists and blastomylonite, in which less-deformed, elliptical to tabular pods, 5 to 10 cm thick, of less-deformed gneiss occur (fig. 7C). The actual contact with the Cavendish can be located to within 2 m in the western part of the Chester quadrangle, where the aluminous Gassetts Schist Member (Ycg) is at the contact. From the crest of Hawks Mountain eastward and into the Cavendish quadrangle, the contact is more difficult to locate because the biotite feldspathic schist of the Cavendish (Ycf) closely resembles and appears to be transitional into the mylonite derived from the tonalite gneiss (fig. 7B and D).

At the east end of Hawks Mountain in the Cavendish quadrangle, the mylonitic foliation turns sharply to the south and undergoes intense plication in later folds that trend N. 25°-30° E. and have steeply southeast-dipping axial surfaces. Mylonitic, porphyroclastic gneiss, like that seen along the lower and intermediate slopes of Hawks Mountain, occurs at the 600- to 700-ft elevation of Quarry Road, 700 m southwest of Perkinsville in the Chester quadrangle. From this point southward, the mylonitic foliation extends southward to the vicinity of Kendricks Corner (K on fig. 4).

The lower contact of the Cavendish Formation on Hawks Mountain is therefore interpreted as a thrust fault here named the Hawks Mountain thrust fault. The Hawks Mountain thrust fault truncates the contact



B-6 -- Figure 7

between the Gassetts Schist Member (Ycg) and the feldspathic schist member (Ycf) of the Cavendish in the hanging wall.

In the northeastern part of the map (fig. 4) the feldspathic schist member (Ycf) unit and a nearly continuous belt of marble and calc-silicate rock (Ycm) occur, forming the southern closure of the large mass of Cavendish Formation exposed on Pine Hill in the adjacent Cavendish quadrangle. The marble is in contact with the feldspathic schist above and the contact appears to be a normal, interbedded one. The contact with the underlying Baileys Mills Tonalitic Gneiss is not exposed but the gneiss is highly mylonitic and porphyroclastic like the gneisses along the lower slopes of Hawks Mountain. The rocks of the Cavendish Formation belonging to the Hawks Mountain mass, extends northward through the Cavendish quadrangle and terminates in a tight fold at the latitude of Felchville due west of the Ascutney pluton.

In summary, the Cavendish Formation in the Hawks Mountain area of the Cavendish and Chester quadrangles is considered either to be intruded by trondhjemite gneiss of the Mount Holly Complex or to be interlayered with the biotite-quartz-plagioclase gneiss of the Mount Holly. The lower contact with the Baileys Mills Tonalitic Gneiss beneath the Cavendish Formation is interpreted locally as a fault, but this also may have originally been an intrusive contact.

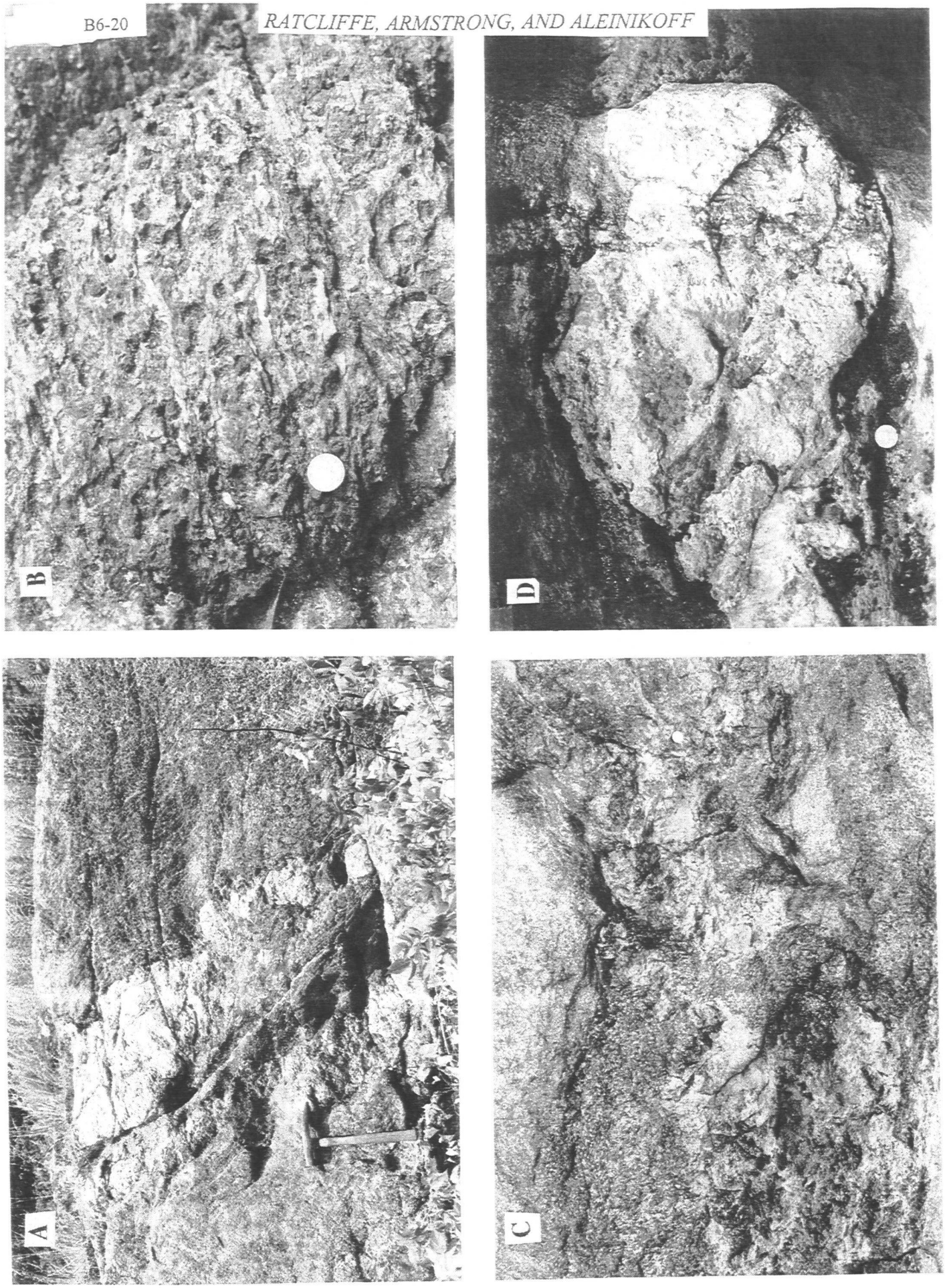
Cavendish Formation in the Star Hill area of the Cavendish quadrangle

The outer belt of aluminous schist (Ycg) and minor calc-silicate rocks of the Cavendish Formation at Star Hill in the Cavendish quadrangle is mapped in the same position as the Star Hill Formation of Thompson (1950) and the Cavendish Formation of Doll and others (1961). Most of this belt is muscovite-rich, garnetiferous biotite-staurolite-quartz±kyanite schist of the Gassetts Schist Member (fig. 8B). East of Hardy Hill Road the belt parallels abundant granitic gneiss (Ygg) and pegmatite (Yp) to the east and extends northward where it widens to about 0.5 km. Along the western margin are two areas of white, tremolite-rich marble and calc-silicate rock (Ycm) in contact with a narrow selvage of feldspathic Cavendish (Ycf). Outcrops of the Cavendish Formation extend for 3 km north of Star Hill where the schist terminates in a complex fold within the surrounding biotite-quartz-plagioclase gneiss (Ybg) of the Mount Holly Complex.

On Star Hill west of Hardy Hill Road, a narrow band of calcite-diopside marble and calc-silicate gneiss (Ycs) 5 m thick is interbedded with pegmatite-bearing biotite gneiss (Ybg) typical of the Mount Holly Complex. This band of calc-silicate rock extends intermittently for about 300 m to the north. To the west of the calc-silicate rocks, and a narrow band of biotite-hornblende gneiss (Ybg), is another band of rusty aluminous schist mapped as unit Yrs that is not quite as coarse grained as the Gassetts Schist Member of the Cavendish mapped east of the road. In figures 4 and 6 all schistose rocks in the Mount Holly are shown as Yrg. The discussion below uses units Yrs and Yrg that are distinguished separately on the

Figure 7. Photographs illustrating the structural (mylonitic) transition from Baileys Mill Tonalite Gneiss into "basal" schists of the Cavendish Formation across Hawks Mountain thrust in Chester quadrangle.

A. Mylonitic tonalite gneiss 50 meters below base of Cavendish Formation on Hawks Mountain view is to north, pencil shows scale.
B. Porphyroclastic plagioclase augen gneiss in Hawks Mountain fault zone 10 meters below Cavendish Formation, specimen is 10 cm long cut perpendicular to long axis of augen and mylonitic schistosity.
C. Porphyroclastic plagioclase augen gneiss like B cut parallel to subhorizontal mylonitic foliation at Hawks Mountain thrust east is to right, north at top. Elongation of porphyroclasts is east-west.
D. "Basal" feldspathic schist of the Cavendish Formation (Ycf) above Hawks Mountain thrust on Hawks Mountain. Relict layers of mylonitic tonalite gneiss are surrounded by darker layers of biotite-plagioclase blastomylonite schist of the Cavendish. Scale is 2 cm long, view is to north.



B-6 -- Figure 8

1:24,000 digital map (Ratcliffe, 1995a). This very narrow belt can be traced more than a kilometer; its northern tip disappears into a mass of coarse pegmatite (Yp). This (Yrg) schist is overlain by typical biotite-microcline-plagioclase-quartz gneiss (Ybg) of the Mount Holly and contains a sill of aplitic gneiss of the Felchville Gneiss. Slightly higher on the south end of Star Hill is a third layer of rusty muscovite-biotite schist (Yrg) similar to but not identical with the lower one; it is less than 100 m long and terminates in biotite-quartz-plagioclase gneiss (Ybg). Pegmatite typical of the Mount Holly Complex intrudes all units on Star Hill (fig. 8A; see Stop 6).

The irregular southern mass of aluminous schist on Star Hill was mapped as the Pinney Hollow Formation by Thompson (1950) and Downie (1982) but is here considered part of the Cavendish Formation. These aluminous garnetiferous schists are mineralogically identical to the lower two aluminous schists (Yrg and Ycg) east of Star Hill (Downie, 1982). The contact with adjacent gneisses of the Mount Holly are well defined on the western slopes of Star Hill. There and south along the powerline, the schist contains thin lenses of calc-silicate gneiss and amphibolite, as well as highly folded pegmatites. In one exposure along the powerline, the schist is crosscut by aplitic gneiss and strongly foliated pegmatite (fig. 8A). The Cavendish Formation in this southern mass is primarily rusty-weathering, highly feldspathic, coarsely garnetiferous schist commonly associated with calc-silicate rock. A lithologic correlation of these schists with those of the Pinney Hollow Formation (Downie, 1982) is unwarranted because the Pinney Hollow lacks calc-silicate rocks, is uniformly chlorite-, muscovite-, and quartz-rich, and has little feldspar. All the garnetiferous schists on Star Hill are similar and are interlayered with gneiss and calc-silicate gneiss of the Mount Holly Complex.

INTRUSIVE ROCKS OF THE MOUNT HOLLY COMPLEX

The well-layered paragneiss units of the Mount Holly and Cavendish Formation are extensively intruded by two different varieties of Middle Proterozoic felsic rocks. The oldest intrusive rocks are tonalitic, and trondhjemite to granodioritic rock mapped as the Baileys Mills Tonalitic Gneiss (Yt), and the Felchville Gneiss (Yft).

The Baileys Mills Tonalitic Gneiss is a coarse-grained, biotite-spotted gneiss locally having relict phaneritic texture resembling coarse-grained tonalite or trondhjemite. The unit is named for exposures along State Route 103 north and south of Baileys Mills in the Chester quadrangle (Ratcliffe, in press). These rocks closely resemble rocks in the Mount Holly Complex in core of the Green Mountain massif dated at 1.35 to 1.31 Ga that were referred to informally as the Rawsonville and Cole Pond facies of an unnamed metatrondhjemite suite (Ratcliffe and others, 1991). In the absence of clear crosscutting relations with other units of the Mount Holly Complex, they were interpreted as older than the paragneiss units of the Mount Holly (Ratcliffe and others, 1991). In the core of the Chester dome, however, both in the Cavendish quadrangle (Ratcliffe, 1995a) and in the Chester quadrangle, these trondhjemites and tonalites intrude the biotite-quartz-plagioclase gneiss (Ybg) and associated units (Ya, Yrs, Yrg). Enclaves of amphibolite (Ya) and biotite gneiss (Ybg) are common within the tonalites and trondhjemites. In the

Figure 8.

A. Deformed Middle Proterozoic pegmatite in schist of Cavendish Formation, powerline exposure Stop 6 near Star Hill. Relict gneissosity dips steeply east, Taconian or Acadian regional foliation is gently north dipping. View to north, hammer 0.2 m long.
B. Coarse garnet schist of Gassetts Member of the Cavendish Formation, Star Hill Stop 7. Dominant foliation dips gently to west, older schistosity is subvertical and strikes N10°E. Penny shows scale.
C. Shredded and retrograded Middle Proterozoic pegmatite pod in Cavendish Formation. Cavendish Gorge 400 feet downstream from dam Stop 8. Chlorite and ilmenite replace relict stellate cluster of coarse biotite or hornblende at right middle. The view is approximately 1 meter wide.
D. Deformed and retrograded Middle Proterozoic pegmatite pod, 0.5 m wide in Cavendish Formation Cavendish Gorge. Pegmatite in both C and D are highly foliated, folded and boudinaged.

Cavendish quadrangle, the Felchville Gneiss (Yft) is a trondhjemite gneiss broadly intrusive into most rocks of the Mount Holly Complex and into the Cavendish Formation (Ratcliffe, 1995a). Two samples of the Felchville Gneiss (Table 1) from the Cavendish quadrangle yielded $^{207}Pb/^{206}Pb$ ion microprobe (SHRIMP) ages for cores of zircons of 1424±22 Ma and 1379±23 Ma. The same zircons have overgrowths dated at 1111 and 950 Ma. These dates are interpreted as the age of intrusion of the Felchville Gneiss, and the times of subsequent metamorphism in the Middle Proterozoic. No data are available from the Baileys Mills Tonalitic Gneiss, but it is reasonable to assume that its age is also approximately 1.4 Ga. Because both the Felchville Gneiss and the Baileys Mills Tonalitic Gneiss intrude the paragneisses of the Mount Holly Complex and the Cavendish Formation, both of the latter are older than 1.4 Ga.

Major-oxide, trace-element, and rare-earth-element chemical analyses of the Baileys Mill Tonalitic Gneiss and the Felchville Gneiss in the Cavendish quadrangle help establish correlation of these rocks. The Felchville Gneiss tends to be higher in SiO_2 (67 to 72 percent) and lower in CaO (0.79 to 3.69 percent) than the Baileys Mills Tonalite Gneiss. Plots of normative An-Or-Ab compositions show the Felchville rocks to be trondhjemite and sodic tonalite, whereas the Baileys Mills samples tend to be more tonalitic (Ratcliffe, in press).

The tonalitic and trondhjemitic gneisses of the Chester dome closely resemble, in major oxide trace-element geochemisty and age rocks from the Green Mountains belonging to the Rawsonville, Cole Pond, and Bondville facies of the Green Mountain massif identified by Ratcliffe and others (1991); therefore, correlation with those rock is strongly indicated. In general, the Felchville Gneiss and its aplitic facies resemble the Rawsonville and Bondville facies of the Green Mountain massif. The Baileys Mills Tonalitic Gneiss is comparable to both the Cole Pond and Rawsonville facies.

Rare-earth element characteristics of the two groups of rocks from the Chester dome are similar. The Felchville Gneiss has chondrite-normalized patterns with La values of 28 to 42 x chrondrite, and Lu values of 1.9 to 3 x chondrite. The Baileys Mills Tonalitic Gneiss samples are similar to those of the Felchville, but have higher chondrite-normalized La values at 112 to 25 x chondrite and Lu values of 1.3 to 3.3 x chondrite respectively. Both sets of rocks exhibit no Eu anomalies and have La_c/Lu_c (chondrite normalized) ratios of approximately 20.

The chemistry of the Bailey Mills and Felchville rocks closely resembles the Rawsonville and Cole Pond trondhjemite and tonalitic facies of the Green Mountain massif as well as similar rocks cropping out on Terrible Mountain in the Weston quadrangle (Ratcliffe and Burton, 1996) which were not included in Ratcliffe and others (1991). Five samples from Terrible Mountain closely resemble, in their coarse-grained biotitic character, the Baileys Mills Tonalitic Gneiss. This is especially significant because Terrible Mountain appears at the easternmost edge of the Green Mountain massif at the same latitude as the Chester dome. The identical rocks reappear in the Chester dome as the Baileys Mills Tonalite Gneiss. Figure 1 of Ratcliffe and others (1991) shows that correlatives of the tonalite-trondhjemite suite of the Green Mountainmassif appear at similar latitude and on strike with rocks identified by McLelland and Chiarenzelli (1990) in the Adirondack Mountains of New York; these rocks are here extended into the Chester dome.

Other intrusive or migmatitic granitic gneisses (Ygg) are found within the Baileys Mills Tonalitic Gneiss, the Felchville Gneiss, and in the paragneiss units of the Mount Holly Complex. These granitic rocks appear locally to intrude other units of the Mount Holly Complex, but in other cases appear to be migmatitic or anatectic rocks perhaps replacing or derived from older microcline-rich, perhaps metarhyodacitic rocks. The origin of these rocks is probably highly complex both in terms of the original protolith and the degree of anatectic remobilization. Similar granitic gneisses are common throughout the Mount Holly Complex of the Green Mountain massif and clearly crosscut both the tonalite-trondhjemite gneisses and the paragneiss units of the Mount Holly. Preliminary data suggest that these granitic gneisses are intruded or formed in place at about 1.25 Ga (Ratcliffe and others, 1991).

Pegmatites in the Mount Holly Complex

Coarse-grained granitic gneiss ranging to pegmatite is widespread in the Mount Holly Complex in the domes. Masses of granitic gneiss contain large microcline crystals 1-2 cm long and a texture that varies from pegmatite to medium grained. Two of the largest masses are east of Star Hill and on the hills east and south of South Reading in Cavendish. Small pods of pegmatite 0.5-10-m thick appear in all of the units of the Mount Holly Complex including the Cavendish Formation (figs. 8C, D). These pegmatites are now chloritized, highly deformed, and clearly predate the regional Paleozoic foliation. Excellent pegmatite exposures occur in the schists of the Cavendish Formation on the east and west walls at the upper end of Cavendish Gorge about 150 m downstream from the dam. These pegmatite pods, which are 0.5-1 m in diameter, contain ghosts of coarse-grained biotite and hornblende that have been totally replaced by fine-grained biotite, chlorite, and ilmenite (figs. 8C, D).

The age of these pegmatites is unknown, but similar kyanite-bearing pegmatites on Ludlow Mountain intrude quartzites, calc-silicate gneisses, and aluminous kyanite-garnet schists (now retrograded largely to muscovite-chlorite-chloritoid-quartz schists). U-Pb SHRIMP ages of zircon from the granodiorite gneiss associated with kyanite-bearing pegmatites are approximately 1314±7 Ma (Table 1). These data and the map relationships on Ludlow Mountain (Ratcliffe, 1992; Walsh and others, 1994) show that aluminous large-garnet schist that resemble the Gassetts Schist Member of the Cavendish Formation is older than 1314±7 Ma. The old, highly deformed pegmatites within the Cavendish Formation in the Cavendish quadrangle (see fig. 8C, D) are markedly different from the younger Devonian pegmatite and two-mica granites that crosscut the regional foliation; the younger straight-walled pegmatites are notably poor in potassium feldspar and are biotite-muscovite granodioritic in composition. The Devonian pegmatites and granite dikes, however, crosscut the strong Paleozoic foliation and are generally weakly foliated, whereas the Proterozoic pegmatites are all highly folded and shredded in the regional Paleozoic foliations (see fig. 8A, B).

The Bull Hill Gneiss Member of the Cardinal Brook Intrusive Suite

The Bull Hill Gneiss of Richardson (1929) is named for exposures of coarse-grained microcline augen gneiss on Bull Hill in Saxtons River quadrangle, 3 ½ km northwest of Cambridgeport. On Bull Hill, the rock is a coarse-grained microcline phenocrystic granite in which the individual microcline crystals may be ovoidal to rectangular and have thin plagioclase rims. Commonly this structure is destroyed because of intense mylonitization (see Ratcliffe, 1991, p. 19-25). In the Saxtons River quadrangle, the Bull Hill Gneiss Member of the Cardinal Brook Intrusive Suite (Ratcliffe and Armstrong, 1995) is discordant to units of the Mount Holly Complex. Cliff exposures of an intrusive breccia containing rotated blocks of gneissic Mount Holly can be seen on the slopes above the Saxtons River 0.8 km west of the crest of Bull Hill. Here the medium-grained aplitic igneous matrix contains small 1 to 2 cm phenocrysts of microcline. Both the gneissic blocks, as much as 10 cm in length, and the igneous matrix are cross-cut by a foliation younger than the gneissosity contained in the rotated xenoliths. In the vicinity of Bear Hill, on the west side of the dome in the Saxtons River quadrangle, and along the west facing slopes to the south, very coarse-grained rapikivi-textured ovoidal microcline forms grains as much as 3 cm in diameter in the Bull Hill Gneiss. These textures are very similar to those present in the much less deformed Somerset Reservoir Granite and Stamford Granite of the Cardinal Brook Intrusive Suite (Ratcliffe, 1991, figs. 2 and 3).

The Bull Hill Gneiss, exposed along the western border of the Athens dome is all highly mylonitic. Augen of microcline define an orientation-shape fabric with the long axes of the augen plunging down the dip of the foliation to the west, northwest, or to southeast, depending upon the direction of dip of the foliation.

The Bull Hill Gneiss is confined to the Chester and Athens domes (Ratcliffe, 1991), where it occurs both at the margin and in the core, principally in the area south of the Chester quadrangle. Although previously mapped as a continuous unit around the Chester dome (Doll and others, 1961), it actually

extends for only a short distance north into the Chester quadrangle. Along the eastern margin of the dome, it extends as far north as Springfield (Walsh and others, 1996a, b). Along the western margin it extends as far north as the village of Gassetts (where it may occur as fault slivers). Previously reported occurrences (Doll and others, 1961) in the Chester and Cavendish quadrangles are not Bull Hill Gneiss but coarse-grained plagioclase-augen gneisses that belong to the Felchville Trondhjemite Gneiss (Ratcliffe, in press).

The Bull Hill Gneiss intrudes the Mount Holly Complex throughout the Athens dome (Ratcliffe, 1991, fig. 6). Two samples of Bull Hill Gneiss from the Athens dome and from southern part of the Chester dome yielded U-Pb zircon upper intercept ages of 945±7 and 955±5 Ma respectively (Karabinos and Aleinikoff, 1990). The latter sample comes from roadcuts in the Saxtons River quadrangle on the south side of Route 103, 300 m east of the intersection with Sylvan Road (see road log). The former sample comes from a point 2 km south of the southern border of the Saxtons River quadrangle from a belt of Bull Hill Gneiss exposed on the western side of Branch Brook and, above a belt of the Hoosac Formation all of which are above the Ober Hill fault in the Townshend quadrangle (Ratcliffe, 1991).

The approximately 955 to 945 Ma age for the Bull Hill Gneiss agrees closely with other ages determined for other members of the Cardinal Brook Intrusive Suite (Karabinos and Aleinikoff, 1990). These other members are clearly intrusive into gneissic units of the Mount Holly Complex. Elsewhere, these rocks lack the strong gneissosity present in the Mount Holly Complex. For this reason the Bull Hill Gneiss and all other members of the Cardinal Brook Intrusive Suite are regarded as post-Grenvillian (Ratcliffe and others, 1988, p. 5-9; Ratcliffe, 1991). This observation is important for structural analysis because the foliation present in the Bull Hill Gneiss must be younger than the Grenville and must, therefore, be Taconic, Acadian, or younger. The comparison between the structure developed in the Mount Holly Complex and that in the Bull Hill Gneiss, therefore, can be a guide to recognizing post-Grenville structure in the core of the Chester and Athens dome.

Cambrian and Ordovician Cover Sequence Rocks of the Chester and Athens domes

A thin, 0.5- to .75-km-thick section of cover rocks overlies the core gneisses of the domes. The section is particularly thin along the northwest, northern and eastern flank of the domes. The ages of these rocks and formation assignments have always been problematic but they are thought to range from Late Proterozoic to Ordovician. The sequence is overlain by Silurian and Devonian rocks of the Northfield and Waits River Formations.

Previous maps (Doll and others, 1961; Thompson and others, 1993) show these cover rocks as continuous, parallel units equated with the standard Hoosac, Pinney Hollow, Stowe, and Moretown units of the normal Eastern Vermont Sequence of Doll and others (1961). Our present mapping (fig. 1) does not support that. Instead we find that the map units are irregular and nonpersistent; in many areas rocks of the Moretown Formation rest or nearly rest directly on basement rock. This is true in the Keyes Mountain area in the Cavendish quadrangle and in the northern part of the dome in general, we place a major fault, the Keyes Mountain thrust, along this basement cover contact. In the Saxtons River quadrangle, on both flanks of the Athens dome, an increasingly thicker section of lower level cover rocks appears beneath the Moretown Formation to the south. However this section beneath the Moretown is highly faulted and also lacks stratigraphic continuity. We have shown the rocks below the Moretown as Rowe Schist and believe that this section is a tectonic melange. Both the Moretown Formation and the Rowe Schist form an allochthonous, highly thrust-faulted tectonic cover, probably Taconian in age, that mantles the Chester and Athens domes.

The Moretown and younger part of the tectonic cover are intruded by tonalitic, mafic and trondhjemitic rocks of the North River Igneous Suite, or rocks of the Barnard Gneiss (of Richardson, 1924). Felsic and mafic metavolcanic rock, coticule, quartzite, schist and amphibolite in fairly regular succession overlie the intrusive rocks and are assigned to the Cram Hill Formation. At present we regard the Moretown Formation, North River Igneous Suite, Barnard Gneiss (of Richardson, 1924) and Cram Hill

Formation as allochthonous to epikinallochthonous units (deposited in transit) on the upper plate of this Taconian tectonic cover.

The Northfield and Waits River Formations overlie the older rocks nonconformably with only a slight but perceptible truncation of units north of the Saxtons River quadrangle. Along the Waits River contact with the underlying Moretown, Barnard Gneiss (of Richardson, 1924), and Cram Hill Formation at the northern end of the dome, a thin section less than 10 m thick of probable recycled felsic and mafic metavolcanic, quartzite and limey schist occur interbedded with the base of the Waits River (Ratcliffe, 1995a, in press). Evidently some felsic and mafic layered gneisses that resemble rocks of the Barnard Gneiss (of Richardson, 1924) intrusive rocks of the North River Igneous Suite and layered metavolcanic rocks of the Cram Hill Formation all Ordovician are interlayered with the Waits River Formation along its base.

STRUCTURAL GEOLOGY

Introduction

Structure of the Chester and Athens domes has figured prominently in interpretations of the models for the tectonic evolution of this part of the New England Appalachians. Previous interpretations, while all recognizing the Chester-Athens dome as antiformal, have depicted the areas to the west of the domes as having been recumbently folded by early Acadian deformation (nappe-stage folds of Thompson, 1950; Rosenfeld, 1968). The cross section E-E' from Doll and others (1961) (fig. 3) familiar to most geologists in New England, illustrates the basic elements of the classic structural interpretation. This inverted sequence involved core rocks of the Butternut Hill fold as well as rocks of the cover sequence west and south of the Butternut Hill fold, and rocks of the Spring Hill area. The early nappe fold, in Rosenfeld's (1954, 1968) view, is the Proctorsville syncline and Townshend-Browington syncline (TBS, fig. 3) with the upper or inverted limb containing Middle Proterozoic rocks in the core of the late Butternut Hill fold. The geology of the Saxtons River quadrangle is critical for evaluation of these ideas. In the view of Thompson (1950) and of Thompson and others (1993) inverted and recumbently folded basement and cover rocks exist in the Star Hill area as well and the Butternut Hill fold (BHF, figs. 3 and 4) and the Star Hill fold (STHF, fig. 3) both of which are viewed as nappe stage folds. The interpretation of Downie (1982) is even more complex and will be discussed at Stop 6.

The hypothesis (J.B. Thompson, Jr., 1950; Rosenfeld, 1968; Thompson, Rosenfeld, and Downie, 1986) that the basement and cover rocks were involved in recumbent isoclinal folds can be tested by mapping the units of the Mount Holly Complex in the core of the domes. A geometric requirement of the isoclinal and recumbent fold model is that the axial surfaces of the proposed nappe-stage folds are the actual planes of symmetry about which the rocks are repeated, as, for example, along the axial traces shown in figures 3 and 4. This symmetry must also be recognizable in the folded pattern of rock units of the Mount Holly Complex in the core of the dome. We have mapped in great detail the distribution and structure of the Mount Holly Complex which had not previously been done (fig. 4). Analysis of the grain in the Mount Holly Complex, in conjunction with careful plotting of the schistosity that might be axial planar to nappe folds provides a particularly useful analytical approach. Form-line maps (fig. 5) of the dominant schistosity in the core of the dome were used to evaluate the extent of pre-dome folding as well as the amount of internal warping during the formation of the dome. Our basic conclusion is that high-amplitude folds corresponding either to the early nappe-stage folds or the dome-stage folds (see fig. 4) proposed are not present in the rocks of the Mount Holly Complex. The dome is a low amplitude structure formed by crenulation folding recognizable in the folded pattern of the schistosity form-line maps (see sections fig. 6).

It is important to note that the above geometric method of analysis is independent of the age assignment of the Cavendish Formation. This is true because the distribution of the Cavendish Formation and its contact with the other units of the Mount Holly Complex do not conform to the proposed patterns for the nappes. For example, in the southern closure of the Chester dome (fig. 4), units of the Mount Holly Complex trend

northeastward across the axial trace of the dome and show little or no deflection in map pattern. The same relationship is evident in the Athens dome. In the Cavendish quadrangle, units in the Mount Holly Complex trend northeast with subvertical dips diagonally across the dome. In the core of the Butternut Hill fold in the Andover quadrangle units in the Mount Holly are not repeated symmetrically as in the core of a nappe but extend east-west across the structure. In general, the internal units in the core of the dome strike and dip at oblique angles to the form of the dome. The general trend of the units and the gneissosity is northeasterly and dips are moderately steep to vertical. Deformation of the core rocks of the domes was not great enough to bring the internal units into conformity with the shape of the dome. Until we mapped the internal structure of the Mount Holly Complex these relationships could not have been appreciated.

The attitude of Paleozoic foliation in the core rocks is at a high angle to the gneissosity in the Mount Holly Complex. The pattern of the folded foliation illustrates well two domal structures with a structural saddle between the two domes (fig. 5). The shape of the dome, as determined by the folded internal foliation, is that of an aysmmetric antiform overturned eastward (fig. 6, section C-C', B-B') to more upright, section A-A'. The projection of the foliation into the line of sections define a dome with a low amplitude, and rather open nonattenuated form. Map relationships like this are now known to be the case throughout the Chester and Athens domes, from areas to the north (Ratcliffe, 1995b; in press) and to the south (Ratcliffe and Armstrong, in press; Ratcliffe, Armstrong and Tracy, 1992).

Structure of the Spring Hill syncline in the Saxtons River quadrangle

Rocks of the Cram Hill Formation form the core of a well-defined synform in the central part of the map (fig. 1). Schistosity in all the rocks of the synform is highly folded and, in fact, the structure is a synform of folded schistosity that has been affected by several sets of crenulation cleavage and cross-folds. A moderately persistent zone of black carbonaceous phyllite, associated quartzite and quartz-pebble conglomerate rim the structure (Ratcliffe and Armstrong, 1995, 1996). At the northern end of the syncline, plunges of minor folds and elongation direction of quartz pebbles plunge gently to the south-southwest, whereas to the south, plunges of these features are north to northwest.

A prominent belt of plagioclase-spotted amphibolite (perhaps metagabbro) appears to intrude the otherwise layered amphibolites. At the top of the amphibolite there occurs a prominent but discontinuous zone of pinkish magnetite-rich coticule, quartzite, dark feldspathic schist and magnetite ironstone within the center of the syncline.

The syncline itself is broadly asymmetric, overturned eastward with a steeply-dipping western flank (C-C', fig. 6). Minor folds near the center of the structure plunge gently southwest, are horizontal, or plunge gently north. The southern closure of the fold plunges generally north, but the bearing and plunge of minor folds are determined by at least four sets of axial surfaces. The oldest set of folds trends generally northeast-southwest and has northwest-dipping axial surfaces. The axial surfaces are broadly folded by subsequent (F_2) northwest- (F_3), northeast-, and by the latest (F_4) north-trending folds.

Although numerous F_2, F_3, and F_4 folds cross-cut the larger synform, the plunges overall define a doubly-plunging structure (see cross section C-C', fig. 6). Detailed mapping of the units has not revealed evidence for complicated repeats of the units, nor is there any evidence at all for inverted strata in the northern part of the syncline. In short, our data do not support the structural interpretation of Rosenfeld (1954) and of Doll and others (1961) that shows the Spring Hill structure as a northerly-plunging, upward closing antiformal syncline corresponding to the lower limb of a nappe as shown in figure 3.

Structure of the Butternut Hill Fold

The axial surface of both the Spring Hill syncline and the Butternut Hill fold in our usage appear to be related to multiple Acadian refolding of older folds. These late folds deform coarse-grained schistosity that is axial planar to folds in the Moretown Formation south of the closure of the Butternut Hill fold. This older and folded schistosity is interpreted as Taconian and the same general age as the composite S_1 and S_2

cleavages in the Hoosac through Stowe belt to the west (OCr unit on fig. 1). Plunges of the folded schistosity are erratic, particularly in the area immediately south of the closure in the Ombs unit of the Moretown Formation where the general trend of the schistosity is east-west and steeply-dipping. Analysis of the plunge of minor folds and of the crenulate cleavages indicates that as many as four separate sets of steeply plunging Acadian folds are present. The Butternut Hill fold does not plunge northward. Instead, plunges of the folded schistosity are predominantly subvertical, steeply north or northwest plunging. Cover rocks of the Hoosac Formation on the southern closure of the Butternut Hill fold (Stop 15) plunge to the south-southwest. Neither the Butternut Hill fold, the Spring Hill fold, nor the Star Hill fold therefore conform to the north-plunging and inverted model of Doll and others (1961). We interpret the structures as normal anticlines and synclines (fig. 6).

Structure of rocks east of the Chester and Athens domes

The eastern flank of the domes consists of a nearly homoclinal, easterly-dipping section of rocks from Middle Proterozoic basement up through the Devonian and Silurian Waits River Formation. The pre-Silurian section seems relatively uncomplicated by folding but is very strongly mylonitic at or near the margin of the domes. Numerous thrust faults such as the Kenny Pond, Townshend, and South Newfane thrusts are mapped along this section and account for the irregular along-strike distribution of units. Rocks of the Moretown Formation appear very close to the basement rocks and in certain places are essentially at the contact with Bull Hill Gneiss (see Stop 12).

Rocks of the Moretown Formation, Cram Hill Formation, igneous rocks of the North River Igneous Suite, as well as the overlying Northfield and Waits River Formations all contain a very strong Acadian schistosity and several younger Acadian crenulation cleavages. Locally, Acadian thrust faults duplicate the section and place Moretown and Cram Hill units onto the Northfield and Waits River Formations. The South Newfane thrust has been traced northward into this area from the Brattleboro and Newfane quadrangles (Armstrong, 1994) where it has been interpreted as a significant Acadian thrust fault. It dies out northward, and strain appears to be transferred to the more easterly Acadian faults.

Clearly, intense Acadian deformation and thrust faulting affected rocks on the eastern side of the domes. A more difficult and, at present, unresolved matter, is the extent of Taconic deformation in the pre-Silurian rocks. We favor the interpretation that all these rocks, with the possible exception of some intrusive rocks in the North River Igneous Suite and perhaps some of the Cram Hill, were affected by Taconic deformation and have a schistosity that is older than the first Acadian foliation in the Silurian and Devonian rocks. We conclude this because at the south end of the Athens dome, the Moretown Formation and older rocks are coextensive with rocks to the west that contain the older Taconic structures (Ratcliffe and Armstrong, in press; Armstrong, 1994). We conclude that the intensity of Acadian structure east of the domes has obscured the older structures and suggest that the schistosities in the pre-Silurian rocks east of the domes have a composite Taconic and Acadian schistosity (fig. 5). This is the result of the nearly homoclinal dip of foliation and the tightly oppressed later folds of foliation along the east flank of the dome. The major deformational front that separates the less deformed Taconic schistosity on the west from the composite Acadian and Taconic schistosity to the east is parallel to the strongly developed northeast-trending Acadian F_4 folds, recognized by Ratcliffe and Armstrong (1995) and shown in figure 5.

This transition is illustrated by the folding of the foliation in the Bull Hill Gneiss as it increases from gentle warping in the core of the dome to intense plication in the eastern 1/3 of the domes. This zone of abundant Acadian shortening is shown in section B-B' and C-C' by prominent folds of the older Townshend, Kenny Pond, and other Taconian thrust faults. Locally Acadian thrust faults are present in this zone. As noted previously, Devonian granite dikes are very highly folded along the eastern margin of the Chester dome, whereas in area to the west these dikes are highly discordant and only weakly folded (Stop 9).

Metamorphism and Tectonism

Ordovician and older cover sequence and basement rocks were all affected by intense deformation and metamorphism perhaps as high as garnet-grade during the Taconian orogeny. The final Taconian

dynamothermal event involved imbricate thrust faulting of basement and cover rocks during the Taconic F_2 event (see Sutter and others, 1985; Ratcliffe and others, 1992). This event was responsible for a regionally developed set of synmetamorphic thrust faults characterized by reclined folds whose hingelines plunge S70°E to E (see Ratcliffe, 1997). The F_2 faults are now the major contacts between units of the eastern Vermont eugeosynclinal sequence (see Trip C-6). These faults and the intense rodding and reclined folds in the Ordovician tectonic cover are folded over the domes. The Taconian F_2 structures are absent from the younger Silurian and Devonian rocks which have a separate deformation plan produced during the Acadian dynamothermal events. The Silurian and Devonian rocks contain multiple schistosities and five generations of folds. The earliest schistosity is bed-parallel; later structures are accompanied by crenulation cleavages that locally pass into more penetrative foliations, especially along the eastern flanks of the domes. The schistosity within the core of the dome (fig. 5) is in part Taconian, especially near the major faults. We do not know to what extent the earliest Acadian foliation is expressed in the core rocks of the dome; and the schistosity may be composite of Taconian $S_s(F_2)$ and the Acadian S_1 foliation. The discussion below describes the intense Acadian metamorphic overprint that has affected all the pre-Cretaceous rocks of this area to some extent. $^{40}Ar/^{39}Ar$ hornblende and biotite and white mica ages from this area record only Acadian cooling ages (Laird and others, 1984; Sutter and others, 1985; Hames and others, 1993; Harrison and others, 1989).

Acadian Metamorphic conditions in the vicinity of the Chester Dome

Figure 9 shows the distribution of Acadian metamorphic isograds around the Chester Dome. Although the Acadian garnet isograd west of the dome continues to the north and south of the closure of the domes, the staurolite-kyanite isograd is somewhat concordant with the shape of the dome. The isograd location is controlled in part by the distribution of aluminous pelite within the basement and cover sequences that conform to the form of the Acadian domes. Petrofabric analyses of Acadian porphyroblasts and deformational fabrics, however, unequivocally show that peak-temperature metamorphic mineral growth occurred after the majority of dome-related deformation. Thus, the geometry, or distribution, of the isograds is in part a function of the distribution of bulk compositions amenable for staurolite-kyanite production (at appropriate P-T) rather than folding of isogradic surfaces subsequent to their formation.

Quantitative estimates of pressure (P) and temperature (T) during Acadian metamorphism have been calculated at several places around the Chester and Athens Domes. Although the number of P-T points is small, it is known that the peak-T (and related P) conditions increase from south to north and west to east across the Athens Dome into the Chester Dome.

Note in figure 9 that the distribution of P-T points as well as the related metamorphic isograds cuts across the dome-related structures, including the Spring Hill syncline and the Chester Dome. This indicates that these different structures were metamorphosed under similar conditions and at similar crustal levels. Therefore we conclude that the variation in crustal level, recorded by the metamorphic P-T conditions, is a consequence of the crustal load that was situated above the Chester Dome and Spring Hill syncline and not relict or folded isobars and isotherms.

Acadian ilmenite-garnet-biotite-chlorite-white mica-plagioclase-quartz assemblages found within pelites of the Rowe Schist (OCr fig. 1) west of the Athens Dome and ilmenite-garnet-biotite-staurolite-white mica-plagioclase-quartz assemblages (some containing kyanite) within rocks of the Moretown and Cram Hill Formations around the Spring Hill synform, immediately west of the Chester Dome, indicate increasing metamorphic conditions from west to east (fig. 9). Kyanite and staurolite along with garnet porphyroblasts overgrow Acadian penetrative schistosity and dome-stage crenulate cleavages within the Spring Hill structure. The staurolite-in isograd (and kyanite-in for appropriate bulk compositions) cross-cuts the Spring Hill synform and along with mineral-fabric relative age relationships, indicates that attainment of peak temperature conditions occurred after the major phases of dome-stage deformation. This relationship is internally consistent with our structural model (C-C', fig. 6) which shows that the low-amplitude structures do not bring significantly deeper crustal levels into juxtaposition with shallower crustal levels. This also implies that the dome is not the cause of the gross distribution of metamorphic isograds, isotherms or isobars.

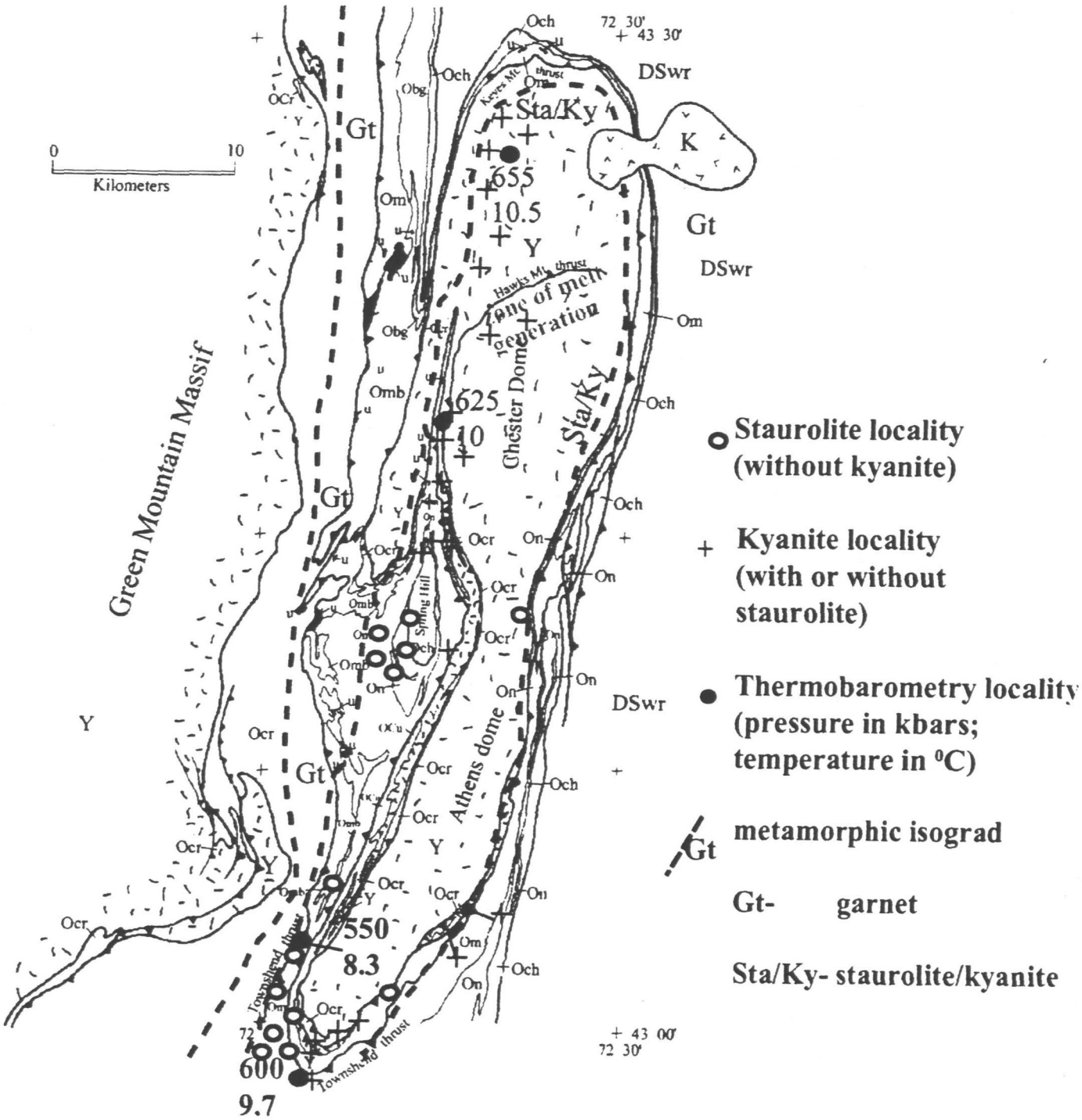


Figure 9. Metamorphic map of the Chester and Athens domes, showing the distribution of the garnet (Gt) and staurolite-kyanite (Sta/Kya) isograds, localities for thermobarometric estimates discussed in text (dark circles) and sample localities containing staurolite (open circles) and kyanite (with or without staurolite; crosses). "Zone of partial melting" refers to area within core of the Chester Dome that contains exposures with kyanite-plagioclase-K-feldspar leucosomes formed by dehydration melt reactions related to Acadian metamorphism. Other symbols on geologic map are explained in Figure 1.

The amplitude of these structures, that we have determined must have been very small in comparison to the thickness of the overlying tectonic load that caused the regional (Barrovian) metamorphism during the Acadian orogeny.

Petrography and Thermobarometry

Thin sections from aluminous pelite in the Gassetts Schist Member of the Cavendish Formation and pelite from the Feldspathic Schist Member of the Cavendish Formation were sampled from the west flank of the Chester Dome (at the type- Gassetts Schist locality) and from Star Hill, in the core of the Chester Dome. Each of these samples contains four or five primary AFM phases, including garnet, biotite, staurolite, kyanite, and in some cases, primary chlorite or chloritoid. In several samples chloritoid is the sixth AFM phase, but only occurs as inclusions within garnet and staurolite porphyroblasts. Garnets from both the Gassetts Schist and the Cavendish Formation show some evidence of two phases of metamorphic growth, which have been previously interpreted to represent Taconian cores with Acadian overgrowths about an unconformable surface (Karabinos, 1984) or possibly two phases of continuous Acadian garnet growth punctuated by a discontinuous phase of garnet resorption (Armstrong and Tracy, 1991). Microprobe analyses, however, have shown that some of the large garnets from the Star Hill locality (Stop 6) are actually "amalgamated" garnets, with relict angular garnet fragments overgrown by a continuous phase of "younger" garnet, forming a single, 1 to 2 cm diameter garnet porphyroblast. Energy dispersive spectrometry of micron-scale inclusions from the early garnet fragments has shown that they are composed of rutile and an aluminum-silicate, which optically, appears to be sillimanite. Sillimanite-rutile assemblages are not found in any of the pre-Silurian cover rocks west or east of the dome, but are present within Middle Proterozoic schists and gneiss within the Green Mountain massif, where the ambient Paleozoic metamorphic grade is Taconian biotite-grade (Ratcliffe, unpublished data). Thus, the sillimanite-rutile assemblages are likely the result of Grenville (Middle Proterozoic) upper-amphibolite to granulite Facies metamorphism. This leads to the conclusion that garnet schists at Stop 6 on Star Hill could be Middle Proterozoic.

Garnets from the Gassetts road cut (optional Stop 10) and Star Hill (Stops 6 and 7) areas display apparently randomly oriented and sized inclusions of margarite, chloritoid, quartz, muscovite, staurolite, ilmenohematite, rutile, and magnetite. The position of different inclusion types with respect to garnet core and rim suggests two phases of garnet growth and one intermediate phase of garnet consumption during the discontinuous reaction:

garnet + chlorite = staurolite + biotite

According to Thompson and others (1977), this discontinuous reaction occurs at a temperature around 590°C (at $P=PH_2O=6$ kbars). Continuous garnet growth could occur at higher temperatures following the total breakdown of chlorite or intersection of a garnet producing discontinuous reaction. The latter occurs within this system (6 kbars) at a temperature of 680°C:

staurolite = garnet + biotite + sillimanite

Since sillimanite is not present within the matrix or outer growth of (Acadian) garnet in the Gassetts Schist, the temperature necessary for this reaction (680°C) was not achieved or pressure was sufficiently higher than 6 kbars such that kyanite was the stable aluminum-silicate phase. At 680°C, X_{FE} of biotite is about .60, whereas during a continuous reaction (at lower temperature) it would be significantly lower. Biotites analyzed from both Gassetts and Star Hill have X_{FE} (Fe/Fe+Mg) around .35. Furthermore, kyanite is present with staurolite, biotite, and muscovite but is not found in contact with garnet. Kyanite development probably occurred by way of the discontinuous reaction:

staurolite + chlorite = biotite + kyanite

This reaction occurs at about 625°C (P=6 kbars; Thompson and others, 1977). The earlier described garnet-consuming reaction (garnet + chlorite = staurolite + biotite) would terminate with the consumption of all chlorite; the kyanite producing reaction requires chlorite.

Multiple garnet and biotite samples were selected from the feldspathic granofels unit of the Gassetts Schist Member of the Cavendish Formation for microprobe analysis. Color contoured, qualitative x-ray maps for Mg, Fe, Ca, and Mn distribution were produced for garnets from the Gassetts Schist samples.

Gassetts Schist samples show a progressive decrease (from core to rim) in the grossular, spessartine, and almandine components with a corresponding increase in pyrope. Locally, some garnet rims show a slight increase in almandine but all garnets consistently decrease in Fe/Fe + Mg from core to rim. Although Thompson and others (1977) reported numerous staurolite inclusions within part of garnet rims where almandine increased, the particular Gassetts samples analyzed in this study did not contain staurolite inclusions.

Garnet-biotite compositions from both the Gassetts and Rowe Schist samples from the Spring Hill area, were used in obtaining temperatures of equilibration for garnet rim and associated biotite. Two models were used in acquiring temperatures; Fe-Mg-Mn activity model, and the Ferry and Spear (1978) almandine-pyrope model. Temperatures were reported using the Ln Kd garnet-biotite equation of Thompson (1976) and the Ln K equation of Ferry and Spear (1978); these equations resulted in a 10 to 50°C temperature variation for a particular analysis. A range of temperatures for each pair was developed by varying the pressure value from 6, 7.5 and 10 kbars for each analysis. This range of pressure is within the range of values reported for this area by Downie (1982; 6 kbars) and Kohn and Spear (1990; 10 kbars). Variation in temperature as a result of maximum and minimum pressure estimates did not exceed 25°C for any analysis. Temperature estimates from the granofels of the Cavendish Formation ranged from 557°C to 682°C; temperature estimates from the Gassetts were between 478°C and 698°C.

Phase equilibria analyses, using the TWQ program of Berman (1991), yielded multiple linearly independent equilibria whose intersections in P-T space gave the following temperatures and pressures: granofels unit of the Cavendish Formation, 655°C (± 55), 10.5 kbars (± 1.5); Gassetts Schist Member, 625°C (± 65), 10.0 kbars (± 1.8); Rowe Schist at Townshend Dam, 550°C (±35), 8.3 kbars (± 1.0); and Rowe Schist from the southern closure of the Athens Dome, 600°C (±44), 9.7 kbars (±1.1). The locations and values of temperatures from these samples, and others reported in Armstrong and others (1992), and Armstrong and Hames (1993) are shown in Figure 9. The large variation in the temperatures from the Gassetts and Cavendish samples is a consequence of variation in garnet rim composition and biotite Fe/Fe+Mg. Garnet composition contour maps in Thompson and others (1977) and Armstrong and Tracy (1992) show severe truncation of isopleths along many of the garnet rims in the Gassetts Schist. In addition, inclusion-free garnet rims observed in samples studied by Armstrong also truncate along regular garnet boundaries. Low-power microscopy shows that these regular boundaries are actually fractures and that many of the garnets are either pieces of a large garnet, possibly Middle Proterozoic or Acadian metamorphic overgrowths upon the fragments. Garnet rim analyses taken from fracture boundaries (on the relict garnet fragments) yielded erroneously low temperature values. Analyses taken from inclusion-free, Acadian garnet rims yielded more consistent temperatures, in the range of 605°C to 695°C.

Petrogenetic Evidence for Temperature Estimates

Temperature values obtained using garnet-biotite thermometry and estimated pressure ranges coincide with P-T estimates obtained from well constrained continuous and discontinuous reaction curves.

The 550°C average for the garnet-biotite-chlorite assemblage from the Moretown Formation (Townshend Dam locality, southwestern part of the Athens dome; Figure 9) is below the staurolite-in reaction:

garnet + chlorite = biotite + staurolite

that occurs at approximately 590°C (at 6kbars). Peak temperature (550°C), however, may have subsequently occurred during a period of isothermal decompression to somewhat more moderate pressure. Thus the corresponding pressure estimate for this sample may be significantly below that experienced by the rocks at peak pressure. This type of P-T path is characteristic of thermal relaxation following tectonic loading in regions of low magmatic activity (Thompson and England, 1984). Gassetts assemblages are better constrained by reaction space. Early, low-grade garnet-chloritoid-chlorite assemblages, upon

continuing prograde conditions, would have given rise to garnet-chlorite-staurolite through the continuous reaction:

chloritoid = garnet + staurolite + chlorite

illustrated in Fe-Mg-Mn space. Continued prograde conditions continues with the continuous reaction:

chlorite + staurolite = Mg richer garnet

until the discontinuous staurolite-in reaction:

garnet + chlorite = staurolite + biotite

Chlorite-rich compositions would result in garnet elimination and continued prograde conditions until the discontinuous reaction:

staurolite + chlorite = biotite + kyanite

This reaction, with the four AFM phases present in the Gassetts samples studied, occurs at about 625°C (at 6 kbars). This coincides well with the acquired garnet-biotite temperatures around 625°C. Chlorite-poor garnet-chlorite-biotite compositions would eliminate chlorite upon reaching the staurolite-in reaction. Continuous garnet producing reactions (toward Mg-richer garnet) would presumably continue to the discontinuous reaction:

staurolite = garnet + biotite + kyanite

at a temperature around 735°. This reaction is above the muscovite breakdown reaction for reasonable pressure estimates (710°C at 6 kbars), and was obviously not achieved in this region. This is supported not only by the abundance of muscovite (and lack of kspar) within these rocks but also by the absence of coexisting adjacent garnet and kyanite.

The presence of significant Acadian pegmatoidal granite and plagioclase-kspar-kyanite leucosomes in schist from the Mount Holly Complex (unit Yrg in fig. 4) from the core of the Chester Dome ("zone of melt generation" in Figure 9) suggests attainment of partial melting at elevated P-T conditions. Although no P-T data has yet been attained from these rocks, it is likely that they will yield P-T values higher than those from the kspar-absent samples of Cavendish Formation from Star Hill (655°C, 10.5 kbars). This is consistent with the structural model presented here, and by Ratcliffe (1995b, c; in press) which show these rocks in the deepest structural level of the Chester Dome (see sections B-B' and C-C', fig. 4).

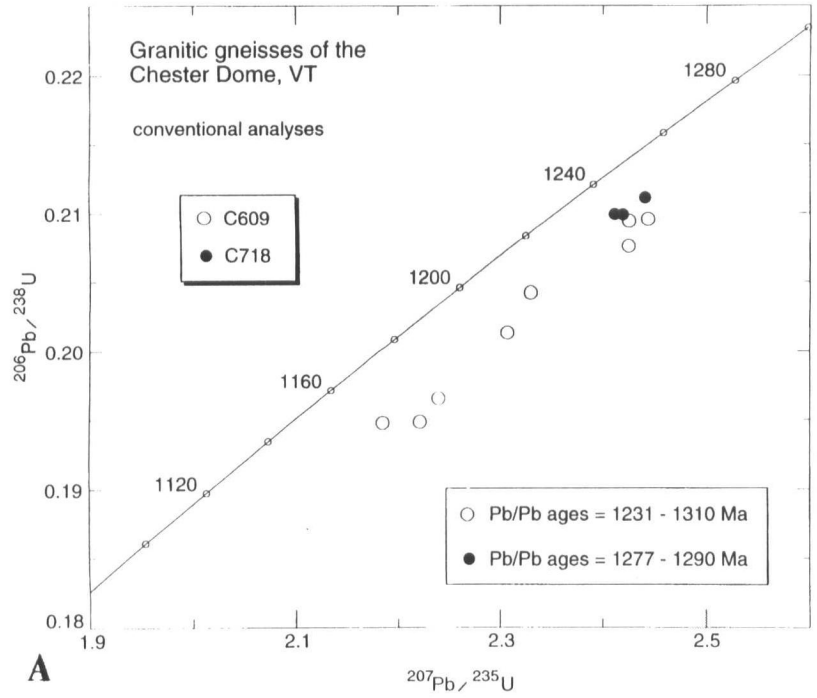
Geochronology

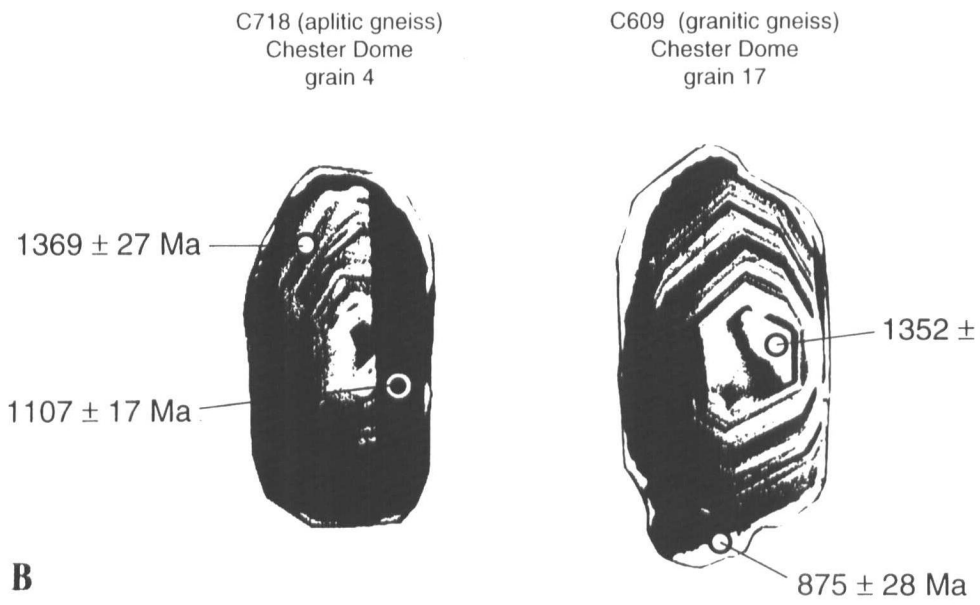
Zircons from two samples of Felchville Gneiss (specimens C609 and C718) were analyzed for U-Pb geochronology. Conventional analyses of splits produced very discordant data (fig. 10A). Ion microprobe (SHRIMP) analyses of growth zones and overgrowths from grains like the two illustrated in figure 10B illustrate the complexity of these zircons and reveals the explanation for the discordant data is figure 10A. Zircons for Middle Proterozoic gneisses have undergone a complex history with times of partial dissolution and(or) subsequent overgrowth. Grain C718 has a dramatically defined clear overgrowth whose $^{207}Pb/^{206}Pb$ age is calculated as 1107±17 Ma (probably a time of Grenvillian metamorphism), while a euhedral growth zone has as an age of 1369±27 Ma. A similar grain from C609 has a core age of 1352±44 Ma and an overgrowth age of 875±28 Ma. Grains analyzed by the conventional U-Pb technique are physical mixtures of multi-age zircons.

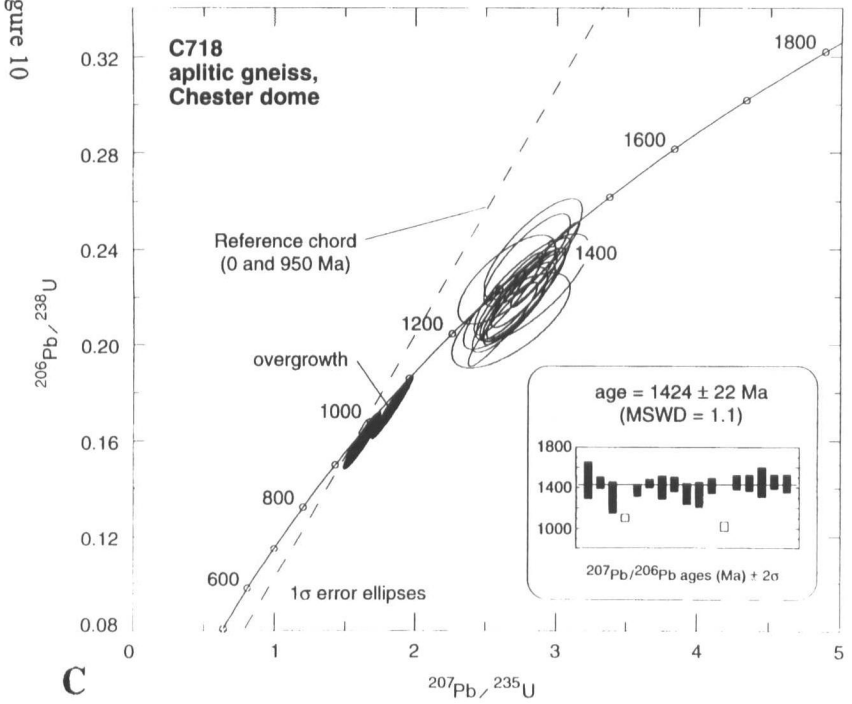
SHRIMP ages from the euhedral growth-zoned parts of the zircons yield $^{207}Pb/^{206}Pb$ ages of 1424±22 and 1379±23 Ma (2 σ weighted average of $^{207}Pb/^{206}Pb$ ages; fig. 10C and D). Overgrowth ages are 950, 1100 and 1200 Ma. From previous work, these are know times of thermal activity in the Middle Proterozoic of Vermont.

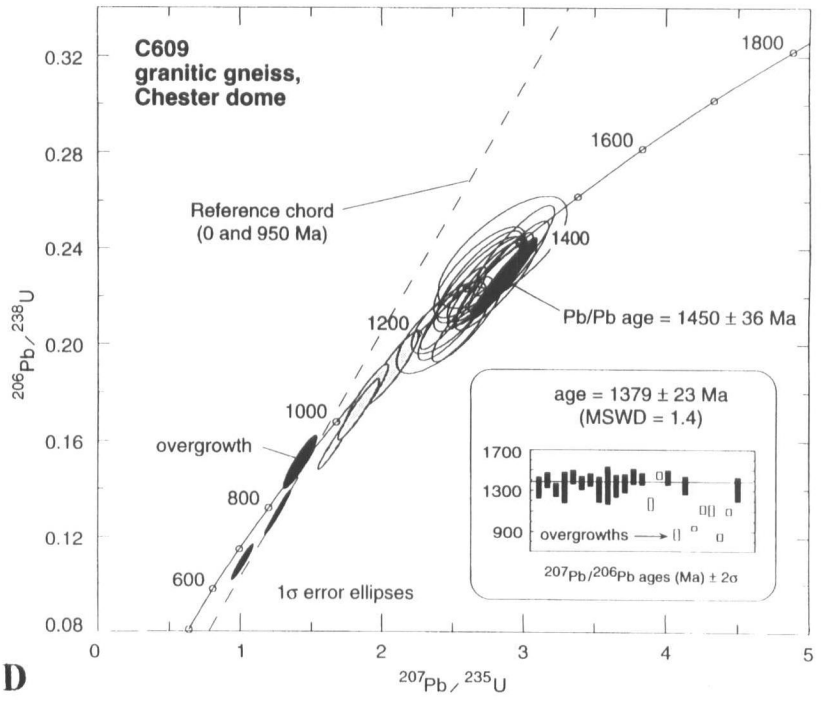
Table 1. U-Pb geochronology of rocks of Mount Holly complex discussed on Trip B-6, ages determined by J.N. Aleinikoff

Sample No.	Rock	Age (Ma)	Comments
VT/Lu 1-91	leucogneiss, Ludlow Mtn. Ludlow quadrangle	1314 ± 7	SHRIMP age, intrudes paragneiss and garnet schist of Mount Holly Complex like Gassetts Schist Member of the Cavendish Formation
C609	Felchville Gneiss Chester dome Cavendish quadrangle	1379 ± 23	SHRIMP age, intrudes paragneiss units of the Mount Holly Complex
C718	Felchville Gneiss from Cavendish Gorge in Cavendish quadrangle Chester dome	1424 ± 22	SHRIMP age, intrudes schists and gneisses of Mount Holly Complex and Cavendish Formation









B-6 -- Figure 10

Figure 10. U-Pb geochronology of the Felchville Gneiss Cavendish quadrangle (see Geochronology section in text)

Table 2. U-Pb geochronology of Paleozoic rocks in southern Vermont discussed on Trips B-6 and C-6 ages determined by J.N. Aleinikoff

Sample No.	Rock	Age (Ma)	Comments
VT/Pl 2-91	2-mica granite in Pinney Hollow Formation in Plymouth quadrangle has xenocrystic Middle Proterozoic zircon	~370	conventional and SHRIMP data [only 1 grain has Acadian age], cross cuts Taconian structure, has Acadian F_3 crenulation cleavage
VT/Br 1-89	Black Mountain granite in core of Guilford dome	373 ± 4	conventional age, foliated granite
Vt/Br 2-89	Trondhjemite of North River Igneous Suite South Newfane	462 ± 6	SHRIMP age, intrudes rocks of Cram Hill and Moretown Formations
SP450	Cram Hill felsic volcanics of Cram Hill Formation near Springfield	484 ± 4	SHRIMP age, layered with mafic volcanics and schist of the Cram Hill Formation
SR3053	tonalite, North River Igneous Suite (of Armstrong, 1994) east side of Chester dome Saxtons River quadrangle	486 ± 3	conventional age, intrudes Moretown Formation
L18	Trondhjemite of Barnard Gneiss of Richardson (1924) Ludlow quadrangle	496 ± 8	SHRIMP age, intrudes Moretown Formation and Proctorsville ultramafic belt and possibly rocks of the Cram Hill Formation

SUMMARY AND CONCLUSIONS

Detailed mapping of the core gneisses of the Chester and Athens domes, in the Cavendish (Ratcliffe, 1995a; in press) Andover (Ratcliffe, 1996b, c), Chester quadrangle (Ratcliffe, 1995b, c; in press), and Saxtons Rivers quadrangle (Ratcliffe and Armstrong, 1995, 1996) indicates that the rock of the Mount Holly Complex closely resemble the Mount Holly of the type area in the Green Mountain massif (Ratcliffe, 1992). The tonalitic to trondhjemitic gneisses of the Baileys Mills Tonalite Gneiss and the Felchville Gneiss closely resemble, both in age and chemical composition, an unnamed suite of metaigneous rocks in
the Green Mountain massif (Ratcliffe and others, 1991). It is noteworthy that these gneisses appear at about the same latitude in both areas and that the occurrences in the domes appear to be an eastward extension of the same east-west belt found in the Green Mountain massif. This relationship suggests that the general east-west structural grain of the Middle Proterozoic rock is consistent for the entire region.

Restudy of the Cavendish Formation in both the Cavendish and Chester quadrangles suggests that these schistose and feldspathic rocks, once considered part of the cover sequence are best interpreted as part of the Mount Holly Complex. Intrusive relations between the Felchville Gneiss and the Cavendish Formation in the Cavendish quadrangle (Ratcliffe, in press) and the U-Pb and $^{207}Pb/^{206}Pb$ zircon ages of the Felchville Gneiss of metasedimentary rocks of the Cavendish Formation are approximately 1.4 Ga or older. The strong physical resemblance between rocks of the Cavendish Formation and widespread chloritoid-chlorite-muscovite-quartz±garnet schist and associated dolomite marble, and calc-silicate rocks in the Mount Holly Complex throughout the Green Mountain massif further strengthens this interpretation.

The internal structure of the Chester and Athens domes indicates that the major folds and distribution of rock units, to a significant extent, predate the regional and strong Paleozoic foliation in the core of the dome. Mapping indicates that the regional Paleozoic foliation is not axial planar to isoclinal and once recumbent folds of large amplitude.

Detailed examination of the Late Proterozoic, Cambrian and Ordovician cover rocks of the dome does not indicate a coherent stratigraphic package, but requires significant internal folding and faulting to explain the distribution of the rock units. A major Taconian fault is proposed to underlie the rocks of the Moretown Formation. The rocks between the Moretown Formation and the core gneisses of the dome are highly tectonized and only locally are original sedimentary contacts preserved. The minor structures, mylonitic foliation and reclined isoclinal folds within the zone are interpreted as Taconian features, consistent with east-over-west imbrication of basement and cover rocks generally exposed along the western margin of the Chester and Athens domes in the Saxtons River and Townshend quadrangles (Ratcliffe and Armstrong, 1996; Ratcliffe and others, 1992; Ratcliffe, 1997).

Formation of the Chester and Athens domes in the Acadian orogeny was the result of multiple compressional fold events and interference among those sets of folds. The peak of the regional metamorphism in the Acadian produced garnet inclusion structures that indicate garnet grew during the later crenulation or F_4 event along the eastern margin of the dome. In this area and within the dome, zones of late shortening occur in localized areas; the deformation of Acadian granite dikes is the greatest in these zones. Comparable but lesser shortening is present along the western margin of the dome where garnet growth also accompanied compression. Rocks within the core of the dome are broadly arched and Devonian granite dikes only weakly deformed there. These data suggest that the level of Acadian strain was greatest along the flanks of the dome and is consistent with an entirely compressional origin for the dome in the Acadian Orogeny.

ROAD LOG

Trip B-6 Saturday Trip

ASSEMBLY POINTS PICO SKI AREA PARKING LOT 7:30 AM

SECONDARY ASSEMBLY POINT ECHO LAKE INN PARKING LOT, TYSON, VT ON RT. 100 (8:00 AM)

Cumulative mileage	
0.0	Road log starts at the intersection of Routes 4 and 100 at Sherburne (for geology to Tyson see Trip C-6)
	Geology along route followed is available in Walsh and Ratcliffe (1994a, b) Walsh and others (1994) and road log of trip C-6
1.7	Tyson at Echo Lake Inn turn left on Boswell Pond Rd.
1.8	Y intersection bear left
3.8	The Alps out crops of Barnard Gneiss (of Richardson, 1924)
5.1	Weld Cemetery Road and Tuttle St. crossroads, now in Cavendish quadrangle
5.6	left turn on Puckerbursh Rd.
6.5	Cross Bartley Rd.
6.8	Top of hill
	For location of stops, see figures 1 and 4.

STOP 1. WAITS RIVER FORMATION WEST FLANK OF CHESTER DOME AND FOLD STYLES WITHIN THE DEVONIAN AND SILURIAN COVER ROCKS. (10 MINUTES) Punky- and gray-weathering quartzose limestone layers in the Waits River dip 40° NW, and strike N70°E. The prominent schistosity, expressed by coarse-grained biotite, is parallel to bedding. F_1 folds are rarely seen but the hinge lines are subhorizontal to gently north plunging and commonly show up from the west, normal drag. The bed-parallel schistosity has a crinkle lineation that plunges N15°E. There are two coaxial crenulation cleavages; the stronger dips west more gently than the schistosity and produces up-from-the-east microfolds, the weaker dips steeply southeast. The acute bisectrix of the crenulation cleavage is subvertical, the obtuse bisectrix subhorizontal. The schistosity and a crenulation cleavage are both folded over the dome. Garnets from other localities have inclusion textures that suggest that rims of the garnets, grew late enough to include microfolds of the crinkled schistosity related to both crenulation cleavages. Here the garnets have ½ to 1/3 of their outer rims replaced by plagioclase.

The photograph (fig. 11) shows minor folds in the Waits River Formation from an outcrop near here within 5 meters above the RMC. The view is to the north. S_0 (bedding) and schistosity (S_1) are parallel and folded by tight F_2 folds. In one area S_1 crosses S_0. F_1 and F_2 plunges are to the north. The fold patterns indicate very nonhomogeneous strain; the later deformation is irregular in orientation and intensity. Note that some F_2 domainal cleavage is subvertical, in other areas it dips west. The straight eastern limb of the western synform preserves S_0/S_1 intersection geometry. The low level of strain and the composite nature of the deformation is characteristic of the Waits River Formation and of the Acadian strain on the west limb of the Chester dome. If viewed in reverse (from the north) the tight synform on the east is a reasonable proxy for the Proctorsville syncline, and the domal area to the west mimics somewhat the form of the Chester dome!

	Turn around and return to Bartley Road
7.4	Turn left on Bartley Road, proceed 0.2 mile and park near contact between Waits River and schists of the Cram Hill Formation and Barnard Gneiss (of Richardson, 1929)

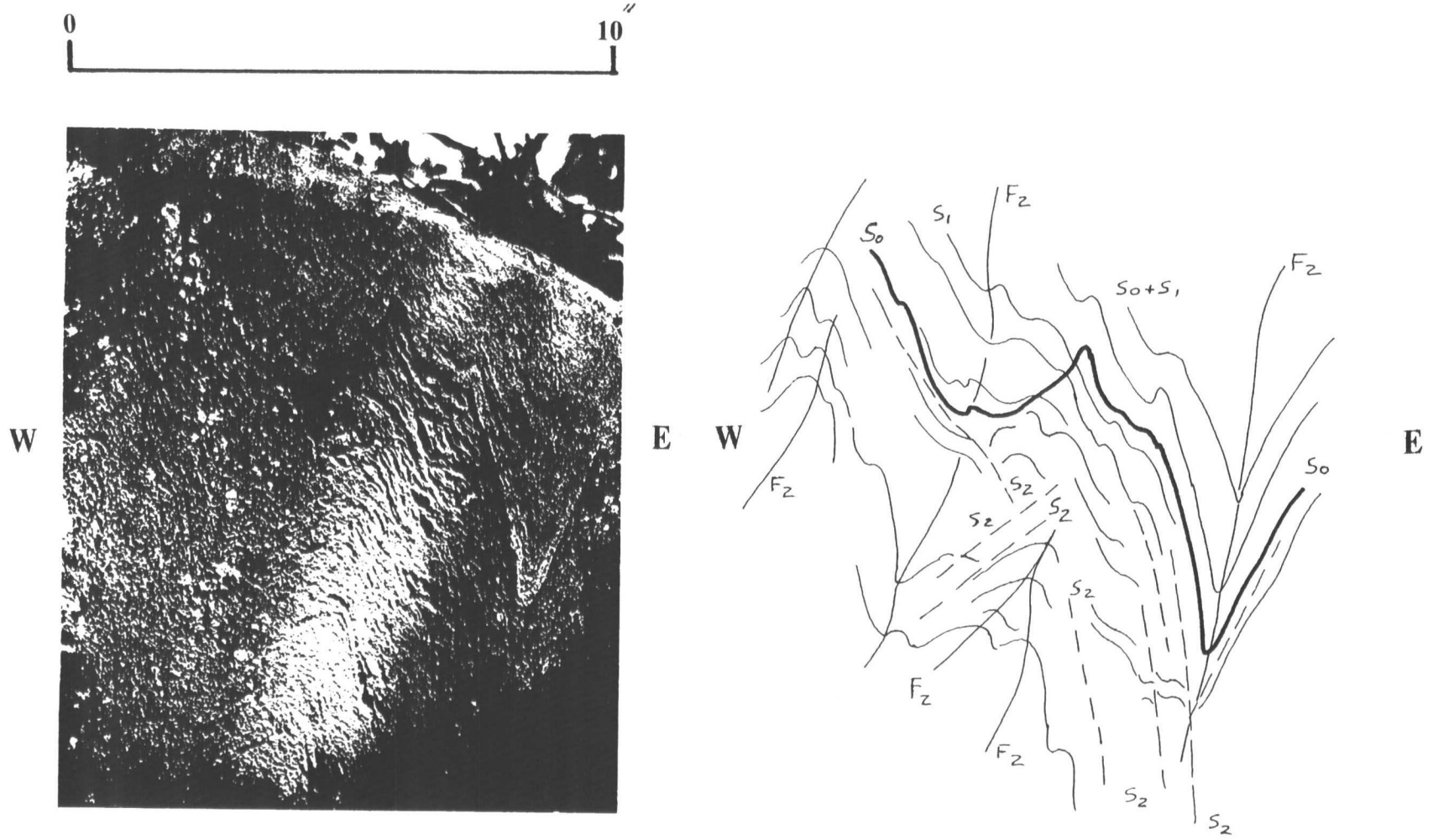


Figure 11. Minor folds in a punky weathering limestone layer near the base of the Waits River Formation on the west limb of the Chester dome near Stop 1. View to north. S_0 bedding and S_1 (schistosity) are parallel except on east limb of western syncline. Note the composite nature of F_1 and F_2 structure, and apparent unfolding of S_0 on the east limb of the small western syncline. Plunges of F_1 and F_2 folds are gentle to north. These minor folds illustrate the low and heterogeneous strain in the Silurian and Devonian rocks. In general this structure mimics the antiformal Chester dome and tightly pinched Proctorsville F_2 syncline, when viewed from the north.

STOP 2. RICHARDSON MEMORIAL CONTACT. (10 MINUTES) This quick stop illustrates the Acadian structure along the outer mantle of the dome and compares deformation in the pre-Silurian rocks to that in the Waits River Formation above the RMC. The first crops of Waits River contain two reasonably well developed north plunging, southeast verging F_1(?) minor folds; a crenulation lineation plunges to the north, here related to the steeply east-dipping late crenulation cleavage. Walk downstream 50 feet to blocks of well-layered felsic and mafic gneiss of the Barnard Gneiss (of Richardson, 1924) . Although the large block is slightly detached and rotated, the Acadian folds of the gneissosity are clear, and a strong southwest intersection lineation between Acadian S_1 and gneissosity in the Barnard plunges southwest not northeast. This kind of variation in plunge occurs at many places across the boundary. The pre-Silurian rocks have structures and schistosities that are discordant to the sheet dip of the dome, although the foliations in both are sensibly parallel. Downstream rusty schists of the Cram Hill Formation or Moretown Formation are exposed. These two short stops should give you an appreciation for the intensity and character of the Acadian deformation that can be compared to the additional complexity in the gneiss in the core of the Chester dome seen at the following stops.

8.4 South Reading

10.4 The "twenty foot hole," a swimming hole excellent exposures of Mount Holly gneiss. Subvertical gneissic layering and two pegmatites trend north from the stream into the yards of the houses north of the road. This is a fine location to see the general trend of the gneissosity (north northeast and subvertical) and the transecting foliation that trends east-west and dips to the north.

10.6 **STOP 3. FELCHVILLE GNEISS** (biotite schlieren, inclusion of hornblende garnet amphibolite) (5 MINUTES) New road cut of biotite trondhjemite gneiss. Gneissosity is folded into a large east verging antiform; the arched Paleozoic axial plane foliation trends east-west to northeast and dips north and northwest; the general trend of folded gneissic layering is northwest and north. Hornblende-garnet gneiss and amphibolite seen here extend north and south from this area for 2 km; the adjacent rusty schist and trondhjemite gneiss parallel this belt to the basement cover contact north of route 44. This is a good demonstration of dominance of the Grenvillian trends in controlling map patterns in the core of the dome, and the relative insignificance of the Paleozoic folds in affecting map patterns. Note how the mapped units trend across the axial traces of the proposed Butternut Hill and Star Hill nappes in figures 2 and 3.

11.1 Curve in road exposing excellent arched Paleozoic axial plane foliation and oppositely verging folds of gneissosity in the Ybg unit of the Mount Holly. These folds are Paleozoic folds (Taconian or Acadian and typical of the amplitudes of pre-dome folding in the core of the dome. Spectacular folds like this are only minor folds that do not affect map patterns in the Mount Holly greatly. The units in the Mount Holly trend north-northeast across the dome at an angle to the east-west trending and arched schistosity.

11.5 Reading (Felchville) intersection 106 turn left

12.5 Intersection Rt. 44 and 106

STOP 4. FELCHVILLE GNEISS AND ZIRCON SAMPLE SITE. (10 MINUTES) These road cuts constitute the type area for the Felchville Gneiss (Ratcliffe, in press). The Felchville Gneiss varies from trondhjemite to granodiorite and intrudes all paragneiss of the Mount Holly Complex. Near contacts the rock is finer-grained, aplitic and trondhjemitic in composition. It encloses calc-silicate rocks of the Cavendish Formation and occurs indiscriminately within or between schist units of the Cavendish. Ghost-like xenoliths of various units of Mount Holly Complex occur within it.

The belt here extends from near the northeastern end of the dome southwestward to the southwest corner of the Cavendish quadrangle and from that point south into the core of the Butternut Hill fold in the Andover quadrangle (fig. 4). Notice that the outcrop belt crosses all the form lines of foliation that define the internal shape of the dome (fig. 5). In cross section A-A' (fig. 6) the Felchville Gneiss extends from a point just above the Cavendish at Hawks Mountain to the roof of the dome.

In this outcrop the axial surface of Paleozoic folds trend east-west and dip north. These folds are the minor folds shown in section A-A' (fig. 6) and refold the axial trace of the older Grenvillian fold that runs diagonally the length of the Cavendish quadrangle and beyond.

Zircon from this outcrop yields a 1379±23 Ma SHRIMP age (table 1 and fig. 10) that we interpret as the crystallization age of the Felchville Gneiss.

Because the Felchville Gneiss intrudes all paragneiss of the Mount Holly the deposition of those rocks are interpreted to be greater than about 1379 Ma. In our interpretation this rock and the paragneisses of the Mount Holly Complex are the oldest rocks yet discovered in the Laurentian crust of the Appalachian.

Continue north on Route 106
13.0 Flat Iron Road, crops of rusty gneiss and schist of the Mount Holly in brook west of bridge
13.3 Crop of amphibolite gneiss (Ya) and biotite gneiss (Ybg) of Mount Holly continuous from Stop 3
13.6 At bend in road, road cuts, park

STOP 5. (Optional Stop) TRAVERSE FROM CORE ROCKS OF DOME UP TO COVER ROCKS. (40 MINUTES) This traverse shows the increase in strain in basement rocks as the Keyes Mountain thrust is approached from the core of the dome. Crops by road and above are granitic gneiss and trondhjemite gneiss, passing up into migmatite gneiss. All rocks contain abundant highly deformed pegmatites. The foliation strikes approximately east-west and dips to the north. Hingelines of isoclinal folds plunge N70° E to N30°E and show both clockwise and counterclockwise rotation senses. Gneisses having abundant and more increasingly deformed pegmatite and augen, continue to about the 1260 foot elevation where a thin 10-m-thick amphibolite occurs. This amphibolite mapped as part of the cover can be traced continuously along the slopes for 2 km to the east and 1 km to the west. Above this a lustrous muscovite-chlorite-garnet-quartz schist about 10 m thick that has chlorite pseudomorphs after biotite, or large cross-biotite. Slabby, orangish gray feldspathic granofels that contains chlorite pseudomorphs after fasicules of hornblende forms the cliffs between 1300' to 1400' elevation. The rocks from the amphibolite upwards have been assigned to the Moretown Formation that here directly overlies the core gneiss of the dome.

The contact is remarkably sharp, easily mapped and absolutely parallel to the mylonitic fabric in the rocks associated with the Keyes Mountain thrust. Cover rocks corresponding to the Hoosac and Pinney Hollow, Ottauquechee are absent.

Turn around, follow 106 south to (Reading-Felchville)
16.2 Felchville
16.9 Turn right on Knapp Brook Rd.
18.1 Road branches, take left branch
19.4 Greenbush Rd. on left, bear right
21.2 Turn right on Old County Road
22.4 Powerline crossing (Optional Stop)

STOP 6. TRAVERSE ACROSS GASSETTS SCHIST MEMBER OF CAVENDISH FORMATION. The first crops in the powerline are muscovitic-biotite-garnet-staurolite-kyanite schists of the Cavendish Formation. Continue walking west to the crest of the second ridge where subvertical coarse-garnet schist of the Gassetts Member of the Cavendish crops out. A pegmatite in the schist is exposed. Subhorizontal late pre-dome schistosity cuts both rocks (fig. 8A). Highly foliated granitic dikes like this are typical of the Mount Holly and indicate that the schist here is in fact schist in the Mount Holly Complex, and not cover rock as suggested by Thompson (1950) and by Thompson and others (1993). In their interpretation this belt of schist is correlated with the Pinney Hollow Formation of the Eastern Vermont Sequence.

Continue north on road to Star Hill

22.6 Parker Road on left, bear right

23.1 Approaching field edge, houses on right, small shed on left. At east side road aplite gneiss (Yap) of the Mount Holly Complex, cliff to west is coarse garnet schist of the Cavendish Formation (Gassetts) about 50 feet thick and is overlain by well layered biotite-quartz plagioclase gneiss of the Mount Holly that contains abundant small pegmatite pods and aplite sills (Yp, Yap) that are too small to map. This sequence is overlain by a thin, second (30-feet-thick) layer of coarse Gassetts schist. Up the hill from this point are abundant crops of aplite gneiss (Yap) in well layered biotite quartz-plagioclase gneiss, pegmatite, and a third layer of schist (Yrs) intruded by a large mass of granitic gneiss and pegmatite at the top of the hill. This traverse is very instructive and demonstrates that this sequence of rocks, including the two belts of the Gassetts Schist member of the Cavendish Formation, belong in the Mount Holly Complex. Downie (1982) referred to these rocks as Hoosac? and noted the resemblance of the interlayered gneisses to normal Mount Holly. We agree with her that they are in fact rocks of the Mount Holly Complex.

23.3 Continue 0.2 farther up hill past houses and park

STOP 7. STAR HILL AND GASSETTS SCHIST MEMBER OF THE CAVENDISH FORMATION and MOUNT HOLLY COMPLEX GNEISSES. (15 MINUTES) This is one of the best places to see the relationship between gneisses of the Mount Holly Complex and coarse garnet schists of the Gassetts Schist Member of the Cavendish Formation. The relationship are clear and all of these rocks are part of the Mount Holly Complex and not cover rocks as proposed by Thompson (1950). Downie mapped this area for her Ph.D. dissertation at Harvard (Downie, 1982). The generalized interpretation shown in figure 2 and 3 is less complex than shown in her thesis where she proposed early Taconian inversion followed by two stages of Acadian folding.

The traverse starts east of the road in coarse foliated pegmatite of the Mount Holly Complex that crops out 100 feet east of the road, below a small cellar hole, excavated parallel to joints, that exposes feldspathic Gassetts Schist. Note the small layer of albite-biotite muscovite feldspathic schist at the lower contact of the pegmatite east of the road. The Gassetts extends across the road where it is overlain by well layered biotite-quartz-plagioclase gneiss that has a lens of coarse-grained calcite marble and well layered biotite quartz-plagioclase diopside calc-silicate in a zone about 25 feet thick. Abundant veins and sills of aplitic gneiss occur within this rock. This rock is absolutely typical of calc-silicate gneiss found throughout the Mount Holly Complex. This is overlain by a second layer of garnet schist mapped as (Yrs on the digital open file report (Ratcliffe, 1995a). On the digital open file report (Ratcliffe, 1995a) the Yrs unit (colored pale bluish green was inadvertently omitted from the simplified Description of Map Units. This rock is the same lithologically as the Gassetts (Ycg); also I find to my chagrin that the system of roads taken from the digital files of the Vermont Center for Geographic information is not properly registered relative to the geology. This will all be rectified on the printed geologic map (Ratcliffe, in press).

On the cliffs from this point south to crops up hill from mileage 23.1 pegmatite occurs in the Ybg unit and the schist (Yrs) all along the hill. Walk up the utility line that runs N60°W up Star Hill. At the base is Gassetts Schist, followed by layered biotite gneiss and pegmatite, and up the hill is the second Gassetts Schist. A pegmatite occurs within the schist 50' south of the powerline followed by a vertical cliff of Gassetts that contains small pegmatites and a gneiss layer. This collection of rocks resembles closely the rocks found in the Mount Holly Complex throughout the Green Mountain massif and in the core of the Chester and Athens domes. We correlate the rocks here with aluminous schistose rocks exposed at the following locations: Blue Ridge Mountain (Chittenden quadrangle), Ludlow Mountain (Ludlow quadrangle), Old Job (Danby quadrangle), Wilder Mountain (Danby quadrangle), College Hill (Jamaica quadrangle), and Greenbriar Brook (Weston quadrangle) to mention a few. At all of these localities variously retrograded, lustrous, coarse- garnet schists contain chloritoid and are associated with abundant pegmatite and calc-silicate gneisses like at Star Hill. At Ludlow Mountain kyanite-chloritoid-muscovite schists are intruded by granodiorite gneiss dated at 1350 Ma by U-Pb zircon (table 1) and at College Hill

by the College Hill Gneiss dated at 1250 MA (Ratcliffe, 1997). The association of coarse-grained garnet schist with minor quartzite, dolomite and calcite marble, and calc-silicate gneiss is widespread throughout the Mount Holly Complex. These metasedimentary rocks are now known to be older than 1350 Ma.

Workers apparently are impressed by the coarse grained nature of the garnet schists here and with similar garnet schists in the Hoosac Formation-- for example unit ЄZhg on Hoosac Mountain in the North Adams quadrangle, Massachusetts, at the type section of the Hoosac (Ratcliffe and others, 1993). The Hoosac garnet schists, however, are never associated with calc-silicate rocks, granite gneisses, pegmatite or other rock types typical of the Mount Holly Complex. We favor the correlation of garnets schists of the Cavendish Formation on Star Hill with similar rocks within the Mount Holly Complex.

Head south

24.6 Bear right on East Hill Road, then immediately turn left down hill

25.9 At intersection with route 131 turn right. Exposures on the right are feldspathic biotite-quartz-plagioclase granofels (Ycfs) of the Cavendish Formation, that here are in contact across the Black River with dark green, coarse grained diopside-graphite calc-silicate gneiss (Ycs). This rock can be traced up the hill to the east for about 1 km where it is in contact with the Felchville Gneiss (Ygp). West of the road, fine grained trondhjemite gneiss of the Felchville intervenes between Ycs and Ycfs unit.

26.5 Entrance to Cavendish Gorge and electric generation plant

STOP 8. CAVENDISH GORGE TYPE SECTION OF CAVENDISH FORMATION. The conventional interpretation (Doll and others, 1961; Thompson and others, 1993) is that the contact exposed here between dolomite marble and gneiss is an overturned west dipping unconformity with rocks of the Mount Holly Complex overlying Cambrian cover rocks of the Cavendish. This interpretation was not arrived at simply, however, as Thompson (1950) originally assigned all these rocks to the Proterozoic. Map relations here, however, indicate that the trondhjemite gneiss exposed at the power plant does intrude the other rocks of the Mount Holly and truncates marble and schist of the Cavendish at numerous places along the gorge. Figure 12 shows the map relations of the trondhjemite and the location of diopside-rich calc-silicate. The fine-grained trondhjemite gneiss of the Felchville Gneiss from location A (fig. 12) contains zircons dated at 1424±22 Ma (table 1 and fig. 10b, c). Chemical composition of the fine grained trondhjemite closely match those of other samples of the Felchville at the dated locality at Stop 4. My present interpretation (Ratcliffe, 1995a, and in press) is that the Cavendish is older than the trondhjemite traverse Stop 8.

Park at the power house and examine the contact between the dolomite marble and trondhjemite gneiss at the power plant (loc. A). There is little evidence for either an unconformity or an intrusive contact. Walk south up the road parallel to the tunnel that extends from the dam to the power plant. Trondhjemite gneiss on the west (by point B) has encroached on the schist and at this point the marble is absent or very thin. Lower ledges west of the tunnel have layers of schist in contact with trondhjemite gneiss. From point B, one can descend easily into the bottom of the gorge, walking entirely on schist. In the gorge 2 thin zones of deformed pegmatite crop out. At point C, where the tunnel and road cross, the schist has layers of gneiss in it including a hornblende-garnet plagioclase rock and more felsic layers. Immediately west near the concrete wall the contact between trondhjemite gneiss, schist and some layered biotite-quartz plagioclase gneiss and garnet gneiss is exposed. The marble is absent. At point D 100' feet to the south on the edge of the gorge, a 4 foot layer of marble intervenes between schist on the east and trondhjemite gneiss in the cliffs to the west.

In walking to the dam (point E) well-layered biotite gneiss, some calc-silicate rock and a white dolomite marble are exposed. Note the finer grained, feldspathic schist that is somewhat similar to the Cavendish is associated with the gneiss near the dam.

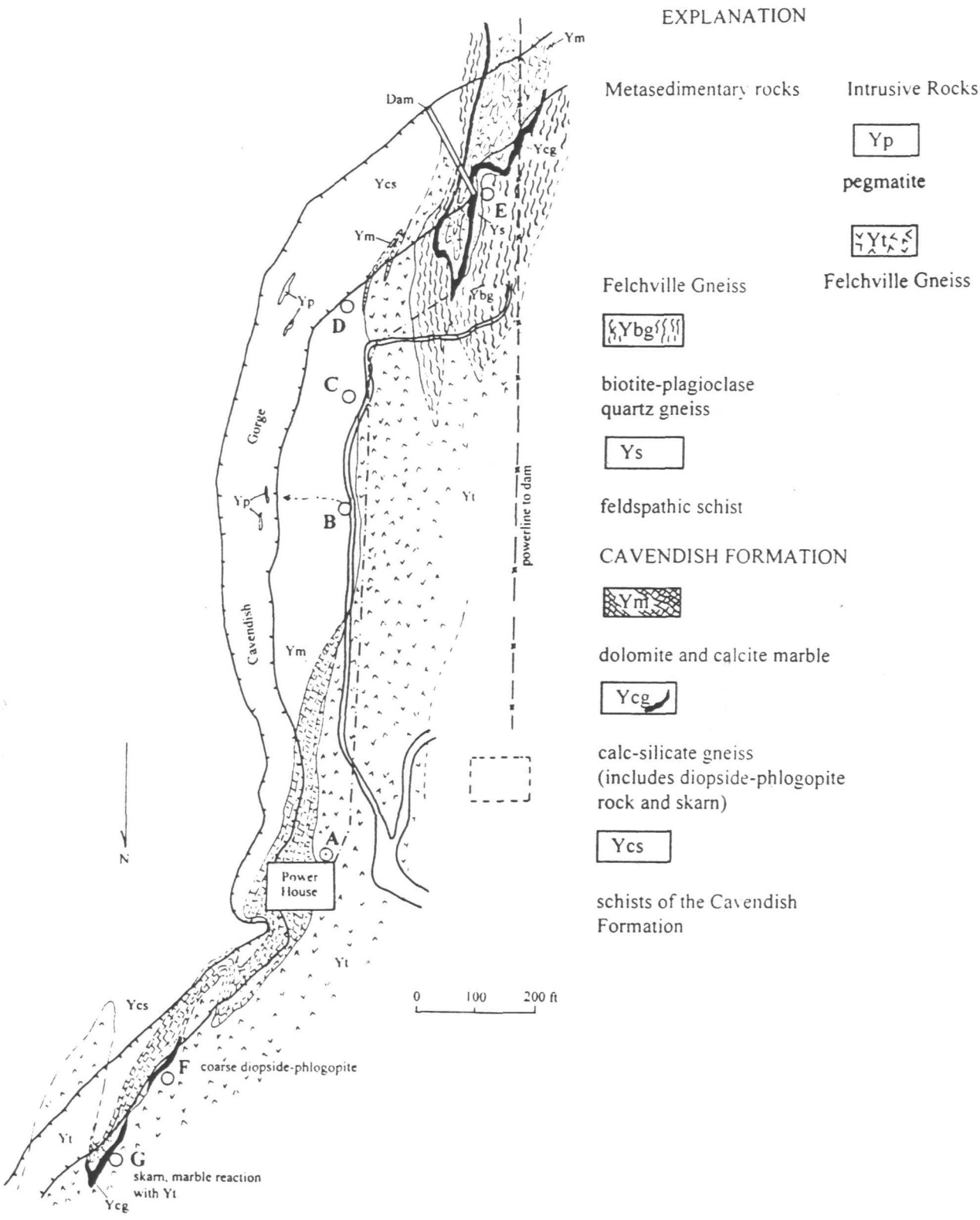


Figure 12. Geologic pace and compass map of Cavendish Gorge, north to bottom. Traverse starts at A and proceeds to E via road. Location A is sample site for C718 aplite of the Felchville Gneiss with $^{207}Pb/^{206}Pb$ zircon age of 1424±22 Ma. Arrow from B shows easy route into Gorge.

From point E descend (single file!) along the northeast trending vertical joint face and examine the rocks. A cross section is given in figure 13. Dolomite marble at the southwest end of the cliff (nearest the dam) passes downward (point 1, fig. 13) through a transitional zone of calc-silicate schist into well-layered biotite-quartz-plagioclase gneiss of the Mount Holly that contains abundant small masses of pegmatite; this is succeeded to the northeast by a more massive gneiss having aplitic layers. This rock is continuous with trondhjemite gneiss at the top of the cliff. At point 2 a transitional zone of mixed calc-silicate gneiss and schist passes down into a 12-foot-thick layer of calcite and dolomite marble. The coarser grained white calcite marble stands out as ridges. The northeastern contact with trondhjemite gneiss at point 3 is sharp. The trondhjemite gneiss narrows to about 2 ft in thickness at foot level where it is in the dolomite marble; a thin 3 foot layer of marble seen at point D marks the contact with the schist of the Cavendish. Note that the trondhjemite forms an upward widening wedge that replaces the marble. The lack of symmetry across the section makes it unlikely that three layers of marble could be structurally repeated. The interlayering of the biotite gneisses with calc-silicate rocks that, in turn, grade into the dolomite marble, suggest that the marble, calc-silicate rock, biotite gneiss and trondhjemite are all part of the Mount Holly Complex. Because the trondhjemite gneiss is discordant to all major units of the Cavendish it seems that the rocks of Cavendish predated intrusion of the trondhjemite. If these relation are correct, and the age of trondhjemite is correctly given by the zircon age of ≈1400 Ma then the Cavendish Formation is clearly part of the Mount Holly Complex and very old.

Similar relations can be seen south of the power station where trondhjemite gneiss has a coarse tremolite-actinolite skarn, and diopside and coarse-phologopite rock along the contact with the marble. Excellent exposure of trondhjemite gneiss in contact with coarse garnet schist of the Cavendish can be found about 500 north of the powerhouse on the east side of the Black River.

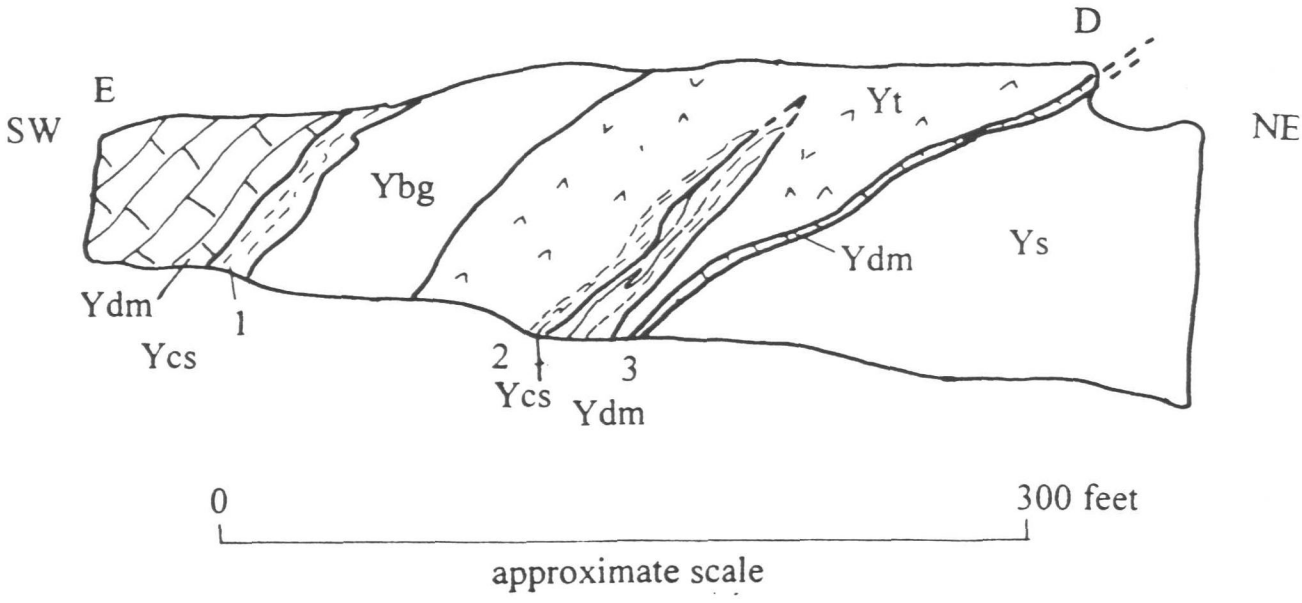


Figure 13. Cross section of joint face at Cavendish Dam, view to northwest scale approximate.

40.5 Turn right on Rt 103, the rocks here are biotite-quartz plagioclase gneiss of the Mount Holly Complex that are continuous with gneisses west of the Cavendish Gorge. Lenses of granitic gneiss intrude this sequence that also includes several layers of rusty garnet-biotite-muscovite±kyanite-quartz schist, white graphite marble, and hornblende-diopside calc-silicate rocks.

40.6 Cross into aluminous, chloritic granofels, hornblende-plagioclase-fasicile schist with amphibolite and coticule. This marks the keel of a tightly pinched syncline of cover rocks of the Moretown that directly overlie the thrust contact with the core gneisses of the dome. F_2 reclined folds dominate and plunge down the dip to the northwest, parallel to the highly lineated mylonitic rocks in the basement. The contact with basement rocks returns at Whitemore Road.

40.7 Granitic gneiss, Ybg and aplitic gneiss of the Mount Holly on the east flank of the Butternut Hill fold

41.5 Turn left on Smokeshire Road on north side of Williams River

41.6 Enter Andover Quadrangle

STOP 9. MOUNT HOLLY COMPLEX EAST FLANK OF BUTTERNUT HILL FOLD. Excellent crops of Mount Holly layered gneisses are cut by a Devonian granite dike that is weakly folded and foliated. The isoclinal folds plunge south and are cut by the granite; late cleavage (NS 80°W) in the granite matches late crenulation cleavages in the Silurian and Devonian rocks of the Proctorsville syncline 1 km to the west. Some Acadian shortening in the Proctorsville syncline postdated granite dike intrusion there as well. The plunges of folds and rotation senses of minor folds around the Butternut Hill fold were used by Thompson (1950) to substantiate that the fold, cored by Mount Holly is a north plunging synform in inverted rocks. Reexamination by Nisbet (1976) and studies by Ratcliffe (1996a, b) in the Andover quadrangle show that plunges are south not north along the southern closures of this fold (Stop 15). Likewise, the Spring Hill syncline, originally interpreted to plunge north in both its north and south closures (Rosenfeld, 1954; 1968), is now known to be a normal doubly plunging syncline (Ratcliffe and Armstrong, 1995). These data reverse the interpretation of the Butternut Hill and Star Hill folds of Thompson (1950) and from that shown in section E-E' of Doll and others (1961).

One kilometer to the west, in the bed of the Williams River, the contact between Mount Holly gneiss and coticule bearing, chloritic and felspathic granofels of the Moretown is exposed. If you are visiting this site, park just before the red house on the right and walk about 150 feet up stream from the house to see crops of the coticule. This is the base of the tectonic cover of the Chester Dome on the west flank of the Butternut Hill fold.

43.2 Turn around and return to Rt 103, and proceed past the previous right turning point from Cavendish Gorge.

43.3 Just before bridge is entrance on right to active flagstone quarry. Previous workers assigned the calc-silicate gneiss and K-feldspar augen gneiss in the quarry to the cover rocks, and correlated them with the exposures at Cavendish Gorge. The augen gneiss and calc-silicate rocks are all typical of the Mount Holly as we map them. A zone of trondhjemite gneiss like the Felchville Gneiss lies between the calc-silicate rocks and the muscovite biotite-plagioclase±quartz schist of the Cavendish and is exposed in the railroad cut just under the quarry. These exposures are contained in the slabby gneisses of uncertain correlation that rim the Chester dome in many places by Thompson and others (1993).

Previously the K-feldspar augen gneiss was called Bull Hill Gneiss (Doll and others, 1961), but the Bull Hill is now dated at 945 to 955 Ma (Karabinos and Aleinikoff, 1990) and should not be considered part of the cover sequence. This age also argues for placing these rocks in the basement, not in the cover.

41.3 **STOP 10. GASSETTS ROAD CUT: (OPTIONAL STOP) CONTACT BETWEEN TONALITIC GNEISS (BAILEYS MILLS TONALITE GNEISS (RATCLIFFE, 1995, IN PRESS)) AND THE CAVENDISH FORMATION.** The contact between progressively more sheared tonalitic gneiss on the east and shredded biotite-porphyroclastic schists of the Cavendish is exposed about 100 feet west of the new road and entrance to the house. This contact everywhere along Hawks Mountain is a tectonic contact (fig. 8). Calc-silicate rock and marble, mapped by Thompson and others (1993) and Doll and others (1961), as a basal unit are not at the base of the schist, are intermittent, are typical calc-silicate rock and marble of the Mount Holly Complex, and at many places are intruded by granite gneiss of the Mount Holly Complex. (See the Open-File Map of the Chester quadrangle (Ratcliffe, 1995b)). Therefore, the older maps are misleading because the basal unit comparable to the marble at Cavendish Gorge is in fact not present in the areas as portrayed by Doll and others (1961) along the base of Hawks Mountain.

Continue past Gassetts roadcut, past junction on left with Rt 10 east, to Brooks Road on left.

42.6 Turn left up Brooks Road, proceed 0.2 miles to last house.

42.8 **STOP 11. BAILEYS MILLS TONALITE GNEISS and CORE ROCKS AND STRUCTURES IN THE CORE OF THE CHESTER DOME.** (15 MINUTES) The pavement exposure and view to the north of Hawks Mountain is an excellent locality to get an appreciation of the internal structure of the Chester dome. Tonalitic gneiss with indistinct hornblende gneiss schlieren trend in subvertical attitude to the north-northeast. The pavement surface is the cross-cutting foliation that makes the form of the dome. This foliation, Taconian, Acadian, or both is parallel to the mylonitic foliation in the Hawks Mountain thrust and everywhere transects gneissic layers and older folds in the Mount Holly Complex (see fig. 5). Devonian muscovite-biotite granite dikes cross cut the foliation as straight walled relatively non-deformed dikes. Within the core of the dome the Devonian dikes are little deformed. A $^{40}Ar/^{39}Ar$ hornblende age of 379 Ma (Harrison and others, 1989) comes from roadcuts in Rt 103, just south of the Brooks Road, and indicate the minimum age of the Acadian metamorphism here. Assuming that the age of the Devonian granite dikes is comparable to that of the Black Mountain granite from the Guilford Dome, given by U-Pb zircon upper intercept age at 373±4 Ma (Table 2), the foliation that is warped to produce the dome is older than 373 Ma (see Stop 6, Trip C-6). Granite dikes near the margin of the dome are weakly to strongly foliated and along the eastern flank of the dome they are highly folded. The dome-stage deformation must, therefore, be in part younger than 373 Ma by the above reasoning. This is consistent with the regional data that indicate that deformation younger than Acadian granites is developed as far west as the east flank of the Green Mountain Massif . (See Stop 6 of Trip C-6)

43.0 Return to Rt 103, turn left; crops on left of Baileys Mills Tonalite Gneiss are cross cut by straight-walled Devonian granite dikes. Several chemically analyzed samples of Baileys Mills Tonalite Gneiss included in the text discussion are from this roadcut.

43.4 Jewett Road on right, crops to west and above on the hill constitute the type section for the Baileys Mills Tonalite Gneiss (Ratcliffe, in press).

45.5 Church Street intersection, continue through "Stone Village", to sharp bend in Rt 103 to right.

45.1 TURN LEFT BEFORE BRIDGE (GO STRAIGHT)

46.0 Follow dirt road to Rt. 11, cross Rt. 11, just past first house on left are outcrops of a lustrous muscovite-biotite-quartz-plagioclase±garnet schist in the Mount Holly. Along strike to the north, this schist contains coarse garnets and it resembles the Gassetts Member of the Cavendish. This rock has been mapped north to exposures in the spillway of North Springfield Reservoir. This rock is clearly within the Mount Holly and is just another example of richly garnetiferous schists that occur at many structural levels in the core of the Chester Dome, but within the Mount Holly Complex. Another wide outcrop belt of coarse kyanite-garnet-muscovite-biotite-plagioclase-quartz schist occurs north of route 10 in the north central part of the Chester quadrangle (see Ratcliffe, 1995b).

Cross into Saxtons River quadrangle.

47.8 Intersection with Douglas Road on right, bear left uphill--don't go down and over RR tracks.

48.6 Crops of Bull Hill Gneiss in old quarry on left. This is the belt of Bull Hill Gneiss that extends northward from the type section on Bull Hill. A U-Pb zircon upper intercept age of 955±5 Ma has been determined (Karabinos and Aleinikoff, 1990) for a sample from the south side of the William River opposite this point. The Bull Hill Gneiss was considered part of the metavolcanic rock in the cover sequence (Doll and others, 1961) because of its interlayering with marbles and schists of the Cavendish Formation. The new data and new mapping indicate it is a discordant intrusive rock, younger than the Grenvillian deformation. In cases where it is apparently interleaved with schist and granofels of the cover sequence, either of two interpretations seem possible: (1) the Bull Hill and other rocks are in fault contact; or (2) the "cover rocks" are really older than the Bull Hill. The next stop illustrates this problem well.

48.7 Pull off to right; park

STOP 12. BULL HILL GNEISS--MYLONITIC EAST BORDER OF THE CHESTER DOME This is not the most scenic spot, but the exposure in the river beneath the accumulated piles of the trash dump are very instructive. **PLEASE BE CAREFUL OF BROKEN GLASS!**

An exceptionally fine strain profile from sheared Bull Hill Gneiss (955±5 Ma) through mylonite gneiss to tectonically layered semi-schist (mylonite) having enclaves of Bull Hill Gneiss is well exposed over a 150-ft section in the river bed. The tectonic layering is intensely developed, so much so that it resembles a bedded rock. This would be a possible interpretation if it were not for the relict zones of prophyroclastic gneiss. Exposures like this may have caused previous workers to regard the Bull Hill Gneiss as interbedded with cover rocks. The age of this mylonitization is unknown and could be Taconian, Acadian, or both. It is representative of the intense shearing along the basement-cover contact that occurs everywhere around the Chester dome.

Continue east on Green Mountain Turnpike.

49.1 Unmarked dirt road; turn left

49.5 Park by small house on left

STOP 13. TONALITIC GNEISS OF THE NORTH RIVER IGNEOUS SUITE: AGE OF THE MORETOWN FORMATION AND EASTERN COVER OF THE CHESTER DOME. We are in the eastern cover sequence, approximately 0.2 miles east of the sheared Bull Hill Gneiss seen at Stop 12. The rock exposed at the stairs to the house is a hornblende-biotite tonalitic gneiss that contains plagioclase phenocrystic mafic dikes and an unusual pod of biotite trondhjemite. The trondhjemite pods appears to be a non-mixing felsic segregations within the more mafic rocks. Grain size variations in the subrounded pod and the smaller felsic segregation within the mafic rocks suggests that the rocks are igneous intrusive rocks rather than volcanic breccias. This tonalite gneiss form a discrete body several kilometers long and 0.5 km wide that is encased in feldspathic schist and granofels of the Moretown Formation. The rock varies from gneissic to a more massive equigranular and hornblende spotted rock with a phaneritic, hypidiomorphic granular texture.

Samples of this rock were collected for zircon dating and a U-Pb upper intercept age of 486±5 Ma was determined by John Aleinikoff (Table2). As discussed in the text, this age is in agreement with that of other dated, more felsic members of the North River Igneous Suite. The data here suggest that the Moretown is at least 486±5 Ma old at this locality (see figure 10 of Trip C-6).

Walk up the hill past the zircon sample locality across a section of Moretown granofels, and amphibolite that extends to within 300 ft of the Bull Hill Gneiss. The remaining 300 ft-wide section consists of lustrous chlorite-muscovite-garnet quartz schist, amphibolite, and a feldspathic granofels 100 to 150 ft wide closest to the Bull Hill. With the exception of the lowest rocks, all other rock types are typical of the Moretown or possibly Stowe Formation. The tectonically abbreviated section places rocks of the Moretown Formation nearly directly on the Bull Hill Gneiss. Ratcliffe and Armstrong (1995) have interpreted this contact as a fault. Return to cars.

51.1 Turn left on Green Mountain Turnpike.

51.2 Turn (bear) right at intersection onto paved road, go through Bartonsville covered bridge to intersection road work.

52.3 Rt 103; turn right

53.8 Small crops on the south side of Rt 103 of Bull Hill Gneiss dated by Karabinos and Aleinikoff (1990) at 955±5 Ma. Note that the intense mylonitization and later structures seen in the Bull Hill Gneiss must be Paleozoic and post Grenville structure as the Bull Hill cross cuts Grenvillian structures.

54.0 Sylvan Road on left; turn left

We are driving south-southwest through the narrow neck of Mount Holly Complex gneisses that separates the Chester dome from the Athens Dome to the south. (See the Digital Open-File Map of the Saxtons River quadrangle (Ratcliffe and Armstrong, 1996) and figure 4). The road follows a central belt of migmatite granitic gneiss (Ygg), and two flanking belts of biotite quartz plagioclase gneiss (Ybg), and amphibolite (Ya) (fig. 4). Grenvillian gneissosity is subvertical, whereas the Taconian and Acadian foliations are gently warped and dip generally to the south producing the south closing end of the Chester foliation dome. The western margin of the dome is strongly foliated and faulted along parallel belts of Bull Hill Gneiss with structure comparable to that seen at Stop 12. The transition from northeasterly trending gneiss in the core to tectonically layered gneisses conformable with the western margin of the dome occurs over a zone of progressively increasing strain that is marked by a fault in basement that Ratcliffe and Armstrong (1995) termed the Kenny Pond Thrust (fig. 6). The thrust takes its name from exposures near Kenny Pond on the southern closure of the Athens dome in the Newfane quadrangle (Ratcliffe and others, 1992) where the fault places slivers of Mount Holly Complex Gneiss over rocks of the Hoosac Formation.

57.3 Intersection Rt 35--turn left, excellent exposures on both sides of the road of tectonically laminated Bull Hill Gneiss, gneisses of Mount Holly Complex, and slivers of Hoosac Formation. Intense down dip elongation lineation plunges N70° to N 80°W. In the core of the dome this lineation is east-west and subhorizontal and on the east limb plunges S75° to 80°E. Hingelines of reclined folds are parallel to the extension lineation. This fault fabric and lineation is folded over the dome. We believe this structure is Taconian and that it dates from a period of Taconian imbricate thrust faulting that affected all the cover rocks (see trip C-6) and all of the Green Mountain massif westward into the Taconic Range.

Continue south to Grafton

60.0 Intersection Rt 121; turn right

60.2 Old Grafton Inn on left. Intersection with Old Grafton Road. Continue straight.

60.3 Intersection of Route 121 with Middletown Road. Proceed straight on Middletown Road, then continue straight at next intersection (on Hinkley Brook Road- a dirt road) where Middletown Road bears right up hill.

Triangle intersection on right. Turn right, onto Houghtonville Road.

62.75 Dirt road on right; location for next stop. NOTE: THIS IS A PRIVATE DRIVE. THE OWNERS HAVE ASKED US NOT TO INCLUDE THIS LOCALITY WITHIN THE GUIDEBOOK. PERMISSION FOR ACCESS TO THIS STOP HAS ONLY BEEN GRANTED FOR THIS TRIP. PLEASE CONTINUE FORWARD IN THE ROADLOG IF YOU ARE DOING THIS TRIP INDEPENDENTLY.

STOP 14. RUGER HILL (Saxtons River 7.5 x 15 minute quadrangle) The discussion here follows terminology in the preliminary Saxtons River map (Ratcliffe and Armstrong, 1995; 1996). Rocks here constitute a bimodal igneous suite which we correlate directly with rocks of the North River Igneous Suite, seen at the last stop. At least five different mafic and felsic igneous rock-types are present in these exposures, including (from oldest to youngest):

(1) Well-layered, garnet-hornblende-cummingtonite-plagioclase amphibolite. Compositional layering in this unit is parallel to regional S2 (Taconian) schistosity, and is defined by epidote-rich layers1 to 5 mm thick. Layering in this unit appears to be bedding, and this mafic rock was probably a water-lain volcanic rock. Similar rocks are found interbedded with Moretown Formation granofels, both east and west of this locality.

(2) Medium-gray-weathering, light-gray, garnet-hornblende-muscovite-quartz-plagioclase felsic metavolcanic rocks that are bedded on a 5 to 20cm-scale. This unit is found in outcrops west of the house about midway between the dirt access road and the woods to the north. These volcanic rocks appear to be interbedded with the mafic varieties described in 1). These rocks are very similar to unit Omfg in the Moretown Formation to the south (see Ratcliffe and Armstrong, 1995).

(3) Dark-green, epidote-rich hornblende-needle, garnet-epidote-plagioclase-chlorite-hornblende amphibolite with distinctive 3 to 6 mm garnet porphyroblasts. Cross-cutting relationships are difficult to recognize locally, but this unit appears to cut both of the volcanic units previously described. This unit is very similar to the mafic dikes found within the Moretown Formation in this area.

(4) Cream-colored to light-gray, massive to slightly layered, garnet-muscovite-quartz-plagioclase gneiss; some layers contain biotite while others contain large hornblende porphyroblasts. Otherwise, this unit is not typically layered. In many places, this unit demonstrably cross cuts the rocks 1 and 2, and in several outcrops on the west side of the house, cross cuts 3 as well. These intercalated hornblende and biotite felsic gneisses are similar to felsic gneisses mapped within the North River Igneous Suite on the east side of the Chester and Athens domes, in the Saxtons River, Chester, Townshend, Newfane and Brattleboro quadrangles, and the Barnard Gneiss (of Richardson, 1924) mapped within the Ludlow, Andover, and Plymouth quadrangles on the west side of the Chester dome.

(5) Very dark-green, fine-grained epidote-plagioclase-chlorite-hornblende amphibolite dikes with relict 4 to 10 mm, rectangular to ovate, plagioclase phenocrysts. This unit cross cuts all of the other units and also a Taconian foliation. Very similar rocks have been mapped to the east and to the south, and are included in the Branch Brook Intrusive Complex (Armstrong, 1994), and are widely present in the Hawley Formation of Kim and Jacobi (1996) in Massachusetts and in the Barnard Gneiss (of Richardson, 1924; see Trip C-6).

All of the rocks at Ruger Hill (except for those in unit 5) contain Taconian S_2, although associated F_2 folds are rare. The pervasive set of north-plunging folds seen at this locality are Acadian F_4 and/or F_5; related to the Spring Hill synform, visible to the northeast from the east side of the house. This generation of folding is responsible for the infolding of these structurally higher-level rocks with the lower part of the pre-Silurian section.

An intriguing point about the distribution of the arc-related igneous rocks (see Trip C-6) is that no similar intrusive rocks are found within the pre-Silurian rocks (including the Middle Proterozoic basement of the domes); regardless of position on either side of the domes, they are restricted to the Moretown and Cram Hill Formations. Does this mean that the feeder system for these igneous rocks was part of a large sill-like system that followed the same structural level across a broad area. Alternatively, is it possible that the intrusives occurred within the Moretown-Cram Hill lithotectonic belt in an area far removed from the rest of the pre-Silurian sequence prior to Taconian accretion and tectonic juxtaposition? We favor the latter interpretation as discussed on Trip C-6.

During the course of our work in southern Vermont, we have used Cram Hill Formation to apply to a complex collection of intercalated and laterally discontinuous metasedimentary and metavolcanic rocks that overlie the Moretown Formation and which are intruded by the North River Igneous Suite or igneous rocks of the Barnard Gneiss (of Armstrong, 1994; Richardson, 1924). The felsic and mixed felsic and mafic varieties of igneous rocks have previously been referred to as the Barnard Gneiss (of Richardson, 1924) or the Barnard Volcanic member of the Missisquoi Formation (Doll and others, 1961). The remaining metasedimentary rocks have been referred to as the Cram Hill Member of the Missisquoi Formation (Doll and others, 1961). Thompson, Rosenfeld, and Chamberlain (1993) applied the name Barnard Gneiss to what they regard as largely volcanic rocks and have used Hawley Formation (Emerson, 1917; Stanley and Hatch, 1988) to refer to a collection of metasedimentary rocks and greenstones above their Barnard. Stanley and Hatch (1988) applied the name Hawley to rocks in Massachusetts which are predominantly felsic and mafic intrusive and extrusive rocks and minor amounts of splintery sulfidic black schist (known as the schist member) that is mapped in the upper third of the formation. Armstrong (1994) has shown that the Barnard Volcanic member of the Missisquoi Formation (of Doll and others, 1961) in

southernmost Vermont contains metaigneous and metasedimentary rocks that are coextensive with the entire Hawley belt in Massachusetts. Both belts consist almost entirely of amphibolite, metatrondhjemite, metatonalite and minor intermediate to felsic metavolcanic rocks that intruded or are interlayered with sulfidic black schist. Thus, the Hawley Formation of Massachusetts is largely equivalent to rocks mapped by others as the Barnard Gneiss and the names Hawley and Barnard are equivalent; the use of both terms, the Hawley Formation and the Barnard Gneiss for rocks in Vermont is inappropriate.

In order to resolve this problem, we use the name Cram Hill Formation to refer to the metavolcanic and metasedimentary rocks in Vermont and equate the Cram Hill with the metavolcanic and metasedimentary parts of the Hawley Formation in Massachusetts, excluding those intrusive rocks in the Hawley Formation, which in Vermont are referred to as the members of the North River Igneous Suite (Armstrong, 1994); or as yet unredefined intrusive rocks of the Barnard Gneiss (of Richardson, 1924).

Continue north on Houghtonville Road. On the right is Spring Hill and the Spring Hill synform, which infolds structurally, and we think, stratigraphically higher rocks of the Moretown Formation, North River Igneous Suite, and Cram Hill Formation within a structural depression between the Middle Proterozoic basement and pre-Silurian cover of the Chester dome (to the east) and the Butternut Hill Fold (to the northwest).

64.4 Junction with Route 121 in village of Houghtonville, turn left.

64.5 Turn right on first road
Rocks on east are rusty schists and amphibolite of the Cram Hill Formation on the wests limb of the Spring Hill Syncline. Plunges reverse from north to south in the middle of the fold.

66.5 Sharp turn to right and head downhill, exposures of quartzite and schist of the Cram Hill in south-plunging folds east of road. Granofels of the Moretown exposed to the west.

67.3 Intersection Popple Dungeon Road, turn left.
Follow the winding road west for two miles to first major intersection and road branching to the right. The exposure in the brook are megacrystic granitic gneiss of the Mount Holly Complex on the east flank of the Butternut Hill fold.

69.3 Continue bearing sharp left still on Popple Dungeon Road.

70.3 Road on right after sharp bend to right

70.6 Park at intersection with small side road on right just before large open fields on left

STOP 15. SOUTHERN CLOSURE OF THE BUTTERNUT HILL FOLD

From the intersection walk due north along old logging road about 700 feet and make way up slopes to the west to crest of small hill.

On the lower part of the northeast facing slopes are crops of the Mount Holly Complex overlain by albitic granofels of the Hoosac Formation. The Hoosac contains epidote amphibolite and rusty weathered schists. Folds plunge moderately steeply to the south-southwest in the cover rocks. To the south, the schistosity in the Hoosac and Moretown Formations on the nose of the Butternut Hill fold plunge steeply south away from the core gneisses of the fold. The geologic relations on the nose of the Butternut Hill fold, and to the south, indicate a south closing, southwest-plunging fold in contradiction to the northerly plunges shown in figure 3.

Our data for both the Spring Hill fold and the Butternut Hill fold support south, not north plunges. These observations change drastically the previous regional interpretations of the garnet rotation axes (Rosenfeld, 1954) used to support the fold nappe model of Doll and others (1961). See our cross sections B-B' and C-C' in figure 6. Our structural observations discussed at Stops 3, 4, 6, 7, 14, and 15 indicate that the Acadian recumbent folds shown in E-E' of Doll and others (1961), and which figure prominantly in prevuous regional interpretation, are not supported by the detailed geology we have mapped.

END TRIP

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TRIP C-1

HOW DOES NEW HAMPSHIRE CONNECT TO VERMONT?

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INTRODUCTION

The aim of this trip is to acquaint interested New England Geologists with several localities that we consider of key importance in understanding the post-Taconian history of eastern Vermont and western New Hampshire. We also present a stratigraphic synthesis showing how we believe the various Silurian and Devonian map units of the two states are related to one another.

The greater part of the known or assumed Siluro-Devonian of Vermont is made up of several closely related map units, covering a large area in the eastern part of the state, that we shall refer to here by means of the collective term "Calciferous Schists" first used by Edward Hitchcock and others (1861). The principal map units in the Calciferous include the Northfield and Meetinghouse "Slates" near the west and east boundaries, respectively, and the Waits River and Gile Mountain Formations (of which the Meetinghouse is often taken as a member) in the central region. Much of the western part of the Calciferous has been mapped as Waits River (highly calcareous), and much of the eastern part as Gile Mountain (sparingly calcareous), but it has become increasingly apparent to us that a boundary mapped on this basis alone can not be a time line and is more likely a boundary between laterally equivalent facies (Fig. 2). All of these units contain the same array of rock types, but in different proportions. As a consequence the mapping of the boundaries between them is dependent to an uncomfortable degree on both the subjective state of mind of the observer and on the nature of the available exposure. Carbonate-rich rocks, for example, are notably more abundant in stream or highway-cut exposures than in pasture or woodland outcrop.

The Calciferous also contains a relatively thin unit of predominantly mafic metavolcanics known as the Standing Pond Volcanics that has proved to be a useful marker in deciphering the internal structure of the Calciferous. Opinions recently expressed to the contrary, we regard the Standing Pond as an essentially coeval time marker, probably containing not more than two distinct geochemical units. Although past compilations, in which some of us have been involved, have used the Standing Pond as an arbitrary boundary between the Waits River and the Gile Mountain, we wish to make it clear, again, that we regret this action and fully recognize that it has been painfully misleading as to the actual distribution of the rock types. This is the principal reason for our revival of the old collective term "Calciferous Schists." It would probably never get past the fussy scrutiny of any commission on stratigraphic names, but it is highly useful in expressing our views as to the nature of the Vermont-style Siluro-Devonian. A second metavolcanic unit, the Putney, has recently been described by us (Thompson and others, 1993) as probably distinct from the Standing Pond. We now suggest that they may be equivalent for reasons to be discussed more fully in connection with Stop 5, the final one of the day.

Except for the revision just mentioned we shall omit here any further discussion of the internal structure of the Calciferous, our views having been expressed fairly extensively in the discussion of Trip B-1 in the guidebook (Thompson and others, 1993) accompanying the 1993 GSA meeting in Boston. Instead, we here focus our attention on narrow zones of rock, also believed to be Siluro-Devonian, that abut the Calciferous on the east and west, and separate it, where present, from the various pre-Silurian units (about which we will also have very little to say here). On the west these units are nearly continuous from the Quebec border to Massachusetts. They include the Shaw Mountain Formation (containing Silurian fossils) in the central, and northern part of the state, and the rocks assigned here to the Hawley Formation. They are separated from the adjacent Calciferous by a sharp, easily mapped boundary, for most of this extent, and also occupy an isolated syncline west of Grafton, Vermont. This is the Spring Hill syncline that we shall visit at Stop 1. Stop 2 will be in the main belt almost directly east of Stop 1 (see Fig. 1).

Similar units are also found adjacent to the east boundary of the Calciferous from our assembly point north (it is also Stop 5), many of the occurrences, as at Stop 3, lying on the New Hampshire side of the Connecticut River. As on the west, these separate the Calciferous from the pre-Silurian, but are far less continuous owing in large part to at least two generations of post-metamorphic faulting. As a consequence they have never received a collective stratigraphic name, and have in some instances been given member status in adjacent pre-Silurian units. These will be visited at Stop 3 with further discussion of their distribution and regional correlations.

The pre-Silurian rocks that separate the Vermont and New Hampshire Siluro-Devonian sequences north of our assembly point lie in the core of the Cornish nappe (Thompson and others, 1968), a west-verging, recumbent fold, the axial surfaces of which have been refolded about roughly north-south axes. Although a fold nappe in the relatively ductile terrane at this latitude, it is probable that the Cornish nappe is floored by discrete thrusts in more northerly areas. Rocks of the Vermont Siluro-Devonian sequence, including Calciferous Schists of the Waits River facies, appear beneath the Cornish nappe in an antiformal area in New Hampshire, extending from Charlestown, north through Claremont and into Cornish. Siluro-Devonian rocks of the Vermont type also appear on the upper limb of the Cornish nappe near its western hinge. This is in fact the most easterly exposure of this sequence if one unfolds the Cornish palinspastically.

Rocks of the New Hampshire sequence appear on the east limb of the antiform in the upper limb of the Cornish nappe (Stop 4). The most westerly units are almost certainly Siluro-Devonian in age, and they rest on probable pre-Silurian units. At Stop 5, farther south, we believe that they lie unconformably on Siluro-Devonian strata of the Vermont sequence. They appear, here, to be on the upper limb of the Cornish nappe near its leading hinge. The rocks at the contact are either quartz conglomerates and quartzites, here assigned to the Hardy Hill Quartzite, or gray schists, typically with thin sandy laminae, belonging to the Devonian Littleton Formation. The Hardy Hill, where present, underlies the Littleton, and may be as young as Devonian. The rocks shown on Fig. 1 as undivided New Hampshire sequence, farther east, also include fossiliferous Silurian and pre-Silurian units associated with the domes and nappes of the Bronson Hill anticlinorium (see trip A-2 in this guidebook).

Several slang designations for some of the key boundaries discussed above may slip from the lips of your leaders from time to time. The term RMC (or Richardson Memorial Contact) refers loosely to the western limit of the Vermont Siluro-Devonian. Depending on the mapper or compiler, it may be anywhere between the western limit of the Hawley or Shaw Mountain, and the western limit of the Calciferous, where these are

different. Monroe Line (ML) refers, in most usage at this latitude, to the eastern boundary of the Calciferous. Northey Hill line (NHL) refers roughly to the boundary between the New Hampshire Siluro-Devonian, and the pre-Silurian west of it, north of our assembly point. The Chicken Yard Line (CYL) takes its name from a formerly fragrant outcrop in Dummerston, Vermont. It is the western boundary of the New Hampshire S-D at and south of our assembly point. Although all of these lines follow faults at some localities, we do not believe this to be everywhere true. Assuming that it is everywhere true is a dangerous state of mind in that it may blind one to the recognition of the existence of potentially important stratigraphic evidence.

ITINERARY

The assembly point for this trip is at the Howard Johnson's restaurant on the north side of Vermont Rte. 11, a few yards west of Exit 7 on Interstate Highway 91. The logistics of this trip will be greatly eased by consolidating into as small a number of high-capacity vehicles as possible. Food and fuel are available here, from 7:00 AM on, but we hope to be underway by 8:00. Parking is available at Howard Johnson's, also at the Mobil and Texaco stations on south side of Rte. 11. We will be back here in the afternoon, ca. 4:30, for Stop 5, the final one for the day. We will also pass by here at about noon so that people who must leave the trip earlier than that may pick up their cars.

We are in the USGS Springfield Quadrangle, and will cross the southern part of the Chester Quadrangle, without stopping, *en route* to Stop 1. Stops 1 and 2 are in the Saxtons River Quadrangle, and on our mid-day return here we will cross the extreme northwest corner of the Bellows Falls Quadrangle without stopping. The afternoon will be spent in the North Hartland Quadrangle (Stop 3), the Ascutney Quadrangle (Stop 4), and back here in the Springfield Quadrangle at Stop 5. The North Hartland and Chester are 7 1/2 x 7 1/2 minute sheets at 1:24,000 with 20 foot contours. The others are all 7 1/2 x 15 minute quadrangles at 1:25,000 with *six meter* contours. So much for going metric!

The stops are widely spaced, and, because most of the roadside outcrops are irrelevant to the purpose of this trip, we will save space by mentioning only a few of the more interesting ones in the following road log. For those who would like more complete descriptions of the roadside occurrences we refer you to three earlier guidebooks, by one or more of us, that have dealt more fully with the local geology in much of the ground we will be crossing. These are here designated GB-1 (Thompson, 1954), GB-2 (Trzcienski and others 1972), and GB-3 (Thompson and others, 1993).

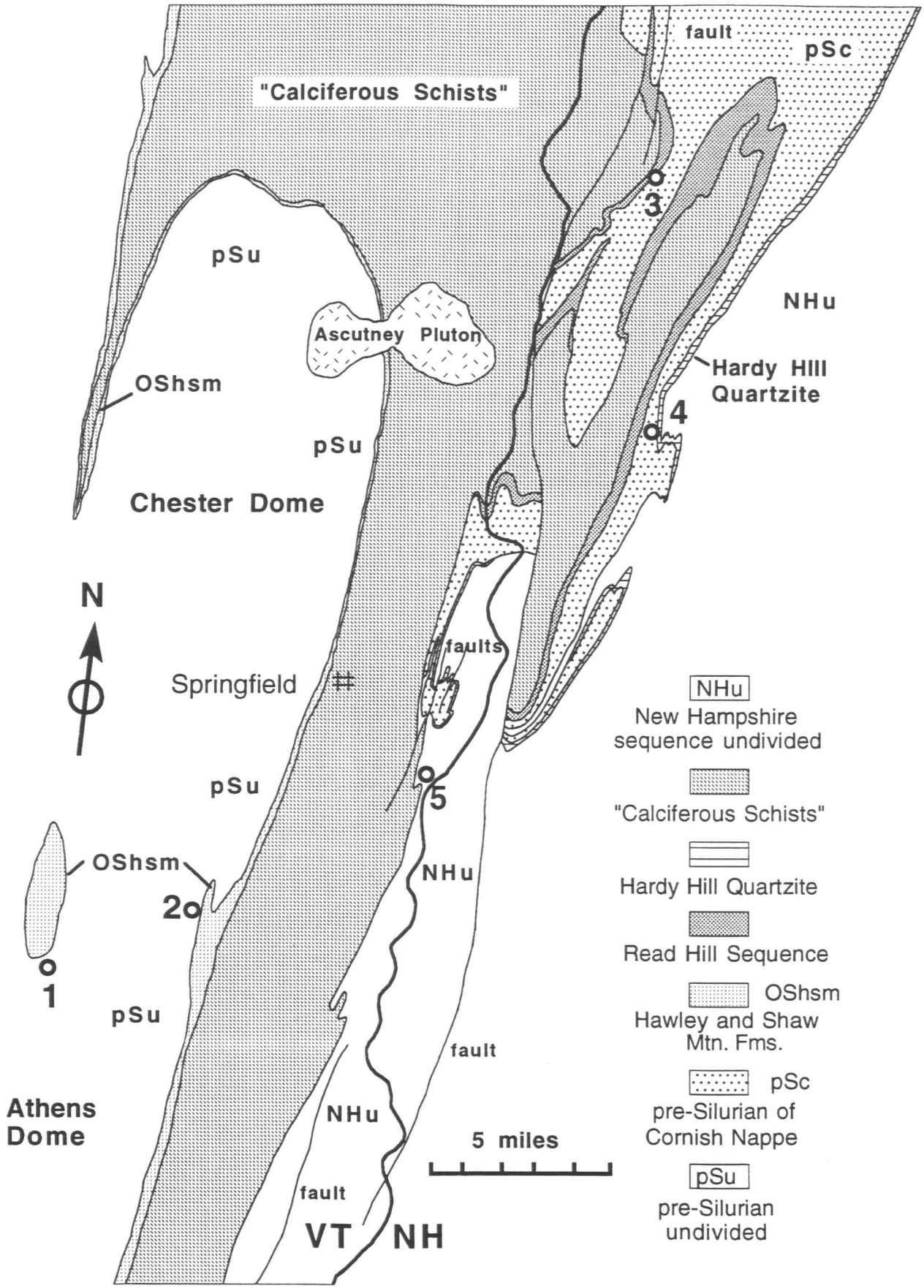
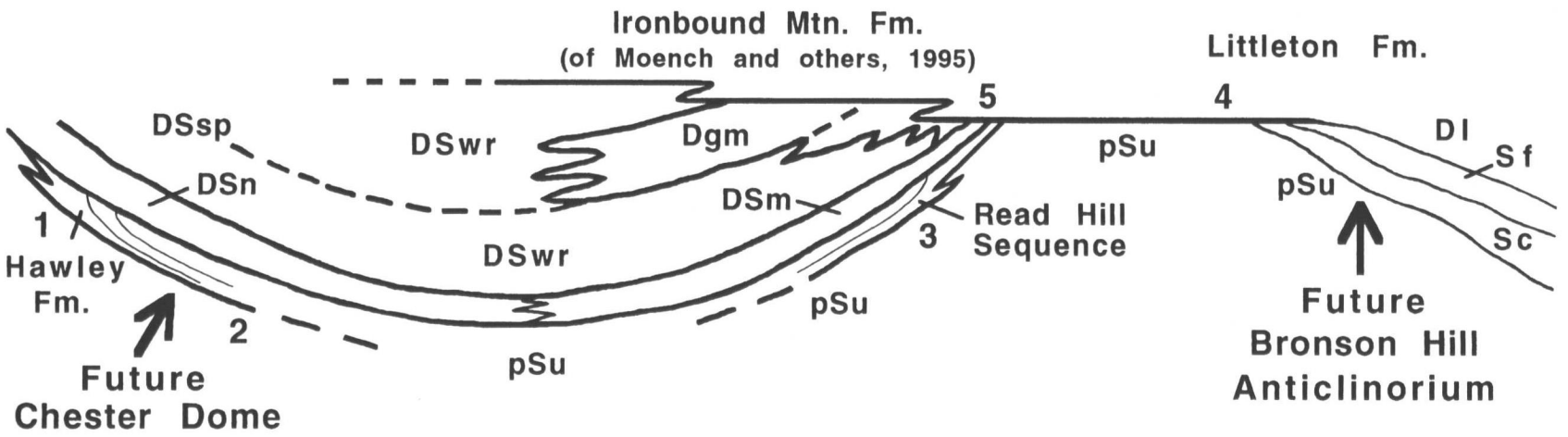


Figure 1. Schematic Geological Map of Field Trip Area

Figure 2. Stratigraphy Prior to Deformation



Dgm	Gile Mountain Fm.	Dl	Littleton Fm.
DSsp	Standing Pond Volc.	Sf	Fitch Fm.
DSwr	Waits River Fm.	Sc	Clough Fm.
DSm	Meetinghouse "Slate"	pSu	pre-Silurian undifferentiated
DSn	Northfield Fm.	Numbers indicate stop locations	

ROAD LOG

Mi.

0.0 Set odometer's zero at Howard Johnson's and proceed west through Springfield on Route 11.

0.01 Don't worry, the first large outcrop on the right will be visited later as part of Stop 5! From there through the town of Springfield we are passing through the "Calciferous Schists" and their associated metavolcanics (see GB-2, Stops 2-4, p.297, 298; and GB-3, Stop 1-3, p. B-11, 12).

4.30 Here, in the west part of Springfield, Rte 106 bears right and we continue left, entering the Chester Quadrangle, on Rte. 11 toward Chester. The rocks in the vicinity of this intersection (GB-2, Stop 4), are a part of the sequence along the RMC that we will see today in more detail at Stops 1 and 2.

8.8 For 0.3 mi. on right there are pre-Cambrian core gneisses of Chester dome. These have pegmatites containing large crystals of epidote, some several inches long, indicating a high-pressure regime, (see GB-3, Stop 3-5, p. B-20).

11.7 Rte. 103 from left coincides for a short distance with Rte. 11. We stay on Rte. 11, entering Chester

12.2 Turn left on Route 35, toward Grafton, soon entering the Saxtons River Quadrangle. For a description (in reverse order) of the outcrops over the next seven miles, see GB-3, p. B-20,21.

14.2 We are entering Grafton Gulf, one of a series of narrow slots in the gneisses of the western border facies of the Chester and Athens domes. These "gulfs" are caused by the erosion of a narrow band of carbonate rock that is but rarely seen.

14.7 A pillar of dolomite, holding up albitic schist on the right is one of these rare appearances. We cross Chester-Grafton town line a short distance beyond.

15.1 Leave Grafton Gulf. Road bends east, leaving line of gulf.

17.1 Road has returned to the line of the gulf, but it is less conspicuous here.

19.5 Intersection with Vermont Route 121 in Grafton. Turn right onto Route 121 and pass through Grafton, a picturesque village where some fine examples of old Yankee architecture have been restored and preserved thanks to the liberal application of dollars by the Windham Foundation. The local Cheddar cheese from the store in the south part of the village on Route 35 is some of Vermont's best.

19.7 Bear left at fork just beyond village center, leaving Rte. 121. Proceed straight ahead past the church on your left. Go straight ahead, do not take the road that bears left up the hill by the church.

19.8 A short distance beyond the church turn left onto the road that follows Hinkley Brook. Upstream, the new topographic map shows a hiatus in the road of about 0.3 mi., but the road, as we shall see, continues through, and has for decades. It is correctly shown on the old fifteen minute map [-- so much for field completion!].

20.8 Make a sharp right off of the Hinkley Brook Road, and go NE about 500 feet to a house where we will park for Stop 1. At this location we are on the south southeast slope of the easternmost knob of Spring Hill.

Stop 1 Landowner permission needed! Barnard Gneiss and Hawley Formation.

We shall here see a section, starting in the Barnard Gneiss, and extending through the lower half of the Hawley Formation. The Barnard is largely metavolcanic, but in part intrusive, giving Ordovician ages of 471 Ma or more (Karabinos and others, 1996; also this guidebook, Trip B-3). We start at elevation 368 m and head NNW up the hill toward the part of the section that will be the center of our attention. Outcrop appears at about 420 m elevation. The first outcrops consist of interlayered amphibolite, hornblende gneiss, and biotite gneiss of the Ordovician Barnard Gneiss. Some of the gneiss is garnetiferous. At about 450 m elevation we pass into a unit consisting primarily of rusty-weathering garnetiferous schist, some graphitic, with interbedded amphibolites. The unit also contains some thin feldspathic quartzites. For the purposes of this trip these schists have been mapped with the Hawley, but they may be more closely akin to the Barnard. At about 480 m elevation you will observe ahead a cuesta of white to gray conglomeratic quartzite about 10 m thick striking N30° E. The quartzite has a few interbeds of quartz-rich garnetiferous muscovite schist, and elsewhere on strike coexisting staurolite and chloritoid have been found in that schist. Notable, especially on the uppermost weathered surface, are highly deformed pebbles oriented about N5° W and plunging northerly about 26°. Above the quartzite and extending to the top of the 514 m E knob of Spring Hill is a thick zone of amphibolite, some of which is blastoporphyritic (characterized by relict phenocrysts of plagioclase, now saussuritic). Proceed WSW toward saddle in Spring Hill. Arriving in the semi-clearing in the saddle, you will note a silvery biotite muscovite schist peppered with small euhedral garnets and characterized by nodules rich in fine-grain rather pale-pink garnets. Much of the schist is graphitic, and in some places on strike, both kyanite and staurolite are present. Fifty years ago, when the saddle was open pasture, gray to creamy-white muscovite-bearing calcite marble was easily observable in the saddle immediately overlying the schist; and, if we have a sizable enough group, we may still be able to find some in outcrop. There are numerous blocks of the same marble in the nearby stone fence. The marble is unlike any marble noted at other horizons in southeast Vermont. An identical marble, at the same stratigraphic position, crops out at about the 528 m contour on the east side of Blake Hill about 3 km to the north. It is somewhat surprising that these were apparently overlooked by the normally very observant and resourceful nineteenth century Vermonters. They also overlooked some magnetite-rich iron formations that occur higher in the Hawley. Both would have been valuable commodities for local agriculture and iron production in that time when such necessary raw materials required transport over considerable distances by primitive means.

The dominant rock type above the marble is amphibolite, much of which is blastoporphyritic. These amphibolites (not to be visited) overlie the marble in the trough of the Spring Hill syncline. Interbedded with the amphibolites in a number of places near the axis of the syncline are strata up to 4 m thick of thinly laminated coticule (derived from the Latin diminutive for whetstone; characterized by very fine-grain pale pink garnet, much of which has thin partings rich in amphibole of the cummingtonite-grunerite series). In some places the coticule is rich enough in magnetite so as to be characterizable as a banded iron formation.

Rocks of the "Calciferous schists" are not present in the Spring Hill syncline but undoubtedly were prior to the last major erosion. The Calciferous Schists do appear farther north, however, in a syncline at Proctorsville Gulf and beyond, where they are flanked by thin Hawley.

Rocks having much in common with the Hawley as seen on Spring Hill occur along the western margin of the Calciferous Schists in central and northern Vermont as the Shaw Mountain Formation. At several localities, the limestones of the Shaw Mountain contain Silurian faunas, notably near Hardwick (Konig and Dennis, 1964), and near Irasburg (Doll, 1951). Similar occurrences along the northwest margin of the Calciferous Schists are found northeast across the Eastern Townships of Quebec. The Silurian Five Mile Brook Formation of Roy (1989, and references therein), crops out in extreme northwestern Maine near the Quebec border, and carries limestone and associated basalt much as at Spring Hill.

Return to cars, and return through Grafton village to the junction of Rtes. 35 and 121.

0.0 Zero your odometers here, and proceed straight ahead (east) on Rte. 121.

0.3 Outcrops on left are gneisses of the Athens dome near its northwest margin.

1.15 Turn left off Rte. 121 onto Fisher Hill Road. Fisher Hill road takes us across gneisses in the waist region between the Chester and Athens domes.

3.70 Turn left on Bartonsville Road.

4.50 Park on right, if time permits, just beyond height of land on Bartonsville Road.

Stop 2 Landowner permission recommended. Barnard Gneiss and Hawley Formation.

Aleinikoff and Karabinos (1990) obtained a zircon age of ca. 524 Ma from an outcrop of Barnard Gneiss near where the vans are parked. More recent work by Karabinos (see refs. at Stop 1), however, has revised this figure downward to 470-475 Ma.

Walk north a few yards, then east over outcrops of the Barnard Gneiss into rusty-weathering and gray schists and granulites here assigned to the Hawley. The schists include some thin coticules and minor amphibolite. Farther east, on the steeper slope, is quartz conglomerate and quartzite overlain by amphibolites to the crest of the ridge, all part of the Hawley. Walk north along the crest of the ridge, then drop west to a woods road and return to the vans.

The stratigraphic sequence here and elsewhere along the east side of the Athens and Chester domes is very similar to that in the Spring Hill syncline except that no marbles are known to be present within the amphibolites. Coticules tens of feet thick and some iron formations are found higher in the formation, nearer the contact with the Calciferous. The coticules and iron formations continue south into the Hawley Formation of Massachusetts, but the amphibolites are less abundant southward, giving way to rusty-weathering, sulfidic schists. A few miles south on Windmill and Putney Mountains a second quartzite and conglomerate appears above the coticules, near the top of the Hawley, at a position corresponding to that of the Russell Mountain Formation of Massachusetts (Hatch, and others, 1970).

5.9 Road turns sharply left, then right immediately.

6.1 Turn right on highway, Rte. 103.

8.5 Leave Rte. 103, by turning left on road to Brockways Mills.

8.9 Bridge and gorge right in Waits River facies of the Calciferous Schists. This was Stop 1-9 in GB-3, p. B-13. From here to Hardscrabble Corner (mi. 14.2) our route coincides, but in reverse order, with that of GB-3, p. B-12,13, passing Stops 1-8, 1-7, and 1-6 of the 1993 trip. These stops visited relatively high-grade Standing Pond with rolled garnets.

We turn right, just across bridge, on paved road.

11.1 Amphibolites of Standing Pond are visible in pastures on right.

11.9 Height of land. Stop 1-6 of an earlier trip (GB-3) was on Cobble Hill, east-northeast of here. We are here in the extreme northwest corner of the Bellows Falls Quadrangle.

13.3 Follow paved road, which bends sharply right.

14.2 Hardscrabble Corner, proceed straight ahead, downhill.

15.5 Continue straight ahead.

15.7 Turn right on highway, Rte. 11.

17.6 Howard Johnson's. Zero odometers.

0.0 Continue straight ahead on Rte 11.

0.3 After passing under I-91, turn right on ramp toward I-91 north.

The cuts on I-91 between here and Exit 8 are through geology a bit too complex to describe thoroughly for 65 mph viewing. The geology over the first six miles was the subject of an early guidebook (GB-1, Thompson, 1954). The outcrops described in that guidebook were prior to the cuts on I-91, which was then little more than a gleam in President Eisenhower's eye. Because this geology would make more sense *after* we have visited Stop 4, we shall defer our brief descriptions of most of the major features until after Stop 4, when we return via the southbound lane to Stop 5. Some highlights better seen in the northbound lane are:

4.0 The large cut and the rocks exposed in a road metal quarry just east all belong to the Fitch Formation (calc-silicate rocks, carbonates, calcareous granulites and amphibolites of Silurian age). These exposures are close to the ones at Stop 6 in GB-1. The Fitch is here in the hinge region of the Skitchewaug nappe, a major fold nappe that carries older rocks many kilometers west over the Devonian Littleton Formation. For further discussion see GB-1 and Thompson and others (1968). This is all we shall see of the Skitchewaug nappe, a major feature of the New Hampshire sequence, that belongs to terranes more easterly than those under consideration today.

7.0 A fine view of Mount Ascutney straight ahead, a Mesozoic stock. The cuts on I-91, from here on, entering the Ascutney Quadrangle, are in the Calciferous Schists, out of range of GB-1.

10.6 Exit 8, right. Calciferous Schists exposed in interchange.

10.7 Right off ramp onto Vt. Rte. 131, east to Ascutney Village.

11.1 Turn left (north) on U.S. 5.

16.3 Turn right off U.S. 5 in Windsor and cross covered bridge over the Connecticut River, the longest covered bridge in the U.S. According to local lore, if you hold your breath and make a wish while crossing, it may come true.

16.6 Left on N.H. Rte 12-A.

17.9 Large outcrop on right is the Waits River facies of the Calciferous. Close to the east edge of the Calciferous, it has been mapped as a calcareous zone in the Gile Mountain.

18.1 Turn right off Rte. 12-A onto St. Gaudens road. Sparse outcrop of the Waits River facies over the next mile, entering the North Hartland Quadrangle. We soon pass, on the left, the St. Gaudens National Historic Site. Augustus St. Gaudens, a noted sculptor, designed the Lincoln penny here. He also created the famous monument to the black soldiers of the Civil war that stands on the Boston Common. When asked how he found a model for Abraham Lincoln in Cornish, New Hampshire, he is said to have remarked: "Everybody here looks like Lincoln."

19.8 Turn left, still on St. Gaudens Road.

20.3 Continue straight.

20.4 Turn right, following St. Gaudens Road.

21.3 Turn left, off St. Gaudens Road, down driveway to:

Stop 3: Landowner permission needed! The Read Hill section, pre-Silurian Post Pond Volcanics to the Calciferous Schists.

Vehicles with low clearance should park near house. Others may proceed northwesterly across dip on old farm road and park at east base of steep hill (el. 1105 ft.). Climb west up hill to summit. Rocks are amphibolites and felsic gneisses of the Post Pond Volcanics of Lyons (1955). In most recent compilations these have been correlated with the Ammonoosuc Volcanics (Ordovician) of Billings (1937). In sections with more complete outcrop the proportion of felsic gneiss is higher than seen here. From the summit proceed diagonally northwest down slope to an east-west property line that is also the Cornish-Plainfield town line. The rocks here (and in sparse outcrop on the descent) are westerly-dipping, rusty, sulfidic schists and granulites.

Westward along the fence line the rusty schists are succeeded by quartz conglomerate, followed by highly mafic amphibolites. The amphibolites are succeeded in turn by a narrow zone of yellowish-weathering schist containing lenses of dark, possibility manganiferous granulite. West of these schists are numerous exposures of beige to blue-gray limestone. The pits are the result of quarrying for lime by early settlers. A single

pelmatozooan columnal was found in the most southerly pit (south of the town line) by JBT and C. G. Doll, in 1965 (see Thompson and others 1968, p. 207). The limestones can be followed only a short distance south, but can be seen in extensive exposures north of Plainfield village, also on the east bank of the Connecticut river north of the Plainfield-Lebanon line, and in a long railway cut about two miles south of White River Junction, Vermont. If the section here on Read Hill is correlative with the Hawley of Spring Hill (Stop 1), the part of the Hawley above the limestones is missing.

The limestones here are succeeded, across the Monroe Line, by gray phyllites probably correlative with the Meetinghouse "Slates" of Doll (1944), the most easterly unit of the "Calciferous Schists." Is the Monroe line here a fault or an unconformity? The phyllites make up the higher summit of Read Hill to the west (not to be visited). West of its summit, downslope in farmland are the phyllites and punky-weathering limestones that were seen on the drive in just north of the covered bridge. Although they have been shown on several maps as part of the Gile Mountain Formation, they are lithologically typical of the Waits River. The sequence of rocks seen at this stop was first described (a slightly scrambled description) by C. H. Hitchcock (1908,p. 149) who recognized their probable significance, and on the basis of the limestones here and north, toward White River Junction, suggested that they might be Silurian! In his day much of the ground here would have been open pasture.

Similar sequences, on a regional scale, may be seen in the extreme northwest corner of Maine (Moench and others, 1995) and on both flanks of the Munsungun anticlinorium of Fitzgerald and Hon (1994, and references therein) still farther northeast. The limestones in these distant occurrences are closely associated with mafic volcanics, and are locally abundantly fossiliferous, containing much crinoidal debris, and other fossils characteristic of the Silurian.

Return to vans and to St. Gaudens road. Zero odometers there.

0.0 Go left (east) on St. Gaudens Road.

0.3 Turn left, downhill.

1.1 Bear right.

1.3 Go right on paved road (Stage Road) in Hell Hollow. Outcrops on left over next mile are Post Pond Volcanics.

3.1 View of Croydon Mountain ahead. It lies in a large hunting preserve known as the Blue Mountain Forest. This was founded early in the century by one Austin Corbin, a native of the town of Croydon, on the other side of the mountain. Corbin built the Long Island Railroad and Coney Island and became a rich man. The preserve contains elk and Ural Mountain wild boar. Prior to World War II, it also contained a herd of bison that was the source of the breeding stock used to repopulate several of our western national parks with bison. Escaped wild boar are not uncommon outside the preserve, one was shot some years ago in the outskirts of Hartford, Connecticut. The white quartzite ledges (Clough) visible to northbound travelers on I-89, are in the preserve, on the east slope of Grantham Mountain, the high ridge just north of Croydon Mountain.

3.3 Turn right on Penniman Road.

3.8 Continue straight ahead. Outcrops over the next 8.5 miles are beneath the axial surface of the Cornish nappe and include Post Pond Volcanics, calciferous schists, and rocks tentatively assigned on Fig. 1 to the Read Hill sequence. Though east of Read Hill, a palinspastic reconstruction, unfolding the Cornish nappe, would place them in a more westerly, presumably deeper water facies than the S-D of the Read Hill area.

6.0 Cornish Flat. Turn right (south) on Rte. 120 in the village. We are now in the Ascutney Quadrangle.

6.1 Felsic gneiss of Post Pond in pasture on right.

6.5 On right is the first of a series of roadside outcrops of problematic, gray, calcareous granulites in which the road stays on strike over the next 2.5 miles. They lie in a narrow zone, bordered on the east by the Post Pond, and are included on Fig. 1 with the Read Hill series, although resembling closely some of the Gile Mountain facies of the Calciferous.

10.4 Post Pond Volcanics on left.

11.9 Cornish-Claremont town line.

12.3 Bear left on Winter Street, leaving Rte. 120. Road bends southeasterly, crossing the Post Pond Volcanics, here carrying the axial surface of the Cornish nappe.

13.6 Park on right (uphill) side of driveway on the left for:

Stop 4: Landowner permission needed! Section from the Post Pond Volcanics to the Littleton Formation, the principal unit of the New Hampshire Siluro-Devonian.

Walk up to pasture gate on right behind house. Close gate behind you and walk south through pasture to power line. The first outcrops east on the power line are felsic gneisses and chloritic mafic rock of the Post Pond Volcanics. These are succeeded uphill to the east by rusty-weathering, carbonaceous schists that were mapped by (Chapman, 1939), as part of the Orfordville Formation of Hadley (1942). More recently (Thompson and others, 1968) these have been correlated with the Partridge Formation of Billings (1937), in either case assigned an Ordovician age.

About two thirds of the way up the hill the schists are overlain by quartz conglomerate and quartzite. This unit, though shown on more recent maps as Clough, is the Hardy Hill Quartzite, originally regarded as a member of the Orfordville Formation. The Hardy Hill was named by C. A. Chapman (1939) for exposures on the Chapman farm on Hardy Hill, northeast of Lebanon, New Hampshire. A few miles north, this unit shows cross-bedding that is consistent with tops east, into the overlying Littleton Formation. A local fault on Northey Hill was mapped by Billings (1937), and later revived, farther south, by Hadley and Chapman as a fault, more or less along the boundary between the Hardy Hill and the Littleton. We now regard this line as a stratigraphic contact, marking a major change in sedimentation, but not a major tectonic break.

The overlying Littleton Formation (Devonian) is here a gray, non-rusty schist with thin sandy laminae. It contains tiny garnets, and farther east large porphyroblasts of staurolite, the latter strikingly displayed in large road cuts on Interstate 89 southeast of Lebanon. About halfway from the quartzites to the crest of the ridge the Littleton contains small, polymict clasts, mainly of vein quartz, but some of feldspathic material and fine

granulites. The Littleton on the ridge crest is strongly folded, and the Hardy Hill makes several reappearances eastward along the power line, in north-plunging folds. Many discordant folds may also be found, however, owing to rupturing of the comparatively brittle quartzite.

The Hardy Hill, at this locality, is regarded by us as the basal conglomerate to the Siluro-Devonian rocks of the New Hampshire sequence. It is not, however, necessarily correlative, directly, with the Clough and Fitch Formations that mantle the gneiss domes farther east, and that also appear in their over-riding nappes (see Thompson and others 1968; also T. Allen, this guidebook, Trip A-2). We suggest that the name Hardy Hill be revived, allowing for the distinct possibility that it may be a Devonian quartzite, basal here to the Littleton, and hence younger than the Clough and Fitch.

The eastward-topping sequence here and the probably westward-topping sequence at Stop 3 are interpreted as on opposite flanks of an antiform in the upper limb of the Cornish nappe of Thompson and others (1968). The two sequences are, however, obviously quite different, and we have the problem of how they relate to one another if both are indeed Siluro-Devonian, as seems likely. Return to Winter Street and zero odometers.

0.0 Continue south on Winter Street, passing beneath power line.

0.7 Four Corners, continue straight ahead on Winter Street. The rocks at and west of the intersection, and in the woods southeast, are similar to the problematic calcareous granulites, noted earlier. Here, however, they are on the east side of the Post Pond, hence on the upper limb of the Cornish nappe. Their presence is a reminder that there is still a problem in tracing the hinge lines in the axial surface of the Cornish nappe.

2.1 Turn right on Rte. 11.

2.3 Bear right on Rte. 120, leaving Rte. 11. (The outcrops in the Sugar River, below the bridge on Rte. 11 provide excellent exposures of volcaniclastic rocks in the Post Pond.)

2.4 Rte. 120 bears right, but we continue straight ahead on unnumbered road. (Exposures on Rte. 120, right, near intersection, are felsic ignimbrites in the Post Pond, characterized by grains of blue, opalescent quartz.)

2.8 Continue straight ahead at four corners, Mount Ascutney visible ahead.

3.5 Bear right on New Hampshire Rtes. 103 and 12. The highway follows the Sugar River, on the left, containing exposures (beneath the Cornish nappe) of the Waits River facies of the Calciferous.

5.5 Pass beneath a high railroad bridge.

Mount Ascutney soon looms ahead. The main stock, largely syenite, is surrounded by Calciferous Schists. The smaller, mafic stock, of Little Ascutney, farther west, is surrounded by pre-Silurian rocks of the Chester dome, and contains a remarkable eruptive breccia. This breccia contains abundant large blocks that are of typical Littleton lithology, quite unlike any of the country rock now exposed at the present erosion surface (Schneiderman, 1989). It is thus analogous to the well-known

breccia (of about the same age) on Mount Saint Helens Island, in Montreal, that contains fossiliferous blocks notably younger than any of the country rocks now exposed nearby.

6.0 Intersection with Rte. 12-A. We continue straight ahead on Rtes. 103 and 12, which after crossing Connecticut River into Vermont, becomes Vermont Rte. 131.

7.0 Junction with U. S. 5. Continue straight ahead on Rte. 131.

7.5 Turn left onto I-91, southbound. The rocks at the interchange belong to the Calciferous Schists as do the numerous highway cuts over the next two miles.

10.9 Bridge over highway.

11.0 Outcrops in Barkmill Brook, below highway, are of Post Pond Volcanics in the Cornish nappe. Cuts ahead on the next hill duplicate the power line section at Stop 4.

11.3 Rusty-weathering, carbonaceous schists on right, followed on right by:

11.4 Conglomerate and quartzite of the Hardy Hill, and it, in turn, by:

11.5 Littleton Formation, gray phyllites with sandy laminae. These are well exposed, mainly in cuts on the right, over the next 2.5 miles.

14.1 Bridge over Skitchewaug Trail. Cuts below bridge are in Littleton, lying beneath the large cuts just ahead in the Fitch Formation of the Skitchewaug nappe that were noted earlier when traveling northbound. The features here are described in GB-1, Stop 6, as they were seen in 1954. The highway construction has helped in some ways, hurt in others. Striking pavement outcrops of Littleton in a pasture on the north side of the Skitchewaug Trail are in part buried, and in part obscured by growth of young pines in (former) pasture. Contact relations between the Littleton and the structurally overlying Fitch were more clearly shown in 1954. In exchange, we have the huge cuts in the Fitch in the northbound lane, and in the road metal quarry east of it.

14.9 Littleton, right.

16.3 Large outcrop in median is phyllite-matrix conglomerate in Littleton.

17.4 Exit 7, turn right on ramp to Rte 11, westbound.

17.7 Howard Johnson's. Park for Stop 5 (also Stop 1 of GB-2, and Stop 1-2 of GB-3).

Stop 5 Park in the Howard Johnson's lot. The Chicken Yard Line.

The Chicken Yard Line (see Trzcienski and others, 1992), may be seen midway in the big cut on the north side of Route 11. It is a nearly vertical contact between gray phyllites and rare sandy beds of the Littleton on the east, and greenstones and related metavolcanic rocks on the west. Several boudins of sheared, pebbly quartzite can be seen along the contact. Is this a major fault, or is it a zone of extreme thinning and attenuation with consequent boudinage of the relatively brittle quartzite? We then walk south on Route 5 to an outcrop in a clump of trees on the south side of a driveway, leading to a YMCA camp. This goes off to the right several meters south of the bridge. The exposure is of a steeply plunging fold in quartz conglomerate and quartzite. The fold opens northward to include some gray phyllite of the Littleton. Return to Route 5 and walk a short distance

south to an outcrop of greenstone on the right. This is in the core of a south-opening fold complementary to the north-opening one in the clump of trees. These define one of the steeply plunging, east-side-north folds, that offset the CYL from here south. There are also (Fig. 1) several post-metamorphic faults that produce offset in the same sense. A few meters farther south the Littleton reappears, east of the greenstone, and includes some phyllite-matrix conglomerates that may be compared to those seen at Stop 4.

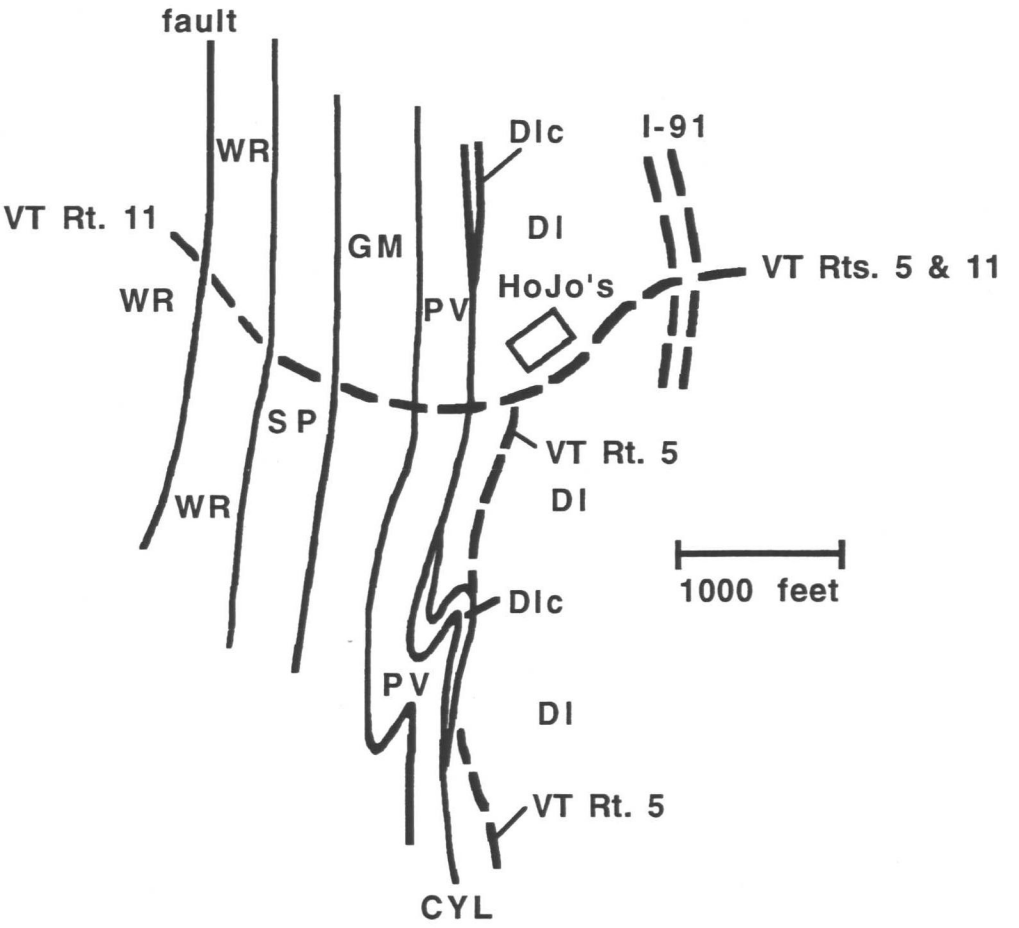


Figure 3. Geological sketch map of the Stop 5 area. WR, Waits River Fm.; SP, Standing Pond Volcanics; GM, Gile Mountain Fm.; PV, Putney Volcanics; Dl, Littleton Fm.; Dlc, quartz conglomerate in the Littleton Fm.; CYL, Chicken Yard Line, separating Vt. and N.H. sequences. HoJo's is Howard Johnson Restaurant.

Those who have time may return to the CYL exposure on Rte 11. The metavolcanics abutting the CYL on the west have been shown on earlier maps as the Putney, but we now think they are probably correlative with the Standing Pond. A number of miles to the south the "Putney" and some associated granulites have primary features (Hepburn and others, 1984) showing tops east. West of the "Putney" in the gully along the north side of Rte. 11 are calcareous phyllites and granulites, a variant of the Gile Mountain facies of the Calciferous. The next large cut (0.3 mi. west of H. J.'s) on Rte. 11 (see Fig. 3) has a nearly vertical contact between gray calcareous phyllites of the Gile

Mountain on the east and greenstones of the Standing Pond Volcanics on the west. We suggest that the Gile Mountain between the two metavolcanics is anticlinal, carrying the trace of the axial surface of the Cornish nappe. If so, the anticlines of Hawley-type rocks in the Whately area of Massachusetts (Trzcienski and others, 1992), may be a re-emergence of these rocks farther south along the same axial trace.

Those who *still* have time may drive west across the bridge on Rte. 11 in order to see outcrops (0.9 mi. from Howard Johnson's) of the Waits River facies of the Calciferous in cuts on the south side of the highway (Stop 2 of GB-2). If our present beliefs are true, the Waits river facies here would be younger than the Gile Mountain facies on the other side of the last Standing Pond. Because the facies boundaries are certainly not time lines this poses no inconsistency with the observations of Fisher and Karabinos (1980) along the White River.

Drive safely on your way home!

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STRATIGRAPHY AND STRUCTURE OF THE ORWELL 7.5' QUADRANGLE, WEST-CENTRAL VERMONT

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INTRODUCTION

The geology of the Orwell 7.5' quadrangle (1:24,000) has never been published in its entirety at any scale below 1:250,000 (the Centennial Geologic Map of Vermont, Doll et al., 1961). Recent mapping (Mango, 1997b) has attempted to draw together past mapping efforts of various portions of the quadrangle, fill in what blanks remained, and simplify the stratigraphy such that formation names used in this region are the same as those used in similar stratigraphies to the north and west. The structure of the quadrangle has also been generalized, so that the major structural features will be identifiable when shrunk to 1:250,000 scale (i.e. the scale of the forthcoming new state geologic map). These simplifications also recognize the limitations of this worker's schedule and ability. The original impetus for undertaking this project was that Professor Andrew Raiford of Castleton State College had begun mapping the Orwell region in the 1970s for a Ph.D. that was never completed. When Professor Raiford died in 1993, it appeared that nobody would finish his mapping in time for the new state map unless this worker volunteered.

Previous mapping and structural and stratigraphic descriptions of the Orwell region include the following published works. Brainerd and Seely (1890) described and mapped the geology of part of East Shoreham and environs, just to the northeast of Orwell. Their "Calciferous formation" is equivalent to the lower half of the stratigraphic section exposed in the Shoreham-Orwell region (Washington and Chisick, 1988). Cady (1945) mapped a large region of western Vermont, reaching just far south enough to include some spotty coverage of the Orwell quadrangle. Unfortunately, blank areas in his map include some of the most puzzling areas of the Orwell quadrangle. Welby (1961) mapped the eastern side of the Lake Champlain valley in great detail; his work extends as far south as the Addison County/Rutland County line, and includes about a third of the Orwell quadrangle. Washington and Chisick (1988) give a very detailed description of the stratigraphy of the Beekmantown Group in the Shoreham region and to the north, but include no map. Professor Raiford left behind various field maps and sketches.

This trip is intended to be a low-key outing that will visit outcrops of the various formations of the Orwell quadrangle, in an attempt to show the changing geologic environment of the region from the late Cambrian through the late middle Ordovician. This geologic history is manifested in a variety of sedimentary rock types (some slightly metamorphosed) and an interesting structural regime.

STRATIGRAPHY

The Orwell quadrangle is at an intersection of previous mapping efforts originating from the south, north and west, with the result that one rock type might have two or even three different names from publication to publication, depending on who was doing the mapping. Therefore, formation names for the current mapping were chosen based on attempts to both simplify the stratigraphy and utilize names most commonly in use. In several cases, the description of lithologies in Orwell more closely matched descriptions of equivalent New York State lithologies (only a few miles away) than they did Vermont lithologies, whose descriptions where often based on outcrops up to one hundred miles away from Orwell. In addition, sparse and discontinuous outcrop in Orwell meant that defining units to the member level would invite misidentification.

The following table gives the names of the formations of the Orwell quadrangle, and those of equivalent formations found elsewhere in Vermont as listed in previous publications (Cady, 1945; Welby, 1961; Doll et al., 1961):

Formation Name	Equivalent Formation(s)
Iberville	Hortonville, Ira
Stony Point	
Glens Falls	
Orwell	
Middlebury	Day Point, Crown Point, Valcour
Providence Island	Burchards, Beldens, Bridport, Weybridge
Fort Cassin	Bascom
Cutting	
Whitehall	Shelburne
Ticonderoga	Clarendon Springs

The formations range in age from late Cambrian to late middle Ordovician (Doll et al., 1961). The Whitehall, Cutting, Fort Cassin and Providence Island Formations belong to the Beekmantown Group, the Middlebury Formation belongs to the Chazy Group, the Orwell Formation belongs to the Black River Group, and the Glens Falls, Stony Point and Iberville Formations belong to the Trenton Group.

Lithologic Descriptions

The following descriptions are based on observation of outcrops within the Orwell quadrangle. Thicknesses of units are taken from Welby (1961).

Ticonderoga Formation
Light and dark gray dolomite and sandy dolomite; fine- to medium-grained; beds 1-4 ft (0.5-1.5 m) thick; contains knots of white quartz, rounded quartz grains and thin quartz sand layers; dull to vitreous luster; can be extremely hard; smooth-weathering; weathers dark gray, bluish gray and yellowish gray; 300 ft (100 m) thick.

Whitehall Formation
Light gray siliceous dolomite; minor limestone beds; coarse to fine grained; partly laminated; massive bedding; 210 ft (70 m) thick.

Cutting Formation
Nondescript, dark to medium gray dolomite; massive beds; cross-bedded quartzite near base (most obvious on weathered surfaces); abundant blue-black chert in upper parts; different beds have either a vitreous or a dull luster; dull-weathering layers have fetid odor; upper part weathers yellowish gray; 440 ft (150 m) thick).

Fort Cassin Formation
Silty and sandy limestone; dark gray and dark bluish-gray; beds 1-2 ft (0.3-0.6 m) thick; some fossils, fossil fragments and pyrite; weathered edges appear banded or ridged; 210 ft (70 m) thick.

Providence Island Formation
Limestone and dolomite; dark gray to light gray; some shale laminae; massive to moderately bedded, with beds 1-3 ft (0.3-1 m) thick; fetid odor; weathers a distinctive orange (these layers are also scored on weathered surfaces) or bluish-gray (these layers weather to look like elephant skin); dolomite is dense and vitreous; 450 ft (150 m) thick.

Middlebury Formation
Dark blue-gray limestone; weathers lighter gray; minor fossils; beds 2-4 in (5-10 cm) thick; looks like slate when deformed; thickness estimated here at 150 ft (50 m).

Orwell Formation
Black limestone; finely crystalline; massive bedding; weathers light gray; contains some fossil fragments in bands, and calcite veins; 40 ft (15 m) thick.

Glens Falls Formation
Black or dark bluish-gray limestone and shale, in couplets; thin bedded, with beds 4-6 in (10-15 cm) thick; beds pinch and swell; some beds highly fossiliferous; weathers bluish-gray to olive gray, with pits and ridges; 450 ft (150 m) thick.

Stony Point Formation
Black, carbonaceous shale, often calcareous; thinly bedded to laminated, with beds up to 3 in (7 cm) thick; "pinstriping" due to interspersed calcite veinlets; 1000 ft (330 m) thick.

Iberville Formation
Black or dark gray noncalcareous shale and slate; can be pyritic, carbonaceous or phyllitic; thin bedded, up to 0.5 in (1 cm) thick and thin cleaving; red-brown weathering; 1000 ft (330 m) thick.

"Taconic" Formations
Equivalent (?) with part of the section are rocks listed here simply as "Taconic". The northwestern part of the Sunset Lake slice of the Taconic Allochthon occupies the southeast sixth or so of the Orwell quadrangle. We will be visiting both the Sunset Lake slice and a newly discovered Taconic sliver just to the north of the Sunset Lake slice. Stratigraphic descriptions of formations within the Taconic allochthon are found in Doll et al., (1961), Zen (1964), Shumaker and Thompson (1967), Rowley and Kidd (1981), Stanley and Ratcliffe (1985), Baldwin and Raiford (1987), and references therein. It is not the goal of this trip to study these rocks in any detail, but rather to show the structural relationship between the Taconic allochthon and the other rocks of the Orwell region, and to point out some of the difficulties encountered while mapping in and around these rocks.

GEOLOGIC HISTORY

The tectonic history of west-central Vermont is described by, among others, Rowley and Kidd (1981), Stanley and Ratcliffe (1985), Mehrtens (1985), Thompson (1990), Washington and Chisick (1988), Zen et al. (1989) and Stanley (1990). The following very brief summary is based on these works and highlights what can be seen in the Orwell quadrangle. By the late Cambrian, a passive margin was established on the eastern coast of North America. (The edge of the Precambrian basement rocks of the Adirondacks is currently exposed just across Lake Champlain from the Orwell quadrangle.) Shallow, warm water allowed the deposition of thick layers of limestone and dolomite; the presence of windblown sand grains and sandy layers indicates that the shore was quite nearby. This environment existed from the late Cambrian until the beginning of the middle Ordovician. During this time, the lithologies changed slowly from mostly dolomite to mostly limestone, perhaps reflecting climatic changes that initially favored dolomitization of limestone but later did not. Stratigraphically, these lithologies are the Ticonderoga through Orwell Formations. The presence of shale layers in the subsequent lower middle Ordovician Glens Falls Formation indicates the onset of orogenic activity to the east of the passive margin (i.e. the beginning of the Taconic orogeny). Downwarping of the margin into an east-dipping subduction zone caused water depths to increase, such that limestone deposition eventually ceased and was replaced by the deposition of great thicknesses of shale (Stony Point and Iberville Formations). In the Orwell region, this orogenic event culminated in the placement of "Taconic" allochthonous rocks on top of the margin's sediments. It is also manifested in the slight metamorphism of some of the units in the eastern part of the quadrangle.

ACKNOWLEDGEMENTS

I am very grateful to Char Mehrtens for spending several days with me in Orwell. Her knowledge of the stratigraphy and her encouragement were both very valuable. I am also grateful that Art Goldstein could spend an afternoon showing me some of the features of the Sunset Lake slice, and helping me puzzle out the placement of the Hibbard Hill klippelet. I also thank Arlen Bloodworth for coming out in the field with me. And through all my Orwell endeavors, I have been truly sorry that I couldn't talk to Andy Raiford; I'm sure sure that a tremendous amount of interesting and useful geologic information died with him, and I miss his humor and good nature.

ROAD LOG

To reach the starting point from Rutland/Killington/Pico, take Route 4 west from Rutland to Exit 2. Take Route 22A north for 14 miles. Turn left (west) on Route 73. After 0.3 miles, Route 73 curves right; leave Route 73 and continue straight on Mount Independence Road. After 3 miles, the road forks; stay right. The road will turn to gravel. 5 miles (total) after leaving Route 22A, the gravel road forks again; take the left fork that bend sharply and goes up a little hill. The Mount Independence parking lot is a few hundred feet further along on the left.

(The following history is condensed from the Mount Independence tourist brochure.) Mount Independence is a Revolutionary War site. A fort was built here to guard against British attack from Canada because Fort Ticonderoga, across Lake Champlain, offered only a poor view to the north. American troops began clearing land on Mount Independence in June of 1776. By fall, three brigades (12,000 troops) had established camps, and had built two batteries and a picket fort. The combined strength of Mount Independence and Fort Ticonderoga caused the British to retreat quickly to Canada in October, 1776. Many American soldiers went home that winter, reducing the force to 2,500. Many of those remaining were sickly, and a number froze to death during the winter. Not many replacements reached the fort in the spring, so when British troops staged an attack in July, 1777, the Americans evacuated Mount Independence. British and German troops remained at the fort until learning of the British surrender at Saratoga, and they retreated back to Canada, burning the structures of Mount Independence as they left.

Mount Independence is separated geologically from the rest of the Orwell quadrangle rocks by a fault that brings these upper Cambrian and lower Ordovician rocks in contact with the younger Providence Island and Middlebury Formations to the east (Figure 1).

Mileage

Begin in the parking lot of the Mount Independence State Historic Site. The first three stops will be visited on foot (see Figure 1). **PLEASE DO NOT USE HAMMERS ON ANY OF THE MOUNT INDEPENDENCE OUTCROPS!** The round-trip distance is about 2.5 miles, on established paths over mostly gentle terrain. Reference will be made to the Mount Independence trail names and stop numbers; a map is available at the visitor's center. Walk 800 ft (250 m) up the trail toward the Outpost. The outcrop is about 20 ft (6 m) to the right in the tall grass.

STOP 1. TICONDEROGA FORMATION (20 minutes) The upper Cambrian Ticonderoga Formation is the oldest lithology we will visit. Here, the Ticonderoga is a massively bedded, medium gray, coarsely crystalline dolomite. A prominent quartz sand layer indicates that bedding is nearly vertical. Quartz grains are also sprinkled throughout the dolomite. The quartz grains are rounded and frosted, suggesting proximity to a quartz sand beach. The beach represented by the Potsdam Formation, a cross-bedded sandstone and quartzite that outcrops in numerous locations close by to the west and south, is an obvious source. Both the dolomite and the presence of beach sand suggest a warm shallow sea as the environment of deposition.

Walk about 700 ft (210 m) up to the Outpost. Turn left on the Blue Trail. Walk 0.7 mi (1.1 km) to Stop 5 of the Blue Trail. (Don't be fooled by the black slate in the trail on the descent; it was brought in to help stop erosion.) The Mount Independence trail guide quotes General St. Clair's papers, referring to a water source "on the west side of Mount Independence, on the low ground, near the lake; it was very steep and about one half a mile distant from the fort". This spring was perhaps the major water source for the fort. The stone slabs are thought to be steps leading down to the spring. Take a quick look at these steps - they are made of Whitehall Formation, showing fine laminations in a mostly massive dolomite. The laminations are due to small variations in grain size.

Walk 300 ft (100 m) to Stop 3 of the Blue Trail. This quarry dates from the 1750s, when the French controlled the Champlain Valley. It is thought that rock quarried here was dragged over the ice to build Carillon, the French fort that the British renamed Fort Ticonderoga.

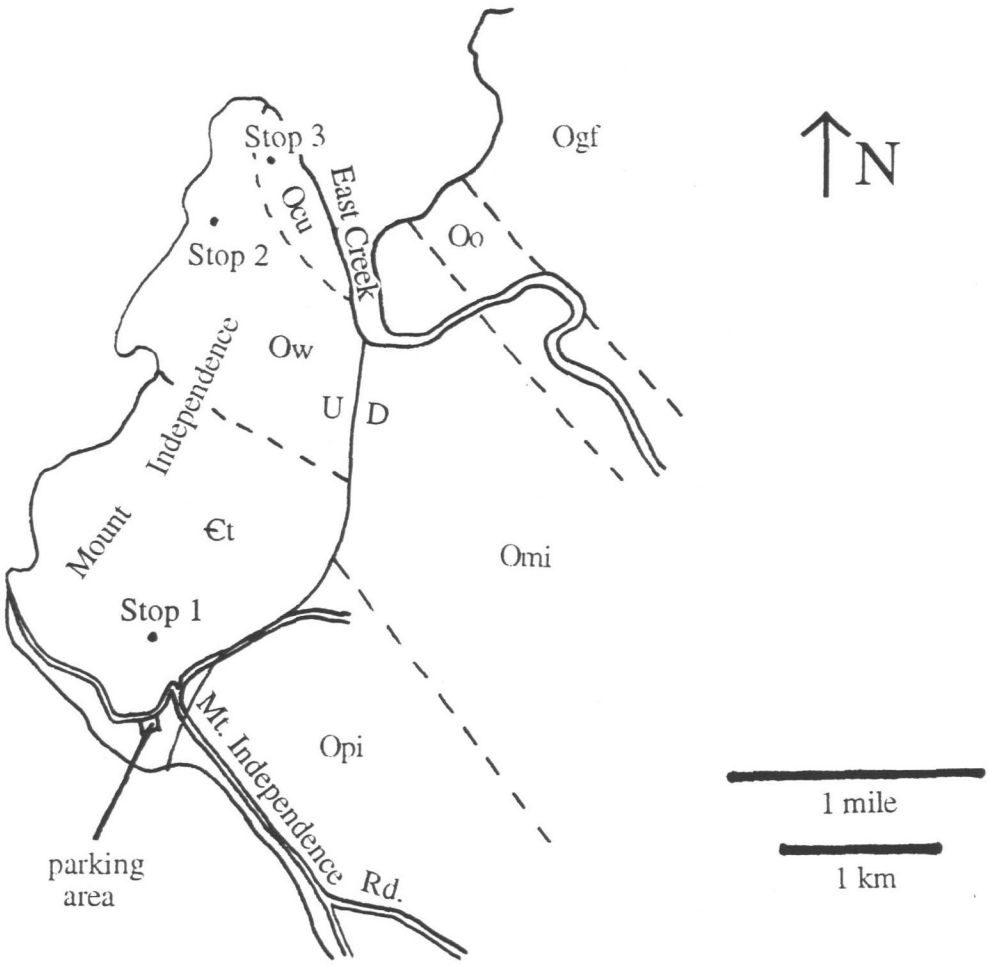


Figure 1. The peninsula of Mount Independence is separated structurally from rocks to the east by a fault that brings Cambrian and lower Ordovician Ticonderoga (Et), Whitehall (Ow) and Cutting (Ocu) Formations up against lower/middle Ordovician Providence Island (Opi) and Middlebury (Omi) Formations. Other formations in the region are the Orwell (Oo) and Glens Falls (Ogf) Formations.

STOP 2. WHITEHALL FORMATION (20 minutes) The Whitehall Formation here is a sparkly, sugary, coarse-grained dolomite. It is thick bedded to massive and medium gray. There are some knots of white calcite, and occasionally quartz grains scattered about. A subtle mottling of lighter and darker patches reflects a slight variation in grain size, with the darker areas being coarser. Washington and Chisick (1988) suggest the Whitehall Formation was deposited in tidal channels, based on sedimentary evidence (such as cryptalgal masses, conglomerates, skeletal hash and cross-stratified silty layers) visible in Whitehall, New York. Therefore, there was a slight regression of the sea after deposition of the Ticonderoga Formation.

Continue 875 ft (275 m) to the end of the Blue Trail. Go right on the Orange Trail. After 150 ft (50 m), come to a T. Go right, toward the Horseshoe Battery. Go 70 ft (20 m) uphill.

STOP 3. CUTTING FORMATION (20 minutes) This small outcrop in the path shows the distinctive cross beds that are indicative of the base of the Cutting Formation (Cady, 1945; Welby, 1961; and others). These cross beds are defined by fine quartz sand in a light gray dolomitic matrix. The presence of cross-stratification suggests a tidal/shoreline environment of deposition. Historical documents and archeological evidence indicate a source of chert along East Creek, which flows just to the east and north of Mount Independence. Chert is common in the middle and upper portions of the Cutting Formation. Since East Creek flows mainly through middle Ordovician Stony Point Formation (a shale) and Orwell Formation limestone, it is likely that the only possible chert locality along East Creek is the northeastern edge of

Mount Independence, which would therefore consist of Cutting Formation. We will see more Cutting Formation at the next stop.

Continue south along the Orange Trail. In 125 ft (40 m) you will arrive at the Horseshoe Battery (Stop 4 of the Orange Trail). From here there is a nice view of Fort Ticonderoga across Lake Champlain. Walk another 1.2 miles to get back to the parking lot.

0.0 Go right out of the parking lot. At the base of the hill is an outcrop of Ticonderoga Formation. Welby (1961) says this is Potsdam Formation, but the presence of dolomite suggests that it is more likely a siliceous layer of the Ticonderoga Formation. It is mapped as Ticonderoga Formation on the Centennial State Map (Doll, 1961).

3.1 Cross the Orwell Thrust. The outcrop at the base of the rise to the right is Glens Falls Formation, and at the top of the rise is Providence Island Formation.

4.7 Go left on Route 73.

6.4 Route 73 bends left; go straight on Hemenway Hill Rd.

8.0 Park along the side of the road, on the west side of Hardigan Hill. **BEWARE OF POISON IVY!**

STOP 4. ORWELL THRUST (60 minutes) The Orwell Thrust is perhaps the major structural feature of the Orwell quadrangle. It is a southern continuation of the Champlain Thrust. In general, it places mostly lower Ordovician rocks on middle Ordovician rocks. The Orwell Thrust here has placed lower Ordovician Cutting Formation on middle Ordovician Stony Point Formation. On the west side of the road, in a little stream gully, there is an outcrop of Stony Point Formation. This is a middle Ordovician black and dark gray, partly calcareous shale with slightly wavy bedding. The Stony Point will be discussed in more detail at Stop 8. Cross back over the road and climb about 750 ft (225 m) uphill through the woods. You will cross over a lot of Stony Point float on your way up. The first decent outcrop is moss-covered Cutting Formation. At this outcrop and others going up to the top of Hardigan Hill, there are good exposures of various parts of the Cutting Formation, including dolomites, sandy limestones and quartz sandstones with limy cement and limestone clasts, with some layers containing chert nodules that have a swiss-cheese-like appearance. Washington and Chisick (1988) suggest a subtidal to supratidal environment for the deposition of the Cutting Formation (they describe "paleokarst" features in the Cutting Formation in East Shoreham).

8.0 Turn around in pull-out on west side of road. Go back south.

9.6 Rejoin Route 73 and continue south.

9.7 Just after crossing a little bridge (Hough Crossing), park along the side of the road. Be careful opening doors and crossing here; there isn't a lot of traffic, but the road isn't very wide and there's not much shoulder. Walk to the northeastern end of the bridge and scramble down the side. **BEWARE OF ELECTRIC FENCE AND POISON IVY!** The outcrop is under the bridge.

STOP 5. FORT CASSIN AND PROVIDENCE ISLAND FORMATIONS (20 minutes) The bed of East Creek marks the contact between the Fort Cassin Formation (below) and the Providence Island Formation (above). The Fort Cassin is a fossiliferous limestone here; on the smooth stream bed outcrop you can find gastropods, cephalopods and lots of worm burrows (Figure 2). Here, the Providence Island is a medium- and dark-gray, fine-grained, medium- to thick-bedded, fairly featureless dolomite. The Providence Island Formation consists of several different rock types, and has often been subdivided into the Bridport, Weybridge, Beldens and Burchards Members (Cady, 1945; Welby, 1961; Washington and Chisick, 1988). The outcrop here is equivalent to Bridport.

9.7 Continue south on Route 73.

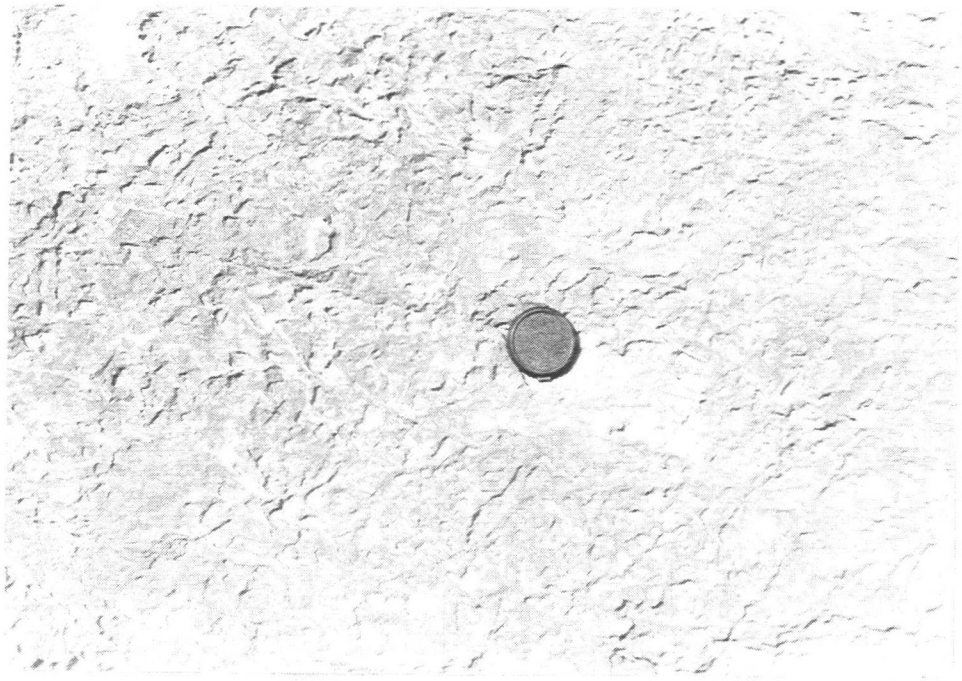


Figure 2. Worm burrows in the Fort Cassin Formation. Lens cap for scale.

11.7 Cross Route 22A.

12.0 Village of Orwell and **LUNCH STOP**. Park along road on either side. There is a small general store.

After lunch, walk from store east along Route 73 for 0.25 mile (0.4 km). The next stop consists of three outcrops along the road, with the first being on the south side as the road curves and goes uphill. The second outcrop is another 575 ft (350 m) east along the road on the south side. The third outcrop is another 300 ft (100 m) along the road, on the north side. **BEWARE OF POISON IVY!**

STOP 6. MIDDLEBURY, ORWELL AND GLENS FALLS FORMATIONS (45 minutes) The first outcrop is Middlebury Formation. The Middlebury is the only Chazy Group formation that occurs in the Orwell quadrangle. It is therefore equivalent to the entire Crown Point section of New York. The Middlebury Formation is a medium-gray, fine-grained limestone. Its most diagnostic feature is the slaty cleavage that forms when the formation has undergone a little deformation (Figure 3). In fact, it is very easy to mistake for slate if you forget your acid bottle. Unlike the Crown Point section, the Middlebury Formation contains very little fossil material. There are some crinoid and brachiopod fragments scattered throughout the outcrop. The second outcrop is the Orwell Formation (Black River Group). This formation is very thin (estimated at about 40'; Welby, 1961) and does not form very prominent outcrop. The Orwell Formation is a medium-gray limestone, with light gray weathering and a massive appearance. Fossils are fairly rare. The third outcrop is of Glens Falls Formation (the base of the Trenton Group). Its characteristic feature is interlayered limestone and shale "couplets". The couplets are usually about half an inch thick, and are evidenced by subtle light and dark striping or by ridging on weathered surfaces. This outcrop also contains a fair amount of fossil fragments, and some dark spots that are probably worm burrows. These three formations represent a change from predominantly dolomite to limestone in the stratigraphy. In terms of depositional environment, the limestones indicate that whatever climatic and depositional conditions that had prevailed throughout the earlier time of dolomite deposition had ceased to exist. Water depth could have increased, thereby decreasing evaporation and related dolomitization. Because of the presence of some fossils in the Middlebury and Orwell Formations, these units were deposited in the photic zone, but subtidal. The presence of shale and fossil fragments in the Glens Falls Formation suggest deposition below the photic

Figure 3. Slaty appearance of Middlebury Formation. Lens cap for scale.

zone, perhaps as turbidites (Baldwin and Mehrtens, 1985; Baldwin, 1987). We will see another part of the Glens Falls Formation at the next stop. This outcrop also illustrates normal faulting: there is a nice set of slickensides showing up-down movement, as well as considerable calcite vein material. This fault is one of several that offset sections of Providence Island/Middlebury/Orwell/Glens Falls (Figure 4).

12.0 Turn around. Go west on Route 73.

12.3 Cross Route 73.

12.6 Route 73 bears right (past intersection with Mount Independence Road).

14.2 Cross Stop 5. Continue on Route 73 past road to Stop 4.

18.0 At T intersection, go left (west) on Route 74.

18.4 Larrabee's Point. Park in little lot next to stone building. Do not continue down to the ferry landing. The outcrop is along the shore of Lake Champlain.

STOP 7. LARRABEE'S POINT MEMBER OF GLENS FALLS FORMATION (20 minutes)
The Larrabee's Point member of the Glens Falls Formation contains layers that are highly fossiliferous. These layers are interspersed with non-fossil-bearing, lime mud layers. The lime mud layer is closer to the parking area, and the more fossiliferous layer is closer to the water.

WARNING: If you plan to visit this site in the spring, check first on the water level of the lake; the fossiliferous layer is often submerged during wet seasons!

Because the fossils are mostly broken and otherwise reworked, Baldwin and Mehrtens (1985) and Baldwin (1987) suggest that these layers represent turbidite graded beds, with material transported from nearer-shore environments. Of the many fossils, the most obvious are brachiopods (orthid and rhynchonellid), bryozoa ("stick" bryozoa looking like little twigs, and Prasopora, looking like large gumdrops) and trilobites (mostly fragments of Cryptolithus, whose identifying features include dots on the head region) (Baldwin and Mehrtens, 1985). Figure 5 contains drawings and a photograph of some of the fossil types. The fossils found at this location are described in detail in Welby (1962) and Baldwin and Mehrtens (1985).

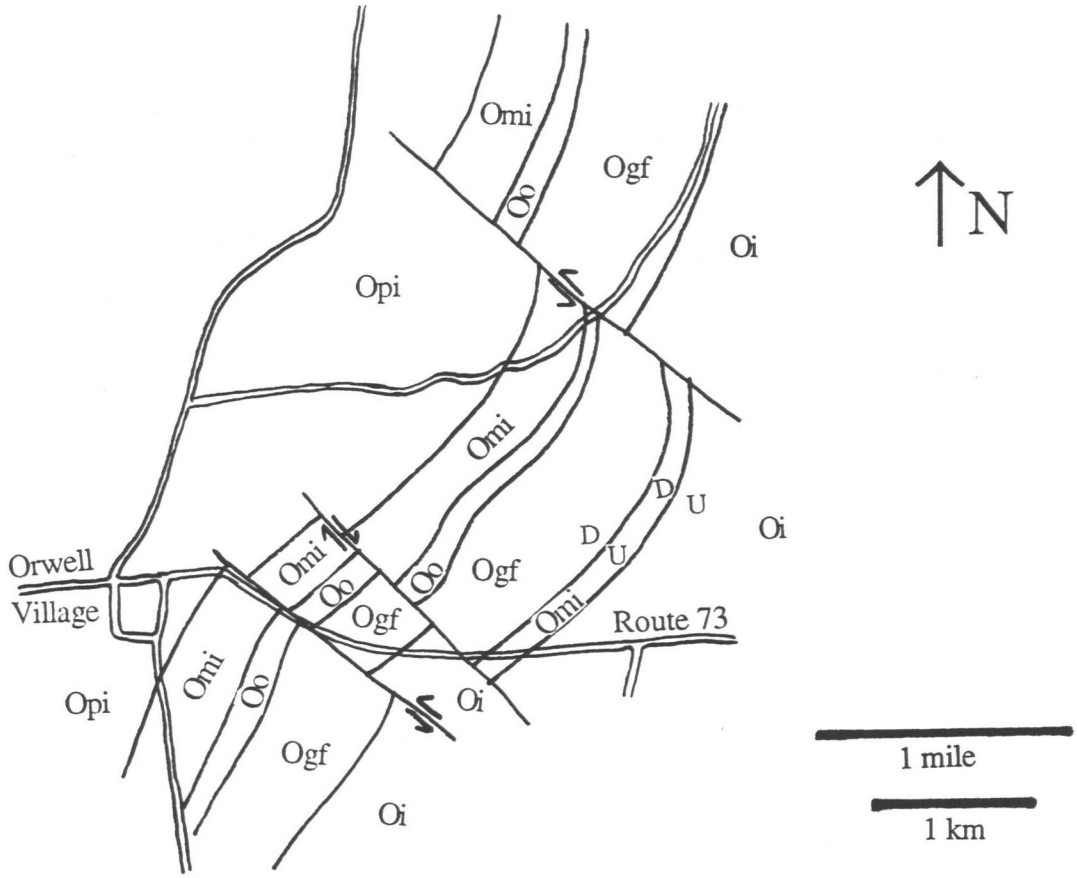


Figure 4. Structure just to the east of the village of Orwell. Three faults offset continuous sections of Providence Island (Opi), Middlebury (Omi), Orwell (Oo), Glens Falls (Ogf) and Iberville (Oi) Formations. There are also what appear to be two dip-slip faults that bring up a sliver of Middlebury Formation.

18.4 Turn around. Go east on Route 74.

18.9 Turn right (south) on Route 73.

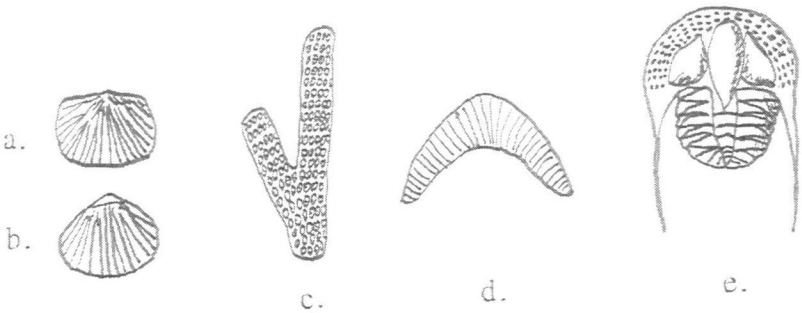
19.5 Turn right on dirt road. There is a sign saying "Vermont Department of Fish and Wildlife Access Area".

19.7 Pull into parking lot. Outcrop is just to south, along a small rise.

STOP 8. STONY POINT FORMATION (20 minutes) This is a good outcrop of typical Stony Point Formation. It is a black, calcareous shale, exhibiting white pinstriping formed by calcite veinlets. The Stony Point is widespread in the Orwell quadrangle, but mostly forms outcrops only along the lake shore. The shales of the Stony Point mark a change from dominant limestone deposition to dominantly detrital deposition. This marks the increase in water depth accompanying the beginning of subduction to the east as part of the Taconic orogeny.

19.7 Go back toward Route 73.

19.9 Turn right (south) on Route 73.



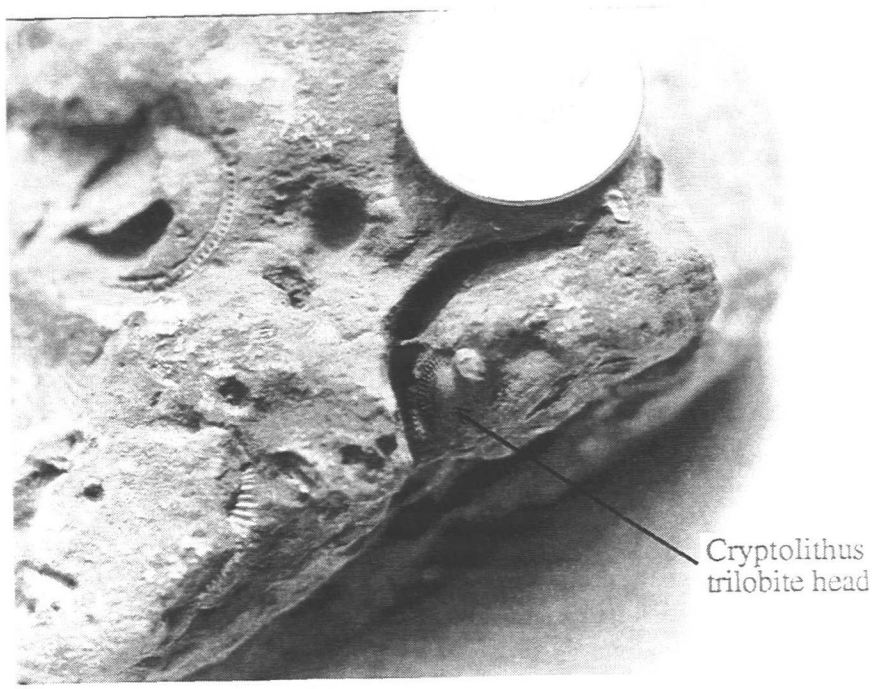


Figure 5. Fossils of Larrabee's Point. The drawing at the top shows some of the major fossil types, at approximately actual size: a. orthid brachiopod, b. rhynchonellid brachiopod, c. "stick" bryozoan, d. Prasopora bryozoan, e. Cryptolithus trilobite. Drawings after Welby (1962) and Baldwin and Mehrtens (1985) The photograph shows many brachiopod fragments and the head of a Cryptolithus trilobite. Penny for scale.

22.2 The white cliff visible to the north across the fields is Cutting Formation, showing the edge of the Orwell thrust. This cliff is slightly to the south of Stop 4.

25.0 Cross Route 22A.

26.4 Turn right (south) on dirt road.

26.8 Park along side of road. The outcrop is in the ditch on the east side of the road.

STOP 9. IBERVILLE FORMATION (20 minutes) Not all Iberville Formation outcrops this badly, but a lot of it does. The Iberville is a black and dark gray, noncalcareous shale. It is often carbonaceous. Minor metamorphism, probably associated with the Taconic Orogeny, has turned some of the formation into slate. The formation often displays rusty weathering. The Iberville Formation is also known as the Hortonville or Ira Formation, depending on whose work is referenced (see, for example, Shumaker and Thompson, 1967; Zen, 1964; Cady, 1945). (The Hortonville Formation is named for the village of Hortonville just to the south; this village is now known as Hortonia. There is another Hortonville in Vermont, located southeast of Rutland.) Cady (1961) estimates the Iberville Formation to be at least 1000 ft thick. This great thickness of fine-grained clastic material is further evidence of a deepening trough as subduction to the east increased as the Taconic arc drew closer.

26.8 Continue south along the dirt road.

27.7 At intersection, turn left (east) on Martin Hill Road.

28.2 Park at the corner of Sunrise Lake Drive.

STOP 10. SUNSET LAKE SLICE OF TACONIC ALLOCHTHON (15 minutes) This is a quick stop, to show an outcrop of Taconic slate (Hatch Hill Formation) at the edge of the Sunset Lake Slice. Note the black slate in this outcrop, and think back to the Iberville Formation of Stop 9. Other allochthonous formations that also contain black slate are the Browns Pond, Mount Merino and Poultney Formations. These formations often exhibit rusty weathering. One difficulty encountered in mapping the Orwell quadrangle was determining when a black, slaty outcrop was allochthonous (i.e. "Taconic") and when it was autochthonous (i.e. Iberville Formation). This is not a new dilemma (see Zen, 1961; Shumaker and Thompson, 1967; and Zen, 1972). Chemical analysis of samples from both allochthonous and autochthonous lithologies (similar-looking rocks from well-identified localities) showed no diagnostic chemical signature that could be used to differentiate one from the other (Mango et al., 1996). Work continues on this subject.

28.2 Turn around. Go west on Martin Hill Road.

28.6 Turn right (north) on Raymond Road (the same road you just came up; it has no road sign at the other end).

30.0 Turn right (east) on Route 73.

30.4 Turn left (north) on Fisher Road.

30.9 Park on side of road. Outcrop is on west side of road, just north of a driveway on the west side.

STOP 11. HIBBARD HILL SLICE OF TACONIC ALLOCHTHON (20 minutes) This is a newly discovered (Mango, 1997a) sliver of Taconic rock. The presence of green slate (Mettawee Formation) makes this outcrop undeniably allochthonous. More Mettawee slate is present in outcrops to the west, at least a quarter of a mile into the woods, making this slice mappable. Defining the extent of the slice is difficult, because of the aforementioned problem in distinguishing one black slate from another. Because of its small size, it was considered better to call the slice a "klippelet" rather than a "klippe" (Paul Karabinos, personal communication). Figure 6 illustrates the location of the Hibbard Hill Slice with reference to the Sunset Lake Slice.

To return to Rutland, go back to Route 73. Turn left (east) and take Route 73 to Route 30. Go right (south) on Route 30 until you reach Route 4. Go east on Route 4 to Rutland.

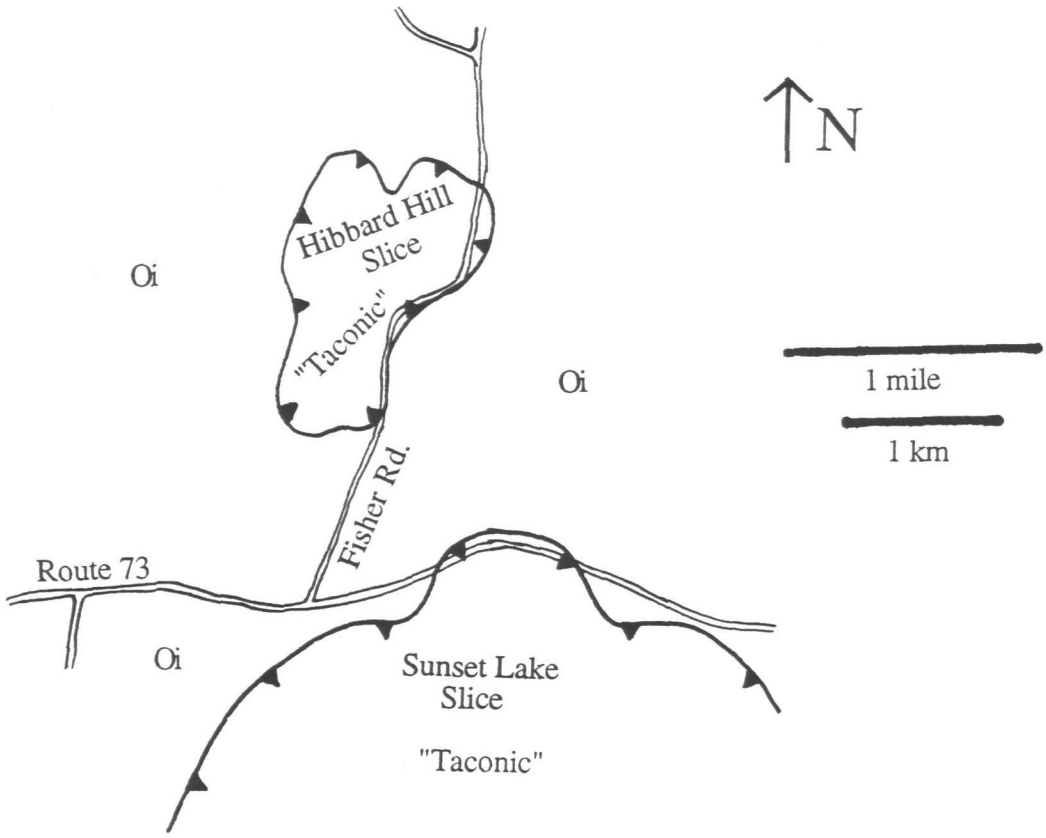


Figure 6. Hibbard Hill Slice of the Taconic Allochthon. It is identified primarily by the presence of green Mettawee slate. This "klippelet" is surrounded by autochthonous Iberville Formation, making delineation of the slice difficult where black slaty "Taconic" rocks are in contact with black slaty Iberville Formation.

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Potsdam Sandstone of the Southern Champlain Valley - Sedimentary Facies, Environments and Diagenesis

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Introduction:

The Potsdam Sandstone is exposed in the circum-Adirondack region of New York State and in adjacent areas of Ontario and Quebec. As the basal unit of the Paleozoic sequence on the craton and Iapetan shelf, its distribution and character are strongly controlled by the geometry of the pre-Potsdam erosional surface. The Potsdam is broadly subdivisible into basal non-marine arkosic facies overlain by quartz sandstone and carbonate-bearing marine facies. The Potsdam Sandstone varies in age from medial Late Cambrian to Early Ordovician from the southeastern Adirondack margin to outcrops in northwestern New York State (Fisher, 1977). The Potsdam Sandstone is the lowermost formation of the Beekmantown Group, which forms the cratonic Sauk supersequence of Sloss (1963) in New York and adjacent areas.

The Potsdam Sandstone in the northern Lake Champlain Valley is considerably thicker than in the area of this field trip (800+ feet near Plattsburgh versus approximately 250 feet maximum thickness in the southern Lake Champlain Valley). In the Plattsburgh area, Fisher (1968) subdivided the Potsdam into a lower arkosic member, the Ausable Member, and an upper quartz arenitic member, the Keeseville. The Ausable consists of terrestrial deposits; the Keeseville, shallow marine deposits (Fisher, 1968). A similar subdivision of the Potsdam is present in the southern Lake Champlain Valley, although the Ausable Member-equivalent strata are found only in a few areas.

The succeeding strata of the Beekmantown Group in the southern Lake Champlain Valley consist of quartz sandstones and carbonates, with the carbonate units often extensively dolomitized. These strata accumulated on a broad, shallow continental shelf which opened to the present east. To the present west, a vast epicratonic platform extended across much of the continent, and shallow marine sandstones and carbonates were deposited through Late Cambrian and Early Ordovician time. High global sea-level stands during this time are documented on other continental landmasses.

The Potsdam Sandstone is approximately time-equivalent to the upper Cambrian Danby and Clarendon Springs Formations of western Vermont (Mehrtens and Butler, 1987). The middle Cambrian Dunham, Monkton and Winooski Formations may be time-equivalent to the non-marine Ausable Member of the Potsdam Sandstone. Deposition of shallow shelf clastics and carbonates began earlier to the north and east in the Champlain Valley and western Vermont, and the total thickness of the Cambrian section is greater in western Vermont, consistent with the earlier subsidence of the continental margin in that area. Thicker sequences in the northern Lake Champlain Valley and in adjacent parts of Quebec and Vermont are linked to earlier, and greater subsidence related to a possible late Proterozoic aulocogen which extended into the craton from the rifting margin of Iapetus (Kay, 1975).

The Potsdam Sandstone acted as an aquifer for the transport of warm 'basinal' brines at various times after its burial. Significant diagenetic alteration of the primary depositional mineralogy has occurred, and Proterozoic basement rocks beneath the Potsdam have been altered these hydrothermal fluids (Whitney and Davin, 1987)

On this field trip we will examine a variety of sedimentary facies within the Potsdam in the southern Lake Champlain Valley. Alluvial fan/braided stream, shallow subtidal and intertidal depositional environments will be discerned. Diagenetic/hydrothermal features can be observed on outcrop scale, and timing of these relative to regional tectonic events are open to discussion and speculation.

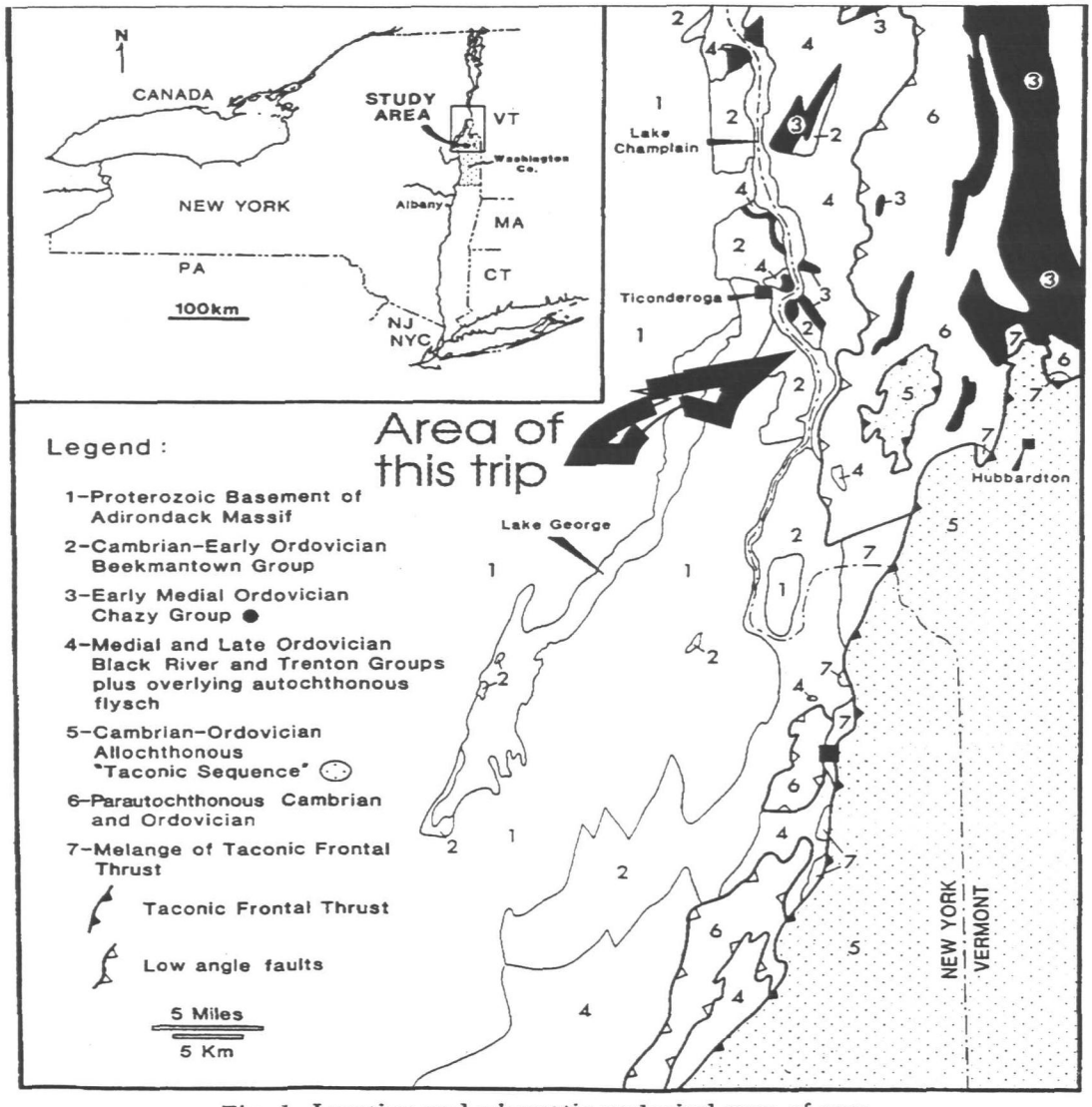


Fig. 1. Location and schematic geological map of area.

Stratigraphy and Sedimentology of the Potsdam Sandstone in the Southern Lake Champlain Valley:

Lower Potsdam Sandstone (=Ausable Member): In the area of this field trip, the Potsdam consists of basal arkosic sandstones and conglomerates (maximum 29 meters thickness) overlain by quartz sandstones with thin carbonate beds (to 40 meters). The basal arkosic unit is discontinuous in distribution, with outcrops generally limited to the area south of Ticonderoga Village in a fault-bounded outcrop belt approximately paralleling NYS Route 22. Elsewhere, the basal Potsdam consists of fine-grained marine facies. The basal arkosic unit is green to light yellow-grey, massive to thick-bedded, crudely laminated and cross-laminated fine pebble conglomerate and sandstone. Reverse and normal graded beds 5-20 cm thick are common, and trough cross-beds are present. The basal arkosic unit of the study area resembles the Ausable Member of the Potsdam Sandstone present in the northern Champlain Valley (Fisher, 1968). The depositional environment of the basal arkosic unit has been interpreted as alluvial fan/braided stream, based upon sedimentary textures, bedding thickness and scale, and primary structures (McRae, et al, 1986). The source areas of the sediment were quite local, since clasts of granitic and calcsilicate gneiss from adjacent Proterozoic basement are commonly found in the conglomerates of the basal unit.

The age of the basal arkosic unit of this area and Ausable Member of the northern Champlain Valley is not well-constrained. Fisher (1977) proposed that the Ausable and related basal arkosic facies of the Potsdam should be considered late Proterozoic (Hadrynian) in age, based on the assumption that these facies accumulated in rift basins developed on the margins of the nascent proto-Atlantic. Quartz sandstones and carbonates of the upper Potsdam Sandstone in the study area contain medial Late Cambrian (Franconian) trilobites (Flower, 1954). These same units contain beds of locally-derived coarse detritus, suggesting that the paleotopography of the source area had not changed dramatically between deposition of the basal non-marine facies and overlying marine units. McRae, et al (1986) used paleomagnetic techniques to determine a poorly-constrained early Paleozoic depositional age for the Ausable Member and related basal Potsdam facies in the circum-Adirondack Region. However, the intense post-depositional alteration of the Potsdam Sandstone in this area suggests that the depositional signal was not preserved, and the remanence measured is likely a diagenetic artifact.

Diagenesis: The green basal Potsdam facies contains ubiquitous iron-rich chlorite (Selleck, 1997) that indicates diagenetic temperatures in excess of 200 degrees C. Quartz cements and overgrowths of feldspar on detrital grains are abundant. Titanium oxide cements (anatase) are common. Associated authigenic REE minerals, sulfides (galena, sphalerite), fluorite and barite suggest that MVT fluids were transmitted through porous and permeable basal sands, causing the extensive hydrothermal alteration. Primary detrital minerals such as hornblende, biotite, plagioclase and ilmenite are commonly altered to complex mixtures of chlorite, anatase, K-feldspar and pyrite. Fluid inclusions in quartz veins which cross-cut the basal Potsdam have homogenization temperatures exceeding 250 degrees C. and salinities near or at halite saturation (Collins-Waite, 1991). Whitney and Davin (1987) suggested that hydrothermal alteration of Proterozoic basement gneisses in the Fort Ann, New York region was the result of expulsion of connate brines from Cambrian-Ordovician sedimentary rocks during the Taconic Orogeny. No absolute age dating of the authigenic phases in the Potsdam Sandstone of the southern Lake Champlain Valley has been done. Rb-Sr and K-Ar dates on authigenic illite from the basal Potsdam of the northern Adirondack margin near Alexandria Bay, New York suggest a late Devonian-early Carboniferous age for hydrothermal activity in that area (Selleck, 1989; Reynolds and Thompson, 1993).

Proterozoic rocks underlying the Potsdam Sandstone were also subject to localized hydrothermal alteration in the study area. Whitney and Davin (1987) recognized authigenic dolomite, celadonite (Fe-illite), and adularia in fracture and vein fillings in Proterozoic gneisses. Selleck (1997) presented fluid inclusion and isotopic data documenting the dolomitization of Proterozoic marbles by Paleozoic brines. These phenomenon can be linked temporally and spatially to the alteration of the basal Potsdam Sandstone above, and record hydrothermal fluid flow of regional scale.

Upper Potsdam Sandstone (=Keeseville Member): The quartz-rich sandstones of the upper Potsdam are divisible into four lithologic packages which are can be traced over the southern Lake Champlain Valley region (Figure 2). These units are:

1. Lower Stratified Unit: Medium- to fine-grained, parallel- laminated and cross-laminated subarkosic arenite. In sharp contact with arkosic facies below, often with a cobble-rich ravinement bed marking contact. Lies unconformably atop Proterozoic basement where basal arkosic facies is absent. Mudcracks and burrows present in carbonate-cemented sandstone near base. Wave-rippled surfaces common. Pyrite cements occasionally present.

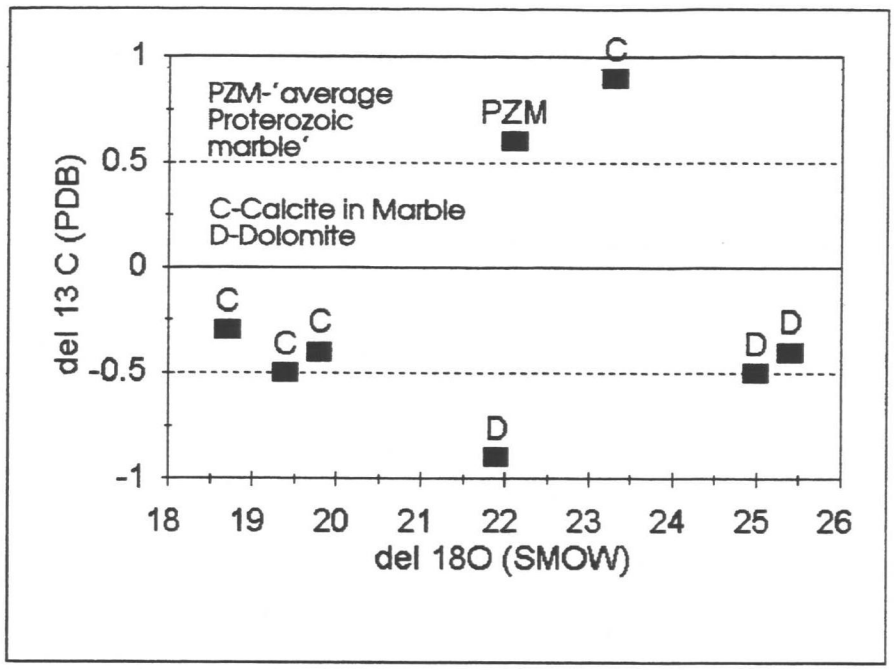
The unusual trace fossil Climactichnites has been described from quarried slabs of the Potsdam Sandstone from Port Henry, New York (Yochelson and Fedonkin, 1993). Although the no specimens have been located in situ, older quarries examined in the Port Henry area are mainly within the lower stratified unit of the upper Potsdam, and the lithologies bearing Climactichites from Port Henry most resemble the lower stratified unit. Yochelson and Fedonkin (1993) conclude that Climactichnites is apparently restricted to the Dresbachian (early Late Cambrian). Assignment of the lower stratified facies to the Dresbachian is consistent with the occurrence of Franconian trilobites in the upper stratified unit (see below).

2. Lower Cyclic Unit: Medium to coarse sandstone and fine pebble conglomerate with thin beds of dolostone. Dolostone beds often contain stromatolites. Fine, carbonate-cemented sandstones contain mudcracks. Bi-directional cross-stratification prominent in trough cross-sets. Pebbles of locally-derived pegmatitic quartz and feldspar common. Contact with underlying lower stratified unit is sharp and marked by abrupt appearance of coarse, pebbly sandstone. Lower cyclic unit unconformably overlies Proterozoic basement in outcrops immediately north of Fort Ann, New York.

3. Upper Stratified Unit: Medium-bedded, coarse to medium quartz sandstone with thin interbeds of calcareous silty shale. Horizontal traces (Rusophycos) and vertical burrows common. Horizontal lamination common; wave-ripple lamination present. Beds lenticular on outcrop scale. Contact with underlying lower cyclic unit is gradational, and marked by decrease in grain size and loss of carbonate beds. Beds within contact zone contain abundant lingulid brachiopod debris and rare, fragmental trilobite material. Intraclast conglomerates are also common in the contact zone.

The upper stratified unit is apparently the origin of the Franconian (medial Late Cambrian) trilobite fauna described by Flower (1964). The field ledges north and east of Fort Ann Village are mentioned by Flower as the source of his collections. These leges lie partially within the lower cyclic unit and include stromatolitic horizons. Phosphatic lingulid debris and Rusophycos traces can be seen on weathered slabs of the basal upper stratified unit, and I assume that these were the source of Flower's collections.

4. Upper cyclic unit: Thick- to massive-bedded medium to fine quartz sandstone with thin (2-10 cm) beds of burrowed, carbonate cemented fine sandstone recurring at 1-2 meter intervals.



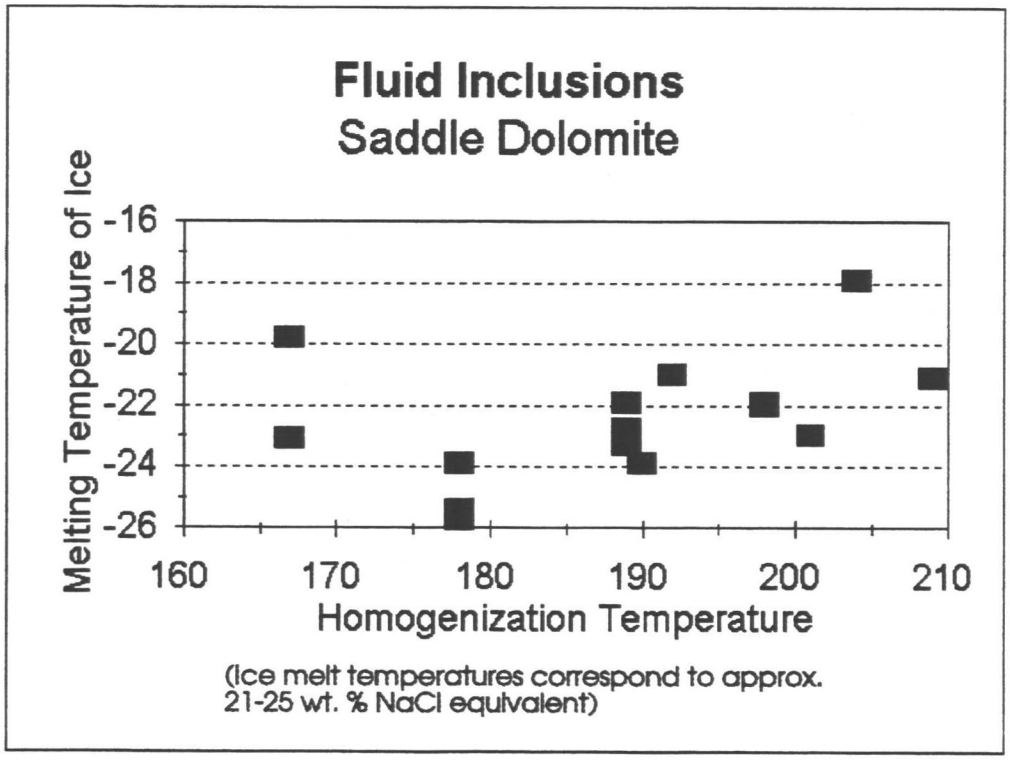


Figure 3. Isotopic and fluid inclusion data from Stop 1. Fluid inclusion data from saddle dolomite crystals in vugs within dolomitized marble. PZM is average of numerous analyses of Adirondack Lowlands marbles (data from Cartwright and Valley, 1993).

hydrothermal event(s) which caused dolomitization of the marbles seen at Stop 1. Fluid inclusion and isotopic work has not yet been carried out on this outcrop.

15.5 Continue north on Route 22.

15.9 **Stop 3.** Park on right; outcrop of left (west) side of Route 22. Please use caution crossing the road!

The uppermost portion of the lower cyclic unit and the succeeding upper stratified unit of the Potsdam Sandstone are well-displayed at this roadcut. Scattered low roadcuts allow a nearly continuous section to be measured from Stop 2 to the base of this exposure.

The lowermost beds in the main roadcut are interbedded sandy dolostones and calcareous silty shales. The dolostone beds contain fragmentary lingulid brachipods (genus Lingulepis), vertical and horizontal traces (including Rusophycos as sole markings), intraclast breccias, and rare prism cracks. This interval is correlative with the horizons sampled by Flower (1964) near the Village of Fort Ann and from which Franconian trilobites were recovered.

The interval transitional between the lower cyclic unit and upper stratified unit displays thin carbonate and shale beds, with intervening bipolar cross-stratified coarser sandstones. Medium bedded sandstones with thin shaley intercalations form the bulk of the remaining section. Note the lensoid character of the bedding within the upper portion of the outcrop. The upper stratified unit is interpreted as lower shoreface and foreshore deposits, documenting a slight deepening following deposition of the tidal flat facies of the lower cyclic unit.

Exposures in the forested areas immediately to the north and west of the roadcut are partly within the upper cyclic unit of the Potsdam Sandstone. We will not examine these on this trip. Note the prominent fault-line scarp to the west and north of the road. We will cross this fault as we proceed north to Stop 4.

15.9 Continue north on Route 22. Note that we cross a high angle fault which brings Proterozoic rock up relative to the Potsdam Sandstone to the south.

16.9 Parking area on right. Roadcuts in Proterozoic gneiss to left display upright banding and foliation with small-scale isoclinal folds.

17.4 **Stop 4.** Park on right side of road adjacent to roadcut south of crossroad.

The angular unconformity separating banded Proterozoic calcsilicate gneiss and the basal arkosic unit of the Potsdam Sandstone is exquisitely presented in these roadcuts. The gneiss below was subject to granulite facies metamorphism some 1.1 billion years ago. The Potsdam Sandstone here is most likely Late Cambrian in age, or 520 million years old. The unconformity surface thus represents one-half billion years of unrecorded geologic time.

Thin- to medium-bedded arkosic wackes and arenites with abundant quartz and feldspar pebbles form the basal arkosic unit of the Potsdam Sandstone. Coarser pockets of pebble conglomerate occur discontinuously above the unconformity, sharply overlain by regularly-bedded greenish pebbly sandstone. These conglomeratic pockets represent an earlier phase of gravel deposition. The main body of sandstone contains common normal and reverse graded beds. The poor sorting, lack of channels and rarity of cross-strata suggest that this facies was deposited by mass flow and/or sheet flood mechanisms on an alluvial fan.

Vertical burrows common. Mudcracks and calcite-filled vugs present in carbonate-cemented beds. Bipolar cross-lamination common. Contact with underlying upper stratified unit is gradational, and marked by reappearance of carbonate-cemented zones.

The uppermost beds of the upper cyclic unit of the Potsdam Sandstone contain progressively more carbonate up-section, and grade into the basal Ticonderoga Dolostone (Rodgers, 1937). This contact, marked by increase in carbonate and loss of quartz sandstone can be seen just east of the canal in the Village of Whitehall, New York, on the west-facing slopes of Skene Mountain. The contact is probably best defined as the level within the sequence where sandy dolostone beds comprise greater than 50% of the rock body. In Ticonderoga Village, basal Ticonderoga Dolostone, marked by cherty dolostone with calcite-filled vugs interbedded with rotten-weathering carbonate-cemented sandstone, can be seen on the northwest side of the village near Mt. Hope cemetery. White-weathering ledges of the quartz sandstones of the upper Potsdam occur in backyards and scattered field exposures downslope from the cemetery.

The Ticonderoga Dolostone and overlying basal Whitehall Formation are Late Cambrian in age. A Trempleauan fauna has been described from limestones within the lower Whitehall Formation (Halley, 1971).

Sedimentary Environments: The upper Potsdam Sandstone units in the study area were deposited in shallow marine shoreface, foreshore and tidal flat settings. The lower stratified unit records regional sea-level rise, with the shoreline trangressing over older terrestrial deposits of the lower Potsdam, where present, and various Proterozoic lithologies where the basal facies is absent. Shoreline stabilization and renewed input of coarse locally-derived clastics are indicated by coarse-grained, rhythmic tidal flat sequences of the lower cyclic facies. A second interval of regional sea-level rise and offshore, subtidal shelf deposition resulted in accumulation of the upper stratified facies, followed by shallowing to a second phase of tidal flat sedimentation in the upper cyclic facies.

Diagenesis of the Upper Potsdam: The intensive diagenetic/hydrothermal alteration seen in the basal arkosic unit of the lower Potsdam is less evident in the more quartzose sandstones of the upper Potsdam. However, the quartz sandstones, carbonates and thin silty shales were subject to similar thermal conditions and widespread effects of hydrothermal fluid flow are present. These include:

1. Pyrite, galena and chalcopyrite in carbonate beds.

2. Pyrite cements in the lower stratified unit.

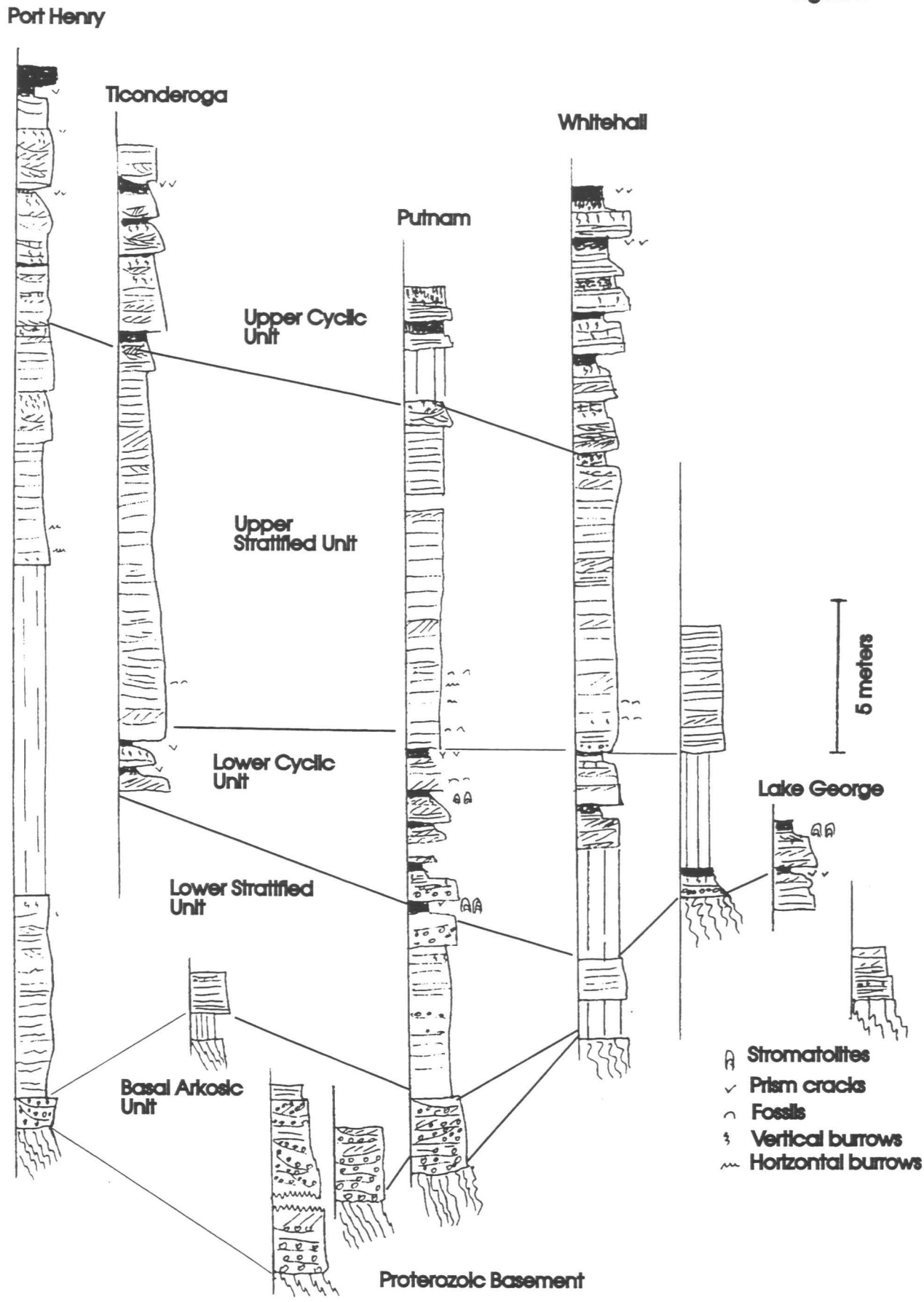
3. Authigenic REE minerals in the lower cyclic unit.

4. Well-crystallized dioctahedral illite in silty shales of the upper stratified unit (Cuprey and Selleck, 1987).

5. Halite-saturated aqueous inclusions with homogenization temperature ranging from 150 to 220 degrees C. in quartz veins within the upper Potsdam Sandstone at Ticonderoga, New York.

The origin of the warm brines responsible for the diagenetic/hydrothermal alteration of the entire Potsdam, dolomitization of the Beekmantown carbonates (Collins-Waite, 1991) and alteration of underlying Proterozoic basement is uncertain. Whitney and Davin (1987) proposed that the fluid evolved from the east to west overthrusting of the Cambrian-Ordovician shelf sequence by coeval rise prism shales and slates during the Taconic (medial-Late Ordovician) Orogeny. However,

Figure 2


Port Henry
Ticonderoga
Whitehall
Putnam
Upper Cyclic Unit
Upper Stratified Unit
Lower Cyclic Unit
Lower Stratified Unit
Basal Arkosic Unit
Proterozoic Basement
5 meters
Lake George
Stromatolites
Prism cracks
Fossils
Vertical burrows
Horizontal burrows

recent work (Geisieke, Goldstein and Selleck, 1997) has shown that vein quartz within the Taconic slates of western Vermont and eastern New York contains fluid inclusions with relatively low dissolved salt content (ranging from 1-8% NaCl equivalent), unlike the highly concentrated, NaCl-saturated brines (20-25% NaCl equivalent) recognized within the Potsdam and nearby basement (Collins-Waite, 1991; Selleck, 1997). It is possible that brines from within the Cambrian-Ordovician shelf carbonates were of high salinity, and thus could have served as the source. Similar high-salinity fluids are found throughout the foreland sequence in the Ottawa-St. Lawrence Lowlands and northern Lake Champlain Valley, and this distribution suggests a regional brine flow episode that may have occurred later in the Paleozoic. Selleck (1997) linked this alteration and associated fluid flow with the deep burial of the entire region during late Devonian time, co-incident with the Acadian Orogeny. Highly evolved and concentrated brines were available from Silurian evaporitic fluids. These fluids may have flowed through regional aquifers like the Potsdam as the foreland was tilted and uplifted while the rocks now exposed at the surface were buried beneath younger Paleozoic cover that has since been eroded. Basin-scale fluid flow linked to orogenic events on continental margins is widely recognized elsewhere in the Appalachians and other mountain belts, (e.g Kesler and van der Pluijm, 1990). Continued study of the fluid signatures and accurate dating of authigenic mineral phases may ultimately resolve the timing of fluid flow events in the southern Lake Champlain Valley.

ROAD LOG

(The road log begins at the intersection of Rt. 4 and Rt. 22 in the Village of Whitehall, New York. To arrive at this point from the Pico/Rutland area, simply take Rt. 4 west to Whitehall)

Cum. Mi.	
0.0	Intersection of Route 4 and Route 22 in Village of Whitehall, NY. Proceed north (turn right) on Route 22.
2.5	Causeway over South Bay, southern Lake Champlain. The relatively shallow lake here is quite murky with suspended sediment.
6.0	Parking area on right gives fine overview of southern Lake Champlain
9.6	Route 22 passes through a series of gravel pits in late Pleistocene terrace or glacial lake delta deposits
11.0-11.6	Series of long roadcuts in Proterozoic gneiss
13.6	**Stop 1**. Park alongside low roadcut on right side of Route 22.

This outcrop displays dolomitized Proterozoic marble. Note the very coarse crystal size of both the unaltered calcite marble and its dolomitized counterpart. Graphite in the calcite marble was apparently unaltered by the dolomitization process. The coarse dolomite crystals contain numerous small (2-5mm) voids that are sometimes partially filled with 'saddle' dolomite rhombs

and calcite spar. Note the vertical fractures in the marble that may have provided conduits for the dolomitizing fluid. Note alteration of calcsilicate minerals in dolomitized marble (best seen in outcrop on west side of Route 22). Pyritiferous, cherty, chlorite-bearing rocks are found adjacent to or within some of the dolomitized beds.

Saddle dolomite within vugs in the dolomitized marbles contains aqueous 2-phase fluid inclusions with T_h of 160-210°C (Figure 3). The relatively high salinities are similar to those seen in quartz veins that crosscut Potsdam Sandstone at Stop 4, and in diagenetic dolomite in carbonate units in the Beekmantown. Stable isotope data from marbles and immediately adjacent dolomite are shown below. Slightly heavier isotopic ratios of oxygen are consistent with dolomitization of precursor calcite at temperatures of 200°C involving fluids that are heavier than normal seawater ($\delta^{18}O$=11-18). This suggests that the fluids were either evaporitic brines or 'evolved' brines that had interacted with country rocks (Selleck, 1997). The 'average Proterozoic marble' data point is based on analyses of Adirondack Lowlands marble by Cartwright and Valley (1991).

The broad valley to the north of this outcrop is underlain by the Potsdam Sandstone (Stop 2 is visible approximately 2 miles to the north along Route 22). This suggests that these marbles lay close to the overlying unconformity with the Potsdam Sandstone, and thus were able to interact with fluids flowing through the Paleozoic sequence. Fractures enhanced downward flow from the sandstone aquifer.

Note that some of the areas of dolomite are bordered by a zone of slightly altered marble that is lightly colored by hematite. What geochemical process might have caused this?

13.6 Continue north on Route 22.

15.5 **Stop 2**. Park on right side of road next to roadcut.

This road cut exposes the upper part of the basal arkosic unit, the succeeding lower stratified unit and the lower cyclic unit of the Potsdam Sandstone. We will examine the basal arkosic unit in more detail at Stop 4. The contact with between the basal arkosic unit and the overlying lower stratified unit is well-displayed on both sides of the road. Note the abundant large clasts at the contact. This is interpreted as a ravinement surface, with the reworking of previously deposited gravels and sands by the encroaching marine shoreline leaving a lag deposit of coarse pebbles and cobbles. The immediately overlying marine sandstones contain rare mudcracks and very rare burrows (seen here only with diligent searching). Note the dominantly horizontal stratification and low-angle foreset cross-strata. This facies is interpreted as a shallow subtidal foreshore deposit. Trough cross-sets are also present. Note also the pyrite cements within the lower stratified unit.

Continuing up the sequence at this outcrop, the base of the lower cyclic unit is marked by reappearance of coarser pebbly sandstone, interpreted as a shallowing event and/or re-activation of local Proterozoic basement source areas. Well-developed bipolar trough cross-strata are readily seen on vertical faces. The coarse, pebbly sandstone intervals are regularly capped by medium sandstones and carbonate beds (mostly ferroan dolomite with late calcite veinlets and vugs). Note that these recur at 1-2 meter intervals, and are thought to represent deposition on the inner portion of tidal flats where fine carbonate mud accumulated. At other outcrops in the area, these carbonate beds contain stromatolites and mudcracks, as well as intraclast breccias. Note the apparent buckling and truncation of some of the carbonate beds. Note also the sulfide mineralization within the carbonate. Sphalerite and galena (microscopic) are associated with the pyrite. The dolomitization and sulfide mineralization here are assumed to be related to the same

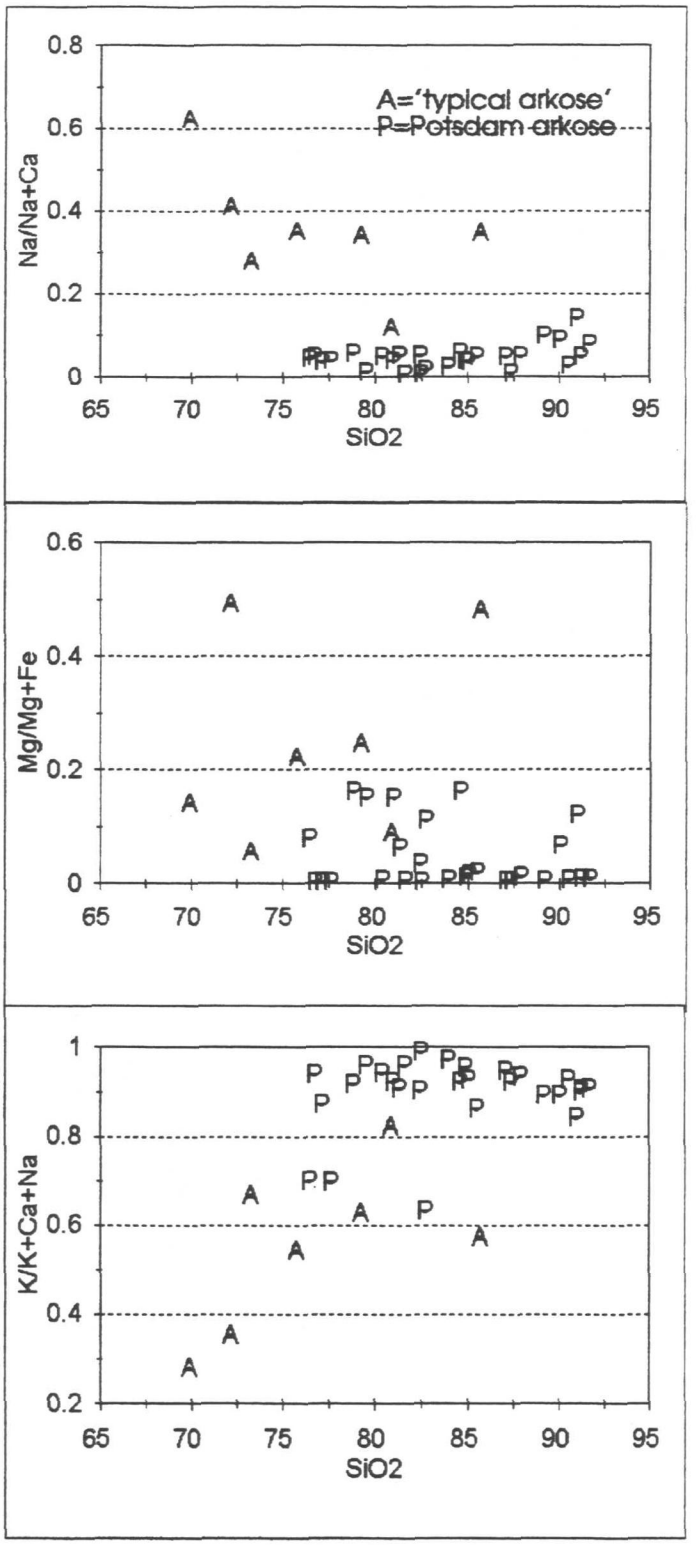
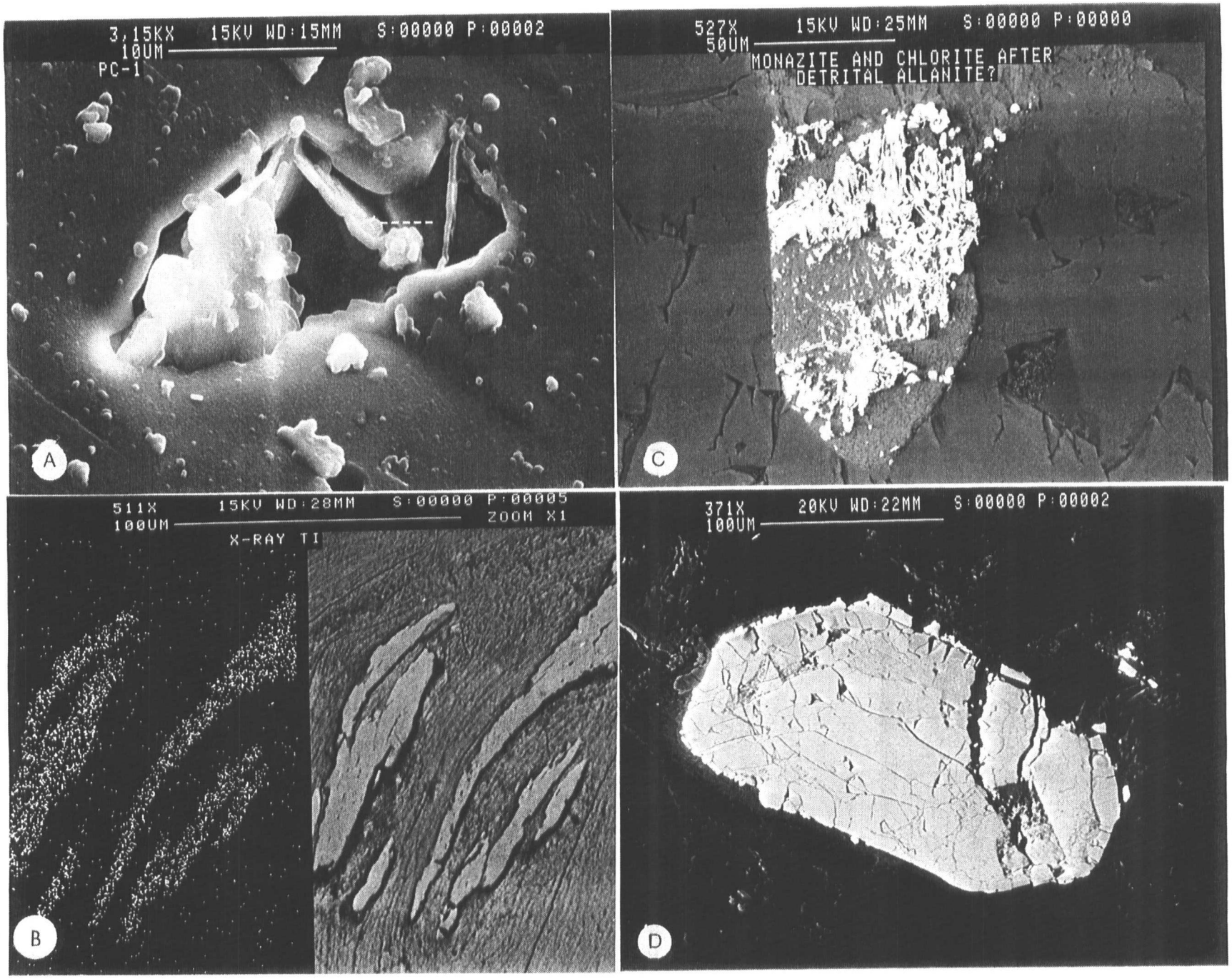


Figure 4. Variation plots of cations (ratios calculated from oxide values) vs. silica. Note relative depletion in Ca, Na, Mg and relative enrichment in Fe, K, compared to "average arkoses"(data from Pettijohn, Potter and Siever, 1973).


3.15KX 15KV WD:15MM S:00000 P:00002
10UM
PC-1
A
527X 15KV WD:25MM S:00000 P:00000
50UM
MONAZITE AND CHLORITE AFTER
DETRITAL ALLANITE?
C
511X 15KV WD:28MM S:00000 P:00005
100UM ZOOM X1
X-RAY TI
B
371X 20KV WD:22MM S:00000 P:00002
100UM
D

The grey-green color of the sandstone is due to the near-ubiquitous presence of iron-rich chlorite (to 25-29 wt. % total FeO). The chlorite formed as a diagenetic product from the alteration of pre-existing ferromagnesian minerals (hornblende and biotite), garnet, plagioclase (now altered to microcrystalline masses of chlorite and minor calcite) and probably also from detrital phyllosilicate precursors such as smectite and vermiculite. Chlorite also occurs as an intergranular cement with quartz, k-feldspar and anatase. Chlorite occurs as tabular crystals within fluid inclusions in vein

Caption for Figure 5:

A. SEM image of breached fluid inclusion in vein quartz from Stop 4. Arrow points to tabular Fe-chlorite that precipitated from fluid as quartz crystal was growing. Fluid (now lost) and solids were trapped within quartz as ccrystal continued to grow. This indicates that Fe-chlorite grew from fluids like those described by Collins-Waite (1991) for this locality (see text).

B. SEM-BSE and X-ray image of area of chlorite with anatase (Ti-phase). This area apparently represents a chlorite-replaced biotite grain.

C. Monazite and chlorite replacing detrital grain (allanite?) - SEM-BSE.

D. Authigenic xenotime as overgrowth on detrital zircon grain - SEM-BSE.

quartz at this outcrop, suggesting that the vein-mineralizing fluids were similar to those which caused the general alteration. Chemical data (Figure 4) indicate that the diagenesis of the basal arkosic facies involved relative loss of magnesium, calcium and sodium, and relative enrichment in potassium and iron, compared to typical arkoses. This suggests that the diagenetic/hydrothermal alteration involved exportation of sodium, magnesium and calcium relative to the probable composition of the original sediment. This is consistent with petrographic observations, notably conversion of plagioclase, hornblende and biotite to iron-rich chlorite, and precipitation of K-feldspar cements.

The magnesium lost from the detrital mineral suite may have provided a source of magnesium for dolomitization of overlying carbonates, and underlying marbles. Mass-balance considerations indicate that this source was only locally significant, given the large volumes of dolomite present in the sequence. However, since similar alteration is seen in some basement gneisses beneath the Potsdam, and so exportation of magnesium may have not been limited to the basal arkosic unit. It is likely that trace metals (Cu, Pb, Zn, etc) present in the original detrital sediment of the basal Potsdam were also mobilized and perhaps exported into the hydrothermal fluids. Scattered occurrences of sphalerite, galena and chalcopyrite in the Potsdam Sandstone and overlying Beekmantown Group carbonates may record the transport and deposition of these metals.

The diagenesis of the basal arkosic unit also resulted in precipitation of an unusual suite of REE-bearing minerals. The yttrium phosphate mineral xenotime is found as scattered grains within chlorite, and as overgrowths on detrital zircon (xenotime is isostructural with zircon). Monazite (Ca-REE-phosphate) is found as isolated crystals within chlorite, and as overgrowths on detrital monazite. Thorite (Th-silicate) occurs as detrital grains and minute crystals within small quartz grains. The source of the rare-earth elements appears to have been within the original sediment, perhaps as grains of allanite (REE-epidote), which is known within the granitic gneisses of the adjacent Proterozoic rocks, or from trace REE in minerals such as garnet, ilmenite or magnetite. Detrital grains of ilmenite and magnetite are often altered to intergrowths of chlorite and anatase.

Diagenetic sulfide minerals are common in the basal arkosic unit at this locality, including sphalerite, chalcopyrite, pyrite and galena. The diagenetic minerals present are all consistent with fluid alteration at temperatures of 200-250°C, with generally reducing conditions. The fluid inclusion work of Collins-Waite (1991) focussed on coarsely crystalline quartz from a near-vertical vein which can be seen in the outcrop on the east side of Route 22. Her results suggest precipitation of quartz at temperatures ranging from 200-300°C, involving halite-saturated brines.

17.4 Continue north on Route 22.

18.0 Roadcut on right (east) exposes lower cyclic unit of Potsdam Sandstone

18.8 Crossing normal fault which brings Proterozoic basement up relative to block to south

19.0 Ledges and low roadcuts on right (east) expose basal arkosic unit and underlying Proterozoic gneiss

20.9 **Stop 5.** Park on right just north of sign for Washington County-Essex County boundary.

The roadcut on the left (west) side of Route 22 is within the basal arkosic unit of the Potsdam. The outcrop is broken by a N-S fault which bounds the higher topography to the west. Good examples of slickenlines and associated features are seen in the exposure. The sandstone here is heavily mineralized with pyrite, chalcopyrite, and sphalerite (observed only in thin-section). The coarse sandstone here is more well-sorted and less chlorite-rich than seen at Stop 4.

20.9 Continue north on Route 22.

22.4 View to northeast of Fort Ticonderoga. Proterozoic gneiss along fault-line scarp to left (west).

24.0 Intersection with Route 74 to Ticonderoga Ferry. Continue north on Route 22.

24.4 Turn right (northeast) onto Shore Airport Road

24.7 **Stop 6.** Park on right before driveway.

The roadcuts on the left (northwest) side of Shore Airport Road are within the Winchell Creek Siltstone Member of the Great Meadows Formation. The stratigraphic assignment of the exposures in the area east of Ticonderoga Village has been somewhat problematic because of fault-related deformation and intense dolomitization and recrystallization. The rocks at this stop are of Early Ordovician age, based on the occurrence of poorly preserved eumophalid gastropods (Yochelson, personal communication). The view southwest from this point shows the fault bounded valley of Lake George to the southwest, and the complex of fault blocks in the immediate are of Ticonderoga Village. The ridge immediately to the southwest of Route 22 is held up by the Ticonderoga Dolostone, and a fault within the valley in the foreground offsets the Paleozoic section.

The Winchell Creek Member at this stop consists of fine dolomitic sandstone and sandy dolostone. Variations in the amount of bioturbation and the scale and style of cross-lamination are well-displayed on weathered, jointed faces. Bipolar cross-lamination and soft-sediment folding of cross-laminated strata are clearly seen. The meter-scale interbedding of cross-laminated units with bioturbated units, coupled with bipolar cross-lamination and prism-cracked silty dolostones (in the upper portion of the outcrop) suggest deposition in a tide-dominated nearshore setting with occasional subaerial exposure.

End of Trip

From this stop, return to Route 22, heading south to Whitehall-Rutland area; head north on Route 22 for Plattsburgh-Burlington area. Take Route 74 east from Ticonderoga for access to Route I87. For those heading south on I87, take Route 9N south from Ticonderoga Village to Lake George. We will make an optional stop near Lake George Village to view excellent exposures of stromatolites in the Potsdam Sandstone.

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LATE PLEISTOCENE–HOLOCENE HISTORY: HUNTINGTON RIVER AND MILLER BROOK VALLEYS, NORTHERN VERMONT

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INTRODUCTION

The areas visited on this field trip are within the Huntington River valley, draining the western slopes of the Green Mountains immediately south of the Winooski River, and the Miller Brook valley, a much smaller, east-flowing drainage basin north of the Winooski River (Fig. 1). Our objectives are to present the results of recently completed and ongoing research at the University of Vermont centered on better understanding the Late Pleistocene and Holocene history of these valleys. Parts of the material presented in this article appear in University of Vermont theses by Zehfuss (1996), Whalen (1997), Church (1997), and Bryan (1995). A comprehensive summary of regional Holocene findings appears in Bierman and others (1997) and a progress report on work in the Miller Brook valley was published by Loso and others (1997).

In the following paper we first outline the Late Pleistocene history of the Miller Brook valley inferred from our current work in the context of earlier studies of the valley. Central to this section is the reinterpretation of glacial landforms previously identified as moraines. In particular, we outline the evidence for an ice contact environment in the valley at the time most of the surficial materials were deposited. We next present a summary of the glacial lake history and later fluvial history of parts of north-central Vermont derived from recent surveying. We follow this with a review of information gleaned from numerous alluvial fans, built on both lacustrine and fluvial terraces, whose history spans the entire Holocene. We present detailed maps, logs of trenches, and ^{14}C dates as the primary data from which we base our interpretations.

The field sites described in the following text are found on the "Huntington" and "Bolton Mountain" 7.5' U.S.G.S. Quadrangle maps. Parts of these maps are reproduced as part of this field guide. Most of the field stops described herein occur on private property and permission needs to be secured before venturing onto these field sites. Contacts are listed in the Road Log.

ICE-CONTACT ENVIRONMENT: MILLER BROOK VALLEY

Stephen F. Wright

Introduction

The Miller Brook Valley, Stowe, Vermont, is a generally ESE-draining valley extending from a low point along the crest of the Green Mountains (Nebraska Notch, Elev. 576 m, 1890 ft) to its confluence with the Little River (Elev. 184 m, 605 ft), (Figs. 1 and 2). The bedrock geology of the area is described by Christman and Secor (1961) although more recent and detailed mapping has been completed and will be incorporated into the new State Geologic map. Rocks along the Miller Brook valley consist of medium grade schists belonging to both the Underhill and Hazens Notch formations. The dominant foliation in these rocks (S_2) strikes N–S and dips moderately to steeply east. Resistant rock units within these formations "V" downstream and control the orientation of tributary brooks to Miller Brook as well as the orientation of Miller Brook itself upstream of Lake Mansfield (Fig. 3). The ESE trend of the Miller Brook valley, and many other river and stream valleys in the Green Mountains, is probably controlled by similarly oriented zones of joints or brittle strike-slip faults, traces of which are readily visible on aerial photographs (see Christman and Secor, 1961, for compilations of brittle structures in the area).

Beginning at Nebraska Notch, Miller Brook flows through three steep-sided, bowl-shaped segments of the valley (labeled "1, 2, and 3" on Fig. 3). Lake Mansfield occupies the lowest of these flat-bottomed, steep sided valley segments and is artificially dammed. A large alluvial fan protrudes into the lake from the north. The stream feeding this fan is currently incised into the fan. Below Lake Mansfield in the area mapped, Miller Brook lies entirely within surficial materials, stepping down through progressively lower lacustrine and fluvial terraces until entering the Little River just upstream from the Waterbury Reservoir. Downstream from the Lake Mansfield dam, the valley contains several distinct ridges lying within or adjacent to fluvial landforms and older lacustrine terraces and standing with up to 30 m of relief. The origin of these ridges and the surrounding sediments is the subject of this paper.

Figure 1: Shaded relief map of north-central Vermont showing the location of field stops discussed in this guide. Drainage basins discussed in the text are indicated on the map. Ice flow direction was generally to the SSE, across the N–S trending ridge of the Green Mountains based on striae preserved along the Green Mountain Ridge (Stewart and MacClintock, 1970). Note that Interstate 89 is not shown traversing through the Winooski Valley on this map. Grid outlines 10 km squares. Upper box outlines Fig. 3 and lower box outlines Fig. 12.

Wagner (1970) first identified several ridges of surficial material in the Miller Brook valley and interpreted these to be moraines produced by a tongue of ice retreating up the Miller Brook Valley. He further conjectured that the Miller Brook moraine (and other moraines he identified in northern Vermont) were not produced by the waning Laurentide ice sheet, but by independent alpine glaciers that occupied the valley subsequent to ice-sheet retreat. Wagner's paper stimulated considerable discussion, mostly centered on whether or not the ridges were produced by a waning tongue of the Laurentide ice sheet or by a local glacier (Stewart, 1971; MacClintock, 1971; Wagner, 1971; Ackerly, 1989; Waitt and Davis, 1988) and raised again the question of whether or not alpine glaciers existed in New England following retreat of the Laurentide ice sheet (e.g. Goldthwaitt, 1916; see review by Waitt and Davis, 1988). Wagner's hypothesis is reevaluated below, using newly constructed maps and soil pit data. At least a portion of the area visited on this field trip was examined on an earlier NEIGC field trip led by Wagner (1972).

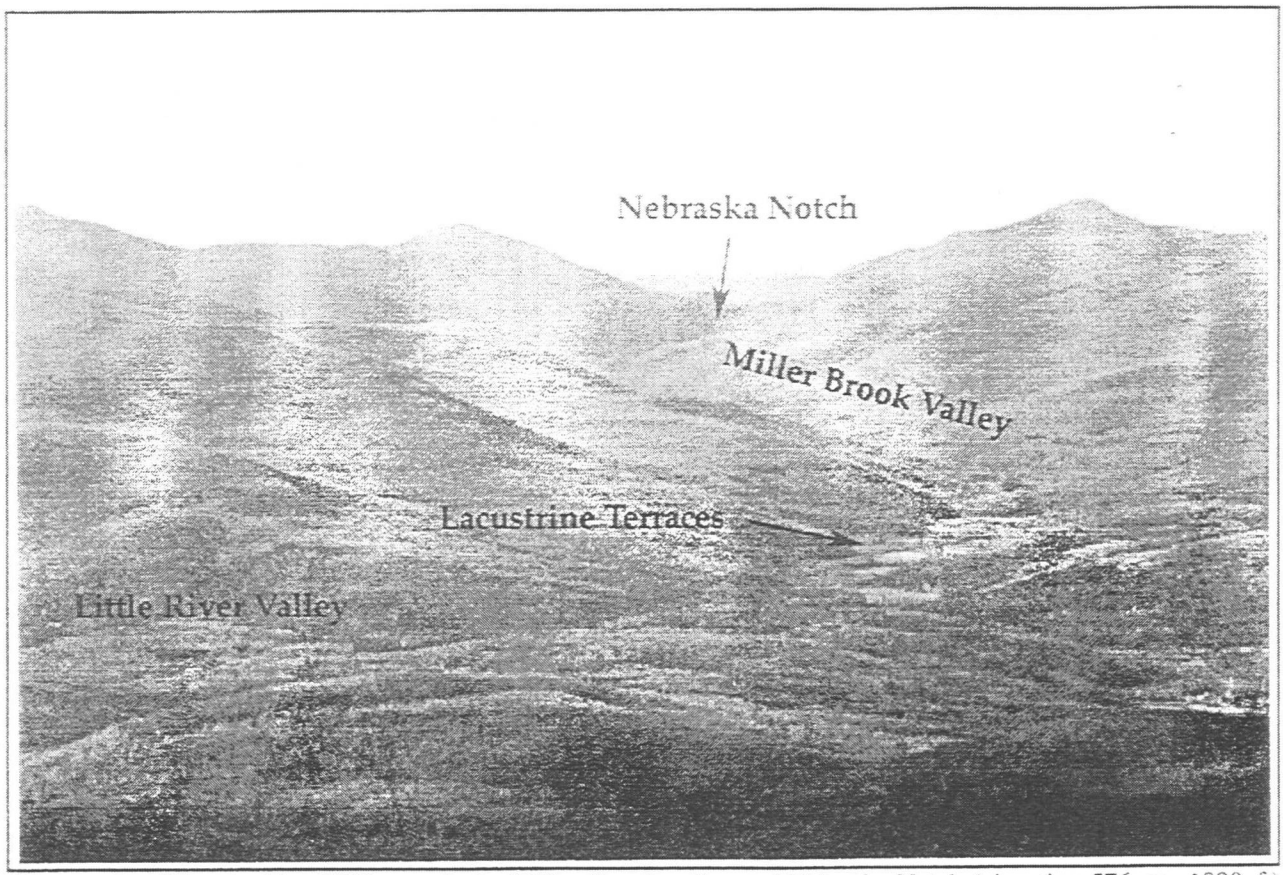


Figure 2: View looking NW towards the Miller Brook valley and Nebraska Notch (elevation 576 m, 1890 ft) from Hunger Mountain. Nebraska Notch is the lowest pass through the mountains between the Lamoille River Valley to the north and the Winooski River Valley to the south. Dewey Mountain (elev. 1024 m, 3360 ft) lies to the north of the Notch and Mount Clark to the south (elev. 902 m, 2960 ft). Small open meadows along the lower part of the Miller Brook valley are on a flight of terraces ranging from 256 to 198 m (840 to 650 ft) corresponding to Lakes Mansfield I and II and the Quaker Springs stage of Lake Vermont.

Detailed Mapping

Interpreting landforms in the Miller Brook Valley has always been confounded by the lack of adequate maps. The ridges that Wagner (1970) identified as moraines are crudely located on his map and his brief descriptions are insufficient to clearly locate their position or morphology. Only portions of the ridges appear on the U.S.G.S. topographic map (Bolton Mountain Quadrangle) and the position of Miller Brook and its principal tributary (unnamed) from the SW are incorrectly shown. During the Fall of 1995, the area immediately downstream of the Lake Mansfield dam was mapped at a scale of 1:2,500 by two students, M. Loso and H. Schwartz, using a Total Station. This map and preliminary descriptions and interpretations of soil pits constructed by P. Bierman's geomorphology class are presented in Loso and others (1997). The mapping area was continued downstream by the author in the fall of 1996, initially using tape and compass methods (1:2,000) with parts later surveyed for elevation

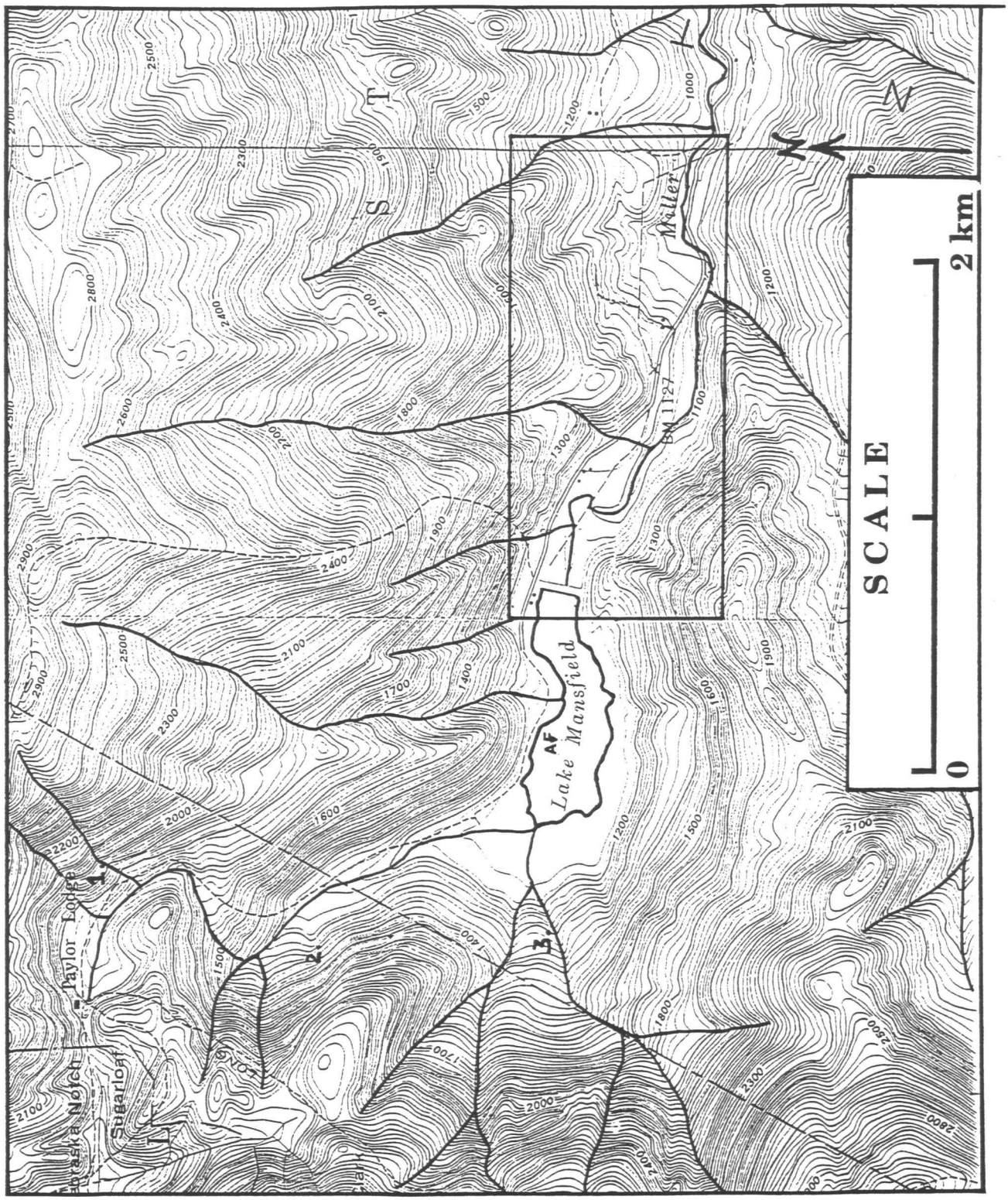


Figure 3: Portion of the Bolton Mountain 7.5' Quadrangle Map (20 ft contours) showing the Miller Brook Valley from its headwaters in Nebraska Notch . The head of the Miller Brook Valley contains three cirque-like landforms (bowl-shaped with steep sides and a gently sloped floor): (**1**) the valley immediately east of Taylor Lodge, (**2**) the NNW-trending portion of the valley NNW of Lake Mansfield, and (**3**) the portion of the valley occupied by Lake Mansfield. Box outlines detailed map of Surficial Landforms (Fig. 4). **AF** = Alluvial Fan

control using the Total Station. The map presented in this field guide is a compilation of these two maps set within an enlarged portion of the U.S.G.S. base map (Fig. 4). The new mapping uses 2 meter contours tied to a benchmark at 1127 ft, whereas the surrounding U.S.G.S. topographic map uses 20 ft contours.

Description of the Landforms

Ridges of surficial material in the Miller Brook valley extend from ~150 m west of the Lake Mansfield dam (not mapped on Fig. 4), across the entire mapped area (Fig. 4), and reconnaissance work indicates that they extend at least another kilometer down-valley. For clarity, individual ridges have been designated with letters, e.g. Ridge C, on Figure 4. The ridges generally follow sinuous pathways, both along the side and across the middle of the valley and are cut by both ephemeral streams and Miller Brook. In places they appear to bifurcate or to show cross-cutting relationships with one another (e.g. Ridges C and D, and Ridges F and G, Fig. 4). The ridges are both sharp-crested (Ridges A, B, the eastern end of C, E, F, and G) and rounded (most of Ridge C and Ridge D), applying terminology employed by Shreve (1985) and show relief ranging from 4 to in excess of 25 m above adjacent fluvial or lacustrine terraces. In general, the crest of Ridges A, B, and C gradually diminishes from 352 m just below the Lake Mansfield dam to 342 m NE of the Mill Pond dam. An exception to this is the abrupt 14 m rise (to 362 m) and fall along a 100 m reach SW of the Middle Pond (Fig. 4).

That the ridges are primary constructional landforms and not erosional remnants is evidenced by several observations: (1) Ridge A surrounds a small bog ("B" in Fig. 4, adjacent to Soil Pit T2H). Sperling and others (1989) retrieved a core from this bog and present a ^{14}C date of 9,280 ± 235 yBP (years before present) from sediment recovered between 2.75 and 2.85 m depth, implying that the bog has been separated from the valley bottom by the intervening ridge for over 9,000 ^{14}C years, within 3,000 ^{14}C years since ice retreat (Lin Li, 1996). (2) Soil pits adjacent to Ridges A (T3R), C, and D clearly show fine to medium grained sand onlapping the coarse sand, gravel, and rounded cobbles comprising the adjacent ridges, indicating that the ridges were partially buried by fluvial and lacustrine sediments after their formation (Fig. 5). (3) A small alluvial fan, fed by an ephemeral stream 100 m NW of the previously described bog, has partially buried Ridge A (Fig. 4). By inference to the active Holocene history of fans elsewhere in the region (see later sections of this paper) we suggest that this fan too has been actively depositing material against the ridge throughout the Holocene.

Ridges A, B, and C

The "classic" Lake Mansfield ridges, the ones Wagner (1970) describes as forming during a later phase of glaciation (his Phase II), are labeled Ridges A and B on Figure 4. Ridge A begins just beyond a broad area of hummocky relief including one well-defined, kettle-like closed depression where the trails intersect, south of Soil Pit RTP. This ridge, interpreted by Wagner (1970) to be a lateral moraine, parallels the south side of the valley. No correlative ridge exists along the north side of the valley. Approximately 600 m below the Lake Mansfield dam the ridge turns abruptly NE (Ridge B, Fig. 4) and crosses the valley where it is cut by Miller Brook. This is the segment interpreted to be an end moraine by Wagner (1970). Based on the map pattern shown in Figure 4, Ridge B does not curve up-valley to meet an unidentified lateral moraine along the north side of the valley, but instead strikes straight across the valley, turns slightly and continues down the middle of the valley (Ridge C; the road is built upon it) until it is cut again by Miller Brook, just east of the Mill Pond dam.

Ten soil pits were dug either along or adjacent to Ridges A and B (e.g. T2M, Fig. 4) as part of the study presented by Loso and others (1997). Several of the pits considered most important to this study were reopened and are reinterpreted here together with observations from new pits (e.g. C-2, Fig. 4). With few exceptions, most of the pits were excavated to approximately 1 m depth.

The near-surface materials comprising Ridges A, B, and C are quite similar to one another and texturally bear attributes of both till and fluvial sediments. Soil Pit C-2, along the crest of Ridge C (Fig. 4), exposes 0.5 m of poorly sorted, poorly to well rounded coarse sand, gravel, and cobble clasts in a fine sand matrix. These materials overlie clean gravel and coarse sand extending to the bottom of the pit (0.7 m). Soil Pit T1M, along the crest of Ridge B (Fig. 4), similarly contains a poorly sorted mixture of moderately to well rounded coarse sand, pebbles, and cobbles in a medium to fine sand matrix. A recent soil slip along the NE side of Ridge B (~5 m downstream from Soil Pit T1R,) reveals clean, moderately sorted and rounded, coarse sand and gravel just above water level that is overlain by at least 3 m of material similar to that exposed in Soil Pit T1M along the crest of the ridge. The gravels occurring in both Ridges B and C are clearly fluvial and the overlying material in both ridges contains abundant rounded clasts, also indicative of fluvial processes. Based on both their map pattern and the fluvial sediments within them, I interpret these ridges to be segments of an esker and not moraines. The poorly sorted veneer of sediment overlying the gravels has likely originated from the low discharge of the stream occupying the esker tunnel towards

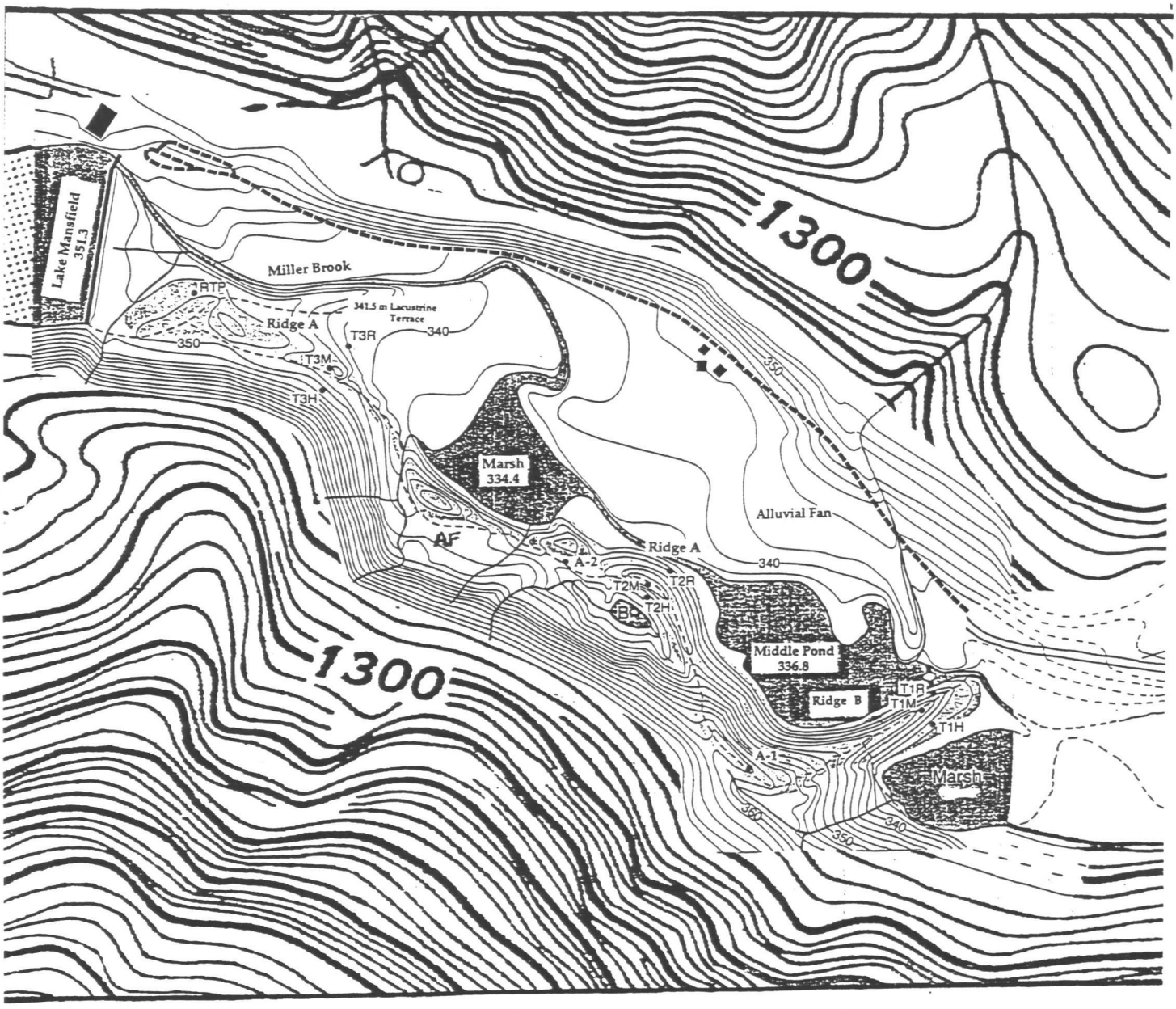


Figure 4: Detailed map of surficial landforms in the upper portion of the Miller Brook Valley (see Fig. 3 for exact location). Western half of map is by M. Loso and H. Schwartz (Loso et al., 1997) and uses 2 m contours. Eastern half of map is by S. Wright and that part surveyed also uses 2 m contours, dashed where inferred. Both maps are placed within an enlarged portion of the Bolton Mountain 7.5' Quadrangle map. Ridges are labeled with letters, e.g. Ridge A, and extend across the entire mapped area. AF = Alluvial Fan.

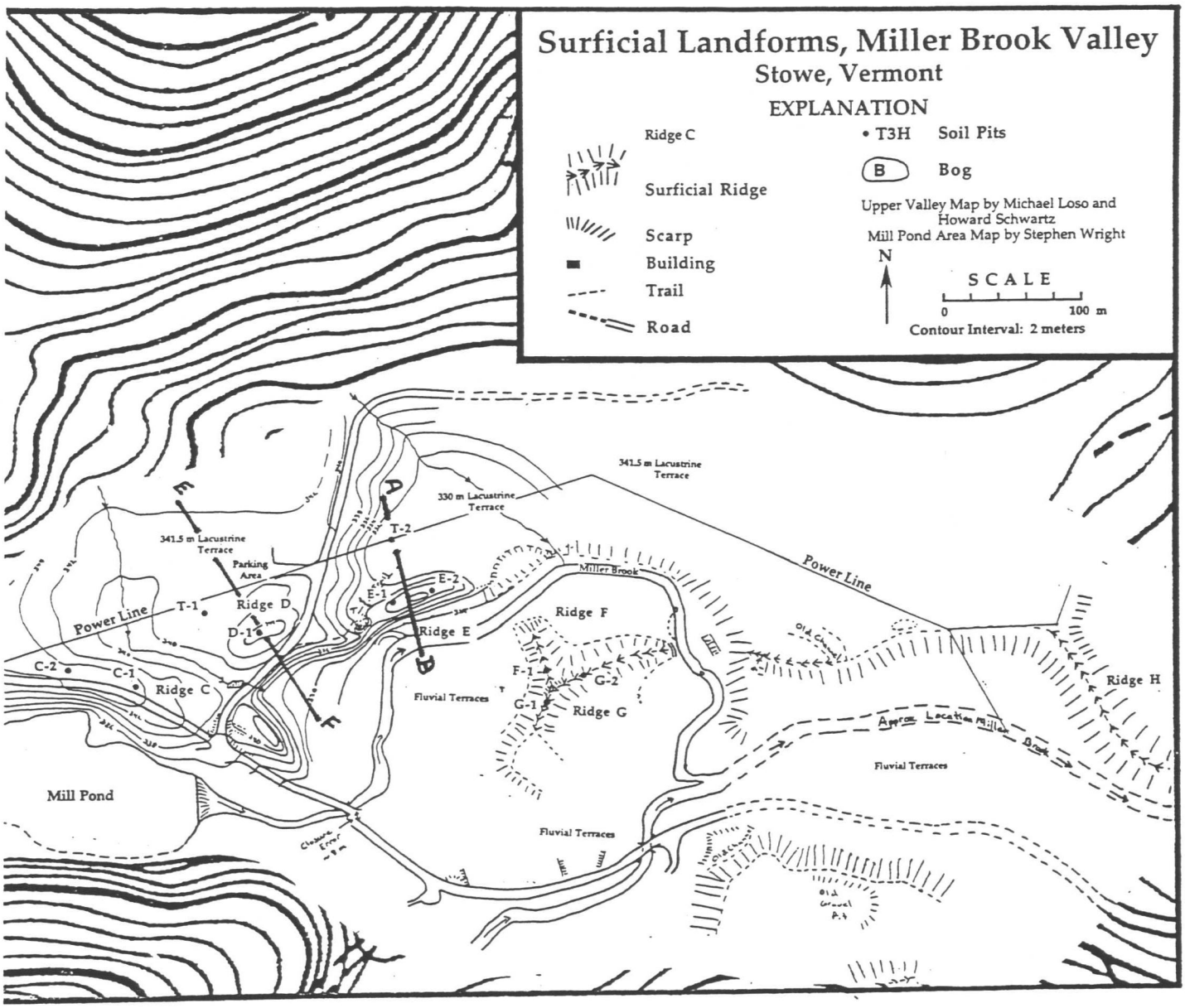


Figure 4: Continued from previous page.

the end of its history. The rounding of most clasts indicates at least some transport of the melt-out till. Soil pit C-1 on Ridge C (Fig. 4) contains coarse diamict including a few faceted, striated cobbles. It seems likely that this diamict melted out of the ice shortly before the ice tunnel was abandoned and the clasts never subjected to significant abrasion.

The transition from Ridge B to Ridge A is marked by an abrupt change in orientation from N60E to N45W, rapid ascent to the highest elevation of any of the mapped ridges (362 m, Fig. 4), and the presence of numerous large, mostly angular boulders, many exceeding 2 m in diameter. Large boulders are also common along segments of Ridge A opposite the "Marsh" (at 334.4 m elev.) and in the vicinity of Soil Pit T3M, but much less common along the segment opposite the Bog ("B" on Fig. 4). Boulders of similar size and frequency are rare along the other ridges shown on Figure 4, all of which occupy positions considerably removed from the sides of the valley. Soil Pits along the crest of Ridge A (A-1, T2M, and T3M) all display unsorted, unbedded coarse sand, gravel, cobbles, and boulders in a medium to fine sand matrix with minor silt (see grain size distribution curves in Loso et al., 1997). Pebbles and cobbles excavated in A-1 and reexcavated in T3M are nearly all moderately rounded (T2M was not reexcavated). Soil textures in these pits are similar to the soil pit (T1M) on the crest of Ridge B, except for the large, usually angular boulders mentioned earlier. These materials differ from the colluvium along the adjacent hillside in that most of the clasts larger than 1 mm are rounded as opposed to angular or subangular. In the top meter of Ridge A none of the pits revealed layers of clean, coarse sand and gravel similar to those observed in Ridges B and C and it is therefore more difficult to claim that fluvial processes are responsible for the deposition of these materials. Nevertheless, the abundant rounded clasts, paucity of silt and clay, and textural similarity to the surficial materials overlying fluvial gravels in Ridges B and C are at least consistent with their deposition in a subglacial stream flowing along the side of the valley. Given the mapped continuity of Ridge A with Ridges B and C, I similarly interpret Ridge A to be an up-valley segment of the same esker. The abundant large boulders in and on it were most likely derived from materials sliding down the adjacent hillside onto the ice surface and later being incorporated into the esker as the tunnel ceiling was melted by the subglacial stream.

Ridges D, E, F, G, and H

Similar to Ridge C, described in the previous section, Ridges D, E, F, G, and H all lie in the central portion of the Miller Brook valley, well away from the valley sides (Fig. 4). Ridge D is a rounded, elongate hill rising in low relief relative (maximum elevation 345.3 m) to the surrounding terrace at 340.5 m elevation. It is oriented approximately N60E and may be continuous with Ridge E. Ridge E is sharp-crested, rises to 338.4 m and shows sharp relief with the 330 m terrace to the north and the <324 m fluvial terraces to the south. Soil pits in these ridges (D-1, E-1, E-2; Fig. 4) all contain clean, well rounded gravel and cobbles in a coarse sand matrix. Bedding was observed in Pit E-1. The sediments in these ridges are clearly fluvial in origin and I therefore interpret Ridges D and E to be segments of an esker. Ridge C appears to cross-cut Ridge D, implying that the former is younger, but it is also possible that the esker bifurcated at this point (Fig. 4).

Ridges F and G are the most striking in the valley forming sharp crested ridges rising at least 25 m above the fluvial terraces that surround them (Fig. 4). Ridge F may be the continuation of Ridge E as it arches across Miller Brook. Ridge F is clearly cross-cut by Ridge G, which rises approximately 5 m above Ridge F at their intersection. Ridge G strikes N15E at its southern end, but swings almost due east where it is cut by Miller Brook. Ridge H on the opposite side of Miller Brook is the continuation of Ridge G, rising to the 341.5 terrace level, where it is completely buried, and then reappears at the eastern edge of the map where it has been excavated by stream erosion (Fig. 4). Soil pits in these ridges (F-1, G-1, G-2) all contain unsorted, moderately rounded cobbles and pebbles in a fine to coarse sand and gravel matrix. Large (up to 1.5 m), relatively unrounded boulders are also incorporated into the ridges. The texture of surficial materials in these pits is very similar to that in the near-surface parts of Ridges B and C (see earlier descriptions). None of the soil pits exceeded 1 m depth and none exposed any clean gravel similar to that observed at the surface in Ridges D and E. Based on the abundance of rounded clasts in these ridges and their sinuous, mid-valley form, I also interpret these ridges to be eskers. Furthermore, it is likely that the easternmost end of Ridge C once connected with the southern end of Ridge G and the intervening esker has been removed by Miller Brook. Ridges D, E, and F may document a former esker tunnel that was abandoned in favor of the tunnel occupied by the eastern end of Ridge C and the southern end of Ridge G.

Lacustrine Terraces

Two distinct terraces exist in the map area, the upper one at ~341.5 m (~1,120 ft) and the lower one at ~330 m (~1,082 ft). Elevations are taken from the middle of these terraces which show ±0.4 m of minor relief. Materials exposed in soil Pits T3R, A-2, T-1, and T-2, excavated in these terraces, are described here as well as materials exposed in recent road-cuts in the vicinity of the Mill Pond dam (Fig. 4). Two meters of well sorted and bedded fine

sand and silt with two thin beds (2–3 cm) of coarse sand onlap the coarse sediments of Ridge A in Soil Pit T3R, located adjacent to Ridge A, on the 341.5 m terrace (0.4 m of colluvium overlie the terrace deposits, Loso et al., 1997). One of the coarse sand layers contains load casts down into the underlying silts. Bedding strikes to 331 and dips 29° NE, parallel to the slope of the buried ridge. Some of the bedding has slumped to the NE, down the slope of the buried ridge. Loso and others (1997) interpret these sediments to be lacustrine, deposited in a shallow area of a lake whose minimum elevation was 342 m. Soil Pit A-2, dug along Ridge A, also reveals medium to fine sand with isolated, matrix supported, moderately rounded pebbles and cobbles. This pit is at the same elevation as Pit T3R and these sediments are probably also lacustrine in origin and onlap Ridge A.

Two topographic profiles (A–B and E–F, Fig. 4) show the terraces adjacent to Ridges D and E and the materials exposed in nearby soil pits and auger holes (Fig. 5). Soil Pit T-1, located on the 341.5 m terrace 40 m west of Ridge D (Fig. 4), was extended to 3.7 m depth using a bucket auger. The upper 1.4 m consists of clean, moderately sorted and well rounded, medium to coarse sand and gravel. The remainder of the auger hole (1.4 to 3.7 m) exposed only fine sand with rare layers of medium to coarse sand and one isolated rounded pebble (Fig. 5). I interpret the fine sands to be lacustrine and the overlying coarse sand and gravel as being fluvial. If this interpretation is correct, the lake elevation in this area is marked by the transition from fine to coarse sand, 1.4 m below the terrace level, ~340.1 m (1,116 ft). Exposures in the vicinity of the road intersection opposite the Mill Pond dam show that the lacustrine sediments are extensively deformed where they are cut by faults and are slumped indicating that the lacustrine and later fluvial sediments in this area were deposited on top of pockets of dead ice. This implies that the lake formed synchronously with the final melting of glacial ice from the valley.

The lower 330 m terrace also appears to record both a lacustrine and fluvial activity. Soil Pit T-2 (along cross-section A–B) was excavated to 1.60 m where it bottomed in coarse rounded gravel and sand. These are overlain by 0.85 m of fine sand and silt which in turn is overlain by 0.75 m of rounded gravel and coarse sand (Figs. 4 and 5). The lake elevation here is interpreted to be at the contact between the fine sand/silt and gravel, or at ~329.2 m (1,080 ft).

The two terraces, corresponding to lake elevations of 340.1 and 329.2 m, may correlate with Lake Winooski, the outlet of which lay 4.0 km south of Williamstown, Vermont at an elevation of 279 m (Larsen, 1987b; see also review of regional glacial lakes later in this field guide). Projecting this threshold elevation N21.5W to the Miller Brook valley with an isostatic gradient of 0.9 m/km gives an elevation of 322.2 m, implying a water depth at the outlet of 7 m for the lower lake level and almost 18 m for the higher lake level, unless the outlet eroded from its current elevation.

Discussion

The direction of ice flow in this part of Vermont was generally from NNW to SSE, across the N–S trending ridge of the Green Mountains (Stewart and MacClintock, 1970; Christman and Secor, 1961; Larsen, 1987a; Ackerly and Larsen, 1987) as evidenced by grooves and striations preserved along the crest of the Green Mountains. Once the elevation of the ice surface dropped below the N–S ridge of the Green Mountains, continued SSE ice flow was funneled through, from north to south, the Lamoille River Valley (elev. 150 m, 490 ft), Smuggler's Notch (elev. 670 m, 2200 ft), Nebraska Notch (elev. 576 m, 1890 ft) and the Winooski River Valley (elev. 103 m, 340 ft), (Fig. 1). Cores taken from Sterling Pond (1.7 km east of Smuggler's Notch, elev. 915 m, 3,000 ft) indicate that the ice was below this elevation at least by 12,700 14 C years ago (Lin Li, 1996). Nebraska Notch is the lowest gap through the Green Mountains between the Lamoille River valley to the north and the Winooski River valley to the south. Therefore, the Miller Brook valley must have contained Laurentide ice well after ice had melted from the eastern flanks of the Green Mountains immediately to the north and south. Once the ice elevation dropped below the elevation of Nebraska Notch, ice supply to the Miller Brook valley was shut off and the ice in the valley stagnated soon thereafter. The esker documented in this report probably formed immediately before and during this stagnant ice stage. While not mapped in detail, reconnaissance work down-valley shows the esker continuing at least 2.6 km down-valley from the Lake Mansfield dam, the crest dropping in elevation from 352 m just below the Lake Mansfield dam to 299 m at its most eastern end. Ice-contact deformation of the lacustrine sediments underlying the 341.5 m terrace imply that that lake occupied the Miller Brook Valley during the final stages of stagnant ice melting. It is unclear whether or not this lake was ever continuous with Lake Winooski, although the lake producing the lower 330 m terrace is low enough to make its correlation with Lake Winooski likely.

Wagner's (1970) interpretation that the ridges described here are moraines stemmed from the following observations: (1) The head of Miller Brook occurs in a bowl-shaped valley, steep-sided and flat-bottomed, a shape typical of glacial cirques; (2) The ridge extending down-stream and along the south side of the valley from the Lake

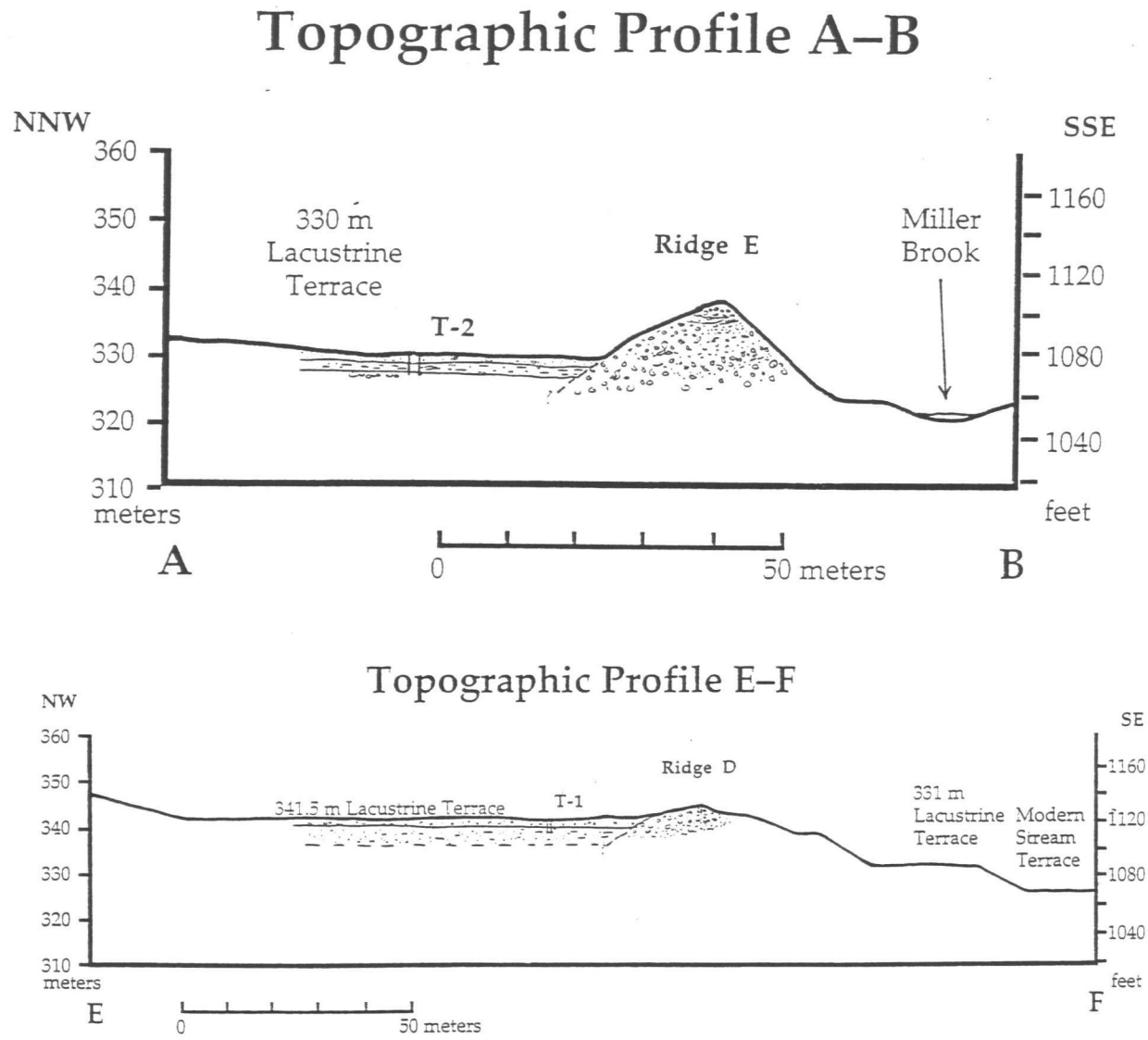


Figure 5: Unexaggerated topographic profiles across terraces and Ridges D and E northeast of the Mill Pond (Fig. 4). **Profile A–B** extends from the 330 m terrace, across **Ridge E**, to Miller Brook. Ridge E is composed of well sorted and bedded coarse sand and gravel. **Profile E–F** extends from the 341.5 m terrace across **Ridge D** to a terrace at 331 m and then down to alluvial terraces adjacent to Miller Brook. Ridge C consists of rounded and bedded gravel and cobbles, similar to Ridge E. Materials exposed in Ridges D and E are interpreted here to be of fluvial origin deposited in an esker tunnel. Soil pits near the border of Ridge D and the 341.5 m terrace show the terrace sands and gravels onlapping the much coarser cobble gravels contained in the Ridge. Soil Pit T-1 in the 341.5 m terrace exposes 1.4 m of medium to coarse sand and gravel that overlie at least 2.2 m of fine sand. Soil pit T-2 in the 330 m terrace exposes 75 cm of coarse sand and gravel overlying 85 cm of grey silt and fine sand, which in turn overlies coarse gravel. The upper sand and gravel in both terraces is interpreted to be fluvial in origin, whereas the underlying fine sand and silt lacustrine deposited in lakes at elevations of 340.1 and 329.2 m respectively.

Mansfield dam turns and crosses the valley in a manner similar to lateral and end moraines produced by alpine glaciers; and (3) A sloping terrace, extending from the end moraine down-valley to a delta was interpreted to be an outwash plain graded from the end moraine down to a pro-glacial lake. I concur with Wagner (1970) that the head of the Miller Brook valley is cirque-like, especially its uppermost reach (Fig. 3), despite arguments by Waitt (Waitt and Davis, 1988) to the contrary. While Loso and others (1997) have determined that the floor of the Miller Brook valley is too low to maintain a valley glacier after the retreat of the Laurentide ice sheet. Given our current understanding of climatic cycling during the Quaternary where gradual cooling over has been followed by rapid warming (e.g. Dansgaard et al., 1993), it is likely that the Miller Brook cirque, and others in New England, formed during the onset of these different glacial episodes, before being overridden by continental ice. Waitt and Davis (1988) have pointed out that alpine glacial landforms can be preserved, despite subsequent cover by continental ice. In this light, the bedrock landform, the cirque at the valley head, formed before the onset of continental glaciation, whereas the ridges of surficial material were formed as the ice sheet thinned and stagnated in the Miller Brook valley.

Conclusions

(1) Detailed mapping reveals that the ridges of surficial material occurring in the Miller Brook valley can be interpreted as a connected system of ridges lying both along the valley edge and valley center.

(2) Ridges B, C, D, and E all contain clean sand and gravel interpreted here to be of fluvial origin. Materials exposed in shallow holes in Ridges A, F, and G are poorly sorted, yet are dominated by relatively coarse rounded clasts. Based on both the map pattern and the fluvial materials exposed within them, I interpret all of these ridges to be eskers.

(3) The map pattern indicates that Ridges D, E, and F were once continuous and are cut by, and therefore younger than, Ridges C and G. It is also possible that the esker tunnel bifurcated and rejoined and that all of the ridges formed synchronously.

(4) Soil pits in the terrace sands commonly expose highly disrupted bedding indicative of collapse following ice melt-out. Closed depressions within and adjacent to Ridges A and B are also likely to be ice melt-out features—kettles. The eskers, kettles, and collapse structures all suggest sediment deposition in contact with glacial ice, most likely stagnant ice.

(5) Onlapping contact relationships of lake sediments over the materials in the Ridges indicate that a lake occupied the valley soon after ice retreat forming terraces at two different elevations. Lake elevations are interpreted to be at 340.1 and 329.2 m, based on the contact between lacustrine and fluvial sediments and may correlate with Lake Winooski.

(6) If Ridges A and B are indeed moraines (Wagner, 1970), they must post-date the eskers, because the eskers are at elevations very close to that of Ridges A and B, necessitating ice, probably stagnant ice, of at least that thickness. If they post-date the melt out of the Laurentide ice sheet, they must have been produced by either (1) a considerable thickening of the Laurentide ice on the West side of the Green Mountains, allowing an ice tongue to extend through the Nebraska Notch, or (2) the establishment of a small cirque glacier. Both require a considerable cooling of the regional climate which is not supported by paleotemperature estimates during the early Holocene (Loso et al., 1997).

(7) If the "cirque-like basins" at the head of the Miller Brook valley are indeed cirques, they most likely formed during the long periods of cooling prior to the advance of the Laurentide and earlier ice sheets.

PROGLACIAL LAKE AND RIVER HISTORY IN THE WINOOSKI DRAINAGE BASIN

Timothy N. Whalen

Introduction

When the Hudson-Champlain lobe of the Laurentide ice sheet retreated from this mountainous region, proglacial lakes developed between the ice and the valley walls, leaving behind a scattered record of deltas and shorelines. As the lakes withdrew from the valleys, rivers began to incise the lake deposits. The subsequent fluvial activity in the Winooski Drainage Basin removed much of the glacial deposits and lowered the valley floors more than 40 m in the lower reaches. The purposes of this paper are 1) to provide an overview of the history of proglacial lakes occupying the Winooski River Valley and its tributaries, with special attention given to the Huntington Valley, 2) to identify when the transition occurred between lacustrine and fluvial processes in each of these valleys, and 3) to describe the mechanisms responsible for river incision in the Winooski Drainage Basin.

Deglacial History

The deglacial record of the Winooski Drainage Basin is one of northwestward ice retreat, ice lowering, and decreasing proglacial lake levels. Larsen (1972; 1987a & b) has previously described the sequence of proglacial lakes in the Winooski Drainage Basin and his five stage model (Table 1) is used as the basis for this discussion. Larsen specifically addressed the chronologies in the Mad, Dog, and Stevens Branch Valleys, but work in the Huntington River Valley (Wagner, 1972; Bryan, 1995) and in the Little River Valley (Merwin, 1908; Connally, 1972) also demonstrates that the history of proglacial lakes in these valleys can be placed within the context of Larsen's model. The following section focuses on the proglacial lake levels in the Huntington Valley and the reader is encouraged to review Larsen (1972; 1987a & b) for specifics concerning the other valleys in the Winooski Basin and Chapman (1937) for the Champlain Valley lake levels.

Table1: Winooski Drainage Basin Proglacial Lake Levels and Ice Positions

Stage	Winooski River	Huntington River	Little River	Mad River	Dog River	Stevens Branch
I	Ice Retreats to Valley Mouths	Lake Jerusalem to Southern Divide	Ice and Local Lakes	Lake Granville to Granville Gulf Divide	Roxbury Divide	Lake Williamstown to Williamstown Divide
II	Ice Retreats to Jonesville	Lake Huntington to Hollow Brook Divide	Lake Winooski above Divide	Lake Winooski to Warren	Lake Winooski to South Northfield	Lake Winooski to Divide
III	Ice Retreats to Richmond	Lake Huntington to Hollow Brook Divide	Lake Mansfield I to Divide	Lake Mansfield I to Moretown	Fluvial	Fluvial
IV	Ice Retreats to Green Mountain Front	Lake Mansfield II to Hollow Brook Divide	Lake Mansfield II to Stowe	Fluvial	Fluvial	Fluvial
V	Lake Coveville to Waterbury	Lake Coveville to Huntington	Lake Coveville to Moscow	Fluvial	Fluvial	Fluvial

Note: The lake positions are cited as the farthest upvalley location for stage and ice postions are cited as position at end of stage.

Stage I lakes were limited to single north-draining valleys because they were bordered to the north by the retreating ice. These lakes drained over southern spillways, located at different elevations, and into the Champlain (Huntington Valley) and Connecticut (Mad, Dog, and Stevens Branch Valleys) Basins. The Stage I spillway in the Huntington Valley is located at the southern divide at an elevation of 460 m (1510 ft), controlling a previously unnamed lake Stage I lake (Fig. 6). Following Larsen's (1972) criteria for naming Stage I lakes based on the nearest village, the name Lake Jerusalem is suggested. Evidence for Lake Jerusalem includes clay deposits found at 240 m (787 ft) south of Huntington Center (Bryan, 1995). During Stage II, the lakes east of the Huntington Valley

coalesced, from southeast to northwest in a time-transgressive manner following the retreating ice, into Lake Winooski and utilized a 279 m (915 ft) spillway south of Williamstown, the lowest Stage I outlet. In the Huntington Valley during Stage II, the ice retreated past the Hollow Brook Valley, allowing Lake Jerusalem to empty into the Champlain Valley and to lower to the Lake Huntington level. Lake Huntington drained across a threshold presently located at 204 m (670 ft) in the Hollow Brook Valley, and its waters entered Lake Coveville in the Champlain Valley at the South Hinesburg delta.

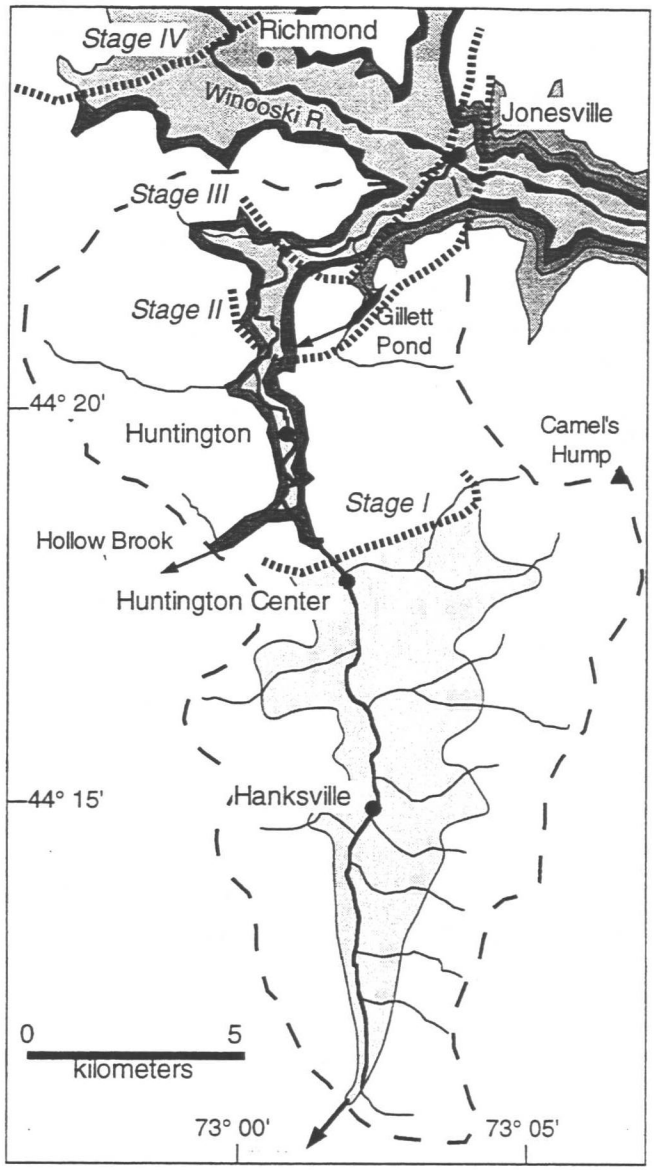


Figure 6: Map of proglacial lake extent and ice positions in the Huntington Valley. The lake levels are shown with the corresponding ice position at the end of each stage.

At the close of Stage II, the ice in the Winooski Valley had retreated to Jonesville, and Lake Huntington had expanded to fill the valley between the Hollow Brook spillway to the southwest and Gillett Pond to the northeast (Fig. 6). However, a tongue of ice remained in the lower Huntington Valley, and when the Gillett Pond spillway (229 m; 750 ft) was uncovered, Lake Winooski drained into the Huntington Valley. During Stage III, the Gillett Pond threshold controlled the level of Lake Mansfield I, and the Hollow Brook threshold continued to control the water level of Lake Huntington (Fig. 7). Stage IV commenced when the mouth of the Huntington Valley became ice free and the waters of Lake Mansfield I were able to drop to the level the Hollow Brook spillway, forming Lake Mansfield II. Once the ice retreated from the Green Mountain Front (Stage V), Lake Mansfield II lowered, Lake Coveville entered the Winooski Valley, and the present configuration of the Winooski River drainage was established.

Fluvial History

The river history in the valleys of the Winooski Basin began when the last proglacial lake shoaled and therefore commenced at different times in each valley. The upper reaches of the Mad River (south of Warren), for example, were clear of lake water during Stage I while the lowest reaches of the Little River remained submerged until the end of Stage V. However, a general valley evolution can be illustrated as in Figure 8, where the initial landscape over which the river flowed was most likely a glaciolacustrine fill terrace. This deposit is generally a coarsening upward sequence of rythmically bedded silts and clays grading into sands (some of which may be deltaic in origin and therefore exhibit foresets) and capped by fluvial deposits. These glaciolucustrine fill terraces have often been mapped as deltas on the basis of their morphology (broad, flat terraces), with little supporting sedimentological evidence (i.e., Wagner, 1972). Precise surveying demonstrates that the elevation of these surfaces decreases downstream in the Huntington River Valley (Whalen, 1997) whereas it should increase (due to glacio-isostatic rebound) if they were all built to a stable lake level. In fact the glaciolacustrine fill terraces are a composite of many depositional environments and record the transition from lacustrine to fluvial deposition.

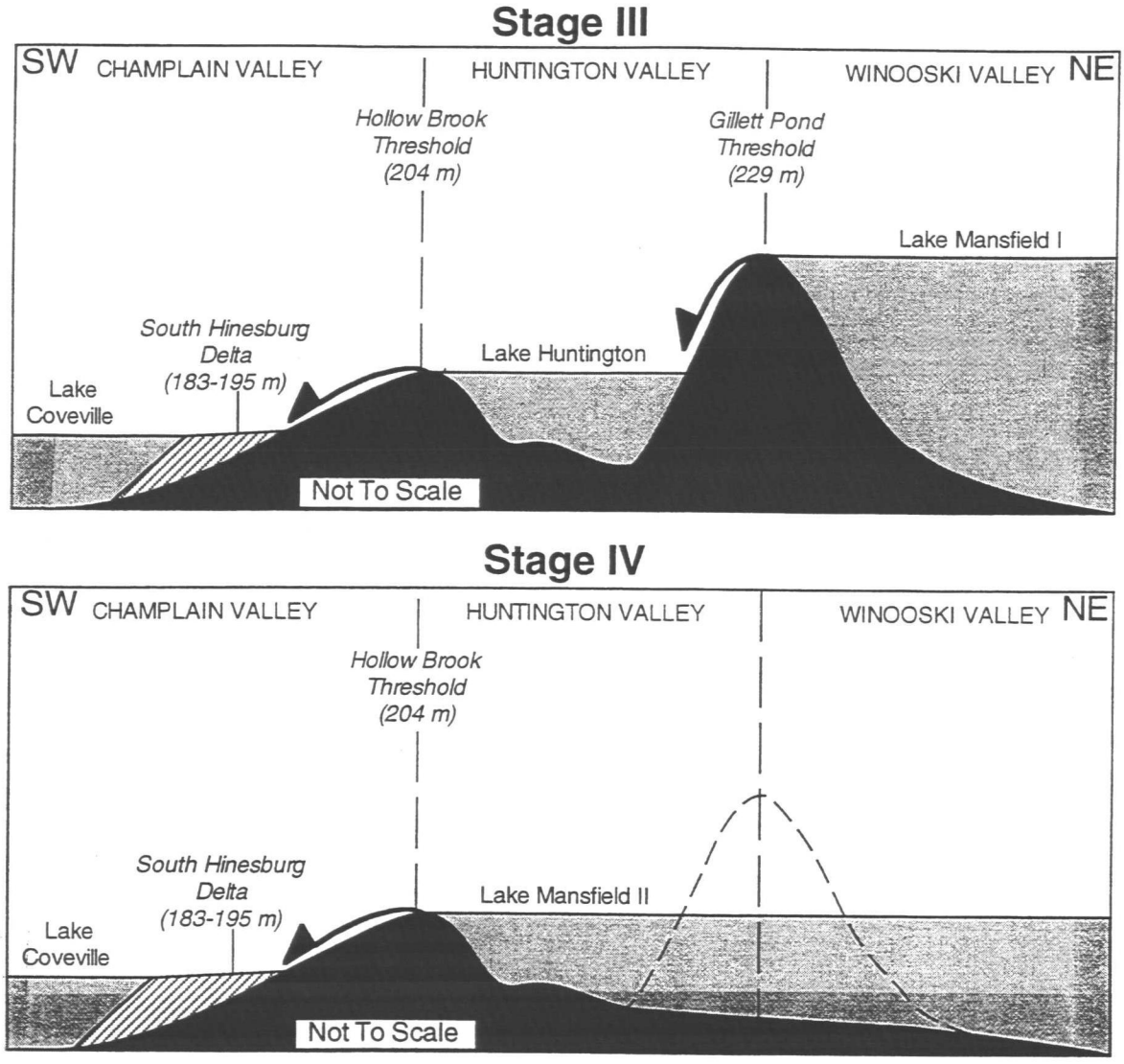


Figure 7: Schematic cross-section through the Huntington Valley during Stages III and IV. During both stages, Hollow Brook was a spillway into the Champlain Valley. During Stage III, when ice blocked the mouth of the Huntington Valley, Lake Mansfield I was forced to drain through the Gillett Pond Spillway. See Figure 6 for ice positions and lake extent.

Subsequent to the valley filling by the glaciolacustrine terrace, dramatic and episodic incision took place. Incision was a response to the proglacial lake base-level drops in the lower reaches of the Winooski Basin. The amount of incision of the glaciolacustrine fill terrace varied along a valley and depended on two factors: 1) the magnitude of base-level change and 2) the distance from the base-level change. Base-level changes continued as first the proglacial lakes, and later the Champlain Sea, in the Champlain Valley lowered to near the present level of Lake Champlain (i.e., Chapman, 1937). Each drop in water level initiated a wave of incision that propagated upstream until a bedrock knickpoint was encountered or until the incision diffused due to a decrease in stream power. Incision resulted in the cutting of the valley fill followed by a period of flood plain development at a lower level; therefore, fill-cut terraces were formed. If a knickpoint lay downstream, the base-level change could not propagate past it and no incision would occur upstream. Base-level changes no longer caused incision after the Upper Marine interval of the Champlain Sea because knickpoints developed along the lower reaches of the Winooski River below Essex.

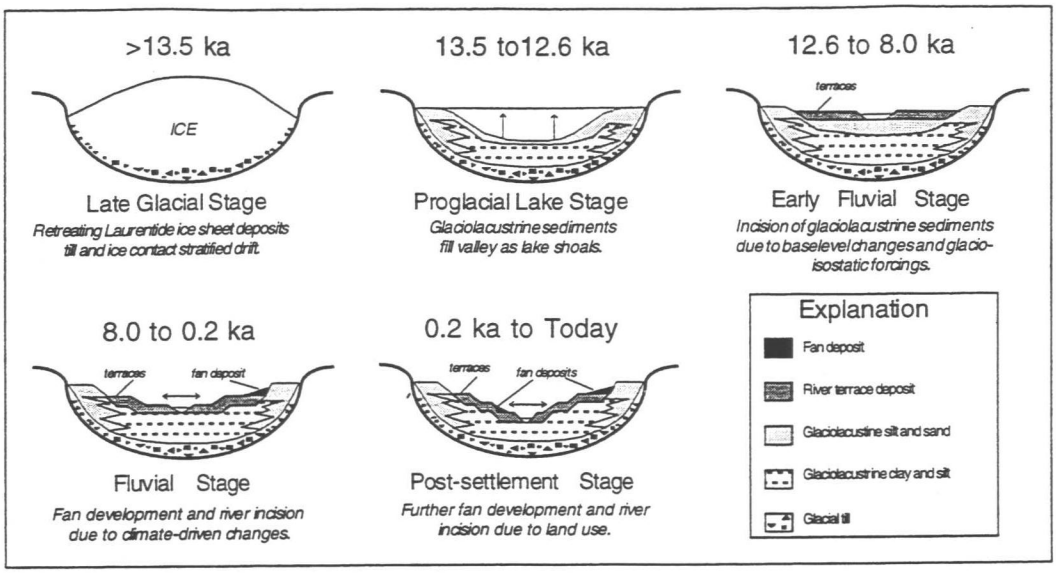


Figure 8: Schematic valley evolution for Winooski Drainage Basin. All ages are in ^{14}C years and are approximate. See text for discussion. Adapted from Church, 1997.

Subsequent to the base-level-driven incision, incision occurred as a response to climate changes and/or internal adjustments of the river system. Widespread climate changes occurred throughout the Holocene and appear to be correlative with periods of fan activity and terrace formation. The relief between terraces decreases after the period assigned to base-level changes which suggests that the mechanism that initiated incision was different. The morphology of the terrace surfaces also changes from a relatively flat feature to terraces dominated by ridges and swales.

Previous Studies of River Terraces in Vermont

One of the only studies of river chronologies in Vermont is that of Brakenridge et al. (1988) who excavated trenches on a point bar of the Missisquoi River, ~60 km north of here, in conjunction with an archeological study. They recovered many datable and identifiable wood fragments from the trenches and they recorded detailed stratigraphy of the terrace deposits. By plotting the channel elevation versus time, their results showed that incision had nearly ceased by 8000 ^{14}C yBP at which point lateral channel migration became the dominant process operating in the river. A period of incision coincides with the time of European settlement in the region (~150 years ago) and presumably the change in river behavior is associated with the clearing of the forests in the watershed for farming and timber.

Brakenridge and others (1988) attributed the decrease in the incision rate at 8000 ^{14}C yBP to be evidence that glacio-isostatic rebound had ended by that time, similar to a time frame proposed by Hutchinson et al. (1981) based on an "undeformed" or horizontal unconformity in the sediments of Lake George that was formed before 7000 ^{14}C yBP. Simple exponential decay models demonstrate that the rate of rebound generally would have decreased to immeasurable amounts during the early Holocene (Whalen, 1997), but episodic incision followed by terrace development continued during the Holocene in the Winooski Drainage Basin. Therefore, the rate of rebound can not in itself control incision, contrary to the suggestion of Brakenridge and others (1988), and other mechanisms must be considered.

Bedrock knickpoints are found in every river, including the Missisquoi below Brakenridge and others' site. Knickpoints stop incision initiated by base-level changes downstream and form local base levels by controlling the river elevation upstream. Even with continued glacio-isostatic uplift during the Holocene, the rate of incision would be dependent on the rate of knickpoint migration upstream. The migration of knickpoints upstream tend to push the fluvial system towards disequilibrium, and periodic internal adjustments, possibly initiated by large floods or widespread climate changes, are evident by the continued formation of terraces during the Holocene. The terrace chronologies in the Winooski Basin reflect a non-synchronous decrease in incision rate among valleys, suggesting that different knickpoints influenced individual valleys at different times.

Moultroup and Aldrich Farms

The flight of terraces at the Moultroup and Aldrich farms (Fig. 9) provides an excellent example of the differences between terraces formed by base level-driven incision and terraces formed by internal adjustments. Eight terraces are identified here and represent three stages of valley evolution, evident by both the relief between terraces and the stratigraphy of terrace deposits. The first stage of the valley evolution is the deposition of the glaciolacustrine fill terrace (T8). The second stage is dramatic incision followed by deposition of fill-cut terraces as a response to base level changes (T7-T6). The final stage is periodic incision related to internal adjustments of the fluvial system triggered by environmental changes (T5-T1).

Alluvial fans present on T5, T2, and T1 in this section of the Huntington Valley were dated by Church (1997) and Zehfuss (1996). The basal dates define the time of initial fan formation, but also provide lower limits to the time of terrace formation. The clustering of dates around distinct periods for fans built on a given terrace and the consistent relationship of younger fans built on lower terraces suggest that terrace formation is at least in part responsible for the timing of fan formation. In addition, limiting ages of the terrace deposits are reported by Whalen (1997) based on terraces correlated to dated base levels and direct dating of the terrace deposits (Tables 2 and 3).

Table 2: Huntington River Terrace and Basal Alluvial Fan Radiocarbon Ages

Terrace	Laboratory #	Material	Depth (m)	^{14}C Age	Calibrated Date (yr B.P.)	1 Sigma Range (yr B.P.)	2 Sigma Range (yr B.P.)
T1*	GX-21329	Wood	4.00	< 100	62-0	128-0	238-0
T2†	CAMS# 30358	Wood	3.60	1900±50	1830	1879-1751	1939-1710
T2*	CAMS# 22994	Wood	4.00	2500±60	2708-2402	2735-2363	2746-2352
T5†	CAMS# 30353	Charcoal	0.75	7790±60	8424	8541-8407	8565-8367
T5**	GX-20276	Wood	4.00	7835±105	8555	8713-8430	8981-8375
T5**	CAMS# 20963	Wood	1.30	8060±60	8981	8993-8764	9194-8662
T5**	CAMS# 20901	Wood	2.50	8530±100	9486	9531-9436	9819-9275
T6†	CAMS# 30347	Charcoal	0.49	8120±60	8991	9189-8980	9240-8736
T6†	CAMS# 30348	Charcoal	0.46	8230±60	9210	9362-900	9380-8988

Note: * Alluvial fan dated by Zehfuss, 1996. ** Alluvial fan dated by Bierman et al., 1997 and Church, 1997.
† Terrace dated by Whalen, 1997. Calibration using CALIB 3.0.3A (Stuvier and Reimer, 1993).

Two pieces of charcoal from the terrace deposits of T6 yielded dates of between 8100-8300 ^{14}C yBP and provide a minimum age for the formation of T6. However, the age of T6 is believed to be much older based on 1) the correlation of it to the Upper Marine of the Champlain Sea, 2)the ages of the fans built on T5, and 3) the age of the T5 deposits. A basal date obtained from wood below the Audubon fan on T5 is 8530 ± 100 ^{14}C yBP (CAMS# 20901) and another basal date from wood found 4.0 m below the Moultroup fan surface across the valley on T5 is 7835 ± 105 ^{14}C yBP (GX-20276). These dates agree well with the date of 7790 ± 60 ^{14}C yBP (CAMS# 30353) from a piece of charcoal recovered from 0.75 m below the T5 surface, suggesting that the T5 was formed by 8500 ^{14}C yBP and was active until at least 7800 ^{14}C yBP. Therefore, the charcoal recovered from T6 appears to been emplaced after the deposition of T6.

The date of 2500 ± 60 ^{14}C yBP (CAMS# 22994) from 4.0 m below Aldrich fan C surface on T2A places an upper limiting age on its formation. The younger alluvial fan, Aldrich B, with a basal age of 1900 ± 50 ^{14}C yBP (CAMS#30358) is located just downstream on T2B. The difference in ages between these adjacent fans points either to the migration of the river channel during this period or another difference in fan initiation. A piece of cloth, presumably a historical artifact, was recovered from the T1 deposits. The Aldrich A fan built on T1 is <100 ^{14}C yBP (GX-2139), consistent with interpreted age of the terrace deposits as historical.

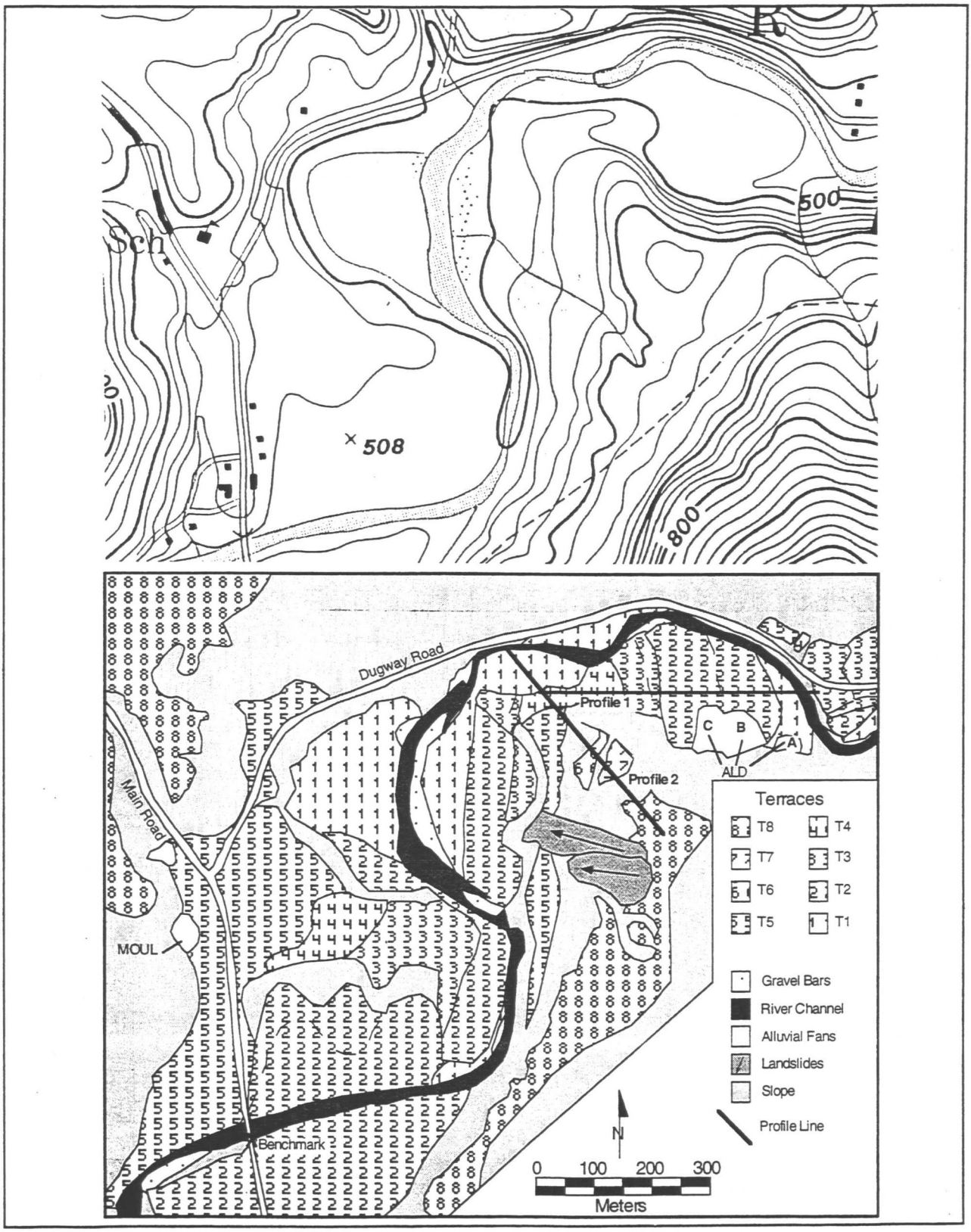


Figure 9: Topographic base map and same-scale map of geomorphic features at Stop 1. See Figures 13 and 15 for detail maps of fans and Figure 10 for profiles of terraces.

Table 3: Summary of Terrace Characteristics at Moultrop Farm

Terrace	Type	River System	Elevation (m)	Incision (m)	Age (^{14}C yBP)
T8	Fill Terrace	Meandering	190.98	-	12.8-12.6
T7	Fill-cut Terrace	Braided	180.44	10.54	12.6-11.7
T6	Fill-cut Terrace	Braided	176.25	4.19	11.7-10.8
T5	Fill-cut Terrace	Meandering	165.57	10.68	10.8-7.8
T4	Fill-cut Terrace	Meandering	162.50	3.07	7.8-6.0(?)
T3	Fill-cut Terrace	Meandering	158.90	3.60	6.0(?)-2.5
T2	Fill-cut Terrace	Meandering	155.47	3.43	2.5-0.2
T1	Fill-cut Terrace	Meandering	151.25	4.22	0.2-Today

Note: Ages based on correlations to dated baselevels in Champlain Valley and dated alluvial fan and terrace deposits (Table 2; Whalen, 1997).

Cross-valley profiles of the terraces clearly show that the magnitude of incision, represented by the relief between terraces, is larger for the highest terraces and decreases towards the river (Fig. 10). The magnitude of incision between both T8/T7 and T6/T5 is over 10 m and for all terraces below T5, less than 5 m of incision took place (Table 3). The obvious differences in relief between the upper (T8-T5) and the lower (T5-T1) terraces corresponds to the change in incision mechanisms and river behavior that is also visible in the terrace stratigraphy.

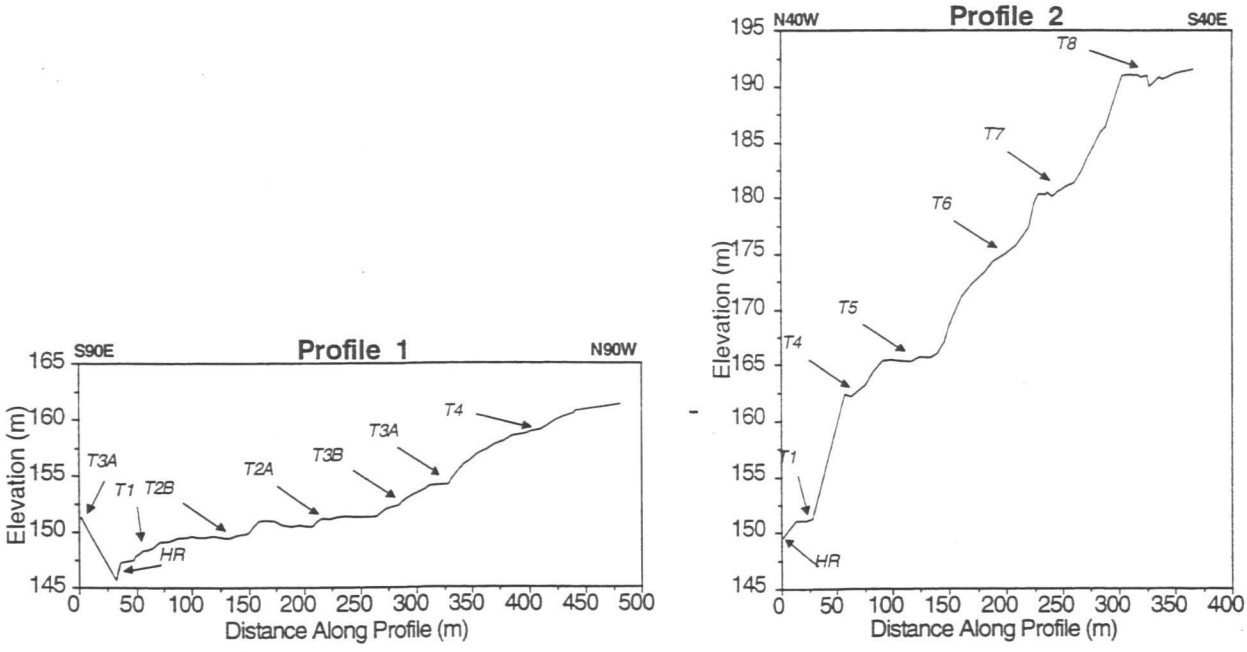


Figure 10: Cross-valley profiles of terraces at the Moultrop and Aldrich farms. Location of profiles is shown on Figure 9. Vertical exaggeration is 10X.

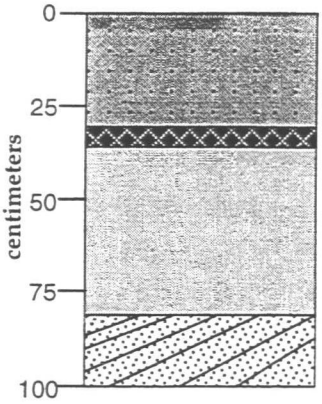
Backhoe trenches and shovel pits, opened on nearly every terrace here, revealed three different terrace stratigraphies (Fig. 11). Common to all terraces is the presence of a 25 cm plow zone capping the fluvial deposits. The plow zone represents the historical land use at this location as tilled fields before the present use as pasture land began about 35-40 years ago (pers. comm., Henry Moultroup). The upper meter of the glaciolacustrine fill terrace, T8, is composed of cross-bedded sand overlain by silty fine sand that suggests deposition by a meandering fluvial system. The stratigraphy of the pits and trenches from T7 and T6 (represented by the T6 log in Figure W5) is consistent with a braided fluvial system. Two fluvial units, channel deposits of imbricated pebbles and longitudinal bar deposits of pebbles and sand, overlie a diamicton interpreted to be glacial till, but overbank deposits are thin or absent on T7 and T6. The terraces deposits from T5 to T1 all show a similar stratigraphy, implying that the fluvial system behaved similarly to today's meandering river since the time of T5. The basal unit of imbricated gravels, deposited in the channel, is overlain by laterally accreted sand which is then capped by overbank deposits of silty fine sand.

The changes in terrace stratigraphy from T8 to T1 demonstrate that the Huntington River has evolved through time. T8 developed at the edge of the shoaling Lake Coveville and therefore probably represents a low gradient river, evident by the abundant fine-grained sediments and meandering-type deposits. However, in upvalley locations the glaciolacustrine terrace is capped with gravel-rich deposits that point towards deposition by a braided river; therefore the proximity to the lake appears to control the type of deposit. The fact that T7 and T6, composed of braided river deposits, formed farther from their respective base levels than T8 is consistent with this relationship. All three terraces developed at a time when incision due to base level changes and hillslope instability due to lack of vegetation would have delivered abundant sediment to the river. Initially, the postglacial landscape was out of equilibrium with prevailing conditions. Vegetation was sparse and due to the instability of the hillslopes, abundant sediment was transported to the river. As vegetation spread across the landscape and the hillslopes stabilized, the sediment supply decreased and equilibrated with the dominant processes, and the gradient of the river decreased as rebound lifted the mouth relative to the river. In addition, the decrease in incision after the initial period of rapid base level changes means that less sediment was eroded upvalley. Together, the changes in the type and amount of sediment supplied to the river led to the development of the fluvial system represented by T5-T1.

T5-T1 formed after the base level changes and glacio-isostatic rebound were substantially complete. The relief between the terraces is less than 5 m at this location, and sublevels are prevalent. Upstream of the Moultroup farm, these sublevels are indistinguishable; so the forcing which is responsible for them was diminished upstream. The proposed mechanism for the formation of these terraces is periodic incision as a response to knickpoint migration. Two kilometers downstream begins the Huntington Gorge. The last terrace to grade across the uppermost section of the gorge is T4. Therefore, the evolution of the gorge through the migration of the knickpoint upstream is a prime mechanism for causing the incision below T5.

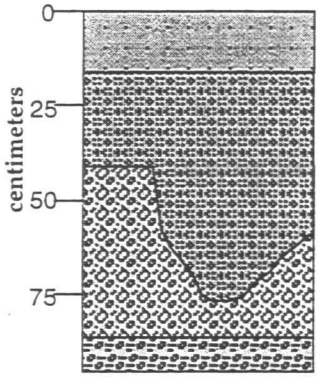
The terraces at the Moultroup and Aldrich farms provide one of the best examples of the different forms of fluvial terraces found in the Winooski basin. The terrace chronologies discussed here are consistent with those studied in the Little and Mad River valleys and may be applicable to other terraced valleys in Vermont.

T8



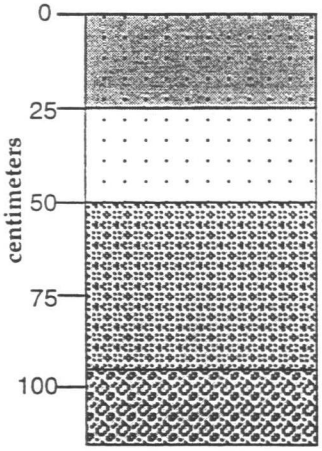
Unit	Sediment	Color	Interpretation
I	Fine Sand with Pebbles	Yellowish Brown 10YR 5/6	Plow Zone
II	Silty Fine Sand	Light Grey 10YR 7/1	Leached Overbank Deposits
III	Silty Fine Sand	Olive Yellow 2.5Y 6/6	Overbank Deposits
IV	Cross Bedded Coarse Sand	Olive Yellow 2.5Y 6/6	Point Bar Deposits

T6



Unit	Sediment	Color	Interpretation
I	Silty Sand with Pebbles	Olive Brown 2.5Y 4/4	Plow Zone
II	Pebbles in a Silty Sand Matrix	Yellowish Brown 10YR 5/8	Longitudinal Bar Deposits
III	Coarse Sand and Imbricated Pebbles	N/A	Channel Deposits
IV	Diamicton: Clayey Silt and Pebbles	Pale Olive 5Y 6/3	Glacial Till

T5



Unit	Sediment	Color	Interpretation
I	Fine Silty Sand with Pebbles	Light Olive Brown 2.5Y 5/4	Plow Zone
II	Silty Fine Sand	Olive 5Y 5/3	Overbank Deposits
III	Coarse Sand and Pebbles	Olive Grey 5Y 5/2	Point Bar Deposits
IV	Imbricated Gravels with Coarse Sand Matrix	N/A	Channel Deposits

Figure 11: Examples of sedimentological differences in the terrace deposits. These three shovels pits along Profile 1, illustrate the changes in river environment through time. See text for discussion.

ALLUVIAL FAN HISTORY, HUNTINGTON, VERMONT

Paul Zehfuss and Paul. R. Bierman

Introduction

Alluvial fans are seldom studied landforms in the Northeast. We have investigated more than twenty alluvial fans in the Huntington River Basin (Bierman et al., 1997; Church and Bierman, 1994 and 1995; Zehfuss and Bierman, 1995), clustered in the northern part of the river valley (Fig. 12). All of the fans are small (<2500 m^2), and have been deposited where Huntington River terraces provide a platform for fan sedimentation. Three sites along the Huntington River have received the most attention and have been used to infer the regional history of hillslope stability and climate change during Holocene. These three sites; Audubon, Moultroup, and Aldrich, include 5 of the 22 alluvial fans identified in the Huntington River Basin, and have supplied information about fan aggradation, and subsequently hillslope erosion, spanning the last 8530 ^{14}C years. This trip will visit the Moultroup and Aldrich fans.

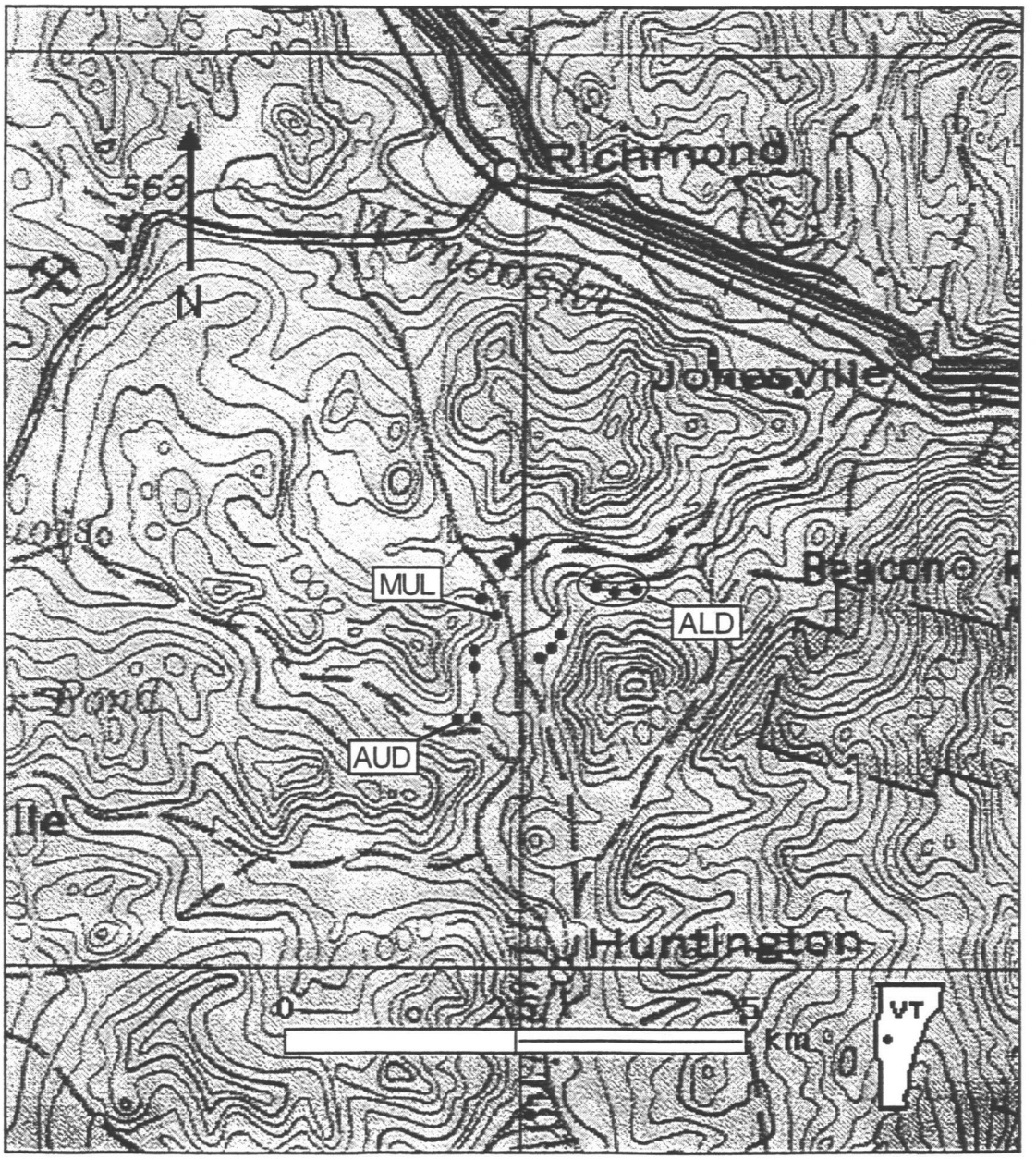


Figure 12: Topographic map showing the location of 22 alluvial fans in the Huntington River Basin. MUL – Moultroup Fan; AUD – Audubon Fan; ALD – Aldrich Fans.

There are similarities between the five fans that we have trenched. Each fan contains a buried soil profile, generally within the upper meter. Radiocarbon dates confirm that the sediment above the organic horizons, often amounting to more than a meter, was deposited coincident with historic destabilization of hillslopes. The soil allows us to identify the period of deposition following the arrival of settlers, who rapidly cleared the hillslopes of trees. A plow zone is usually present above the buried soil, representing the period of time following deforestation during which the fields were used for farming. The abundance of organic material in the fan deposits, such as wood and charcoal, facilitates the use of radiocarbon dating methods for obtaining ages of sediment and estimating rates of deposition.

Most fan sediments are poorly sorted, although there are occasional thin (<10 cm) beds of well sorted, clast-supported gravel, as well as black laminae that may represent decomposed leaf mats or concentrations of finely disseminated charcoal. Material deposited in the fans originates from distinct drainage basins located either on the surrounding mountains, or on slopes created by the edges of higher Huntington River terraces. Very little deposition appears to be occurring on the fans today, although thin (<1 mm) beds of fine silt have been deposited by ephemeral stream activity during severe storms and winter thaws within the past few years.

We have radiocarbon dated 14 samples of wood and charcoal from five alluvial fans in order to determine the timing of aggradation. These data show that two fans began to aggrade in the early Holocene (between 8530 and 8060, and between 7835 and 7360 ^{14}C yBP), two in the late Holocene (2500 and 1900 ^{14}C yBP), and one fan aggraded over 4 m during historic time (<100 ^{14}C yBP). During the field trip, we will view the fans from which we have obtained aggradation histories, as well as to introduce the types of fans common to this region of the Northeast.

Moultroup Fan

The Moultroup fan is seen on the west side of the Richmond–Huntington road, just south of the right-hand turnoff onto Dugway Road (Fig. 12). There is a second, larger fan to the north, with a house built on its surface. The Moultroup fan was trenched extensively by Church and Bierman in the summer of 1995 (Fig. 13). The fan radius is approximately 65 m, with an apex height of 5 m. A band of tall grass in the center of the field marks the approximate toe of the fan, an area of groundwater discharge. The Moultroup fan has accumulated sediment derived from deposits of glacial till and glaciolacustrine deposits that cover the adjacent bedrock highlands.

Trenching of the Moultroup fan revealed a complex stratigraphy. The slightly asymmetrical appearance and surface morphology of the fan suggest that it may be composed of multiple lobes. On the Moultroup fan, there is sufficient radiocarbon and stratigraphic data to estimate sediment accumulation rates over much of the Holocene (Fig. 14). Aggradation rates were rapid in the early Holocene when the fan was beginning to form, between 7835 to 7360 ^{14}C yBP. Sedimentation slowed considerably in the mid-Holocene (7360 to 100 ^{14}C yBP). The stratigraphy of the Moultroup trenches reveals a distinct soil profile near the fan surface buried by about 0.5 to 0.9 m of poorly sorted sediment. A charcoal date (<100 ^{14}C yBP) just above the buried soil indicates that the overlying sediment postdates European settlement and is likely related to land clearance and agricultural practices.

The Aldrich Fans

The Aldrich fans are found on the south side of the Huntington River along Dugway Road, about 1/2 mile east of the Moultroup Fan (Fig. 12). From the river looking south, two fans are readily apparent in the center of the far side of the Aldrich field (Fans B and C, Fig. 15). The fans extend almost 80 meters into the field, and coalesce with one another in the mid-fan portions. Farther east, the smaller Fan A is less obvious (Fig. 15). The Aldrich fans have been deposited on a terrace of the Huntington River that is characterized by uneven terrain that varies in relief by about one meter. The Huntington River marks the northern boundary of the Aldrich farm, where the river has incised nearly 2 m below the surface elevation of the field. Swales between higher deposits of sediment are often filled with puddles following rainstorms and snow melting events.

A steep slope, formed by a higher Huntington River terrace, is the source of fan debris. Drainage basins cut into this higher terrace contain ephemeral streams leading to the fans below and are filled with ice and flowing water in the winter and spring months. Groups of very small stream beds on the terrace top converge at the head of the drainage basins. During winter, thin layers of fine, tan silt are suspended on top of ice which can be seen coating the surfaces of the fans. Clasts ranging in size from very large cobbles to small pebbles line the ephemeral stream beds leading to the fans. Deposits of sand and silt have been exposed along the steep sides of the drainage basins, and in some places have plunged into the stream beds as small land slides (5–10 m^3) containing wet and cohesive sediment. The scarps of the slides appear as hollows below a thick (≈ 25 cm) mat of vegetation.

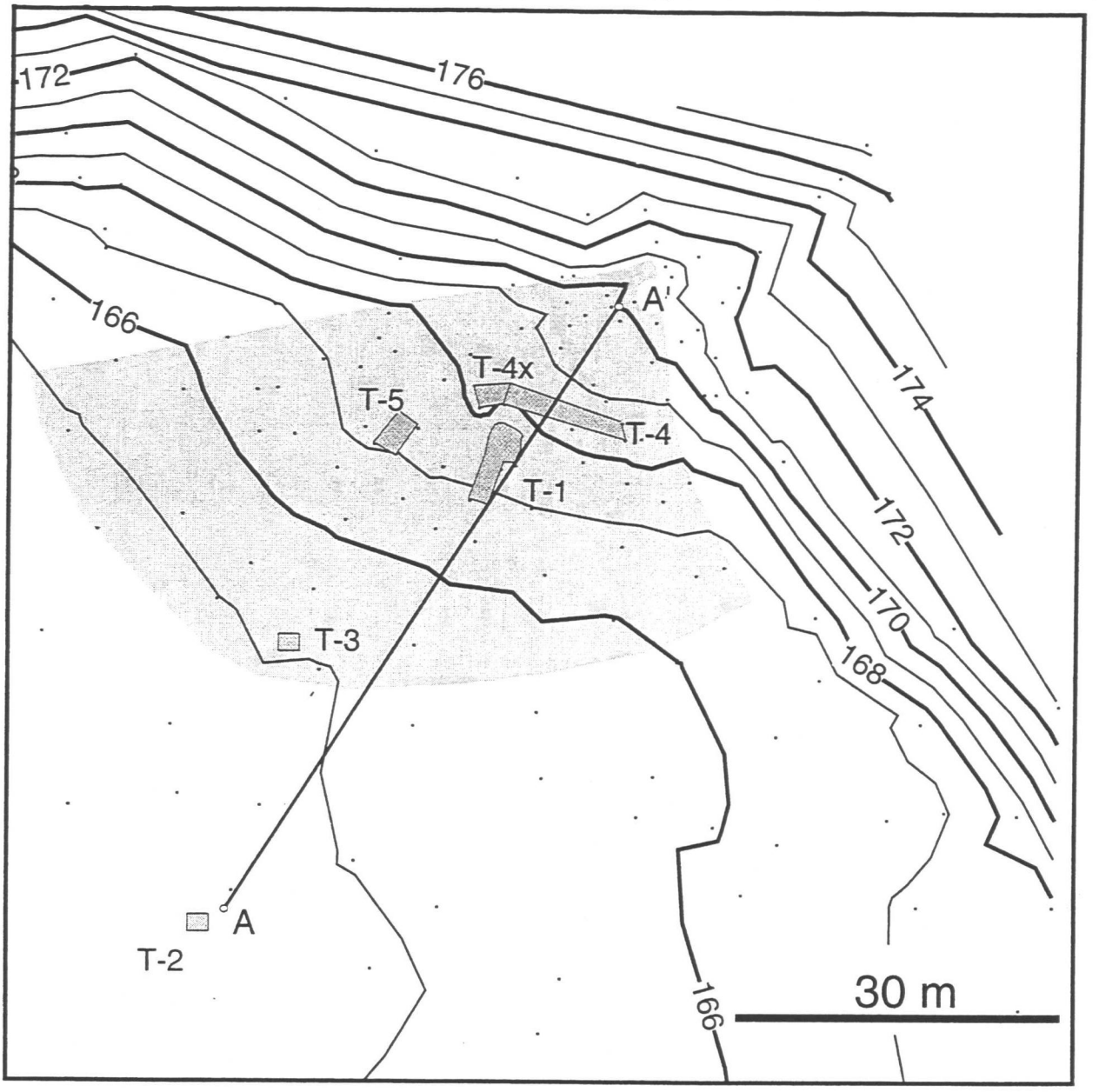


Figure 13: Topographic map of the Moultroup Fan (shaded) showing total station survey points, location of trenches (T-1 to T-5), and line of cross-section (A to A'). North is to the bottom of the map. Contours are in meters above sea level.

Fan A

Fan A is the closest to the Huntington River of the group of three Aldrich fans (Fig. 15) and is predominately composed of sediment deposited after European settlement. A large chunk of wood recovered from four meters below the surface of Fan A was radiocarbon dated at < 100 ^{14}C yBP. Thus, the four meters of sediment above the wood is interpreted to be post-settlement fan deposition. The trench was positioned along the fan surface where there is approximately 5 meters of total fan deposition. This implies that nearly 80% of the vertical aggradation at this location on fan A is composed of post-settlement fan deposition.

Fans B and C

Five trenches were excavated into fans B and C and logged (Fig. 15). Radiocarbon dates provide basal ages for the fans (Fan B, 1900 ^{14}C yBP; Fan C, 2500 ^{14}C yBP), as well as time markers within the stratigraphy exposed by the fan trenches. Figure 16 is a trench log from Fan B. A significant period of relative fan surface stability is marked

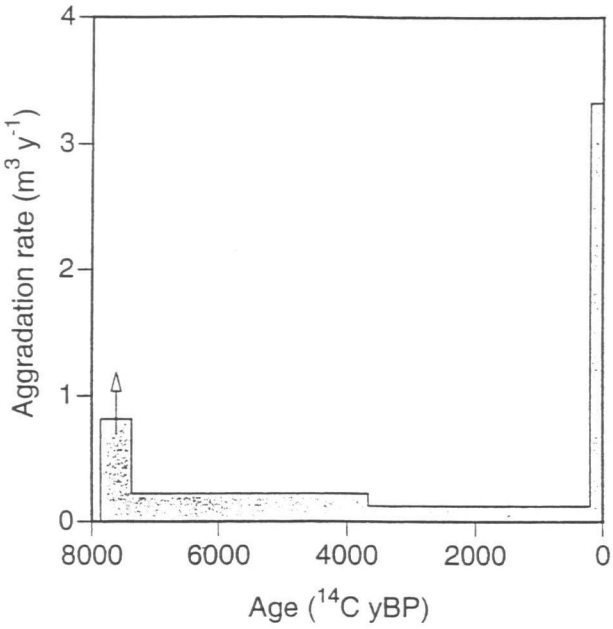


Figure 14: Estimated aggradation rates in ^{14}C years to the Moultroup Fan (assuming that fan geometry is that of a right circular cone) demonstrate response of fan during historic period (180 years, based on settlement history).

by the paleosol A horizon identified in most Aldrich trenches. The buried soil represents a period when vegetation became well established on the fan surface, and also indicates that hill slopes were supplying little sediment to the fan. Radiocarbon dates of 230 and 310 ^{14}C yBP obtained from wood and charcoal in the paleosol A horizon (Fig. 16) allow the soil horizon to be correlated between trenches (Fig. 17) and further demonstrate the fan sedimentation during the time of European settlement. We suggest that the extensive deforestation that occurred during settlement in the late eighteenth century destabilized the hillslopes. Without stability provided by root systems, the soils and sediments were more easily eroded from drainage basins, increasing sedimentation on the fan surfaces.

Following deforestation, sediment continued to be deposited at a relatively rapid rate as compared to depositional rates previous to settlement. As the farming industry began to develop in Vermont, the Aldrich field was utilized for agricultural purposes. Verbal communication with residents of the Huntington and Richmond regions has verified that the field was plowed repeatedly previous to 1960. A cross-section of Fan B (Fig. 17) shows the location of the plow zone in the trenches. Towards the toe of Fan B, the paleosol A horizon is incorporated into a cumulative plow zone. It can be inferred that the paleosol A horizon may have existed throughout the entire Aldrich field at one point, but was combined, where it was not buried by alluvial fan sediments, with the cumulative plow layer during farming. Following the cessation of row-crop farming on the Aldrich meadow and its subsequent and continued use for hay and pasture, the cumulative plow layer has been buried by continued fan deposition (Fig. 17). The recovery of forest growth on the hillside has returned some, but not complete, stability to slopes in the drainage basin.

Conclusions

Our investigation of alluvial fans in the Huntington River Basin has shown that fans provide an accurate, datable record of post-glacial hillslope stability. The sensitivity of fans to their environment, climatic conditions and vegetation, make them useful when quantifying the landscape response to shifts in environmental conditions. When assessing the impact of human-induced landscape modifications, fans are sobering examples of land-use impacts, and remind us of the need for prudent land management.

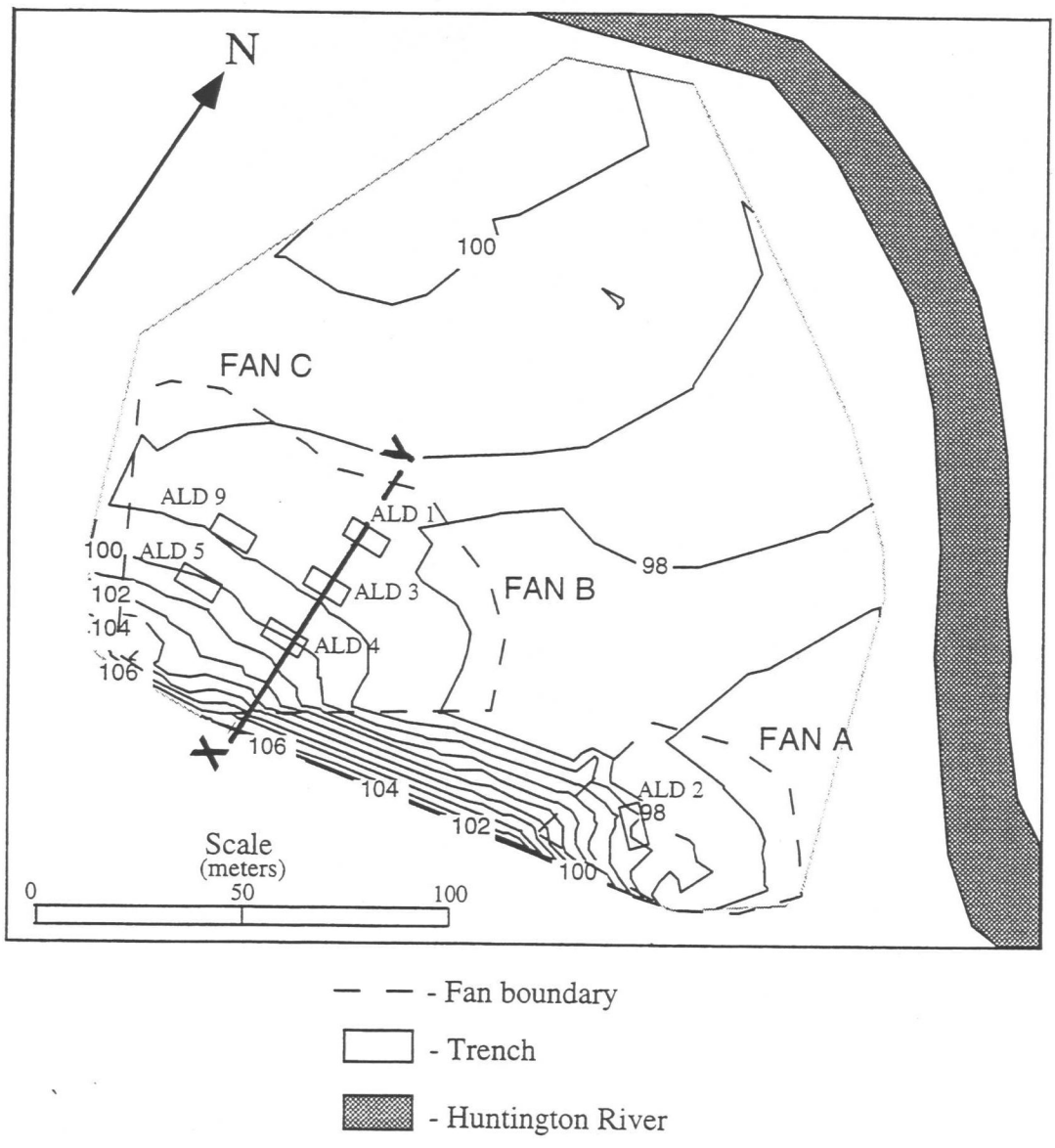


Figure 15: Contour map of the Aldrich field showing the location of 6 backhoe trenches (e.g. "ALD 3"). Alluvial Fans A, B, and C are outlined by dashed lines. Cross-section X–Y through Trenches ALD 1, ALD 3, and ALD 4 (Fig. 17) is shown with heavy line. Contours are in meters above an arbitrary datum.

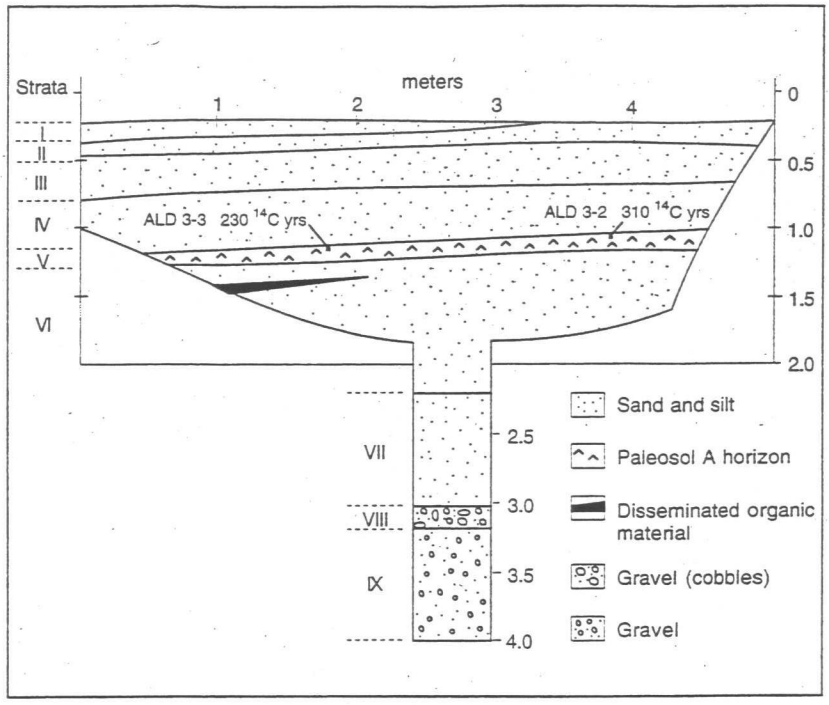


Figure 16: Log of Trench ALD-3 in Fan B (Fig. 15) showing river gravels (Units VIII, IX) underlying fine-grain overbank (Unit VII) and post- and prehistoric alluvial fan deposits (Units I through VI) along with radiocarbon ages on single pieces of charcoal from the paleosol horizon. View is to the south. Vertical scale is in meters and there is no vertical exaggeration.

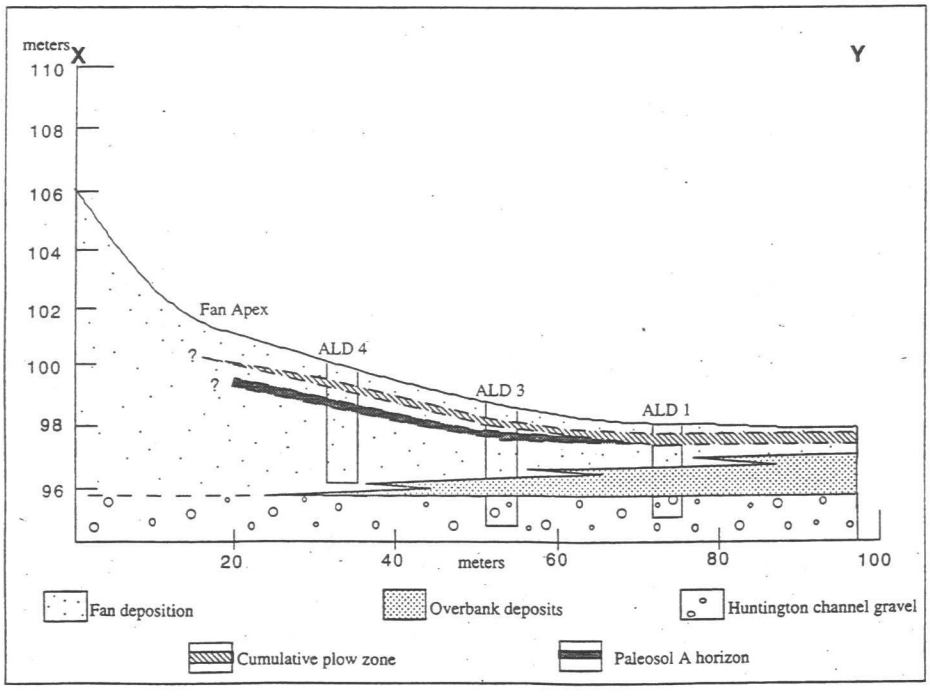


Figure 17: Cross-section of Aldrich Fan C, looking West (Fig. 15), incorporating information from Trenches ALD 1, ALD 3, and ALD 4. Cross-section is drawn with 3.7X vertical exaggeration and elevations are relative to an arbitrary datum.

ACKNOWLEDGMENTS

We are deeply indebted to the willing access granted us by the Moultroup and Aldrich families of Huntington and the Lake Mansfield Trout Club for allowing not only us, but many UVM students to traverse and poke at the fascinating landforms on their properties. We also thank Michael Loso and Howard Schwartz for allowing us to use their detailed map of portions of the Miller Brook valley as a basis for some of the work presented here. Thanks are also extended to Lori Barg for assisting with surveying in the Mill Pond area.

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ROAD LOG

Pico Ski Area to Richmond Exit I-89

Mileage: Cummulative from Pico	Mileage: Cummulative from South Hinesburg	Mileage: Since last turn	
0.0	-	0.0	Start mileage when leaving Pico parking lot, Left on US 4 west
8.8	-	8.8	Right on US 7 north in Rutland
37.3	-	28.5	Right on VT 116 north/ VT 125 east
37.8	-	0.5	Left turn off of VT 125 east, follow VT 116 north straight after stop sign
49.6	-	11.8	Right at stop sign onto VT 116 north, VT 17 west
64.2	0.0	14.6	Right onto Hollow Brook Road, Alternative start at Blaises Country Store in South Hinesburg
67.4	3.2	3.2	Hollow Brook Divide
69.4	5.2	2.0	Left onto Main Raod at stop sign
73.0	8.8	3.6	**Moultrop Farm, STOP 1** (Henry Moultroup: 434-2279)
			Walk 0.6 miles north from the Moultrop Farmhouse, down Dugway Road to crossing.
73.2	9.0	0.2	Dugway Road
73.4	9.2	0.2	Top of T8
76.7	12.5	3.3	Left before round church, following (To US 2 sign)
77.3	13.1	0.6	Left at light onto US 2 west
78.8	14.6	1.5	I-89 on ramp in Richmond

Richmond Exit I-89 to Miller Brook (Stop 3)

Mileage

0.0 **Richmond Exit, I-89:** Travel east along I-89 and exit at Waterbury.

13.7 **Intersection of I-89 and Route 100:** Take Route 100 north 7.1 miles until reaching the turnoff (left or west) for Moscow. The turnoff is marked by a small green sign and arrow.

20.8 **Intersection of Route 100 and Town Road to Moscow:** Follow this road west through the village of Moscow. The road follows the Little River valley west and then south.

22.8 **Bridge over Miller Brook:** Follow the paved road around a sharp right turn immediately after crossing the bridge. A dirt road continues straight along the river valley and eventually dead ends at the Waterbury Reservoir. The road now follows Miller Brook. Continue up the road 0.4 miles. An active alluvial fan is forming on the meadow along the left-hand side of the road. Above map is an enlarged portion of the Stowe 7.5' Quadrangle (1:12,000, North up) showing the location of Stop 3. Miller Brook and the Little River have been highlighted for clarity.

23.2 **STOP 3: ACTIVE ALLUVIAL FAN**
Request permission to visit this site from the occupants of the farmhouse below the fan. One can usually park on the meadow or along the side of the road.
Continue up the Miller Brook Road.

24.5 **Road to gravel pit takes off to left.** Continue straight ahead on the Miller Brook road which turns to dirt shortly after the intersection.

25.0 **Gravel Pit:** Time permitting we will visit this pit on our return down the valley. Continue up the road. Use caution as the road winds and is extensively used by cyclists and pedestrians.

26.4 **Intersection with road to right opposite dam on Miller Brook:** Turn right and follow this road (Old County Road) approximately 0.1 mile to pull out on left that leads to a parking area along the edge of a small meadow.

26.5 **STOP 4: MILLER BROOK LANDFORMS**
Most of the area traversed on this stop belongs to the Lake Mansfield Trout Club. Please request permission before visiting the Trout Club property (253-7565).

END OF TRIP

TRIP C6: BASEMENT, METASEDIMENTARY, AND TECTONIC COVER OF THE GREEN MOUNTAIN MASSIF AND WESTERN FLANK OF THE CHESTER DOME

by
Nicholas M. Ratcliffe, U.S. Geological Survey, 926A National Center, Reston, VA 20192
Gregory J. Walsh, U.S. Geological Survey, 926A National Center, Reston, VA 20192
John N. Aleinikoff, U.S. Geological Survey, Box 25046, 963 Denver Federal Center, Denver, CO 80225-0046

ABSTRACT

Middle Proterozoic rocks of the Mount Holly Complex (1.35 to 1 Ga) in the Green Mountain massif (Ratcliffe and others, 1991) are unconformably overlain by autochthonous Late Proterozoic and Cambrian cover rocks of the Dalton, Tyson, Hoosac, and Plymouth Formations. West of the massif, this basal section passes conformably upwards into Cambrian through Lower Middle Ordovician shelf-sequence rocks. Along the eastern side of the massif, the basal autochthonous sequence is overlain by Late proterozoic through Middle Ordovician rocks of the Pinney Hollow, Ottauquechee, Stowe, Moretown, and Cram Hill Formations. Middle Ordovician conodonts found in a limestone sliver beneath the Pinney Hollow Formation require major faults in the eastern cover sequence. U-Pb zircon ages of 496 to 462 Ma (Lower Ordovician) from island-arc intrusive and volcanic rocks in the Moretown and Cram Hill Formations indicate that the upper part of the eastern cover sequence is much older than previously recognized. It formed as a Taconian accretionary package that was partially assembled before 496 Ma and finally emplaced by synmetamorphic thrusting before about 445 Ma in the Taconic orogeny. This tectonic cover now mantles the Chester and Athens domes, and the eastern flank of the Green Mountain massif.

INTRODUCTION

New geologic mapping of the eastern cover sequence as defined by Doll and others (1961) east of the Green Mountain massif from the Massachusetts border to Rochester, Vermont, indicates that the previous portrayal of the rocks as a continuous, east-facing and eastwardly younging succession is not correct. Mapping in the southern part of the state by T.R. Armstrong and N.M. Ratcliffe has shown that south of Jamaica, Vermont, the cover sequence rocks are so highly thrust faulted and folded that any semblance of continuity of stratigraphic units above the Hoosac Formation is lost (Ratcliffe, 1993, 1997; Ratcliffe and Armstrong, in press). Careful lithologic mapping has shown that units between the base of the Moretown Formation and the top of the Hoosac Formation in southern Vermont are coextensive with the Rowe Schist of Massachusetts. The Rowe Schist in Massachusetts has been interpreted as a heterogeneous Taconian tectonic mélange (Stanley and Hatch, 1988). Armstrong and Ratcliffe agree with that interpretation (Ratcliffe, 1997; Ratcliffe and Armstrong, in press; Ratcliffe and others, 1992). For a more complete list of our published and in press geologic maps see the introduction to trip B-6.

Our regional mapping (Walsh, Armstrong, Ratcliffe) indicates that the Pinney Hollow, Ottauquechee and Stowe Formations are continuous south as far as the the Saxtons River quadrangle, where additional faulting and folding disrupts the section. When traced southward through the Townshend quadrangle into the West Dover and Jacksonville quadrangles (Ratcliffe and Armstrong, in press) only the rocks of the Stowe Formation immediately beneath the Moretown are still continuous with rocks to the north. The bulk of the Rowe Schist in Massachusetts maps northward into the area occupied principally by the Stowe to the north (Ratcliffe and Armstrong, in press). Rocks assignable to the position of the Ottauquechee Formation are absent in the Townshend, West Dover and Jacksonville quadrangles and carbonaceous schists occur in many different structural positions in what we map as Rowe Schist. Thus the structural sequence of Pinney Hollow,

Ottauquechee and Stowe is no longer mappable. For a cross section that illustrates the complicated tectonic stacking in the cover section, see the north-south section A-A of the Jamaica-Townshend U.S. Geological Survey I-series map (Ratcliffe, 1997).

Since 1992, we have mapped the Pinney Hollow, Ottauquechee, and Stowe Formations north and south of the Plymouth area and have determined the detailed internal structure of the units (fig. 1). We find that marked north to south thinning is accompanied by map scale truncation of internal units at the contacts of the Pinney Hollow, Ottauquechee, Stowe, and Moretown Formation (Walsh and Ratcliffe, 1994a, b). Near major contacts, transposition of older schistosity is intense and a characteristic set of reclined (F_2) folds and pronounced down-dip rodding is present. These structures are the same features described by Ratcliffe (1993) in the Mount Snow-Readsboro area and by Ratcliffe and others (1993) at Whitcomb Summit thrust in Massachusetts, which places Rowe Schist on similarly deformed rocks of the Hoosac Formation at the northern termination of the Berkshire massif. This deformation is the regional (F_2) Taconian structure dated in Massachusetts and southern Vermont by $^{40}Ar/^{39}Ar$ biotite and hornblende ages (Sutter and others, 1985). In Massachusetts the Taconian F_2 fold and thrust fabrics are synmetamorphic and were active during peak Barrovian metamorphism. This metamorphism post dates emplacement of the Taconic allochthons and affects rocks as young as the Middle Ordovician Walloomsac Formation. $^{40}Ar/^{39}Ar$ ages of hornblende at $\simeq$445 Ma from the Middle Ordovician Walloomsac Formation in western Massachusetts provide the best available age for the time of the Taconian Barrovian M_2 metamorphism in western New England (Hames and others, 1991).

Based on the structural criteria mentioned above, we (T.R. Armstrong, G.J. Walsh, and N.M. Ratcliffe) have interpreted the contacts between the Pinney Hollow, Ottauquechee, Stowe, and Moretown largely as Taconian thrust faults (fig. 1). These formed during a period of late Taconian synmetamorphic imbricate thrust faulting that also affected Laurentian sialic basement of the Green Mountains as well as its autochthonous cover rocks. Locally, in the east, still younger Acadian thrust faults are recognized to mark major faults between units.

The above interpretations have been based largely on physical criteria supported by collateral arguments about the paleotectonic setting and origin of igneous rock suites found within the eugeoclinal section (see geochemistry section below). These data suggest a change upwards within the Hoosac, Pinney Hollow, Ottauquechee, and Stowe from alkaline within-plate rift basalts to more MORB-like oceanic basalts. In southern Vermont the transition to MORB chemistry, however, occurs within the metabasalts of the Hoosac Formation (Ratcliffe, 1993). Metabasalts of the Pinney Hollow, Ottauquechee, Stowe and Rowe Formations we have analyzed are in general light-REE-depleted like those reported by Coish and others (1986, 1987). The transition to bimodal island-arc volcanism occurs within rocks of the Moretown Formation. This information and the distribution of slivers of ultramafic rocks (dunite, serpentinite, and talc-carbonate rock) at various places within the section led Stanley and Ratcliffe (1985) to conclude that the Pinney Hollow, Ottauquechee, Stowe, and Moretown Formations represent a tectonically assembled package of rocks formed by accretionary collapse and eventual collision of Taconian island-arc(s) against the rifted Laurentian margin in the Taconian orogeny.

The discovery of Middle Ordovician conodonts (reported here) in a dolostone lens in phyllite previously defined as the Upper Member of the Plymouth Formation (Ratcliffe, 1994) indicates that rocks as young as approximately 480 to 460 Ma underlie the Pinney Hollow Formation. Conventional U-Pb and ion microprobe ages of zircon from felsic intrusive and volcanic rocks in the Moretown and Cram Hill Formations range from 460 Ma to about 496 Ma (reported here, see geochronology section). This indicates that the rocks tectonically at the top of the section are as old or older than some rocks within the lower part of the section. These results provide the first bit of paleontologic and geochronologic data supporting the structurally derived tectonic model proposed for the eastern Vermont eugeoclinal sequence by Stanley and Ratcliffe (1985).

We propose (following Stanley and Ratcliffe, 1985) that the eastern Vermont eugeoclinal sequence from the base of the Pinney Hollow through the Moretown Formation, and comparable rocks in southern Vermont and

Massachusetts (the Rowe Schist and Moretown Formation) are a tectonically assembled package and that the units are all fault bounded. The rocks structurally below the Moretown Formation have tectonic and paleodepositional affinities to late Proterozoic and Cambrian rift sediments of the Hoosac Formation, deposited on and east of the rifted Laurentian margin, and to Late Precambrian and Cambrian rocks in the Taconic allochthon. The Moretown and younger rocks represent rocks of the accretionary prism, fore arc, and arc-related volcanic rocks. This eastern belt was accumulated as an accretionary wedge and intruded by arc volcanic rocks before emplacement in the Taconic orogeny.

Purpose and Design of Trip

We will visit outcrops within the autochthonous Laurentian cover immediately east of the Green Mountain Massif (Stops 1 to 6) and will demonstrate that even this section is cut by numerous F_2 Taconian generation thrust faults. Two traverses will then be taken across the allochthonous Pinney Hollow, Ottauquechee, and Stowe Formations in the Plymouth and Ludlow quadrangles where we will present arguments for major thrust faults, both Taconian and Acadian, along formation boundaries.

In the Ludlow and Andover quadrangles the structural geology and age of the Moretown Formation, the Cram Hill Formation and of the Proctorsville ultramafic belt will be shown to be Lower Ordovician or older based on new U-Pb dating of zircons from igneous rocks of the Barnard Gneiss (of Richardson, 1924) and of the North River Igneous Suite (of Armstrong, 1994) that intruded and are interlayered in the Moretown and Cram Hill Formations.

Our final conclusion is that the allochthonous units in the eastern Vermont eugeoclinal sequence, the Pinney Hollow, Ottauquechee, Stowe, Moretown and Cram Hill Formations and the Rowe Schist in southernmost Vermont are all bounded by Taconian thrust faults and that this section is a structural amalgamation rather than a lithostratigraphic sequence.

STRATIGRAPHY

Major Lithotectonic Units

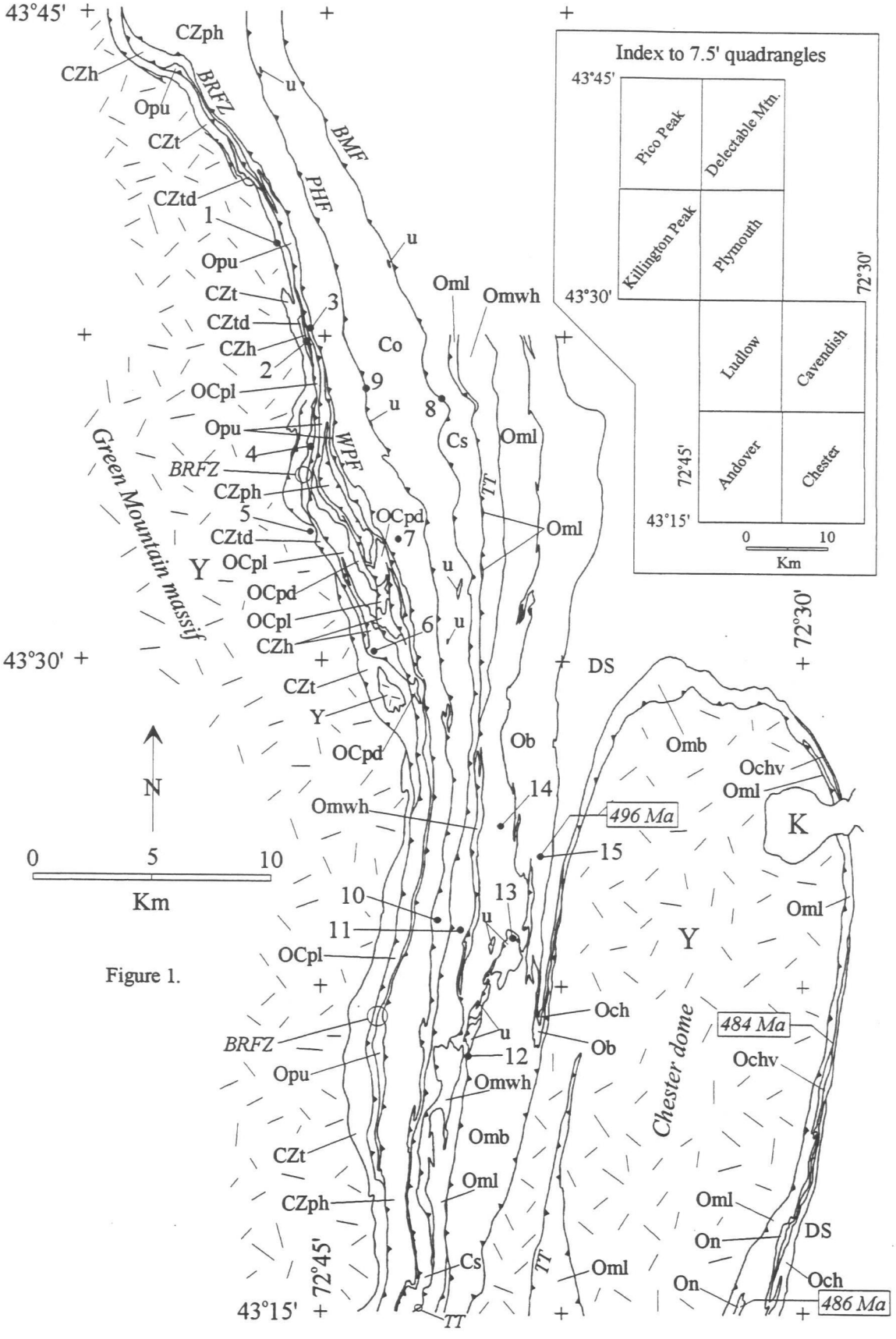
In very general terms, the rocks on the eastern flank of the Green Mountain massif can be assigned to the following lithotectonic assemblages following Stanley and Ratcliffe (1985):

(1) Rocks of the Laurentian plate. These include the Middle Proterozoic Mount Holly Complex and its immediate unconformable cover of the Tyson, Hoosac, and Plymouth Formations that together formed the basement and cover affected by ancient Late Proterozoic rifting to form the Laurentian passive continental margin in the Cambrian and Ordovician.

(2) Eugeoclinal rocks of the late-rift and oceanic-drifting stage. These include the Late Proterozoic and Cambrian Pinney Hollow, Ottauquechee, and Stowe Formations, plus the Rowe Schist of southern Vermont. Although these rocks were deposited in a deepening and widening Iapetan ocean basin in the Late Proterozoic and Cambrian, sediment polarity was largely from the Laurentian margin and, therefore, these rocks are not truly exotic rocks. Their deposition probably spanned the transition from rift subsidence to ocean basin formation and drifting.

(3) Rocks of the oceanic arc and fore-arc, and accretionary prism. These include the rocks of the Ordovician Moretown Formation and the Cram Hill Formation, and may include felsic and mafic volcanic or intrusive rocks of the Ordovician Barnard Gneiss of Richardson (1924) and the North River Igneous Suite of Armstrong (1994). These rocks and the contained ultramafic rocks are exotic to the Laurentian plate. They form the upper plate of the eastern Vermont eugeoclinal sequence that was tectonically assembled in the Taconic orogeny.

(4) Post-accretionary and late-Taconian through post-Taconian metasedimentary and volcanic rocks. These may include some rocks in the uppermost part of the Ordovician Cram Hill Formation and the Silurian and Devonian rocks of the Waits River Formation of the Connecticut Valley sequence.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BMF
PHF
WPF
Green Mountain massif
Chester dome
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484 Ma
486 Ma
Index to 7.5' quadrangles
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Killington Peak
Plymouth
Ludlow
Cavendish
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Figure 1.

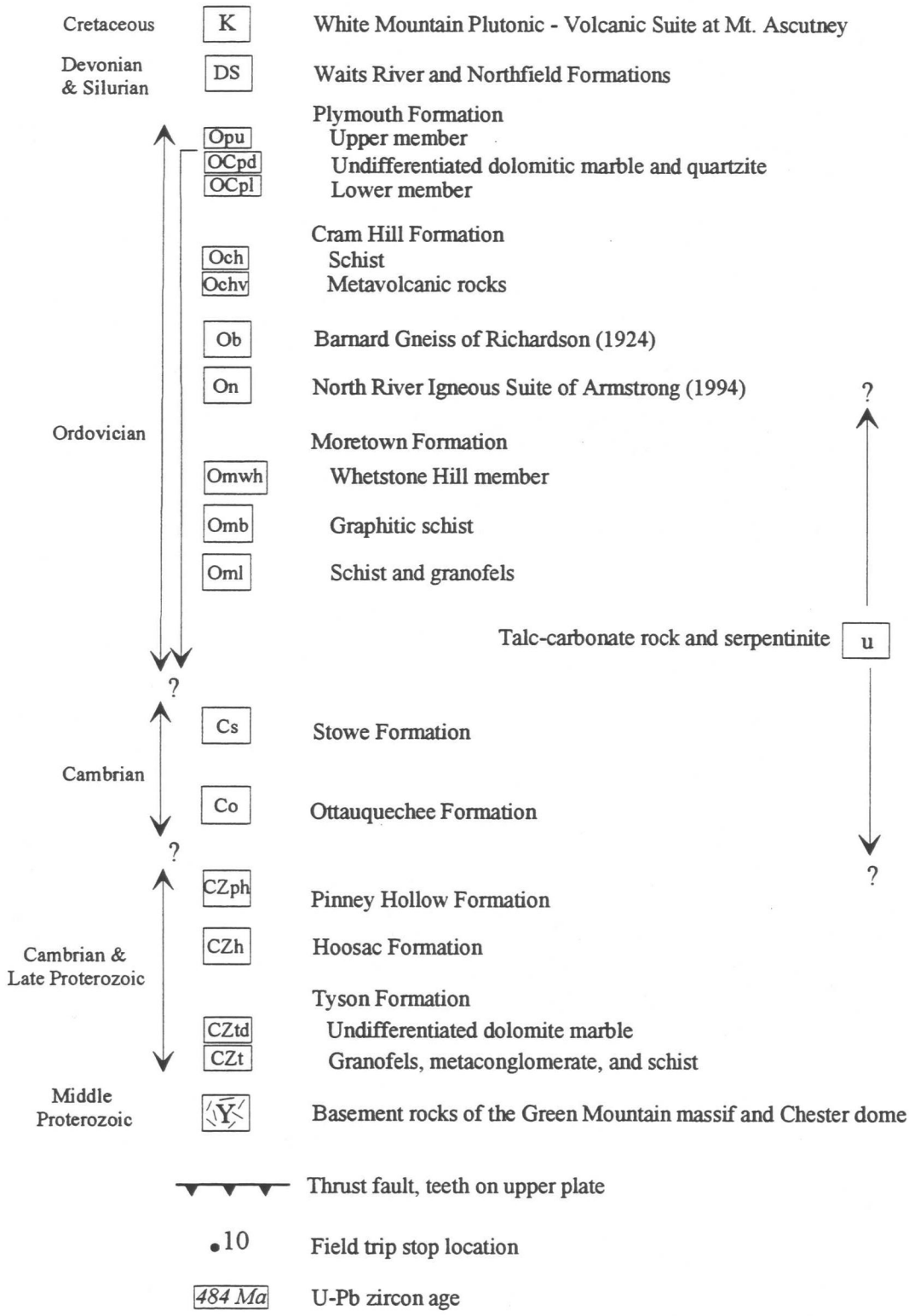


Figure 1. Simplified geologic map of the eastern cover sequence. Geology compiled from Ratcliffe (1992); Walsh and Ratcliffe (1994a, b); Walsh and others (1994); Ratcliffe (1995a, b, c); Ratcliffe (1996a, b); and Walsh and others (1996a, b). BRFZ = Black River fault zone, WPF = Wood Peak fault, BMF = Bald Mountian fault, TT = Townshend thrust, PHF = Pinney Hollow fault.

Minor Devonian granite dikes and small plutons intruded late in the Acadian orogeny. Abundant post-tectonic Cretaceous lamprophyre, diabase and syenitic dikes and plutons are related to a major episode of intraplate alkaline plutonism.

Autochthonous rocks

Figure 2 (reproduced from Ratcliffe, 1994) shows the previous usage of stratigraphic nomenclature west of and beneath the Wood Peak fault at the base of the Pinney Hollow Formation. For a description of the units see Table 1. The symbology used in table 1 is taken from the Open-File Report 94-225 of this area by Walsh and Ratcliffe (1994a). In figure 3 the interpreted restored section from (Ratcliffe, 1994) shows the relationship between the clastic and carbonate rocks east of the Green Mountain massif with rocks of the miogeoclinal shelf sequence near Rutland. The approximate stratigraphic position of the fossiliferous dolostone (d) at Stop 3 in black phyllites of the Plymouth Formation. Ratcliffe still favors the idea that the carbonate rocks of the Tyson Formation and those of the Plymouth Formation are laterally equivalent and were brought into juxtaposition by thrust faulting on the Black River thrust zone in the Taconian orogeny. However, the black phyllite (upper member of the Plymouth) that contains Middle Ordovician conodonts is no longer considered Cambrian as shown in figure 2 (see below).

Rocks of the autochthonous sequence occur intermittently along the eastern side of the Green Mountain massif where the basal unit, either the Tyson Formation or the Hoosac Formation, rest unconformably on the Mount Holly Complex.

Rocks of the Plymouth Formation extend for 2 km north of Stop 3 and south 21 km to the southern border of the Ludlow quadrangle, where rocks of the Pinney Hollow Formation directly overlie the Hoosac Formation on the Wood Peak fault (fig. 1)

Discussion of conodonts at Stop 3 and stratigraphy of the Plymouth Formation

We report the finding of Anita Harris (personal communication to Ratcliffe, 1997) that a black carbonaceous dolostone that we sampled from just beneath the base of the Pinney Hollow Formation contains poorly preserved but identifiable Lower Middle Ordovician conodonts. The black dolostone, approximately 10 meters thick, is interbedded with black carbonaceous phyllite that is continuous to the south with black phyllite previously mapped as the Upper Member of the Cambrian Plymouth Formation as redefined, apparently incorrectly by Ratcliffe (1994). Formerly, this unit was defined as black schist of the Plymouth Member of the Hooosac Formation (Doll and others, 1961).

The lower member of the Plymouth Formation consists of feldspathic quartzite, and quartzite. The Middle Dolostone Member contains breccia beds, and laminated blue-gray dolostone and more massive cream to beige dolostone (Ratcliffe, 1994). The upper member of the Plymouth is a black, carbonaceous phyllite that contains beds of blue-gray dolostone and thin layers of dolomitic quartzite. For the distribution of these units, see Walsh and Ratcliffe (1994a, b.)

It now appears that the rock mapped as the upper member of the Plymouth contains the Middle Ordovician fossils referred to above. However, attempts to obtain conodonts from dolostones in the type Plymouth for comparison to those at the fossil locality were not successful as all additional samples were barren.

The black phyllite of the Plymouth, as previously defined (Ratcliffe, 1994), overlies dolostone beds that contain sedimentary breccia beds similar to those occurring within the Rutland Dolomite of Brace (1953) that occur at the base of the Cambrian through Ordovician shelf sequence of the Vermont Valley (Thompson, 1972).

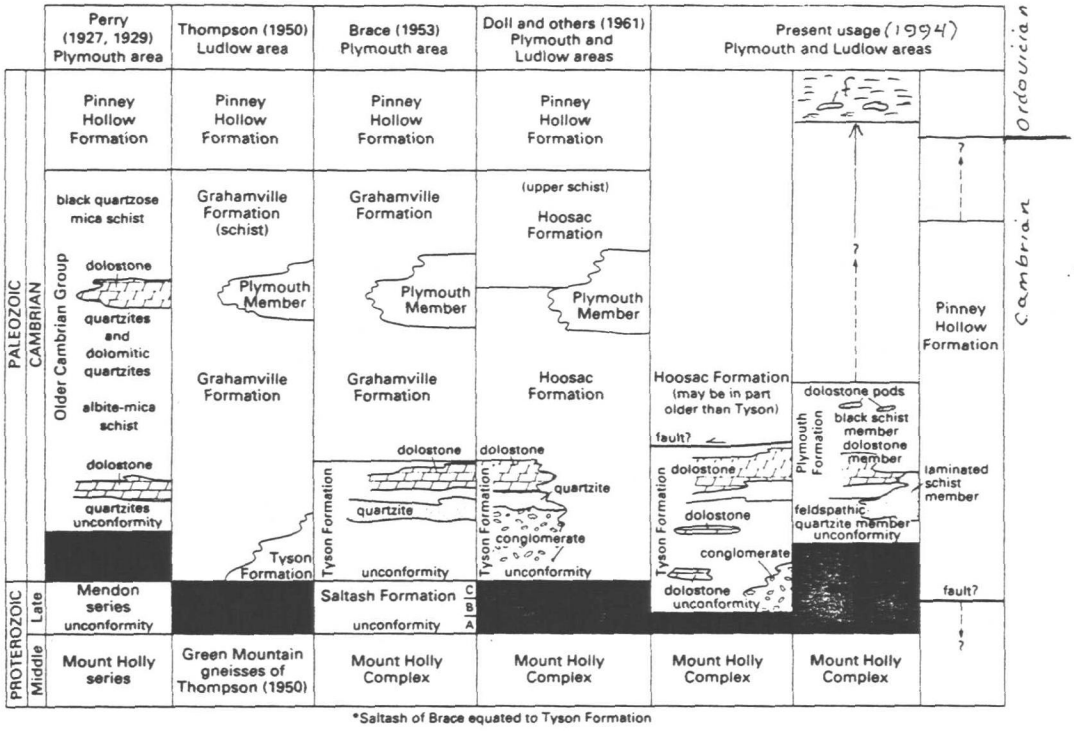


Figure 2. Correlation chart from Ratcliffe (1994) showing previous usage; f denotes revised (1997) position of the unnamed fossiliferous black schist of the former Upper Member of the Plymouth Formation into the Middle Ordovician. Revised time scale shown on the right side of diagram.

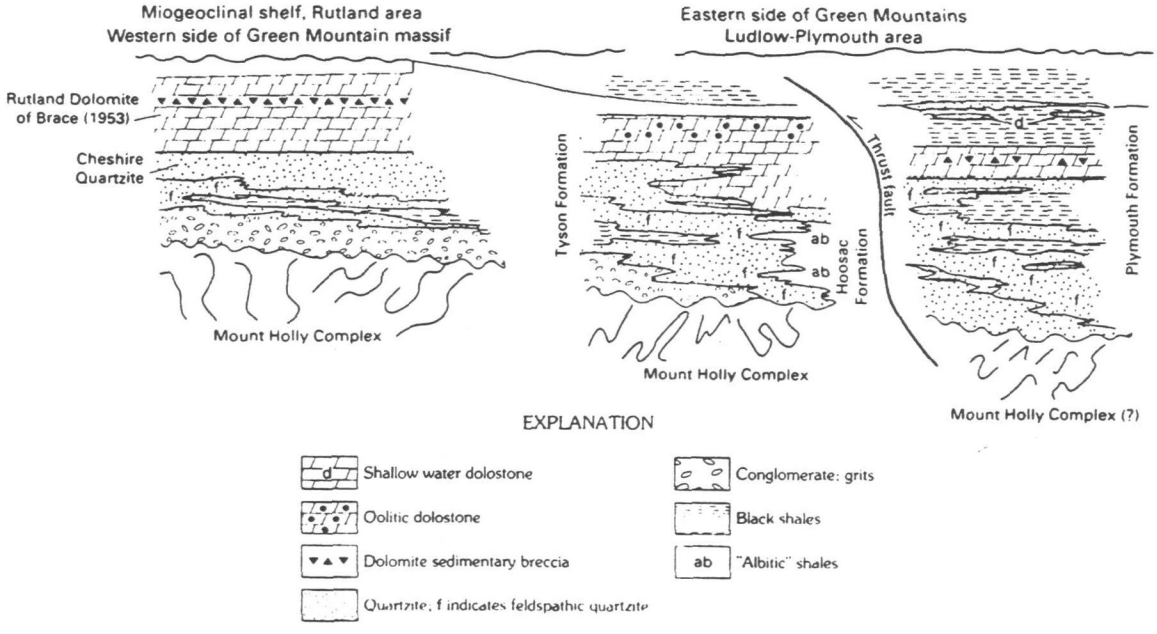


Figure 3. Diagrammatic west-to-east section showing inferred depositional relations between basal autochthonous cover sequence rocks east and west of the Green Mountain massif from Ratcliffe (1994, fig. 1.6). Black shales of the Plymouth Formation are known at least in part to be Middle Ordovician. The "d" denotes projected position of conodont-bearing rock seen at Stop 3. The black phyllite shown transgressing section westward could be equivalent to the exogeosynclinal Ira Formation of the Vermont Valley sequence west of the Green Mountains.

Ratcliffe (1994) accepted the correlation of Thompson, Stanley, and Ratcliffe (1985) pointed out the significance of this correlation, namely that rocks of the Cambrian and Ordovician shelf sequence must have extended this far to the east; the root zone for the Taconic allochthon should lie east of this point. Stanley and Ratcliffe (1985) identified the base of the Pinney Hollow as the present westernmost limit for deposition of the source rocks of the Taconic allochthon.

We have mapped the east side of the Sherburne Valley (northwest of the fossil locality) in great detail, attempting to define the relationship of the fossiliferous dolostone and interlayered black phyllite to the Pinney Hollow and underlying albitic rocks of the Hoosac Formation. The problem is complicated because dark graphitic schists and associated granofels that resemble the black phyllite of Plymouth appear in different structural positions in rocks of the Tyson and Hoosac Formation in the Pico Peak quadrangle within the Black River fault zone (fig. 1). Black carbonaceous schists, possibly correlative with the fossiliferous rock at Stop 3, extend to the south to the southern border of the Ludlow quadrangle where the Black River fault zone and the Wood Peak fault appear to merge (fig. 1). South of this point neither carbonaceous phyllite nor carbonate rocks of the Plymouth are present in the footwall of the Wood Peak fault. At the time of preparation of this guidebook we have not entirely resolved the extent and position of the Middle Ordovician black phyllite.

The contact of the Plymouth Formation with the Pinney Hollow on the Wood Peak fault is, however, everywhere sharp, and well-defined. It is marked by a contact-parallel second generation schistosity (S_2) and reclined folds (F_2) typical of the regional S_2 (F_2) fabric. Carbonaceous schist with dolostone and pods of dolomitic quartzite in irregular and highly folded thin zones are intermixed with more feldspathic gray phyllites, and with zones of tectonically laminated feldspathic quartzite typical of the lower part of the Plymouth. Massive dolostone typical of the Middle Member of the Plymouth is absent. The base of this intermixed zone extends downwards to a prominent zone of tectonically laminated feldspathic quartzite and schistose mylonite that marks the lower contact with albitic rocks of the Hoosac. We do not know the explanation for interlayering of the carbonaceous rocks--it may be entirely tectonic. In any case, the definable belt of black phyllite traceable to the fossiliferous locality only extends 2 km north of the sampled locality in the Pico Peak quadrangle. Faults of the Black River fault zone however extend through the Pico Peak quadrangle to the northwest.

The Middle Ordovician fossils, lithic similarity with and stratigraphic position of the black phyllites of the Plymouth now suggest a possible correlation of this Upper Member with the Ira Formation of the Vermont Valley sequence. It now appears that some or all of the black phyllites of the Upper Member of the Plymouth Formation should be renamed and redefined as Middle Ordovician and be removed from the Plymouth Formation. The lower two members of the Plymouth should be retained as probable Lower Cambrian rocks, although their age could range up into the Lower Ordovician.

Assuming that the black phyllite at the top of the Plymouth is Ira, or nearly Ira equivalent, the sequence from dolostones of the Plymouth up through black phyllite or Ira correlative, to Pinney Hollow in the central syncline at Plymouth duplicates the relationships between the Taconic allochthon and the autochthon west of the Green Mountain massif. That is, the contact between the Pinney Hollow and the black phyllite there may mark the actual sole and root zone of the Taconic allochthon. The Wood Peak fault beneath the main belt of Pinney Hollow is clearly a Taconian F_2 fault and younger than the contact between the Pinney Hollow Formation and black phyllite in the syncline at Plymouth.

Allochthonous rocks of the Pinney Hollow, Ottauquechee and Stowe Formations

The base of the main belt of the Pinney Hollow Formation is the Wood Peak thrust which extends the length of figure 1. Overall the unit is a rather uniform green, ilmenite-chlorite-muscovite-quartz schist that locally contains abundant chloritoid and garnet. Semicontinuous greenstones of basaltic composition are abundant throughout. Limited chemical data indicate they are MORB-like with slight LREE depletion (fig. 5A). Locally belts of more feldspathic, almost gneissic rocks, occur as do restricted belts of dark sulfidic

Table 1. Simplified Description of Map Units for inset maps in roadlog (figures 13 to 17). This list is greatly abbreviated and was taken from the bedrock geologic map of the Plymouth quadrangle by Walsh and Ratcliffe (1994a). The age of the Barnard Gneiss (of Richardson, 1924) is revised from Silurian.

LATE-METAMORPHIC AND POST-METAMORPHIC INTRUSIVE ROCKS

Kd	Lamprophyre and diabase dikes (Cretaceous)
Keb	Epidosite breccia dike (Cretaceous)
Kds	Sausseritized lamprophyre dike (Cretaceous)
Kdsy	Syenitic dike (Cretaceous)
Kdo	Ouachitic breccia dike (Cretaceous)
Dg	Granodiorite (Devonian)

ROCKS OF THE CONNECTICUT VALLEY SEQUENCE

Undifferentiated Northfield and Waits River Formations

DSnw	Gray phyllite (Silurian-Devonian)
DSnwc	Quartzite and conglomerate (Silurian-Devonian)
DSnwl	Siliceous limestone (Silurian-Devonian)

Undifferentiated schist and quartzite

DSs	Schist and quartzite (Silurian?-Devonian?)

ROCKS OF THE TACONIAN ACCRETIONARY PRISM

Barnard Gneiss (Ordovician)

Obd	Hornblende metadiorite dike
Obbf	Biotite metatrondhjemite gneiss
Obhf	Hornblende metatrondhjemite gneiss
Obm	Layered metatrondhjemite, amphibolite, and mafic schist

Cram Hill Formation

Och — Rusty schist (Ordovician?)

Whetstone Hill Member of the Moretown Formation

Omwhb — Black phyllite (Ordovician?)

Omwhbq — Quartzite (Ordovician?)

Omwhg — Biotite-garnet phyllite (Ordovician?)

Omgq — Garnet-quartz granofels (Ordovician?)

Omq — Feldspathic quartzite (Ordovician?) -

Omwhv — Volcanic conglomerate (Ordovician?)

Omwha — Amphibolite and greenstone (Ordovician?)

Moretown Formation

Omgt — Garnetiferous schist (Ordovician)

Omb — Black phyllite (Ordovician)

Omak — Ankeritic greenstone (Ordovician)

Oml — Laminated pinstripe schist (Ordovician)

Oma — Amphibolite (Ordovician)

Omgr — Green phyllite and phyllitic quartzite (Ordovician)

Omgrc — Conglomerate (Ordovician)

Ultramafic Rocks

OZu — Ultramafic rock (Late Proterozoic to Ordovician)

ALLOCHTHONOUS RIFT- THROUGH DRIFT-STAGE ROCKS OF THE LAURENTIAN MARGIN

Stowe Formation

Є sgt — Garnetiferous schist (Cambrian)

Є s — Silvery-green schist (Cambrian)

Є sa — Amphibolite and greenstone (Cambrian)

Ottauquechee Formation

Є ob — Black phyllite (Cambrian)

Є oqb — Black quartzite (Cambrian)

Є oq — Tan quartzite (Cambrian)

Є oc — Carbonate-bearing schists (Cambrian)

Є og — Gray-green feldspathic cross-biotite schist (Cambrian)

Є oa — Greenstone and amphibolite (Cambrian)

Pinney Hollow Formation

ЄZph — Silvery-green schist (Late Proterozoic? to Cambrian)

ЄZpha — Amphibolite and greenstone (Late Proterozoic? to Cambrian)

ЄZphb — Black schist (Late Proterozoic? to Cambrian)

ЄZphc — Chloritoid-quartz granofels (Late Proterozoic? to Cambrian)

AUTOCHTHONOUS COVER ROCKS OF THE LAURENTIAN MARGIN

Tyson Formation

Ctd	Dolostone (Cambrian)
Ctq	Quartzite (Cambrian)
Ctdq	Quartzose dolostone and dolomitic quartzite (Cambrian)
Ctbs	Carbonaceous phyllite (Cambrian)
Cts	Phyllite (Cambrian)
Ctqc	Phyllitic quartzite (Cambrian)
Ctc	Conglomerate (Cambrian)

Plymouth Formation

Cpbs	Phyllite (Cambrian) *Note some or all of the unit is now regarded as Ordovician (see text)
Cpfqd	Dolomitic quartzite and dolostone (Cambrian)
Cpdq	Dolomitic quartzite and carbonaceous dolostone (Cambrian)
Cpbd	Dark blue-gray dolostone (Cambrian)
Cpwd	Dolostone (Cambrian)
Cpdb	Dolomite breccia and conglomerate (Cambrian)
Cpd	Dolostone and dolostone conglomerate (Cambrian)
Cpq	Vitreous quartzite (Cambrian)

Є pfd — Dolomitic quartzite and ferruginous dolomite (Cambrian)

Є pfq — Pinstriped quartzite (Cambrian)

Hoosac Formation

Є Zh — Schist (Late Proterozoic? to Cambrian)

ROCKS OF THE GREEN MOUNTAIN MASSIF AND EASTERN OUTLIERS

Mount Holly Complex

Undifferentiated Rocks

Ymh — Undifferentiated gneiss (Middle Proterozoic)

Ymg — Mixed gneiss (Middle Proterozoic)

Intrusive and Migmatitic Rocks

Ygg — Biotite granite gneiss (Middle Proterozoic)

Yap — Aplite (Middle Proterozoic)

Ymig — Migmatite gneiss (Middle Proterozoic)

Yfg — Felsic magnetite gneiss (Middle Proterozoic)

Metasedimentary Rocks

Ybg — Biotite-quartz-plagioclase gneiss (Middle Proterozoic)

Yq — Quartzite (Middle Proterozoic)

Ycs — Calc-silicate rocks (Middle Proterozoic) -

Ym — Marble (Middle Proterozoic)

Yrs — Rusty aluminous schist (Middle Proterozoic)

schist. Internal units within the Pinney Hollow exhibit a low angle truncation with the bordering Ottauquechee Formation.

Black carbonaceous schists, black and gray quartzite and belts of feldspathic granofels characterize the Ottauquechee Formation at its type section in the Plymouth quadrangle. Internal structure in the Ottauquechee is complex; north to south internal units are truncated against its upper and lower contacts. Greenstones that we have analyzed are LREE depleted MORB basaltic rocks (fig. 5B). Serpentinite and talc-schist occurs within the Ottauquechee at four localities in figure 1.

Rocks of the Stowe Formation are predominantly lustrous green chlorite-muscovite-quartz schists that contain abundant magnetite; more feldspathic varieties contain abundant cross-biotite and locally the rock is abundantly garnetiferous. Ankeritic greenstone and associated talc-carbonate rock occurs in the Stowe Formation at two localities in the Ludlow quadrangle which are too small to show on figure 1. The lower (western) contact of the Stowe is the Acadian Bald Mountain fault in the Plymouth area. To the south that fault merges with an unnamed fault that probably is Taconian. One analyzed greenstone (Fig. 5A) from the Stowe Formation at Stop 11 is LREE enriched and is similar to the alkaline to transitional basalts that occur along with more abundant MORB-type basalts in the Stowe of central Vermont (Coish and others, 1986).

Exotic rocks of the Moretown Formation

The base of the Moretown everywhere we have mapped it is a major fault known as the Townshend thrust (Ratcliffe, 1997; Ratcliffe and Armstrong, 1995, 1996) that truncates stratigraphy and structure in the Moretown. Internal units of the Moretown are very complexly folded and everywhere contain abundant reclined minor and map scale Taconian F_2 folds whose nearly east-west hingelines are arched over the Chester and Athens domes to the east. We interpret the Moretown Formation as part of a Taconian accretionary wedge that contains olistolithic and tectonically incorporated blocks of altered oceanic mantle.

Barnard Gneiss (of Richardson, 1924) in the Plymouth and Ludlow quadrangles

The Barnard Gneiss in the Ludlow area consists largely of plagioclase-rich felsic gneisses, intermediate plagioclase-rich rocks, and less common diabasic or hornblende-plagioclase dikes. Cross cutting relationships are common between the felsic and intermediate rocks, indicating that the bulk of the Barnard here is intrusive. Included screens of hornblende-plagioclase amphibolite, feldspathic granofels, ankeritic greenstone and biotite-rich schist typical of the Moretown Formation occur within the more felsic rocks. The map relations suggest that the Barnard Gneiss forms a southward thinning intrusive sill-like body that crosscuts various units of the Moretown Formation and the contact within the Moretown associated with the Proctorsville belt of ultramafic rocks. Certain feldspathic (dacitic) layers in the Moretown Formation and the ankeritic greenstones in the Moretown (Stop 14) represent volcanic and/or volcaniclastic felsic and mafic rocks similar to rocks which elsewhere intrude the Moretown Formation.

The Barnard Gneiss of the Ludlow quadrangle (Fig. 4) contains three main felsic units. The main unit in the central and northern area consists of very leucocratic massive, white-weathering, magnetite-garnet-biotite-metatrondhjemite (unit Obf), confined to the core of the belt. Numerous exposures show that this rock type, expressed by analyses 7, 8, 9, 10 in table 2, is an intrusive rock that encloses schlieren of more hornblende-rich amphibolite, or pockets of mixed amphibolite and hornblende felsic trondhjemite or tonalitic gneiss (Obh). The magnetite-biotite trondhjemite (Obf) contains euhedral zoned plagioclase, having epidote-rich cores and albitic rims. Relict plagioclase twins indicate a composition of approximately An_{25} for the cores. The average grain size of the plagioclase grains is approximately 2 to 3 mm, but larger glomeroporphyritic clumps of euhedral plagioclase as much as 1 cm in diameter are present. Myrmekite intergrown with the albitic rims is common, as are relict but chloritized igneous biotite, and larger clumps of primary(?) epidote as much as 7 mm in length. The lack of layering and the granitic texture of plagioclase and quartz indicate solidification in situ and a plutonic non-extrusive origin for this facies.

Major oxide compositions

Sample No.	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
Field ID No.	L-10	L-536	Lu-127	L-612	L-493	L-12B	L-12A	L-13	Lu-160	2682	Lu-67
SiO_2	48.1	51.3	57.0	74.3	53.0	50.7	74.2	75.6	77.5	75.6	77.5
TiO_2	1.54	1.75	1.25	0.23	0.17	2.88	0.21	0.21	0.18	0.17	0.13
Al_2O_3	15.3	14.3	14.0	13.2	9.15	13.4	13.2	12.5	12.0	12.8	12.5
Fe_2O_3	2.81	3.054	4.187	1.709	1.710	3.798	0.892	1.728	1.194	1.260	0.841
FeO	7.4	6.79	8.380	1.090	6.110	8.550	1.600	1.10	0.95	1.20	0.620
MnO	0.18	0.18	0.24	0.06	0.18	0.23	0.07	0.08	0.07	0.03	0.03
MgO	7.16	6.45	3.71	0.53	14.80	4.22	0.83	0.54	0.37	0.51	0.24
CaO	11.2	9.4	5.79	3.06	11.50	7.98	1.98	2.11	1.30	1.78	0.92
Na_2O	3.42	4.05	3.05	4.59	1.28	2.58	4.25	4.86	4.76	5.21	6.27
K_2O	0.26	0.32	0.76	0.42	0.13	1.03	1.28	0.33	0.73	0.25	0.19
P_2O_5	0.16	0.21	0.15	0.07	<0.05	0.54	0.05	0.06	0.05	0.13	<0.05
H_2O+	1.5	1.43	1.48	0.41	1.48	1.60	0.71	0.41	0.55	0.56	0.32
H_2O-	0.06	<0.05	0.05	<0.05	0.05	0.05	<0.05	<0.05	<0.05	<0.05	<0.05
CO_2	0.84	0.17	0.04	0.17	0.04	2.20	0.28	<0.01	<0.01	0.15	<0.01
Total	99.99	99.45	100.1	99.45	100.1	99.76	99.6	99.59	99.71	99.50	99.67
CIPW normative composition											
Q	0.0	0.0	13.98	39.11	2.193	12.42	39.56	41.10	43.97	39.86	32.28
Or	1.561	1.930	4.564	2.50	0.7833	6.204	7.653	1.967	4.353	2.09	1.131
Ab	29.40	34.98	26.23	39.13	11.04	22.25	36.38	41.49	40.64	44.48	53.43
An	26.04	20.31	22.63	14.28	19.21	22.36	7.817	10.10	6.114	7.62	4.204
C	0.0	0.0	0.0	0.0	0.0	0.0	2.015	0.4827	1.169	1.09	0.4533
di	19.419	20.027	4.454	0.3896	30.759	0.1776	0.0	0.0	0.0	0.0	0.0
hy	3.975	13.138	21.245	1.4688	33.076	19.015	4.100	1.7991	1.5258	2.23	0.8889
ol	10.186	0.8144	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
mt	4.139	4.52	4.051	2.496	2.469	5.613	1.308	2.501	1.747	1.84	1.228
il	2.971	3.392	2.412	0.4401	0.3292	5.575	0.4035	0.4024	0.3449	0.33	0.2486
ap	0.385	0.5077	0.3610	0.1670	0.1207	1.304	0.1198	0.1434	0.1195	0.12	0.1193
cc	1.94	0.3946	0.0924	0.02291	0.02319	5.10	0.6442	0.02294	0.0295	0.34	0.0229
Total	100.01	100.01	100.02	100.00	100.00	100.02	100.00	100.00	100.01	100.00	100.00

Table 2. Partial chemical analyses of major oxides (in weight percent) and normative-mineral compositions of Barnard Gneiss from the Ludlow Quadrangle, VT

Table 2 (continued).
(1) L-10, very fine-grained, epidote-plagioclase amphibolite, roadcuts east side of South Reading road approximately 4 km north of Proctorsville, 300 m south of powerline crossing; (2) L-536, coarse-grained epidote-actinolite-plagiocase metadioritic dike, sill associated with amphibolite like sample (1) at or near top of the Barnard Gneiss sample contains relict diabasic textured consisting of interlocking well-twinned, subhedral plagioclase laths 3 to 7 mm long; (3) Lu-127, coarse-grained biotite-quartz-hornblende-plagioclase gneiss or schist, containing approximately 40 percent of black metamorphic hornblende grains as much as 1 cm long in decussate structure, sample comes from a small lens 50 m thick within the hornblende-felsic gneiss facies of the Barnard near the western border of the Barnard, 800 m east of Colby Pond. Sample may be a raft of amphibolite from the Moretown Formation or a metamorphosed mafic dike intruding this Barnard Gneiss but rock contains no relict igneous texture or structure; (4) L-612, medium- to fine-grained, white-weathering, hornblende-biotite-epidote-plagioclase trondhjemite gneiss containing well-twinned oligoclase in subhedral grains as much as 5 mm in diameter interpreted as relict hypidiomorphic granular texture; (5) L-493, medium-grained, greenish gray to dark-green, quartz-clinozoisite-plagioclase actinolite rock, consisting of 80 percent pale green, chunky rectangular grains of actinolite, having cloudy cores. Texture suggests replacement of original pyroxene; rock interpreted as a clinopyroxenite or augite-rich wehrlite, perhaps cumulate derived from diorite sill or dike into which rock grades; (6) L-12B, medium-dark gray, fine-grained, biotite-rich, chlorite-magnetite-plagioclase tonalite that cross cuts schlieren and indistinct layering in L-12A; (7) L-12A, medium grained, very leucocratic, muscovite-epidote-biotite trondhjemite gneiss, containing subhedral well-twinned oligoclase 3 to 5 mm in diameter and anhedral quartz grains in hypidiomorphic granular texture from powerline exposure 600 m west of South Reading Road, approximate 4 km north of Proctorsville; (8) L-13, medium- to coarse-grained, very leucocratic, biotite-magnetite-epidote-garnet-plagioclase trondhjemite, containing large 3-7 mm well-twinned grains of plagioclase; (9) Lu-160, medium- to coarse-grained, biotite metatonalite, containing rectangular well-twinned plagioclase 0.7 mm to 1 cm long, and larger glomeroporphyritic clumps of plagioclase as much as 1.5 cm in diameter, host rock that includes or surrounds pyroxenite sample 5, 300 m north of Cavendish-Reading border north of Prospect Hill ; (10) 2682, indistinctly garnet-biotite fine-grained, biotite-epidote-plagioclase trondhjemitic gneiss, lacking distinctive hypidiomorphic texture of samples 4, 6, 7, and 8 from roadcuts on northside of road, at west of hill near The Alps in northeastern corner of the map, rock at sample site may contain inclusions or relicts of several types of igneous rocks, considered typical of the indistinctly mixed variety of felsic Barnard Gneiss; (11) Lu-67, white-weathering, very leucocratic massive medium-grained, biotite ± hornblende metatrondhjemite, containing myrmekitic intergrowths of quartz and plagioclase as well as minor antiperthic plagioclase; sample is representative of the large areas of massive biotite-trondhjemite from the northern outcrop belt of Barnard Gneiss in the map, sample 1.6 km north of the high point in the road passing through the Alps.

The second most prominent facies of the felsic type of Barnard Gneiss is a leucocratic tonalite-trondhjemite (Obh) characterized by small amount, (<5 percent) of hornblende and biotite. This unit also contains abundant (greater than 55%), euhedral to subhedral oligoclase having distinct igneous intergrowths with other plagioclase grains and large 0.7 to 1 cm anhedral areas of quartz. The unit also is a non-layered but foliated rock. Sample 4 in table 2 is typical of the hornblende tonalite. The hornblende tonalite is gradational with the Obf unit and very commonly contains thin amphibolitic layers or schlieren of more hornblende-rich (10 to 15% hornblende) tonalitic gneiss. The hornblende tonalite unit forms two elongate belts within Obf and appears to form more mafic borders around screens of amphibolite or schist.

The leucocratic facies (units Obf and Obh) are weakly- to moderately-well-foliated; the deformation appears to be more intense near the borders of the intrusive rocks. Metamorphic muscovite, chlorite and strings of small epidote and albite form thin, foliated domains 0.5 mm thick, spaced 1-2 cm apart in the rock and impart the gneissic structure to the rocks. Original igneous quartz forms elliptical clots having lengths of 1-2 cm and thickness of 1 to 0.5 cm in the plane of the foliation. The spaced nature of the foliation, the rather low axial

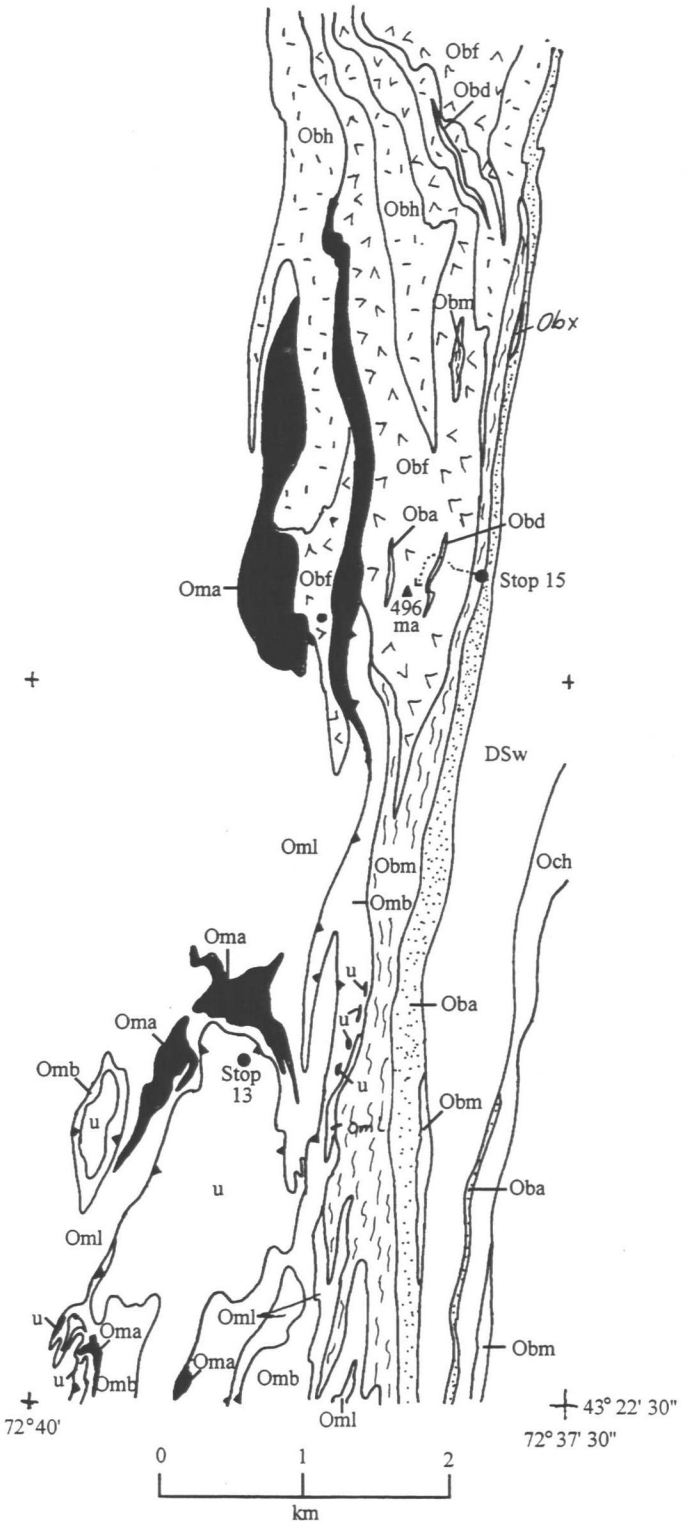


Figure 4. Geologic map showing different facies of igneous rocks in the Barnard Gneiss of Richardson (1924) near dated locality at Stop 15. Petrographic and chemical characteristics of units Obf, Obh, Obm, Oba, Obd, and Obx are discussed in text. For other units see explanation of figure 12.

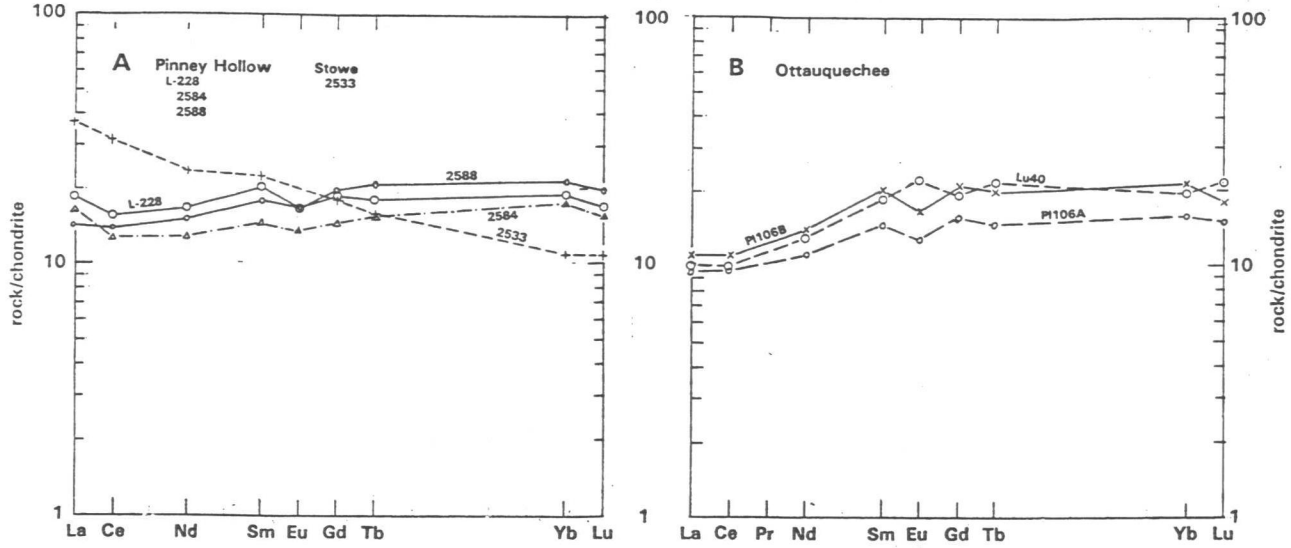


Figure 5. Chondrite normalized rare earth element diagrams of amphibolites from the Pinney Hollow and Stowe Formations (in A) and of the Ottauquechee (in B) in the Plymouth and Ludlow quadrangles. Data normalized to C1 chondrite of Anders and Ebihara (1982) in this and other diagrams following.

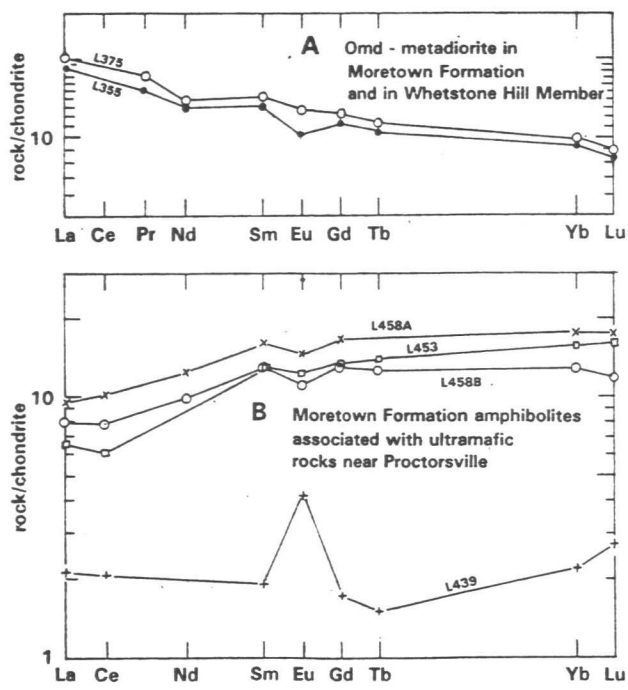


Figure 6. Chondrite normalized rare earth element diagram of mafic rocks of the Moretown Formation in the Ludlow quadrangle (samples discussed in text).

ratios of the deformed quartz clots, and the well preserved igneous intergrowth structures suggest low strain for the bulk of the Barnard within the core of the intrusive.

Along the eastern side of the felsic rocks, and extending to the south, intimately interlayered black- and white-layered gneisses occur. This mixed unit, mapped as Obm (fig. 4), consists of alternating layers of 0.5 to 1 m thick of hornblende-rich (15-20%) plagioclase-quartz gneiss, and black fine-grained amphibolite. Locally the felsic layers are hornblende-free fine-grained biotite trondhjemite resembling the Obf unit. In most exposures, the layers are strictly parallel, except in a few cases where leucocratic layers truncate the more hornblende-rich layers. This well-layered mixed unit passes upwards into more massive amphibolite (Oba) that contains beds of rusty sulfidic schist, coticule, and quartzite. Locally higher layers of the mixed variety (Obm) occur interlayered in the Oba amphibolite near the contact with the overlying Waits River Formation.

A prominent belt of amphibolite (black in fig. 4) occurs between the two masses of felsic gneiss along the projection of the thrust fault between the pinstripe granofels (Oml) and the Omb units of the Moretown. North of the amphibolite the Obh unit is more hornblende-rich and heterogeneous than normal and appears to contain partially digested screens of amphibolite.

Throughout the Barnard Gneiss as well as in adjacent country rocks, mafic dikes of two varieties are common. Those within the Barnard are mapped as Obd (fig. 4). One variety is even-textured, medium-grained, diabasic rock, which occurs as straight-walled features, having chilled borders. The second variety is similar but contains white-weathering, rectangular to ovoidal phenocrysts or deformed ocelli rich in plagioclase in a more hornblende-rich, fine-grained matrix. Both varieties cross cut adjacent rocks and are only weakly foliated. Although commonly metamorphosed to actinolite, epidote, chlorite and plagioclase, some rocks retain a distinct diabasic texture, and either chlorite or actinolite forms rectangular clots replacing an earlier mafic igneous mineral probably clinopyroxene.

Geochemistry of igneous rocks of the eastern Vermont eugeoclinal sequence

We have limited chemical data from rocks in the Ludlow and Plymouth quadrangles. In all 24 samples were analyzed for major, trace and rare-earth elements. A summary of some of that information will be presented. Amphibolites in the Pinney Hollow (3 samples), Ottauquechee (3 samples), and Stowe (1 sample) are basaltic rocks that on chondrite normalized diagrams (fig.5A & B) are MORB-like except for a more alkaline, perhaps transitional basalt, from the Stowe Formation. These amphibolites are similar to rocks we have analyzed from the Rowe Schist in southern Vermont shown in figure 7 and quite similar to rocks from the same belt in central Vermont (Coish and others, 1985, 1986). Metabasalts from the Hoosac Formation from southern Vermont, in contrast, are alkalic to transitional within-plate basalts characteristic of rift basalts. The uppermost basalts within the Hoosac however are LREE depleted and indistinguishable from metabasalts in the Rowe Schist (Ratcliffe, 1997; Ratcliffe and Armstrong, in press) (fig. 8).

Six mafic rocks from the Moretown in these quadrangles (fig. 6) have been analyzed. Two samples of dioritic dike rocks are moderately LREE enriched. Three LREE depleted samples are amphibolites associated with the Proctorsville ultramafic belt near at Stop 13. Sample L439 comes from the outermost mantle of the ultramafic rocks at Stop 13. The three samples of amphibolite (Oma) (see Stop 13) are MORB-like and similar to many amphibolites in the Rowe Schist and the Ottauquechee Formation.

Amphibolites of basaltic composition we have studied in the eastern Vermont eugeoclinal sequence (including the Rowe Schist) are tholeiitic within-plate to N-type MORB. Minor MORB alkalic to transitional types also occur in the Stowe or Moretown (fig. 7).

Major Element Chemistry of the Barnard Gneiss of the Ludlow Quadrangle

Mafic rocks

Four mafic samples of Barnard Gneiss (nos. 1, 2, 3, 5, table 2) have been analyzed. Sample 1 is a typical fine-grained amphibolite from the uppermost parts the Barnard (Oba) that is associated with diabasic segregations and dikes of metadiabase like sample 2. These two samples are probably serially related through fractional crystallization, as indicated by their proximity, and the major oxide data indicating SiO_2, TiO_2, P_2O_5 total alkali increase and decrease in CaO, MgO between samples 1 and 2. Sample 3, a hornblende-rich gneiss or amphibolite, from the base of the Barnard Gneiss is iron and SiO_2 enriched, relative to samples 1 and 2, but contains markedly lower alkalis, TiO_2 than 1 and 2 and is probably not serially related to them. The fourth sample no. 5, an unusual rock, is strongly enriched in MgO (14.8%), and contains low Al_2O_3, has normative quartz and abundant normative pyroxene. Texturally the rock resembles a metapyroxenite or metagabbro as it retains original igneous plagioclase as intergrowths within a framework of coarse pyroxene (now replaced by actinolite). No opaque minerals are present. Sample 5, based on major-element data, may be a clino-pyroxene cumulate, suggesting that clinopyroxene was a fractionating phase in some mafic rocks of the Barnard Gneiss. Based on major element data the four samples of mafic rock from the Barnard Gneiss may be comagmatic but none of the analyzed samples appear to represent the parental liquids from which the mafic rocks themselves evolved.

Felsic rocks

Five felsic samples (no. 4, 6, 7, 8, 9, and 10 in table 2) represent the range of siliceous rocks in the metatrondhjemite-metatonalite facies of the Barnard Gneiss. The rocks range from 50.7% to 77.5% SiO_2, contain low Al_2O_3 (12.0 to 13.4 percent) and are all peraluminous, having molecular $Al_2O_3 > CaO + (Na_2O + K_2O)$. Some samples are moderately corundum normative. Normative An-Ab-Or values indicate trondhjemite and tonalite. Based on SiO_2 variation diagrams, few of the samples appear to be serially related. However, the samples higher in percent normative albite (Ab/Ab+AnX100) have positive correlation with increasing SiO_2, and with decreasing MgO, suggesting fractionation of plagioclase. Textures seen in thin sections also suggest in situ crystallization of plagioclase.

As a group, the felsic rocks have a Peacock alkali-lime-index of approximately 65 to 70% SiO_2, and therefore are calcic, having weight percent $SiO_2 > 61$ where $CaO/Na_2O + K_2O = 1$. However, this intercept is not well constrained because of the lack of samples near unity.

Rare-earth element and trace element data

The 11 samples were analyzed for trace and REE by INAA, Energy Dispersive X-ray Fluorescence and Nb by ICP-AEC. Three mafic Barnard samples (fig. 9) have flat REE chondrite-normalized patterns at 10 to 20x chondrite and exhibit slight positive or negative Eu anomalies. Sample no. 2 (L-536, in table 2) has slightly higher total rare-earth elements and has slightly elevated chondrite-normalized values and may be related to sample 1 (L-10) by fractionation of plagioclase. Sample 3 (Lu-127), however, has a lower chondrite-normalized pattern despite its more differentiated major element chemistry. Although similar in pattern, sample 3 cannot be derived from a parent such as sample 1. Sample 5 (L-493) is extremely depleted and has a chondrite-normalized pattern (fig. 9A) suggestive of a pyroxene cumulate. Pyroxenites or altered pyroxenite may contain very low REE contents (McDonough and Frey, 1989) although most reported clinopyroxene from pyroxenite contain significantly higher REE than sample 5 (Grauch, 1989, p. 152). La/Ta values for samples 1, 2, 3 are greater than 30 and suggest the possibility of crustal contamination.

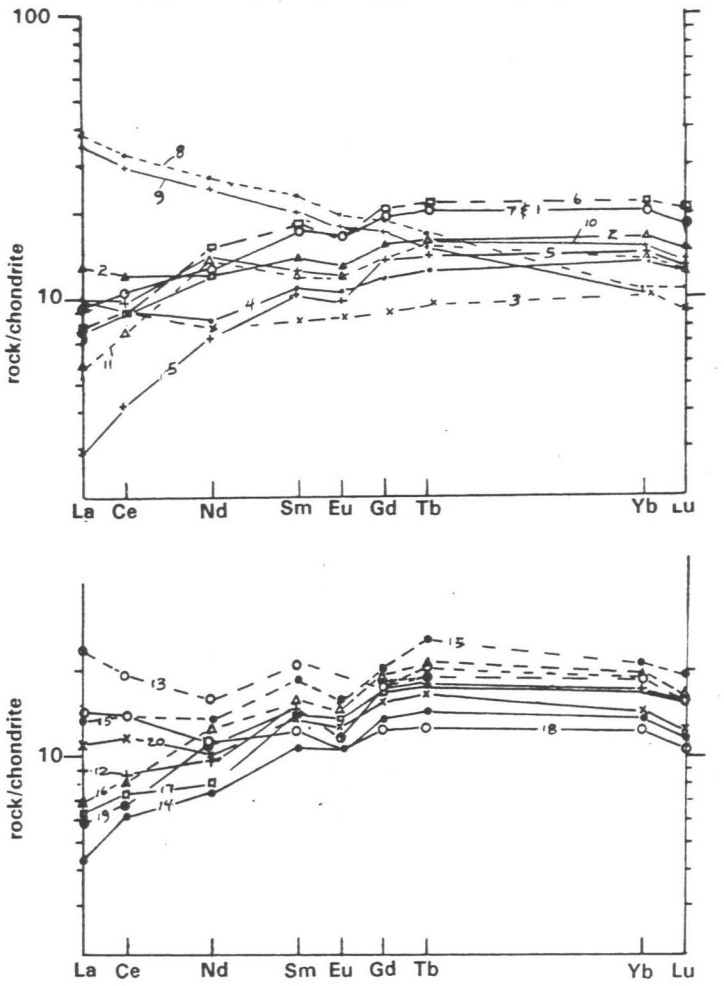


Figure 7. Chondrite normalized rare earth element diagram of amphibolites from the Rowe Schist in the West Dover and Jacksonville quadrangles of southern Vermont; generally equivalent to amphibolites in the Pinney Hollow and Stowe Formations to the north. Data from Ratcliffe and Armstrong (in press); numbers refer to specimens in that report.

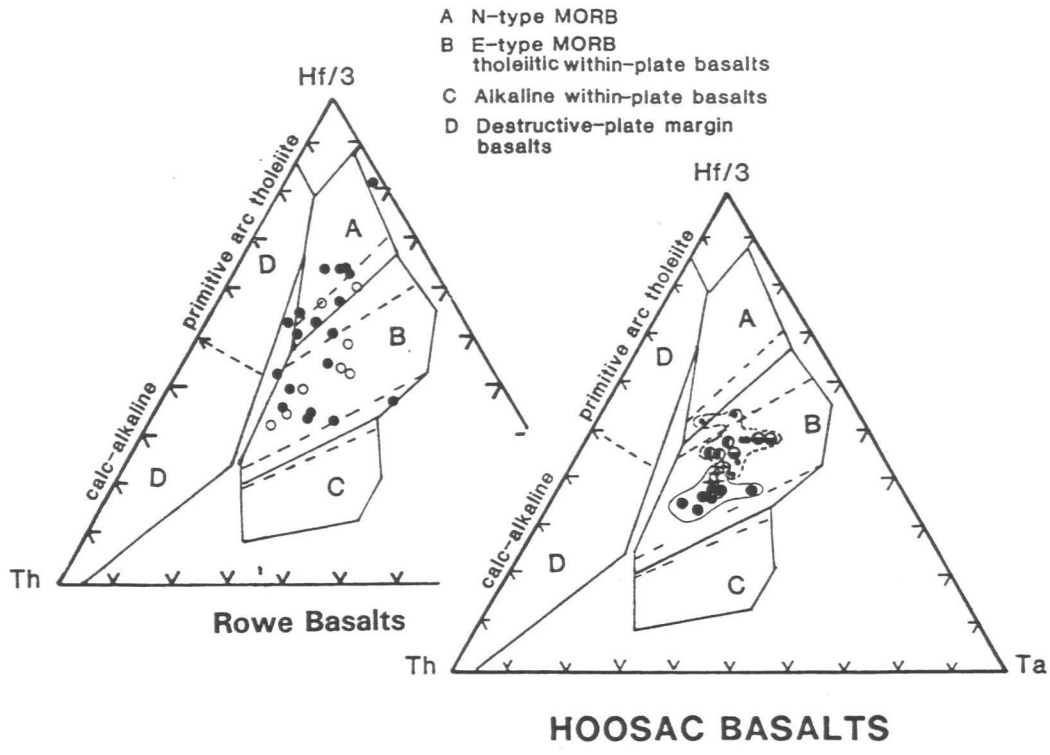


Figure 8. Samples of amphibolites from the Rowe Schist, Pinney Hollow, and Stowe Formations compared to amphibolites from the Hoosac Formation on the tectonic discrimination diagram of Wood (1980). In Hoosac diagram large solid dots are stratigraphically low in the Hoosac, half circle and small dots are from highest units. Open circles in the Rowe diagram come from the Pinney Hollow and Stowe Formations.

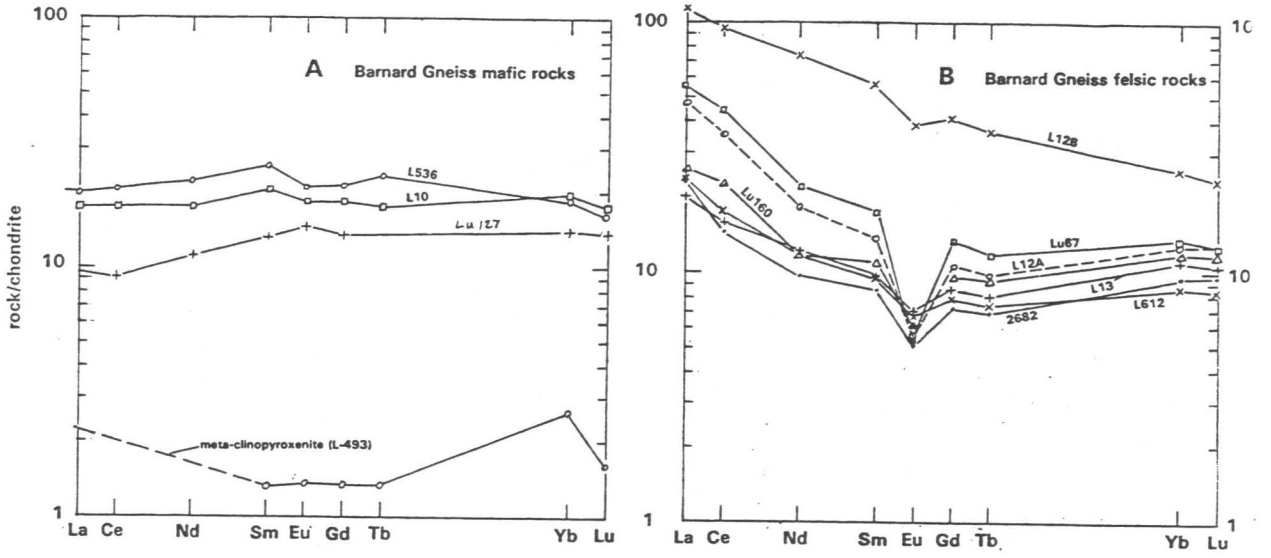


Figure 9. Chondrite normalized rare earth element diagrams of mafic rocks (A) and felsic rocks (B) of the Barnard Gneiss (see table 2 for number used in text discussion).

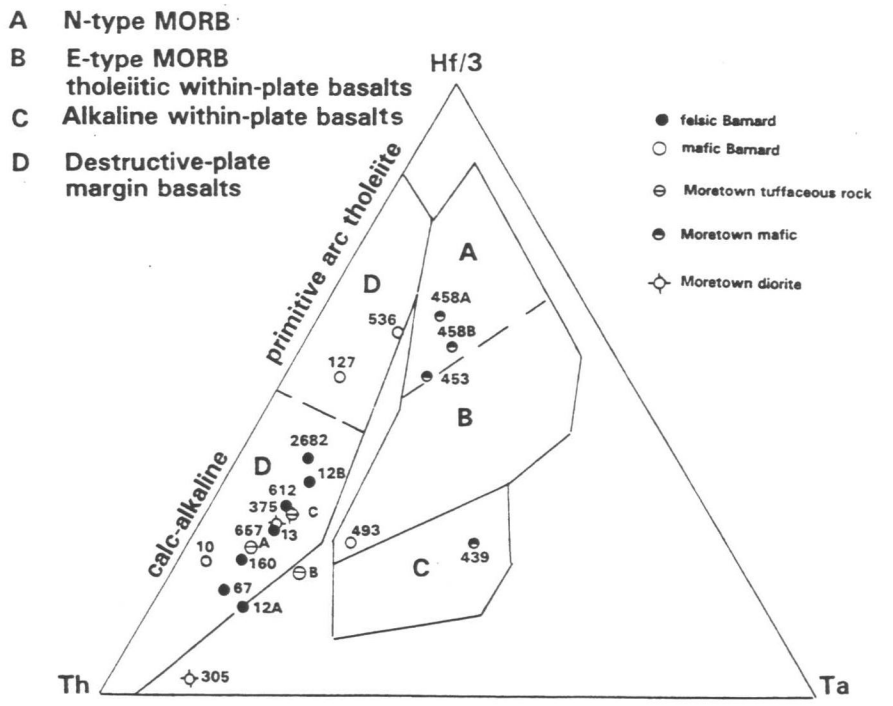


Figure 10. Samples of igneous rocks of the Moretown Formation and Barnard Gneiss on the tectonic discrimination diagram of Wood (1980). Felsic rocks are plotted following the usage of MacDonald and others (fig. 13, 1992).

The felsic Barnard Gneiss samples exhibit, for the most part, very comparable chondrite normalized patterns (fig. 9B). All except no. 6 (L-12B) have heavy rare-earth normalized abundances at $\simeq$10X chondrite and marked light rare-earth element enrichment. La/Sm values range from 3.08 to 5.59. All samples exhibit a moderate to strong negative Eu anomaly. Increased La/Sm values, indicative of increased fractionation of light rare-earth elements have a strong positive correlation with the increasing magnitude of the Eu anomalies, and with increasing percent of normative Ab, suggesting plagioclase fractionation as a dominant control. Samples 4, 9, 11 (L-612, Lu-160, Lu-67 from the tonalite-trondhjemite (4, 9 and 11) have sequentially increasing negative Eu anomalies, lower relative chondrite normalized Eu values and relative increases in the other rare-earths on a chondrite normalized diagram. These samples could have been related by serial fractionation of plagioclase, and perhaps pyroxene. Evidence for fractionation of hornblende is lacking, as increase in the Lu/Sm values is not compensated by decrease of the heavy rare earths.

Sample 6 (L-12B) and other samples are not related as derivatives of the other rocks. Sample 6 has elevated rare-earths, lacks, light-rare earth enrichment relative to heavy-rare earths and the fact that it crosses the patterns of more differentiated rocks such as sample 7 (L-12A) indicates that the felsic units of the Barnard Gneiss are in part a composite of several types of felsic intrusive rock, not related to one another by in situ crystallization and fractionation.

Summary

The geologic relations indicate overlapping intrusions of variously differentiated basaltic, dioritic and trondhjemitic rocks. The complex field relations and rock chemistry suggest multiple intrusions of these different rocks. In chemical characteristics, the tonalites and trondhjemites are low Al_2O_3 and high Yb trondhjemites, generally thought to be oceanic types of trondhjemite (Arth, 1979, p. 125). However, the strong light-rare earth enrichment relative to heavy and middle rare-earth elements, and the strong negative Eu anomalies are unlike high-silica dacites from oceanic areas (Gill and Stork, 1979) or from oceanic dacites (Arth, 1979, fig. 1; MacDonald and others, 1992, figs. 62 and 63).

The Barnard data from the Ludlow area and some selected data from the underlying Moretown plot largely in the destructive plate margin fields on the Hf/3-Th-Ta tectonic discrimination diagram of Wood (1980) (fig. 10). Felsic samples plot as mature island arc rocks. The Barnard rocks plot in the field for mature island arc or for continental margin rocks in the diagram of MacDonald and others (1992, fig. 13) that contains low-K obsidians similar to the composition of the Barnard rocks. Mafic rocks of the Barnard plot in both the primitive-arc tholeiite field and in the calc-alkaline field (fig. 10). Samples L-10 and L-536, judged to be cogenetic, plot in two different fields. Despite these uncertainties, both the mafic and felsic rocks have Th-Hf-Th characteristics that suggest derivation of the Barnard rocks in either a continental or oceanic arc. Nb-Ta values are markedly low for all Barnard Rocks, and this relationship is characteristic of island or volcanic arcs (MacDonald and others, 1992, fig. 17). Felsic rocks of the Barnard Gneiss are similar chemically to metatrondhjemite layers and an intrusive body in the Hawley Formation (Kim and Jacobi, 1991), however, our samples tend to show greater LREE enrichment, but similar overall REE abundances.

Regional Correlation of Barnard Gneiss (of Richardson, 1924)

We used Barnard Gneiss in the original sense of Richardson (1924) to apply largely to intrusive rocks found at the type locality about 15 km north of Proctorsville near Bethel, Vermont. The rocks we have mapped are continuous on the ground with those of the type locality. Doll and others (1961) referred to these rocks as the Barnard Volcanic Member of the Missisquoi Formation. They show the unit as a continuous belt extending from near Bethel, Vermont, around the Proctorsville syncline and Chester dome to the Massachusetts border, where the unit is coextensive with rocks of the Hawley Formation (Stanley and Hatch, 1988). Because new studies suggest that significant parts of the Barnard, as used by Doll and others (1961), and of the Hawley

Formation are intrusive rocks (Walsh and others, 1994; Armstrong, 1994; Ratcliffe and Armstrong, 1995, 1996; Kim and Jacobi, 1996), the nomenclature needs to be revised. Igneous rocks of the old Barnard Volcanic Member of the Missisquoi Formation are referred to as members of the North River Igneous Suite in Vermont, east of the Chester and Athens domes (Armstrong, 1994). These rocks, including some possible volcanic rocks, for example unit Obm on figure 4, west of the domes, which are traceable to the type locality, we refer to as Barnard Gneiss (of Richardson, 1924). Where metasedimentary rocks contain volcanic rocks, the volcanic rocks are referred to as volcanic units bearing the formation name of the including strata. We recognize Barnard-like felsic volcanic rocks within the Moretown and Cram Hill Formations.

The four dated samples discussed below come from the Barnard Gneiss at Proctorsville (L-18) at Stop 15, from a metatonalite of the North River Igneous Suite (SR305), a metavolcanic layer in the Cram Hill Formation (SP450) and from a metatrondhjemite dike of the North River Igneous Suite near South Newfane, Vermont. Samples L18 and SR305 and SP450 are similar to intrusive rocks of the Hawley Formation described by Kim and Jacobi (1991). The dated rocks all come from intrusive and extrusive felsic rocks in the outcrop belt of the Barnard Volcanic Member of the Missisquoi Formation as previously shown by Doll and others (1961).

Aleinikoff and Karabinos (1990) report a 418$\pm$1 Ma U-Pb zircon concordia intercept age for a felsic layer in Barnard Volcanic Member of the Missisquoi Formation near West Bridgewater in the Woodstock South quadrangle. The rock sampled occurs in the upper third of the Barnard Gneiss as we map it, and resembles the mixed felsic-mafic unit we map as unit Obm. The rock is also very similar to layered metavolcanics in the Cram Hill Formation at our sample site SP450 discussed below.

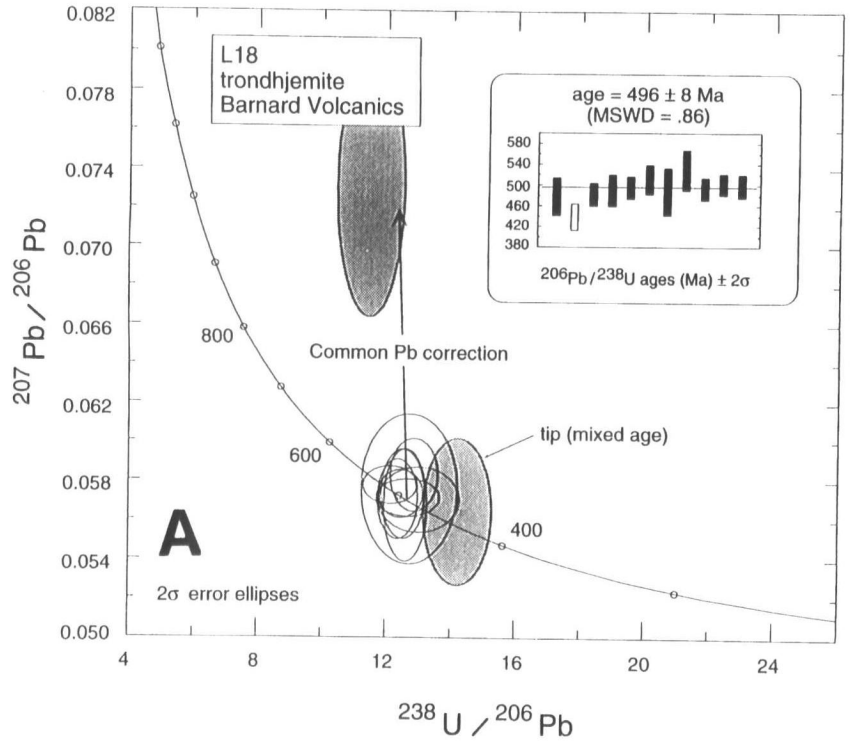
Although we do not know how to interpret this 418Ma, or Late Silurian age, it does not agree with all subsequent data from the Barnard Gneiss and related rocks. At West Bridgewater, the felsic layer sampled is about .0-3 m thick and chemically is identical to trondhjemitic gneiss of our Barnard Gneiss in major and REE abundances, and contains very low Zr (35-50 ppm). Near the top of the sampled layer, a narrow 15-cm-thick more quartz-rich irregular felsic layer contains 900-1000 ppm Zr. This narrow zone has very elevated REE abundances, about 100 times that of the main low zirconian layer and is enriched in Na_2O and SiO_2 relative to the main layer. We speculate that this may be a mobilized vein generated at a later time than the formation of the dated trondhjemitic layer and that this layer contributed the zircon analyzed. We, therefore, think that the 418$\pm$1 Ma age does not date the time of crystallization of the trondhjemitic rock of the Barnard Gneiss.

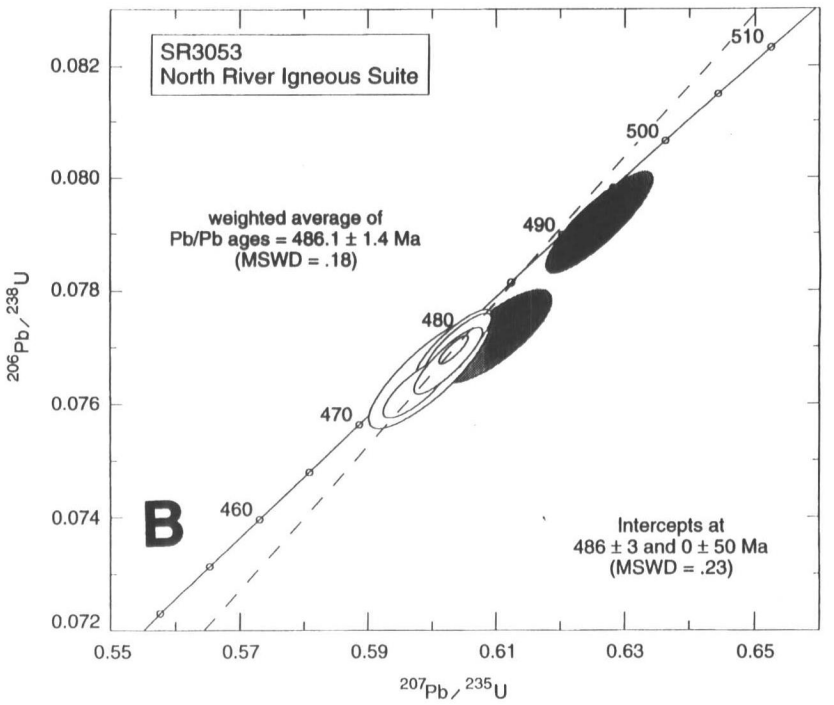
Geochronology

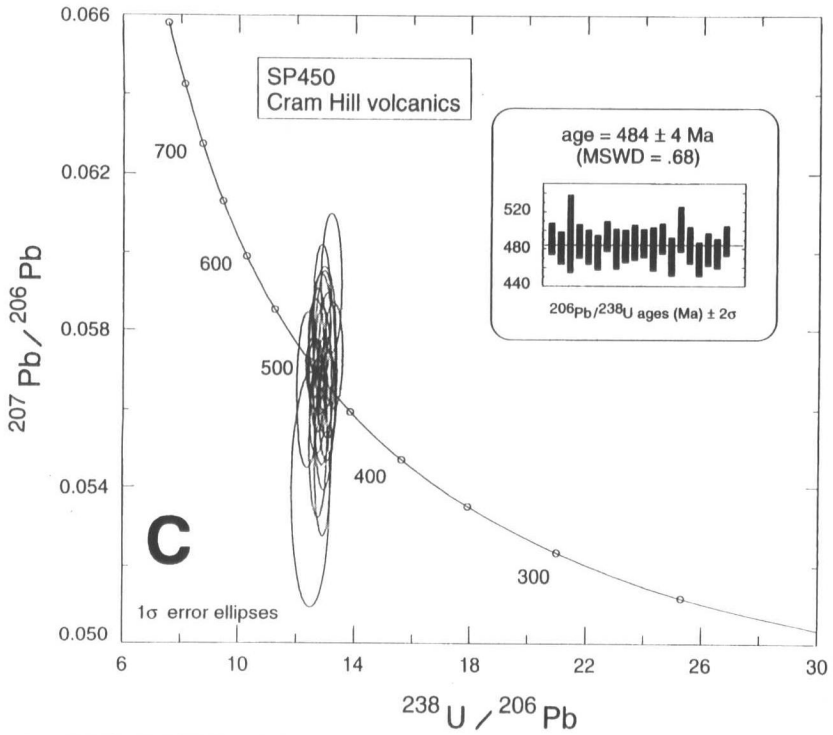
Four samples (fig. 11 and table 3) of Early to Middle Ordovician igneous rocks were dated by zircon U-Pb methods (either conventional or ion microprobe). Sample locations and dates are shown on Figure 1, except for VT1BR2-89, which is located south of the map border on the eastern side of the Athens dome. The SHRIMP ion microprobe was used when we suspected, on the basis of crystal morphology and color, that the zircons contained multiple age components. Data from SHRIMP analyses of three samples are displayed on Tera-Wasserburg plots (fig. 11A, C, and D); the calculated age is the weighted average of $^{206}Pb/^{238}U$ ages. Conventional analyses of zircon from one sample are plotted on a standard concordia plot where the calculated age is the upper concordia intercept (fig. 11B).

Sample L18, a medium-grained metatrondhjemite (Obf) from the Barnard Gneiss (of Richardson, 1924) at Stop 15, is the oldest dated Paleozoic rock in the study. The age of 496$\pm$8 Ma is based on averaging data from 10 grains (fig. 11A). One analysis of about 435 Ma, from near the tip of a grain, probably is a mixture of components from the igneous core and a metamorphic overgrowth.

Zircons from sample SR3053 comes from a coarse-grained metatonalite of the North River Igneous Suite that intrudes the Moretown Formation on the east side of the Chester dome (see Stop 13-Trip B-6) are euhedral, prismatic, and light tan. Because of the relatively simple zircon morphology, this rock was dated by







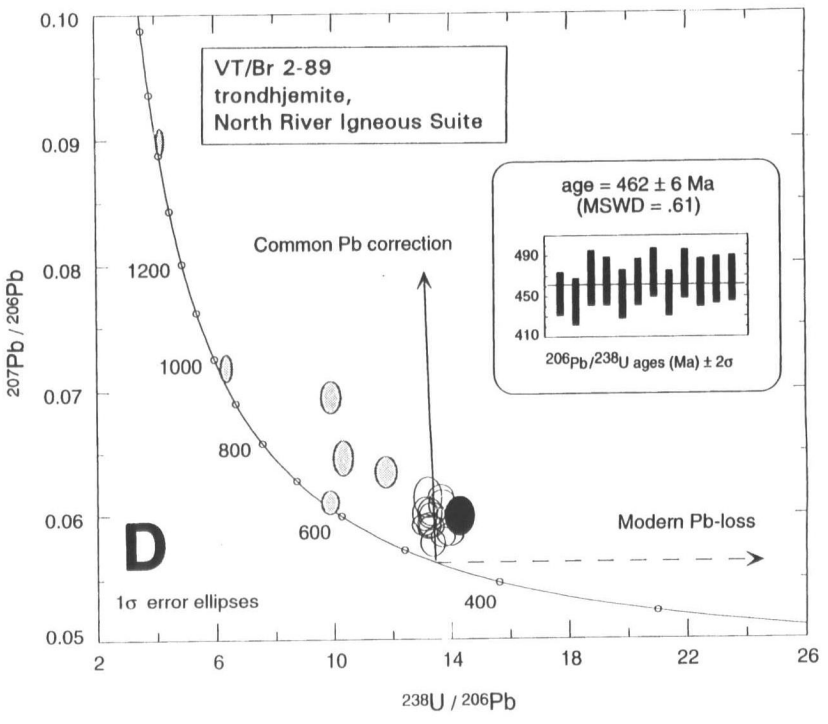


Figure 11. U-Pb geochronology of felsic igneous rocks of the Barnard Gneiss, North River Igneous Suite, Cram Hill volcanics discussed in text and shown in table 3.

Table 3. U-Pb geochronology of Paleozoic rocks in southern Vermont discussed on Trips B-6 and C-6 ages determined by J.N. Aleinikoff

Sample No.	Rock	Age (Ma)	Comments
VT/Pl 2-91	2-mica granite intruding Pinney Hollow Formation in Plymouth quadrangle has xenocrystic Middle Proterozoic zircon	~370	conventional and SHRIMP data [only 1 grain has Acadian age], cross cuts Taconian structure, has Acadian F_3 crenulation
VT/Br 1-89	Black Mountain granite in core of Guilford dome	373 ± 4	conventional age, foliated granite
Vt/Br 2-89	Trondhjemite of North River Igneous Suite South Newfane	462 ± 6	SHRIMP age, intrudes rocks of Cram Hill and Moretown Formations
SP450	Cram Hill felsic volcanics of Cram Hill Formation near Springfield	484 ± 4	SHRIMP age, layered with mafic volcanics and schist of the Cram Hill Formation
SR3053	tonalite, North River Igneous Suite (of Armstrong, 1994) east side of Chester dome Saxtons River quadrangle	486 ± 3	conventional age, intrudes Moretown Formation
L18	Trondhjemite of Barnard Gneiss of Richardson (1924) Ludlow quadrangle	496 ± 8	SHRIMP age, intrudes Moretown Formation and Proctorsville ultramafic belt and possibly rocks of the Cram Hill Formation

conventional U-Pb techniques (fig. 11D). Six fractions of zircon yielded nearly concordant data. A standard best-fit calculation is not appropriate for this data set because there is no spread amongst the individual data points. Forcing the calculated discordia through a reasonable lower intercept of 0±50 Ma results in an upper intercept age of 486±3 Ma. Alternatively, the weighted average of the $^{207}Pb/^{206}Pb$ ages yields an age of 486±1 Ma (fig. 11B). Two additional fractions have slightly older $^{207}Pb/^{206}Pb$ ages of 503 and 504 Ma, suggestive of mixing a small amount of inherited Proterozoic material with Ordovician zircon.

Sample SP450 (fig. 11C) is from the east side of the Chester dome in the Springfield quadrangle and comes from a thin (1-m-thick) felsic layered within a layer sequence of amphibolite, and felsic metavolcanics 50 m thick. The mixed felsic and mafic unit is very well-layered and contains layers of black schist, coticule and arkeritic greenstone. From the bedded nature of the rocks and the interbedded black schist, we have assigned

this rock to the Cram Hill Formation. The sample site is within 100 m horizontally of the base of the Waits River Formation. We think the rock sampled is a metavolcanic rock that dates the age of sedimentation of pelitic rocks and interbedded volcanics at the top of the eastern Vermont eugeoclinal sequence. Isotopic data from sample SP450 are relatively simple (fig. 11C). Twenty data points were averaged to yield an age of 484±4 Ma. Inherited radiogenic Pb was not found in any analyzed grains. The rock dated is very similar to the West Bridgewater discussed above and to the Obm unit shown on figure 4.

Sample VT/Br2-89 (fig. 11D) comes from a metatrondhjemite sill of the North River Igneous Suite near South Newfane on the east side of the Athens dome. Isotopic data from sample VT/Br 2-89 are complex; these zircons contain multiple ages of inheritance, plus an igneous population. Twelve analyses were averaged to yield a crystallization age of 462±6 Ma (one analysis was excluded from the calculation because the isotopic systematics suggest that it has lost some radiogenic Pb recently). Six other analyses have ages ranging from about 530-1400 Ma. Ages in the range of about 530-620 Ma may reflect analysis across age zones (i.e. a mixture of Proterozoic and Ordovician ages). Two older ages of about 960 and 1400 Ma probably represent inheritance because these are ages of rocks in the Laurentian margin. In summary, arc rocks in eastern Vermont apparently formed throughout most of the Early to Middle Ordovician, from 496±8 to 462±6 Ma.

Proctorsville ultramafic belt

One of the most extensive belts of ultramafic rocks (serpentinite, talc schist and talc-carbonate rocks) in western New England occurs in the vicinity of Proctorsville (fig. 12). Large to small masses of ultramafic rock extend intermittently for about 30 km south of Proctorsville to Townshend Reservoir (Ratcliffe, 1997). Along this entire distance the ultramafic rocks are interspersed at or near the contact between two major units of the Moretown Formation. Well-laminated pinstriped biotite granofels (Oml unit) tends to occur to the west beneath the ultramafic rocks, and dark gray carbonaceous sulfidic biotite schist (Omb unit) occurs to the east. The ultramafic rocks in many cases have a thin selvage of the black schist encasing them. Although the serpentinites are highly foliated and affected strongly by Acadian faulting, the ultramafic rocks and schist share much older complex folds that probably are Taconian. Hingelines of the refolds plunge very steeply thus producing quite irregular and almost chaotic map patterns. The complex highly refolded nature of the contact between ultramafic and enclosing rocks makes the talc-rich deposits in the keels of folds very difficult to project from surface mapping alone.

Massive to well-bedded amphibolites and ankeritic greenstones occur near the ultramafic rocks. They are encased in either the granofels of the Moretown (Oml) or the darker schist (Omb). The association of the ultramafic rocks at or near the contact between the two varieties of Moretown is a remarkably consistent and suggests localization along a fault. The fault, however, is not expressed by any clear cut fault fabric and the fault, if present at all, could well be premetamorphic. Nevertheless we show a fault along this contact, and portray the ultramafic rocks as fault slivers (fig. 12). We cannot rule out the alternative interpretation that the ultramafic blocks are olistoliths deposited with an euxinic part of the Moretown depositional basin. These blocks, however, would have had to been very large, as single masses at present are greater than 2 km long.

Igneous rocks of the Proctorsville ultramafic belt

Rare, foliated mafic to intermediate dikes, centimeters to several meters thick intrude the Moretown Formation. There are three varieties of mafic dikes. The most widespread are plagioclase-rich hornblende amphibolites that have chill borders and a distinct dioritic appearance; however, much of this texture may be metamorphic. Less abundant are fine grained biotite-quartz-plagioclase rocks or black medium-grained amphibolites that contain distinct white plagioclase phenocrysts.

Layered felsic rocks and amphibolite occur interbedded in the Moretown in both the pinstriped granofels and in the black schist units (fig. 12). The frequency of felsic layers increases upwards within the

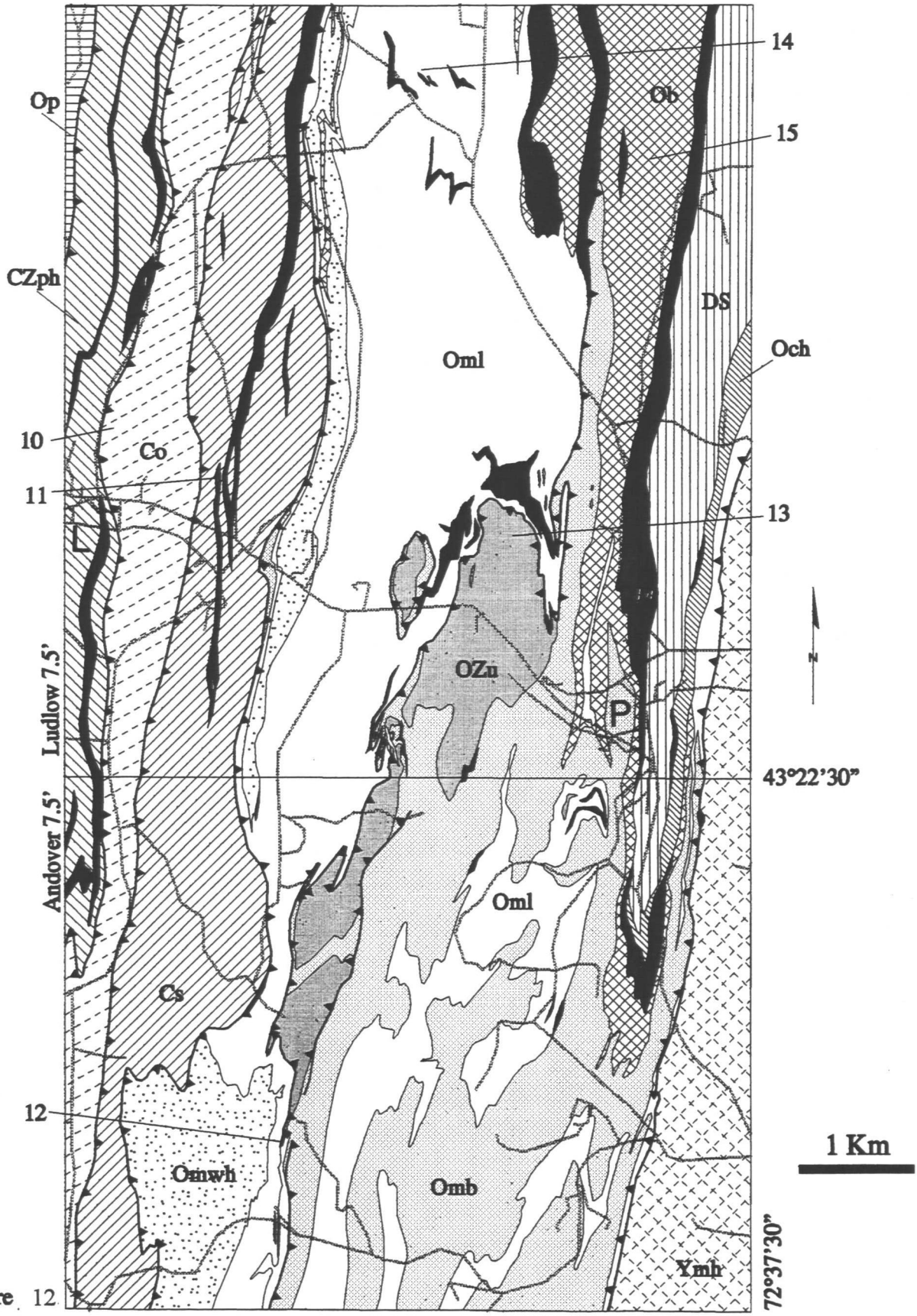

Op
CZph
10
11
12
14
15
13
Ob
DS
Och
Oml
Co
OZu
P
Cs
Omwh
Omb
Ymh
Ludlow 7.5'
Andover 7.5'
43°22'30"
72°37'30"
1 Km
N

Figure 12

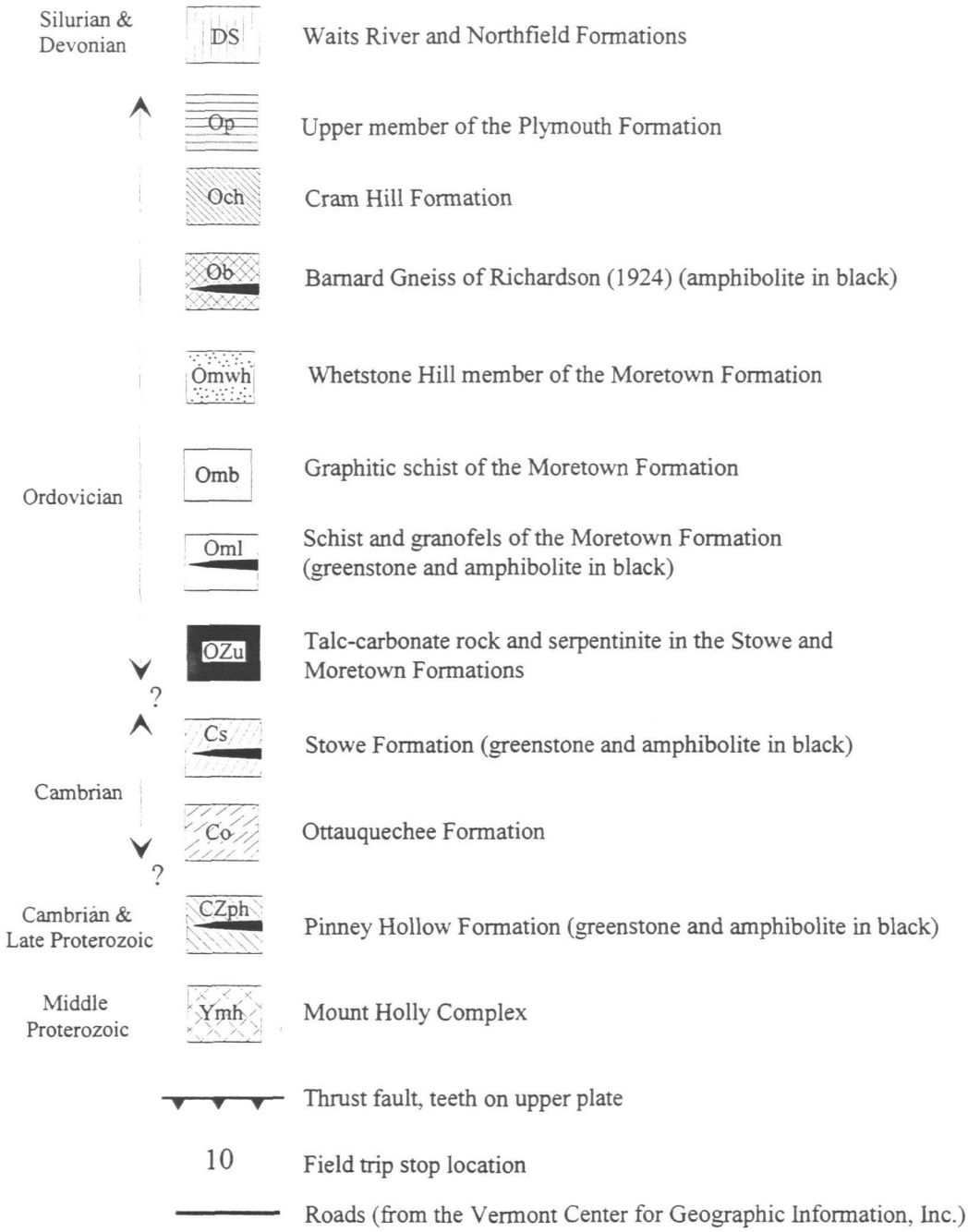


Figure 12. Generalized geologic map of the southeastern part of the Ludlow quadrangle and northeastern part of the Andover quadrangle. Geology simplified from Walsh and others (1994) and Ratcliffe (1996a, b).

black schist unit and in the Cram Hill Formation as the main mass of the Barnard Gneiss (of Richardson, 1924) is approached. Felsic layers and mixed felsic and mafic layers of the Barnard Gneiss are in depositional and (or) intrusive contact with upper parts of the black schist unit of the Moretown Formation. Felsic, mafic and intermediate igneous rock, both volcanic and intrusive, are common withIN the Moretown package of rocks that constitutes the host for the Proctorsville ultramafics. The abundance of these igneous rocks increases upwards toward the overlying Cram Hill Formation (fig. 12). North of Proctorsville the contact between the Oml and Omb units of the Moretown pass into the large mass of felsic Barnard Gneiss (of Richardson, 1924), that appears to intrude across the Proctorsville ultramafic belt.

Interpretation of the origin of the Proctorsville ultramafic belt and the Moretown lithotectonic unit

We interpret the Moretown lithotectonic unit as an allochthonous accretionary complex produced on the upper plate of a Taconian oceanic subduction zone following Stanley and Ratcliffe (1985). The abundance of arc-like bimodal volcanic rocks bedded within the sequence is evidence for a nearby volcanic-arc source that produced dacitic to andesitic volcanic rocks similar in composition to the trondhjemites of the Barnard Gneiss (of Richardson, 1924) that later intruded the preassembled accretionary package. The new geochronologic data indicates that the age of the felsic to tonalitic volcanism and plutonism is remarkably old. If our interpretation of the geologic map relations is correct, the assemblage of the ultramafic rocks within the Moretown Formation predated 496 ± 8 Ma. Similar volcanism continued during deposition of Cram Hill Formation until 484 ± 4 Ma. Intrusion of additional trondhjemite into this package apparently continued until 462 ± 6 Ma based on the age of the dike or sill that intrudes the Moretown and Cram Hill Formation near South Newfane (VT/Br 2.89 in table 3).

The presence of intrusive volcanic-arc igneous rocks within the accretionary prism or fore-arc setting is not normally expected. We speculate that plate geometry changed following the formation of the accretionary wedge, perhaps by "westward" jumping of the subduction zone to a new position closer to the Laurentian plate. This new consuming plate margin then become the active locus of closure and the one figured in Stanley and Ratcliffe (1985) which lead to the final assembly of the other lithotectonic units that make up the allochthonous units of the eastern Vermont eugeoclinal sequence.

Based on the probable Lower Medial Ordovician age of the conodonts at the top of the autochthonous cover rocks at Plymouth (Anita Harris, personal communication to Ratcliffe, 1997). The final closure and resultant Taconian dynamothermal events occurred between Middle Caradocian (about 460 Ma) and about 445 Ma (Ashgillian) the time of regional Taconian (M_2) Barrovian metamorphism in western New England (Hames and others, 1991), using the revised Lower Paleozoic times scale of Tucker and McKerrow (1995).

It is interesting that an age of 496 ± 8 Ma for the Barnard Gneiss of Richardson (1924) near Proctorsville or lower Tremadocian, is about the same age as the medium-high pressure M_1 Taconian metamorphism first suggested to be about 490 ± 8 Ma from a total fusion $^{40}Ar/^{39}Ar$ age on barroisite from Belvidere Mountain in northern Vermont (Laird and others, 1984). The intrusive activity in the Barnard Gneiss (of Richardson, 1924) and in the Cram Hill Formation, which we place between 496 ± 8 and 484 ± 4 Ma and perhaps ranging to as young as 462 ± 6 Ma, overlaps with the estimated time of M_1 Taconian metamorphism in northern Vermont.

Conodonts of late Arenig to early Llanvirn age in the Melburne Formation at the top of the Oak Hill Group in southern Quebec (Marquis and Nowlan, 1991) and the conodonts from the black phyllite at Stop 3 are younger than the proposed medium-high-pressure metamorphism of Laird and others (1984). These data support the contention of Stanley and Ratcliffe (1985) that the older metamorphism in northern Vermont formed in an outboard rather than continental-margin setting and was transported later to the west on thrusts such as the Underhill thrust.

In some fashion the subduction zone metamorphism in northern Vermont and the volcanic-arc activity responsible for igneous rocks of the Barnard Gneiss (of Richardson, 1924) are closely related in time and must reflect different aspects of the paleotectonic setting in the Lower Ordovician.

In recent years there has been growing recognition that arc volcanism preserved in western New England spans a much longer period of time than previously appreciated. The Barnard Gneiss (of Richardson, 1924), volcanics in the Cram Hill Formation and North River Igneous Suite have ages between 496 ± 8 and 462 ± 6 Ma, that are roughly comparable to the 488 and 474 Ma ages of the Collinsville Formation from the core of the Shelburne Falls dome (Karabinos and others, 1996). These ages are in total agreement with the proposition of Stanley and Ratcliffe (1985) that collision with an island arc or multiple arcs in Late Middle to Late Ordovician produced the Taconic orogeny. Stanley and Ratcliffe (1985) identified the Bronson Hill arc as the culprit, not knowing any better at the time. They were taken roundly to task by Tucker and Robinson (1990) who showed that volcanic rocks of the Bronson Hill arc only range back to the final culminating phase of the Taconic orogeny ($\sim$453 to 449 Ma) and do not coincide in time with the proposed Cambrian through middle Ordovician subduction event proposed by Stanley and Ratcliffe. We see in the evidence from the Barnard Gneiss at Proctorsville and from the data of Karabinos and others (1996) that arc activity was ongoing over much of the Ordovician. The collision with a large island arc built on a continental substrate, still pumping out volcanic rocks at 449 Ma (Tucker and Robinson, 1990), is appropriate in time and scale to produce the tectonic overburden and inertia to cause the extensive amount of imbricate thrusting of Laurentian crust and high-grade Barrovian metamorphism in western Massachusetts at about 445 Ma (Hames and others, 1991). One thing we can be certain of is that the preserved arc rocks of the Barnard Gneiss, as exposed, cannot conceivably be the tectonic cover responsible for the tectonic overburden, which caused the Taconian orogeny. These rocks probably only represent flotsam accreted prepartory to the real Taconic collision for which the Bronson Hill arc is still very much a candidate.

TRIP C-6
Road Log
Sunday, September 21

Assembly point:
Pico ski area parking lot, Route 4, 0.5 mile west of Deer Leap Mountain (Pico Peak Quadrangle)
Time: 7:30 am

Cumulative mileage

0.0 Ski Shack Intersection Routes 100 and 4, Killington, Vermont

Large cuts behind building are in Middle Proterozoic granitic gneisses of the Mount Holly Complex; black layers 3-5 m thick of chlorite-biotite-albite-epidote schist and granofels are screens of country rocks in the gneiss. Both the gneiss and the biotite rocks are strongly foliated, and retrograded. A strong retrogressive Taconic foliation and mylonitic structure is folded into rather broad north-plunging folds.

Fractures, veins of quartz and ankerite, chloritic shear surfaces and zones of breccia crisscross the rock. These late faults and fractures are of unknown age; they may be Acadian (indeed microcline-epidote-quartz does cement some faults near here), or they may be Cretaceous and related to a regional system of oblique-slip normal and strike-slip faults that appear to have accompanied intrusion of abundant (largely mafic) alkaline dikes of the White Mountain Intrusive Suite. Our trip will not dwell on the Middle Proterozoic rocks, so examine them before we depart.

Head east (downhill) on Routes 100 and 4.
1.9 Park along road before church at Sherburne Center (Pico Peak Quad)

STOP 1. MOUNT HOLLY CONTACT WITH TYSON FORMATION (contact is exposed at third restricted parking sign uphill from the church). Uphill 300 feet are abundant pegmatites in biotite-rich plagioclase gneiss of the Mount Holly Complex. Light-green vitreous, muscovite quartzite of the Mount Holly (Yq) is interlayered in the gneiss to the south. The contact between quartzite and chlorite-muscovite-albite-quartz granofels of the Tyson is sharp and most undistinguished. Note the complexity of folds in the quartzite despite the general conformity of foliation; both rocks were obviously deformed together after deposition, producing an apparent conformity. Telltale signs of the original angular unconformity still exist; the compositional rodding in the Y rocks reflècts the intersection of the unconformity-parallel foliation and compositional layers in the gneiss. Another is the large cm-sized muscovite plates in the quartzite that are absent in the granofels.

The lowest beds of the Tyson are often nondescript albitic granofels like this, although spectacular boulder-, cobble-, and pebble-conglomerate intermittently occur, commonly 10 to 50 m above the contact. Overall, the basal unit is an albitic granofels like this or a olive-tan to gray well-laminated quartz-grit, and gritty quartzite.

Despite the high angularity of the unconformity as seen in the map relations, most exposures are like this one and appear to show conformity, not a happy situation for those seeking easy recognition of an unconformity representing a hiatus of about 400 million years!

Go south on Routes 100 and 4.
Just after the first intersection, gneiss of the Mount Holly returns.
2.7 Restaurant on right, just beyond Tyson touches down for 60 m, then Mount Holly continues.

3.2 East limb of a syncline of Tyson meets the road, above this point at 1500' is a dipslope exposure of some of the best deformed cobble conglomerate of the Tyson. The elongation direction of the cobbles is N80° E, and lies within the AC plane of the dominant foliation. This elongation direction is also the plunge direction of the reclined folds in the cover sequence rocks, and of the mullions on the exposed thrust surfaces in the cover rocks. We think this is the Taconic S_2 (F_2) deformation plan.

4.2 Ledges of quartzite and dolostone of the Tyson cut by a marked quartz vein. This quartzite and dolostone is the first dolostone above the albitic grits and granofels and underlies the black carbonaceous phyllite seen at the next stop.

Leave Pico Peak and enter Killington Peak Quadrangle

5.3 Bear Mountain Road and the Killington Sky Ship lift, park on left side of road near foot bridge over Ottauquechee River.

STOP 2A. BLACK PHYLLITE OF THE TYSON, DOLOSTONE OF THE TYSON AND CONTACT WITH THE ALBITIC HOOSAC FORMATION (KILLINGTON PEAK QUADRANGLE) (fig. 13). Walk first across Rte. 100 and examine ledges of carbonaceous phyllite of the Tyson exposed north of the lift. Excellent exposures of this rock may be seen in large roadcuts 0.3 mi up Bear Mountain Road, where similar dolostone and quartzite are exposed at mileage 4.2 underlying the black phyllite.

This black phyllite is mappable along the west side of Ottauquechee River valley to a point 0.4 mi north of Shaw Brook in the Pico Peak quadrangle. To the south, this phyllite can be mapped to the slopes just north of optional Stop 5 where it is cut out by a thrust fault in the Tyson.

Walk east across footbridge to ledges of dolostone and exposed contact of dolostone of the Tyson with overlying albitic granofels and schist of the Hoosac Formation.

STOP 2B. THRUST FAULT OF HOOSAC AND TYSON. This outcrop and others like it at Plymouth Union have been cited for evidence of the conformity and stratigraphic succession of the Hoosac above the Tyson (Thompson, 1972). Magnetite concentrated near the contact here and at the Plymouth Union mine were attributed to pre-Hoosac erosion and terra rosa developed on the dolostone. Our interpretation of this contact is quite different; we map it as a fault.

Examine the planar contact, marked by a lineated vein of chlorite-quartz and magnetite that extends for 50 meters along the base of the cliff. Near the contact the dolostone is a well-foliated highly-disarticulated rock in which irregular knots of coarser grained dolostone are surrounded by a finer-grained more strongly foliated dolostone. Note the structure in the overlying granofels, the numerous noses of isoclinal folds of quartz and the absence of any feature that could reasonably be called bedding.

The hingelines of the minor folds of quartz knots in the dolostone and in the granofels plunge down the dip of the foliation. This is the regional Taconic F_2 structure present in all of the cover rocks near the eastern cover of the Green Mountain massif in Vermont and all along the east and western side of the Berkshire massif in Massachusetts. One of us (NMR) has mapped this structure from southern Massachusetts where this structure is dated by $^{40}Ar/^{39}Ar$ dating of hornblende and biotite at staurolite grade to be approximately 455 to 445 Ma (see Sutter and others, 1985; and revision in Hames and others, 1991). Walsh has mapped the F_2 structures northward to the Lincoln massif.

Preliminary $^{40}Ar/^{39}Ar$ data (Burton and others, 1990) from the Devils Den area in Danby quadrangle suggest that cooling ages of muscovite in the regional (F_2) foliation in the core of the Green Mountain massif also is older than Acadian. Based on these data, we regard the F_2 (S_2) structure age as Taconian.

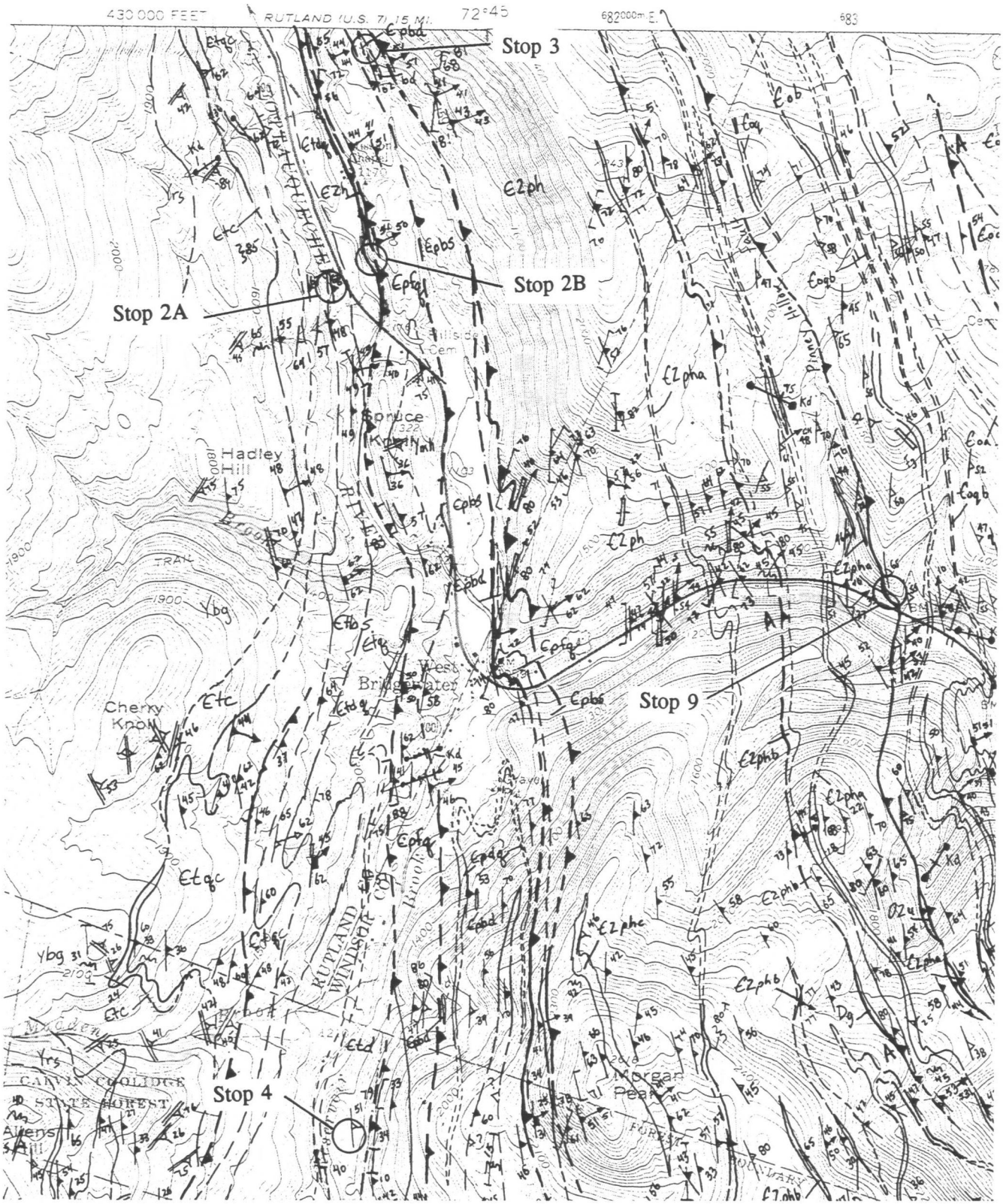


Figure 13. Inset map from Walsh and Ratcliffe (1994a) showing locations of Stops 2A, 2B, 3, 4, and 9. Stops 2 through 9 are shown in figures 13 through 17. See table 1 for the description of map units. Scale 1:24,000 (see figure 15 for bar scale).

We fail to see the argument for a terra rosa because the contact is sharp, the magnetite is disseminated in the albitic (Na-rich) granofels, which is not a good analog for terra rosa developed on dolostone. However, this section from dolostone, containing quartz wacke beds, into magnetite-rich granofels that is gradational into flaser-bedded quartzites (our next unit) is very similar to relationships between the Forestdale Marble and overlying Moosalamoo Phyllite as developed on and along the west side of Mount Nickwacket in the Mount Carmel quadrangle. There disseminated magnetite is present for 50 meters above the Forestdale Marble, in thinly laminated mudstone and crossbedded quartzite. A thin bed of magnetite-cemented dolomite and quartz wacke bed about 5 cm-thick does occur at one locality at the contact with the underlying Forestdale Marble on the southwest flank of Mount Nickwacket. Although we place a fault at the contact in this outcrop and elsewhere in the section, we want to stress the point that the Tyson, Hoosac, and Plymouth may be in general stratigraphic order; however, many of the contacts must be faults, and repetition by faulting is common. Units are cut out along strike in various strands of the Black River fault zone (fig. 1).

Drivers, please return to cars; participants, begin walking 0.7 mi north past Mission Chapel to the start of our traverse.

Log continues from parking lot.

5.4 Turn left on Mission Farm Road

6.1 Park at bridge over Ottauquechee River at north border of Killington Peak quadrangle.

STOP 3. SECTION THROUGH ALBITIC GRANOFELS, INTO FELDSPATHIC TECTONITE OF THE PLYMOUTH UP TO BLACK CARBONACEOUS PHYLLITE OF THE PLYMOUTH(?) TO FOSSILIFEROUS DOLOSTONE IN CARBONACEOUS BLACK PHYLLITE BENEATH THE BASE OF THE PINNEY HOLLOW (fig. 13). (60 MINUTES) Walk north on dirt road to trail on left and head east up slopes.

STOP 3A. Albitic granofels of Hoosac Formation overlain by what we informally call spaghetti rock. This highly tectonized rock contains a thoroughly transposed schistosity and abundant reclined folds and a strong downdip lineation, veins of quartz and albite forms pods in the rock. This rock can be traced all along the east side of the Sherburne Valley as a linear belt despite the highly folded rocks adjacent to it.

STOP 3B. Above spaghetti rock, feldspathic quartzite of the Plymouth, note the transposed layers and abundant folds plunging downdip. This rock continues to the top of the hill where a more quartzose and less deformed rock occurs.

Walk west across a valley without outcrops, go upslope and watch for float of graphitic schist. Climb vertically up approximately 100 ft to exposures of black dolostone.

STOP 3C. Pod of carbonaceous dolostone and graphitic phyllite. This dolostone is interbedded at the southeast end of the crop with carbonaceous phyllite. The dolostone-phyllite contact is folded into a south-verging reclined fold; the axial surface is the regional S_2 foliation.

Early Middle Ordovician conodonts have been obtained from this rock. These results were unexpected as the phyllite was thought to be the uppermost member of the Lower Cambrian Plymouth Formation! However, no fossils have previously been found in any rock on the east flank of the Green Mountain massif. Similar age conodonts occur in dark limestone pods in carbonaceous phyllites of the Melborne Formation of southern Quebec at the top of the Oak Hill Group (Marquis and Nowlan, 1991). Unfortunately, all our other attempts to obtain conodonts from six other sample localities were futile.

Carbonaceous phyllites extend up the hill for about 100 feet, then the green magnetite-chlorite-muscovite-quartz phyllites, having the intense S_2 transposition structures, are found. We have interpreted this contact as a

major fault. It separates autochthonous, but faulted Late Proterozoic and Lower Cambrian cover sequence rocks of the Tyson, Hoosac, and Plymouth and an at present unrenamed Lower Middle Ordovician black phyllite from the allochthonous sequence of the Pinney Hollow, Ottauquechee, Stowe, Moretown, Barnard Gneiss, and Cram Hill to the east. From the U-Pb zircon ages of igneous rocks within the Moretown Formation, which range between 496 ± 8 and 462 ± 6 Ma, the Moretown Formation near the top of the allochthonous section is older than the dolostone and black phyllite beneath the Pinney Hollow Formation.

6.2 Turn left on Routes 100 and 4
7.0 Feldspathic quartzite of the Plymouth Formation on right
8.0 Turn right (south) on Route 100
9.0 Town line
9.5 Park by entrance to quarry on left, or drive in, if gate is open.

STOP 4. QUARRY IN DOLOSTONE OF THE TYSON FORMATION AND EXPLANATION OF TECTONIC BRECCIAS SEEN AT STOP 2 (fig. 13) (10 MINUTES). East-dipping, well-bedded dolostone with quartz wacke beds occur in the west side of the quarry; on east wall of the quarry is a zone of mylonitic dolostone containing the tectonic flow breccia seen at Stop 2B. Note that the coarser grained knots of dolostone are actually sheared off noses of F_2 reclined folds; the ductile matrix is finer grained and strongly foliated. A strong (F_2) downdip lineation can be seen. Veins of specular hematite 2 to 5 cm thick vein the rock and are folded. Iron mineralization accompanied faulting. Dolostone extends to the upper tree line at the south end of the quarry; the next outcrop above that is magnetite albitic granofels like that exposed at Stop 2B. This shear zone is located within the Tyson dolostone, not at the top. This interesting exposure helps clarify our interpretation of the flow breccia at Stop 2B, and helps support the existence of thrust faults within the Tyson section.

Continue south on Route 100.
10.5 Crops of the non-mylonite quartzite and dolostone of the Tyson.
11.3 Road cuts in dolostone, quartzite on west
11.8 Begin outcrop of quartzite and retrograde calc-silicate of the Mount Holly.

12.0 Park on left side of road just after stream; cross south of Black Pond.

Optional STOP 5. ALBITIC GNEISS AND CALC-SILICATE ROCKS OF THE MOUNT HOLLY COMPLEX (fig. 14.) This area contains some fascinating exposures that illustrate the complexities of basement-cover relations in the Green Mountains. The rocks exposed in the roadcut were mapped by Brace (1953) as his A member of the Saltash Formation, that he regarded as cover rocks older than the Tyson. Doll and others (1961) assigned these rocks to the basal part of an expanded Tyson Formation. Mapping here (Ratcliffe, 1994; Walsh and Ratcliffe, 1994a) show these rocks as part of the Mount Holly Complex. The graphitic schist member of the Tyson (the same rock seen at Stop 2A) is faulted against the gneisses, just north of these exposures. At this point rocks of the Mount Holly Complex extend far east into the valley of the Black River. The sequence of rocks exposed here consists of quartzite, carbonaceous, chloritic calcite schists and whitish-gray salt- and pepper-textured magnetite-chlorite-biotite-epidote-albite quartz gneiss.

These units extend in nearly continuous outcrop south of the brook for 600 feet to the adit of an old magnetite mine. The walls of the adit are altered chloritic calc-silicate gneiss, aplite and salt and pepper gneiss. One hundred feet or less south of the adit are numerous exposures and caves in massive, not particularly foliated crops of dolostone of the Tyson. Brace mapped a small northeast-trending fault here. We envision a southeast dipping normal fault that drops down the dolostone that is riding on an east dipping thrust fault. Three hundred feet north of the mine is a second small open pit that exposes the magnetite ore in quartzite, rusty, deeply weathered schistose gneiss and deeply weathered calc-silicate gneiss. The magnetite veins are associated with pegmatite pods and aplite. The entire section of rock extending from the road cuts south to the

Figure 14. Inset map from Walsh and Ratcliffe (1994a) showing locations of Stops 5 and 7, and sample VT/Pl2-19. Scale 1:24,000 (see figure 15 for bar scale).

southern magnetite mine can be traced up the hill to the top of Burnt Mountain 2 km west where aplite and pegmatite can be seen intruding calc-silicate rocks and salt and pepper-textured magnetite-biotite-oligoclase quartz gneiss of the Mount Holly Complex.

The ore zone exposed in the northern opening can be followed by distinctive dark chocolate-brown weathering soil and scattered outcrop to the low pass at the northern end of Burnt Mountain. Similar magnetite-rich salt and pepper gneiss extends farther west and south to North Shrewsbury where it forms a belt about 1 km wide north of Saddle Mountain, in the center of the Green Mountain massif.

Despite the high magnetite and albite content of the rocks exposed in the road cut, and the general resemblance to cover rocks, they are part of the Mount Holly Complex!

13.5 Continue south past the intersection with 100A
15.7 Around bend to left pull into sand pit

STOP 6. DEVONIAN GRANITE DIKES, AGE OF S_2 (F_2) AND ACADIAN DEFORMATION (fig. 15) (Plymouth quadrangle). We are here with the permission of the owner; he asked that we take special care that no one is hurt. Because of the unstable and unsure footing in the gravel, no one is to climb on the gravel-strewn outcrops above the 30-foot drop into the sand pit, please!

Numerous granite dikes intrude the S_2 (F_2) deformed schists. The dikes are known to be Acadian based on two lines of evidence. Dikes like this cut Acadian schistosity in the Waits River of the Proctorsville syncline just east of our final stop (Stop 15). There they are weakly folded by kink folds that crisscross the Waits River. These kink folds have garnet inclusion structures that indicate garnet growth (Acadian) outlasted the crenulation events. The second line of evidence comes from U-Pb dating of zircon from the Black Mountain granite in the core of the Guilford dome where Hepburn and others (1984) noted that the granite cross cuts schistosity in the country rocks, but is also foliated. John Aleinikoff's U-Pb date from the Black Mountain Granite shows that it has an age of 373 ± 4 Ma (table 1 and fig. 16). Attempts to date the dikes in this pit were not successful because of lack of zircon, but zircon from a larger body of granodiorite in the Pinney Hollow Formation in this quadrangle (Sample VT/P12-19, fig. 14) suggest an Acadian age based on SHRIMP ages from one grain in a collection of euhedral xenocrystic Middle Proterozoic zircons! These granites are not the favorite rocks of U-Pb geochronologists.

Dikes here cross cut the F_2 (S_2) fault-related structures we have seen so far. Zones of spaghetti rock seen at Stop 3A are cross cut and the F_2 lineation is rotated in the granite! The dikes are weakly to strongly foliated. At one point just east of the vertical face the dikes are faulted and prophyroclastic. The fault strikes N25°W and dips 55°NE; excellent structures resembling a "C" and "S" fabric, or ramp and flat thrust feature as well as actual offsets show this is a thrust fault, probably Acadian.

We have, therefore, in this exposure evidence for the pre-granite age of S_2 (F_2) and our fault contacts in the region and also evidence for Acadian thrust faulting and foliation that is younger than S_2 (F_2). We have mapped two Acadian thrust faults near here (Walsh and Ratcliffe, 1994a) and several smaller ones in the Pinney Hollow and Ottauquechee belts to the east. In this area, the Acadian strain appears to be domainal, and spaced along discrete zones. The crenulation cleavages in the area are regionally developed and record weak Acadian strain through most of the area. To the east, Acadian strain must increase. On to the next stop!

Turn right and back to Plymouth Union.
Plymouth Hollow on left.

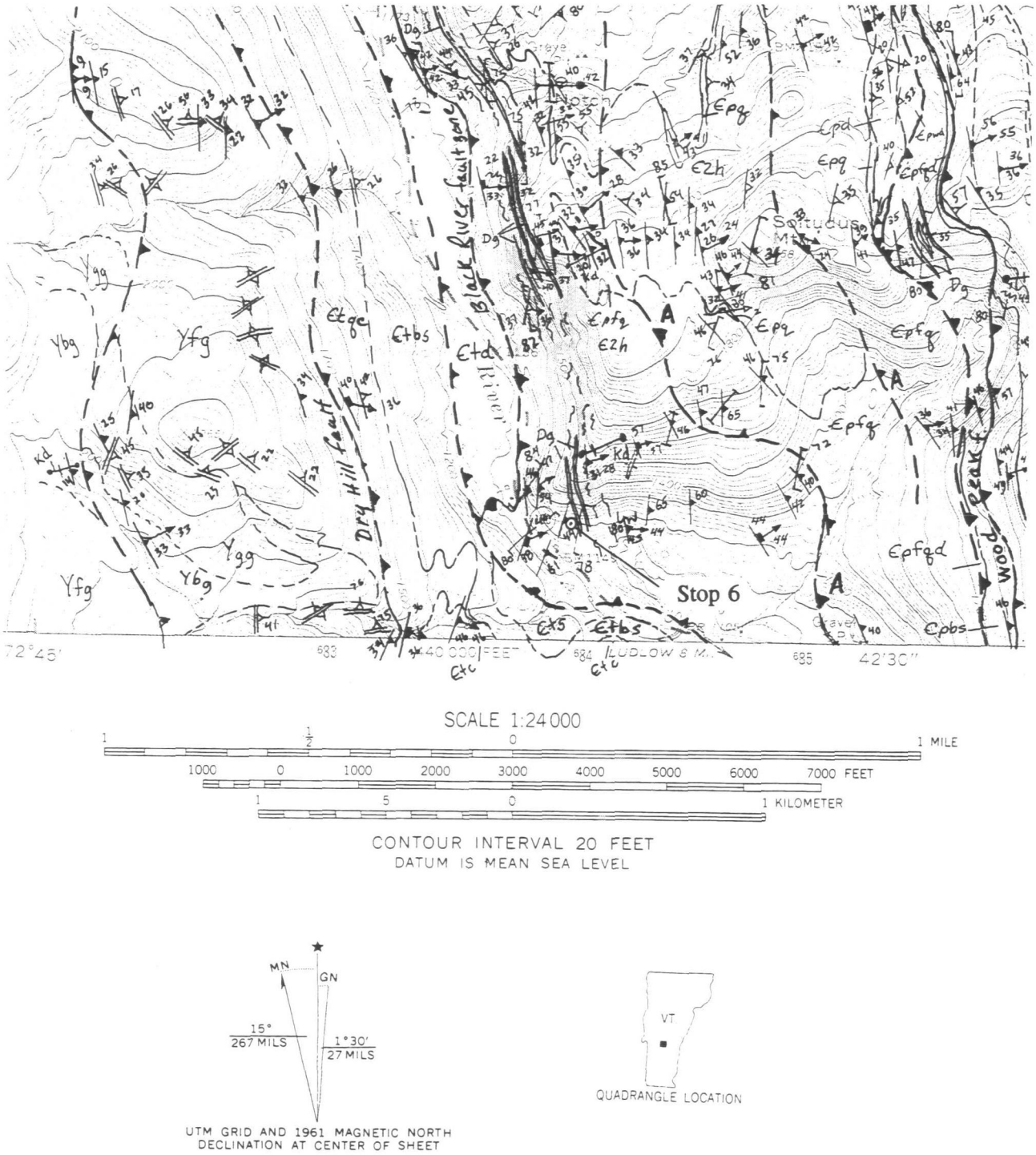


Figure 15. Inset map from Walsh and Ratcliffe (1994a) showing location of Stop 6.

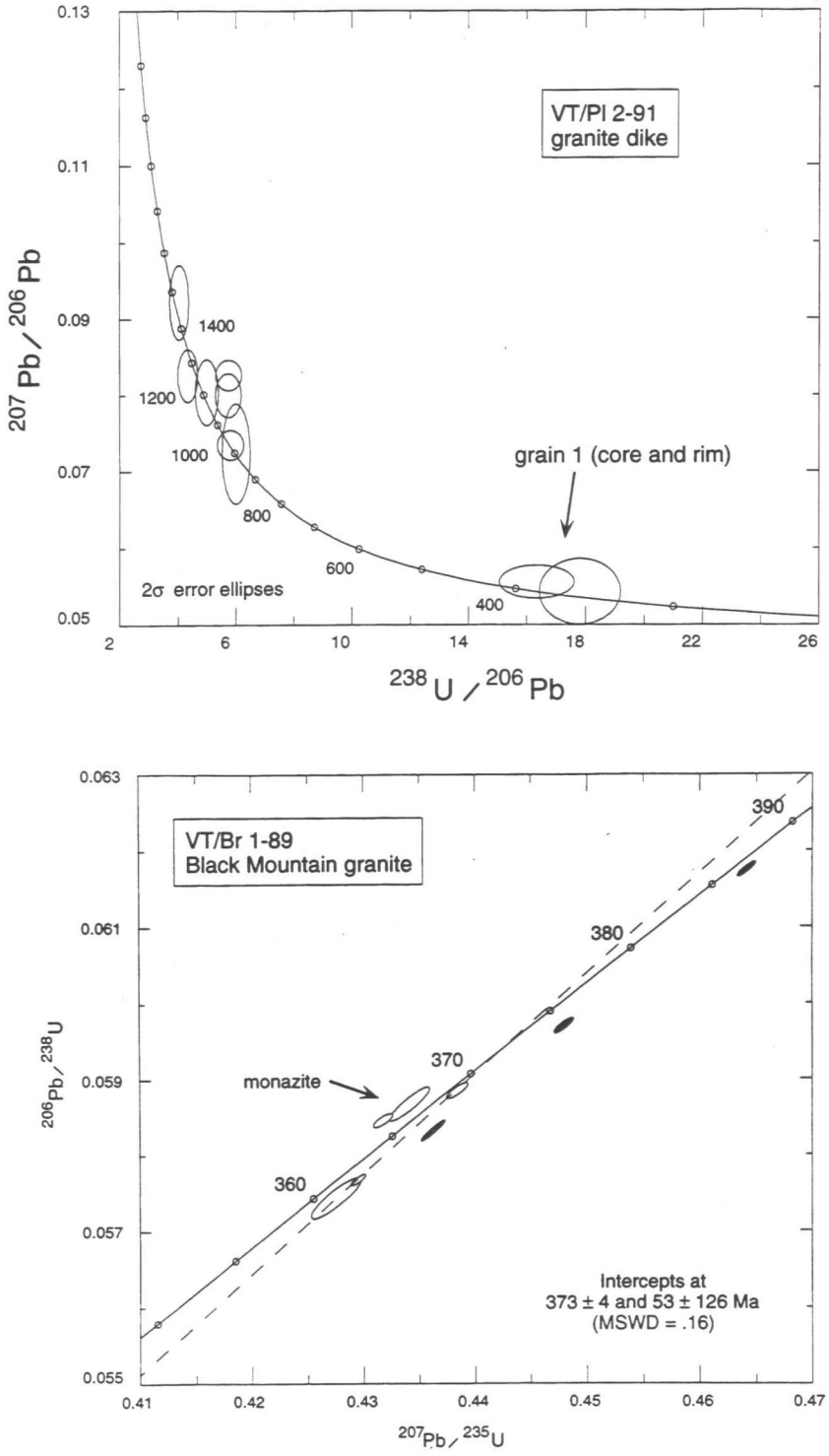


Figure 16. U-Pb geochronology of Devonian granites in table 3, PL2-91 granodiorite stock in Pinney Hollow Formation (see fig. 14 for location); Brl-89 Black Mountain granite in Guilford dome, Brattleboro 15 minute quadrangle.

18.1 Turn right of 100A
On the left behind the Saltash Inn is the Plymouth magnetite mine, where the sharp contact between Tyson dolostone and albitic Hoosac can be seen in a small 30-foot long underground exposure. The contact looks like that seen at Stop 2B. There is no evidence of interbedding or of a regolith. Outcrops along the road are feldspathic pinstripe of the Plymouth seen at Stop 3B. Contrary to assertions by (Ratcliffe, 1994) that there are exposures of granite gneiss along the cliffs north and south of this cut, he now no longer believes that is correct. We still think, however, that there are faults in this section.

19.1 Lands Hill Road on right; Calvin Coolidge homestead on the left. The central ridge to the left is cored by a synclinal mass of garnetiferous Pinney Hollow; the lowland is occupied by shelf sequence dolostones of the Plymouth. Thompson (1972, Stop 14) drew special attention to the similarity between sedimentary dolostone breccias in the Plymouth dolostone to basal parts of Brace's Rutland Dolomite (Brace, 1953) or the Dunham Dolomite of Doll and others (1961). He suggested that the shelf sequence extended this far east in the Cambrian.

A black phyllite that was previously mapped as the upper member of the Plymouth Formation immediately underlies the Pinney Hollow Formation in a central syncline. The phyllite is identical to that seen at Stop 3B. It also contains pods and thin beds of dark-blue gray dolostone like the fossiliferous rock. Attempts to obtain conodonts from beds at Plymouth have so fare been futile.

The contact between Pinney Hollow Formation and black phyllites duplicate relations between the Taconic allochthon and the Vermont Valley sequence to the west. If this is correct the contact with the Pinney Hollow Formation marks either the actual sole of the Taconic allochthon east of the Green Mountains or the best approximation of the root zone for the allochthon (Stanley and Ratcliffe, 1985). The eastern contact of the Plymouth consists of two imbricate thrusts, the eastern one, the Wood Peak thrust, carries Pinney Hollow over feldspathic quartzite and black phyllites of the Plymouth and/or the unnamed Ira correlative. This fault traces north to the fossil locality at Stop 3C.

20 Cross Wood Peak thrust
20.2 Typical Pinney Hollow

STOP 7. TYPE PINNEY HOLLOW FORMATION (fig. 14). (10 MINUTES) Chlorite-muscovite-quartz±magnetite±chloritoid phyllite with well-developed S_2 (F_2) structure, chlorite pseudomorphs after garnet and two subvertical crenulation cleavages. One strikes northwest to east-west, the later one strikes N.15°E.

20.4 Greenstone in Pinney Hollow
20.6 Abundant garnet and coarse magnetite that is concentrated on joints occurs on hill north of road. The late magnetite mineralization possibly is related to Acadian granites.
20.9 Entrance Coolidge State Park
21.2 Greenstone and a carbonaceous phyllite cross the road.
21.7 Ottauquechee contact, here a late (Acadian) fault marks the contact, whereas to the north and south the F_2 (S_2) generation Pinney Hollow fault is the contact. Along road the Ottauquechee is 1.2 mi wide and consists of approximately 60% black carbonaceous phyllite, black and gray quartzite, 30% gray or green feldspathic schist and granofels, and no amphibolite. A large tear drop-shaped mass of amphibolite, bordered by feldspathic granofels occurs just north of 100A in about the middle of the formation. Four bodies of serpentinite occur south of 100A, associated with amphibolite or isolated in the black phyllite. Coticule occurs at one locality (see Walsh and Ratcliffe, 1994a, b).

23.5 Cross Bald Mountain Fault. Contact of Stowe on black phyllite of the Ottauquechee. This is interpreted as a Acadian thrust marked by a well-developed eastwardly increasing Acadian strain gradient. Acadian

garnet overgrows Acadian folds of Taconian S_2. The Stowe consists of two major units of coarsely garnetiferous schist and a non- or sparingly garnetiferous chlorite-biotite-muscovite±magnetite quartz-knotted schist. Unit is 1 mile wide along 100A, and thins to about 0.3 mile at the southern border of the Plymouth quadrangle.

23.9 Contact with informal Whetstone Hill Member of the Moretown Formation
24.5 Volcanic breccia in the Whetstone Hill Member
24.7 Moretown Formation laminated pinstripe schist
25.3 Intersection with Route 4; turn left on Route 4.
25.8 Whetstone Hill crops
26.4 Stowe-Moretown contact
27.0 Rogers Road on right; turn right
Log resumes at Rt 100 and Rogers Road
Turn right (if you can, up dirt road to right)
(Stop log resumes on return)
Park at end of drive

STOP 8. THE BALD MOUNTAIN FAULT: THE CONTACT BETWEEN THE OTTAUQUECHEE AND STOWE FORMATIONS (fig. 17) (PLYMOUTH 7.5' QUADRANGLE). From the camp, walk approximately 100 meters west-southwest along slope and across the meadow to an exposure in the woods. The rock on the east side of the exposure is gray-green biotite-garnet schist of the Stowe Formation. Rocks on the west side of the exposure include black graphitic schist and phyllite and interlayered rusty-weathering, biotite-garnet schist of the Ottauquechee Formation. The contact between these rocks is interpreted as a fault. The contact was named the Bald Mountain fault for exposures just west of the summit of Bald Mountain in Bridgewater (Walsh and Ratcliffe, 1994a, b) (fig. 17). A sharp break in slope, easily visible from the Ottauquechee River valley to the north and from Pinney Hollow to the south, coincides with the contact on Bald Mountain. Earlier authors characterized the contact as either a transition zone of interlayered black and green rocks 200 to 300 feet wide (Perry, 1929), or an abrupt upward change from black phyllite to pale-green schist (Chang and others, 1965). Keith (1932) was the first to state that the Ottauquechee was terminated at the top by a fault. Our mapping demonstrates that there are interlayered black, graphitic rocks and gray-green, non-carbonaceous, feldspathic rocks throughout the Ottauquechee Formation, and that the gray-green feldspathic schists are present as mappable layers within the black phyllite right up to the contact with the overlying Stowe Formation. The gray-green schists in the Ottauquechee are lithologically similar to rocks in the overlying Stowe Formation, but they truncate along the contact with the Stowe. Furthermore, map units in the Stowe (greenstone, garnet-schist, and silvery green chlorite-muscovite-quartz schist) are also truncated along the contact.

In the lower-plate exposures along the contact the second generation schistosity (S_2, Taconian) is progressively transposed by the third generation foliation (S_3, Acadian) as the fault is approached. The fault itself is characterized by a planar phyllonitic to mylonitic fabric parallel to the S_3 foliation. The planarity of the contact and the moderate dip (45°-60°) provide one of the few places in the eastern cover sequence where any contact obeys the rule of V's (fig. 17; Walsh and Ratcliffe, 1994a, 1994b; Doll and others, 1961). East of the fault, the foliation is commonly a composite S_2-S_3 schistosity. Where S_2 and S_3 are a composite fabric, the older schistosity is transposed into the S_3 orientation on the limbs of F_3 folds, and a relict S_2 schistosity is often preserved in the hinges of such folds. West of the fault the S_2 schistosity is characterized by a down dip, east-to-south-east trending lineation (L_2) with sub-parallel isoclinal reclined fold hinges (F_2), and the younger S_3 foliation varies from a crenulation cleavage with associated tight to open, upright, north-to-northeast plunging folds to a penetrative fault-related schistosity (see Stop 6). Where S_3 is a penetrative schistosity anywhere west of the fault, S_2 and S_3 become a sub-parallel composite foliation. The east-to-southeast trending L_2 lineation is preserved in places east of the fault, but more prominent are a northeast trending lineation subparallel to F_3 fold hinges and a north-trending lineation subparallel to F_4 fold hinges.

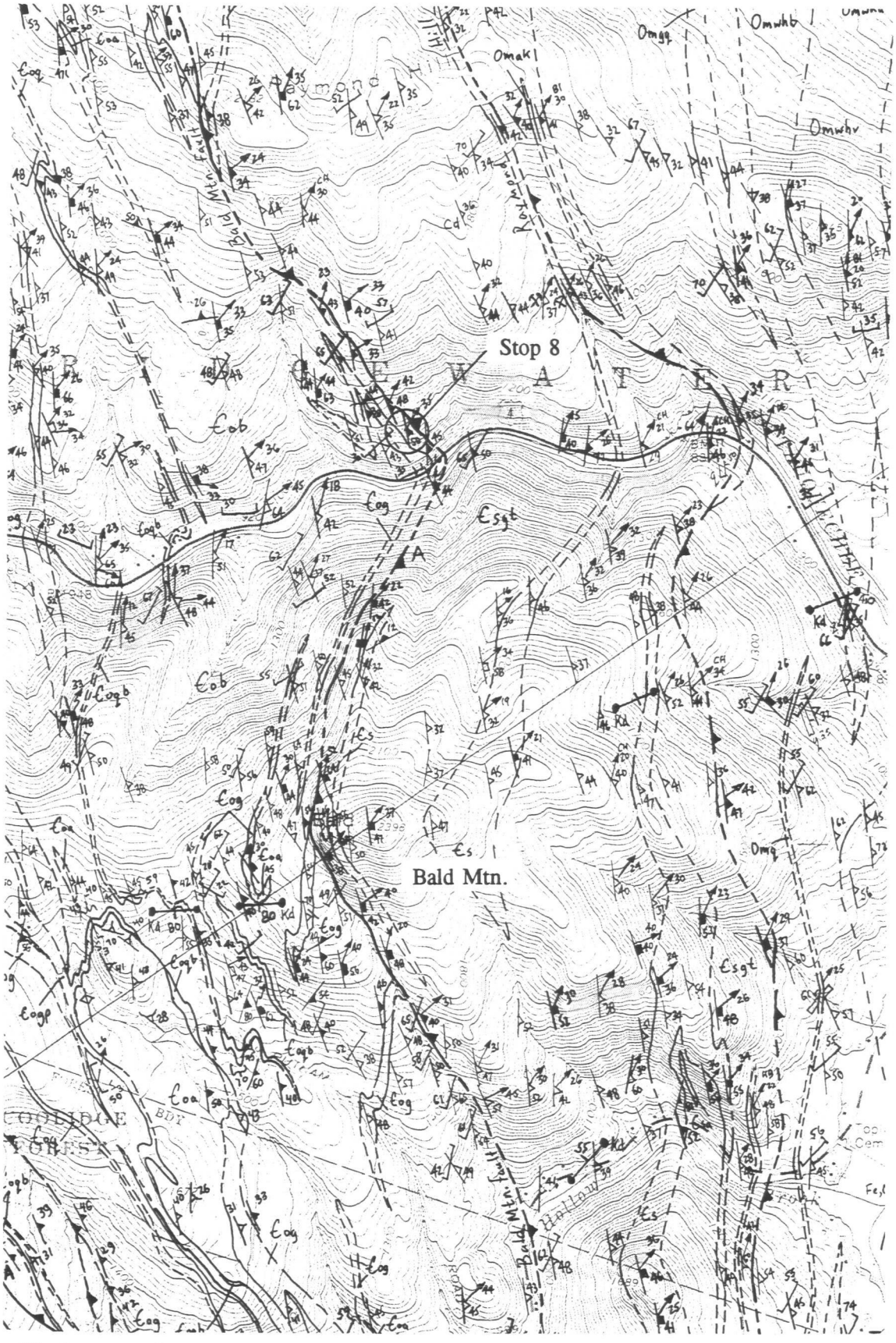


Figure 17. Inset map from Walsh and Ratcliffe (1994a) showing location for Stop 8. Scale 1:24,000 (see figure 15 for bar scale).

The presence of mappable upper- and lower-plate truncations and a penetrative S_3 foliation support the interpretation that here the Ottauquechee-Stowe contact is an Acadian fault.

27.0 Return to Route 4; turn right.
The Ottauquechee in its type section along the river is 2 miles wide and contains about the same proportion of units as along the 100A section. Units are substantially parallel to the Pinney Hollow fault although there is low angle truncation of greenstones in the Pinney Hollow. More pronounced truncations occur along the Bald Mountain fault.
28.4 Gray feldspathic granofels and schist.
28.7 Feldspathic granofels, greenstones, and black graphitic schist.
29.5 Black graphitic phyllite and abundant black vitreous quartzite.

Optional STOP 9. TYPE OTTAUQUECHEE FORMATION (fig. 13). Black phyllites of the Ottauquechee and four vitreous quartzites occur in the lower third of the formation. The contact with the Pinney Hollow is marked by a strong S_2 (F_2) fabric near the Pinney Hollow fault. Discontinuous greenstones occur along the contact. Because of the general lack of greenstones within the Ottauquechee, we place the contact east of the greenstone, but it is possible they are all slivers. A thin zone of talc-carbonate rock several meters thick occurs along the contact 1 mile south of Route 100.

Continue west through the Pinney Hollow.

31.2 Turn south on Route 100 and proceed south through Plymouth Union and past STOP 6.
38.5 Pass into Ludlow quadrangle.
40.4 Crops on west opposite Amherst Lake of deformed S_2 (F_2) elongated pebbles of conglomerate of the Tyson or the east side of the basement-cored Dry Hill anticline. Pebble elongation direction is N70-75°E, and is parallel to hinglines of F_2 reclined folds. Feldspathic pinstripe tectonite of the Plymouth Formation occurs east of the Black River fault in outcrops across the lake.
42.0 Tyson Village, green albitic granofels of the Hoosac Formation forms the highest unit along the west flank of the Dry Hill anticline and is carried westward on an unnamed thrust fault that overrides basement gneiss of the Green Mountain massif. One mile south of Tyson the fault passes down into basement and dies out.
45.6 Turn left on Routes 100 and 103.
46.5 Exposure of pinstriped feldspathic tectonite of the Plymouth west side of road with strong downdip F_2 rodding, on the upper plate of the Black River fault. The last exposures of the pinstripe occur in the southern part of the Andover quadrangle, just west of the point that 100 turns west and heads up the hill toward Weston.
47.5 Intersect Routes 100 and 103 at Ludlow; turn left and follow Route 103 to the east.
47.8 Cross bridge over Black River and proceed to second left turn.
47.9 Turn left up hill on Commonwealth Avenue. The first crops are lustrous green phyllites of the Pinney Hollow within 50 feet of the contact with the Ottauquechee.
48.4 Turn right into the "Views" development park near entrance.

STOP 10. FELDSPATHIC CROSS-BIOTITE GRANOFELS AND BLACK CARBONACEOUS PHYLLITE, QUARTZITE OF THE OTTAUQUECHEE. On the Route 103 crossing, the Ottauquechee is 0.5 miles wide, having thinned from about 0.8 miles at the Plymouth-Ludlow quadrangle border. These exceptional and fresh exposures illustrate the abundance of isoclinal and rootless F_2 folds developed near major contacts. In the crops closest to the road, cross-biotite, garnet-plagioclase-muscovite-quartz schist and granofels occur; one 1-cm thick layer of coticule is exposed near the east end of the first crop. Black graphitic phyllite overlies this, intruded by two narrow felsic Cretaceous dikes above this black quartzite and interbedded granofels and more black graphitic phyllite is exposed. There are several northwest-trending slickensided brittle

faults, exhibiting left-oblique normal faulting.

The garnet and biotite are clearly post S_2 and all garnets studied here and throughout this area show static overgrowth of F_2 microfolds. The garnet is interpreted as Acadian.

Return to Route 103 via Commonwealth Avenue

48.9 Turn left on Route 103

49.8 Turn left up Preedom Hill Road

50.0 Turn left onto Okemo Pines; go 0.2 miles and park past intersection.

STOP 11. TALC-CARBONATE ROCK, ANKERITIC GREENSTONE AND THE "PAC MAN" GARNET MEMBER OF THE STOWE. These crops are about 0.1 mile east of the contact with the Ottauquechee Formation on the west. The Stowe is 0.5 miles wide in this section. Ankeritic greenstone occurs here in the lower third of the formation, but traces to the upper contact 2 miles north of here where the faulted Moretown and adjacent Whetstone Hill member overlie the Stowe, implying a low- angle truncation of units of the Stowe.

In the roadcuts a central talc-carbonate lens several feet wide is bordered by ankeritic greenstone and these in turn are bordered east and west by the most remarkable "Pac Man" garnet member of the Stowe. Where this unit occurs, it has garnets of irregular shape and inclusion structures that suggest irregular growth in the quartzofeldspathic F_2-deformed matrix. The garnets clearly are post S_2 (F_2) as they grow along noses or overgrow F_2 hinges, and could not possibly have been rotated during growth!

F_2 structures are well developed, but a later (Acadian) cleavage cross cuts S_2 as do several spaced faults and associated folds, which fold the F_2 lineation. Note the generally north-plunging Acadian folds of the Taconian S_2.

Twenty feet east of the entrance to the house, on north side of the road a zone of disarticulated S_2 foliation can be seen in the "Pac Man" unit; near the base of the crop a small drag fold and fault is visible along which euhedral garnet is concentrated in a cm-wide zone. Clearly the late fracturing, minor faulting, and garnet mineralization is Acadian. This stop is instructive because it shows the rather low level of Acadian strain seen here, and the partition of that stain into discrete zones.

Return to Preedom Hill Road via the lower loop road.

50.5 Turn left (east) on Route 103.

51.6 Smithville. Turn right on East Street, head south past Luzenac America talc mine on left. We are driving over laminated pinstripe rock of the Moretown Formation (Oml) east of a fault that places the Moretown Formation and the informal Whetstone Hill member of the Moretown on the Stowe Formation to the west.

52.6 Enter Andover quadrangle. The ridges to the east are underlain by ultramafic rocks of the Proctorsville ultramafic belt, containing one of the largest talc-producing bodies in the state. New open pit operations on the left, in the newly expanded Argonaut pit, provide excellent exposures of contacts of the ultramafic rocks (serpentinite, talc schists and talc carbonate rocks) with country rock. The ultramafic rocks (OZu) of the Proctorsville belt form large to small bodies localized along the contact between laminated pinstripe Oml (below) and dark carbonaceous schists and feldspathic quartzite on the east mapped as the Ombs unit of the Moretown. Amphibolite or ankeritic greenstone (Oma) form large to small layers or more irregular masses in either Oml or Ombs near the contact with the serpentinites, but some kind of metasedimentary rock always intervenes between the ultramafic rock and the amphibolite. Just opposite the quarries the road crosses onto green laminated metasiltstone and phyllite of the Stowe Formation. Continue in the Andover quadrangle south on East Street to sharp bend in the road.

53.7 Gate to quarry (road log will resume at exit back onto East Street).
We cross immediately back into the Moretown Formation across a fault that carries Moretown, and the informal Whetstone Hill member of the Moretown Formation over the Stowe Formation. Drive about 1 mile south past old talc quarries to large water-filled abandoned talc quarry.

STOP 12. SOUTHERN TERMINATION OF THE MAIN PART OF THE PROCTORSVILLE ULTRAMAFIC BELT. A central belt of talc schist is in sharp contact on the west with a large-garnet schist of the Moretown Formation and a gritty feldspathic granofels. East of the talc are complexly refolded biotite feldspathic granofels, dark biotitic schists, and minor ankeritic greenstone of the Omb unit of the Moretown Formation. From this point south, talc deposits occur as small pods and stringers in the Moretown Formation, generally bordered on the east by the dark biotite schist (Ombs) unit. The next large occurrence is in the northern part of the Saxtons River quadrangle in the area of Hamm Pit, (see Trip B-6) where the association of units is the same; namely Oml, with or without amphibolite, serpentinite, and black schists on the east. The Proctorsville ultramafic belt and the associated black schist extends south to the area west of Townshend Dam, where a small talc deposit occurs in the middle of the Moretown in the core of the Townshend-Brownington syncline (Ratcliffe, 1997).

53.7 Log resumes at exit from quarry, turn right on East Street, and proceed north to Route 103.
54.8 Turn right on Route 103; pass by road cuts and nearby outcrops of serpentinite on the north.
Continue east to Route 131.
56.0 Turn left on Route 131 east past rusty weathering, splintery sulfidic garnet-biotite-plagioclase quartz schist of the Moretown and associated amphibolite, on east side of the ultramafic rocks.
56.2 Opposite Singleton's Store, felsic and mafic volcanic rocks, interbedded dark schist at the base of the Barnard gneiss of Richardson (1924).
Go through Proctorsville past the school.
56.8 Turn left on Twenty Mile Stream Road, pass amphibolite and black schists of the Cram Hill Formation on the east side of the road, or the west limb of the Proctorsville syncline. Interlayered felsic metavolcanic rock and a thick fragmental-appearing amphibolite mark the uppermost part of the Cram Hill in many localities near the contact with the overlying Waits River Formation that is exposed 0.1 mile east.
57.6 Turn left on Quarry Road; go 0.3 miles to gate to quarry.
58.1 Stop 13

STOP 13. VERDE ANTIQUE QUARRY IN THE MORETOWN FORMATION AT THE CONTACT WITH THE CRAM HILL. This stop is at the north end of a large antiformal, north- plunging exposure of serpentinite. The verde antique (serpentine-carbonate rock) located in the small quarry, is overlain to the north by feldspathic granofels of the Moretown Formation that encloses several large masses of amphibolite. The transition from granofels into the amphibolite is seen to be gradational in cliffs north of the powerline. This interbedding between amphibolite and granofels is common and many of these amphibolites may be in part basaltic volcaniclastic deposits in the Moretown. In the powerline to the north and east Moretown granofels contains garnet. Thin 5- to 10-cm thick mafic, foliated biotitic dikes crosscut schistosity in the Moretown. Similar mafic dikes are present within the upper part of the Moretown, felsic rocks of the Barnard Gneiss of Richardson (1924) and the Cram Hill metavolcanic and metasedimentary rocks. These dikes are only present within the belt of Moretown near the Proctorsville ultramafic belt and in the Cram Hill east of the Chester dome. They are similar to dikes Armstrong (1994) has mapped as mafic dikes of his Branch Brook dike and sill complex that intrudes rocks on the North River Igneous Suite along the eastern side of the Chester dome.

Just over the crest of the hill rusty weathering schists of the Moretown (Omb) crop out. From that point down the hill there are few outcrops but near the base a thin serpentinite occurs within the Ombs unit, followed at the base by felsic trondhjemite gneiss of the Barnard Gneiss (of Richardson, 1924), a mafic or ultramafic

tremolitic dike, followed by a well exposed section of interlayered trondhjemite gneiss, amphibolite, hornblende trondhjemite gneiss, and well-layered felsic and mafic gneiss, all interlayered or intruding black sulfidic schists of the Cram Hill Formation. This section, less than 200 m wide, extends to the highway visible below. East of the road are limited crops of massive amphibolite, interspersed with layers of dark sulfidic schist or more lustrous, muscovite-garnet-studded schist, followed by punky weathering limestone and schist of the Waits River Formation exposed at the high point in the powerline east of the road. The entire section from the serpentinite across to the base of the Waits River is less than 400 m thick.

To the north the contact between the package of Oml unit of the Moretown, associated amphibolite, serpentinite and black schist (Omb unit of the Moretown) passes into a main mass of Barnard Gneiss (of Richardson, 1924) almost 2 km wide. The map relationships suggest the Proctorsville ultramafic belt and associated metasedimentary rocks of the Moretown are intruded by metatrondhjemite of the Barnard Gneiss.

58.4 turn left on Twenty Mile Stream Road
58.9 Davis road on right
60.5 Intersection on left, turn left
60.9 Entrance on right to electric substation
Go to crops behind station

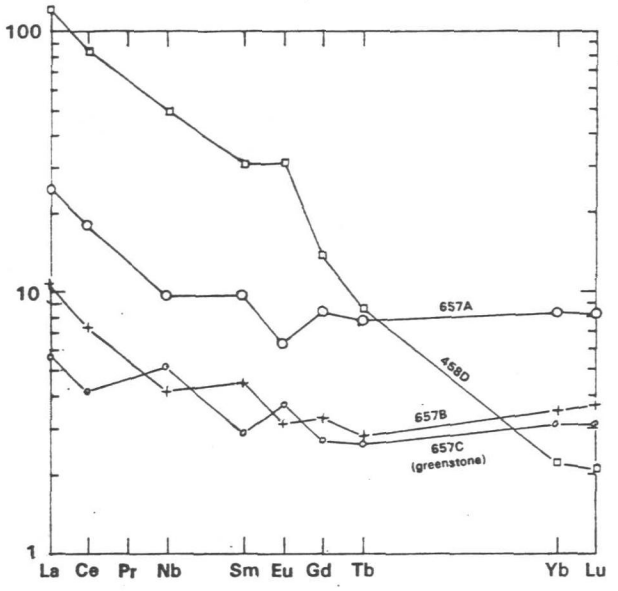
STOP 14. FELSIC METAVOLCANIC AND ASSOCIATED ANKERITIC GREENSTONE IN MORETOWN FORMATION. Behind power substation in rock cuts are exposures of excellent quartzite and green granofels of the Moretown Formation. A felsic layer 3 to 4 m thick grades at its top into a 3-m thick layer of ankeritic greenstone. The contact between the two layers contains a thin zone of reworked felsic rock that in thin section contain cross beds. The felsic layer is dacite of the same major- and trace-element chemistry as felsic rocks of the Barnard Gneiss of Richardson (1924) (Table 2 and figure 18), and is identical to the Obf unit of the Barnard Gneiss at Stop 15. The erosional contact with the overlying greenstone here indicates the layer is metavolcanic. The Moretown contains many layers of felsic and mafic volcanic rocks like this. The chemistry of these suggest that they are arc-related (see geochemistry section and figures).

Attempts to recover zircon from the felsic layer failed. Much of the felsic gneiss of the Barnard at and near its type locality near Bethel, Vermont is interbedded felsic volcanic rock in granofels otherwise typical of the Moretown. Felsic and mafic volcanic rocks, less common "diabasic" dikes and/or sills are common in the Moretown Formation.

61.3 Return to Twenty Mile Stream Road, turn left
61.4 Davis Road on left, turn left
61.5 Turn left on South Reading Road
Crops east of the road for the next 1.5 miles are amphibolite that has beds or inclusions of rusty weathering, sulfidic schists, minor coticule and quartzite. This amphibolite has a fragmented texture consisting of knots and clumps of more plagioclase-rich material and may have been a fragmental andesitic or basaltic andesite. Amphibolites like this are present throughout the areas we have mapped. Crops west of the road are interlayered mixed amphibolite and felsic gneiss (unit Obm).
63.4 Just before power line turn line onto dirt road (4 wheel drive vehicle only) and proceed above 0.2 mi west to top of second knob.

STOP 15. DATED METATRONDHJEMITE OF THE BARNARD GNEISS OF RICHARDSON (1924) IN POWERLINE NORTH OF LUDLOW. Walk west in powerline to second knob, and walk south 300 feet along west side of ridge to ledges of Barnard Gneiss of Richardson (1924). Medium to coarse grained garnet-biotite-magnetite-trondhjemitic gneiss intruded by mafic dikes crop out almost continuously across south-facing low scarps at south end of hill. Both the dikes and the trondhjemite are folded and foliated. Some amphibolitic layers may be screens rather than dikes.

A



B

Major oxide compositions

Sample No.	(1)	(2)	(3)
Field ID No.	657A	657B	657C
SiO_2	72.9	55.7	50.5
TiO_2	0.25	0.40	0.31
Al_2O_3	13.4	15.4	15.4
Fe_2O_3	2.144	7.876	2.782
FeO	1.310	6.680	5.73
MnO	0.70	0.11	0.15
MgO	0.73	3.72	6.72
CaO	2.75	3.24	8.15
Na_2O	4.82	3.59	2.06
K_2O	0.47	0.18	0.12
P_2O_5	0.06	0.07	0.06
H_2O+	0.53	2.53	3.50
H_2O-	<0.05	<0.05	<0.05
CO_2	<0.01	<0.01	4.78
Total	99.49	99.56	100.3

CIPW normative composition

Q	36.08	13.18	19.72
Or	2.809	1.04	0.7336
Ab	41.25	31.52	18.03
An	13.34	16.14	10.16
C	0.1306	3.735	8.567
hy	3.183	30.484	28.08
mt	2.566	2.858	2.715
il	0.4802	0.7883	0.6091
ap	0.1437	0.1720	0.1470
cc	0.0230	0.0236	11.25
Total	100.0	99.9	100.0

Figure 18. A. Chondrite normalized rare earth element diagram of greenstones (657C) transitional volcaniclastic rocks (657B) and 657A dacitic metavolcanic layer at Stop 14. B. Chemical analyses of 657 C, B, and A. Sample 458D is a biotite granite dike in the Proctorsville ultramafic south of Stop 13.

The dated zircon comes from ledges in about the middle of the main belt of biotite metatrondhjemite. The age of 494±6 Ma is very significant as it indicates a minimum age for the Moretown and for incorporated ultramafic rocks of the Proctorsville ultramafic belt in the Moretown. This age is nearly Cambrian, older than the Taconic orogeny and older than sedimentary rocks in the Taconic allochthon, the shelf sequence rocks and the lower part of the Eastern Vermont cover sequence dated to be as young as Middle Ordovician by conodonts.

A 496±8 Ma intrusive age for arc igneous activity and of contemporary felsic volcanic rocks in the Moretown and Cram Hill Formations indicates that formation of the Moretown accretionary wedge and subduction-related volcanism had started by the very lowest Ordovician or latest Cambrian.

The assemblage and some structure within the Moretown must be very old. It must have formed outboard of the Pinney Hollow and Stowe and was later accreted to the underlying rocks after about 460 Ma, but before about 445 Ma the accepted age of the Taconian medium pressure metamorphism (the M_2 Taconian metamorphic event of Sutter and others (1985)), as revised by Hames and others (1991).

END OF TRIP

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