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 1987

Guidebook for Field Trips in Vermont
Volume 2





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Cover: Western half after Stanley and Ratcliffe (1985); eastern half after Doll and others (1961). For explanation of symbols see p. 110.

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FORWARD: FIFTEEN YEARS OF PROGRESS IN UNDERSTANDING THE GEOLOGY OF VERMONT

Introduction:

Since 1972, when NEIGC was last held in Vermont, advances have occurred in all branches of geology. The most notable local applications of these advances involve applying the model of plate tectonics to Paleozoic and Mesozoic geologic history of the Vermont Appalachians. The products of sedimentation, faulting, folding, metamorphism, and igneous intrusion preserved in the rock record are now understood in relation to the processes of rifting, seafloor spreading, subduction, obduction, transform faulting, and collision of outboard terranes. A new level of clarity in picturing the ancient history has resulted from old fashioned field work mixed with new ideas.

Literature concerning Vermont geology is becoming increasingly abundant. References in this guidebook alone call attention to a large part of the material published during the last 15 years, but numerous works have gone unmentioned. The Vermont State Geologist, Dr. Charles A. Ratte, is making an effort to maintain an up-to-date record of work done in Vermont. He started by publishing *Bibliography of Vermont Geology* (Ratte and others, 1980) which has been followed by a supplement (Ratte and Vanecek, 1983).

As our understanding of the local geology advances, it is still convenient to talk about geographic regions of the state using familiar terms, but it is important that the names we use are free of false implications. A case in point is the term "Connecticut Valley Synclinorium" which Hatch (Trip B-3) would like to see replaced by "Connecticut Valley Trough". Until it is determined with certainty whether the fine-grained rocks of the Northfield Formation and Meetinghouse Slate (constituting the flanks of that terrain) are the oldest or the youngest rocks of the belt, we won't know if the structure is anticlinorial or synclinorial.

New data and new interpretations of old data are abundant in the articles of this guidebook, and the remainder of this forward will be used to call attention to this information (see Figure 1 for trip locations). A familiarity with the classical ideas regarding Vermont's geologic history is assumed, and frequent reference is made to the locations within the guidebook where fuller coverage can be found.

Grenvillian Basement Massifs:

For the past several years, Karabinos (Trip C-7) has been mapping in the northern part of the Green Mountain massif, a structure cored by the Mount Holly Complex which was first deformed during the Grenville orogeny in Middle Proterozoic time. He has recognized two units in the massif, a tectonically higher unit (northeast) overlain by an eastern cover sequence and a tectonically lower unit (southwest) overlain by a western cover sequence. The fact that the grade of the Proterozoic metamorphism there was higher than the grades of metamorphism during Paleozoic time helps distinguish the basement rocks from those of the cover.

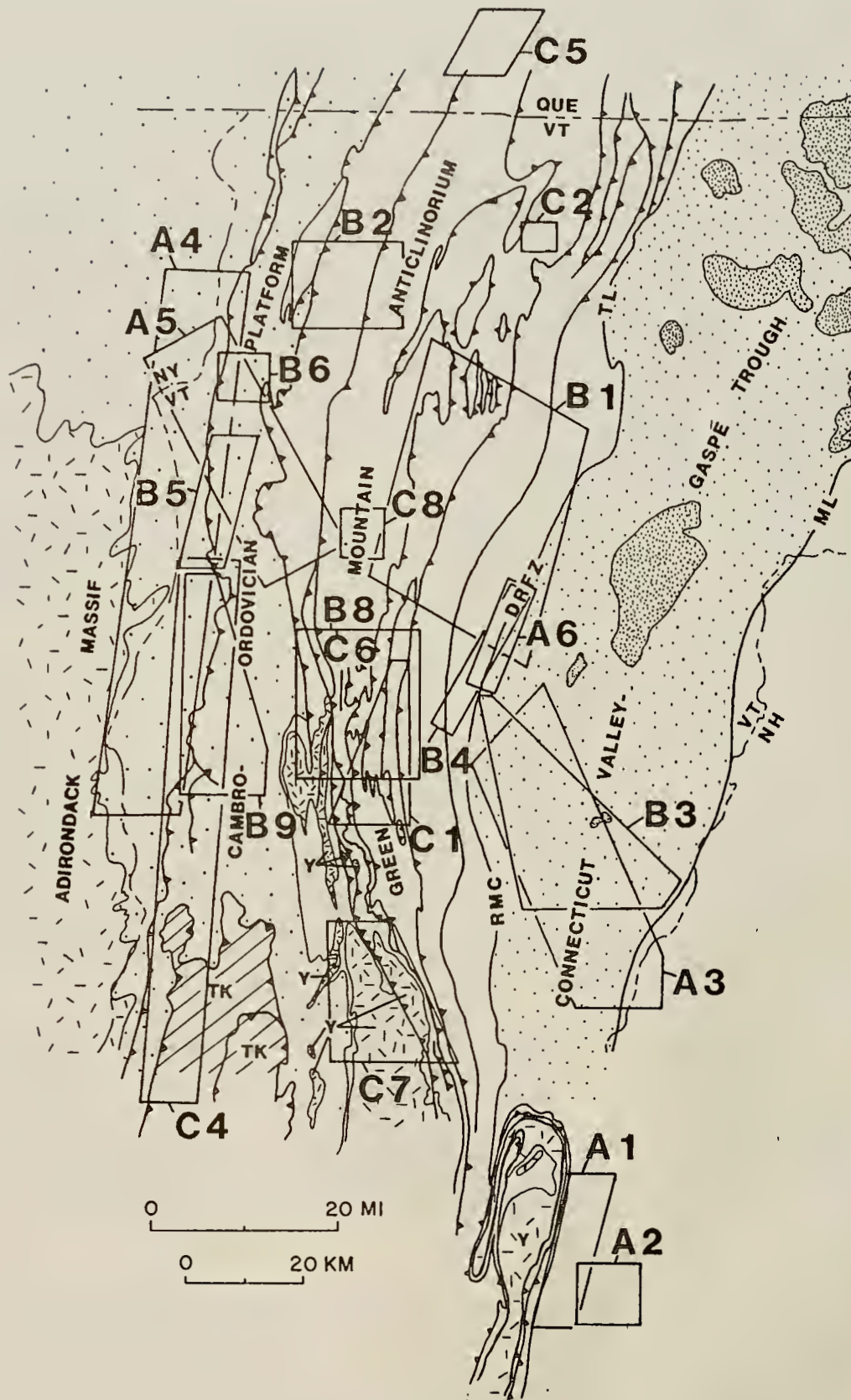


Figure 1. Location map for NEIGC '87 Field Trips. Western half of base after Stanley and Ratcliffe (1985); eastern half after Doll and others (1961).

Stanley and others (Trip B-8) have been studying basement gneisses further to the north and they report Proterozoic mafic dikes which crosscut the Grenvillian foliation along the western margin of the Eastern Lincoln massif. Coish (p. 345 and following) has provided the results and interpretations of his geochemical studies of these rocks which he is able to classify as transitional basalts. These studies support a model in which Late Precambrian rifting initiated the development of an ocean basin (Iapetus Ocean of Rodgers, 1968).

Deposition During Late Precambrian Rifting and Early Paleozoic Time:

Deposition of the earliest sediments on the rifting erosion surface of Middle Proterozoic basement is discussed in several places in the guidebook. Karabinos (Trip C-7) has examined the cover sequences resting unconformably on the Middle Proterozoic Mount Holly complex in the northern part of the Green Mountain massif. After studying the rocks north of the exposed portion of the massif, he finds that a facies change is not observed between the eastern and western cover sequences, but rather they are in fault contact. Stanley and others (Trip B-8) discuss the characteristics of basal conglomerates of the Pinnacle Formation which overlie the gneisses of the Mount Holly complex along the western boundary of the Eastern Lincoln massif.

Late Precambrian/Early Cambrian stratigraphy in southern Quebec has been studied by Colpron and others (Trip C-5) who have been able to determine that subsidence there during the rifting stage remained slow during deposition of the Pinnacle Formation but then accelerated in White Brook time. This contrasts with the more rapid subsidence recorded in Vermont throughout this whole period. Their explanation involves the proximity of the Sutton area to the paleo-position of the triple junction that generated the Quebec reentrant.

Following deposition of the initial rift facies in the newly developing Iapetus Ocean, a stable platform sequence developed during Cambrian and Ordovician time as sedimentation kept pace with the passive subsidence of the shelf (Rodgers, 1968). Mehrtens, Parker and Butler (Trip B-6) describe the facies and evolution of the Cambrian platform sequence in northern Vermont. Study of the three-dimensional spatial distribution of the sequence showed that after the rift clastics were covered by the sands and distal equivalents of the Cheshire Formation, part of the platform collapsed along a listric fault. On the "upthrown" block, the section developed vertically to form the platform sequence. The whole region was experiencing thermal subsidence, but platform sedimentation kept pace with subsidence from Dunham to Danby time. This study provides an excellent example of how the detailed work by geologists trained in one field (here stratigraphy and sedimentation) can be interpreted in the light of other fields (here structure and tectonics) with the results being a more exciting and visible reconstruction of the historical record.

Looking higher in the platform sequence, Chisick, Washington and Friedman (Trip C-4) have been working to establish the stratigraphic and temporal relations of the Middle Beekmantown Group in central and southern Vermont. They review the regional setting in which the shelf sedimentation was occurring during Ordovician time in that area. Their work includes the results of applying an understanding of the regional structural style of thrust faulting to resolve the locally complex arrangement of the stratigraphy.

MacLean (Trip A-4) has studied the Middle Ordovician Glens Falls Limestone, located stratigraphically above the Beekmantown Group discussed above. The alternating limestone and shale beds of this formation record a portion of the depositional history in a rapidly subsiding foreland basin located between a continental massif to the west and an active island arc with impinging thrust nappes coming in from the east. He recognizes that prior conclusions regarding lateral continuity of the section were oversimplified. Pinching of formational thicknesses to the south and facies changes within the Trenton Group to the northeast support tectonic interpretations that such changes may be the result of syndepositionally active block faults. The results of MacLean's work are indicative of how new understanding of tectonic setting provides new perspectives on the history of sedimentation.

Deformation and Metamorphism in the Foreland, Transitional Zone, and Pre-Silurian Hinterland:

Sedimentation in the early Paleozoic Iapetus Ocean continued throughout the spreading and collisional stages until closure of the ocean occurred. Histories of deformation and metamorphism differ along the length of the orogen, and several articles in this guidebook call attention to those differences. Stanley, Leonard and Strehle (Trip A-5) have studied the structural history of the foreland and transitional zone in northwestern Vermont where they recognize north-south trending folds and easterly dipping imbricate thrust faults. They report that along the western margin of the orogen, the Champlain thrust differs in its geometry from the smaller Hinesburg thrust which developed along the overturned limb of a large recumbent fold (Dorsey and others, 1983). They also note that seismic studies have shown that these thrusts extend eastward under the Green Mountains and that the major folds of western Vermont are formed by duplexes and related structures. High-angle Mesozoic faults cut the eastern part of the Platform sequence. Detailed petrographic studies of rocks from the fault zones have increased the confidence in interpretations regarding directions of movement and the conditions within the faults at the time of movement. Regional considerations reviewed in Stanley and Ratcliffe (1985) lead to the conclusion that most of the movement occurred during the Taconic orogeny, but it is important to recognize that available information does not place constraints on the amount of post-Ordovician movement.

Deformation of the foreland has also been studied further to the south in the central Champlain Valley by Washington (Trip B-9) who reports here on the temporal and spatial relations between secondary structures and thrusting. He reviews the various types of thrust systems (duplex and imbricate fan) and the types of folds recognized to be related to these systems (fault-bend folds and related ramp-bend fold trains). Washington has found that the relationship of these folds to the faults that produce them, and the cleavage and joints that form in association with them, can be used to locate thrust faults which are not exposed.

Numerous thrust faults which imbricated the basement and cover rocks of the Green Mountain massif have been recognized by Karabinos (Trip C-7). These westward-directed thrusts have locally transported basement onto the western cover sequence, have separated the massif into two recognizable tectonic units each having its own cover sequence, and have juxtaposed the eastern and western cover sequences north and south of the massif so as to erase the opportunity for determining if the sequences are related by facies changes. He suggests that a large duplex structure involving both western cover and basement may help account for the anticlinorial structure of the massif. The evidence preserved does not allow him to determine if the thrusting occurred during the Taconic orogeny, during the Acadian orogeny, or, as was the case with metamorphism, during both.

Stanley and others (Trips B-8 and C-6) describe a complex history of synmetamorphic thrusting in pre-Silurian rocks located between the platform to the west and the Moretown Formation to the east in central Vermont. They have concluded from their detailed but widespread mapping efforts that the stratigraphy may have been quite simple prior to a complex history of deformation. For example, they propose that the Battell Member of the Underhill Formation, the Granville Formation, and the carbonaceous schists of the Hazens Notch, Pinney Hollow, and Stowe Formations were part of a once-continuous deposit along the eastern edge of North America. They also propose several other correlations between very similar lithologies which have been previously mapped as separate units, suggesting possible physical continuity of these rocks at the time of their deposition.

The deformation of the pre-Silurian rocks is seen by Stanley and his students to be the result of eastward-dipping subduction and the associated westward-directed thrusting. Laird (p. 339 and following) has placed some constraints on the conditions during this synmetamorphic deformation by determining that an earlier medium-high pressure facies metamorphism was followed by a lower greenschist facies metamorphism in mafic schist near a contact between the Pinney Hollow and Ottauquechee Formations. Coish (p. 345 and following) reports on geochemical studies of metamorphosed mafic volcanics in Vermont. Such rocks from the western part of the belt show evidence of having formed within a continental plate, those further to the east have characteristics of having formed near an ocean ridge,

and mafic volcanics found in between show characteristics reflective of both environments. These results support a rifting model with the sequential formation of mafic volcanic units later followed by their accretion during compressional tectonism.

To the north of the area being studied by Stanley and his students, Thompson and Thompson (Trip C-8) have done detailed mapping of structures along the axial trace of the Green Mountain anticlinorium (GMA) north and south of the Winooski River. They recognize two early episodes of folding and interpret them to be of Taconian age, but the third and youngest folding episode is interpreted to have produced the anticlinal structure during the Acadian orogeny. This interpretation conflicts with that of Colpron and others (Trip C-5). Thompson and Thompson have also mapped folded fault surfaces which predate the anticlinal folding, much like those described in rocks to the south by Stanley and others (Trips B-8 and C-6).

Doolan, with Mock and McBean (Trip B-2), provides a complex model of accretion during Ordovician subduction to explain the structures of the Camels Hump Group in northern Vermont. Most of these rocks were deposited as rift-related clastics, laid down prior to the development of a platform sequence. Following the early stages of westward-directed thrusting, some of the supercrustal rocks are proposed to have been backfolded eastward out over the oceanic lithosphere. The model culminates with the deeply deposited Stowe Formation to being thrust westward over the backfolded rocks below. In the process it emerged to provide clasts to the unconformably overlying Umbrella Hill Conglomerate (located at the base of the Moretown Formation). Part of Doolan's model is the separation of the Taconian deformation, which concluded with eastward-directed backfolding, from the collisional stage when island-arc terranes were accreted. Doolan suggests that this final stage was perhaps Acadian and produced the regionally prominent cleavage, the Green Mountain anticlinorium, and the further westward imbrication of the foreland region. Arguments presented below concerning deformation along the western margin of the Connecticut Valley trough support the idea that the regionally prominent cleavage to which he refers is probably Acadian.

Colpron and others (Trip C-5), working to the north in Canada, recognize three phases of deformation in rocks of the Oak Hill Group, all thought to be of Taconic age. They interpret the dominant structural features in that area to have resulted from the second phase, but the third phase was responsible for the formation of the Green-Sutton Mountain anticlinorium.

Working east of the area discussed above, Bothner and Laird (Trip C-2) report that high-pressure facies series metamorphism (Laird and Albee, 1981) is recorded in mafic schists of the Belvidere Mountain Amphibolite member of the Hazens Notch Formation at Tillotson Peak. Their mapping and detailed petrographic analysis of these rocks have led them to postulate that peak metamorphism and

earliest folding were subduction related. This was followed by a second metamorphism and deformation before rapid ascent on westward directed thrusts brought the rocks close enough to the surface to cool quickly and preserve the products of the early metamorphism. Also discussed by Bothner and Laird is the complex history of refolding in the area and the preservation of large E-W fold structures, a condition quite rare in Vermont.

Anderson (Trip B-1) recognizes four generations of veins containing primary metamorphic assemblages in rocks of the Green Mountains north of Interstate 89, each having developed as grade was decreasing from peak condition of the event with which it was associated. This contrasts with the timing of host rock mineral growth during the period while conditions were rising toward peak. The relation between these various generations of veins and the multiple metamorphisms reported by Laird (p. 339 and following) remains to be worked out.

Deposition in the Connecticut Valley Trough (CVT):

Questions abound regarding the source, age, and time of deformation of rocks in the Connecticut Valley trough (CVT). The view popularized by Doll and others (1961) of a synclinal basin floored by an erosional unconformity is in revision. Westerman (Trip A-6) reviews the evidence for and implications of the faulted nature of the western boundary of the trough. He concludes that the western margin of the CVT is not an erosional unconformity or a faulted erosional unconformity except very locally where isolated lenses of the Shaw Mountain Formation are present. In most places the fine-grained rocks of the Northfield Formation are the western unit of the CVT and Westerman argues that they were faulted into place rather than having been deposited against the units with which they are now in contact.

A possible revision of the previously published stratigraphy for the units within the belt is proposed by Hatch (Trip B-3). He calls for a reversal of part of the sequence based both on observations of graded beds and on a sedimentary model involving proximal and distal facies of units deposited from an eastern source. He retains the stratigraphic sequence of the Waits River Formation under the Gile Mountain Formation, but he interprets the Northfield Formation as the western, distal facies of the Gile Mountain Formation.

Concerning the age of rocks in the CVT, Westerman (Trip A-6) discusses the reasons for separating the rocks of the Shaw Mountain Formation from the rest in the trough, based on sedimentological arguments. The Shaw Mountain rocks are known to be Silurian or Early Devonian as shown by fossil evidence (Doll, 1984). The time of deposition of the bulk of the rocks of the CVT (exclusive of the isolated lenses of the Shaw Mountain Formation along the western margin) is uncertain, but reports of graptolites from rocks in the trough (Bothner and Berry, 1985; Bothner and Finney, 1986) indicate that at least part of the trough is of Ordovician

age. Since some of the graptolites come from the Northfield Formation which Hatch suggests may be the youngest unit in the section, the entire CVT may be pre-Silurian and the Shaw Mountain Formation may represent slivers of near-shore Silurian sediments caught between two accreting Ordovician terranes.

Deformation and Metamorphism in the CVT:

Based on their study of rocks in the CVT in southeast Vermont, Boxwell and Laird (Trip A-1) report evidence for two pulses of prograde metamorphism separated by a deformational event, all followed by retrograde metamorphism which locally reached biotite grade. This occurred in rocks thought to have experienced tectonism only in the Devonian. An important result of their study is the recognition that equilibrium assemblages indicate that conditions during the dominant metamorphism gradually increased from east to west across the study area. This is in contradiction to the pattern of isograds shown on the Centennial Geologic Map of Vermont (Doll and others, 1961).

Anderson's work (Trip B-1) on metamorphic vein development involved rocks from both the GMA and the CVT in north-central Vermont. He reports two prominent generations of metamorphic vein growth in rocks of the CVT with much of the mineral development having occurred as grade was decreasing from the peak conditions of the metamorphic event with which it was associated. His picture of multiple metamorphisms is compatible with earlier work and with the work of Boxwell and Laird discussed above. It is also quite compatible with the multiple episodes of isoclinal folding reported for rocks of the Northfield Formation by Westerman (Trip A-6).

No overall structural model has been proposed to explain the complexity of deformation preserved in the rocks of the CVT. After recognizing the faulted nature of the margins of the basin, and thereby removing the constraint that the marginal rocks need be basal units of a stratigraphic sequence, Hatch (Trip B-3) has developed a structural model of an anticlinal arch that fits his sedimentary model.

Regarding the Question of Taconic vs. Acadian Ages for Structures:

Many geologists have noticed that as they drive eastward across Vermont from Burlington to Montpelier, they cannot with confidence identify the ages of the structures which they observe in the roadcuts. Strongly cleaved rocks occur a few miles east of Burlington, with the cleavage dipping at moderate angles to the east. As one travels eastward, the prominent cleavage steepens, passes through vertical, and in the vicinity of Montpelier it dips steeply to the west. At that location, when passing over a boundary known as the Taconian unconformity of Cady (1960), the Taconian Line of Hatch (1982), and the Richardson Memorial contact (RMC - informal), the transition is made from the Cambrian-Ordovician terranes of the Green Mountains to the rocks of the Connecticut Valley trough. As can be seen at

the I-89 outcrop exposing this boundary (Trip A-6, Stop 1), it is a fault zone, and the dominant structures in the rocks on both sides are the same.

Along the length of this fault zone it is common to find exposures of the Shaw Mountain Formation whose Silurian age has been firmly established from fossils (see Westerman, Trip A-6). These Silurian rocks frequently exhibit a well-developed cleavage which is parallel to that on both sides of the RMC, so it seems safe to conclude that the dominant, pervasive cleavage seen in rocks adjacent to the RMC must be Silurian or younger (i.e. Acadian). The outcome of current discussions referred to above regarding the age of rocks within the Connecticut Valley trough does not affect this conclusion.

For the highly deformed and strongly metamorphosed rocks of the Green Mountains, a strong case has been made (Stanley and Ratcliffe, 1985) that westward-directed Taconian thrusts produced the map pattern seen today, and the timing of this major deformation is supported by the work of Sutter and others (1985). Where the cleavage is synmetamorphic and metamorphism occurred during the Ordovician, the cleavage is clearly a Taconic structure.

Assuming from the arguments above that the prominent cleavage seen in rocks on the western flank of the Green Mountains is Ordovician in age and the prominent cleavage seen in the Montpelier area is Devonian in age, and given that it is not readily apparent that these cleavages aren't the same one, then is it reasonable to consider that the development of the prominent cleavage is diachronous and perhaps best described as Tacadian?

Mesozoic Intrusions and Rift Features:

Application of the concepts of plate tectonics is not restricted to the Paleozoic history of Vermont. McHone has been studying Mesozoic dikes and related structures throughout New England, and here (Trip B-5) he discusses his reasons for proposing that the Champlain Valley is a structural basin formed by Cretaceous rifting. This is a model which he sees as inviting comparison to other younger and better-studied rift basins. Radiometric dating of the intrusions shows them to be of undoubted Cretaceous age and their geochemical character supports an intra-plate origin. The high-angle faults of the area also have orientations which match well with such a model, but constraints on the timing of the faults remain limited.

Glacial history:

Studies of past glacial activity in Vermont continue primarily through the efforts of a small group of energetic workers. Three papers in this guidebook, two by Larsen (Trips A-3 and B-4) and one by Ackerly and Larsen (Trip C-1), cover many of the ideas that have come forward in the past 15 years. Although no overwhelming new theory of glaciation has been applied which might be analogous to the application of plate tectonics to ancient mountain building, re-examination of

old ideas, applications of ideas from different fields, and a steady search for new information has produced results.

Larsen (Trip A-3) reports on the deglaciation of Vermont and the relative ages of glacial Lake Hitchcock, glacial lake Winooski, and the Champlain Sea. After defining the precise location of the boundaries of Lake Hitchcock, Koteff and Larsen (1985) were able to use elevations of shoreline indicators to calculate the current slope of the ancient lake surface. It is planar and rises 4.74 feet/mile in a direction of N21.5W. The planar character is interpreted to suggest that no rebound had occurred prior to the draining of Lake Hitchcock. Larsen uses evidence of high energy gravel deposits in the valley formerly occupied by Lake Hitchcock to conclude that it drained while Lake Winooski still existed. Since Lake Winooski drained before ice retreated to the St. Lawrence Valley and allowed marine water to flood the Champlain Valley, he further concludes that Lake Hitchcock must have drained long before any marine sediments were deposited in the Champlain Sea about 12,500 BP. Larsen (Trip B-4) uses the glacial and postglacial sediments in the Dog River Valley to test the model of deglaciation described above. Also on this trip he evaluates his conclusions regarding postglacial rebound as determined from the current elevations of features indicative of former lake levels.

Ackerly and Larsen (Trip C-1) report here on evidence indicating glacial ice flow directions in the Green Mountains. Examination of an old set of data indicating southwest-directed ice flow in the area of Middlebury Gap confirms that the data are real and more widespread than previously reported. Alternative explanations include 1) a local ice cap in the Green Mountains with ice flowing southwestward into the Champlain Valley, or 2) reversal of ice flow direction as a result of drawdown of ice thickness in the Champlain Valley. Given an absence of evidence for a local ice cap in the region and anticipation of the ice flow reversal by the theoretical ice surface reconstructions of Hughes and others (1985), Ackerly and Larsen opt for the latter explanation.

Hydrology:

Although there are numerous studies going on in Vermont and other New England states involving efforts to understand hydrologic systems, only one hydrology-related field trip (Caldwell and others; Trip A-2) is available this year (and it's on the "wrong" side of the river). It is unfortunate that the results of most hydrologic studies do not become available for field review by the public since they are conducted for clients in the business sector rather than as pure research.

A catastrophic flood in the Cold River of southwestern New Hampshire produced erosional and depositional features, and had a pronounced effect on the local groundwater system in a small kettle. Recharge far exceeded the total rainfall as a result of runoff into the kettle from the nearby steep hillside. The results of this study indicate that estimates of recharge of an aquifer can be significantly underestimated if secondary recharge is not considered.

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September, 1987

CHRONOLOGICAL LISTING OF MEETINGS OF THE
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Meeting	Year	Location	Organizer
1st	1901	Westfield River Terrace, Mass.	Davis
2nd	1902	Mount Tom, Massachusetts	Emerson
3rd	1903	West Peak, Meriden, Conn.	Rice
4th	1904	Worcester, Massachusetts	Emerson
5th	1905	Boston Harbor and Nantasket	Johnson, Crosby
6th	1906	Meriden to East Berlin, Conn.	Gregory
7th	1907	Providence, Rhode Island	Brown
8th	1908	Long Island, New York	Barrell
9th	1909	North Berkshires, Mass.	Cleland
10th	1910	Hanover, New Hampshire	Goldthwait
11th	1911	Nahant and Medford, Mass.	Lane, Johnson
12th	1912	Higby-Lamentation Blocks	Rice
13th	1915	Waterbury to Winsted, Conn.	Barrell
14th	1916	Blue Hills, Massachusetts	Crosby, Warren
15th	1917	Gay Head and Martha's Vineyard	Woodworth, Wigglesworth
16th	1920	Lamentation and Hanging Hills	Rice, Foye
17th	1921	Attleboro, Massachusetts	Woodworth
18th	1922	Amherst, Massachusetts	Antevs
19th	1923	Beverly, Massachusetts	Lane
20th	1924	Providence, Rhode Island	Brown
21st	1925	Waterville, Maine	Perkins
22nd	1926	New Haven, Connecticut	Longwell
23rd	1927	Worcester, Massachusetts	Perry, Little, Gordon
24th	1928	Cambridge, Massachusetts	Billings, Bryan, Mather
25th	1929	Littleton, New Hampshire	Crosby
26th	1930	Amherst, Massachusetts	Loomis, Gordon
27th	1931	Montreal, Quebec	O'Neill, Graham, Clark, Gill, Osborne, McGerrigle
28th	1932	Providence-Newport, R.I.	Brown
29th	1933	Williamstown, Massachusetts	Cleland, Perry, Knopf
30th	1934	Lewiston, Maine	Fisher, Perkins
31st	1935	Boston, Massachusetts	Morris, Pearsall, Whitehead
32nd	1936	Littleton, New Hampshire	Billings, Hadley, Cleaves, Williams
33rd	1937	New York City & Dutchess Co.	O'Connell, Kay, Fluhr, Hubert, Balk
34th	1938	Rutland, Vermont	Bain
35th	1939	Hartford & Conn. Valley	Troxell, Flint, Longwell, Peoples, Wheeler
36th	1940	Hanover, New Hampshire	Goldthwait, Denny, Shaub, Hadley, Bannerman, Stoiber
37th	1941	Northampton, Massachusetts	Balk, Jahns, Lochman Shaub, Willard
38th	1946	Mt. Washington, NH	Billings
39th	1947	Providence, Rhode Island	Quinn
40th	1948	Burlington, Vermont	Doll
41st	1949	Boston, Massachusetts	Nichols, Billings, Schrock, Currier, Stearns
42nd	1950	Bangor, Maine	Trefethen, Raisz
43rd	1951	Worcester, Massachusetts	Lougee, Little
44th	1952	Williamstown, Massachusetts	Perry, Foote, McFadyen, Ramsdell
45th	1953	Hartford, Connecticut	Flint, Gates, Peoples, Cushman, Aitken, Rodgers, Troxell

Meeting	Year	Location	Organizer
46th	1954	Hanover, New Hampshire	Elston, Washburn, Lyons, McKinstry, Stoiber, McNair, Thompson
47th	1955	Tigonderoga, New York	Rodgers, Walton, MacClintock, Bartolome
48th	1956	Portsmouth, New Hampshire	Novotny, Billings, Chapman, Bradley, Freedman, Stewart
49th	1957	Amherst, Massachusetts	Bain, Johannson, Rice, Stobbe, Woodland, Brophy, Kierstead, Webb, Shaub, Nelson
50th	1958	Middleton, Connecticut	Rosenfield, Eaton, Sanders, Porter, Lungren, Rodgers
51st	1959	Rutland, Vermont	Zen, Kay, Welby, Bain, Theokritoff, Osberg, Shumaker, Berry, Thompson
52nd	1960	Rumford, Maine	Griscom, Milton, Wolfe, Caldwell, Peacor
53rd	1961	Montpelier, Vermont	Doll, Cady, White, Chidester, Matthews, Nichols, Baldwin, Stewart, Dennis
54th	1962	Montreal, Quebec	Gill, Clark, Kranck, Stevenson, Stearn, Elson, Eakins, Gold
55th	1963	Providence, Rhode Island	Quinn, Mutch, Schafer, Agron, Chapple, Feininger, Hall
56th	1964	Chestnut Hill, Massachusetts	Skehan
57th	1965	Brunswick, Maine	Hussey
58th	1966	Katahdin, Maine	Caldwell
59th	1967	Amherst, Massachusetts	Robinson, Drake, Foose
60th	1968	New Haven, Connecticut	Orville
61st	1969	Albany, New York	Bird
62nd	1970	Rangeley Lakes, Maine	Boone
63rd	1971	Concord, New Hampshire	Lyons, Stewart
64th	1972	Burlington, Vermont	Doolan, Stanley
65th	1973	Fredericton, New Brunswick	Greiner
66th	1974	Orono, Maine	Osberg
67th	1975	Great Barrington, Mass.	Ratcliffe
68th	1976	Boston, Massachusetts	Cameron
69th	1977	Quebec City, Quebec	Beland, LaSalle
70th	1978	Calais, Maine	Ludman
71st	1979	Troy, New York	Friedman
72nd	1980	Presque Isle, Maine	Roy, Naylor
73rd	1981	Kingston, Rhode Island	Boothroyd, Hermes
74th	1982	Storrs, Connecticut	Joeston, Quarrier
75th	1983	Greenville & Millinocket, Me.	Caldwell, Hanson
76th	1984	Danvers, Massachusetts	Hanson
77th	1985	New Haven, Connecticut	Tracy
78th	1986	Lewiston, Maine	Newberg
79th	1987	Montpelier, Vermont	Westerman

ACKNOWLEDGEMENTS

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METAMORPHIC AND DEFORMATIONAL HISTORY OF THE
STANDING POND AND PUTNEY VOLCANICS IN SOUTHEASTERN VERMONT

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INTRODUCTION

Recent investigations of rocks in eastern Vermont, notably those by Chamberlain et al. (1985), Karabinos (1984), Bothner and Finney (1986), Hepburn et al. (1984) and Laird and Albee (1981), have inspired questions concerning the interpretation of the metamorphic and structural histories recorded by the rocks. The Standing Pond and Putney Volcanics occupy a position near a major boundary separating two metamorphic terranes, the Connecticut Valley-Gaspe synclinorium (CVGS) and the Bronson Hill anticlinorium (BHA; Chamberlain et al., 1985). Because of their stratigraphic position and mafic component, the Standing Pond and Putney Volcanics contain critical evidence relevant to both the structural and metamorphic histories of the area. The purpose of this field trip is to examine the metamorphism and deformation of the Standing Pond and Putney Volcanics in southeastern Vermont.

STRATIGRAPHIC FRAMEWORK

The Standing Pond and Putney Volcanics are included in a northeast trending, steeply dipping homocline (the "Vermont Sequence"; Currier and Jahns, (1941) within the CVGS (Figure 1) in southeastern Vermont. Other papers in this volume address the nature of the western margin of the CVGS (David S. Westerman, Trip A-6) and the stratigraphic order of units within the CVGS (Norman L. Hatch, Jr., Trip B-3). Fisher and Karabinos (1980), Hepburn et al. (1984), Hatch (1986) and Bothner and Finney (1986) present arguments regarding stratigraphic succession within the CVGS. A consensus appears to be emerging that the stratigraphic order of rocks of the Vermont Sequence are (seen on this trip from oldest to youngest): Waits River Fm., Standing Pond Volcanics, Gile Mountain Fm. and Putney Volcanics. Facing directions seen at Stop 6 this trip (Figure 2) support the hypothesis that the Gile Mountain Fm., overlies the Waits River Fm.

The Waits River Formation consists of micaceous schist and calcareous quartzite containing lenses and pods of impure marble. A gradation from micaceous schist to impure quartzite occurs in some areas, but generally, layering is distinct. The schists weather rusty brown to medium to dark gray depending upon the amount of carbonate present; alignment of mica grains define a well developed schistosity in the rocks. The massive impure quartzites weather punky brown, and where the peak of metamorphism attained garnet grade the rock is pocked by garnet knobs.

The Standing Pond Volcanics form a time stratigraphic unit between metasediments of the Waits River and Gile Mountain Fms. (Hepburn et al., 1984) and vary considerably across strike. Mafic to felsic varieties of rock are all included within the Standing Pond Volcanics, but in most areas greenstone or amphibolite predominate. A large pelitic component is present in some rocks and results in a very distinctive faculitic layer, examples of which may be seen at Stops 3 and 6. Contacts of the Standing Pond Volcanics with the Waits River and Gile Mountain Fms. are commonly gradational and at low grade (Stop 2) not readily apparent.

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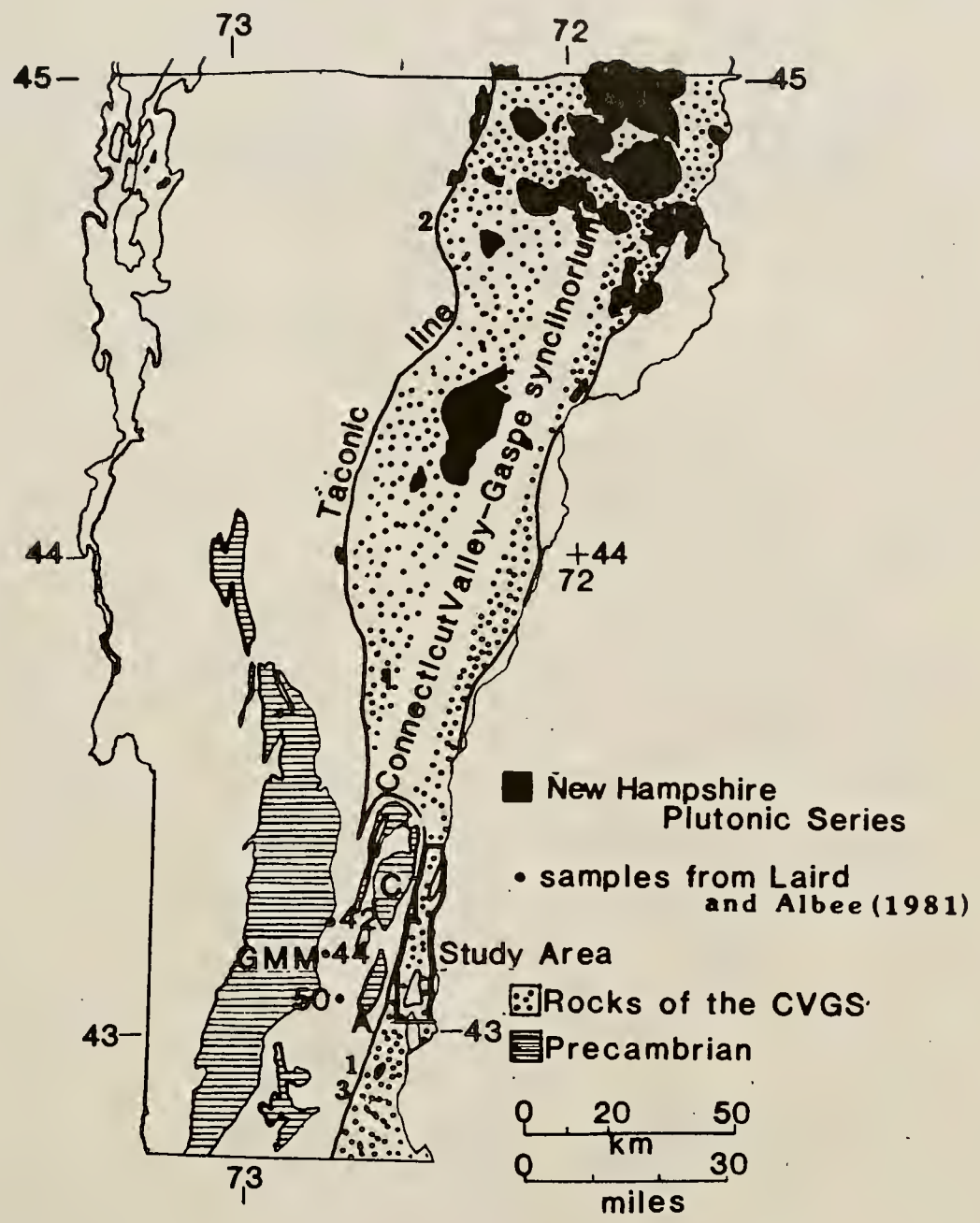
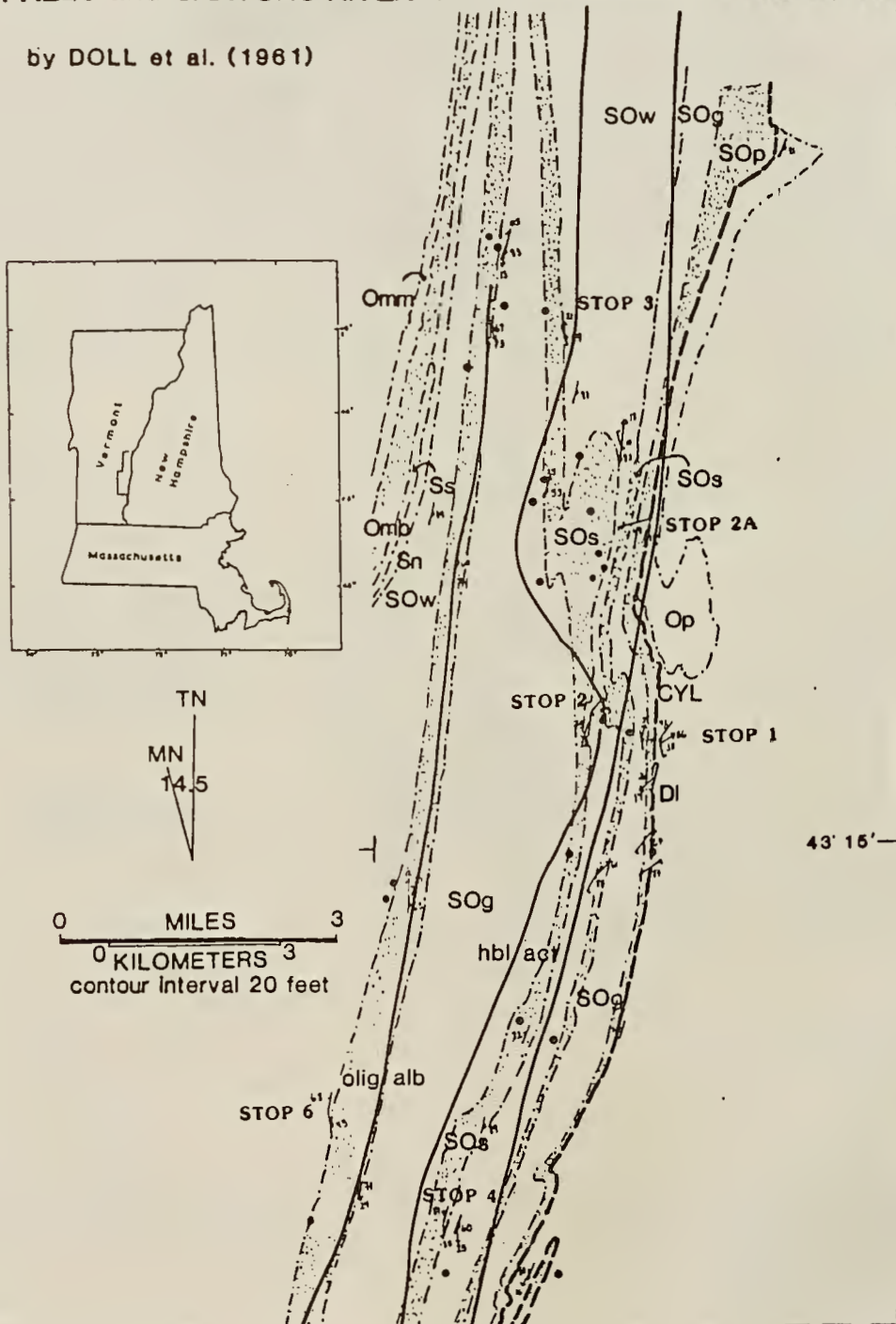


Figure 1. Simplified geologic map of Vermont after Doll et al. (1961) showing the area of this trip (outlined). Numbers refer to localities discussed in the text. Green Mountain massif (GMM), Chester (C) and Athens (A) domes shown for reference.

GEOLOGIC MAP of PART of the CLAREMONT, BELLOWS FALLS and SAXTONS RIVER 15' QUADRANGLES In VERMONT

by DOLL et al. (1981)



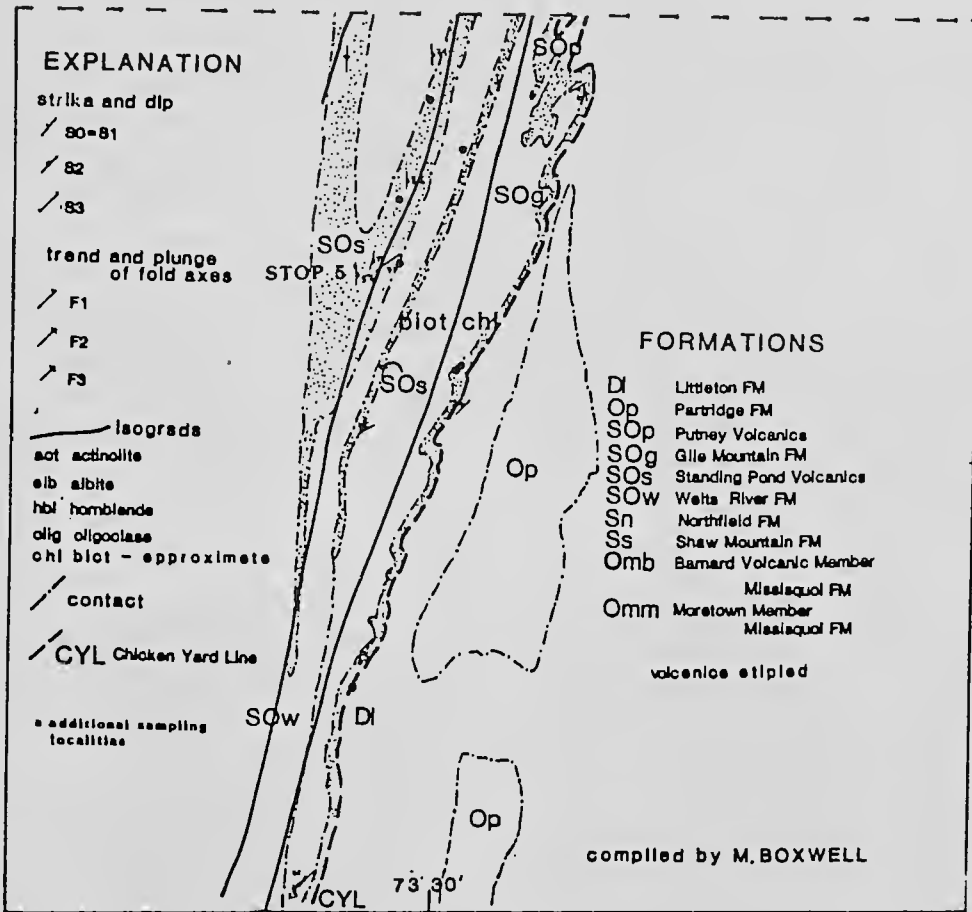


Figure 2. Geologic map of the study area with the stop localities of this trip indicated.

The Gile Mountain Formation in contact with the Standing Pond Volcanics consists of interlayered pelitic schist and quartzite in about equal proportions. Both layers weather medium to dark-gray but are only rarely rusty in contrast to schists and quartzites of the Waits River Fm. Compared with the Waits River Fm., the Gile Mountain Fm. contains a smaller carbonate component and a more monotonous repetition of schist and quartzite. Within the garnet zone, garnet porphyroblasts are abundant in the schistose layers.

The Putney Volcanics are fine-grained, massive greenstones found within the chlorite and biotite zones. At their eastern boundary, the Putney Volcanics are in sharp contact with rocks of the Littleton Fm. The western contact is not nearly as sharp because mafic rocks of the Putney Volcanics grade into chlorite and biotite grade pelitic rocks. Hepburn (1982) distinguishes the Putney Volcanics from the Standing Pond Volcanics on the basis of bulk rock chemistry, and consistent with Hepburn et al., (1984), they are here considered as a formation separate from the Standing Pond Volcanics.

Along the eastern margin of the field area, the Putney Volcanics are separated from the Littleton Fm., included within the Bronson Hill anticlinorium, by a zone of sheared rock (Stop 1). The Littleton Formation consists of dark-gray, massive, micaceous schist and phyllite that contains only a small carbonate component and much less quartz than schists of the Waits River and Gile Mountain Fms. Rocks of the Littleton Fm. generally weather dark-gray or rusty where the sulfide content is high, and are extremely fissile. These rocks are included within the "New Hampshire Sequence" (Currier and Jahns, 1941) which, in southeastern Vermont, is separated from the Vermont Sequence by the Chicken Yard Line (Hepburn et al., 1984).

AGE DATA

The stratigraphic age of the Waits River Fm. as shown on the Centennial Map of Vermont (Ooll et al., 1961) is Devonian. However, recent graptolite rediscoveries near Montpelier, Vermont (Bothner and Finney, 1986) support Richardson's (1916) assignment of this unit to the Ordovician (Middle to Upper) period. Bothner and Berry (1985) report the presence of Upper Ordovician graptolites in rocks equivalent to the Gile Mountain Fm. in Quebec, further supporting an older age for the rocks.

Isotopic data constraining the age of the Vermont sequence include a 423 Ma U-Pb age on zircon from a felsic rock mapped within the Standing Pond Volcanics at Stop 2 (Aleinikoff, 1986, pers. comm.) which implies that the Standing Pond Volcanics are early Silurian or older. Crystallization ages of cross-cutting plutons of the New Hampshire Plutonic Series (Figure 1) provide a minimum age. Naylor (1971) obtained a 375 Ma Rb/Sr age on coarse muscovite and whole rock samples from the Black Mountain granite twenty miles south of the field area (Figure 1, location 1). (Data from Naylor (1971 and 1975) and Harper (1968) reported herein are recalculated using standards reported by Steiger and Jager, 1977.)

Intrusion of the New Hampshire Plutonic Series postdates regional metamorphism and formation of recumbent folds in the CVGS (Naylor, 1975) thereby constraining these "events" to pre-Middle Devonian time if it may be assumed that the deposition of the rocks post-dated the Taconic Orogeny. Metamorphic ages, including a K/Ar whole rock age of 369 ± 8 Ma near the western contact of the Waits River Fm. with the Shaw Mountain Fm. (Harper 1968; Figure 1, herein location 2) and a $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 358.5 ± 4.3 Ma from hornblende of the Standing Pond Volcanics near Brattleboro, Vermont (Sutter et al., 1985; Figure 1, location 3) indicate that the rocks were metamorphosed during a late Devonian event.

STRUCTURE

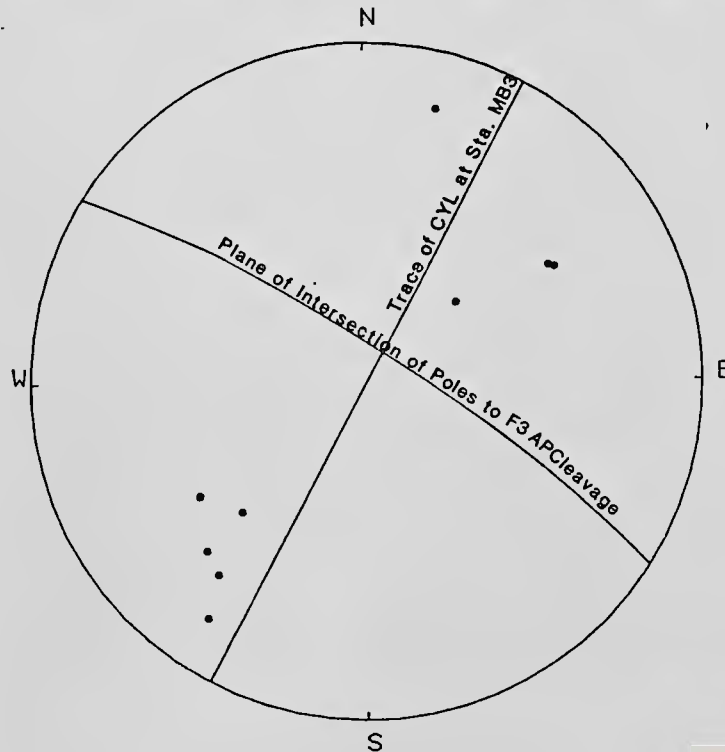
At least three fold generations are recognized in rocks examined on this trip. Bedding, S0, is rarely visible, but where present is parallel to an S1 schistosity. S1, inferred to have formed during an F1 fold event, is only rarely preserved, and is crosscut

by the main schistosity. Axes of F1 folds trend northerly or southerly with a moderate to shallow plunge, and F1 axial planar surfaces dip steeply to the east.

The dominant folding event, F2, is coaxial with F1. Where present, the foliation axial planar to F2 folds varies from a widely spaced axial plane cleavage to a well developed schistosity, S2. S2 is usually undeformed or only broadly warped or kinked by F3 folds. F2 folds are identified in the field by undeformed to gently deformed S1 surfaces. An "average" S2 dips steeply to the west. The plunge of F2 folds is generally moderate to shallow although some southwest trending folds plunge steeply. The variability in trend and plunge of F2 folds may result from porpoising of shallow plunging F2 axes.

F3 is defined by folds of S0, S1, and S2. The majority of F3 folds likewise trend northeast and southwest but plunge moderately to steeply in several directions. F3 folds appear as rounded folds in competent layers and kink bands in schistose layers. A poorly developed fracture cleavage, defined as S3, is axial planar to F3 folds formed in schistose layers.

F1 folds identified during this study are equated with major recumbent folds described by Hepburn et al. (1984) and Hepburn (1975). These folds may also be equivalent to the nappe-stage folds identified by Thompson and Rosenfeld (1979) which occur directly east of the study area. F2 folds described above are similar in style and orientation to folds formed during back folding associated with dome stage deformation (Rosenfeld, 1968).



• F3Axial Plane Cleavage

Figure 3. Stereographic projection of poles to F3 kink band axial plane foliation. Arc labelled "Plane of intersection..." is the intersection of contoured F3 axial plane points. Note the orthogonal relationship of this plane to the trace of the Chicken Yard Line measured at Stop 1 (referred to as Sta. MB3 above).

F3 folds cannot be confidently correlated to folds described by previous workers. As Hepburn et al. (1984) state, post-F2 deformation was brittle and affected the same S-surface resulting in a variety of fold styles and orientations which are not easily correlated between outcrops. In addition, F3 folds observed in this study may result from more than one period of folding. Both of these factors may contribute to the apparent random orientation of F3 fold axes observed on this trip.

Separate plots of F3 kink bands, however, suggest an association between the kink bands and faulting coincidence with the Chicken Yard Line. Figure 3 suggests that the kink bands may be conjugate to the trace of some of the faults which occur along this horizon and therefore, may be genetically related. The significance of the horizon mapped as the Chicken Yard Line is a current source of debate. South of the area of this trip the horizon mapped as the Chicken Yard Line is interpreted as an unconformity (Hepburn et al., 1984). However, extensive shearing of rocks along this horizon exemplified at Stop 1 is consistent with the interpretation that one or more faults occur along this contact. Further work is necessary to determine the extent, relative motion and amount of offset along these surfaces.

METAMORPHISM

Three metamorphic events are documented by evidence obtained from rocks of the Standing Pond and Putney Volcanics. The first two "events" appear to represent a continuum in metamorphic conditions, separated by a deformational event. The third event is a retrograde event, the effects of which vary locally.

M2, the second metamorphic event, resulted in the equilibrium assemblages present in the rocks. This event was a medium-pressure facies series event which resulted in the gradual increase in metamorphic grade observed from east to west across the study area. Easternmost rocks, Stops 1 and 2, preserve evidence of greenschist facies metamorphism, while rocks to the west preserve epidote-amphibolite facies metamorphism (Figure 2). Pelitic rocks intercalated with the mafic rocks show evidence of chlorite through garnet grade metamorphism from east to west.

Between Stops 1 and 2, and below the garnet isograd of Doll et al. (1961) amphibole changes composition from actinolite to hornblende. Farther west, and within the garnet zone, plagioclase changes composition from albite to oligoclase (Figure 2). These compositional changes are consistent with changes in rocks metamorphosed in medium-pressure facies series conditions as stated by Miyashiro (1973, p. 250).

As the effects of M2 increase from east to west across the area covered by this trip, the mafic rocks become noticeably more coarse-grained and generally darker in color. Modal changes in the equilibrium assemblage include notable increases in the abundances of amphibole and hematite/ilmenite solid solution laths, and decreases in abundances of chlorite and titanite (Table 1). Biotite, epidote, plagioclase, quartz and carbonate also change in modal abundance; however, changes in modal amounts of these minerals may be a function of bulk composition and/or influences of different metamorphic reactions which occurred in the rocks.

Chemical trends concomitant with increasing metamorphic grade were determined by electron microprobe analysis. In the vicinity of Stop 2 and below the garnet isograd of associated pelitic rocks, amphibole changes composition from actinolite to hornblende (Figures 2, 4 and 5). Specifically, Al^{VI} , Fe^{3+} , A-site occupancy (Na(A) and K contents), Ti and Na (M4) increase while Fe^{2+} , Mn and Mg contents decrease (Boxwell, 1986, Figures 25-31). Farther west and within the garnet zone, plagioclase changes composition from albite ($\leq An_04$) to oligoclase ($An_{14} - An_{25}$). Three samples from the area contained both albite ($\leq An_05$) and oligoclase ($An_{14} - An_{25}$), allowing tight constraint of the oligoclase isograd (Figure 2).

TABLE 1
RANGE OF MODAL PERCENTAGES FROM SAMPLES OF STANDING POND
AND PUTNEY VOLCANICS DETERMINED FROM THIN SECTIONS

STOP:	1	2	2(int)	2A	3	4	5	5(A1)	6	6(A1)
QUARTZ	10-15	8-24	5-30	10-22	10-15	10-25	4-15	10-13	5-16	10-22
FELOSPAR	20-26	2-36	5-44	23-30	20-25	5-40	12-34	13-23	15-34	13-25
CARBONATE	2-5	5-25	12-32	10	3-20	0-25	5-20	7-20	3-12	2-19
CHLORITE	13-30	10-30	0-19	7-30	2-17	4-25	4-20	5-28	1-8	4-15
WHITE MICA	1-2	0-15	0-40	3-5	0-3	3-15	-	0-5	0-12	0-9
BIOTITE	-	0-18	0-15	0-3	1-16	5-30	1-10	10-15	2-7	5-10
EPIOTTE	7-15	0-20	0-5	2-15	3-6	<1-20	2-8	<1-7	1-8	<1-5
AMPHIBOLE	2-10	0-15	0-5	0-2	0-40	-	5-40	0-24	25-45	14-27
GARNET	-	0-2	-	-	0-10	-	3-10	0-5	0-10	1-10
OPAQUES	1-15	<1-12	1-7	<1	2-5	<1-5	3-7	3-6	2-7	2-5
TITANITE	8-12	0-15	0-5	13-15	2-7	0-8	2-5	2-3	0-10	0-2
ZIRCON	-	0-1	0-1	-	-	0-<1	0-<1	0-2	0-<1	0-1
TOURMALINE	-	-	0-1	-	-	-	-	0-<1	0-<1	-
APATITE	-	0-1	0-1	-	0-1	0-<1	0-<1	0-2	0-2	0-2
OTHERS	ST	P,R	R	-	-	-	-	F	R	-
Number	4	8	9	2	3	4	6	3	9	4
An Content	≤4	≤3			≤6				17-20	

Note: "Number" refers to the number of thin sections from which the range was derived. The symbol "-" is used if the mineral was not identified. Abbreviations used are ST- stilpnomelane, P- pennantite, R- rutile, F- fluorite. The opaque minerals identified were: iron oxide- ubiquitous, sulfide minerals in all samples except rocks from Stop 5, magnetite - Stops 1 and 5, pyrite- Stop 1, chalcopyrite- Stops 1, 3, 6 and hematite/ilmenite laths- all Stops except Stop 1. The letters "A1" refer to aluminous-rich rocks present at Stops 5 and 6. The term "int" refers to rocks at Stop 2 which are gradational between the volcanics and adjacent metasedimentary rocks. The range in An content was determined by electron microprobe analysis.

Other changes in mineral composition with increasing metamorphic grade are summarized as follows: the average Ti and Al (mostly Al^{VI}) contents in biotite increase. Fe³⁺ content in epidote decreases and Al^{VI} component increases. Although the composition of chlorite is variable across the field area, the composition of chlorite in equilibrium assemblages is more restricted and changes regularly with increasing metamorphic grade. In general, Mg, Al^{IV}, Fe³⁺ and Al^{VI} contents increase, whereas Fe²⁺ and Mn contents decrease.

As shown in Figure 4, data from this study plot within the medium- and low-pressure facies fields delimited using data of other mafic schists in Vermont (Laird et al., 1984; the fields and bars are amended as shown in Figure 4 to incorporate data from this study).

The fact that amphibole changes composition from actinolite to hornblende at lower grade than the change in plagioclase from albite to oligoclase occurs indicates that M2 was a medium-pressure facies series event similar to metamorphism preserved in rocks west of the CVGS in southeastern Vermont (Laird, 1980; Laird and Albee, 1981). North of the field area in rocks of the CVGS, low pressure facies series metamorphism is preserved.

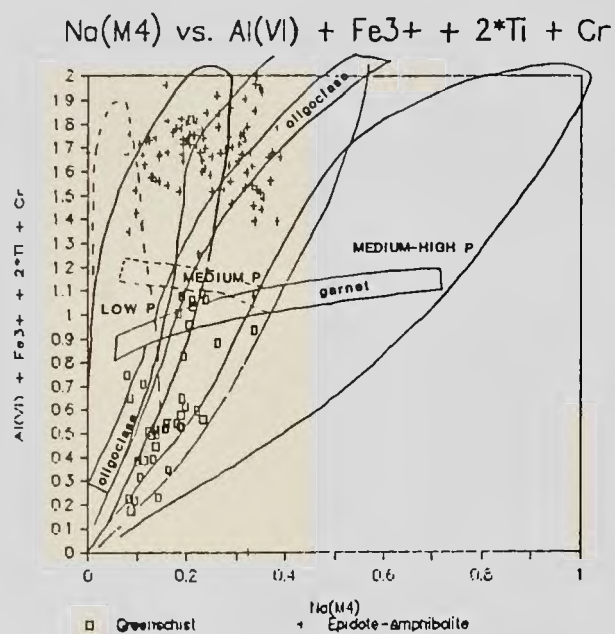


Figure 4. Plot of Al(VI) + Fe³⁺ + 2*Ti + Cr vs. Na(M4) formula proportion units in amphibole showing that analyses from the study area plot within the medium-pressure facies field delimited by Laird et al. (1984). Analyses plotting to the right of the oligoclase isograd are amphibole which coexist with albite; those plotting to the left of the oligoclase isograd represent amphibole which coexist with albite and/or oligoclase. (Dashed lines represent suggested revisional boundaries for the low-pressure facies field and garnet isograd bar based on data from this study.)

Figure 5 summarizes the compositional changes in the major phases described above. The increase in the Y-component ($AF_2O_3 - 3/4CaO - Na_2O$) is dominated by an increase in the Tschermak substitution ($Al_2Mg_{-1}Si_{-1}$). As can be seen in Figure 5, the MgO/FMO ratio in chlorite and amphibole spans a range of bulk composition yet the tie line orientations are consistent within samples of the same metamorphic grade. This observation, coupled with the fact that the change in compositions linked by the tie lines is similar to that first observed by Wiseman (1934) and later by Laird (1980, 1982) regardless of bulk composition, supports the idea that the relative increase in Fe²⁺/Mg ratio in amphibole and the decrease in the same ratio in chlorite is a function of metamorphic grade and not bulk composition.

A sample from Stop 2 (Figure 5b) appears to be intermediate in composition between greenschist facies and epidote-amphibolite facies zone samples. The gradual change in amphibole composition contradicts information supporting a miscibility gap in amphibole compositions between actinolite and hornblende suggested by many workers (e.g. Miyashiro,

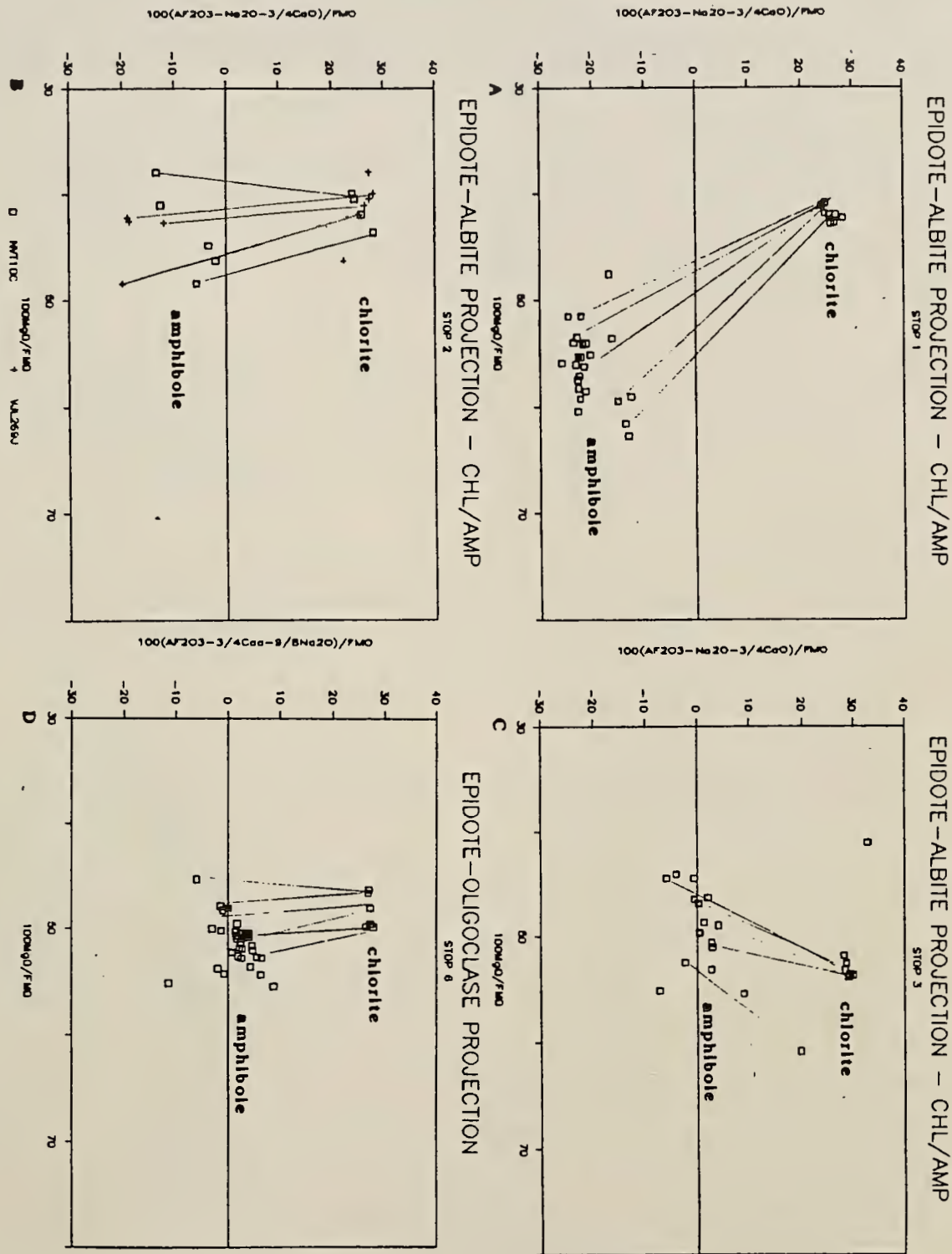


Figure 5. Projection of amphibole and chlorite from epidote and albite (or epidote and oligoclase) for sample from Stop 6 where the plagioclase is oligoclase). Lines connect points from the same sample, and are shown for illustrative purposes. Note the change in orientation of the tie lines from greenschist (5a and sample VJL269J - 5b) to epidote-amphibolite facies (sample NVT10C - 5b, 5c and 5d).

1958; Shido and Miyashiro, 1959; Brady, 1974; Cooper and Lovering, 1970; Doolan et al., 1973). As can be seen by data from sample NVT10C the miscibility gap is "bridged" by this composition. Textural evidence, such as exsolution lamellae, which might support the theory that actinolite and hornblende are immiscible (Grapes and Graham, 1978) is likewise lacking in these samples.

The pressure temperature regime inherent during the formation of the equilibrium assemblage may be discerned by comparing analytical data from rocks of the study area with information in the literature. Because none of the samples analyzed contained andesine, it may be assumed that conditions during M2 were not as intense as the upper limit of the "transition zone" described by Liou et al. (1974). The maximum temperature and minimum pressure of formation of the M2 epidote-amphibolite facies zone samples are therefore approximated at 560°C and 3.4 kbar.

The estimated conditions of metamorphism are consistent with estimates obtained using the plagioclase-hornblende geothermobarometer of Plyusnina (1982). The geothermobarometer yielded a temperature range from 405°C to 540°C and pressure from <2 to >8 kbar. The pressure of metamorphism inferred from the amount of aluminum present in amphibole in equilibrium with plagioclase of the respective composition ranges from <2 to >8 kbars. The maximum temperature is estimated for samples containing no titanite and is consistent with the 500°C to 540°C temperature range of the upper stability limit of titanite suggested by Moody et al. (1983).

Mineral growth in samples west of the Chester and Athens domes from the Moretown member of the Missisquoi Fm. and the Pinney Hollow Fm. (Laird and Albee, 1981) yield a temperature estimate of 500°C ± 25°C for the garnet-albite zone and 550°C ± 50°C for the garnet-oligoclase zone. These data are correlative with M2 mineral assemblages from the study area and empirically consistent with temperature and pressure estimates referred to above. Data from the study area are likewise consistent with solid piezothermometry studies of kyanite grade rocks from Gassetts, Vermont to the northwest which yielded 5.6 kbar and 545°C ± 20°C (Adams et al., 1975).

Evidence for the first metamorphic event, M1, is preserved within garnet and amphibole grains. Coarse-grained examples of these phases contain cores which formed prior to an F2 fold event. Textural evidence (e.g. abundance of inclusions in cores) and differing optical orientation of the cores suggest that the cores formed discretely from the rims.

Conditions inherent during M1 are not entirely discernible from the assemblages present. Equilibrium assemblages relict from M1 are incomplete, and therefore attempts to discern the conditions of metamorphism are hampered. However, based upon the fact that the composition of amphibole cores is the same as the composition of amphibole rims, one might suggest that the conditions of metamorphism were the same during M1 and M2.

A late retrograde metamorphism, M3, locally affected rocks within the study area. This event reached a maximum of biotite grade in western samples but affected rocks on a bed-to-bed basis and in some cases only affected microlayers within a specific rock. M3 occurred under less intense metamorphic conditions than M2, as evidenced by the lower grade assemblages resultant from M3. Specific conditions of M3 metamorphism may not be discerned, however, because of the lack of complete equilibrium assemblages.

An important conclusion to be drawn from this study is that the equilibrium assemblages indicate that the conditions inherent during the dominant metamorphism, M2, gradually increased across the study area. This hypothesis does not support the pattern of isograds presented on the Centennial Geologic Map of Vermont (Doll et al., 1961). The isogradic pattern on the state map suggests that metamorphic grade decreases from east to west across the study area, before gradually increasing to the west.

RELATIONSHIP OF METAMORPHISM TO DEFORMATION

The relationship between the timing of metamorphism and deformation within rocks of the field area is summarized in Table 2. Formation of an S1 schistosity is assumed to

have accompanied an F1 deformation. This early foliation is preserved by oriented inclusions within mineral grains of the equilibrium assemblage (e.g. amphibole and garnet) suggesting that the F1 fold event preceded the main metamorphism (M2).

Table 2

<u>Deformation</u>	<u>Metamorphism</u>
	M3 post-dates all deformation; sheet silicates aligned parallel to relict cleavage planes in pseudomorphs
F3 Kink folding in schistose rocks, open folds in more competent layers; folds F2 folds	
F2 Formation of dominant schistosity, folds F1; M2 minerals aligned parallel to F2 fold axes	M2 Formation of equilibrium assemblage. Varies from greenschist in the east to epidote-amphibolite facies in the west (chlorite to garnet grade)
F1 Formation of early isoclinal preserved in less competent rocks may predate M1 or be contemporaneous with it.	?? M1 Formation of amphibole and garnet cores which contain a foliation which predates F2

Table 2. Summary of the relationship of metamorphism to deformation.

The main metamorphic event, M2, was contemporaneous with an F2 fold event. The long axes of amphibole grains are parallel to F2 fold axes measured in the field (Stop 6). Rotated garnets (Stops 3 and 6) contain concentric inclusion trails indicating that the grains formed during deformation as suggested by Rosenfeld (1968 and 1972). Radial splays of amphibole found in garbenschiefer may have formed in an area of reduced pressure during this deformation. The fact that some of the amphibole splays are curved as well suggests that the grains formed contemporaneously with folding. Also, the main foliation is deformed around porphyroblastic grains of the equilibrium assemblage while other grains of the equilibrium assemblage overgrow the main foliation suggesting that the main metamorphism and deformation were simultaneous.

F3 folds deform the main schistosity which formed during F2; therefore, F3 must post-date F2 and M2. The late retrograde metamorphism, M3, appears to post-date all deformation. The position of sheet silicate grains parallel to relict mineral cleavage planes in pseudomorphs (e.g. chlorite and biotite after amphibole) attests to the interpretation that the M3 retrograde metamorphism occurred after deformation ceased. Sheet silicates which overgrow foliation are interpreted as products of M3 metamorphism.

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Mileage0.0 STOP 1

The first stop consists of three roadcuts which are located on the north side of Route 11, and extend almost continuously from the driveway into Howard Johnson's restaurant westward for approximately .3 miles to Paddock Road. On the east end of the first roadcut are chlorite grade rocks of the Littleton Fm. The fine-grained, gray-green weathering, epidote-carbonate schists display primary layering and a west over east fold sense. F3 generation kink bands are superposed on F2 folds of primary layering and S1. The orientation of primary layering in the Littleton Fm. here is northeast trending, westward dipping.

Approximately 200 feet west of the eastern end of this outcrop is the horizon referred to as the Chicken Yard Line (CYL) which marks the boundary between the Vermont and New Hampshire Sequences in southeastern Vermont. In this roadcut, the CYL occupies a zone less than 6 feet wide and is a light colored, tan to gray, fine-grained, sugary textured rock in which distinct lithons less than four inches in length may be seen. The trace of the CYL here trends north-east and dips nearly vertical and is clearly at a high angle to layering observed in rocks of the Littleton Fm. The nature of the CYL at this locality appears to be a shear zone, and is expected to inspire some debate on this trip.

The Putney Volcanics make up the western end of the first roadcut as well as the entirety of the two western roadcuts of this stop. Here the Putney Volcanics are fine-grained, massive, dull green-gray weathering greenstones. Color and texture vary due to subtle changes in mineralogy and mode (e.g. amount of epidote, presence of carbonate and amount and species of sheet silicates). Sulfide porphyroblasts and epidote knots are abundant in some layers. The amphibole is actinolite and very fine-grained. Layering in the Putney Volcanics is indistinct, similar to Stop 2 and different from Stops 2A to 6, where well defined layering is visible.

Continue west on Route 11 towards Springfield.

0.75 The eastern roadcut of Stop 2 begins on the south side of Route 11.

1.0 Turn right (north) off Route 11 onto road leading over the Black River. Park here on the side of road and cross Route 11 to examine rocks of Stop 2.

STOP 2

Two roadcuts extending 0.25 miles along the southern side of Route 11 comprise Stop 2. Here metasediments of the Waits River and Gile Mountain Fms. are intercalated with the Standing Pond Volcanics. According to Doll et al. (1961) the Waits River Fm. occupies the western end of the eastern roadcut and the Gile Mountain Fm. occupies the western end of the western outcrop. Contacts of both of these formations with the Standing Pond Volcanics, which comprise the remainder of both outcrops, are gradational and therefore not easily pinpointed. The metasedimentary units at this locality are biotite grade. The Standing Pond Volcanics here are similar to the Putney Volcanics seen at Stop 1 with respect to lack of distinctive layering and massive appearance. The Standing Pond Volcanics are weathered light to medium gray-green or brownish and commonly contain rusty pits where carbonate and sulfide grains have weathered out. The fresh surface varies in color due to differences in composition; lighter layers are comparatively felsic, green layers, chlorite- and biotite-rich and punky brown layers are carbonate-rich. Unlike the Putney Volcanics seen at the previous stop these rocks have a smaller mafic component and lack epidote stringers. Amphibole from the eastern roadcut is hornblende which allows classification of the rocks as

epidote-amphibolite facies (e.g. after Miyashiro, 1972). The fold style of F2 here is similar to that seen at Stop 1; however, F2 trends southwest instead of northeast. Because the plunge of F2 is shallow ($\leq 30^\circ$) the change in trend may result from porpoising of F2. F3 folds here appear as large warps of layering with little or no widely spaced crenulation cleavage axial planar to the fold surface. The rock which Aleinikoff (1986, pers. comm.) dated is approximately 100 feet from the western end of the western outcrop.

- 1.3 Turn left (north) up steep hill, road turns into a dirt road near the top of the hill.
- 2.7 Turn left (northwest) on Old Crown Point Road just after pavement begins.
- 3.6 Turn left (west) at intersection onto Old State Route 10.
- 3.8 Turn right (north) onto Eureka Street.
- 4.0 Farmhouse on west side of road belongs to Rufus Estes. If you are following this road log after the NEIGC trip, please ask his permission to park off the road and look at the outcrops in the pasture on the east side of the road.

STOP 2A (Optional Stop)

Many different layers of epidote-amphibolite facies zone Standing Pond Volcanics exist in the field which are continuous with those seen at Stop 3 (see below). The purpose of this stop is to examine the nappe-stage fold displayed in cross section in one of the pasture outcrops. The fold can be viewed on the southern face of an outcrop of medium gray-green, massive greenstone near the southeastern edge of the height of land in the pasture. The fold is an isoclinal upright fold with an undeformed axial planar cleavage. It appears to be similar in style to the F1 nappe-stage folds of Rosenfeld (1968 and 1972) yet has an undeformed axial planar cleavage suggesting that it might be an F2 generation fold.

- 5.5 Turn left (west) on Barlow Road.
- 6.2 Turn right (north) on Old Crown Point Road.
- 6.3 Outcrops on the east side of road are biotite-grade metasedimentary rocks.
- 7.1 Pull off on the east side of the road where a dirt road leads into the pasture. Walk back to farmhouse on west side of road and ask permission of Dick and Helen Moore to enter the pasture to examine the rocks, if you are following this road log after the NEIGC trip.

STOP 3 [PLEASE! NO HAMMERS AT THIS STOP]

Outcrop on the west side of Old Crown Point Road is biotite grade Gile Mountain Fm. The knots on the surface are identified petrographically as chloritized garnet. The chloritization is interpreted as having occurred during M3. The garnets may be M1 or M2 generation. Continuing into the pasture, a wide variety of layers of Standing Pond Volcanics can be seen. Westernmost layers are felsic, light colored, and contain biotite and chlorite but no garnet or amphibole. Farther east in the field are dark-colored, medium-grained amphibolites that may or may not contain garnet. The amphibole is hornblende, and the plagioclase albite. Based on the mafic rocks therefore, the rocks here are classified as epidote-amphibolite facies. One must wonder why the pelitic rocks have been affected by the late retrograde metamorphism, M3, yet the mafic rocks do not appear affected by M3 at all.

The amphibole in most samples is aligned parallel to the foliation. The main schistosity here is F2 and because the amphibole lies in the plane of schistosity one may conclude that the amphibole formed contemporaneously with the

formation of the S2 schistosity.

- Continue north on Old Crown Point Road.
- 7.3 Turn left (west) at the T-intersection. At the southwest corner of this intersection is another outcrop of Gile Mountain Fm. which contains garnet replaced by chlorite.
 - 7.9 Turn left (south) at the T-intersection and follow this road into downtown Springfield.
 - 10.4 Turn left (south) on Main Street - Route 11.
 - 10.8 Cross over the Black River and stay to the right (west; do not follow Route 11).
 - 10.85 Go straight (west) at the lights - up the hill.
 - 11.1 At the top of the hill bear left (south).
 - 13.2 Continue south at the intersection at Hardscrabble Corner.
 - 13.3 Bear right at the fork in the road; continue on the paved road.
 - 18.3 Pass under the railroad overpass and pull off the road to the east. The best outcrops are on the east side of the bridge along the river.

STOP 4 Beware of the Poison Ivy.

The rocks along the river bank are primarily mafic schists of the Standing Pond Volcanics. Units vary from dark greenish-gray, weathering, schistose mafic rocks to black and white laminated, massive "amphibolites" with felsic porphyroblasts. In thin section such differences are obscured by the effects of M3 metamorphism. All of the amphibole has been replaced by biotite, chlorite, epidote group minerals, carbonate and felsic minerals. Despite the pervasiveness of the effects of the retrograde metamorphism these rocks may have been subjected to epidote-amphibolite facies metamorphism during M2. The coarse-grained amphibole pseudomorphs may indicate that the amphibole was hornblende in composition. In the absence of analytical data and because the surrounding rocks are greenschist facies zone rocks, these rocks are included within the greenschist facies zone also.

Downstream from the mafic rock is a fine-grained, massive, calcareous, micaceous quartzite of the Waits River Fm. This biotite grade sample is located near the contact with the Standing Pond Volcanics according to Doll et al. (1961), but unfortunately the contact is not visible.

F2 generation folds are closed folds of layering and S1 which have a cleavage developed that is axial planar to F2. F2 folds plunge moderately to the southwest.

- Continue south across the bridge.
- 12.6 Turn right (west) onto Route 103.
 - 20.2 Turn left (south) onto Pleasant Valley Road.
 - 26.6 Turn left (east) at the T-intersection onto Route 121 towards Saxtons River.
 - 26.9 Continue straight on Route 121 (don't turn right to go over the river).
 - 27.0 Turn right (south) onto side street and head towards the river.
 - 27.05 Turn right (west) and park in the parking area near the river.

STOP 5 (Stop 10 of Rosenfeld, 1972).

The outcrops along the river here are spectacular and afford almost a complete summary of both the structure and metamorphism seen thus far on the trip. The Standing Pond Volcanics at this stop, as mapped, form the southern hinge of a recumbent fold (Doll et al., 1961; Figure 2) and therefore, units should be repeated across the axis of the fold. However, an increase in

metamorphic grade within these exposures makes assessment of lithic continuity across the fold difficult to discern.

Beginning in the outcrop southwest of the parking area downstream from the bridge, three generations of folds are visible. F1 is seen in a carbonate-rich layer folded back upon itself. F2 appears as open to closed folds of layering and S1. F1 and F2 seen here are nearly parallel; F2 plunges moderately to the south. F3 open folds and kinks of S2, S1 and layering are visible in this outcrop also. F3 plunges moderately also, but in a southwesterly direction.

Rocks east of and downstream from the bridge are biotite grade, greenschist facies samples. The metasedimentary rocks are predominantly light-colored, micaceous schists. The contact between the metasedimentary rocks and the Standing Pond Volcanics is gradational. Rocks containing a predominant mafic component are present west of (upstream from) the bridge. West of the bridge the metamorphic grade changes to garnet grade or epidote-amphibolite facies.

The Standing Pond Volcanics, as seen in samples upstream from the bridge, are medium- to coarse-grained, brown-weathering and predominantly massive although some layers appear schistose. Some of the massive rocks appear to contain discrete laminae of mafic and felsic compositions. Others contain garnet, and still others contain approximately 20% carbonate minerals. A question of many workers in this area is whether the layers represent original layering or are a product of metamorphism.

Back track on Pleasant Valley Road to Route 103.

- 27.10 Turn left (north) back towards the center of Saxtons River.
- 27.15 Turn left (west) onto Route 121.
- 27.5 Turn right (north) onto Pleasant Valley Road.
- 33.9 Turn left (west) on Route 103.
- 34.0 Turn right (north) towards Brockways Mills.
- 34.4 Turn right (east) at T-intersection.
- 34.8 Turn off the road at the clearing, park and cross the railroad tracks into the field. Cross the field and go into the woods at the southernmost corner of the field. Follow the path down to the river.

STOP 6

Streamcuts here are Standing Pond Volcanics in contact with rocks of the Waits River Fm. to the west (upstream) and rocks of the Gile Mountain Fm. to the east (downstream). The contact with garnet grade calcareous schist of the Waits River Fm. is very sharp and well-exposed. The contact between Standing Pond Volcanics and garnet grade, micaceous schist of the Gile Mountain Fm. is not exposed.

Directly east of the contact with the Waits River Fm. is a spectacular example of fascicular schist or "garbenschiefer". Splays of amphibole cover the rock, some of which emanate from two-inch diameter garnet porphyroblasts and radiate in 360° . Some amphibole grains within individual fascicles appear curved also. Within garnet porphyroblasts concentric and sigmoidal inclusion trails are visible. These patterns have been described by Rosenfeld (1968, 1972) and are used to interpret the rotational directions of the rocks during deformation.

Downstream from the fascicular schists are other well defined layers of Standing Pond Volcanics. Some layers are very micaceous and appear similar to the garbenschiefer yet contain no garnets. Farther downstream massive, laminated, green-gray and white weathering amphibolites, some of which contain rusty pits where carbonate has weathered out, are present. Near the falls, is a layer which weathers orange colored and is very light on the fresh surface. Plagioclase crystals are easily visible and predominate in this felsic rock. This layer appears similar to the felsic layer observed at Stop 3 yet is farther east in the layering sequence at this stop than was the felsic layer seen at Stop 3.

Downstream from the falls is a layer of mafic rock which is laminated black and white. On the eastern side of the pool (downstream) are micaceous schists of the Gile Mountain Fm. Medium-grained garnet knobs are present in this rock.

Microprobe analyses of a sample of laminated, dark green-gray weathering amphibolite that does not contain garnet allowed classification of the amphibole as hornblende and the plagioclase as oligoclase ($An_{17} - An_{20}$) which is consistent with epidote-amphibolite facies zone classification (Figure 2).

Many of the laminae within layers at this stop appear to be cut by an S2 schistosity. F2 folds plunge moderately to the northeast, and are generally open folds. The long axes of amphibole grains lie parallel to the plane of S2 and in some localities the long axes of amphibole grains are parallel to axes of F2 folds. The curved fascicles of amphibole and rotated garnet grains are interpreted as indicating that F2 and M2 were contemporaneous.

Primary layering is visible in micaceous schists of the Waits River Fm. upstream from the fascicular schist. Graded beds within the layers indicate that rocks to the east stratigraphically overlie those to the west. This is consistent with interpretations of Fisher and Karabinos (1980) in which they conclude that the Gile Mountain Fm. overlies the Waits River Fm.

To get to Northfield, return to Route 103 and turn left (east) on Route 103. Follow Route 103 to Interstate 91 North. Take I-91 North to the Northfield Exit.

GEOLOGIC AND HYDROLOGIC EFFECTS OF A CATASTROPHIC FLOOD IN THE COLD RIVER, SOUTHWESTERN NEW HAMPSHIRE

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INTRODUCTION

The southwestern portion of New Hampshire is characterized by small, steep-gradient watersheds which drain into the Connecticut River. The valley sides in these watersheds are exceptionally steep, are often composed of fine-grained impermeable till, and are subject to frequent slumps and slides. Although there are terraces of sand and gravel locally, they have little moderating effect on the flashy characteristics of the watersheds. Intense summer rainstorms have produced the record runoffs in this area. One such event occurred in August, 1986, and the effects of this storm are the subject of this field trip.

On August 7, 1986, about 6 inches (15 cm) of rain fell within a 2-hour period in a portion of the Cold River watershed in the towns of Acworth, Alstead, and Langdon, New Hampshire (Fig. 1). This intense rainfall was concentrated in roughly circular area of about 30 mi² (78 km²) located in the middle third of the 100 mi² (260 km²) watershed of the Cold River.

GEOLOGIC EFFECTS OF THE 1986 FLOOD

Surface runoff from the steep hillsides of this watershed formed new gullies, widened and deepened existing streams. Numerous road were washed out (Stops 1 and 3) and at least one home was severely damaged.

In one such gully, erosion exposed a complex section of saprolite overlain by till (Stop 5). The saprolite is developed on sulfide-bearing andesite (Ammonoosuc volcanics) and is overlain by a deeply weathered diamicton, probably colluvium, although it does somewhat resemble till. Exposures of saprolite do exist in other parts of New England and Quebec, but are rare enough to be noteworthy (LaSalle et al., 1985). The concensus of opinion about these saprolite exposures, especially those such as that at Stop 5 that underlie till, is that they were formed during the Tertiary Period. The clay mineralogy of the sub-till saprolites described by LaSalle et al. (1985) suggested to these authors that they were formed under a warmer and wetter climate than either interglacial or Holocene conditions. These same authors presume that saprolite was widespread in New England and Quebec prior to glaciation and that the known exposures are remnants of this extensive soil cover.

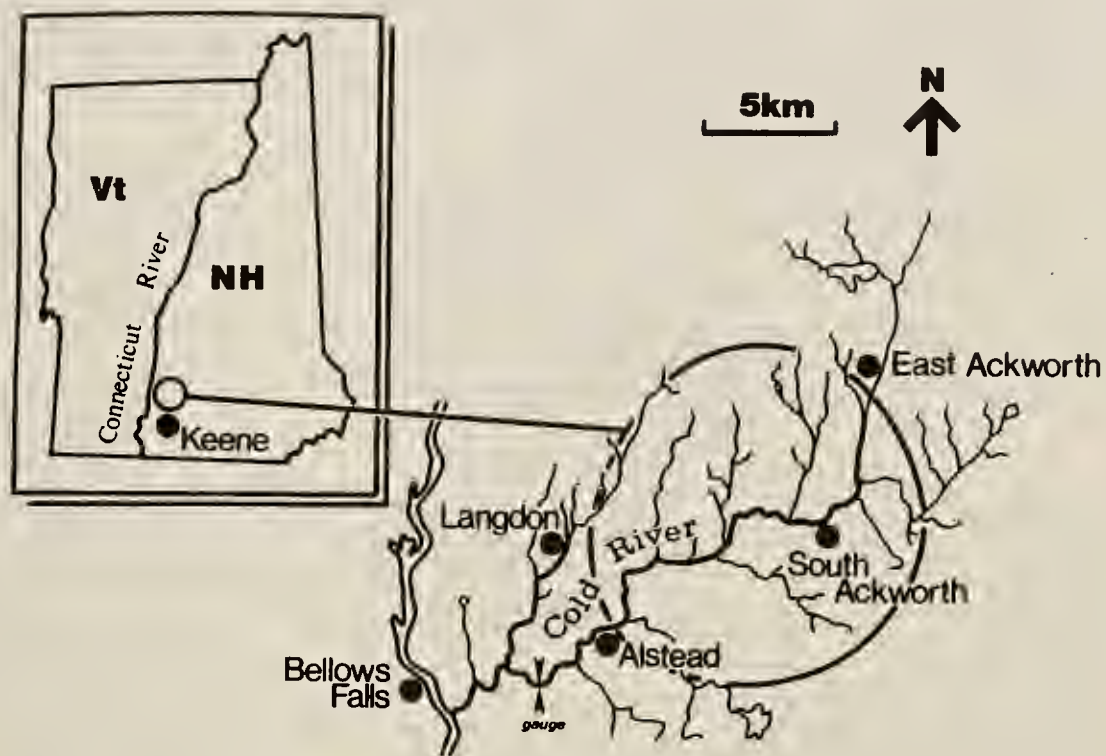


Figure 1. Location of the Cold River watershed. Area of intense rainfall during August, 1986 flood is roughly outlined by circle.

Deposits of the 1986 flood

At the base of many hillside gullies, coarse gravel was deposited on alluvial fans on the edges of flood plains (Stop 4). The fans existed prior to the 1986 flood and appear to consist largely of material similar to that deposited in that flood. It may thus be possible that events similar to the 1986 flood were responsible for the formation of these fans.

Coarse gravel nearly fills the channel of at least one tributary stream near its juncture with the Cold River (Stop 6). This gravel was graded to the flood level of the Cold River, and forms a kind of delta within the channel of the tributary and is one of a number of records of the stage of this flood. The gravel in this delta was eroded from within the watershed of the tributary and thus represents a deposit that is different from the slack-water deposits described by Kochel and Baker (1982) in which sediment is carried into a tributary from the trunk stream.

The flood waters generated within the central portion of the Cold River had major effects far downstream. At the mouth of the Cold River, flood waters estimated to be about least 10-feet (3 m) deep, with a velocity 5 feet (150 cm) per second, moved into the Connecticut River. Gravel clasts up to 15 inches (400 mm) in intermediate diameter was deposited as a delta which prograded nearly three-fourths of the way across the Connecticut (Stop 8). When the flood waters lowered and the delta was exposed, the Connecticut was confined in a channel about 100 feet (30 m) wide. The very high flow and associated turbulence of this flow caused extensive bank erosion on the Vermont side of the river.

On the delta surface, a number of features indicated that the flow during this flood was in a direction that combined the flow from the Cold River and the Connecticut (Fig. 2). These flow indicators include shingled gravel, oriented trees and scour pits around the base of the trees, and gravel bedforms with crests spaced about 8 feet (2.4 m) apart (Fig. 2). Subsequent floods in January and April, 1987 have moved the trees, wiped out the bedforms, and reoriented the clasts of the delta surface. The delta has not moved measurably downstream, but the distal end of the delta has been eroded, widening the channel. At this writing (July 6) the delta has reappeared, as spring and early summer high water has receded. There is a small channel between the delta and the river bank (Fig. 2), and the delta can be reached only by a knee-deep wade.

HYDROLOGICAL EFFECTS OF THE 1986 FLOOD

About 2 weeks prior to the intense rainfall, an observation well had been placed near the bottom of a kettle in a small delta. A steep bedrock hill rises above the delta surface. Figure 3 is the hydrograph of this well that covers the period of the August, 1986 flood. As may be seen in the summary of rainfall in Figure 3, abundant rain had fallen prior to the major storm. This figure also shows that the water table when the well was installed was over 40 feet (12 m) below the ground surface. Within a few weeks of the August storm, recharge had found its way from the surface to the water table and eventually added about 9 feet (2.7 m) to the saturated thickness of the aquifer and as quickly the water was dissipated. The aquifer is abruptly terminated to the

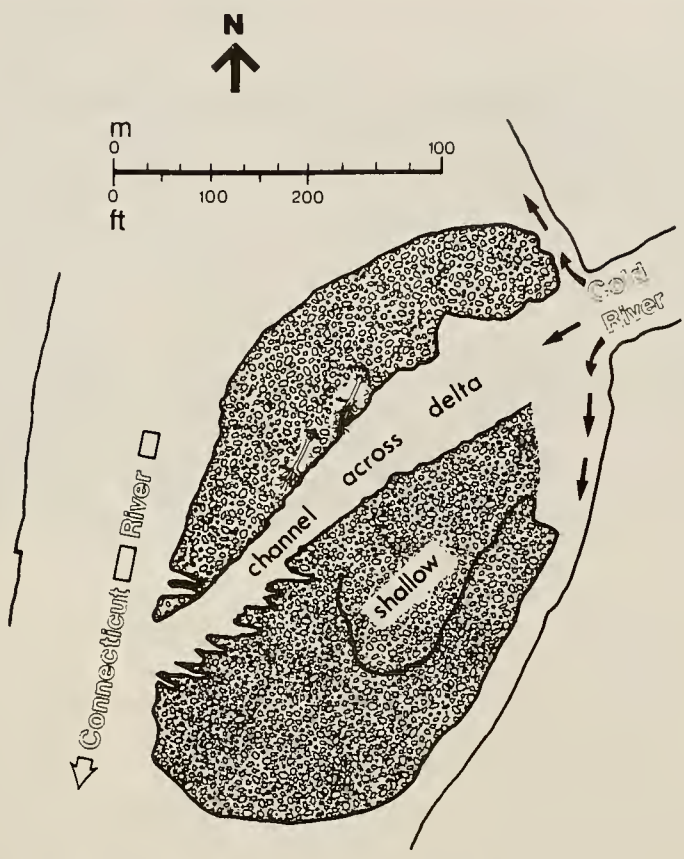


Figure 2. Map of the delta deposited at mouth of Cold River during August, 1986 flood (See Figure 1). Trees on delta surface and gravel bedforms (jagged edges of distal end of delta) indicate flow direction during flood. Plane table and alidade map by D.W. Caldwell and Ed Kelly, September 20, 1986.

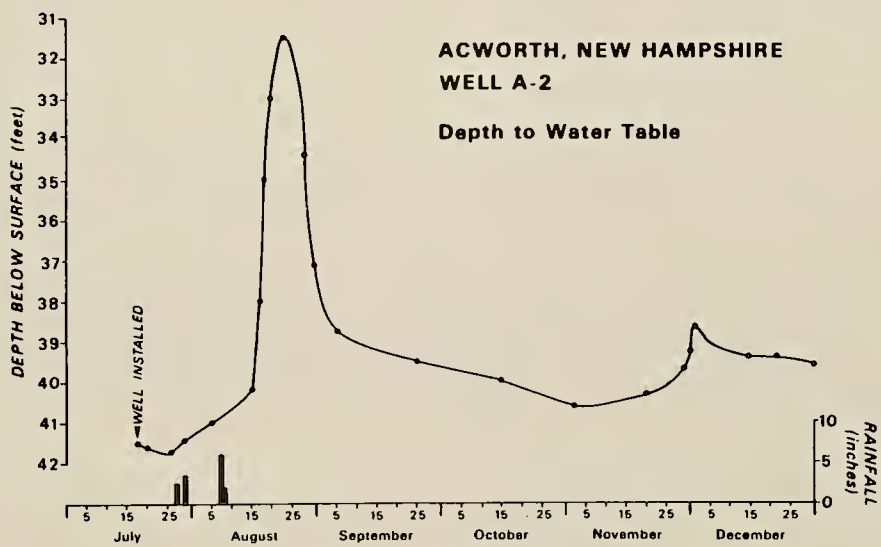


Figure 3. Hydrograph of observation well showing recharge during flood of August, 1986. Bar graph at bottom of figure indicates rainfall during period July 26 - August 8, 1986. Note the very rapid rise and recession of water table. The recharge of the aquifer was a result of direct rainfall combined with runoff from steep hillside that was concentrated into a kettle in a glacial lake delta.

north, and groundwater is discharged in a series of contact springs along the base of the valley wall, more than 100 feet (30 m) below the delta surface. Using the concept of Specific Yield (Sy), we may estimate the amount of water recharged during the event:

$$S_y = \frac{V_{wa}}{V_a} \quad (1)$$

where S_y is Specific Yield (or effective porosity), V_{wa} is the volume of water added (or released) to storage, and V_a is the volume of aquifer into which the water is added or released. This analysis conceives of a unit prism of the aquifer 1 foot (or other unit) square in cross section and with a length equal to the change in water table elevation. Because this analysis uses a unit prism, volume and length are equal. Rearranging equation 1 to solve for V_{wa} , and assuming a reasonable Specific Yield of 0.20,

$$V_{wa} = S_y \times V_a \quad (2)$$

and $V_{wa} = 0.2 \times 9 \text{ feet} = 0.2 \times 2.7 \text{ m}$
 and $V_{wa} = 1.8 \text{ feet or } 21.6 \text{ inches} = 548 \text{ mm}$

This analysis indicates that the aquifer in the vicinity of the kettle received about 21 inches (548 mm) of recharge while about 6 inches (150 mm) of rain were falling. It should be remembered that other rain preceded and followed the principal storm on August 7 (Fig. 3). The total of all the rainfall in late July and early August was about 12 inches (305 mm). We believe the excess recharge occurred when runoff from the steep hillside above the aquifer was concentrated into the kettle. This secondary recharge was not observed in other parts of the watershed. We believe that in situations like this one, unless secondary recharge from steep hillsides is considered total aquifer recharge may be underestimated.

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- LaSalle, Pierre, DeKimpe, C.R., and Laverdiere, M.R., 1985, Sub-till saprolites in southeastern Quebec and adjacent New England, in Borns, H.A., LaSalle, P., and Thompson, W.B., eds., Late Pleistocene history of northeastern New England and adjacent Quebec: Geological Society of America Special Paper 197, p. 13-20.

ROAD LOG, COLD RIVER FIELD TRIP

Field trip meeting place is on Route 12 in North Walpole, New Hampshire, about 1 mile south of Bellows Falls, Vermont. From the north or south, take exit 6 from Interstate 91, go south on Route 5 to Bellows Falls and follow signs for Route 12. Meeting place is 1/2 half mile south of Green Mountain Railway roundhouse at traffic light. Turn right on off-ramp to Connecticut River. From Boston area take Route 140 off Route 2 near Gardiner to Route 12. Follow Route 12 through Keene to North Walpole. From eastern New Hampshire and from Maine follow Route 101 from Portsmouth to Keene and then Route 12 to North Walpole. As parking is somewhat of a problem, we may want to consolidate into as few vehicles as possible. We will pass this way to Montpelier this afternoon and can retrieve vehicles left here.

Mileage	Road Log
0.00	Leave off ramp, south on Route 12.
0.20	North Walpole Gauge. Mouth of Saxon's River across Connecticut.
0.50	Turn left on Route 123. Mouth of Cold River.
0.70	Turn left onto Cold River Road. Glacial Lake Hitchcock delta across valley.
1.5-1.9	Whitcomb Sand and Gravel across Cold River.
2.6	<u>STOP 1.</u> Mouth of Great Brook. Cold River Road was washed away in flood by large eddy in river at this point. We are nearly 3 miles downstream from area of intense rainfall in August, 1986 storm. There was no evidence of large flows from Great Brook itself. Continue along Cold River Road.
3.5	Intersection with Route 123. Cross Route 123 and park on right for Optional stop. <u>STOP 2.</u> (Optional Stop). Drewsville Gauge, discontinued by USGS in 1978. Operated by Department of Geology, Boston University since 1980. Inclined staff gauge allowed estimation of 1986 flood stage of 13.00 feet. By extrapolation of stage - discharge rating curve this is equivalent to about 10,000 cfs. This exceeds previous record stage of 12 feet, and compares with the stage of 11 feet during April 1987 floods. Note gorge below bridge and numerous bridge abutments. Return to cars and proceed south (east) along Route 123.
5.5	Juncture of Route 123 with Route 12A. Turn right, cross Cold River and bear left through village of Alstead. Note till cut in stream channel on left, behind playground. This is near the downstream (western limit) of intense rainfall and erosion during August 1986 flood.
6.1	Turn left at Gulf station to Route 123A.

- 6.50 STOP 3. Park on left. Note gravel deposits in woods beyond pool. This is result of both the 1986 and 1987 floods. Walk a few hundred feet along road. Note abrasion of trees in ditch on left which resulted from transport of coarse sediment in floodwaters. Cross Route 123 to juncture with dirt road. This road was completely washed away in flood, leaving a channel about 30 feet wide and 10 feet deep. The road was rebuilt and the road replaced with a 12-inch culvert. Return to cars and proceed along Route 123A.
- 8.60 Cross Cold River, find parking on shoulder.
STOP 4. Alluvial fan on edge of valley. Cobble size gravel was deposited during intense runoff in August, 1986. Numerous fans like this one were also covered with gravel in this flood.
- 9.40 Turn right on Route 123A. Note small slide scars on left.
- 9.50 STOP 5. Find parking along right side of road and return to clearing on left side of road with blue sap lines. Large gully cut to bedrock during August, 1986 storm. Exposure on east side of gully near large maple tree shows the following stratigraphy:
3 feet laminated grey till
2 feet reddish-brown diamicton, probably colluvium
1 foot yellow-brown saprolite
bedrock (Ammonoosic volcanics)
Other exposures of saprolite along gully channel up the hill.
Return to vehicles and continue along Route 123A.
- 10.00 Turn right. Home of Dr. and Mrs. George Hanson. We will have lunch here and also visit Stop 6. Pull past house and barn and park in indicated area.

Stop 6 is within walking distance of the Hanson house. Walk along path to Cold River and along bank in upstream direction to juncture with Millikens Brook.
STOP 6. Cobble and boulder gravel was deposited by Milliken Brook in its own channel as the flow was slowed by flooding in the Cold River. This channel deposit was graded to the level of the flood stage of the Cold River during August, 1986 flood. This deposit is in effect an in-channel delta. This stream now flows between coarse channel deposit. During spring flood in 1987, the channel was so diminished in capacity that water was diverted onto flood plain.
- 10.01 Return to Hanson's for afternoon stops (2). Return to 123A, turn right.
- 11.4 Turn right and cross one way bridge.
- 12.0 Park in lot of Acworth highway garage. Washed out road on left leads to Beryl Mountain pegmatite quarry. Continue along road and descend dirt road on right.
STOP 7. Washover fan from steep slopes of Beryl Mtn. covers road and continues to floor of kettle on right. Observation well beyond

fan experienced rise in water table of about 9 feet within a few weeks of the flood. Analysis of the hydrograph of this well (see Fig. 2) and assuming a specific yield of 0.2, indicates as much as 1.8 feet of water were recharged into the aquifer from the kettle. This suggests that water from the steep slopes above was concentrated into the kettle and this water plus the direct rainfall accounts for the high recharge.

Return to vehicles. Continue ahead on road.

- 12.2 Acworth dump. Uncollapsed delta surface exposed in rear. Bedrock on left side of road.
- 13.0 Turn right at mineral shop.
- 14.2 Turn left on Route 123A
- 17.0 Juncture with Route 123 and Route 12A. Straight ahead.
- 17.8 Turn left on Route 123.
- 19.6 Turn left at blinking light on Route 123.
- 19.9 Lake Hitchcock delta surface.
- 20.3 Pits on right expose foreset beds in Hitchcock delta.
- 21.2 Descend foreset slope of Hitchcock delta.
- 21.8 Turn right.
- 22.0 Cross Cold River.
- 22.3 Turn right on Route 12 and almost immediately turn left onto grassy area beside road. Be careful of crossing Route 12. Walk south along roadbed of abandoned railroad. Upon reaching bridge abutment, descend slope to right. Careful of poison ivy. Reach bank of Connecticut River.
- STOP 8. Delta in Connecticut River was deposited from coarse sediment carried down the Cold River during the flood of August, 1986. "Paleo" flow indicators included oriented trees and gravel bedforms, as well as shingled gravel (Fig. 2). Debris in fallen trees indicated water in Cold River was about 10 feet deep when delta was deposited. Subsequent floods in January, 1987 and April, 1987 have eroded the distal end of the delta, but the delta has stayed pretty where as it was formed. The channel between delta and Vermont side of river has been widened by erosion on both sides. Soon after delta was formed, this channel was about 100 feet wide. Return to vehicles.

On Friday trip: turn north (left) on Route 12. At second light, turn left over Connecticut River, then right and find Route 5. Follow Route 5 for about 3 miles to I-91.

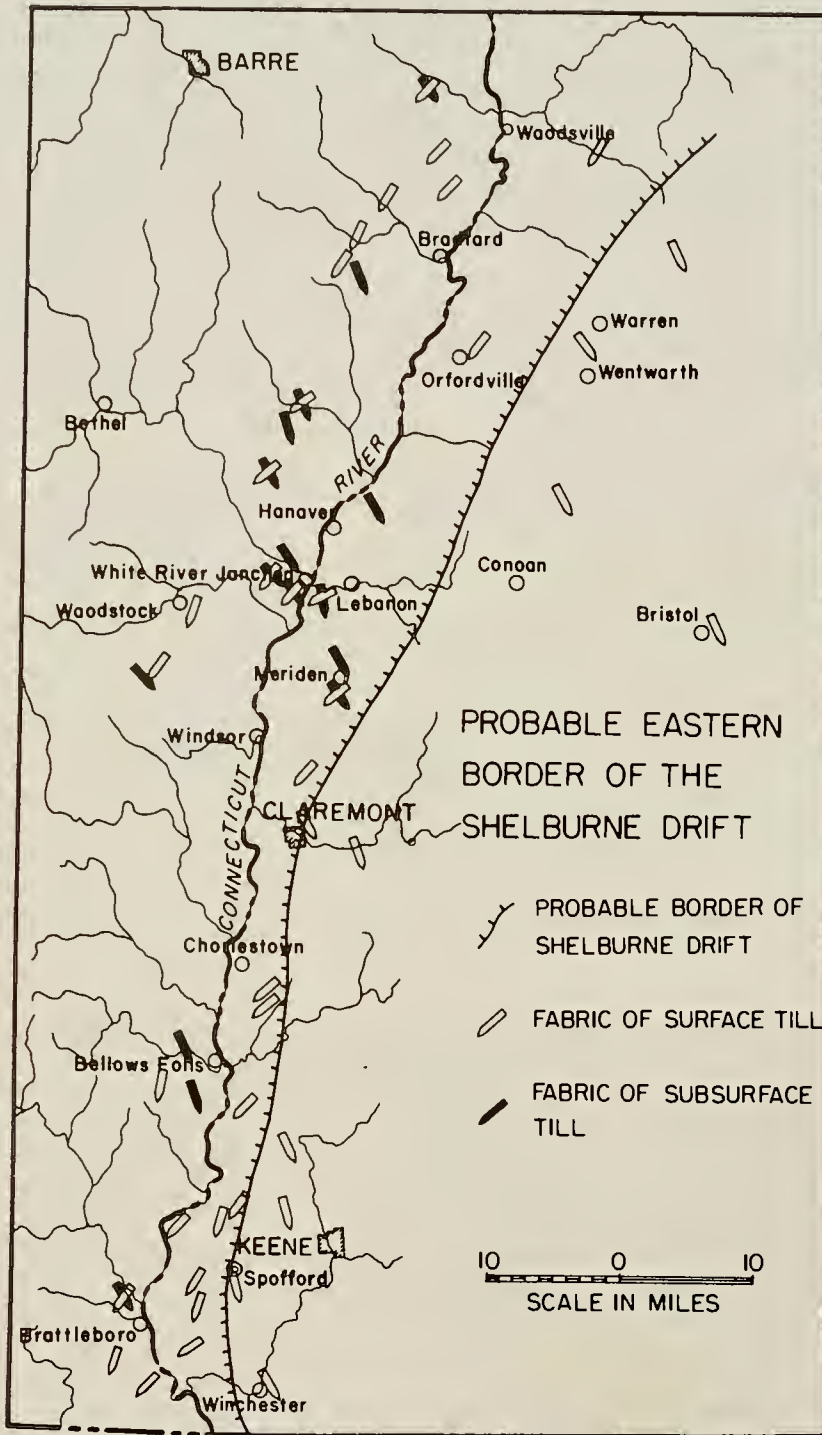


Figure 1. "Probable eastern border of the Shelburne drift". According to Stewart and MacClintock (1969, p.99), the Shelburne drift presumably was derived by ice moving from the northeast. Note that surface-till fabrics west of the border between Winchester and Charlestown indicate that the Shelburne drift came from older drift east of the border. That situation would be highly unlikely if not impossible.

GLACIAL LAKE HITCHCOCK IN THE VALLEYS
OF THE
WHITE AND OTTAUQUECHEE RIVERS,
EAST-CENTRAL VERMONT

by

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INTRODUCTION

There are two main purposes of this field trip. One is to review Late Wisconsinan glacial stratigraphy at two exposures in West Lebanon, New Hampshire, where outwash older than the last glacial advance can be observed. The other is to address the question of the relative ages of glacial Lake Hitchcock, glacial Lake Winooski, and the Champlain Sea. Field trip stops will be made on the following U.S.G.S. 7.5-minute quadrangles: Hanover, Vt-NH; Quechee, Vt; Sharon, Vt; Randolph, Vt; and Brookfield, Vt. In addition, the field trip passes through the South Royalton, Vt, and Randolph Center, Vt, quadrangles. Erosion of complex metamorphic rocks by streams and continental ice sheets has produced a rugged, hilly topography with 800 to 1,000 feet of local relief. Drainage is controlled by the Connecticut River which flows south-southwest through the area. Three major tributaries of the Connecticut River in this area are the White, Ottauquechee and Mascoma Rivers.

The area probably has been covered by ice sheets several times but specific evidence of multiple glaciation in east-central Vermont and west-central New Hampshire has not been demonstrated. Multiple-till exposures representing two separate glaciations are known in northern, east-central, and southern New Hampshire (Koteff and Pessl, 1985), as well as in southern Quebec and southern New England. The margin of the last ice sheet retreated northward from Long Island at least by 19,000 years ago (Sirkin, 1982), and the Quebec Appalachians were deglaciated by 12,500 years ago (McDonald and Shilts, 1971). Therefore, the ice margin retreated through the field trip area between 19,000 and 12,500 years ago. Using linear interpolation and assuming a steady rate of ice margin retreat, we can guess that the ice margin retreated through the area between 14,000 and 15,000 years ago. However, the ice probably had an increasing rate of retreat through this area.

Stewart (1961), and Stewart and MacClintock (1964, 1969, and 1970) recognized three separate drift (till) sheets in Vermont and westernmost New Hampshire. They named (1) a northwest-derived Bennington drift, (2) a northeast-derived Shelburne drift, and (3) a northwest-derived Burlington drift. Most of

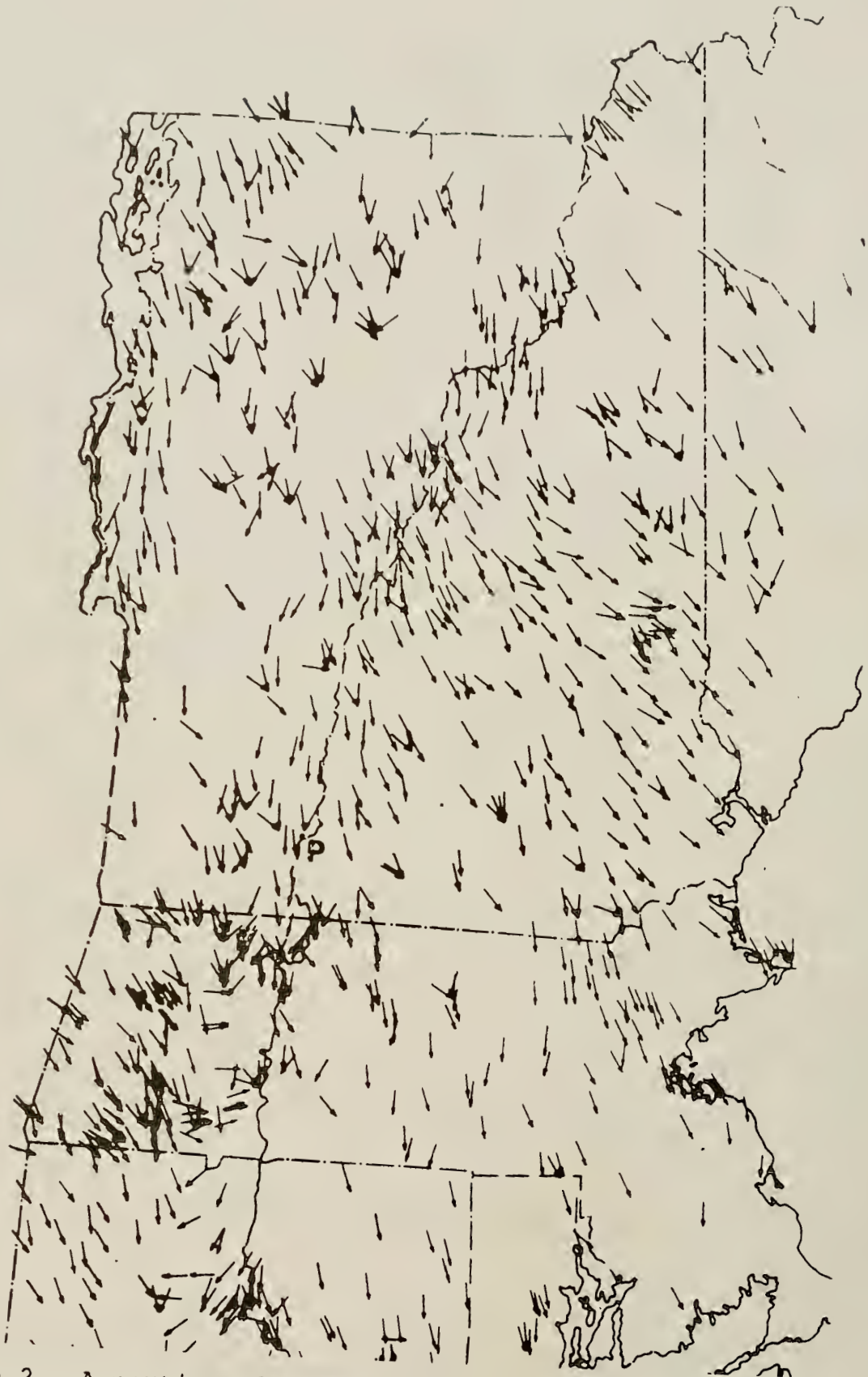


Figure 2. A portion of James W. Goldthwait's compilation of glacial striations in New England; P, Putney, Vt., (Flint, 1957, p. 60).

the area of this field trip lies in the area of the so-called Shelburne drift (Fig. 1). The Shelburne drift was found not to exist at its type locality (Wagner, Morse and Howe, 1972) and 8 of 9 indicator fans mapped in the area of the Shelburne drift are oriented to the south-southeast. According to Stewart and MacClintock, they should be oriented to the southwest of their source areas. I believe that the model of three drift sheets in Vermont and New Hampshire is untenable and that the surface till of New England resulted from the advance and retreat of one ice sheet, the Late Wisconsinan (Laurentide) ice sheet.

During retreat of the ice sheet in this area, the ice margin was accompanied by a northward-expanding glacial Lake Hitchcock (Lougee, 1939, 1957). Lake Hitchcock developed a stable outlet over a bedrock threshold at New Britain, Connecticut, and drainage down the present-day course of the Connecticut River was blocked by a large ice-contact delta at Rocky Hill, Connecticut (Stone and others, 1982). Recent work by Koteff and Larsen (1985 and in prep) indicates that the former shoreline of Lake Hitchcock now rises toward about N21.5°W with a gradient of 0.90 m/km (4.74 ft/mi). The lake formed during ice retreat when the land was still depressed by the weight of the ice, and it extended 320 km from its spillway to West Burke, Vermont, before uplift due to the removal of the weight of the ice sheet commenced. Deltaic and lake-bottom deposits associated with Lake Hitchcock will be observed at several stops on this field trip.

At its maximum extent, an arm of Lake Hitchcock extended up the valley of the Second Branch of the White River and into Williamstown Gulf in central Vermont. Gravel bars located in the valley bottom near East Brookfield appear to have been formed by the outlet stream from glacial Lake Winooski after Lake Hitchcock drained (Larsen, 1984). This would seem to indicate that the dam for Lake Hitchcock had been breached while glacial ice still blocked the northwest-draining Winooski River. However the level of Lake Hitchcock lowered, whether it was by a rapid breaching of its dam or by slow rebound of the crust, or by both, the Connecticut River eventually cut down through the sediments of Lake Hitchcock leaving former flood plains elevated as stream terraces above the level of the modern flood plain. Stream-terrace deposits will be observed at Stops 1 and 2.

No mention has been made of glacial Lake Upham in the above account. The reason for that is twofold. First, that portion of the map of Lake Upham shown by Lougee (1957) to be north of the so-called "Algonkian hinge line" essentially is the same as the shoreline shown by Larsen (1984) and Koteff and Larsen (1985 and in prep) to be the northern part of Lake Hitchcock. Secondly, although many data points (delta elevations) collected by Koteff and Larsen (in prep) fall below the level of Lake Hitchcock, at this time there is no clearly defined single shoreline that one might call "Lake Upham". For these two reasons, I suggest that the term "Lake Upham" be dropped.



Figure 3. Indicator fans in Vermont, New Hampshire, and southern Quebec: 1. Barre; 2. Braintree; 3. Brocklebank; 4. Knox Mt.; 5. Lebanon; 6. Glover; 7. Ascutney; 8. Cuttingsville; 9. Mt. Hereford; solid black, igneous source rocks; black lines down-glacier from plutons are 10% isopleths showing percent of source rocks in till samples. (Sources: 1, Larsen, 1972; 2-8, Great Pebble Campaigns at Norwich University, 1974-78 and 1980-81; 9, McDonald, 1967)

ADVANCE OF THE ICE

During advance of the last ice sheet in this part of the Connecticut Valley, braided meltwater streams issuing from the ice dropped tons of sediment in front of the ice as outwash or valley train deposits. As time went on, the ice sheet overrode these advance outwash deposits and in most areas removed them and recycled them into till or into younger advance outwash further downstream. However, there are three places in east-central Vermont and western New Hampshire where advance outwash deposits can be observed. They are (1) along the Jail Branch southeast of Barre (Larsen, 1972, Stop 1), (2) on the south side of the White River east of Bethel, and (3) in two pits visited on this field trip in West Lebanon (Stops 1 and 2). Advance outwash probably occurs at other sites but exposures are not common.

DIRECTION OF ICE MOVEMENT

The direction of movement of the former ice sheet can be ascertained by a study of striations, roche moutonnée forms, crag-and-tail features, indicator fans, and the orientation of elongated stones in till (till-fabric analysis). Figure 2 shows a portion of a map of glacial striations compiled by James W. Goldthwait (Flint, 1957). Note the fact that along the Connecticut River of east-central Vermont and western New Hampshire striations trend between south-southwest and southeast. Away from the river both on the east and the west the striations are oriented between south-southeast and southeast. I attribute the pattern to one glaciation and to the facts that striations are made at different times during a single glaciation and that the direction of movement of bottom ice may be quite different than the regional gradient on the surface of the ice due to rugged topography.

Support for the idea that ice movement in east-central Vermont was to the south-southeast as shown in Figure 2 came from a series of "Great Pebble Campaigns" at Norwich University. Students in physical geology labs each collected 100 pebbles at selected sites around and south of granitic plutons and other unique source areas. When the percent of indicator clasts at each site was plotted it was found that the highest concentrations were between due south and southeast of each source area studied (Fig. 3).

When we compare the south-southeast orientation of the indicator fan derived from the Lebanon dome (5, Fig. 3) with the southwest orientation of the till-fabric arrows shown in Figure 1 there is an obvious discrepancy. Note on Figure 1 the symbol at Lebanon that shows southwest-oriented surface till over southeast-oriented subsurface till. The location of that Stewart and MacClintock study is taken to be the West Lebanon Sand and Gravel pit (Stop 2) where, today, apparently only one till can be observed. To approach a solution to the problem, till-

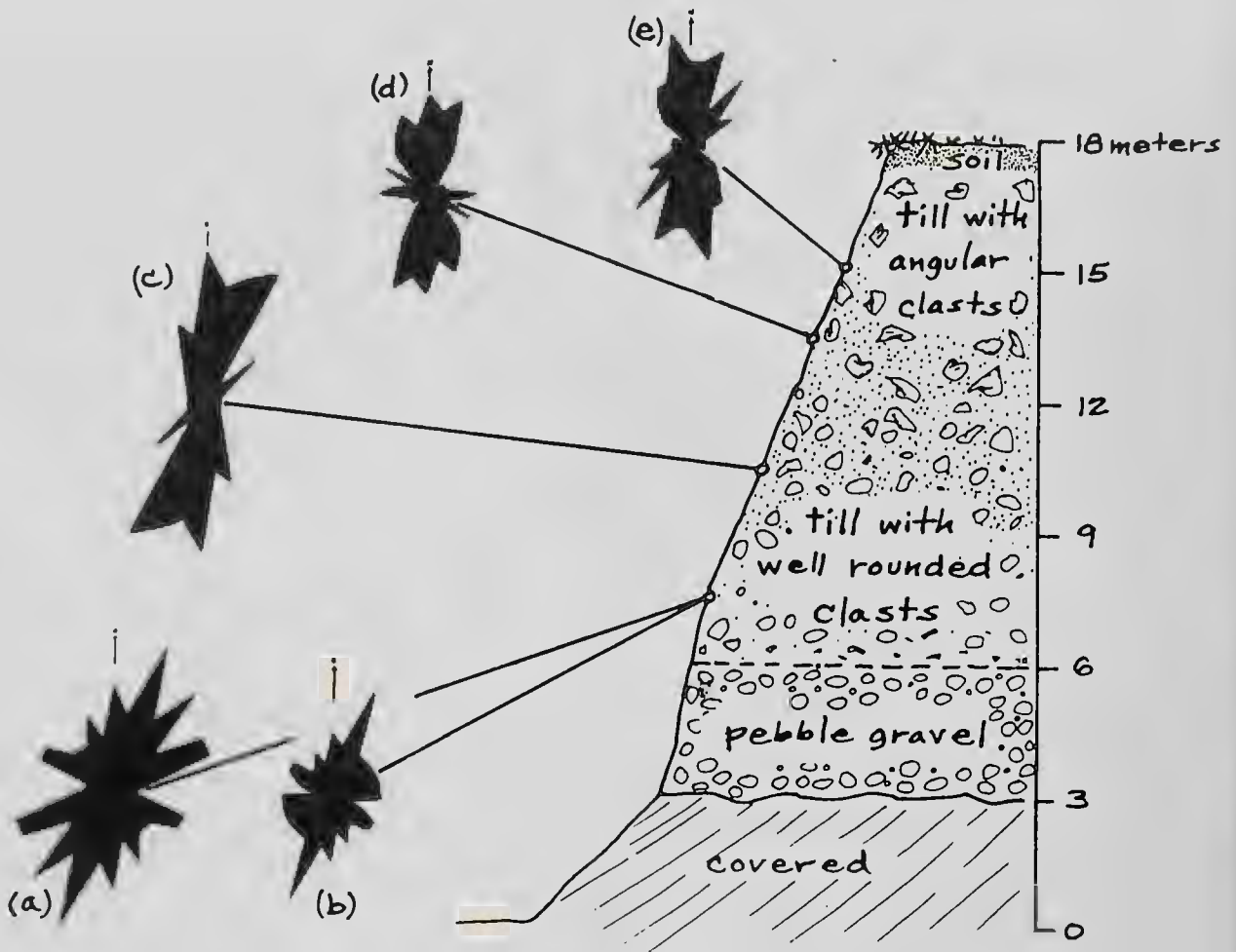


Figure 4. Stratigraphic section at Stop 4, West Lebanon Sand and Gravel pit, showing location of till-fabric studies made in July, 1973, by the following: (a) Steve Tenney, (b) John Cleary, (c) Julian Green, (d) Jim Reynolds, and (e) Tom Lyman. Note the upward change in shape and orientation of the till-fabric diagrams. The more rounded diagrams reflecting more spread of data are at the bottom and the narrower diagrams indicating less spread of data are at the top. The change in orientation from north-northeast/south-southwest at the bottom to north/south at the top reflects a change in the direction of ice movement during deposition of the till.

fabric studies were made during the summer of 1973 by students in a geomorphology class at Dartmouth College (Fig. 4). The till-fabric diagrams in Figure 4 show a change in orientation from north-northeast/south-southwest at the bottom to north/south at the top which reflects a change in the direction of ice movement during deposition of the till. The question remains: when were different parts of the till section deposited? The north/south orientation of the upper two fabric diagrams shown in Figure 4 is more consistent with the south-southeast trend of the Lebanon indicator fan than the southwest trend of "surface till" shown by Stewart and MacClintock.

DEGLACIATION

As mentioned above, based on interpolation, the ice margin retreated through the West Lebanon area between 14,000 and 15,000 years ago. Evidence for that statement comes from the fact that Mirror Lake, NH (altitude 212 meters) was deglaciated by about 14,000 years ago (Davis, Spear, and Shane, 1980). Because Mirror Lake is located 61 kilometers (38 miles) northeast of West Lebanon, it seems reasonable to assume that the West Lebanon area was deglaciated at least by that time.

In Massachusetts, deglaciation of the Connecticut Valley was by an active lobe of ice that readvanced several times (Larsen and Hartshorn, 1982). The active lobe is also shown by a radial pattern of striations stretching across the valley and the distribution of erratics of Jurassic-Triassic rocks transported both east and west of their source area in the Connecticut Valley. Inspection of the Goldthwait compilation (Fig. 2) reveals southwest and west-southwest striations located west of the Connecticut River in both Massachusetts and Connecticut.

A basic question then is how far north was the Connecticut Valley ice margin an active, spreading lobe. I believe the answer lies on the Goldthwait map near Putney, Vt (P, Fig. 2), at the site of a striation that trends about S30°W. On the west side of the Connecticut Valley north of Putney there are no striations that indicate a radial pattern of movement by an active ice lobe. I interpret the lack of a radial pattern of striations to indicate that deglaciation of the Connecticut Valley north of Putney was by a stagnant tongue of ice. In my view the width of the stagnant zone was many kilometers wide and, up-glacier from the stagnant ice, the active ice was sluggish at best, showing no sign of lobate flow in late-glacial time.

An important glacial feature that formed in late-glacial time in this area, and one that typically forms in stagnant ice, is the Connecticut Valley esker. It probably was built in segments over a north-south distance of at least 40 kilometers from Windsor, Vt, on the south, to Lyme, NH, on the north. We will not visit the Connecticut Valley esker on this field trip, but excellent exposures in the esker are available for study in

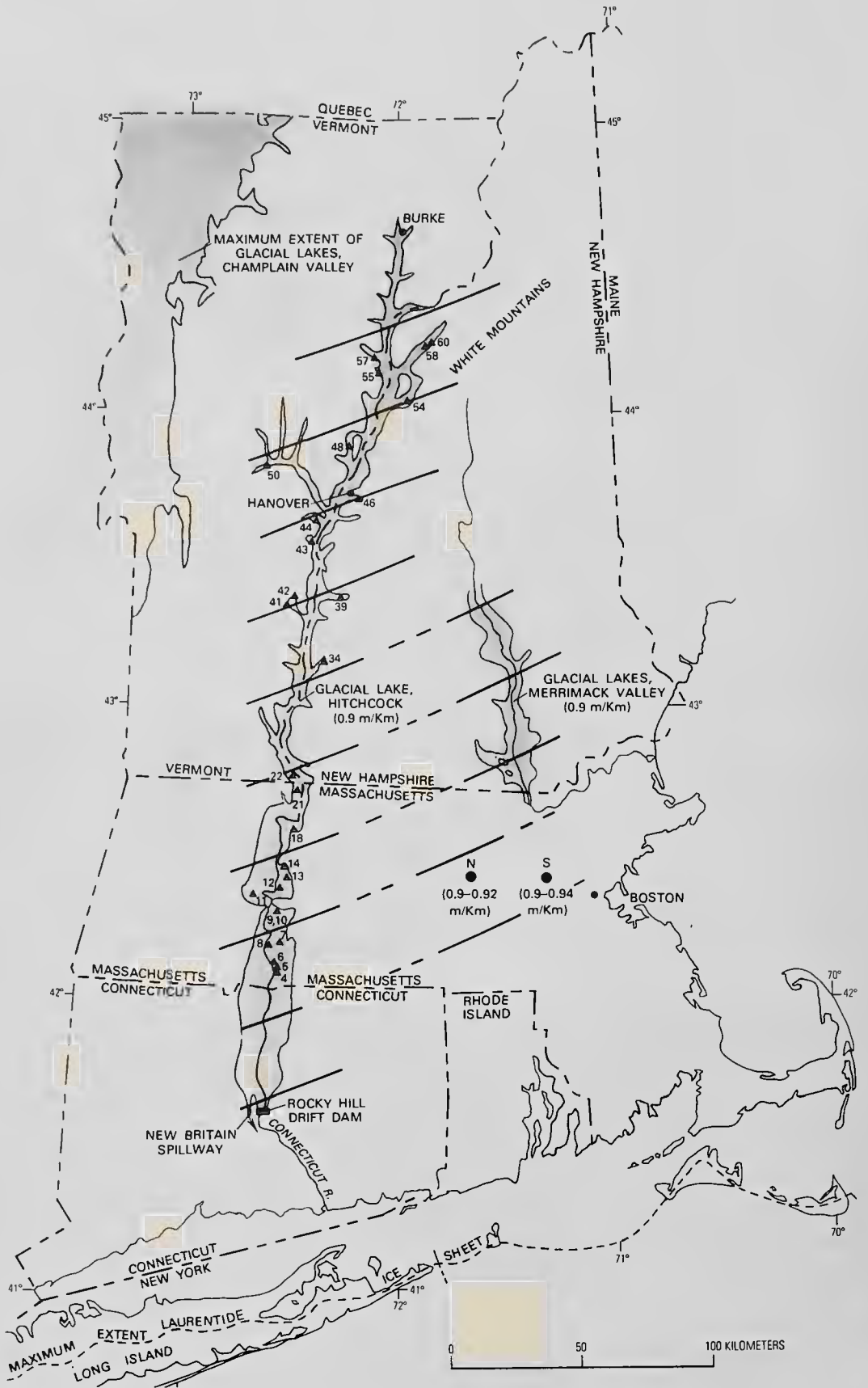


Figure 5. Generalized outline of glacial Lake Hitchcock and selected other glacial lake areas in western New England. (N) glacial Lake Nashua; (S) glacial Lake Sudbury; solid triangles denote location of altitude obtained from unmodified, ice-marginal, or meltwater-derived delta; uplift isobase interval is 25 meters. (Figure from Koteff and Larsen, in prep.)

Hartland, Vt, where the top of the esker rises 40 meters above what was once the floor of Lake Hitchcock. Sediments in the Pike pit, 1.2 kilometers east of Hartland village, display a wide range of grain sizes from interbedded pebbly sand and pebble-cobble gravel in south-dipping crossbeds to fine and very fine sand with ripple crossbedding dipping to both north and south. Large angular blocks, over 1 meter on edge, of bedded fine sand, silt, and clay appear to have dropped into the esker sequence from overlying ice. At the very least, the presence of the Connecticut Valley esker tells us that there was a large subglacial, meltwater stream flowing south into Lake Hitchcock just prior to retreat of the ice margin through this area. Although we do not stop in the Connecticut Valley esker on this trip, sediments of the Sharon esker in the White River valley will be observed at Stop 5.

GLACIAL LAKE HITCHCOCK

As the ice margin retreated through the West Lebanon area it was accompanied by a northward expanding Lake Hitchcock. For an up-to-date treatment of the origin and early history of Lake Hitchcock see Stone and others (1982) and Koteff and others (1987). By the time the ice margin was in the vicinity of the Holyoke Range in Massachusetts, the level of Lake Hitchcock had become stable because downcutting at the New Britain spillway had reached bedrock and ceased. Koteff and Larsen (in prep.) have established the location and orientation of the stable shoreline of Lake Hitchcock by determining the elevation of the topset/foreset contact in many deltas (Fig. 5). The highest 28 deltas are ice-marginal features, consecutively built northward, and define the former shoreline as a plane with a gradient of 0.90 m/km (4.74 ft/mi) toward N21.5°W (Fig. 6). Because the former shoreline appears to be planar, as opposed to being curved, we believe that postglacial rebound did not commence in New England until the ice margin had retreated north of West Burke (Koteff and Larsen, in prep.). If rebound had caused the spillway to rise while ice still occupied the northern part of the Lake Hitchcock basin the youngest deltas would have formed in a rising lake. That would have produced a concave-up projected profile instead of the linear projected profile that we see in Figure 6.

The sediments of Lake Hitchcock that will be seen on this field trip consist mainly of deltaic and proximal lake-bottom deposits. The latter consist of varves with winter clay layers less than 2 centimeters thick and summer layers of laminated very fine sand and silt that typically are 30, 40, 50, or more centimeters thick. These proximal varves occur directly above esker deposits and are often collapsed. I attribute the thickness of the summer layers to the rapid rate of deglaciation of stagnant ice. Proximal varves are well exposed at Stops 1 and 5.

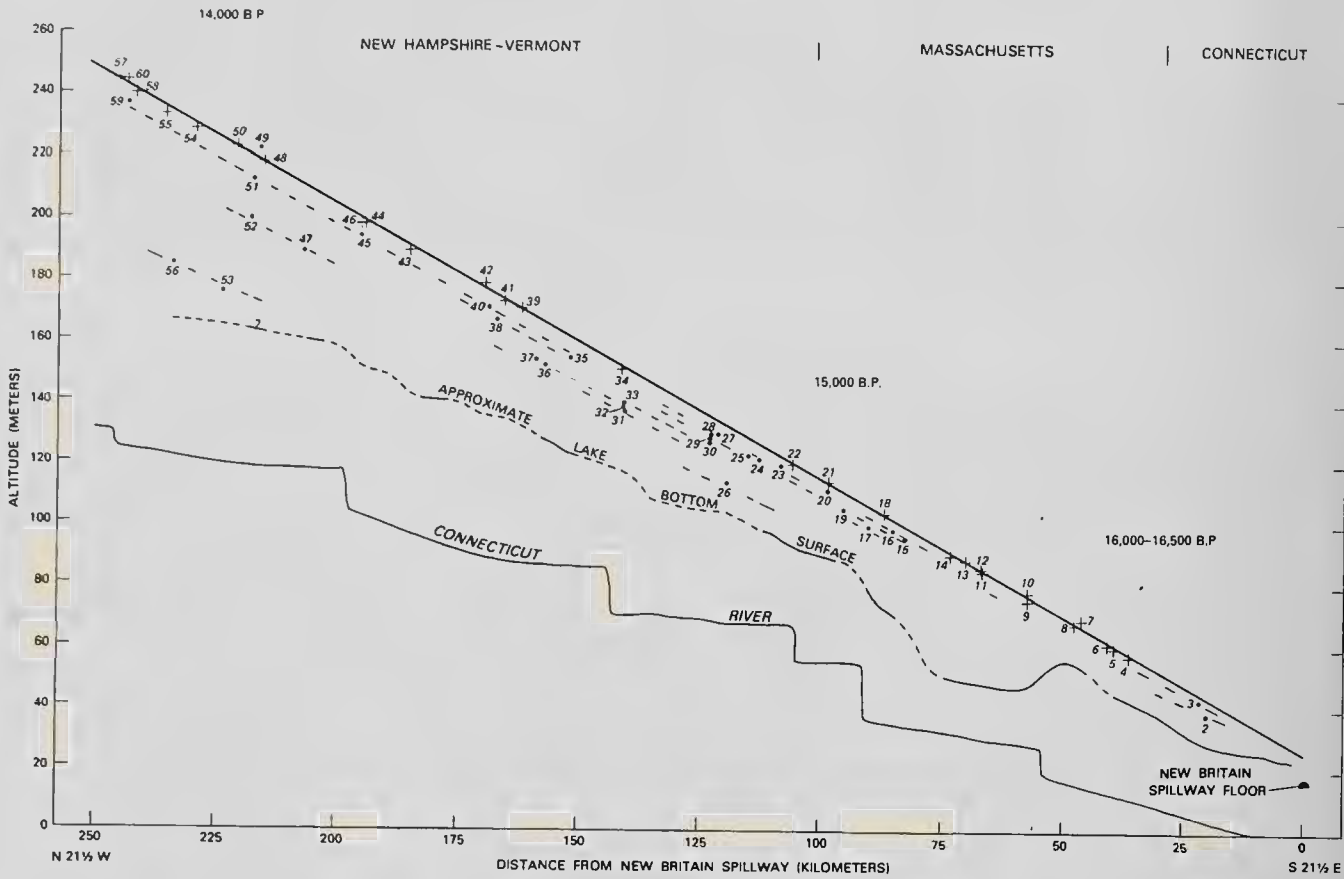


Figure 6. Ordinary least squares regression profile based on altitudes of topset/foreset contacts of 28 unmodified, ice-marginal, or meltwater-derived deltas (+) in glacial Lake Hitchcock. (.) other altitudinal data. Dashed profiles are diagrammatic only. Lake-bottom profile estimated from previous publications and topographic maps. (Figure from Koteff and Larsen, in prep.)

QUECHEE GORGE

When the retreating ice margin was located just north of the present site of Quechee Gorge, an ice-contact delta was formed by meltwater streams flowing into an arm of Lake Hitchcock that extended into the valley of the Ottauquechee River. The original delta extended completely across the valley and was slightly higher than the sandy plain crossed today by U.S. Route 4 just east of Quechee Gorge. Later, after either Lake Hitchcock had drained or postglacial uplift had begun, meteoric water from upstream crossed over the delta plain and lowered the surface by as much as 3 meters. This is shown by fluvial beds overlying lake-bottom sediments at Stop 3.

In time, Lake Hitchcock drained when the Rocky Hill dam was breached. The stream that we know today as the Ottauquechee River was flowing south on the west side of the Quechee delta. As Lake Hitchcock lowered, the Ottauquechee River was easily able to erode down through sand, gravel, and till until it struck ledge. Downcutting slowed abruptly when the ledge was encountered but nonetheless it continued to the present day to produce one of the top geologic and scenic features of Vermont.

The preglacial valley of the Ottauquechee River is located under the east side of the Quechee delta. A well, located near U.S. Route 4, 1.1 kilometers east of Quechee Gorge, has over 35 meters of unconsolidated sediment and a second well, 366 meters east of Quechee Gorge and 30 meters south of U.S. Route 4, has 42 meters of fine sand overlying 5 meters of till (James W. Ashley, 1987, pers. commun.). The Ottauquechee River could not slip laterally down along the bedrock ridge it encountered during downcutting and back into its preglacial valley because of a bedrock high located 198 meters N75 E of the east end of the bridge over Quechee Gorge. The bedrock is exposed above the level of U.S. Route 4 and has faint striations trending due south.

WILLIAMSTOWN GULF

Lake Hitchcock eventually extended up the valleys of the White River and its tributaries. Its maximum northward extent in the White River basin was up the valley of the Second Branch and 1.6 kilometers into Williamstown Gulf, a narrow V-shaped valley about 2.5 kilometers long. Located just south of the lowest drainage divide (279 meters/915 feet) between the Winooski River basin and the White River basin, Williamstown Gulf must have been occupied by an outlet stream from a glacial lake whenever the Winooski River was blocked on the northwest by glacial ice. If we accept two major glacial advances and retreats between the St. Lawrence Lowland and southern New England during the Wisconsinan (Koteff and Pessl, 1985), then by necessity Williamstown Gulf was eroded by a major outlet stream on four occasions during the Wisconsinan, that is, during each advance and retreat of the ice sheet. Williamstown Gulf is,

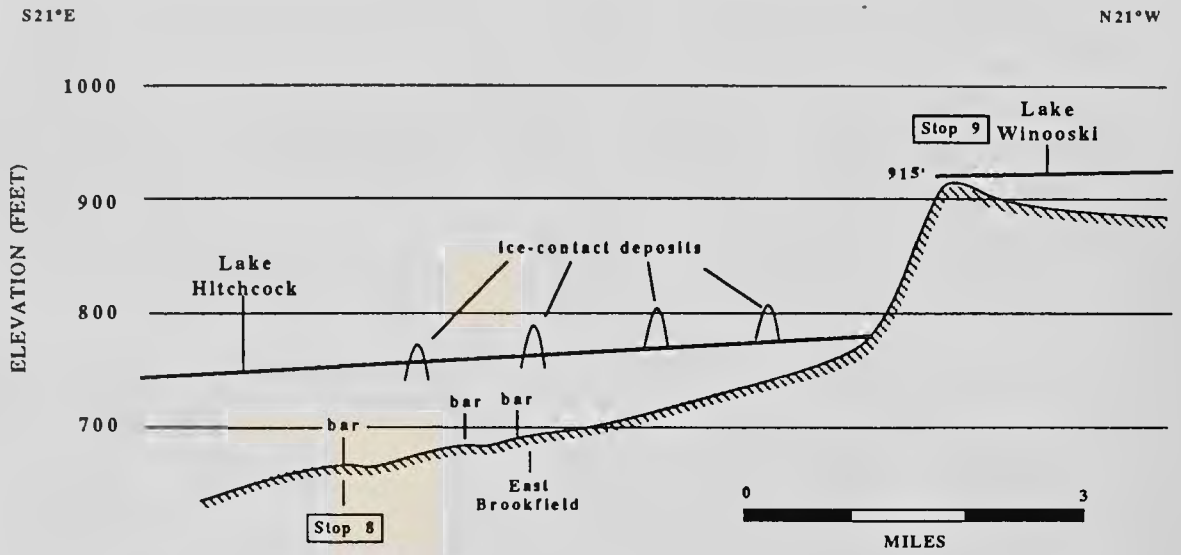


Figure 7. Projected profile oriented N21°W-S21°E of Lake Hitchcock shoreline in relation to gravel bars on the valley floor of the Second Branch of the White River.

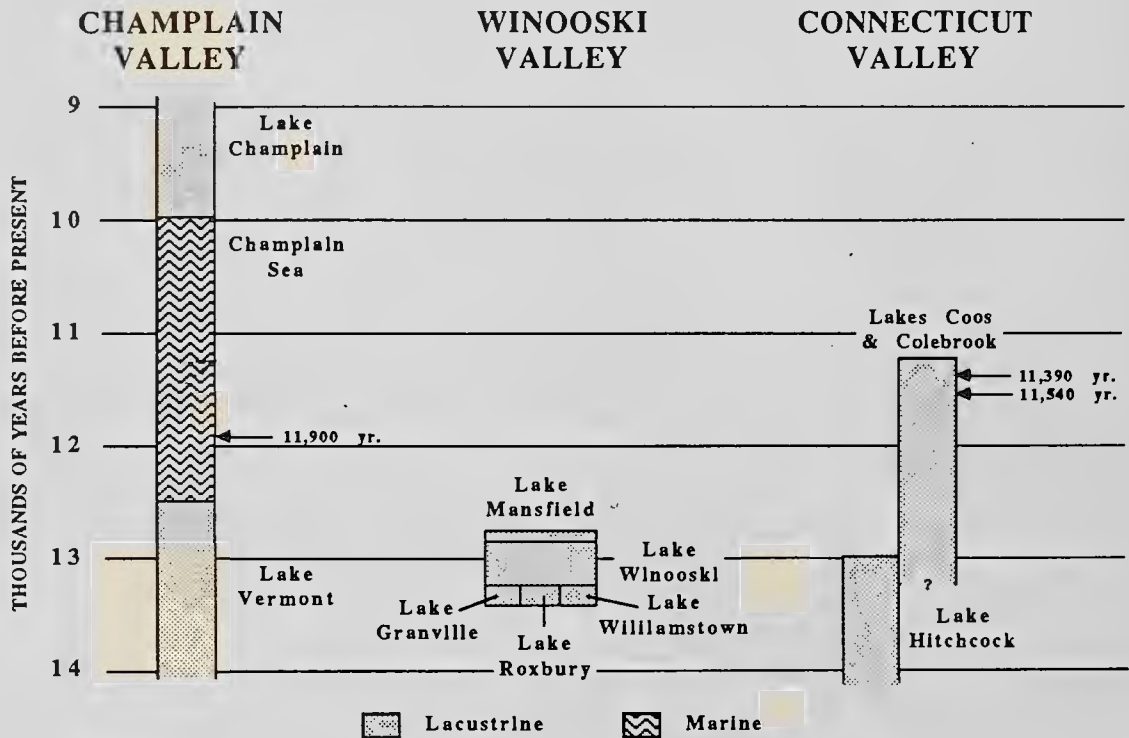


Figure 8. Chronology of glacial lakes in the Champlain, Winooski, and Connecticut valleys in relation to C-14 dated Champlain Sea.

therefore, a relic landform. It bears a special fluvial imprint in an otherwise glacial landscape.

RELATIVE AGES OF LAKE HITCHCOCK,
LAKE WINOOSKI, AND THE CHAMPLAIN SEA

Glacial Lake Winooski developed in the northwest-draining Winooski valley after the margin of the ice sheet retreated north of the 279-meter threshold located 4.0 kilometers south of Williamstown, Vt. Meltwater from the ice sheet passed through Lake Winooski, over the threshold and into Williamstown Gulf where, at an elevation of about 237 meters (777 feet), it joined Lake Hitchcock (Fig. 7). Four small ice-contact deposits (deltas?) are located on the valley side between the south end of Williamstown Gulf and a point 1.9 kilometers south of East Brookfield. These four deposits fall on or near the projected level of Lake Hitchcock, and are believed to have been deposited directly into Lake Hitchcock (Fig.7).

Longitudinal gravel bars up to 300 meters in length occur along 7 kilometers of valley floor in the vicinity of East Brookfield. The bars, difficult to see on the ground, are easily viewed on aerial photographs. The bars are composed of pebble gravel with some cobbles in horizontal layers or in sheets that are parallel to the bar surfaces that are convex-up in both longitudinal and transverse profile. The bedding is indistinct except that some gravel layers have a sand matrix. Sand layers and sedimentary features formed in the lower-flow regime are absent. The gravel layers were all deposited in the upper-flow regime by a large stream under flood conditions. I interpret that stream to have been the outlet stream from Lake Winooski after Lake Hitchcock had drained, because the minimum projected level of Lake Hitchcock is 25 meters above the bar surface at Stop 8, 3.0 kilometers south of East Brookfield (Fig. 7).

If that is true, then it is possible to establish a minimum age for the draining of Lake Hitchcock by inspecting the chronology of glacial lakes in the Champlain, Winooski, and Connecticut Valleys in relation to the Carbon-14-dated Champlain Sea (Fig. 8). Lake Winooski continued to exist while the ice margin retreated down the Winooski valley to the northwest toward the Champlain Valley. When the ice margin reached the vicinity of Jonesville, Vt, on the west side of the Green Mountains, Lake Winooski drained and water no longer flowed south over the 279-meter threshold in central Vermont.

A sequence of four separate ice-marginal lakes formed in the valley of the Huntington River as Champlain Valley ice melted from the lower Winooski valley (Wagner, 1972). Eventually, an arm of glacial Lake Vermont occupied the lower Winooski valley as far east as Waterbury. With retreat of the ice margin in the Champlain Valley, glacial Lake Vermont extended northward into Quebec. Ice in the St. Lawrence Valley disintegrated and

about 12,500 years ago (Gadd and others, 1972), marine waters of the Champlain Sea replaced Lake Vermont (Fig. 8). The Champlain Sea existed from about 12,500 to 10,000 years ago at which time crustal rebound raised the threshold of the Champlain basin to sea level, which caused the freshening of Lake Champlain.

The conclusion I reach is that the draining of Lake Hitchcock occurred while Lake Winooski existed and several hundred years before any marine sediments were deposited in the Champlain Valley. If we accept the date of 12,500 BP as the time of marine incursion, then the draining of Lake Hitchcock could easily have occurred at or before 13,000 BP.

ACKNOWLEDGMENTS

Literally hundreds of students at Norwich University participated in one or more "Great Pebble Campaigns". Without their help, the map of indicator fans in Vermont (Fig. 3) would not have been possible. Five Dartmouth College students produced the till-fabric diagrams shown in Figure 4. James H. Reynolds, III, supplied the computer drawings in Figures 7 and 8. Carl Koteff reviewed the paper and made suggestions for its improvement. To all these people I extend my sincere thanks.

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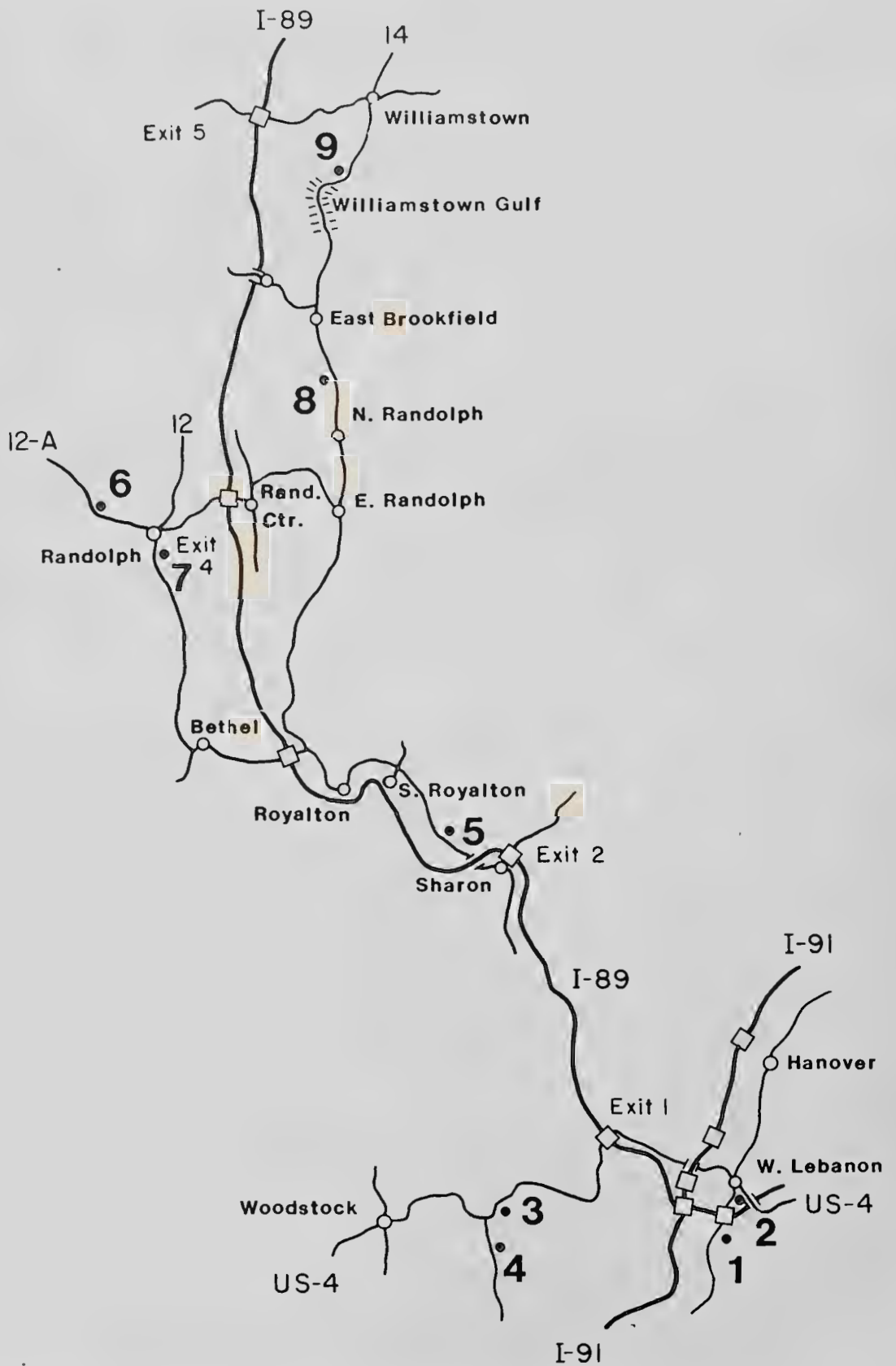


Figure 9. Location of field trip stops.

Road Log

START AT PARKING LOT AT McDONALD'S RESTAURANT, N.H. ROUTE 12A, 0.25 MILE SOUTH OF INTERCHANGE 20, ROUTE I-89, WEST LEBANON, NEW HAMPSHIRE

Mileage

- 0.0 Begin mileage and turn right (east) on Interchange Drive
- 0.1 Turn left on Plaza Heights
- 0.3 Turn right between two large buildings, park in pit area

STOP 1. MOULTON CONSTRUCTION CO. PIT.

EAST FACE: The owners request that visitors stay away from the steep east face which has a total height of about 23 meters. From the bottom up the section consists of 8.4 meters of pebbly coarse sand in trough crossbeds that consistently dip to the south. Next is a massive layer, 2.7 meters thick, composed of pebble gravel with cobbles. Above the massive gravel layer is a 4.6-meter unit similar to that at the base with trough crossbeds that dip to the south. Capping the section is 4.6 meters of gray, compact till that is overlain on the south by 1 meter of brown fine to very fine sand. The sequence is interpreted to be advance outwash at the base with the overlying till representing the advance and retreat of the Late Wisconsinan ice sheet. The sequence does not represent a readvance of the last ice sheet because during retreat the margin of the last ice sheet was accompanied by a northward-expanding glacial Lake Hitchcock. A readvance in a glacial lake should result in a section with till overlying lacustrine sediments, not fluvial sediments as seen here.

SOUTHWEST FACE: At the base 1.0 meter of loose coarse sand and pebble gravel is overlain by 3.2 meters of compact diamict made up of pebble gravel with cobbles and boulders. The clasts in the diamict are rounded to angular and are up to 1.0 meter in diameter. Above the diamict is 5.0 meters of laminated very fine sand, silt, and clay in proximal varves up to 60 centimeters thick. In places, up to one half meter of loose pebble gravel lies between the diamict and the varves. Stream terrace gravel overlies the varved sediments at the south end of the pit. The 1.0 meter of loose gravel at the base is interpreted to be advance outwash. The diamict is believed to be the Late Wisconsinan till that here was derived from coarse advance outwash and the varves are interpreted to be the bottom sediments of Lake Hitchcock. At the top, thin stream-terrace deposits represent a stage in downcutting through older sediments by the Mascoma River or the Connecticut River.

Retrace route to NH Rt 12A

- 0.8 Turn right (north) on Rt 12A
- 1.0 Proceed north under Interstate I-89

- 2.3 Turn right (east) on U.S. Route 4 and immediately
- 2.33 Turn right (south) on Elm Street, proceed south
- 2.7 Gate to Twin State Sand and Gravel
- 2.9 Near office turn left and descend to active pit

STOP 2. TWIN STATE SAND AND GRAVEL PIT.

SOUTHEAST FACE: Walk southeast to face with 10 meters of till overlying pebble gravel. The section is interpreted to be Late Wisconsinan till overlying advance outwash. The lower portion of the till is rich in well rounded pebbles. Measurements for the till-fabric diagrams shown in Figure 4 were made at this site.

ACTIVE PIT: At the west end the section consists from the bottom up of: (1) 2.5 meters of pebbly sand in south-dipping trough crossbeds, (2) 3 meters of pebble gravel in flat to gently dipping beds, (3) 3.2 meters of pebbly medium sand with trough crossbeds, (4) 0.5 to 3.0 meters of pebble-cobble gravel in well developed channels, (5) up to 1.0 meter of fine sand with a well developed soil profile, and (6) 4.0 meters of artificial fill made up of well bedded fine sand that probably was deposited in a man-made settling pond similar to that now found just to the south. I believe that the lower 8.7 meters is advance outwash formed in a proglacial braided stream. The poorly sorted pebble-cobble gravel in channels and the fine sand constitute stream-terrace deposits that were formed as alluvium by the Mascoma River as it cut laterally to the north into a scarp underlain by till. The channels represent flood events in the Mascoma drainage basin while lateral cutting took place.

Proceed southwest up and out of pit

- 3.2 Turn right on tar road, proceed north out of pit
- 3.8 CAUTION. Turn left (west) on US Route 4 and turn left (south) on NH Rt 12A
- 4.8 Turn west on I-89 North at Interchange 20
- 5.3 Cross Connecticut River and enter Vermont
- 5.6 DO NOT EXIT ON I-91, CONTINUE WEST ON I-89
- 6.6 Projected shoreline of Lake Hitchcock passes through the site of the V.A. Hospital at the right
- 8.8. Turn right from I-89 at Exit 1
- 9.1 Turn left (west) on US Route 4
- 11.9 Surface of Quechee delta

12.3 Red Pines Restaurant on right was the site of an excavation where the topset/bottomset contact was exposed in 1982

12.4 Turn right and drive behind Restaurant-Dana's at the Gorge

STOP 3. QUECHEE ICE-CONTACT DELTA.

The topset/bottomset contact of the delta was measured to be 194.4 meters (638 ft) ASL. Pebbly coarse sand is exposed at the surface and very fine sand and silt are found draping the bedrock knobs. The projected level of Lake Hitchcock at this site is about 197.5 meters (648 ft) ASL. Because there are no deltaic foreset beds exposed below the pebbly coarse sand, it is presumed that they were removed by at least 3.1 meters (10.2 ft) of erosion at this site.

Quechee gorge resulted from the superposition of the Ottauquechee River which was located on the west side of the delta when Lake Hitchcock drained. The preglacial course of the river is at the east end of the delta.

Proceed west on US Route 4 across the Quechee Gorge bridge

13.3 Turn left (south) on Quechee Road at blinking light

13.6 Channel on right (now occupied by man-made pond) leads to saddle between two hills; melt-water that fed the ice-contact delta at Stop 4 passed through this saddle to the southeast

13.9 The bank between the pond on the left and the road marks the ice-margin position when the ice-contact delta was being built

14.1 Turn left (east) into pit road, park next to gate

STOP 4. ROBERT SEERY PIT.

Note the pebble-cobble gravel at the top of the north face of the pit. The section at the south face from the bottom up consists of 2 meters of light brown fine sand covered by 2 meters of pebbly coarse sand and pebble gravel in trough cross-beds. Fine sand with pebbles, about 0.3 of a meter thick, caps the gravel and is interpreted to be eolian in origin. The contact between the gravel and the underlying fine sand is thought to approximate the topset/foreset contact and was measured to be 197.5 meters (648 ft) ASL.

Retrace route north to US Route 4

14.9 Turn right (east) on US Route 4 at blinking light

15.6 Pass over Quechee gorge

- 18.9 Turn right, enter Interstate I-89 North at Exit 1
- 24.0 Sharon Rest Area
- 28.8 Turn right, leave I-89 North at Exit 2
- 29.1 Turn left at end of ramp
- 29.3 Turn right on Route 14 in Sharon, proceed west
- 30.3 Pass under I-89 bridge over the White River
- 31.0 Thick varves of Lake Hitchcock exposed on right
- 31.7 CAUTION, TURN LEFT ACROSS TRAFFIC, PARK IN REST AREA

STOP 5 SHARON ESKER

The face is oriented northwest-southeast and is about 240 meters long and up to 35 meters high. A section was measured 88 meters northwest of the culvert in the v-shaped notch and had 3 units. At bottom unit had 20.6 meters of interbedded pebble-cobble gravel, pebble gravel, and coarse sand. Open-work structure in the gravels and dune crossbedding dipping to the southeast are common. The environment of deposition for the lower unit was a high energy stream in a subglacial tunnel. The second unit is 9.8 meters thick and consists of varves up to 50 centimeters thick that represent bottom deposits of Lake Hitchcock. The third unit, exposed at the top, is 2.2 meters thick and is made up of stream-terrace deposits. Most of the varves dip 30 to 40 degrees to the northeast indicating they were deposited on or adjacent to ice that later melted.

Proceed northwest on Route 14

- 34.6 Village of South Royalton
- 35.4 Excellent exposure of graded beds in the Gile Mountain Formation (Devonian) at river level
- 36.1 Narrow railroad underpass
- 36.6 Village of Royalton
- 37.4 Narrow railroad underpass
- 38.1 Turn left (west) on Route 107
- 38.7 Turn right, enter I-89 North at Exit 3
- 47.3 Turn right from I-89 at Exit 4, turn left (west) on Route 66

- 50.3 Proceed straight at blinking red lights, Route 66 ends, join Route 12
- 50.4 At junction leave Route 12 and proceed straight on Route 12A west and northwest up the valley of the Third Branch of the White River
- 53.5 Turn right on dirt road and immediately turn right again on road to pit

STOP 6 PIT IN LOWER BRANCH DELTA

The topset/foreset contact seen at the north end of this pit has only recently been exposed. The topset beds are 1.1 meters thick and are composed of poorly sorted pebble-cobble gravel. The foreset beds are at least 2.4 meters thick and are composed of clean pebble gravel and pebbly coarse sand with open-work structure being common. The foreset beds dip toward N55°E indicating that they were deposited by melt-water streams flowing from a stagnant tongue of ice in the Third Branch valley to the west. The topset/foreset contact has been trimmed indicating that some erosion of the foresets has taken place. The topset beds are poorly sorted in contrast to the foresets and were derived by a stream flowing from north to south down over the delta surface. The elevation of the topset/foreset contact was measured to be 228.3 meters (749.0 feet) ASL.

Retrace route to Randolph

- 53.8 Turn left on Route 12A
- 56.8 Turn right (south) on Route 12
- 57.2 Railroad tracks in center of Randolph, proceed south and rise up onto surface of Randolph ice-contact delta
- 57.7 Pull U-turn in vicinity of Gifford Memorial Hospital and park on east side of Route 12

STOP 7 RANDOLPH ICE-CONTACT DELTA

The surface elevation is about 228.7 meters (750 feet) ASL at the intersection of Route 12 and Highland Avenue. The surface slopes to the southeast and is underlain by pebbly sand. A similar surface lies 0.8 of a mile to the north. I interpret the landform to be an ice-contact delta that was formed in Lake Hitchcock by meltwater issuing from a stagnant tongue of ice the margin of which was located just west of the village of Randolph.

Proceed north on Route 12

- 58.5 Bear right on Route 12
- 58.6 Blinking red lights, turn right onto Route 66

- 61.3 Proceed east over I-89
- 62.0 Turn sharp left (north) in Randolph Center
- 62.6 Turn right, follow Route 66 to East Randolph
- 66.2 Junction with Route 14, turn left (north) in center of East Randolph
- 66.6 Pit starts on left, possible camera stop, geology is similar to that at Stop 5 with Lake Hitchcock bottom deposits overlying esker gravels
- 66.9 Pit ends on left
- 67.9 Pit on left with collapse structures
- 68.9 Road follows remnant of southernmost gravel bar
- 70.6 Turn left into pit 0.1 mile north of large brick house on left

STOP 8 WHEATLEY PIT IN GRAVEL BAR

Pebble gravel with cobbles in flat beds was formed in the upper-flow regime. The topography has been modified by man but inspection of aerial photographs of this area shows elongated landforms that are interpreted to be longitudinal bars formed by the outlet stream from Lake Winooski after Lake Hitchcock drained. This locality is 25.6 meters (84 feet) below the minimum projected water plane of Lake Hitchcock.

- 70.9 Turn left Route 14, proceed north
- 71.3 Landform on west side of valley is interpreted to be an ice-contact delta built directly into Lake Hitchcock at an approximate elevation of 230 meters (754) feet ASL.
- 72.0 The house and barn west of Route 14 at the cemetery are located on a streamlined landform interpreted to be a longitudinal bar formed by the outlet stream from Lake Winooski after Lake Hitchcock drained
- 72.4 Village of East Brookfield
- 72.8 Route 65 on left
- 73.5 Note streamlined appearance of plowed field on right
- 74.4 Red house on right is on a post-Lake Hitchcock fan formed by the dissection of an ice-contact delta that was graded to Lake Hitchcock
- 75.2 Enter Williamstown Gulf, a V-shaped valley that probably was occupied by an outlet stream draining the Winooski

River basin during each advance and retreat of several ice sheets

- 76.5 House on left marks approximate spot where Lake Hitchcock projection intersects the topography
- 77.0 CAUTION, TURN LEFT ACROSS TRAFFIC, ENTER DIRT ROAD. The road follows the margin of the outlet channel from Lake Winooski
- 78.0 Park on right near junction with Route 14

STOP 9 THRESHOLD OF GLACIAL LAKE WINOOSKI

The area behind the white house east of Route 14 represents the drainage divide between the Winooski River and the Connecticut River drainage basins. This locality represents the lowest spot on the margin of the Winooski drainage basin east of the Green Mountains. As long as the ice sheet blocked the Winooski River from draining to the west this threshold was occupied by an outlet stream draining to the south by way of the valley of the Third Branch of the White River. The elevation is about 279 meters (915 feet) ASL.

Turn left (north) on Route 14 to proceed to Williamstown (2.5 miles). From Williamstown: Barre is 6 miles north on Route 14 and Exit 5 of I-89 is 4 miles west on Route 65

Turn right on Route 14 for points south

FACIES RELATIONSHIPS WITHIN THE GLENS FALLS LIMESTONE
OF VERMONT AND NEW YORK

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INTRODUCTION

The Trenton Group in the northeastern United States represents one of the most thoroughly studied rock units in the world. However, the most northeasterly equivalent of the Trenton Group, the Glens Falls Limestone, has only received cursory attention in previous works that include Kay (1937, 1953), Erwin (1957), Welby (1961), and Fisher (1968). This is unfortunate since the Glens Falls Limestone provides a valuable link between more thoroughly studied Trenton Group sections of central New York and Quebec. Furthermore, this formation is located within 30 km (along depositional strike) of the Taconic orogenic events. With these thoughts in mind, this field trip has two purposes: first and foremost, the major lithologies of the Glens Falls Limestone will be described and placed into a facies model; secondly, the tectonic significance of the facies distribution will be discussed. These two goals will now be clarified.

The Glens Falls Limestone contains alternating limestone and shale beds which record the initiation of a large transgressive sequence noted in Middle Ordovician sections throughout the Appalachian Mountains. These limestone/shale cycles record shelf sedimentation in an elongate, rapidly subsiding, foreland basin bounded by the Adirondack basement to the west and an island arc with impinging thrust nappes to the east. Temporally, these lithologies represent an important transition from quiescent, shelf sedimentation recorded as massive Black River Limestone to calcareous and non-calcareous shales of the Utica and related shales which are generally interpreted as flysch deposits (Rowley and Kidd, 1981; Teetsel, 1984). These flysch sediments represent deep water deposits associated with increased subsidence near the eastwardly dipping subduction zone.

Within the Glens Falls Limestone, two members have traditionally been differentiated: the lower, massive Larrabee and the upper, shalier Shoreham. These members were defined by Kay (1937) and were primarily based on correlations to quarries in Larrabee and Shoreham, Vermont. The presence of the trilobite *Cryptolithus* served as the primary criteria for recognizing the upper Shoreham Member. More recent compilations by Fisher, (1977) have rejected the Shoreham Member since it defines a biostratigraphic zone and is misleading in defining lithologic boundaries. Fisher reclassified this upper shaley member as the Montreal Member on the basis of similarity to the Montreal Formation in southeastern Quebec. These lithologies are overlain by the Cumberland Head Argillite, a massive argillaceous unit with interbedded shale.

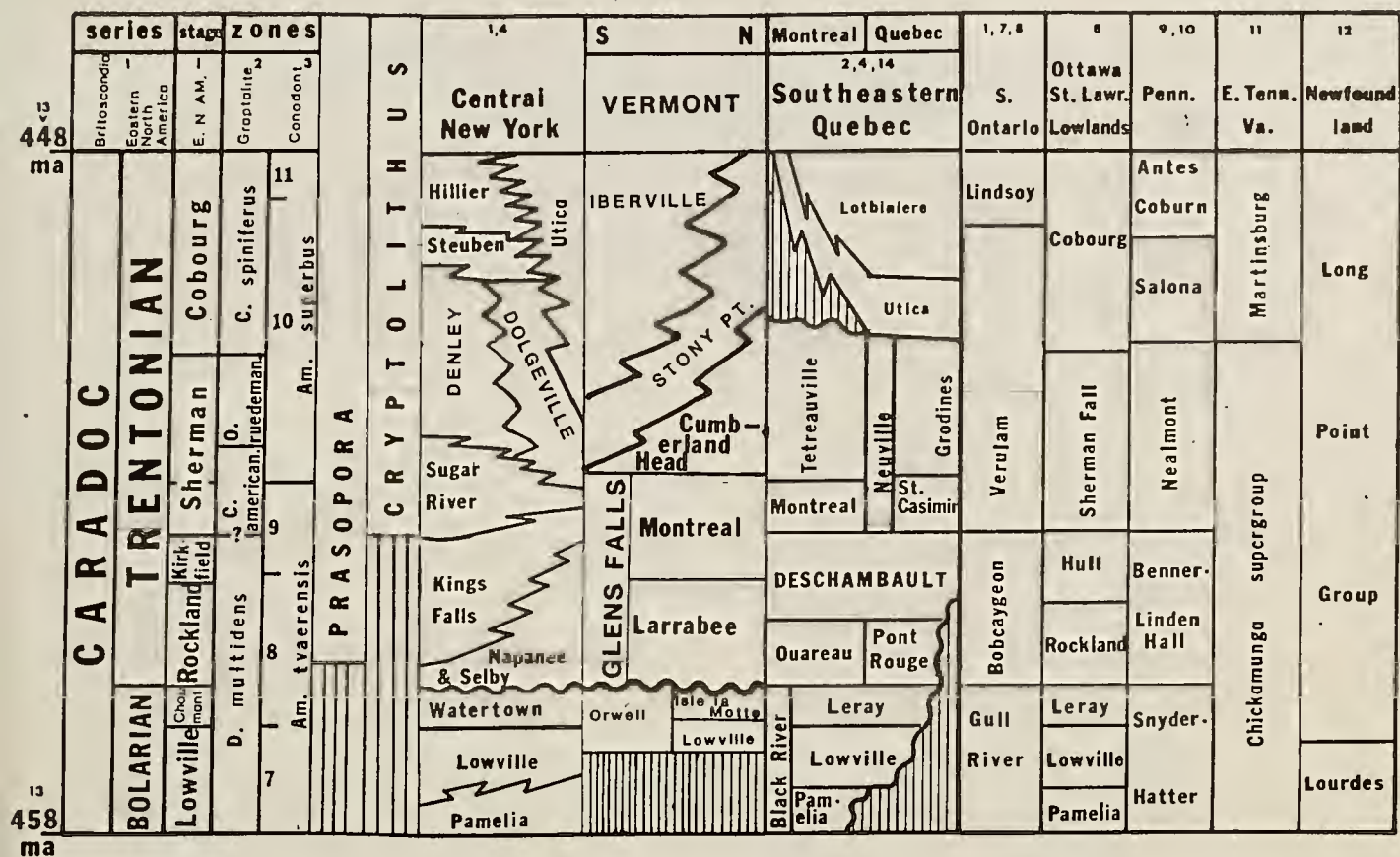


Figure 1. Time stratigraphy of the Middle Ordovician in the northeast. References listed are as follows: 1. Fisher, 1977., 2. Riva, 1974., 3. Sweet and Bergstrom, 1971., 4. Mehrtens, 1978., 5. Teetsel, 1984., 6. Fisher, 1968., 7. Brett and Brookfield, 1984., 8. Schopf, 1966., 9. Berg and others, 1980., 10. MacLachlan, 1967., 11. Ruppel and Walker, 1984., 12. Klappa and others, 1980., 13. Harland and others, 1983., 14. Belt and others, 1979.

Until recently, the lithologic units within the Glens Falls as well as other Trenton Group lithologies were thought to be laterally continuous and isochronous throughout the northeast. Largely through the work of Marshall Kay (1937, 1953) this supposed stratigraphic uniformity allowed the Trenton Group to become the standard stratigraphic section for the Middle Ordovician in North America. However, recent detailed studies by Cisne and Rabe (1978) and Cisne et al. (1982) have shown that the Trenton Group in the northeast contains localized and laterally discontinuous facies producing a rather complicated stratigraphy. Cisne et al. (1982) suggest that vertically abbreviated stratigraphic sections developed on downthrown fault blocks which were syndepositionally active. These block faults may be related to plate flexure associated with compression along the subducted slab in a manner suggested by Chappel (1973). During the closure of ocean basins, this flexure will pass under shelf deposits producing syndepositional block faults (Cohen, 1982). Cisne et al. (1982) note that a similar plate flexure and associated block faulting is currently occurring in the Sahul Shelf

and the Timor Trough, North of Australia. These block faults are the dominant control on the vertical distribution of sediments within this active subduction zone (Veevers et al. 1978).

Similar abbreviated vertical sequences and laterally discontinuous facies relationships can be documented in the Champlain Valley. Criteria useful for recognizing syndepositional block faults were developed by Cisne et al. (1982) and were summarized by Mehrtens (in press) including: 1. numerous unconformities within the sequence, 2. attenuated and compressed stratigraphic sequences, 3. major facies changes over short distances and 4. apparent anomolous juxtapositions of shallow and deep water sediments.

During this field trip, five stratigraphic sections will be analyzed and tested for these features. Time control is provided the first appearance of Cryptolithus and by the trepostome bryozoan Prasopora. Both Cryptolithus and Prasopora make their appearance in a wide variety of lithologies, suggesting that their first appearance is not facies controlled. Bathymetric control is obviously very important if shallow and deep water facies are to be differentiated. Bathymetry for each facies is determined by three methods: 1. recognition of bathymetric indicators used by Shanmugam and Walker (1978) and Benedict and Walker (1978) including algae, wave structures, graptolites, pyrite among other depth sensitive sedimentologic, biologic and chemical features, 2. recognition of trace fossil trends developed by Seilacher (1967, 1978) and 3. recognition of storm deposit/turbidite successions used by Kriesa (1981), Aigner (1982), and Handford (1985). Details and conclusions derived from these methods will be discussed for each outcrop.

OUTCROP LOCALITIES

The field trip is roughly divided into two parts. The first three stops will concentrate on the northern sections and the last two stops will concentrate on the southern sections. The general trend to remember is that the northern sequence is complete and gradational while the southern sequence is abbreviated and punctuated by rapid facies changes.

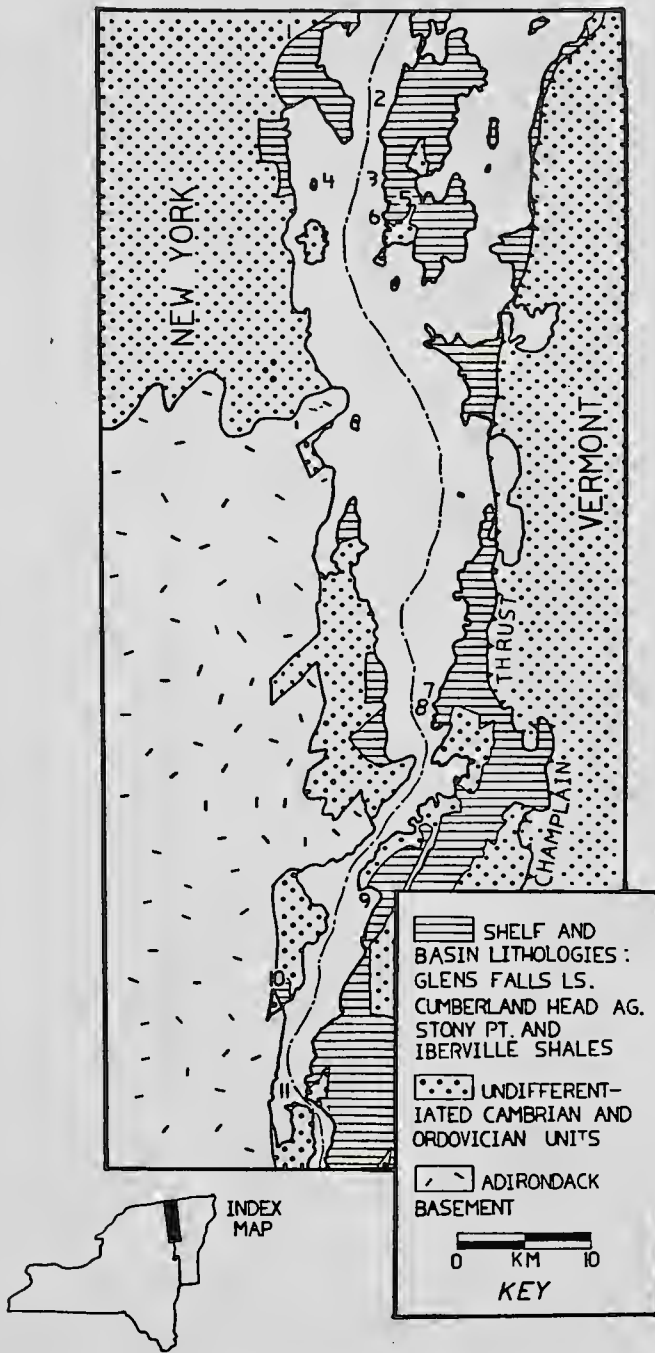
1. NORTHERN SEQUENCE

The northern sequence of the Glens Falls Formation is best exposed along the western shore of Grand Isle and in the Plattsburg area. This sequence records the gradual transgression and deepening of the foreland basin represented by a gradual succession from proximal grainstone facies, here termed facies A, to distal mudstone facies, here termed facies E.

Stop One: McBride Bay

A. Black River/Trenton Contact

The field trip begins on the Black River/Trenton Contact at the base of the McBride Bay section. The Black River Limestone is made up of massive oncolitic grainstone occasionally showing crossbedding. The transition to the thinner bedded Glens Falls Formation is sharp and, although no iron



LOCATION OF MEASURED SECTIONS

Figure 2. Outcrop localities measured as a part of this study. The field trip will make stops at 6) McBride Bay, 3) Rockwell Bay, 5) Lessor's Quarry, 8) Charlotte and 9) Button Bay.

mineralization has taken place on this surface, the abundant undercut surfaces and abrupt lithologic change suggests the presence of an unconformity. Above this contact, a unique, three meter, shallowing upward sequence grades from thin, nodular micritic mudstones and wackestones with occasional whole fossil grainstone lenses to more massive, algal rich grainstone. This interval is important for regional correlation since it compares well in scale, timing and lithology with the Selby Limestone of central New York described by Cameron and Mangion (1977).

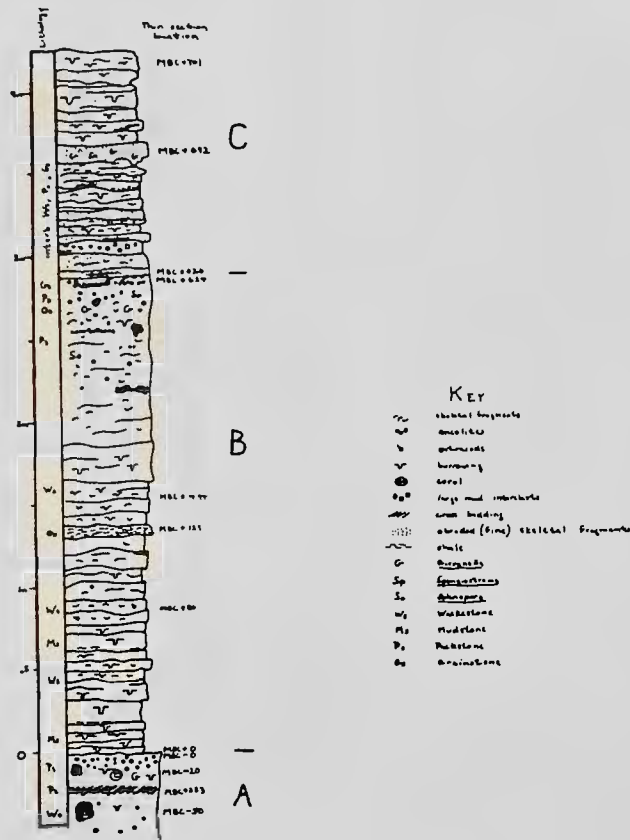


Figure 3. Measured stratigraphy immediately above the Black River/Glens Falls contact. The 0 locates the contact.

B. Grainstone - Facies A

Facies A begins above the massive, algal rich grainstone of the Selby-like transition interval located on the northern shore of McBride Bay.

Lithologies: Facies A consists of fine-grained pelbiosparite and pelbiomicrite grainstone, pelonkobiosparite grainstone and thin shale partings. Onkoids occur as algal coated trilobite and brachiopod fragments varying in size from 2mm to 2cm in diameter. Peloids occur as much smaller,

rounded, algal grains ranging from one to three phi in diameter with no encrusting bioclasts visible. Sub-rounded to angular mud clasts with diameters varying between 3mm and 1cm are present, but rare. Equant, fresh water cements are most common but rims of early, marine phreatic cements are seen in shelters of larger bioclasts.

Bedding Style: Bedding thickness ranges from 3cms to 6cms with thin shale partings less than 1cm thick punctuating bed contacts. Bedding is generally continuous but extremely wavy with abundant pinch and swell structures. Bedding planes are commonly scalloped, displaying sharp microtopography with relief locally up to 3cm in places.

Sedimentary Structures: Peloids and bioclasts of facies A are rounded and sorted, commonly displaying small scale cross lamination with wave amplitudes up to 4cm. Scours are especially common and bioclasts and onkoids are commonly bevelled at bed tops.

Biota: A typical, open shelf fauna is indicated by conspicuous amounts of brachiopods, gastropods and trilobites. Bryozoans and crinoids are rare. Larger onkoids appear to be constructed of the algae Spongiostroma and the smaller onkoids and peloids are constructed of the algae Girvanella. Solenopora is present but in much smaller quantities than other algae. Trace fossils are dominantly epifaunal including Helminthopsis Cruziana and Chondrites A. Boundaries between these traces and the surrounding sediment are generally sharp, occasionally noted by a change in sediment color.

Interpretations: These lithologies record moderately high energy, shallow water environments well within the photic zone and above normal wave base. The widespread algal growth in this facies substantiates depths in the photic zone and the wavy bedding style and sorted, abraded bioclasts and peloids suggests active wave reworking by currents or waves.

The thin grainstones were deposited during periods of high energy possibly related to storms. Hine *et al.* (1981) noted that movement of sand on the Bahamas platform requires storm energy to disrupt sediment binders and initiate particle movement.

During periods of low energy, thin shale rinds were deposited as the ambient sediment. Scalloped bedding planes were produced during these breaks in carbonate deposition with early marine cementation forming firmgrounds and hardgrounds.

C. Grainstone/Wackestone - Facies B

Lithologies: Facies B consists of skeletal biospararite grainstones and fine-grained grainstones surrounding horizons of nodular wackestones. Nodular lithologies consist of fine grained biosparite grainstone (calcsiltite) to biomicrite mudstone (average grain size less than three phi) and biomicrite wackestone. Bioclasts make up 10 percent of this nodular lithology, usually consisting of whole fossil debris. Matrix material consists of fine-grained bioclastic silt or micrite.

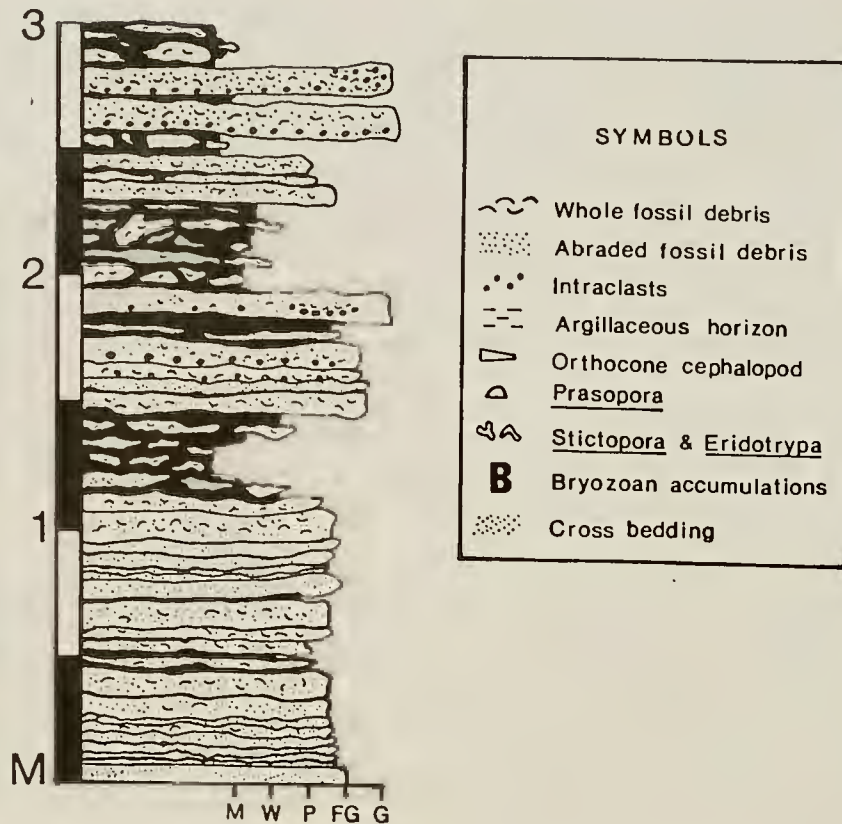


Figure 4. Lithologies and bedding style along the Facies A/Facies B transition: McBride Bay.

Grainstone horizons punctuate the nodular beds every 0.5 to 1.5 meters. These grainstones vary in thickness from around 7cms to 12cms. An idealized grainstone consists of a poorly washed, whole fossil lag base abruptly overlain by a laminated, finely ground grainstone. More commonly, either the basal lag or the laminated grainstone is absent. Rounded intraclasts are usually associated with the lags. The overlying biosparite grainstones are dominantly composed of abraded bioclasts and algal peloids averaging three phi in diameter. Scours are never seen separating the basal lag from the overlying laminated grainstone, suggesting that they were deposited as a coherent unit. Shale partings with an average thickness of 2cm are seen separating most beds.

Bedding Style: Nodular beds are generally discontinuous over a distance of 20cms occurring as a disrupted layer 10cms thick. In some cases, no bedding can be defined with the nodular beds chaotically occurring throughout a 20cm to 30cm thick package.

Compared with the nodular beds, grainstone beds are relatively continuous and well defined in outcrop. However, thickness within one bed varies dramatically, showing well defined pinch and swell structures and possible rippled surfaces. Basal lags are discontinuous over a few meters with some intraclastic rich lags defining shallow scours. Bedding surfaces can be either scalloped with abrupt microtopography or smooth and undulatory.

Sedimentary Structures: Nodular beds lack interpretable sedimentary structures. The nodules appear to be generally mottled and homogenized by bioturbation.

Grainstone commonly exhibit gradual distribution grading but exceedingly abrupt size grading. The lags are frequently ungraded with respect to allochem size but show a slight increase in mud percentage in upper laminae. Shells are dominantly oriented concave down in hydraulically stable position protecting muddy patches and displaying abundant shelter porosity.

Overlying finely-ground grainstones contain abundant parallel laminations of alternating dark silty and lighter bioclastic laminae varying from 5mm to 0.5mm in thickness. At the base of a laminated grainstone unit, the bioclastic laminae are thick and are separated by thin silty laminae. These bioclastic units become thicker and more frequent towards the top of a laminated unit. Irregularities such as low angle truncations are frequently seen. Transitions from lower plane laminae to upper hummocky laminae are also displayed. Hummocky laminae tend to parallel concave down truncation surfaces.

Biota: Nodular limestones appear intimately associated with the trilobite Isotellus, with large, complete specimens exposed at the margins of nodules. Gastropods and brachiopods are common and cephalopods are occasionally found. Bryozoan biostromes are sometimes seen in the shaley intervals separating beds. Grainstones contain much the same fauna as the nodular beds. Small peloids of the algae Grivanelia between two and three phi are common. Whole fossil gastropods and large Isotellus fragments are common. Shell lags contain dominantly disarticulated, whole fossil brachiopod valves and large gastropod and trilobite fragments.

Ichnofauna assemblages are generally identical to facies A. Cruziana, Helminthopsis, Chondrites A and to a lesser extent, Chondrites B are noted as well defined, sharp traces on bedding planes of grainstone beds. Evidence of infaunal deposit feeding is rare. Traces in the nodular limestones are difficult to identify since these beds are mottled and generally homogenized. Cruziana, however, appears quite commonly.

Interpretations: Abundant algal peloids and whole fossil gastropods suggest that these lithologies were deposited within the photic zone (Benedict and Walker, 1978).

Grainstone lags and overlying laminated grainstone represent event deposits separating periods of ambient carbonate mud sedimentation represented by the nodular wackestones.

Trace fossils found in this facies comprise a shallow water assemblage based roughly on the work of Seilacher (1967, 1978) and Pickerall and Forbes (1979). These traces are produced as habitation burrows and locomotion burrows as opposed to deposit feeding burrows.

Nodular beds are produced by a combination of bioturbation and differential compaction. Original carbonate muds were winnowed from facies A and deposited in a low energy environment and were subsequently churned by large, trowel shaped Isotelid trilobites. A small amount of early cementation sufficiently adhered micrite particles together so that the nodules remained intact rather than becoming fluidized sediment. Differential compaction due to loading of overlying sediments further enhanced the production of nodular bedding. A similar scenario has been suggested for the development of nodules in subtidal Holocene sediments in the Bahamas by Mullins et al., (1980).

Whole fossil lags and overlying laminated grainstones (sandy couplets) represent proximal, storm winnowed tempestites. Passage of large wavelength waves during storms disrupted the sediments and entrained shells, sand and mud into suspension. Shells settled out first forming the lags while sand and silt were deposited as the storm waned. Protected mud patches indicate that storm flushing and entrainment of shelly material was often incomplete and suggests generally in situ development (Kriesa, 1982; Aigner, 1982).

Laminated portions of the grainstones represent reworking and redeposition of fine-grained material during the waning flow of a storm event. Infiltration structures, fine sand and silt passing down into the flushed lag, are commonly produced during this time. The sequence of thinning bioclastic laminae is strikingly similar to the graded rhythmites noted in cores of Holocene shallow shelf (less than 40 meters of water) sediments by Reineck and Singh (1972). These authors suggest that graded rhythmites form from the deposition of suspended fine material during waning storm events. The presence of parallel lamination, low angle truncations and hummocky cross stratification further substantiates the hypothesis that these deposits formed from suspension clouds created during high energy storm events. The initial settlement of suspended particles was followed by traction currents which frequently mobilized these sediments and formed occasional ripples similar to those found in the grainstones of facies.

Stop Two: Rockwell Bay

This second stop exhibits muddier lithologies that have traditionally been grouped into the Montreal Member and are here termed mudstone Facies E. Slightly more skeletal lithologies, termed Facies D, will not be seen on this field trip.

D. Mudstone Facies E.

Lithologies: Facies E consists of laminated biomicrite mudstones with very thin shelly bases occasionally present. In some cases, the mudstones are subtly graded in grain size from fine bioclastic lime silt to lime mud. Bioclastic grainstone bases are rare and are generally thin (less than 2cm) consisting of coarsely fragmented fossil debris with an occasional whole fossil bioclast (typically brachiopods or trilobites).

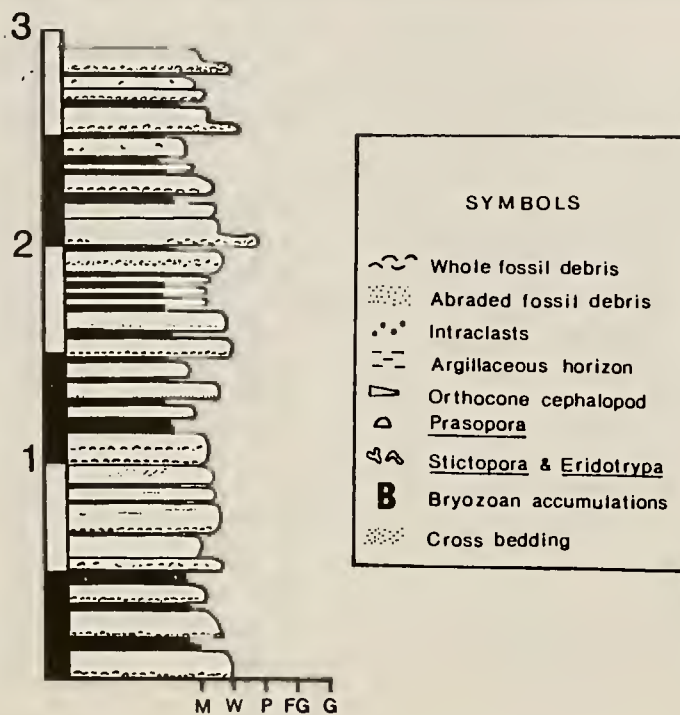


Figure 5. Lithologies and Bedding Style of Facies E exposed at Rockwell Bay.

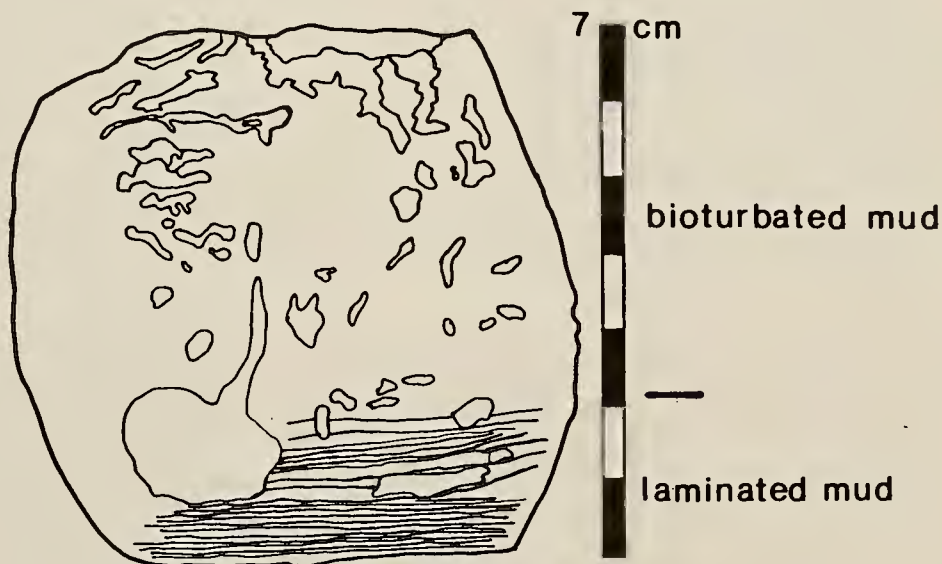


Figure 6. Line diagram of an acetate peel taken from Rockwell Bay. Note the concentrated bioturbation at the bed top and the preserved lamination at the base. These structures indicate rapid episodic deposition.

Bedding Style: Bedding is tabular and continuous with the bed thickness varying little over an outcrop. The basal lags are generally discontinuous over one to three meters.

Sedimentary Structures: These micrites appear structureless but upon closer inspection, they reveal several interesting features. Both distribution and grain size grading are present. Laminae thicknesses generally grade from greater than 1mm to less than 0.1mm at the bed tops. This lamination scheme is generally continuous across one bed but small, low angle truncations can be detected. Occasionally, these laminae are highlighted by small amounts of euhedral, authigenic pyrite. The laminae are generally irregular in thickness making tracing of laminae difficult across the scale of a large thin section. These irregularities are sometimes related to bioturbation but in other cases the pinching and swelling is undoubtedly primary.

Lamination is rarely visible in the coarser bioclastic grainstone bases, although subtle bands of sorted material are sometimes observable. Larger bioclasts are generally seen in hydrodynamically stable, concave-down position. Escape burrows are occasionally seen extending up from the muddy tops of lower beds into the coarser fraction of overlying beds.

Biota: Although shelly forms are common in facies E, a general decrease in fossilized forms is seen in comparison to facies A and B. Bryozoans, crinoids and brachiopods are most common along with the notable introduction of graptolites and the trilobite Triarthrus. Trace fossils are restricted to infaunal forms with the trace fossil Teichichnus and Chondrites B overwhelmingly dominant. Traces seem to be restricted to the uppermost few cms of the beds.

Interpretations: These beds represent storm generated turbidity current deposits. These turbidites are base absent containing BDE Bouma cycles (mudstones with thin shelly bases) and DE cycles (graded laminated mudstones). The presence of escape burrows commonly associated with these deposits suggests the deposition of these suspended muds was rapid (Kriese, 1981).

Storm generated turbidites were emplaced below storm wave base in a low energy, low oxygen environment as indicated by graptolites, the presence of the deep water trilobite Triarthrus and pyrite. The predominance of trace fossils over relative shelly, in situ forms suggests that nutrient rich currents were no longer available to filter feeding organisms and the remaining life forms were detritus feeders.

Based on the works of Fursich (1975), Pickerill and Forbes (1979), and Seilacher (1967, 1978), the deposit feeding burrows roughly comprise a deep water assemblage.

Stop Three: Lessor's Quarry

Although no true biohermal buildups are visible in the Glens Falls Formation, evidence of extensive biostromes and flanking bioclastic deposits are visible here at Lessor's Quarry.

E. Bryozoan Facies F.

Lithologies: Three lithologies are associated with the bryozoan facies:

1. Bryozoan rich biomicrite wackestone, packstones and baffle-stone containing large dendrils over 2cms long of the ramose bryozoans Stictopora and Eridotrypa. Commonly these bryozoans are fragmented and intermixed with an open shelf fauna. The fragmented bryozoans can make up over 90 percent of the packstone. These packstones vary from one to five cms in thickness and are laterally discontinuous over several meters.

2. Discontinuous and thin, abraded grainstones and packstones varying from less than one cm to four cms in thickness. Bioclasts are rounded and abraded to medium fine and fine sand size. These fine skeletal beds are discontinuous over the scale of one meter.

3. Laminated and burrowed wackestones and mudstones composed generally of very fine silt and mud forming weakly defined but laterally continuous beds.

Bedding Style: In contrast to the other facies, interbedded shales are not found. Instead, the rock is massive and devoid of bedding plane partings. Bases of bryozoan accumulations are sharp and sometimes concave up, defining broad, thin lenses. Grainstones for much smaller, sharply defined scours. Mudstone laminations are generally continuous when not disrupted by abundant bioturbation.

Sedimentary Structures: The bryozoan bafflestones are structureless. Matrix material is chaotic and lamination free. Fossil debris intermixed within the bafflestones is dominantly whole fossil with no hydrodynamic orientation or shelter cement.

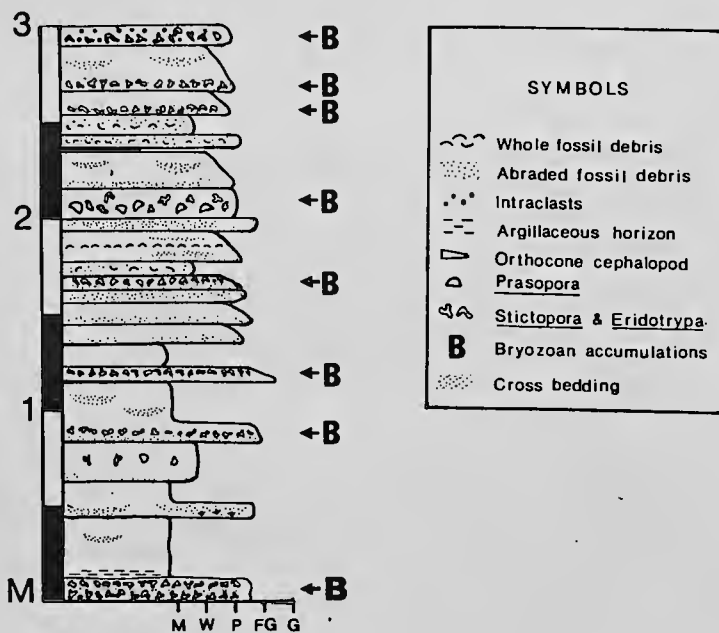


Figure 7. Bedding style of Facies F taken from the Lessor's Quarry measured section.

Biota: This facies contains possibly the widest diversity of fauna in the Glens Falls Formation. Bryozoans are dominant but trilobites, gastropods and brachiopods are also present. Notably absent are crinoids, cephalopods and graptolites. Traces are abundant and diverse including dominantly infaunal forms, which may reflect more of a preservational bias than an environmental characteristic. Bedding planes needed to check for epifaunal trace development are rarely exposed due to the massive nature of the bedding. Elongate thin borings of Trypanites are commonly seen in large Prasopora colonies.

Interpretations: This facies accumulated a short distance from a large bryozoan accumulation. Bryozoans acted primarily as sediment contributors (bryopackstones) and secondarily as sediment inhibitors (bafflestones) (Cuffey, 1977). Currents transported material across bryozoan-rich biostromes and deposited much of the skeletal material in flanking areas. Occasionally, rigid, ramose, bryozoans grew up from shelly pavements to produce minor, laterally discontinuous biostromes. Organic and skeletal material was produced in abundance in these biostromes and depositional rates in areas fringing larger biostromal masses must have been fairly high and constant. The fairly constant accumulation of skeletal matter precluded the accumulation of shale interbeds.

These bryozoan accumulations accumulated at moderate depths between normal and storm wave base. Algae are rare suggesting depths below the photic zone. Pickerill et al. (1984) suggests that the Trypanites borings were occupied by filter feeding organisms suggesting that waters were well circulated and oxygenated. Strong currents commonly swept these areas producing the abundant scouring and channeling. Laminated mud overlies these lithologies, possibly representing periods of slightly lower energy, open shelf sedimentation.

F. Northern Sequence: Conclusions

Based on the addition of incomplete sections and their stratigraphic position relative to Prasopora and Cryptolithus time lines, the northern sections of the Glens Falls Formation reaches a thickness of at least 90 meters.

The bathymetric indicators, trace fossil succession and storm/turbidite succession all show a gradual replacement of shallow water characteristics by deep water characteristics. This is further corroborated by the decrease in shelly bed thickness and frequency with increased height in the sequence. In conclusion, the features of the northern sequence suggest deposition of a slowly subsiding ramp with cyclisity of facies occurring in the shallower water facies.

2. Southern Sequence

The southern sequence, exposed along a narrow band paralleling Lake Champlain from Charlotte to Bridport, Vermont, is much thinner in comparison reaching only 50 meters. Facies transitions are more abrupt relative to the northern sequence with several facies omitted from the succession suggesting more rapid subsidence in this area.

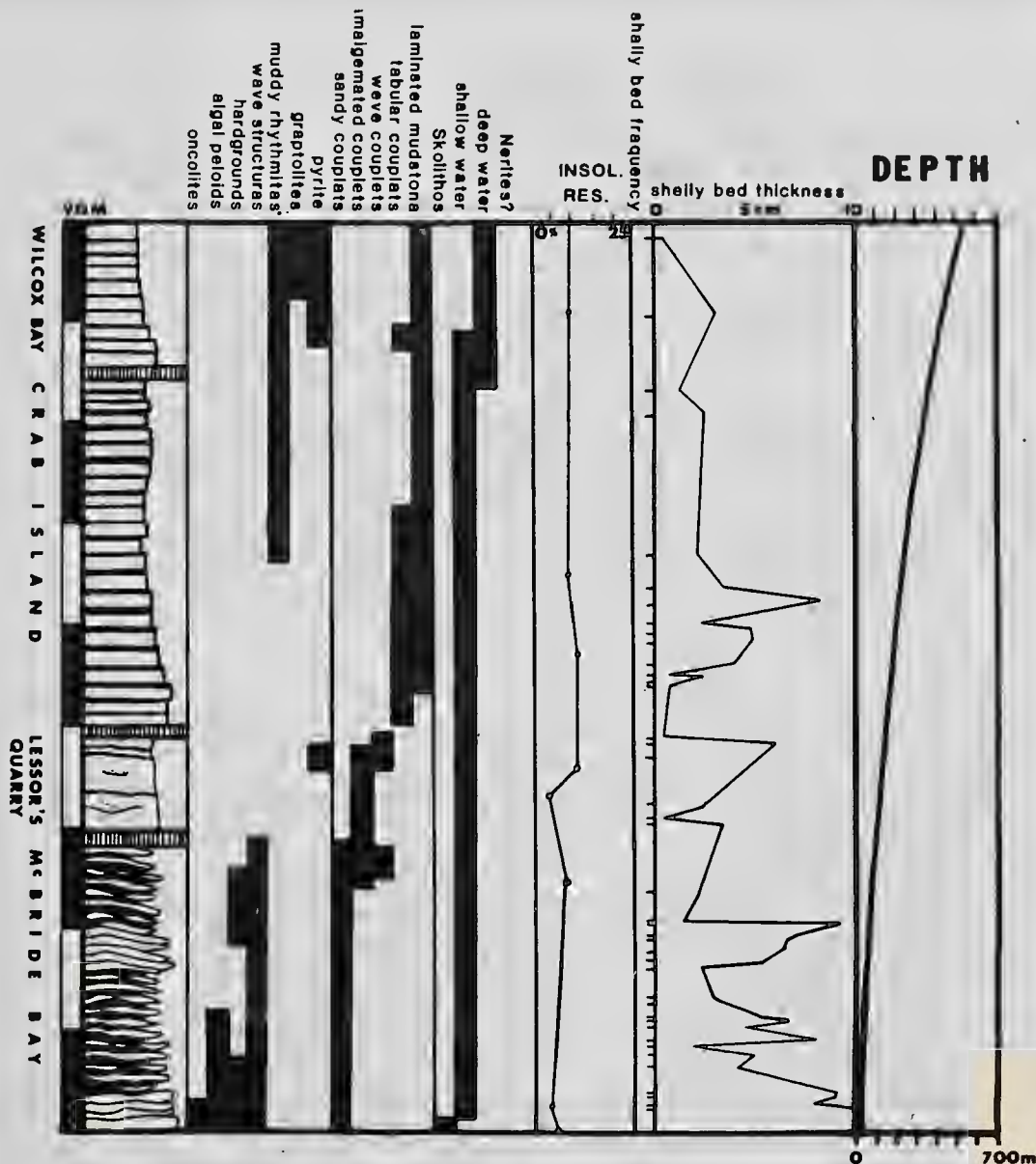


Figure 8. Bathymetric trends within the Northern Sequence. A gradual succession of bathymetric indicators suggests a gradual deepening within the sequence.

Locality Four: Charlotte

Here at Charlotte some differences in the regional stratigraphy become apparent. The Black River Formation, consisting of light grey massive, structureless micrite is exposed on the small peninsula to the west of the Glen Falls section. Facies A and B still show up here but their vertical extent is limited and no cyclisity is noted. Facies A is exposed as a small weathered bench of massive limestone on the southern end of the outcrop. The laterally continuous, finely ground grainstones separating nodular intervals typical of facies B can be seen above this bench. These beds are similar to facies B with the exception that these beds contain more micrite and their upper surfaces are more undulatory suggesting more wave reworking. With increased height in the section, argillaceous content sharply increases and an argillaceous facies, facies G, is introduced.

A. Facies G, Argillaceous Facies

Lithologies: Facies G consists of argillaceous mudstones and wackestones varying from 10 to 20 cms with thick shale interbeds. Insoluble content in these wackestone is high ranging from 10 to 24 percent relative to the other facies with insoluble contents averaging around 6 to 8 percent.

Bedding Style: Beds are massive and generally tabular. Occasional lensing of beds is common but basal scours are absent.

Sedimentary Structures: Subtle changes in grain size produce a laminated texture visible in some weathered beds. These laminations are generally parallel and regular, showing thickening and thinning in bed thickness. Unlike the mudstones of facies E seen at Rockwell Bay, these laminations show no rhythm or grading. Occasionally, pyrite will highlight these laminations. Often bioturbation will completely obliterate these laminations, homogenizing and mottling the micrite.

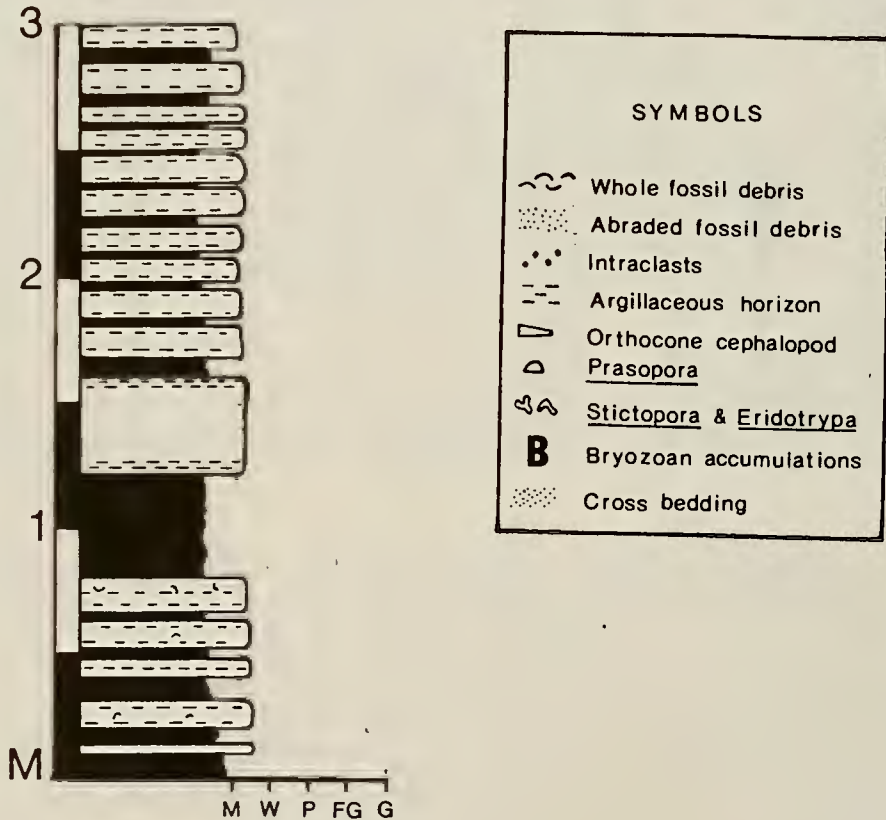


Figure 9. Bedding style and lithologies of facies G - argillaceous facies (taken from Westport, New York section not on this field trip).

Biota: Biota are generally scarce. Occasionally, bryozoans will be found in conjunction with this facies and laminae of abraded shells may be associated with the more calcareous beds. The interbedded shales are frequently more fossiliferous containing abundant bryozoans and other abraded bioclasts.

Interpretations: These argillaceous beds represent an increase in pelagic to hemipelagic sedimentation separating storm generated, fine grained turbidites. As noted by Hesse (1975) turbiditic deposits are bioturbated at bed tops while pelagic beds are either unburrowed or wholly burrowed throughout. No grading or recolonization extends down from the upper surfaces suggesting that deposition was continuous and not episodic. Apparently, these beds originated as a sudden influx of terrigenous material into the basin.

This facies becomes more distinct in southern sections as we will see in the next stop.

Locality Five: Button Bay

In this last outcrop, the only known complete section of the Glens Falls Limestone is exposed stretching from massive micrite of the Black River Formation to the argillite and shale of the Cumberland Head Argillite. Like the northern sequence and the Charlotte section, the Black River/Glens Falls contact is sharp possibly representing a minor disconformity. Undercut surfaces are rare suggesting that hardground development is minimal. The shallowing up interval associated with the contact in the northern sequence is not exposed and facies A is poorly developed. Oncolites and abraded algal grains are not present in either the Charlotte or the Button Bay sections.

In comparison to the gradual increase in mudstone percentage noted in the northern sequence, the lithologies of the southern sequence remain more or less constant. The repetitious pattern of finely ground grainstones punctuating sequences of nodular micrites and wackestones is no longer seen. Instead, the great majority of the Glens Falls Limestone in the southern sequence consists of muddy couplets with localized bryozoan mats here termed facies C.

B. Facies C

Lithologies: Facies C consists of three lithologies: 1) thin, fine grained and whole fossil grainstones overlain by 2) sparsely fossiliferous biomicrite wackestone, bryozoan packstone and mudstones and 3) thin interbedded shale. Whole fossil and finely ground fossiliferous bases are thinner than the grainstone lags of facies B, ranging from 2cms to 4cms in thickness and are laterally discontinuous over several meters. Laminated wackestones and mudstones frequently lack the thin grainstone bases seen in facies B. Bryozoan beds occur as discrete units between the grainstone/mudstone couplets and less fossiliferous mudstones. Occasionally the tops of a grainstone/mudstone couplet will be colonized by bryozoans.

Bedding Style: Beds are laterally continuous but pinching and swelling of beds is common. Hummocks occur as beds with planar bases and concave down tops. Fine-grained shell debris frequently fills troughs and swells at the bases of beds. The bryozoan beds also show similar pinch and swell structures. Bedding planes are even and swaley as opposed to the scalloped bedding planes associated with the facies A and B.

Sedimentary Structures: Parallel lamination is visible in the grainstones bases as slight variations in bioclast grain size. Size grading is abrupt between shell material and laminated mud. As in the case with the grainstones of facies B, distribution grading appears gradual with mud content progressively increasing upwards. Bryozoan packstones appear to have little internal stratification and bryozoan branches appear to float in a lime mud matrix.

Biota: Important faunal characteristics include a decrease in algal peloids and an increase in bryozoan content. Brachiopods and gastropods are still abundant but cephalopods are more rare. Stictopora and Eridotrypa appear to be the most common bryozoans in the muddy packstones.

Epifaunal traces, especially Chondrites A, appear less dominant in this facies. Infaunal traces such as Chondrites B appear much more commonly, frequently burrowing the upper parts of grainstone/mudstone couplets. Helminthopsis frequently appears on bedding planes but trace walls are noticeably less distinct.

Interpretations: The limestones of facies C record deposition in slightly deeper water than facies A and B as suggested by rarer algal peloids. Also, the predominantly finely-grained grainstones of facies A and B have been replaced by muddier lithologies. This fining-upward trend is to be expected in transgressive sequences with deeper water environments becoming progressively more mud-rich (Aigner, 1982). Of course, basinward sediment transport, local facies relations and available sediment sizes can greatly modify this trend.

Grainstone/mudstone couplets represent slightly more distal storm deposits when compared to the proximal storm deposits of facies B. The abraded and sorted bioclastic bases are formed by dominantly traction currents related to storm surge ebb flow. The overlying mud, deposited during waning flow conditions, suggests that the lags were emplaced in a quiet, deeper water setting than facies B.

Wave action still played a major role by shaping beds into the hummocks and pinch and swell features that characterize this facies.

Bryozoans periodically inhabited sea floor forming patchy, bryozoan mats. These mats appeared as sediment baffles and localized biostromes. Studies on recent bryozoans suggests that their distribution on the sea floor is dominantly controlled by sedimentation rates with maximum growth occurring in areas of little siltation (Lagaaji and Gautier, 1965). Therefore, these bryozoan beds represent subtidal deposition relatively removed from the shoreline and clastic sources.

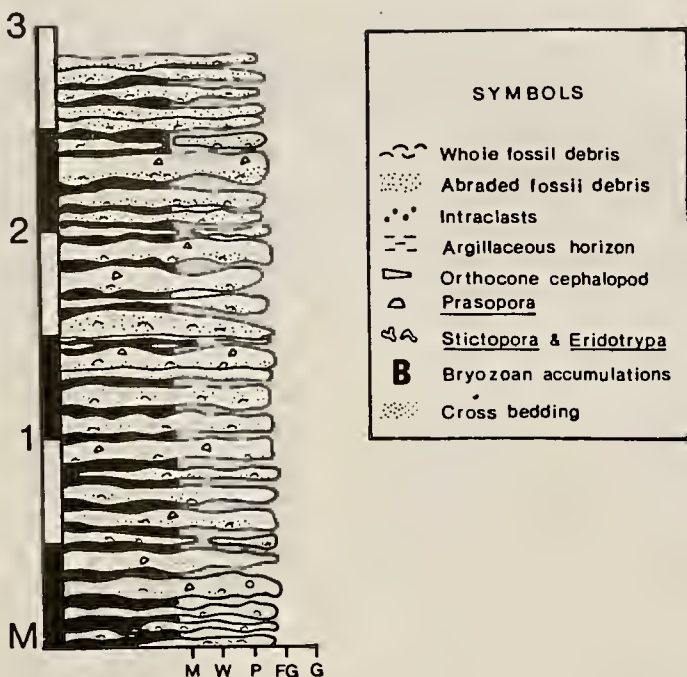


Figure 10. Bedding style and lithologies of facies C. taken from the measured section of Button Bay.



Figure 11. Bathymetric trends within the southern sequence exposed in the Button Bay Section.

Argillaceous micrites of facies G punctuate the lithologies in facies C and these argillaceous tongues can be traced north to Charlotte and south to Westport, New York.

The contact between the Glens Falls Limestone is noticeably sharp with muddy couplets in facies C passing into thick argillite and shale interbeds. No evidence of the deeper water carbonate turbidites of Rockwell Bay are seen.

Southern Sequence: Conclusions

The abbreviated nature of the Glens Falls Formation may represent deposition of a rapidly subsiding fault block. Cisne *et al.* (1982) note that abbreviated sections in the Trenton Group from the central New York area may represent deposition on a down thrown fault block, with nodular and wavy lithologies rapidly overlain by Utica Shale. A similar mechanism may be controlling the abbreviated distribution of lithologies noted in the southern sequence of the Glens Falls Formation.

Basin evolution is envisioned to involve the following steps. Sometime after the passing of the peripheral bulge and deposition of the Black River Group, the Champlain Valley began to rapidly subside. This is recorded as the sharp transition from massively bedded, shoal water limestone to interbedded, open shelf limestone with thin shale interbeds. Rapid downfaulting in the southern Champlain Valley caused muddy couplets representing storm surge triggered turbidites to accumulate rapidly. Less rapid subsidence in the north allowed the development of shelf cycles seen at the McBride Bay section. As the rapid subsidence continued in the south, the thin Glens Falls sequence was overlain by a thick sedimentary pile of shale. The northern sequence accumulated a more complete sequence of proximal storm deposits to distal storm surge turbidites overlain by a thick Cumberland Head Argillite sequence. Eventually, flysch deposits of the Stoney Point and Iberville filled the entire basin and molasse sedimentation began.

Revised Stratigraphy

The facies descriptions of Chapter Two, the Time-Space diagram and these conclusions suggest that a re-evaluation of the Champlain Valley stratigraphy is necessary. It is possible that facies were laterally discontinuous in contrast to the extreme lateral continuity of Trenton facies suggested by Kay (1937, 1953). The use of the Montreal and Larrabee member names are still recommended although lithologic revisions are noted in these members. Primarily, the Larrabee encompasses facies C in the south while to the north, the Larrabee encompasses facies B which is lithologically similar to the Deschambault as described by Parker (1986). The Montreal Member includes tabular facies which appear lithologically similar to the Montreal Formation and the overlying Tetreauville Formation. The shallowing up interval at the base of the McBride Bay section is lithologically similar to the Selby Formation on the basis of Cameron and Mangion's 1977 description, and like the Montreal Member, the presence of the Selby Formation is restricted to the northern part of the basin.

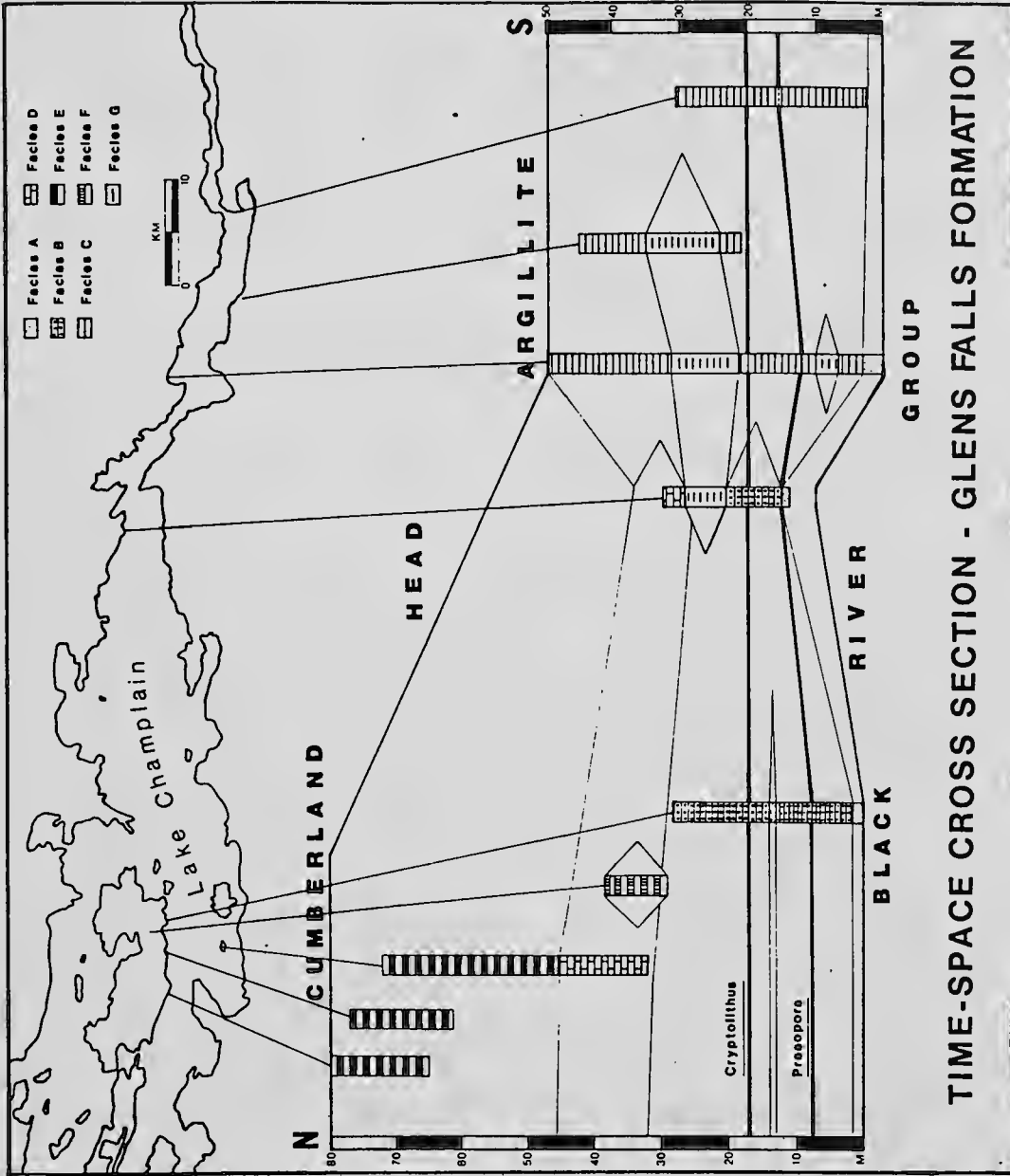


Figure 12. Bathymetric trends within the southern sequence exposed in the Button Bay Section.

Thickness data suggest that the Glens Falls and the overlying Cumberland Head Formation pinch out to the south. This facies relationship is poorly constrained in time and extensive graptolite work along the Cumberland Head/Glens Falls formational boundary could produce some interesting results. The southern pinch out, most likely produced by deposition on a subsiding fault block, coincides well with the facies relationships of the overlying shales suggested by Fisher (1977) and Teetsel (1984). With the conclusions of these works coupled with information compiled in this study, a more realistic picture of deposition in a tectonically active basin is beginning to unfold.

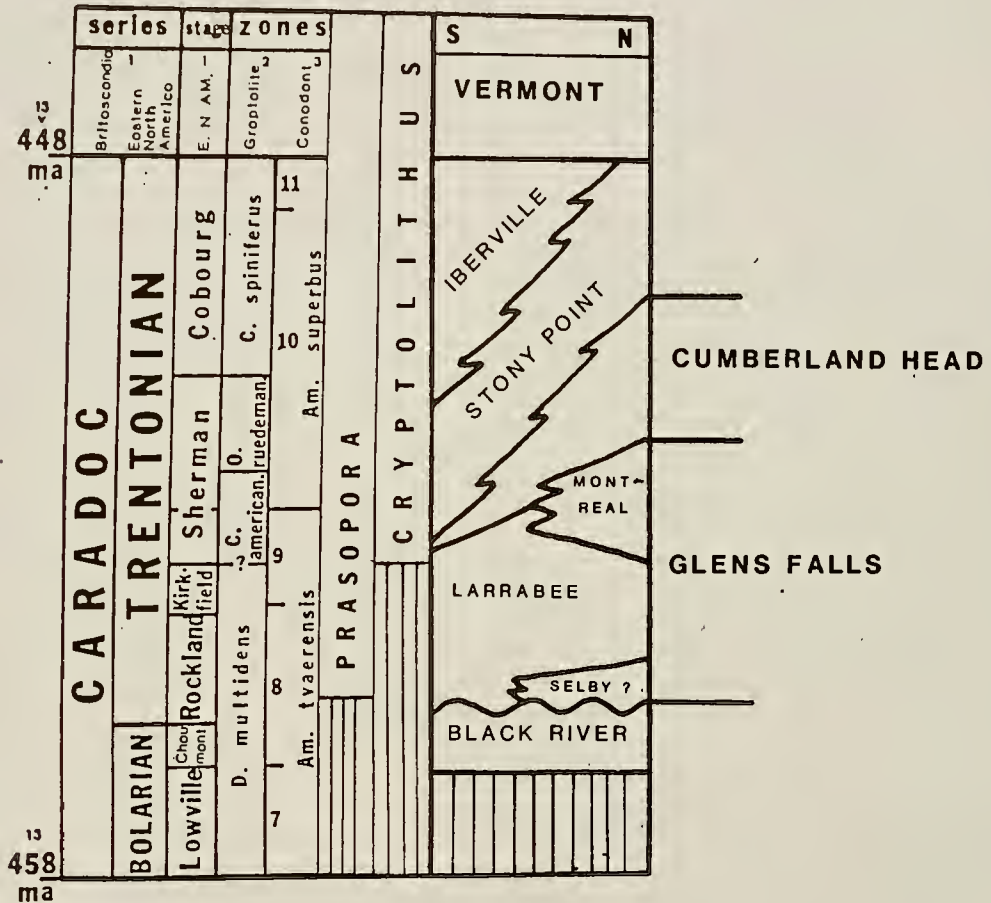


Figure 13. Revised stratigraphy of the Middle Ordovician in the Champlain Valley of Vermont and New York showing the facies relationships within the Glens Falls Limestone.

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Acknowledgements

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ITINERARY

- 0.0 Assembly point is at the Allen Apple Barn in the town of South Hero on Route 2.
- 0.5 Turn left onto Station Road.
- 0.7 Bear left at the fork and continue on Station Road.
- 2.3 Turn right onto West Shore Road.
- 2.4 LOCALITY 1: McBride Bay - park cars on the side of West Shore Road and continue down dirt driveway to the lakefront. Refer to the text for information concerning the locality. Private residence; NO HAMMERS, PLEASE.
- Return to cars and proceed north on West Shore Road.
- 4.5 LOCALITY 2: Rockwell Bay - park cars on the left side of the road and continue north along the shoreline to the limestone outcrops on the point. Refer to text for information concerning the locality. Proceed south on West Shore Road.
- 5.9 Turn left onto Sunset Hill Road.
- 6.5 LOCALITY 3: Lessor's Quarry - either pull into the field through the barbed wire gate or park along the side of Sunset Hill Road. The quarry lies approximately 200 yards to the south of the barbed wire gate.
- Return to cars and proceed east on Sunset Hill Road.
- 7.1 Turn right onto Route 2. Proceed off Grand Isle towards Burlington, Vermont.
- 16.7 Turn right heading south on Interstate 89.
- 26.1 Take the first Burlington Exit heading west on Route 2.
- 26.4 Take first left heading south on Spear Street.
- 33.3 Turn right heading west down hill on Irish Hill Road.
- 34.8 Turn left at the blinking light heading south on Route 7.
- 39.0 Turn right at the blinking light onto Route F-5 towards the ferry to New York State.
- 41.4 Keep straight onto dirt road past signs pointing south to the ferry to New York State. Route F-5 banks sharply to the left.

- 41.6 LOCALITY 4: Charlotte - turn into Charlotte Children's Center and park in the parking lot. Follow the path behind the children's center to the lakefront.
- Return to the cars and proceed west onto the dirt road.
- 41.8 Continue straight onto the paved road (Route F-5).
- 43.1 Turn right onto Route 7 south.
- 52.2 Turn right on Route 22A heading south towards Vergennes.
- 53.9 Turn right onto Panton Road heading west towards Basin Harbor.
- 55.5 Turn right onto Basin Harbor Road.
- 59.6 Turn left at the T and turn left onto Jersey Road (road is unmarked).
- 61.4 Turn right onto dirt road directly after the First Season Greenhouses.
- 62.3 LOCALITY 5: Button Bay - proceed into the driveway on the right; the house is a large, modern log home surrounded by large pine trees. Walk down the path to the lakefront.

END OF FIELD TRIP

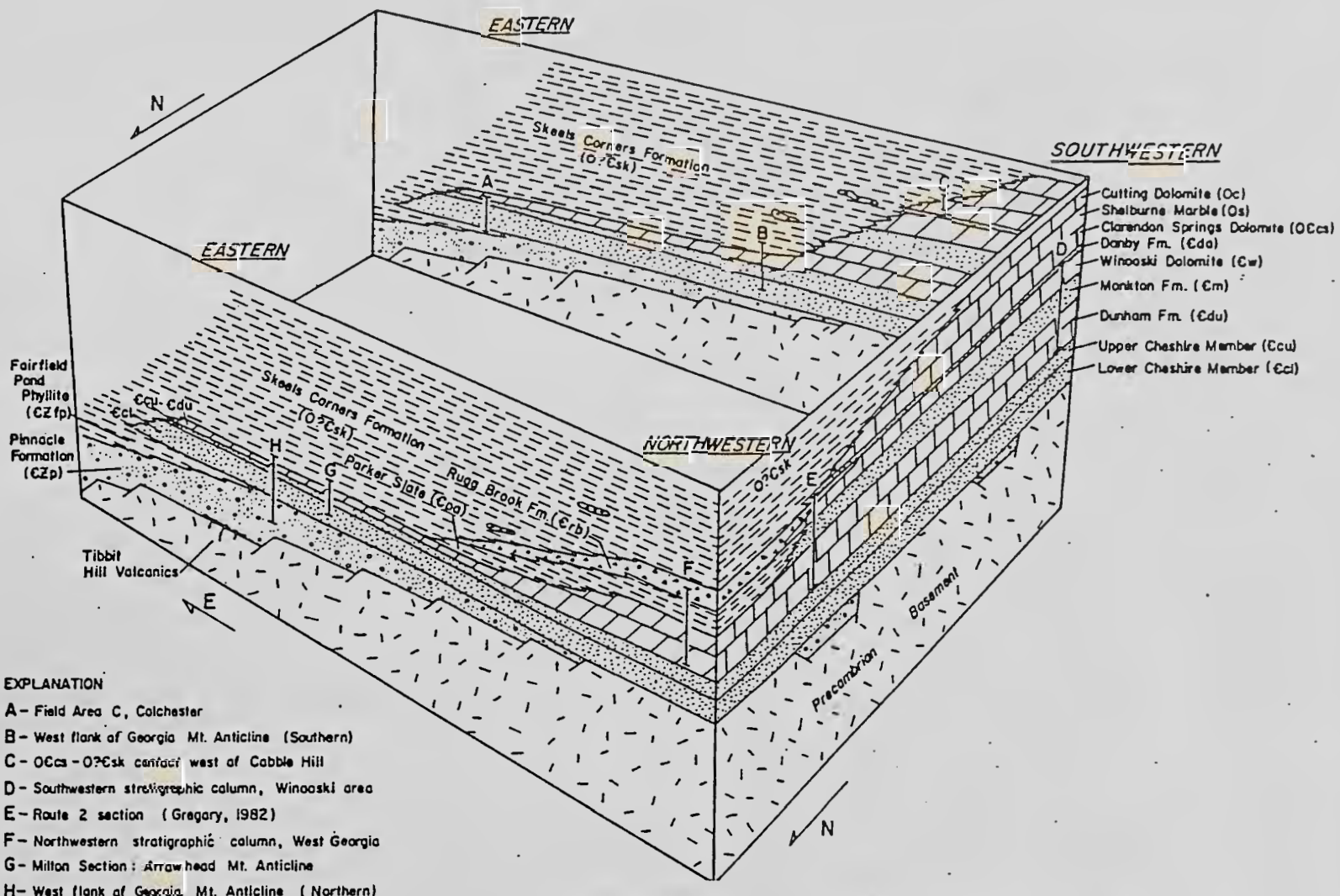
A TRANSECT THROUGH THE FORELAND AND TRANSITIONAL ZONE OF WESTERN
VERMONT

by
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Department of Geology
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INTRODUCTION

Western Vermont is underlain by three distinctive sequences of rocks that range in age from Late Proterozoic to Middle Ordovician and are typical of the western part of the Appalachian Mountains. The lowest most sequence, which rest with profound unconformity on the Middle Proterozoic of the Green Mountain and Lincoln massifs, largely consists of metawackes, mafic volcanic rocks and phyllites that represent a rift clastic sequence. These rocks grade upward into siliciclastic and carbonate rocks of the platform sequence which in turn are overlain by Middle Ordovician shales of the foreland basin sequence. The boundary between the two sequences is the base of the Cheshire Formation (fig. 1). North of Burlington, Vermont the platform sequence grades into shales, breccias and conglomerates of the ancient platform margin and eastern basin. These sequences have been studied by a number of workers in the past (Cady, 1945; Cady and others, 1962; Hawley, 1957; Erwin, 1957; Welby, 1961; Stone and Dennis, 1964, for example) and are receiving current attention by Mehrtens (1985; in press) and her students (Gregory, 1982; Myrow, 1983; Teetsel, 1985; Bulter, 1986; MacLean, 1986). Agnew (1977), Carter (1979), Tauvers (1982), DiPietro (1983) and Dorsey and others (1983) have reexamined parts of the rift clastic sequence while Doolan and his students are currently working in the same sequence in northern Vermont and Quebec. Figures 1 and 2 illustrate the representative stratigraphic columns for western Vermont north of the latitude of the Lincoln massif where the field conference will be held. Additional information can be found in Welby (1961) and Doll and others (1961).

The structure of western Vermont is dominated by major, north-trending folds and imbricate thrust faults which are well displayed on the Geologic Map of Vermont (Doll and others, 1961). The rift clastic and platform sequences have each been displaced westward on major thrust faults that extend through much of Vermont. The larger of the two, the Champlain thrust, extends from southern Quebec to Albany, New York and places the older platform sequence over the younger foreland basin sequence. Estimates of westward displacement range from 15 km to 100 km. The smaller of the two, the Hinesburg thrust, places transitional and rift clastic rocks over the platform sequence and as such forms a boundary between the synclinal rocks of western Vermont and the Green Mountain anticlinorium. Dorsey and others (1983) have demonstrated that the Hinesburg thrust developed along the overturned limb of a large recumbent fold (fault-propagation fold of Suppe, 1985) and therefore is quite different from the geometry of the Champlain



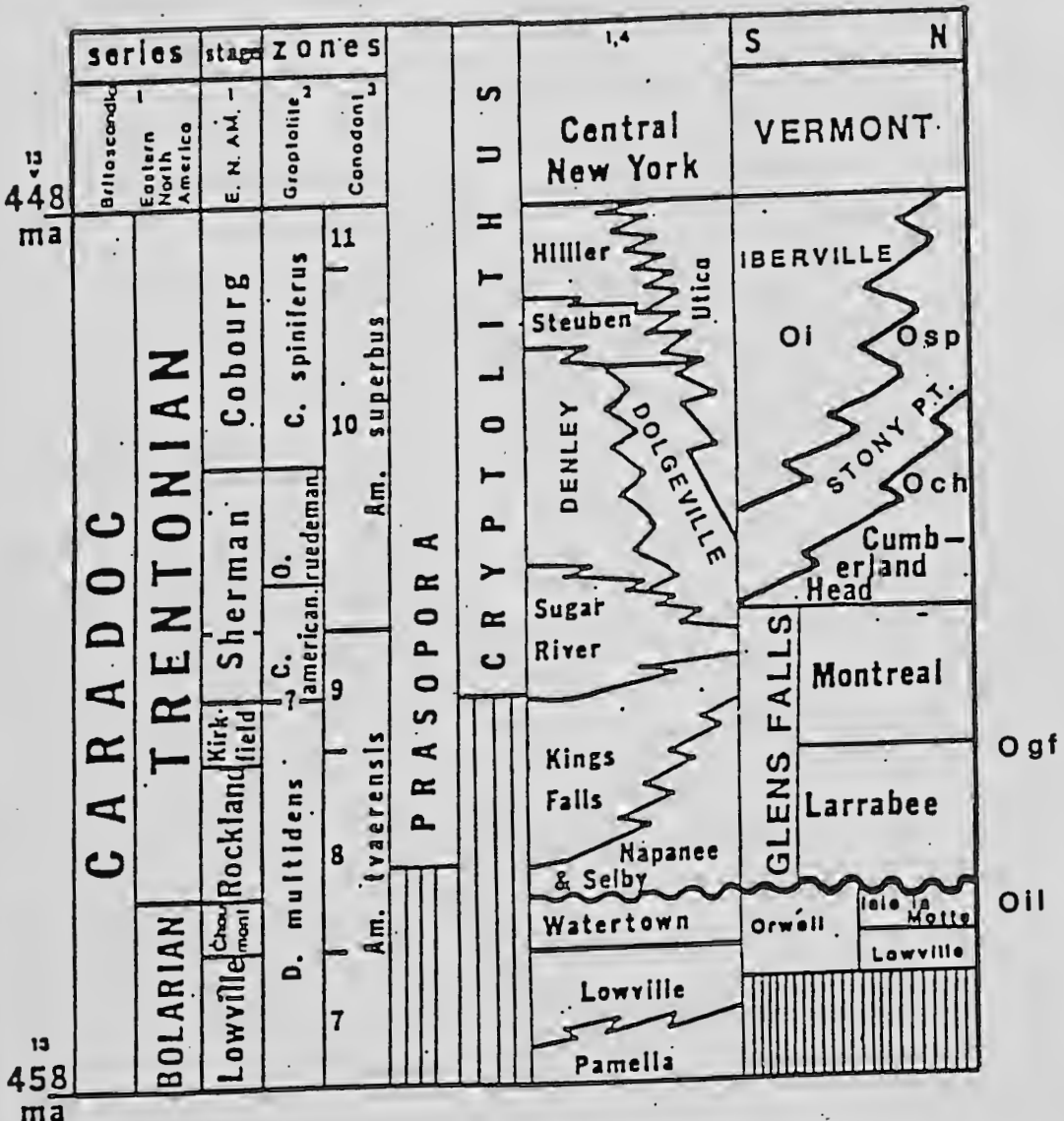
EXPLANATION

- A- Field Area C, Colchester
- B- West flank of Georgia Mt. Anticline (Southern)
- C- OCcs - O?Csk contact west of Cabbie Hill
- D- Southwestern stratigraphic column, Winooski area
- E- Route 2 section (Gregory, 1982)
- F- Northwestern stratigraphic column, West Georgia
- G- Milton Section: Arrowhead Mt. Anticline
- H- West flank of Georgia Mt. Anticline (Northern)

- Diagrammatic representation of the original depositional protolith for the Milton quadrangle reconstructed from stratigraphic information in the indicated areas.

Figure 1

Dorsey and others (1983)



MacLean 1986

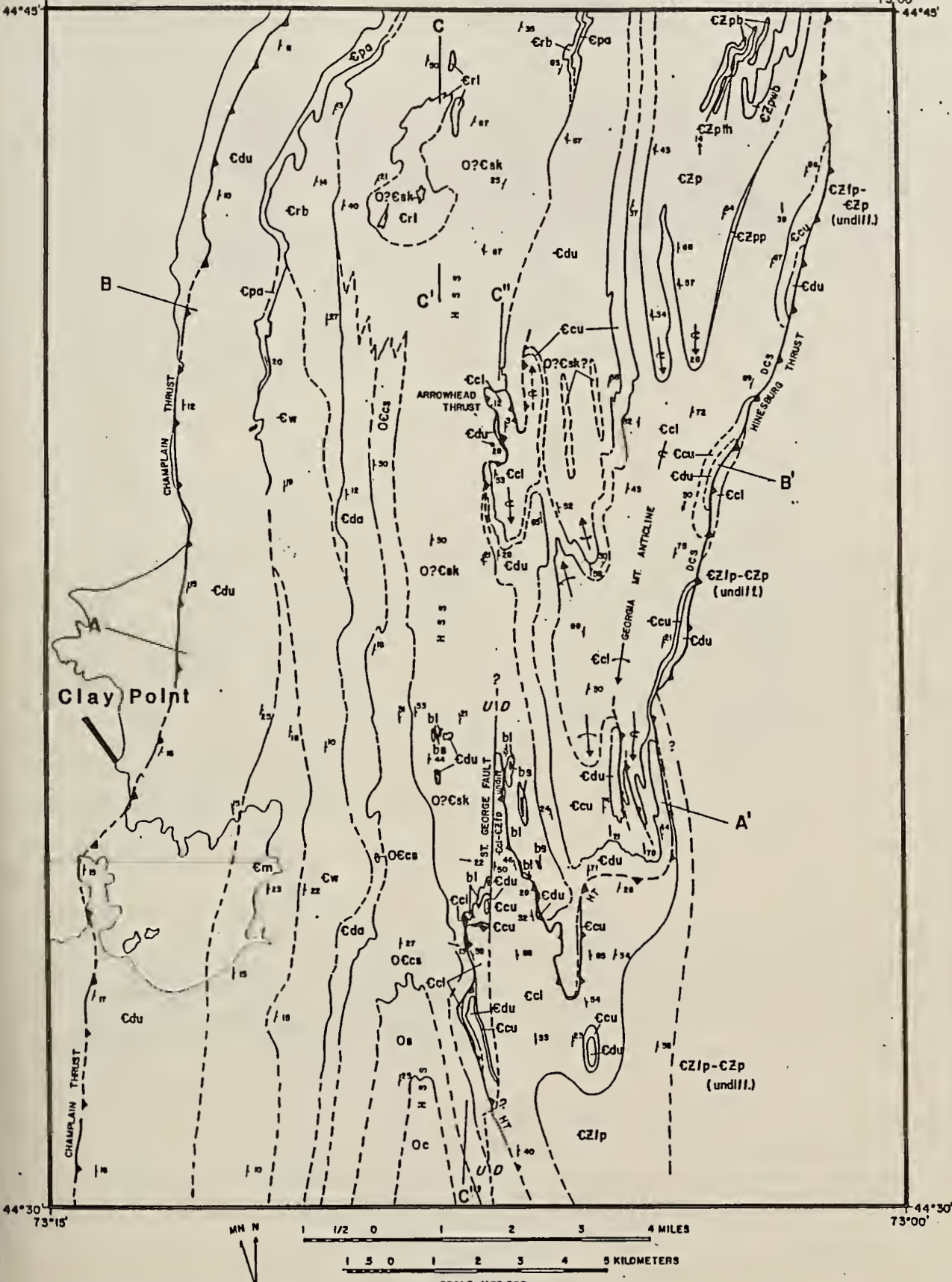
Figure 2 - Stratigraphic column for the western Vermont and Central New York. For further stratigraphic discussion read Erwin (1957) and Hawley (1957, 1972).

thrust fault. Westward displacement is estimated to be 6 km.

Recent seismic traverses across western Vermont demonstrate that the Champlain thrust dips eastward at approximately 15 degrees beneath the Green Mountain anticlinorium and that the major folds of western Vermont are formed by duplexes and related structure on this and other thrust faults which are present both within the platform and the foreland basin sequences. High angle normal faults have been mapped along the Champlain thrust fault and in many parts of the foreland basin sequence (Welby, 1961; Doll and others, 1961). Seismic information shows that some of the faults are older than the Champlain thrust fault whereas others are younger. Stanley (1980) has shown that several of the faults that cut the eastern part of the platform sequence are Mesozoic in age.

The structure of western Vermont in the Burlington region is best illustrated by the recent work of Dorsey and others (1983) in the Milton quadrangle (fig. 3) and Leonard (1985) in South Hero Island. The cross sections for this region show that the Champlain thrust fault is essentially planar which is consistent with recent seismic studies at this latitude. Detailed surface mapping, however, has shown that the overlying thrust faults are folded. Dorsey and others (1983) therefore suggest that these folds are related to duplexes in the Champlain slice (fig. 4). Based on this configuration the highest fault, the Hinesburg thrust, is the oldest and the Champlain thrust is the youngest. Thus the thrust sequence developed from the hinterland (east) to the foreland. This conclusion is consistent with the character of each of the fault zones. The Champlain thrust fault is marked by gouge, welded breccias and pressure solution features (Stanley, 1987). The Arrowhead Mountain thrust fault is marked by cataclasites, dislocation glide and creep features, pressure solution, and limited recrystallization. Oriented sericite is formed along cleavages (Strehle and Stanley, 1986). The Hinesburg thrust fault is marked by extensive recrystallization with the development of protomylonites and ultramylonites. Sericite, chlorite, and stilpnomelane are present (Strehle and Stanley, 1986). Thus the fabrics developed under progressive more ductile conditions generated at higher temperatures and pressures. Shortening as measured between the pin points in section B-B' (fig. 4) is in the order of 55 percent with 6 km of displacement on the Hinesburg thrust and 0.85 km of movement on the Arrowhead Mountain thrust. The Milton cross sections clearly show the change in structural style and fabric that occurs as one crosses from the foreland to the hinterland.

Most of the deformation in western Vermont occurred during the Taconian Orogeny. This conclusion is based on regional considerations (Stanley and Ratcliffe, 1985) and the analysis of available isotopic data (Sutter and others, 1985). Mesozoic deformation in the form of extensional faults and igneous activity clearly affected the region (McHone and Bulter, 1984; Stanley, 1980). Available evidence, however, can not rule out limited Acadian or even Alleghenian deformation.

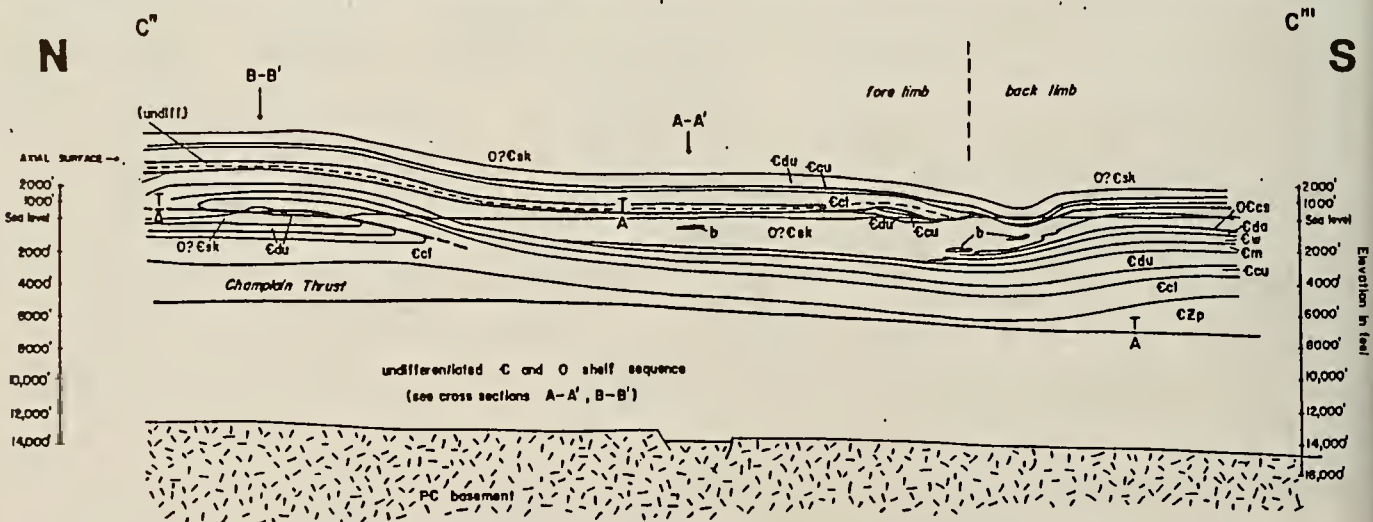
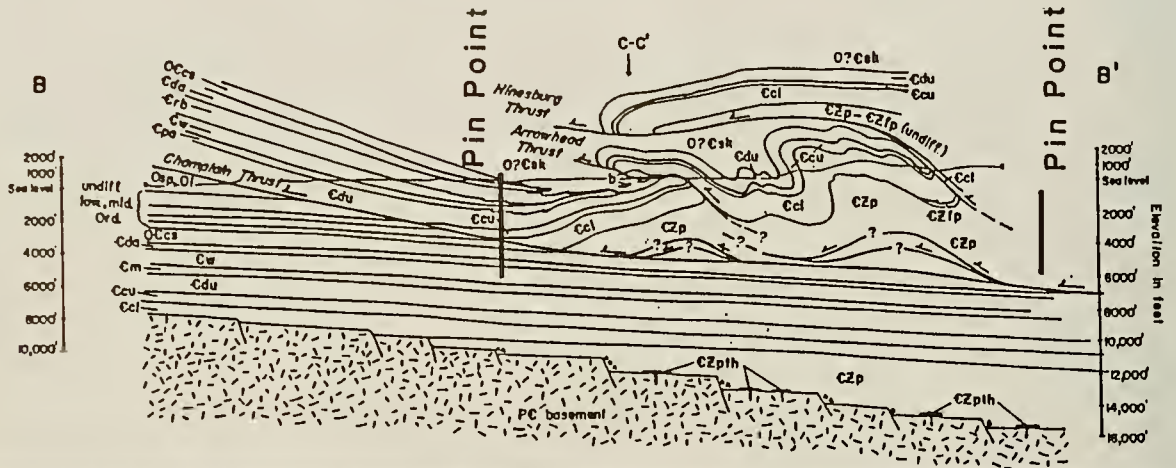
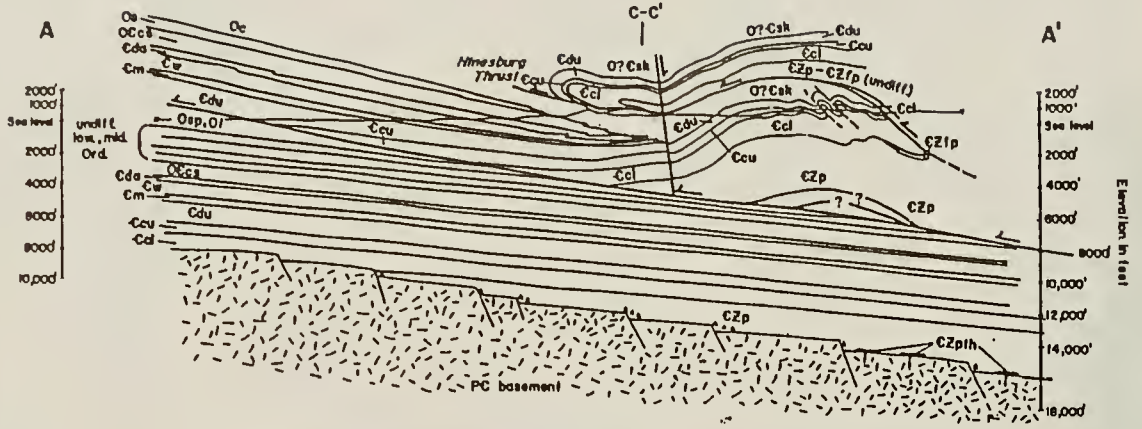


**BEDROCK GEOLOGY OF THE MILTON QUADRANGLE
WESTERN VERMONT**

Figure 3

Dorsey and others 1983

W



Dorsey and Stanley (1983)

CROSS SECTIONS OF THE MILTON QUADRANGLE WESTERN VERMONT

Figure 4

This field trip will begin in the western part of the foreland in South Hero Island and end along the western margin of the hinterland at the Hinesburg thrust fault at Mechanicsville, Vermont. The trip across central Vermont will continue across the pre-Silurian hinterland through the northern extension of the root zone for the Taconic allochthons Stanley and Ratcliffe, 1985). Participants will therefore be able to observe the change in structure and fabric from the foreland to the hinterland.

ITINERARY

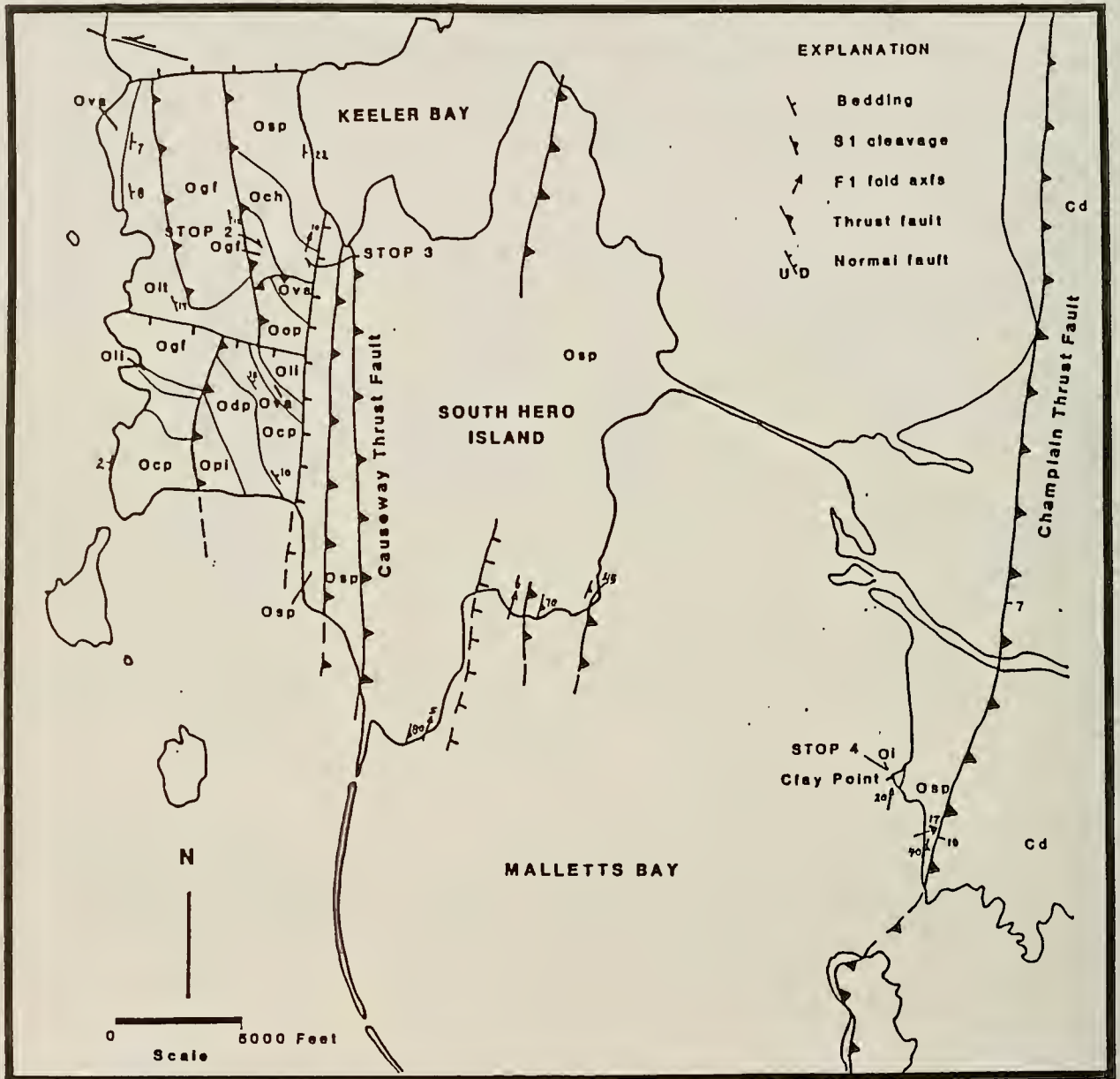
ALL STOPS ARE LOCATED ON GEOLOGIC OR TOPOGRAPHIC MAPS. A MILEAGE LOG IS NOT INCLUDED.

Assemble at the Apple Store on the south side of Route 2 in the village of South Hero at 8:30 AM. The first three stops will be on South Hero Island (fig. 5). PLEASE DO NOT USE HAMMERS ON THIS TRIP. LEAVE THEM IN THE CAR.

STOP 1 - West Shore of South Hero Island. This stop illustrates the low level of deformation that characterizes much of the west shore of the Champlain Islands and the eastern shore of New York in the area of Plattsburg. Local areas of moderate deformation, however, do exist where bedding plane faults and high-angle faults cut the bedrock. These outcrops of shale and thin micrite beds are in the Stony Point Shale. Note that the bedding, which is nearly horizontal, is cut by a poorly developed cleavage that is only developed in the shale beds. The cleavage dips very gently to the east and is interpreted to result from simple shear parallel to the bedding. Thin layers of calcite are present on some of the beds. Those layers that are marked by prominent slickenlines are bedding plane faults. As we will see at other outcrops today, the larger faults are marked by thicker layers of lineated calcite.

STOP 2 - Lessor's Quarry (fig. 6) - This quarry is located in the fossiliferous Glens Falls Limestone. The quarry contains some of the finest evidence of pressure solution in western Vermont. The cleavage (S1), which is discontinuous and wavy, is a classical pressure solution feature with well developed selvages that truncate fossils and offset bedding. A small anticline at the south edge of the quarry contains adjustment faults at its hinge that end along cleavage zones with thick clay selvages.

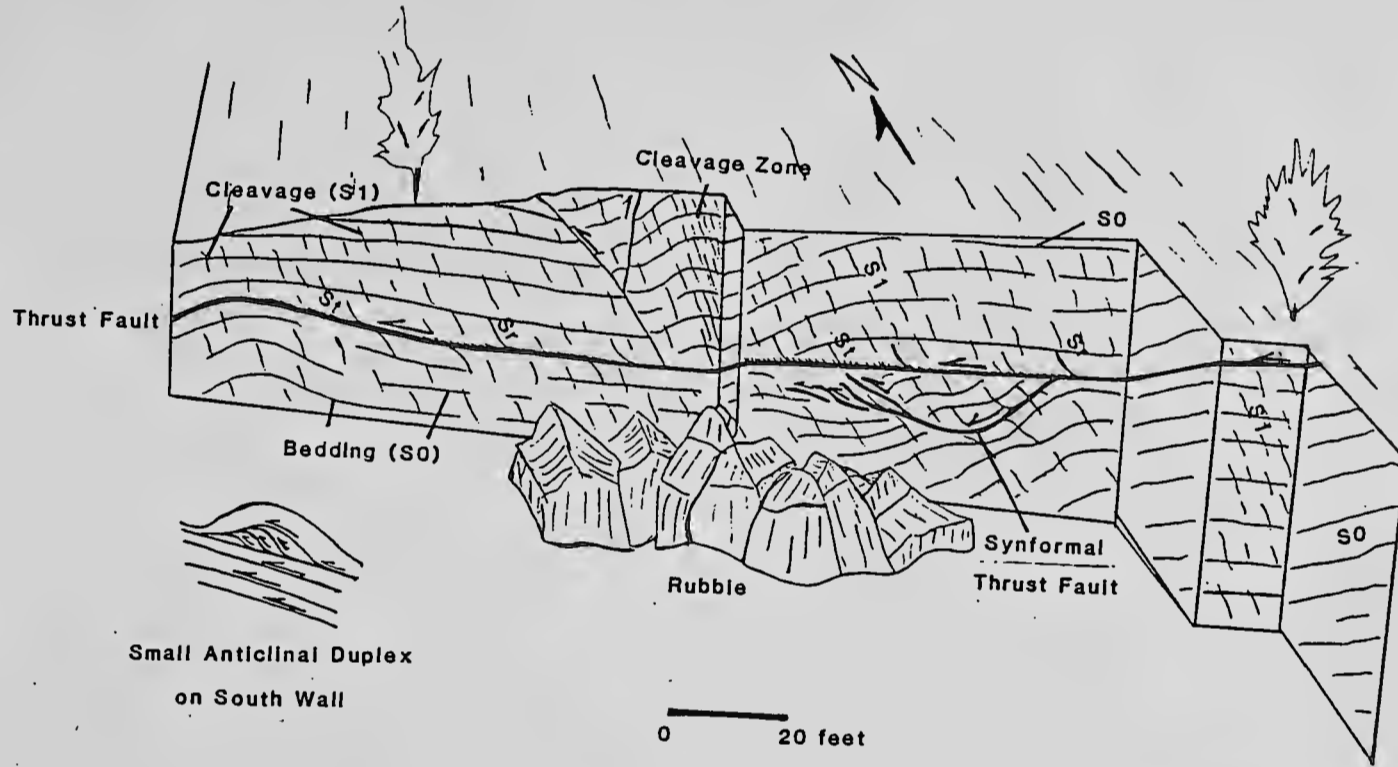
The major structures in the quarry are bedding-plane thrust faults. These faults are marked by calcite layers with west-trending slickenlines and a fault-zone cleavage (St). Near the larger faults the S1 cleavage is rotated (Sr) toward the plane of the fault. Note that both St and Sr dip gently to the east and indicate that movement on the bedding faults was to the west. The St cleavage forms as a result of simple shear on the faults. The anticline along the southwall and edge of the quarry is formed from a small duplex. Unfortunately, the best evidence for this duplex has been excavated.



Bedrock Geology of South Hero Island, Vermont

Leonard (1985), Erwin (1957)

Figure 5



Stanley (1987)

**THRUST FAULTS AND RELATED STRUCTURES
AT
LESSOR'S QUARRY, SOUTH HERO ISLAND, VERMONT**

Figure 6

On the northeast side of the quarry (fig. 6) a syncline and an associated blind, synformal thrust fault are truncated by the major thrust fault that is continuous across the north wall of the quarry. The origin of this structure is not clear, but it is thought to be associated with a duplex or ramp below the level of the quarry floor.

STOP 3 - "THE BEAM" - THIS IS A SUPERB OUTCROP THAT SERVES AS A FIELD LABORATORY FOR RESEARCH AND TEACHING OF FORELAND DEFORMATION. PLEASE STUDY IT. USE YOUR CAMERAS BUT NOT YOUR HAMMERS. REFER TO FIGURES 7 AND 8.

The outcrop is located in the Cumberland Head Formation approximately 5 miles west of the exposed front of the Champlain thrust fault or approximately 4600 feet below the restored westward projection of the thrust surface. The major questions that will be discussed are: 1. How do ramp faults form?, 2. Are there criteria to determine if imbricate thrust faults develop toward the foreland or hinterland?, 3. What is the relation between faulting and cleavage development?, 4. What processes are involved in the formation of fault zones?, 5. Are there criteria that indicate the relative importance and duration of motion along a fault zone?, 6. Is there evidence that abnormal pore pressure existed during faulting?, and finally 7. What is the structural evolution of the imbricate faults? The first six questions will be largely addressed by direct evidence at the outcrop. The last question will be answered by palinspastically restoring the imbricated and cleaved sequence to its undeformed state (fig. 8).

THE OUTCROP

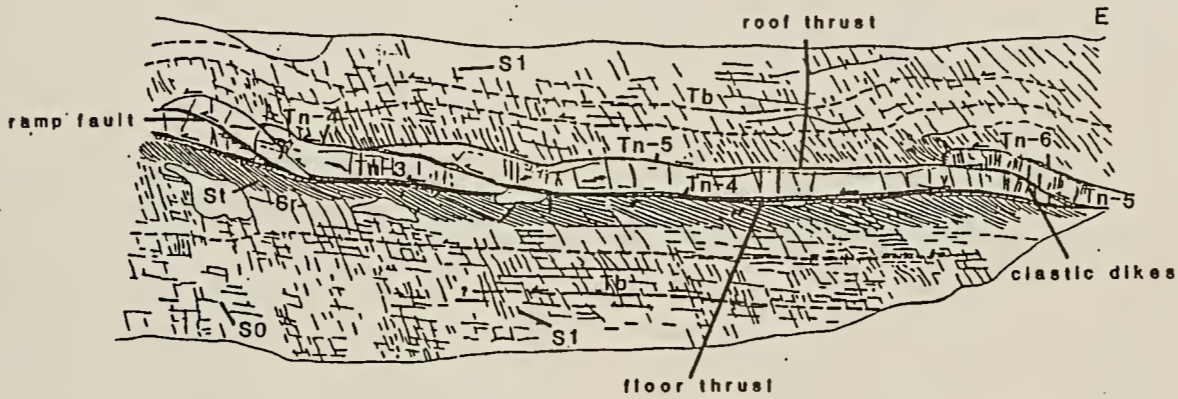
Five imbricate thrust faults and associated ramps are exposed in profile section in a foot thick bed of micrite that extends 45 feet along an azimuth of N 80 E (fig. 7a). The imbricate fault can be further classified as a central duplex of three horsts that are separated from two simple ramps at either end of the outcrop by approximately 8 feet of flats. The micrite bed is surrounded by at least 5 feet of well-cleaved calcareous shale. Bedding plane faults are present along the upper and lower surface of the micrite where they merge with ramp faults that cut across the micrite bed at an angle of approximately 30 degrees. The lower bedding plane fault or floor thrust is relatively planar and the fault zone is thick. In comparison the upper bedding plane faults are folded in the ramp areas, cut by ramp faults, and the fault zones are thin. The upper bedding plane fault forms the roof thrust for the central duplex. Along the intervening flats the upper faults are generally planar although they are cut by the S1 cleavage in many places. The lower bedding plane fault is the major decollement across the outcrop. Older bedding plane faults are also present throughout the shale and are offset by the penetrative S1 cleavage.

All the fault surfaces are covered by layers of sparry calcite that vary in thickness from several mm to 4-6 cm. The thickest zone is



EXPLANATION

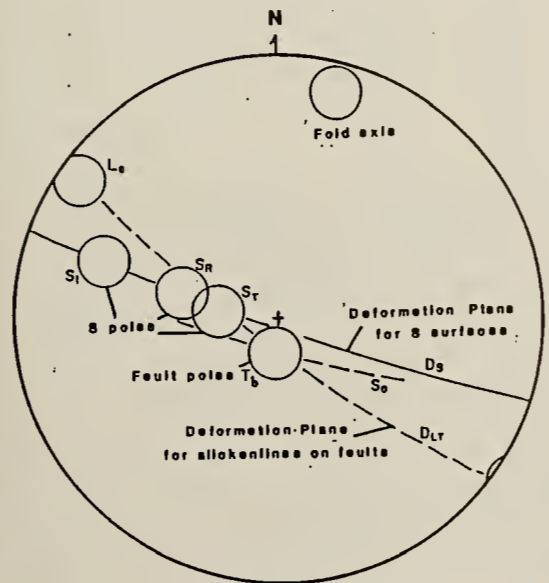
- S₀ Bedding
- S₁ Pressure solution cleavages
- S_r Thrust zone cleavages
- T_b Bedding plane thrust in shale
- Calcite veins & layers
- En echelon fractures
- R Rubble
- Thrust fault, showing relative motion



Stanley 1985

Figure 7a

DEFORMATION OF THE CUMBERLAND HEAD FORMATION
SOUTH HERO, VERMONT



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 7b

found along the decollement or floor thrust whereas the thinnest zone is found along the older bedding plane faults in the calcareous shale (fig. 7a). In all but the thinnest layers, the calcite is arranged in distinct layers that are separated by discontinuous selvage of dark gray shale. Each of the layers are marked by grooves or slickenlines that trends N 56 W and are essentially parallel from layer to layer (fig. 7b). In sections oriented perpendicular to the layering and parallel to the slickenlines the shale selvages are more parallel to each other than they are in sections cut perpendicular to the slickenlines where the selvages either anastomose or conform to the cross section of the grooves. In the parallel sections, however, some of the selvedge layers are truncated by more continuous surfaces. One of these can be traced for 5 feet or more along the decollement. At a number of places along the different fault zones small dikes of sparry calcite have intruded the lower part of the calcite-shale layers. The fault-zone fabrics are most clearly displayed along the decollement where the calcite-selvage layers are more abundance.

Fractures are common throughout the micrite bed where they are oriented at either a high angle or low angle to the bedding. Most fractures are filled with sparry calcite. The most prominent fractures are arranged in en echelon arrays that climb either to the west or the east. Most of these arrays are located in ramps areas and along the west-facing limbs or small flexures. A few are present in the flat regions of the bed. The fractures in many of the arrays in the ramp regions are folded and some are cut by younger generations of en echelon fractures. In the eastern ramp the hangingwall and footwall are cut by near-vertical fractures that are filled with fragments of the surrounding micrite embedded in sparry calcite so as to form clastic dikes.

Two, well developed, pressure-solution cleavages are present throughout the calcareous shale. The first and most conspicuous one, S1, strikes N 20 E and dips 60 to the east (figs 7b). This orientation is an average based on measurements taken across the vertical face of the outcrop because the individual S1 surfaces are quite wavy along strike. As a result they form a distinct diamond-shape pattern on the bedding surface. The acute angle of the diamond pattern is approximately 30 degrees in the shale and 50 degrees in the micrite bed. This geometry indicates a moderate level of cleavage development. S1 surfaces are covered by a black, carbon-rich selvedge of illite and kaolinite which is less than two tenth of an inch thick. Although many of the cleavage surfaces are vertically continuous through the shale, some of them are short and discontinuous with tapered ends. The thickest selvedge occurs on the most continuous surfaces. The surfaces of the selvedge are not lineated although some are polished. The S1 cleavage offsets bedding and the older bedding-plane thrust faults with a down-to-the-east sense throughout much of the outcrop. This displacement is greatest where a selvedge is the thickest and it is gradually reduced to zero as a selvedge thins toward the tapered ends of the shorter cleavage surfaces. The average width of the

microlithons between the S1 surfaces is 2.2 inches thick. The S1 cleavage not only cuts the older bedding-plane faults but cuts across most of the roof faults on top of the micrite bed.

A second, well developed cleavage, St, is restricted to a foot-thick zone directly below the decollement (fig. 7a). The individual cleavage surfaces are thinner (about a mm thick), more closely spaced (about a cm or more), and are covered with a thin selvedge (less than 1 mm). Furthermore, St dips to the east at only 6 degrees compared to the steeper dip of S1. St is also developed along the roof thrust of the micrite bed, but the zone is thinner and it is more difficult to recognize because the roof faults have been folded in the ramp areas and deformed by S1. The St cleavage is definitely related to movement of the thrust faults because it is only found near the faults and it is absent away from them.

In the zone near the decollement and the roof faults the S1 cleavage is rotated eastward so that the steeper 60 degree dip in the shale is reduced to 25 degrees (fig. 7b.). The strike of the rotated S1 (hereafter referred to as Sr) is the same as S1 away from the faults. The counterclockwise rotation of S1 through an angle of shear of 35 degrees indicates that movement on the decollement was east-over-west along a direction of N 56 W as indicated by the slickenlines along the decollement. This sense of displacement is consistent with the orientation of St since the normal to St would correspond to the direction of maximum finite compressive strain.

Cleavage, similar to S1 and St, is totally absent from the micrite bed. In a few places, however, a very thin (less 1 mm), stylolitic to very irregular cleavage is oriented perpendicular to bedding in the micrite. This cleavage is not developed uniformly throughout the micrite. Where it is formed the cleavage surfaces are separated from each other by at least 6-10 cm. The form, orientation and limited distribution of this cleavage indicates that it formed very early in the deformation sequence while compression was essentially parallel to the planar micrite bed.

OTHER IMPORTANT RELATIONS

Ramp Faults

All the east-dipping ramp faults form from arrays of west-climbing, en echelon fractures. These arrays appear to nucleate along the west-facing limbs of slightly asymmetrical buckle folds. The sequence continues with the rotation of the individual fractures by distributed shear strain along the west-climbing array. The individual fracture may continue to grow as the older central parts of the fractures are rotated counterclockwise to produce S shaped fractures. New arrays of planar fractures develop over the older arrays and produce a weakened zone along the trend of the array. As generations of fractures are superposed they coalesce and develop into ramp faults as demonstrated by deformed fracture arrays in the footwall and hangwall that are truncated by the ramp

faults. The presence of sparry calcite and cavities in the fractures indicates that the fractures opened rapidly and at a shallow enough level in the crust so that the micrite bed was strong enough to support the shape of the cavities. East-climbing, en echelon fracture arrays are also present in the micrite where they are abundant in the ramp regions.

Fault Chronology

The bedding plane faults in the surrounding shales are clearly older than these faults because they are cut and offset by the S1 cleavage. Although the bedding plane faults that are in direct contact with the micrite bed may have formed originally at this time there is abundant evidence that they certainly were active long after those in the shale. Their average age therefore is younger.

The evidence for the relative age of each of the faults is found in the regions where the ramp faults merge with the roof and floor faults. For example, in the western part of figure 7a the ramp fault (Tn-1) cuts across the upper-bedding plane fault (Tn-2) of the hangingwall block but merges asymptotically with the floor fault (Tn) of the footwall block. Furthermore, the roof fault of the hangingwall block is folded and cut by S1 cleavage. At the junction of the ramp fault and the floor fault, the quasiplanar slip surfaces in the calcite-shale fault zone cut across the fault zone of the ramp fault. The relative age relations therefore are clear - the oldest fault of the three is the roof thrust and is designated Tn-2. The youngest fault is the floor thrust (Tn). This same relative chronology applies to all the ramps and duplexes to the east.

Several important conclusions result from this analysis. First, the imbricate faults which comprise the roof, ramp, and floor fault system become younger to the west or the foreland. Second, the floor fault is continually reactivated during the evolution of the fault system. The average age of this fault zone therefore becomes younger to the west. Furthermore, it is the most continuous fault in the outcrop. This second conclusion explains why the fault zone for the floor fault is much thicker than the fault zones along the ramp or roof faults. The floor fault has been active for a longer period of time than any of the other faults. The thickness and continuity of a fault zone therefore are directly related to the duration of motion of a given fault.

Fault-Zone Deposits

The important facts that bare on the evolution of the fault-zones are the following:

- 1) All the faults zones are filled with veins of sparry calcite and minor quartz which are generally oriented either parallel or at a low angle to the fault surface. Vertical veins, which commonly

join sills higher in the fault zone, are more common in the lowest layers of the fault zone and in the shale directly beneath the fault zone.

2) Discontinuous, dark, stylitic clay laminae with concentrations of quartz adjacent to or within the laminae are interlayered with the calcite in all but thinnest zones. The laminae are identical in appearance to the selvages on the S1 cleavage surfaces. Some of the clay laminae are continuous with chips of shale which are either completely enclosed within a calcite layer or occur at the boundary between two calcite veins. These shale chips preserve varying degrees of pressure solution. For example, some of the chips are similar to the undeformed shale in microlithons away from fault zones whereas others have dark, thin selvages within the chip and along their edges.

3) A relative planar surface decorated with shale laminae cuts older surfaces in the fault-zone deposit and is continuous for 5 or 10 feet along the floor fault. Such surfaces as this are called slip surfaces.

4) The size of the sparry calcite is directly proportional to vein width. Most grains are bladed in form, but their long axes is not preferentially aligned. The calcite in all the layers is twinned with the greatest density occurring in the thinner layers between shale laminae where the grains are turbid and small. The larger grains in thicker layers nearest the planar slip surface are generally more twinned than are those grains in veins further away.

5) The slickenlines on each of the vein layers are formed by grooves rather than calcite fibers.

6) The calcite-shale layers tend to be more parallel in sections cut parallel to the slickenlines rather than they are in sections cut perpendicular to the slickenlines where the layers are irregular or anastomose.

It is clear from the foregoing information that the calcite has been intruded along the faults after the initial cohesion had been broken along the shale-shale or shale-micrite contacts where the strength contrast is the greatest and the cohesive strength the weakest. Once such a zone has developed and is filled with calcite, it becomes a zone of weakness. Subsequent failure likely occurred along the calcite-shale interface and resulted in scabs and chips of the shale being incorporated into the fault zone. Other shale fragments may have been sheared in along faults or carried in along thick veins. During renewed movement the calcite and quartz were dissolved from the shale to form the black selvages which are interlayered with the calcite and decorate the slip surfaces. These boundaries then formed weak planes along which subsequent movement occurred within the fault zones. As the fault zone thickened movement could occur along planar surfaces (slip surfaces) which smoothed out the irregular geometry formed by

ramp zones and broad folds in the intervening flat regions of the micrite bed. Movement was no longer restricted to the calcite-shale boundaries of thin fault zones, but could occur along any favorably situated calcite-selvedge boundary. The resulting clay selvedge then acted as a catalyst that facilitated solution of calcite from the selvedge-calcite boundary of the surrounding veins. This process resulted in the stylolitic form of the selvedge. We suggest that preferential solution in the direction of fault movement produced the slickenline grooves and the stylolitic selvedges best seen in sections cut perpendicular to the slickenlines.

Because the floor fault was continually active during the evolution of the imbricate system, It is not surprising to find evidence for repeated vein injection in the form of numerous crosscutting veins in the thick fault zone deposit. During each of these events the influx of fluid and the subsequent crystallization was relatively rapid so that sparry calcite formed rather than fibered calcite. As the fault zone thickened with layers of calcite and clay selvedge, new veins could form along any surface of weakness within the fault zone rather than being confined to the outer borders with the country rock. As movement continued across the fault zone, the calcite in the older veins became heavily twinned and severely strained. Repeated solution of calcite along their boundaries with the adjacent clay selvedges reduced their thickness and produced the common observation that calcite in many of the thinner veins are heavily twinned.

The senario that has been inferred from the fault zone fabrics and the relative age relations among the faults suggests that fault movement was intermittent with each event occuring rapidly. During the intervening time deformation in the fault zone may have been restricted to twinning in the calcite.

Shortening

The shortening across the outcross is conveniently recorded by the folds in the ramps areas and structural overlap across the faults. The displacement on each of the faults in the micrite bed ranges from 3 to 19 inches which adds up to a total displacement of 52.8 inches or 4.4 feet over a present horizontal distance of 35 feet. The five anticlines over ramps and the broad folds along the flats account for approximately 5 inches of additional shortening so that the total shortening equal 57.8 inches or 4.8 feet. These values correspond to a shortening of 13.7 percent.

In the shale the corresponding shortening is provided by volume reduction across the cleavage surfaces. In order to see if the shortening determined from the micrite bed is comparable to the shortening in the shale, an independent estimate was made for the shale by determining the percentage of insoluble material. Samples of suitable material from four different microlithons where immersed in hydrochloric acid until all the soluble material was eliminated. The final average residue was 36 percent of the

original mass (a range of 32% to 39% for 4 samples). The number of cleavage selvages across a present width of 26 feet was then counted. The total width of the selvages (a range of 1.16 to 1.78 foot) was then multiplied by 2.8 to give an estimate of the original width now represented by the cleavages (3.25 to 4.98 feet). The original length of the present 26 foot width was then estimated to be 29.3 to 31 feet. Shortening was then calculated to be in the range of 7.3 to 10.6 percent.

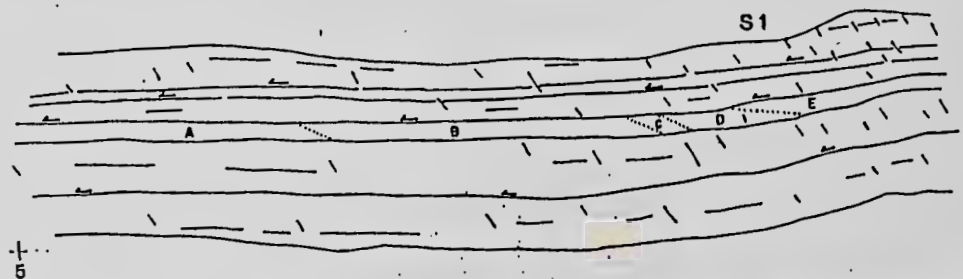
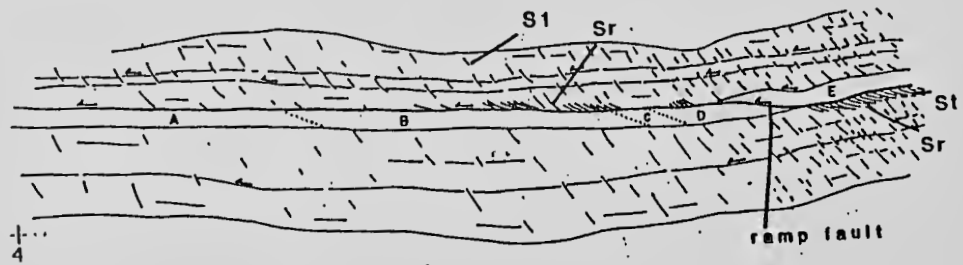
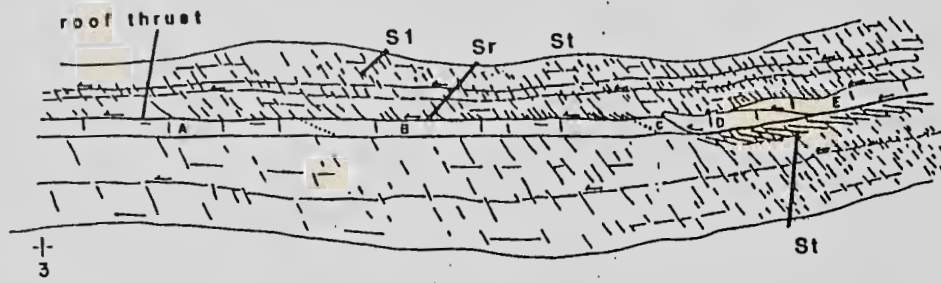
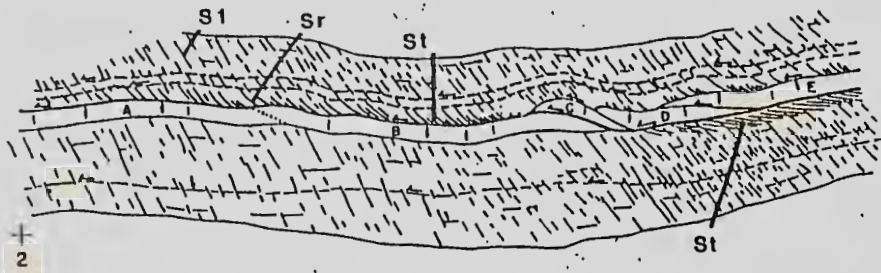
This difference in shortening values in the "beam" and the shale is not considered to be significant because the method for estimating shortening in the shale is less accurate than the method for the micrite bed. For example, the number of cleavage surfaces in the shale and their thickness were underestimated since many thin cleavage selvages were overlooked in the original count. Furthermore, if the lowest value of 32 percent were used along with all the existing data the total shortening would be a little over 14%. Thus the range of values overlaps the shortening value calculated for the micrite bed.

We therefore conclude that the formation of the cleavage and consequent shortening in the shale occurred during imbricate faulting in the micrite bed. Although this conclusion may seem intuitively obvious, it does have important implications for the evolution of the cleavage. Because we have already proven that the faults represent a time-transgressive sequence that developed from east to west, we must also conclude that the cleavage in the surrounding shales must have developed in a similar manner. Unlike our earlier conclusion, this relation is far from obvious from the relations within the cleaved shale. In fact it is the existence of the micrite bed and its fault geometry that allows us to conclude that the cleavage in the shale is indeed a time transgressive phenomena. The evolution of the cleavage and its relation to imbricate faulting is best shown by retrodeforming the faulted micrite bed to its original predeformational condition (fig. 8).

The next problem is the origin of Sr, the rotated cleavage, and St the finely spaced cleavage below the floor thrust. Because Sr is simply the dominant S1 cleavage throughout the shale and is only present near the floor and roof faults, it had to form after the ramp faults developed and during subsequent displacement on the roof and floor faults. During this time the S1 cleavage near these faults is rotated in simple shear in the direction of fault displacement. The fact that the Sr cleavage below the floor fault is rotated more than it is along the roof fault is consistent with our earlier conclusion that the floor thrust was active throughout deformation whereas the individual roof faults are short lived.

The St cleavage, which is only present along the floor thrust, clearly formed after Sr because it cuts across Sr at a low angle and is not folded or rotated. Its absence along the roof faults is consistent with their short history of displacement. St is a true fault zone cleavage because it is restricted to a thin region below

A-5



RETRODEFORMED SECTION OF THE CUMBERLAND HEAD FORMATION

SOUTH HERO, VERMONT

Stanley, 1987

Figure 8

the floor fault. Furthermore, the St cleavage is in the orientation predicted by the Ramsay equation $\tan 2\theta = 2/\gamma$ (Ramsay and Huber, 1984) where γ of 36 degrees is the rotation of S1 (Sr) as it is traced into the fault zone. This relation proves that St is the result of simple shear along the floor fault.

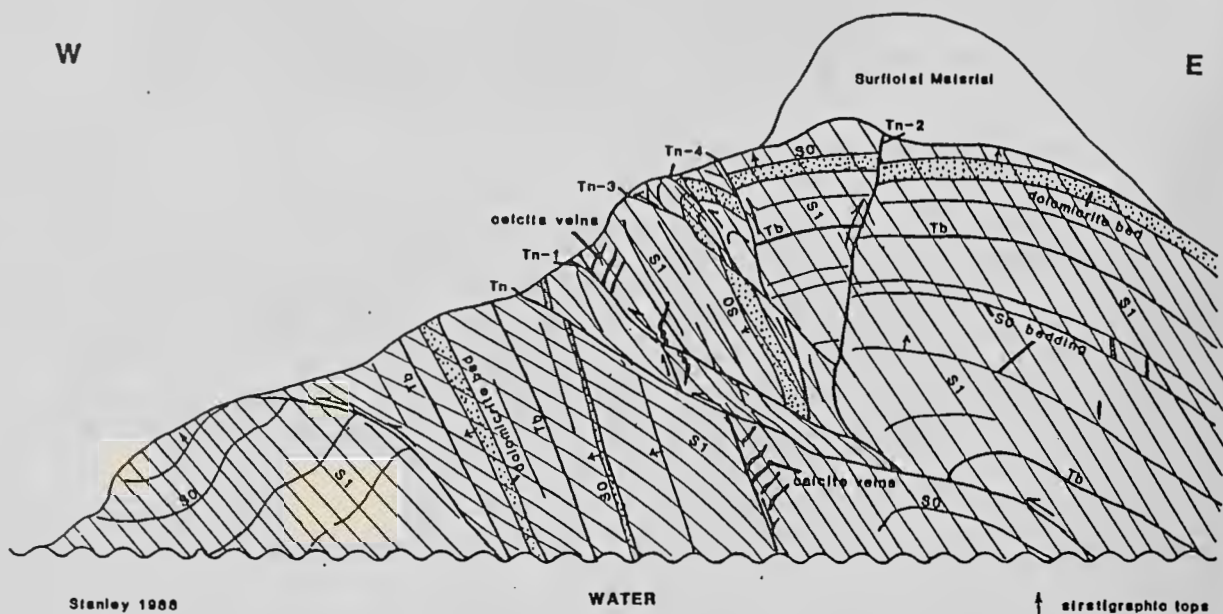
Are St and Sr time transgressive? Because we have demonstrated earlier that the floor fault and the respective roof faults are time transgressive, it must follow that both Sr and St are also time transgressive to the west. This conclusion suggests that the amount of rotation of Sr and the intensity of St should also increase to the east where the displacement on the floor fault has been longer. We could detect no such relation which, in turn, may suggest that there is a limit beyond which Sr can be rotated.

EVOLUTION OF STRUCTURES

The evolution of the imbricate faults and the various cleavages described in the foregoing section is illustrated in a series of retrodeformed sections in figure 8. Section 1 shows the "beam" in its present state. Section 2 is developed by reversing the deformation associated with the youngest ramp fault at the western part of the outcrop. For example, the rocks of the hanging wall block (B, section 2) are unfolded as they are returned to their original position east of the footwall block (A, section 2). During the time represented by section 2 the active floor fault in the eastern part of the diagram climbs section along the ramp fault below block C and continues along the top of block A and B. As a result the St and Sr cleavages below blocks A and B are absent. Furthermore, the S1 cleavage is shown to be more abundant in the upper plate than the lower plate because it is actively moving.

Sections 3, 4 and 5 show the retrodeformation continuing to the east and are constructed in the same manner as section 2. Thus the evolution of the imbricate system and its associated structures can be seen by studying diagrams 5 through 1.

STOP 4 - CLAY POINT (fig. 9) - Clay Point is located several 1000 feet west of the trace of the Champlain thrust fault (fig. 3 and 5). The rocks consists of medium to dark gray, noncalcareous shales, dolomitic siltstones and beds of brown-weathered dolomicrite of the Iberville Formation, a foreland basin flysch deposit. This sequence is rhythmically layered with the base of each cycle marked by yellowish-brown weathered, dark gray laminated siltstone which contains ripple laminations (Hawley, 1972). These beds grade upward into dark grey, well cleaved shale that are marked by thin layers of lineated calcite. These layers represent bedding plane faults (Tb, fig. 9a). which are identical to the bedding plane faults at the "beam". Like the bedding, these faults are folded into a west-facing, overturned anticline that is cut by a prominent, east-dipping, pressure solution cleavage (S1, fig. 9b). This cleavage has the same orientation as the S1 cleavage at the "beam" (STOP 3) and is part of the same generation although it



DEFORMED IVERVILLE FORMATION AT CLAY POINT, VERMONT

Figure 9a

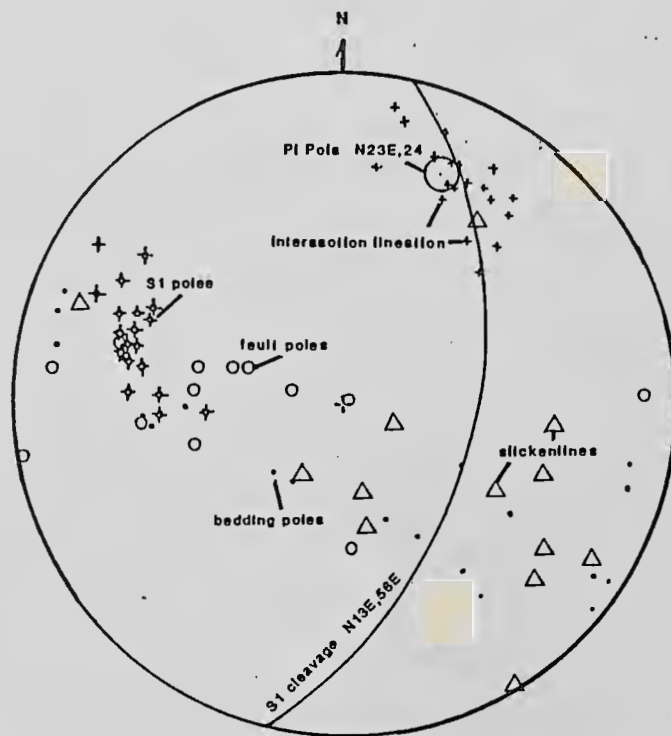


Figure 9b - Lower hemisphere equal area projection showing the orientation of structural elements at Clay Point.

formed earlier because it is situated farther to the east and closer to the Champlain thrust fault. The overturned limb of the anticline is cut by a series of thrust faults (Tn through Tn-4) that are clearly influenced more by the cleavage than they are by the bedding. The sense of displacement on these faults is provided by the rotated S1 cleavage adjacent to the fault surface. We believe that faults of this type that are controlled by the cleavage are younger than the faults seen at the "beam". This relation can be seen at the east end of the "beam" where a younger fault has developed from the older ramp and is climbing through the upper plate shale.

The other features that you should study are:

1) Calcite fractures (fig 9a). Many of these fractures contain cross fibers and indicate slow extension parallel to the fiber direction. Several of the veins are compound with an outer border of fiber calcite and an inner core of sparry calcite, which indicates rapid extension. These veins therefore opened more rapidly after an earlier period of slow extension.

2) Several of the faults on western side of the outcrop have two directions of slickenlines. Although this fact suggests a change in fault motion during the evolution of the structure at Clay Point, another explanation is suggested by the fact that many of these faults are bedding plane faults (Tb). During folding the older northwesterly slickenline direction was rotated out of the plane containing the transport direction (deformation plane). Subsequent movement along the same northwesterly direction produced the new slickenlines which cut across the older direction.

3) A west-trending Mesozoic dike cut the anticline just north of the cross section.

4) Study the laminate calcite along the faults. Look at the fabric perpendicular and parallel to the slickenline direction.

STOP 5 - THE CHAMPLAIN THRUST FAULT AT LONE ROCK POINT, BURLINGTON, VERMONT - The following discussion is reprinted from The Centennial Field Guide, Volume 5, of the Geological Society of America in 1986. All the figure numbers for this stop refer to those figures in the reprint. The reprinted discussion appears in Appendix 1.

STOP 6 - THE HINESBURG THRUST FAULT AT HINESBURG, VERMONT - This is the classic and best exposed locality for the Hinesburg thrust fault. It contains many fault related fabrics that have recently been studied by Strehle (1985) and published by Strehle and Stanley (1986) in a bulletin of the Vermont Geological Survey (Studies in Vermont Geology No.3). This publication also contains analysis of other fault zones of western Vermont which will be seen during this NEIGC. The reader is referred to this paper or an earlier NEIGC trip by Gillespie and others (1972).

The Hinesburg thrust fault separates the Cambrian-Ordovician rocks

of the platform sequence from the older, highly deformed metamorphic rocks of the eastern hinterland. As shown in figure 4, the Hinesburg thrust fault developed along the overturned, sheared limb of a large recumbent fold. This fault probably broke out from the overturned limb of a fault-propagation fold (Suppe, 1985) and therefore is similar in origin to the Arrowhead Mountain thrust fault. To the south the Hinesburg thrust fault dies out somewhere in the overturned limb of the Lincoln massif (Tauvers, 1982; DiPietro, 1983; DelloRusso and Stanley, 1986).

At the Mechanicsville locality the lower 40 m. of the Cheshire Quartzite is structurally overturned along the base of the upper plate of the Hinesburg thrust fault. Higher up the cliff the quartzite grades into the Fairfield Pond Formation of Tauvers (1982). The lower plate rocks, which are poorly exposed, consist of carbonates of the Lower Ordovician Bascon Formation. Slivers of dark gray phyllite of the Brownell Mountain Phyllite are found at several localities along the fault trace. Chlorite, muscovite, and stilpnomelane are present in the quartzite. Muscovite and chlorite are present in the schist.

The following features should be studied here:

- 1) The change in fabric as the fault surface is approached. The quartzite grades from a protomylonite to an ultramylonite along the fault surface.
- 2) The presence of east-over-west asymmetrical folds. These folds are related to simple shear along the fault.
- 3) The prominent mineral lineation consisting of elongate quartz and quartz clusters.
- 4) "Z" shaped quartz veins that are associated with beds of quartzite.
- 5) Rare east-dipping shear bands.
- 6) Late fractures and associated en echelon fracture arrays.

The interpretation of these structures and the thin sections fabrics are discussed in Strehle and Stanley (1986).

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APPENDIX 1 - LONG ROCK POINT - REPRINT.

The Champlain thrust fault, Lone Rock Point, Burlington, Vermont

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LOCATION

The 0.6 mi (1 km) exposure of the Champlain thrust fault is located on the eastern shore of Lake Champlain at the north end of Burlington Harbor. The property is owned by the Episcopal Diocesan Center. Drive several miles (km) north along North Avenue (Vermont 127) from the center of Burlington until you reach the traffic light at Institute Road, which leads to Burlington High School, The Episcopal Diocesan Center, and North Beach. Turn west toward the lake and take the first right (north) beyond Burlington High School. The road is marked by a stone archway. Stop at the second building on the west side of the road, which is the Administration Building (low rectangular building), for written permission to visit the field site.

Continue north from the Administration Building, cross the bridge over the old railroad bed, and keep to the left as you drive over a small rise beyond the bridge. Go to the end of this lower road. Park your vehicle so that it does not interfere with the people living at the end of the road (Fig. 1). Walk west from the parking area to the iron fence at the edge of the cliff past the outdoor altar where you will see a fine view of Lake Champlain and the Adirondack Mountains. From here walk south along a footpath for about 600 ft (200 m) until you reach a depression in the cliff that leads to the shore (Fig. 1).

SIGNIFICANCE

This locality is one of the finest exposures of a thrust fault in the Appalachians because it shows many of the fault zone features characteristic of thrust faults throughout the world. Early studies considered the fault to be an unconformity between the strongly-tilted Ordovician shales of the "Hudson River Group" and the overlying, gently-inclined dolostones and sandstones of the "Red Sandrock Formation" (Dunham, Monkton, and Winoski formations of Cady, 1945), which was thought to be Silurian because it was lithically similar to the Medina Sandstone of New York. Between 1847 and 1861, fossils of pre-Medina age were found in the "Red Sandrock Formation" and its equivalent "Quebec Group" in Canada. Based on this information, Hitchcock and others (1861, p. 340) concluded that the contact was a major fault of regional extent. We now know that it is one of several very important faults that floor major slices of Middle Proterozoic continental crust exposed in western New England.

Our current understanding of the Champlain thrust fault and its associated faults (Champlain thrust zone) is primarily the result of field studies by Keith (1923, 1932), Clark (1934), Cady (1945), Welby (1961), Doll and others (1961), Coney and others (1972), Stanley and Sarkisian (1972), Dorsey and others (1983), and Leonard (1985). Recent seismic reflection studies by Ando and others (1983, 1984) and private industry have shown that the Champlain thrust fault dips eastward beneath the metamorphosed rocks of the Green Mountains. This geometry agrees with earlier interpretations shown in cross sections across central and

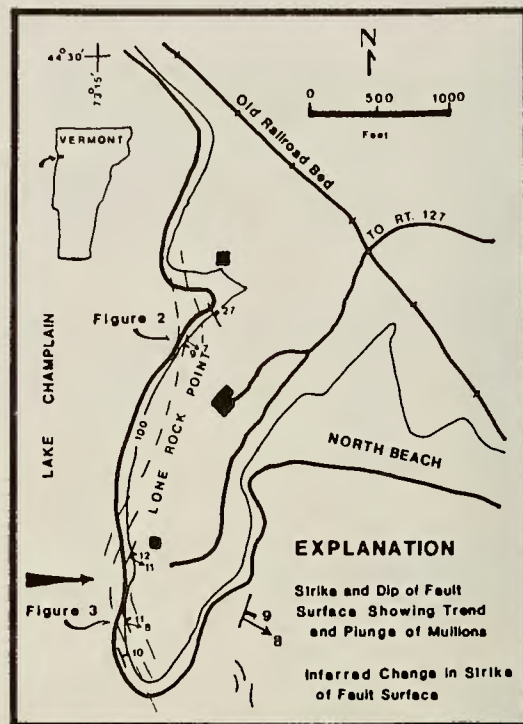


Figure 1. Location map of the Champlain thrust fault at Lone Rock Point, Burlington, Vermont. The buildings belong to the Episcopal Diocesan Center. The road leads to Institute Road and Vermont 127 (North Avenue). The inferred change in orientation of the fault surface is based on measured orientations shown by the dip and strike symbols. The large eastward-directed arrow marks the axis of a broad, late syncline in the fault zone. The location of Figures 2 and 3 are shown to the left of "Lone Rock Point." The large arrow points to the depression referred to in the text.

northern Vermont (Doll and others, 1961; Coney and others, 1972). Leonard's work has shown that the earliest folds and faults in the Ordovician sequence to the west in the Champlain Islands are genetically related to the development of the Champlain thrust fault.

In southern Vermont and eastern New York, Rowley and others (1979), Bosworth (1980), Bosworth and Vollmer (1981), and Bosworth and Rowley (1984), have recognized a zone of late post-cleavage faults (Taconic Frontal Thrust of Bosworth and Rowley, 1984) along the western side of the Taconic Mountains. Rowley (1983), Stanley and Ratcliffe (1983, 1985), and Ratcliffe (in Zen and others, 1983) have correlated this zone with the Champlain thrust fault. If this correlation is correct then the Champlain thrust zone would extend from Rosenberg, Canada, to the Catskill Plateau in east-central New York, a distance of 199 mi (320 km), where it appears to be overlain by Silurian and Devonian rocks. The COCORP line through southern Vermont

shows an east-dipping reflection that roots within Middle Proterozoic rocks of the Green Mountains and intersects the earth's surface along the western side of the Taconic Mountains (Ando and others, 1983, 1984).

The relations described in the foregoing paragraphs suggest that the Champlain thrust fault developed during the later part of the Taconian orogeny of Middle to Late Ordovician age. Subsequent movement, however, during the middle Paleozoic Acadian orogeny and the late Paleozoic Alleghenian orogeny can not be ruled out. The importance of the Champlain thrust in the plate tectonic evolution of western New England has been discussed by Stanley and Ratcliffe (1983, 1985). Earlier discussions can be found in Cady (1969), Rodgers (1970), and Zen (1972).

REGIONAL GEOLOGY

In Vermont the Champlain thrust fault places Lower Cambrian rocks on highly-deformed Middle Ordovician shale. North of Burlington the thrust surface is confined to the lower part of the Dunham Dolomite. At Burlington, the thrust surface cuts upward through 2,275 ft (700 m) of the Dunham into the thick-bedded quartzites and dolostones in the very lower part of the Monkton Quartzite. Throughout its extent, the thrust fault is located within the lowest, thick dolostone of the carbonate-siliciclastic platform sequence that was deposited upon Late Proterozoic rift-clastic rocks and Middle Proterozoic, continental crust of ancient North America.

At Lone Rock Point in Burlington the stratigraphic throw is about 8,850 ft (2,700 m), which represents the thickness of rock cut by the thrust surface. To the north the throw decreases as the thrust surface is lost in the shale terrain north of Rosenberg, Canada. Part, if not all, of this displacement is taken up by the Highgate Springs and Philipsburg thrust faults that continue northward and become the "Logan's Line" thrust of Cady (1969). South of Burlington the stratigraphic throw is in the order of 6,000 ft (1,800 m). As the throw decreases on the Champlain thrust fault in central Vermont the displacement is again taken up by movement on the Orwell, Shoreham, and Pinnacle thrust faults.

Younger open folds and arches that deform the Champlain slice may be due either duplexes or ramps along or beneath the Champlain thrust fault. To the west, numerous thrust faults are exposed in the Ordovician section along the shores of Lake Champlain (Hawley, 1957; Fisher, 1968; Leonard, 1985). One of these broad folds is exposed along the north part of Lone Rock Point (Fig. 2). Based on seismic reflection studies in Vermont, duplex formation as described by Suppe (1982) and Boyer and Elliot (1982) indeed appears to be the mechanism by which major folds have developed in the Champlain slice.

North of Burlington the trace of the Champlain thrust fault is relatively straight and the surface strikes north and dips at about 15° to the east. South of Burlington the trace is irregular because the thrust has been more deformed by high-angle faults and broad folds. Slivers of dolostone (Lower Cambrian Dunham Dolomite) and limestone (Lower Ordovician Beekmantown Group) can be found all along the trace of the thrust. The limestone represents

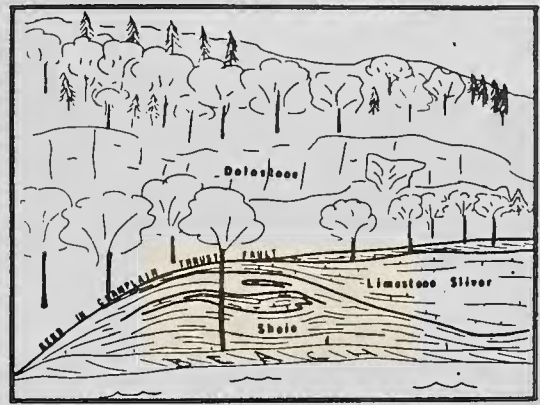


Figure 2. A sketch of the Champlain thrust fault at the north end of Lone Rock Point showing the large bend in the fault zone and the slivers of Lower Ordovician limestone. The layering in the shale is the S1 cleavage. It is folded by small folds and cut by many generations of calcite veins and faults. The sketch is located in Figure 1.

fragments from the Highgate Springs slice exposed directly west and beneath the Champlain thrust fault north of Burlington (Doll and others, 1961). In a 3.3 to 10 ft (1 to 3 m) zone along the thrust surface, fractured clasts of these slivers are found in a matrix of ground and rewelded shale.

Estimates of displacement along the Champlain thrust fault have increased substantially as a result of regional considerations (Palmer, 1969; Zen and others, 1983; Stanley and Ratcliffe, 1983, 1985) and seismic reflection studies (Ando and others, 1983, 1984). The earlier estimates were less than 9 mi (15 km) and were either based on cross sections accompanying the Geologic Map of Vermont (Doll and others, 1961) or simply trigonometric calculations using the average dip of the fault and its stratigraphic throw. Current estimates are in the order of 35 to 50 mi (60 to 80 km). Using plate tectonic considerations, Rowley (1982) has suggested an even higher value of 62 mi (100 km). These larger estimates are more realistic than earlier ones considering the regional extent of the Champlain thrust fault.

Lone Rock Point

At Lone Rock Point the basal part of the Lower Cambrian Dunham Dolomite overlies the Middle Ordovician Iberville Formation. Because the upper plate dolostone is more resistant than the lower plate shale, the fault zone is well exposed from the northern part of Burlington Bay northward for approximately 0.9 mi (1.5 km; Fig. 1). The features are typical of the Champlain thrust fault where it has been observed elsewhere.

The Champlain fault zone can be divided into an inner and outer part. The inner zone is 1.6 to 20 ft (0.5 to 6 m) thick and consists of dolostone and limestone breccia encased in welded, but highly contorted shale (Fig. 3). Calcite veins are abundant. One of the most prominent and important features of the inner fault zone is the slip surface, which is very planar and continuous throughout the exposed fault zone (Fig. 3). This surface is marked by very fine-grained gouge and, in some places, calcite slickenlines. Where the inner fault zone is thin, the slip surface is located

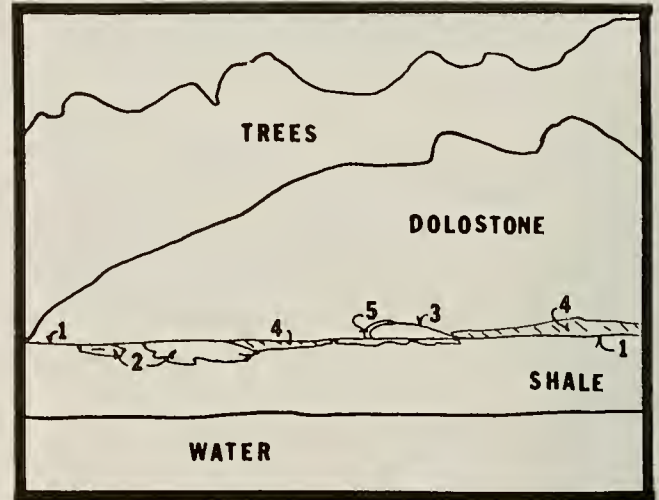


Figure 3. View of the Champlain thrust fault looking east at the southern end of Lone Rock Point (Fig. 1). The accompanying line drawing locates by number the important features discussed in the text: 1, the continuous planar slip surface; 2, limestone slivers; 3, A hollow in the base of the dolostone is filled in with limestone and dolostone breccia; 4, Fault mullions decorate the slip surface at the base of the dolostone; 5, a small dike of shale has been injected between the breccia and the dolostone.

along the interface between the Dunham Dolomite and the Iberville Shale. Where the inner fault is wider by virtue of slivers and irregularities along the basal surface of the Dunham Dolomite, the slip surface is located in the shale, where it forms the chord between these irregularities (Fig. 3). The slip surface represents the surface along which most of the recent motion in the fault zone has occurred. As a consequence, it cuts across all the irregularities in the harder dolostone of the upper plate with the exception of long wave-length corrugations (fault mullions) that parallel the transport direction. As a result, irregular hollows along the base of the Dunham Dolomite are filled in by highly contorted shales and welded breccia (Fig. 3).

The deformation in the shale beneath the fault provides a basis for interpreting the movement and evolution along the Champlain thrust fault. The compositional layering in the shale of the lower plate represents the well-developed S1 pressure-resolution cleavage that is essentially parallel to the axial planes of the first-generation of folds in the Ordovician shale exposed below and to the west of the Champlain thrust fault (Fig. 4). As the trace of the thrust fault is approached from the west this cleavage is rotated eastward to shallow dips as a result of westward movement of the upper plate (Fig. 4). Slickenlines, grooves, and prominent fault mullions on the lower surface of the dolostone and in the adjacent shales, where they are not badly deformed by younger events, indicate displacement was along an azimuth of approximately N60°W (Fig. 4; Hawley, 1957; Stanley and Sarkesian, 1972; Leonard, 1985). The S1 cleavage at Lone Rock Point is so well developed in the fault zone that folds in the original bedding are largely destroyed. In a few places, however, isolated hinges are preserved and are seen to plunge eastward or southeastward at low angles (Fig. 4). As these F1 folds are traced westward from the fault zone, their hinges change orientation to

the northeast. A similar geometric pattern is seen along smaller faults, which deform S1 cleavage in the Ordovician rocks west of the Champlain thrust fault. These relations suggest that F1 hinges are rotated towards the transport direction as the Champlain thrust fault is approached. The process involved fragmentation of

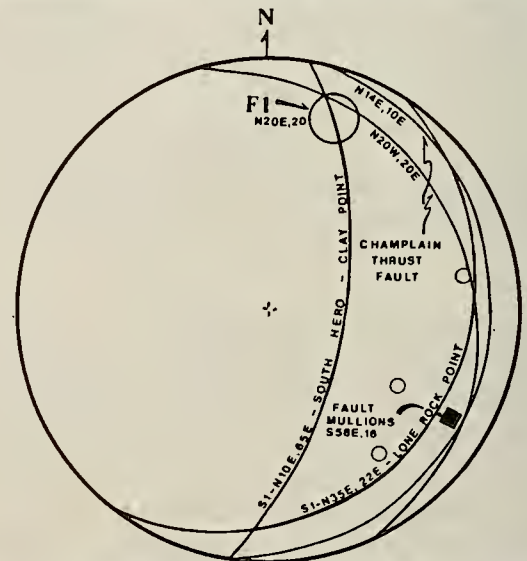


Figure 4. Lower hemisphere equal-area net showing structural elements associated with the Champlain thrust fault. The change in orientation of the thrust surface varies from approximately N20°W to N14°E at Lone Rock Point. The orientation of S1 cleavage directly below the thrust is the average of 40 measurements collected along the length of the exposure. S1, however, dips steeply eastward in the Ordovician rocks to the west of the Champlain thrust fault as seen at South Hero and Clay Point where F1 hinges plunge gently to the northeast. Near the Champlain thrust fault F1 hinges (small circles) plunge to the east. Most slickenlines in the adjacent shale are approximately parallel to the fault mullions shown in the figure.

the F1 folds since continuous fold trains are absent near the thrust. Much of this deformation and rotation occurs, however, within 300 ft (100 m) of the thrust surface. Within this same zone the S1 cleavage is folded by a second generation of folds that rarely developed a new cleavage. These hinges also plunge to the east or southeast like the earlier F1 hinges. The direction of transport inferred from the analysis of F2 data is parallel or nearly parallel to the fault mullions along the Champlain thrust fault. Stanley and Sarkesian (1972) suggested that these folds developed during late translation on the thrust with major displacement during and after the development of generation 1 folds. New information,

however, suggests that the F2 folds are simply the result of internal adjustment in the shale as the fault zone is deformed by lower duplexes and frontal or lateral ramps (Figs. 1, 2). The critical evidence for this new interpretation is the sense of shear inferred from F2 folds and their relation to the broad undulations mapped in the fault zone as it is traced northward along Lone Rock Point (Fig. 1). South of the position of the thick arrow in Figure 1, the inferred shear is west-over-east whereas north of the arrow it is east-over-west. The shear direction therefore changes across the axis of the undulation (marked by the arrow) as it should for a synclinal fold.

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STRUCTURES IN THE DOG RIVER FAULT ZONE BETWEEN NORTHFIELD AND MONTPELIER, VERMONT

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INTRODUCTION

The purposes of this trip are to examine the lithologic and structural character of rocks exposed in the Dog River fault zone (DRFZ) along a traverse between Northfield and Montpelier, Vermont, and in the process to better understand their tectonic significance. The rocks exposed within the DRFZ have long been recognized as significant from a stratigraphic standpoint, but recognition of their structural character is recent. This author has done detailed mapping along only a 10-mile section of the zone, but reconnaissance studies suggest that its structural character is similar north to the Quebec border and south as far as Randolph, Vermont, a distance of about 80 miles. Support for this work has come from the Vermont State Geologist (Dr. Charles A. Ratte) and from Norwich University.

The location of the DRFZ (Figure 1) corresponds with a previously identified geologic entity to which several names have been applied. These include the "Richardson Memorial Contact" (RMC), a pre-Silurian/Silurian boundary, the Green Mountain Anticlinorium/Connecticut Valley-Gaspe Synclinorium transition, and the Taconian Line (TL). For a variety of reasons, none of these names is entirely appropriate for the regional structure to which they all refer.

The "Richardson Memorial Contact (RMC): Geologists working in Vermont during the past several decades have recognized that Richardson and his associates had identified a geologic boundary which could be traced the entire length of the State (and beyond). For purposes of informal communication, this boundary has been referred to as the RMC (Richardson Memorial Contact) with no implication as to the nature of the boundary. Rocks on the western side of the boundary were designated by Doll and others (1961) as the Missisquoi Formation of Ordovician age, but Richardson and Camp (1919) considered these rocks to be of Cambrian age. East of these "Cambrian" rocks, Richardson (1919) reported the identification by Dr. Rudolf Ruedemann, State Paleontologist, Albany, N.Y., of numerous graptolites from rocks of the Northfield and Waits River Formations. The ages of these fossils were reported to be Lower to Middle Ordovician. Richardson's mapping led him to conclude that this boundary was an erosional unconformity. Certainly, his identification of what he considered a basal conglomerate in the Shaw Mountain

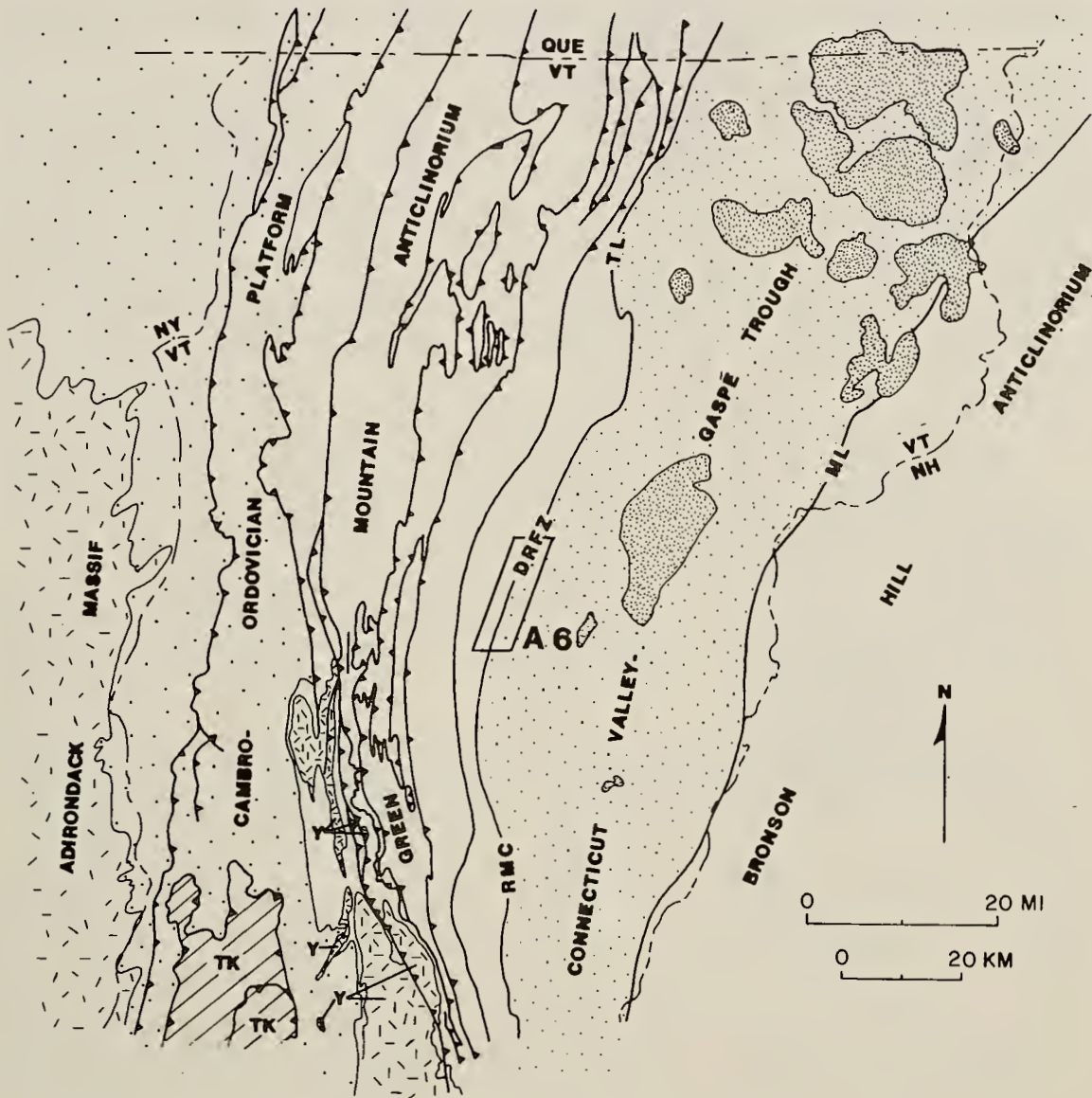


Figure 1. Tectonic map of northern Vermont. Western half after Stanley and Ratcliffe (1985). TK=Taconic klippe; Y=Lincoln Mountain and Green Mountain basement; RMC=Richardson Memorial Contact; DRFZ=Dog River Fault Zone; TL=Taconian Line (after Hatch, 1982); ML=Monroe Line

Formation of Currier and Jahns (1941) strongly influenced his conclusion that this boundary was an erosional unconformity.

The primary problem with calling this boundary a "contact" is the implication of lithologic constancy along one side or the other, whereas formations on opposite sides of the RMC change frequently as a result of tectonic imbrication. Locally, conglomerates and crinoid-bearing, quartz-rich calcareous sandstones of the Shaw Mountain Formation are in fault contact with various rocks of the Missisquoi Formation, but most often the contact is between the dark gray slates and phyllites of the Northfield Formation and the wide variety of units of the Missisquoi Formation. Locally rocks of the Waits River Formation occur in contact with Missisquoi units. All of these contacts are also invariably fault contacts, but where the conglomerates are absent, the fine-grained rocks lack the lithologic basis for assuming that they rest on an erosional unconformity. Instead, rocks of the Northfield Formation appear to represent a distal facies of the Waits River Formation, often containing turbiditic, fine-grained graded beds. Intraformational conglomeratic horizons occur within rocks of the Northfield Formation, probably representing submarine slide conglomerates, but they do not appear to be localized along the western margin.

Pre-Silurian/Silurian Boundary: Currier and Jahns (1941) reported the discovery of crinoidal limestone in the Shaw Mountain Formation near Northfield. They concluded at that time that the rocks were of Middle Ordovician age, but subsequent, more detailed studies have shown them to be most probably of Middle Silurian age (Boucot and Thompson, 1963; Doll, 1984). These determinations are supported by recent conodont studies (Denkler, K.E. and Harris, A.G., written comm., 1986). They identified 3 Pa element fragments of *Ozarkodina* sp. indet. of Silurian-Early Devonian morphotype and 1 P, 1 Sb, & 1 Sc element fragments of an oulodontiform apparatus from a 73.2 kg sample collected on Winch Hill Road off Route 12A just south of Lover's Lane (see Stop 10, this trip).

The interpretation of Currier and Jahns (1941) that the eastern Vermont stratigraphic section started with the Shaw Mountain Formation above an unconformity and continued up to the east with the Northfield Formation and the Waits River Formation, was concurred with by others and was adopted on the Centennial Geologic Map of Vermont (Doll and others, 1961). The recognition of the Silurian age of the fossils from the Shaw Mountain Formation had been made by the time of publication of that map. By accepting the assumption regarding the stratigraphic section and by recognizing the age of the Shaw Mountain Formation, it became widely popular to refer to the RMC as a pre-Silurian/Silurian boundary.

The assignment of an Ordovician age to rocks of the Missisquoi Formation (includes the Moretown Formation of Cady (1956)) on the western side of the RMC is very tenuously tied to a long distance correlation with fossiliferous rocks in Quebec. At this stage there is no reason to assume that an age assignment of

Ordovician is incorrect, but it is important to remember that it is an assumption rather than an established fact. Regarding the age of the rocks of the Shaw Mountain Formation, all evidence suggests that the assignment of a Middle Silurian age is correct. The problem with referring to the name of the boundary in question using ages of rocks stems from newly reported graptolite data (Bothner and Berry, 1985; Bothner and Finney, 1986) which suggests that at least parts of the Northfield, Waits River and Gile Mountain Formations are of Ordovician age. If these age determinations hold up to further scrutiny, then the discontinuous lenses of the Middle Silurian Shaw Mountain Formation may be sandwiched between packages of older rocks and the RMC would only very locally be a pre-Silurian/Silurian boundary.

Green Mountain Anticlinorium/Connecticut Valley-Gaspe Synclinorium Transition: Cady (1960a, 1960b) and Doll and others (1961) popularized the notions that much of the western half of Vermont is anticlinorial in character (with the Precambrian basement of the Green Mountain massif having been uplifted and its previously continuous cover removed by erosion), while the eastern half of the State is synclinorial in character (a folded basin presumably resting on the same basement as is exposed to the west). Stanley and Ratcliffe (1985) have quite successfully replaced the simplified notion of a breached anticlinal structure with an elegant model in which the dominant structures are westward directed thrusts (Figure 1), but no such solution has been developed for the complex structures within the Connecticut Valley trough.

The sedimentological and structural history of the Connecticut Valley rocks is still uncertain. Hatch (1986; this volume) has suggested that the pelitic rocks in the section (including the Northfield Formation and the Meetinghouse Slate) may be the youngest. This would create a situation in which the rocks in the core of the belt are oldest and those on the margins are youngest - an anticlinorial structure rather than synclinorial. Both the western margin (Westerman, 1985) and the eastern margin (Hatch, 1985; this volume) of the CV trough are recognized as fault zones, but the senses and extents of motion in these zones remain to be fully studied.

The Taconian Line: Cady (1960b) described the Connecticut Valley synclinorium as being bounded on its western margin by the Taconian unconformity which would correspond to what has been described above as the RMC, a pre-Sil/Sil boundary and the GMA/CV-GS transition. Cady concluded from his studies that the rocks west of the boundary had experienced the Taconic orogeny during Ordovician time while the rocks east of the line, being Silurian or younger in age, had not experienced that orogeny. Hatch (1982) used stratigraphic, structural and metamorphic bases to define this line which he named the Taconian Line. Hatch's work preceded the recent reports of Ordovician graptolites in rocks from southeastern Quebec within the CV-GS (Bothner and Berry, 1985) and from the Montpelier area in the Northfield and Waits River Formations (Bothner and

Finney, 1986). It also preceded recognition of the extent of strain which has occurred along that line in central Vermont.

It is significant to note that radiometric dating of metamorphic minerals, the method used for assigning ages to metamorphic and orogenic events in western New England, has not produced Taconian ages from rocks of the Moretown (Missisquoi) Formation (see Sutter and others, 1985; Jo Laird, person. comm., 1987). This opens up an additional possibility, namely that the rocks of the Moretown Formation belong on the eastern side of the conceptual Taconian Line. No evidence as yet been found in central Vermont by this author to indicate that rocks of the Moretown Formation experienced more orogenic activity that did those of the Northfield and Waits River Formations (Westerman, 1987). Osberg and others (1971) and Hatch (1975) reported small folds in rocks of the Moretown and Hawley Formations on the west side of the TL which have not been found on the east side. Although these folds are rare and not large enough to effect the map pattern at a scale of 1:24,000 (Hatch, 1982), Osberg and Hatch suggest that they represent pre-Silurian (i.e. Taconian) deformation.

STRATIGRAPHY

The earliest detailed mapping in the Dog River Valley was done by Richardson and Camp (1919). They reported as one of their reasons for selecting the area for study to be that "it falls in the line of the erosional unconformity between the Upper Cambrian and the Ordovician formations represented in the eastern half of the State". The next generation of study of these rocks was organized by Richard H. Jahns (Currier and Jahns, 1941; White and Jahns, 1950). It was the work of Jahns and others that provided the location of geologic contacts in the Barre West 15' quadrangle for the Centennial Geologic Map of Vermont (Doll and others, 1961). At the time of publication of the Centennial map, the rocks of the area were all assigned formational names. The age of each formation was thought to be known based on either 1) local fossils, 2) correlation with fossiliferous rocks elsewhere (generally along strike to the north in Canada), or 3) presumed stratigraphic relations with adjacent, although perhaps tenuously dated, formations.

The most recently published map of the study area remains that of Doll and others (1961) and Figure 2 shows the distribution of formations as they are represented on that map. In their interpretation, the geologic section shown here started far to the west at the basal unconformity of the Paleozoic section resting on the Grenville basement. It has younged eastward through the Cambrian section, here reaching the Missisquoi Formation (Ordovician). The members of that formation include the Moretown member at the base (west) containing two horizons of carbonaceous phyllite and slate. Capping that member are the Harlow Bridge Quartzite and the Cram Hill member. All of these rocks are reported

BEDROCK GEOLOGY OF THE DOG RIVER VALLEY
(after Doll and others, 1961)

Key to Letter Symbols

- Dw - Waits River Formation
- DSn - Northfield Formation
- Ss - Shaw Mountain Formation
- Missisquoi Formation
 - Och - Cram Hill Member
 - Ohb - Harlow Bridge Quartzite
 - Ocp - carbonaceous pelite
 - Om - Moretown Member

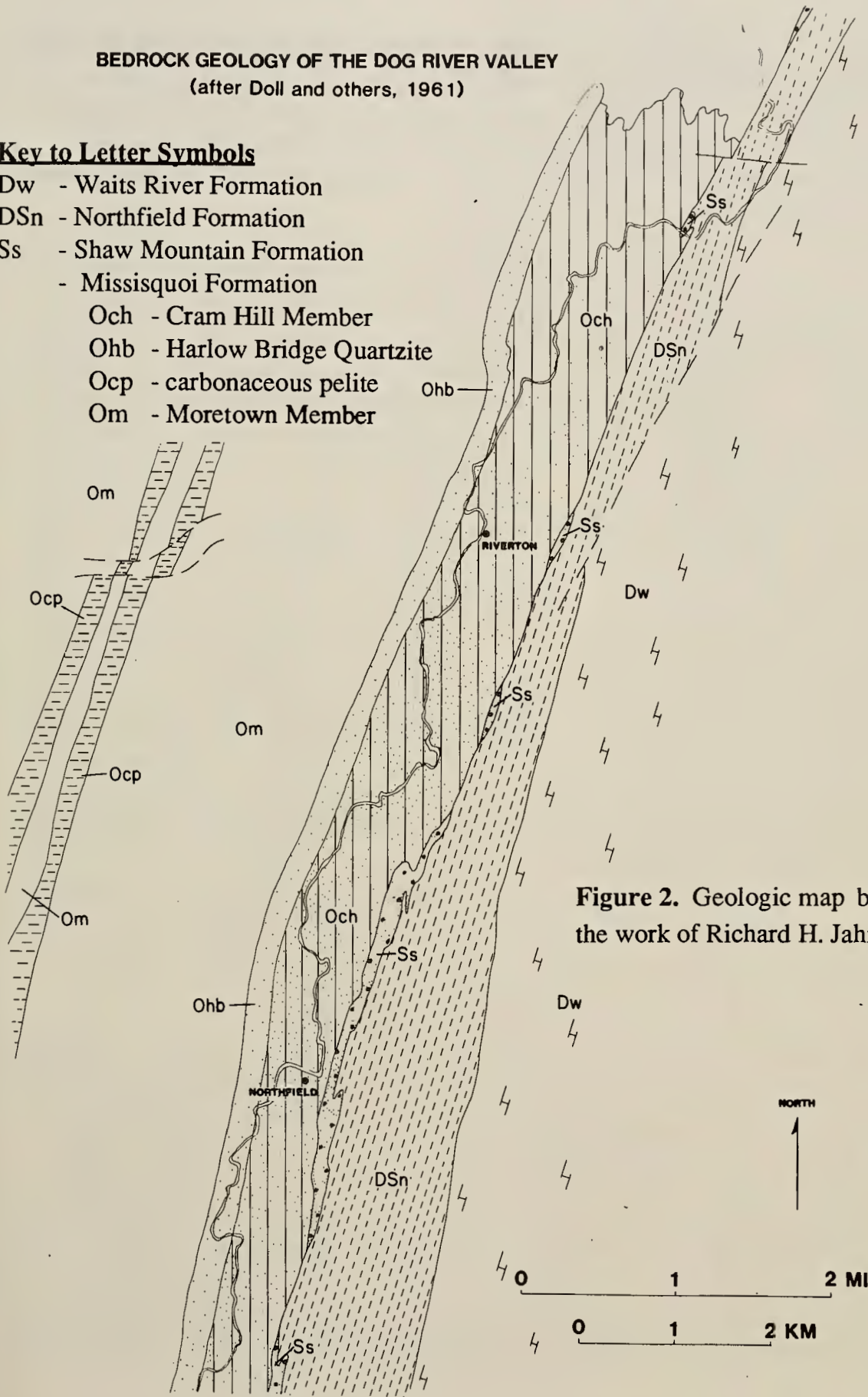


Figure 2. Geologic map based on the work of Richard H. Jahns

throughout the literature to be of Middle Ordovician age, based on correlation with fossiliferous rocks of the Magog Group in the Eastern Townships of Quebec (Cooke, 1950; Berry, 1962). The Missisquoi Formation is shown in Figure 2 after Doll and others (1961) to be unconformably overlain by the conformable sequence of discontinuous Shaw Mountain Formation, the Northfield Formation and the Waits River Formation.

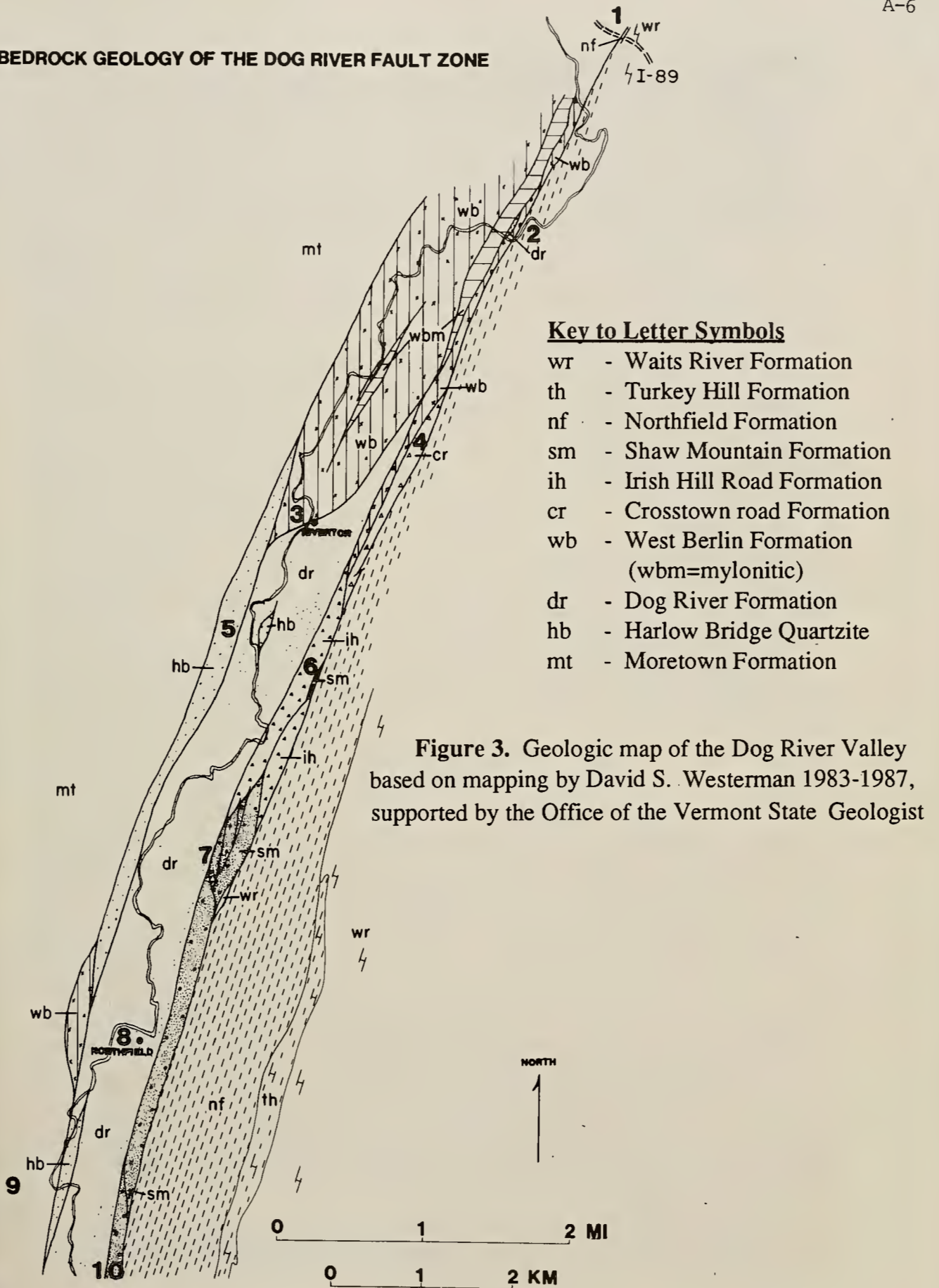
Results of more recent mapping are shown in Figure 3 in which the basic structural character of the DRFZ can be seen. A braided pattern of fault traces has been produced by the imbrication of various slices within the zone. This same pattern can be seen at individual outcrops, in hand specimens, and in thin sections. The only formational contacts that appear to be depositional are those between the Northfield, Turkey Hill and Waits River Formations. The lithologic character of each of the mappable units in the study area is briefly discussed below. The members of the Missisquoi Formation of Doll and others (1961) shown in Figure 2 have been remapped in part with some names being dropped and others added. All the subdivisions of that formation are here informally given formational status. The lithologic character of each of the mapped units is discussed below.

Moretown Formation: Cady (1956) named this unit based on mapping in the Montpelier quadrangle. Where shown on Figure 3, it is composed primarily of interbedded, blue-greenish gray quartzite and phyllite. The quartzites commonly exhibit well-developed pressure solution laminae. Also included here are carbonaceous phyllites and slates which occur as discontinuous lenses within the chloritic quartzites, but their distribution lacks the continuity or regularity shown in Figure 2. More detailed mapping will allow these rock types to be shown as a separate unit. Greenstone dikes are also present in the Moretown Formation, along with a variety of metamorphosed volcanic and volcanoclastic rocks.

Harlow Bridge Quartzite: This unit consists of massive, buff weathering beds of pale green quartzite containing minor interbeds of greenish-gray phyllite. Bedding thicknesses are typically greater than 15 cm and often approach 1 m. Locally associated are pyrite-spotted, medium-grained granulitic greenschist dikes composed primarily of chlorite, albite, epidote and quartz.

Dog River Formation: This formation replaces most of the rocks in the southern half of the study area mapped as Cram Hill member by Jahns and shown as such on the Centennial Map of Doll and others (1961). Non-volcanic, rusty-weathering quartzites and gray phyllites dominate the formation. Bedding is commonly preserved in these rocks but is often dismembered on well-defined shear planes which parallel the axial planes of drag folds. Thicknesses of beds range from a few centimeters up to 1.5 meters, and quartzites dominate over phyllite more commonly than the reverse. This may well be due to the greater resistance of quartzite-rich zones and their tendency to form outcrops. At two locations within the main belt of this formation, thinly bedded "limestone" occurs with muscovite-

BEDROCK GEOLOGY OF THE DOG RIVER FAULT ZONE



rich quartzites. To the north this same assemblage is exposed as a fault-bound slice of the Dog River Formation within the West Berlin Formation (Figure 3).

West Berlin Formation: In the northern half of the study area, much of the area previously mapped by Jahns as Cram Hill member is here designated the West Berlin Formation. These rocks are dominantly intermediate and mafic metavolcanics (commonly porphyritic) along with light green metaquartzites and phyllites similar to those found in the Moretown Formation. Ankeritic greenschist is a common lithology in this formation and is thought to be restricted to it. These rocks weather with a characteristic thick, brick-red rind over a partially weathered orange horizon.

Zones of severe strain in this formation are common and are characteristically deeply weathered with pinkish-buff colors. Mineralogy and textures are highly variable, and "out of place" lithologies tend to show up in these zones. These include some talc-bearing rocks and brecciated zones (occasionally quartz rich).

Crosstown Road Formation: Greenschists in this formation lack ankerite and weather to a greenish color. They are composed mostly of plagioclase, chlorite, epidote and calcite and occasionally have porphyritic textures. An irregularly shaped plug of metagabbro is included in this unit and has overall chemical characteristics similar to the surrounding greenschists. The formation does not include quartzites and phyllites as do the other units described above.

Irish Hill Road Formation: These rocks are predominantly of pyroclastic and/or volcanoclastic origin and are locally interbedded with light green quartzites and dark gray phyllites. The volcanic rocks are quite readily recognized by their blue-green color and well-developed schistosity, and the frequent presence of 1-2 mm crystals of blue quartz and white plagioclase. Quartz streaks 0.5 mm wide and several mm long are common where the rocks are strongly sheared. Sharp contacts between various lithologies of this formation probably reflect original bedding. This layering typically parallels cleavage, but in some cases the layering shows considerable deformation and dismemberment by shear.

Shaw Mountain Formation: The two diagnostic lithologies found in this formation are quartz-pebble (cobble) conglomerate, and crinoidal, calcareous, white quartz metasandstone. Pinkish-buff weathering, calcareous quartzites are also common but their similarity to some rocks of the Dog River Formation make them less diagnostic. Currier and Jahns (1941) previously included felsic tuffs in this formation, but those rocks are now recognized as mylonites derived from the other lithologies of the formation. These yellowish-white schists consist almost entirely of quartz and muscovite, making an igneous origin unlikely. Textural variations grade through the full range from very fine-grained schists to only moderately deformed conglomerates.

Northfield Formation: Dark gray slates and phyllites dominate this formation. Where pyrite is absent and the texture is slatey, the rock has been quarried for

shingles and other construction products. Many areas, in particular along the western margin of the formation, are characteristically phyllitic and typically crenulated. Bedding is commonly seen, particularly on glacially scoured surfaces and quarry walls. Silty beds, often with sandy bases and pelitic tops, have been found throughout the formation. Brown weathering, calcareous sandy beds are rare. Several horizons of intraformational conglomerate have been observed in the old quarries, most notably in the quarry north of the Norwich University ski slope and the quarry on the north side of Lover's Lane just west of Route 12.

Turkey Hill Formation: In the southern half of the study area there is a mappable unit separating the Northfield Formation from the Waits River Formation. The field criteria for recognizing this unit are the presence of brown-weathering, calcareous metasandstone and the absence of the "punky brown" limestones which characterize the Waits River Formation. The phyllitic beds in this formation are indistinguishable from those in the adjacent formations. Bedding thicknesses are generally thin, unlike the thick beds typical of the Waits River Formation. The gradational character of the Turkey Hill Formation suggests that its contacts are depositional.

Waits River Formation: Dark gray phyllites, brown-weathering, calcareous metasandstones, and dark brown, punky "limestones" are the three lithologies which characterize this formation. Bedding is generally measured in 10's of centimeters, and is usually readily observed. The word limestone is here in quotation marks because it is actually a calcareous metasandstone probably representing calcareous flysch with the calcite having been derived as detritus.

STRUCTURE

Structural relationships in the study area are most quickly reviewed by examining the map view presented in Figure 3 and the stereograms shown in Figure 4. The stereograms show at a glance that all of the structures are geometrically related to the orientation of the DRFZ. The N23E trend of that zone with the steep westerly dips of the fault surfaces defines a plane to which all other structures can be related.

Bedding: Bedding is common in most of the formations of the study area but is generally not recognized in the greenschists and greenstones of the West Berlin and Crosstown Road Formations, or in the conglomerates of the Shaw Mountain Formation. From a structural perspective, it is important to note that nowhere have individual beds been traced between outcrops. At most large outcrops, bedding in all formations is commonly dismembered by shear at shallow angles to the plane of bedding.

Orientations of bedding are typically somewhat east of north with steep westerly dips (Figure 4A). Significant variations from the dominant trend

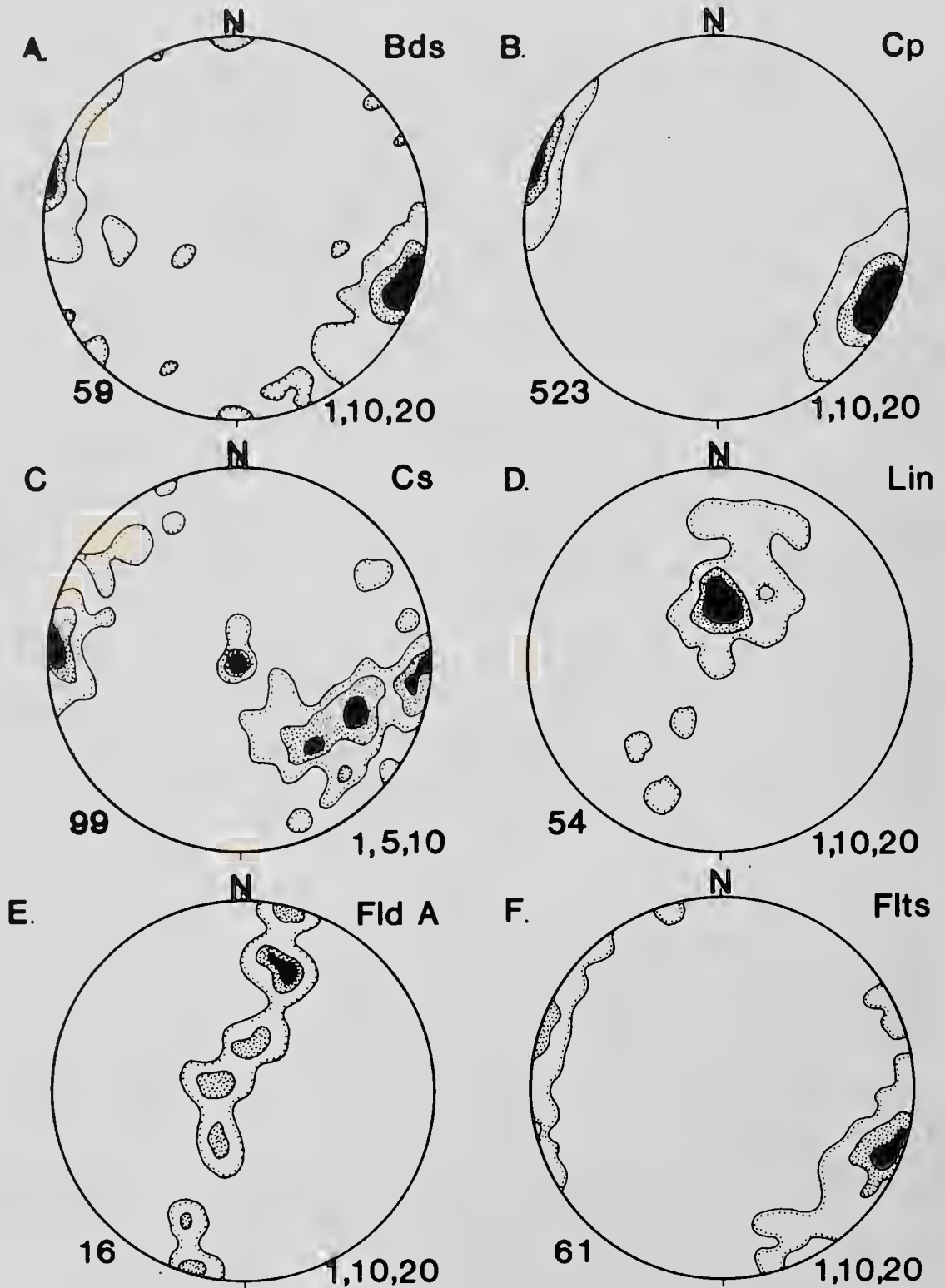


Figure 4. Equal area projections of structural elements. **A** - poles to bedding; **B** - poles to pervasive cleavage; **C** - poles to spaced cleavage; **D** - lineations including elongated minerals and clasts, cleavage intersections, and fold axes of crenulations; **E** - all fold axes; **F** - poles to fault surfaces; # measurements lower left; % contours lower right

represent measurements in the limbs or on the crests of folds. Bedding tends to parallel the major and minor fault traces of the DRFZ. This may be the result of either rotation of bedding during faulting or development of faulting parallel to the bedding orientations. Exceptions to the general parallelism of bedding and the dominant regional structure are local, such as in the noses of dismembered folds of quartzite.

Pervasive cleavage: Nearly all of the rocks in the study area have a pervasive foliation derived from the preferred orientation of platy minerals. Massive quartzites, containing only minor muscovite and chlorite, have only weak foliations, but foliation is well developed in most rocks. Even the quartz-pebble conglomerates of the Shaw Mountain Formation have a strong foliation as a result of both the significant muscovite content in the matrix and the preferred orientation of the distorted pebbles. The pervasive cleavage shows a more constant orientation than does bedding (Figures 4A and 4B), reflecting that it is axial planar to the large-scale folds in the beds.

Exceptions to the presence of a pervasive foliation are restricted for the most part to some fossiliferous layers in the Shaw Mountain Formation. Here the cleavage parallels the pervasive regional trend, but is spaced. Crinoid columns are well preserved locally, apparently having been located between shear surfaces which absorbed the strain. Their existence brings to mind the image of the composed gentleman in the Colt 45 advertizement, seated at his floating table and contented by his beer at hand, heading for the top of the waterfall.

Rare folds in the Northfield and Moretown Formations suggest that the pervasive cleavage does not everywhere represent the earliest cleavage. What appears as an older cleavage surface has been observed in isoclinal folds paralleling the tightly folded bedding. It is because of these observations that neither the $S_{0,1,2}$ or $S_{n, n+1, n+2}$ notation is applied here.

Spaced Cleavage: A spaced cleavage is common in rocks throughout the study area except in many of the rocks of the shear zones. This spaced cleavage tends to dip less steeply (to the west) than does the pervasive cleavage, and strikes tend to be more easterly (Figure 4C). Most of the measurements used to construct Figure 4C are of spaced cleavage associated with crenulations. Orientations of this cleavage are similar in all of the formations, but the sense of motion indicated by the geometry of the associated crenulations is not consistent. Indications of motion with a west over east sense are dominant. Also represented in Figure 4C are both closely spaced, near vertical fractures probably associated with the high-angle faulting of the DRFZ, and some nearly horizontal spaced cleavage.

The age of the spaced crenulation cleavage is constrained. It must postdate the juxtaposition of all of the units since it is present in them all with the same orientation, but it must predate the bulk of the shearing since it is commonly

absent in many of the fine-grained papery rocks of the shear zones. It is interesting to speculate that the spaced cleavage may be the product of a stress field in rocks which failed to "let go" and deform by pervasive strain and faulting. The similarity in orientation of crenulation fold axes, small-scale asymmetric fold axes, pervasive and spaced cleavage intersection lineations, mineral lineations in sheared pyroclastics, and long axes of pebbles (Figure 4D) suggests that all these structures either formed in the same stress field or have been rotated into their current orientations by movement in the DRFZ.

Folds: Folds identified to date include only those seen in individual outcrops. No folds have been recognized as mappable structures at a scale of 1:24,000. In most cases where outcrop size exceeds several meters, folds are observed to be dismembered by shear parallel to their limbs. As noted earlier, a few isoclinal folds have been found in the Moretown and Northfield Formations in which it appears that an early cleavage parallels the bedding and has been folded along with it. This suggests that the isoclinal folds observed are a second generation of such folds, the axial planes of which are parallel to both the bedding and the pervasive cleavage of the region.

Asymmetric, small-scale folds have axial plane orientations parallel the spaced cleavage noted above. They are simply a "blow up" of the crenulations described above and their fold axes nearly parallel the intersection lineation of the pervasive and spaced cleavage throughout the region. Figure 4E shows the orientations of all generations of fold axes in the study area.

Faults: Most of the rocks exposed in the Dog River Valley between Northfield and Montpelier have been deformed to some extent by faulting. The main traces of the faults in this zone occur in a belt about 1 mile wide, referred to in this paper as the Dog River fault zone (DRFZ). The zone is located geologically between the Moretown Formation on the western side and the Northfield Formation on the eastern side. Caught in the DRFZ are the rocks of the Dog River, West Berlin, Crosstown Road, Irish Hill Road, and Shaw Mountain Formations. Orientations of the faults in this zone are concentrated around N23E, 81W, with a second concentration near N51E, 85W (Figure 3 and Figure 4F).

Within the DRFZ the range of deformation is extensive. At one extreme, crinoid columnals are occasionally preserved in undeformed lenses of the Shaw Mountain Formation, and graded beds can be found in rocks of the Dog River Formation. At the other extreme, coarse, quartz-pebble conglomerates have been milled to produce papery, yellowish-white mylonites previously identified as felsic tuffs (Currier and Jahns, 1941).

Lineations of various origins are common in the sheared rocks of the DRFZ and they consistently plunge moderately to the north (Figure 4D). This corresponds with the measured orientations of deformed pebbles by Currier and Jahns (1941) in the conglomerates of the Shaw Mountain Formation. Study of oriented thin

sections of such sheared conglomerates show imbrication of lithons indicating west over east movement. Rotated plagioclase crystals in sheared tuffaceous rocks of the Irish Hill Road Formation show the same sense of motion. Crystals have well-developed, asymmetric tails of quartz and fibrous chlorite. These studies, coupled with the consistent northward plunges of other lineations, suggest that movement in the DRFZ was both high-angle reverse and left lateral.

DISCUSSION

The first conclusion reached from this study is that the Connecticut Valley trough in east-central Vermont is bounded on its western margin by a high-angle fault zone. On the Geologic Map of Massachusetts (Zen and others, 1983) a decollement surface is indicated for the same boundary, but in central Vermont the most recent movement in the zone appears to have been west over east with a left-lateral component. Perhaps this surface 1) once dipped gently to the east, 2) experienced westward thrusting, 3) was steepened by continued compression, 4) rotated through vertical, and 5) experienced a reversal of the movement direction.

The second conclusion is that the idea of an erosional unconformity at the western margin of the CV trough being overlain by a conformable stratigraphic sequence of Shaw Mountain, Northfield, Waits River, etc. seems highly unlikely. It is difficult to visualize a sedimentological succession from the lithologies of the Shaw Mountain Formation to those of the Northfield Formation. Clean, coarse, quartz-pebble conglomerates and crinoid-bearing "limestones" with abundant white quartz sand suggest a near shore, locally high energy, shallow water environment for the Shaw Mountain Formation. The morphotypes of the conodonts studied by Denkler and Harris (written communication, 1986) indicate that they "lived in a relatively shallow water, high energy environment". On the other hand, rocks of the Northfield, Turkey Hill, and Waits River Formations have deep water affinities as suggested by the thick turbiditic sequences and great volumes of pelitic sediments. Fred Larsen (pers. comm., 1984) has puzzled over the lack of calcite in the muddy tops of the thick, graded beds in these units. He speculates that deposition occurred below the CCCD with the calcite in the sands being detrital and preserved only because of rapid burial under the pelitic tops to the beds. Whatever the actual depth of deposition, it was probably much greater than that for the Shaw Mountain lithologies. The fact that contacts between Shaw Mountain rocks and Northfield rocks are always faults supports the idea that these units were deposited in different environments at some distance from each other.

After one eliminates the idea of a conformable stratigraphic sequence from the Shaw Mountain to Waits River Formation, and recognizes both the deep, quiet water environment for the Northfield Formation and faulted western margin to the CV trough, there's not much reason to postulate the presence of an erosional unconformity at most locations between these two terranes.

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ITINERARY

- 0.0 Starting at the Commuter Lot on the west side of I-89 at the Northfield exit (Exit 5), head left (east).
- 0.3 Turn left (north) on I-89. The outcrops between here and the first stop are in the Waits River Formation. Note the characteristic colors of the well-bedded phyllites and calcareous metasandstones.
- 4.5 A thick horizon of phyllite is exposed along the west side of I-89 here.
- 7.7 The hills to the east are the Barre and Knox Mountain plutons.
- 10.0 The color change in the rocks on the right corresponds to the Taconian Line of Hatch (1982)
- 10.1 Stop 1: Park well off the road before the Exit 8 offramp. At the east end of the large roadcut here are exposed the contacts between the Waits River, Northfield, and Missisquoi Formation (of Doll and others, 1961). Permission to stop here has been obtained from the Vermont Traffic Committee, Department of Public Safety. Future users of this guide who choose to stop here should anticipate being asked to move on immediately if they are observed by the Vermont State Police.

Start at the east end of the outcrop in the Waits River Formation. The brown weathering of the calcareous metasandstones is characteristic. Observe the pinstripe spaced cleavage which truncates the more steeply dipping earlier pervasive cleavage (possibly parallel bedding). Small-scale faults here trend N40E, 55W, segregating zones showing the older deformations. These faults show evidence of normal displacement.

The contact of the Waits River and Northfield Formations is obscured by extensive faulting (N30E, 83W). The Northfield Formation is here only 75 feet thick and evidence of structural features older than those of the main fault zone has been obliterated. Younger faults are present trending somewhat more easterly than the main zone and dipping moderately to the west. Abundant lenses of quartz are mixed in with the rocks of the Northfield Formation and have been faulted by the youngest faults.

The contact between the gray-black shiny phyllites of the Northfield Formation and the green rocks of the Missisquoi Formation is a fault, and the green rocks are strongly mylonitic in character. A laminar texture gives these rocks a gneissic appearance along the eastern margin of the unit. Further west, the rocks are still granulitic but lack the gneissic texture. The 300-foot section of light green schists and granulites has been cut by numerous faults which parallel the pervasive fabric and dip steeply to the west. The younger faults described in rocks to the east are present here as well.

- 10.2 Exit the interstate, slowly turning north and eventually east. The rocks here are undifferentiated lithologies of Cady's Moretown Formation (1956) and Doll and others' Missisquoi Formation (1961).
- 11.1 Continue straight at the lights, now driving east on Route 2. Cross the TL.
- 11.5 Turn south on Route 12, heading toward Northfield.
- 13.9 Turn left on a gravel road. Cross the Dog River and bend right (south), staying as close to the river as you can.

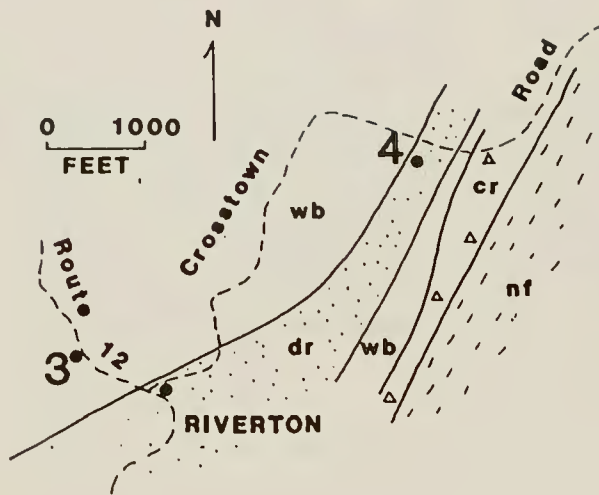
14.4 Stop 2: Stop at the steel gate and walk west to the river's edge. Outcrops in the small stream and north along the shore of the Dog River are phyllites of the Northfield Formation. After examining these rocks, work your way westward across strike. You will see that pale, rusty mylonites (apparently derived from a quartz-rich source) separate the main outcrop belt of the Northfield Formation from fault-bound slices of that unit further west at the water's edge. These slices of Northfield even include minor calcareous metasandstone beds, indicating the potential for such structures to survive when located between shear surfaces which are accommodating the strain.

West of the slices of Northfield are mylonitic rocks of uncertain origin. Currier and Jahns (1941) interpreted these rocks to be Shaw Mountain lithologies, but they are probably severely altered products of the West Berlin Formation since they occur elsewhere out in the main belt of that formation. Just west of the main corner of the river is an outcrop containing metamorphosed porphyritic andesite, mixed with the pale, ankerite-spotted mylonites, supporting the association of these rocks with the volcanic sequence of the West Berlin Formation.

Return to the cars, turn around and return to Route 12.

14.7 Turn left (south) on Route 12. Outcrops along the route here are primarily metavolcanics and metasediments of the West Berlin Formation.

17.8+ Stop 3: Park on the right in the pullout area. A variety of rocks of the West Berlin Formation are exposed in this area. At the northernmost outcrop to the west (through the bushes) are micaceous, rust-spotted quartzites and medium-grained greenschists. Further south and west (in the open pit) are folded metasediments, greenschists, and intermediate porphyritic flows, sills, or dikes. The small asymmetric folds seen here are characteristic of the rocks throughout and adjacent to the DRFZ. On the south side of the entry road to the quarry is ankeritic greenschist with the characteristic thick, brick-red weathering rind.



Key to Letter Symbols

- nf - Northfield Formation
- cr - Crosstown Road Formation
- wb - West Berlin Formation
- dr - Dog River Formation

Continue south on Route 12, cross the tracks, and cross the bridge.

18.0 Turn left (east) onto Crosstown Road.

19.0 Turn right and drive up Parker's driveway.

19.1 Stop 4: Park along the side of the driveway, walk east across the field about 200 feet and then south uphill to the boxed in end of the field. Just into the woods at the southwest corner of the field is an outcrop of dark gray phyllite of either the Dog River Formation or possibly the Northfield Formation as a slice caught within the DRFZ. Outcrops to the west are convincingly Dog River Formation, but here the phyllite lacks the characteristic rusty weathering habit of that unit. Just to the east, under an uprooted tree, is ankeritic greenschist of the West Berlin Formation. This lithology underlies the width of the field.

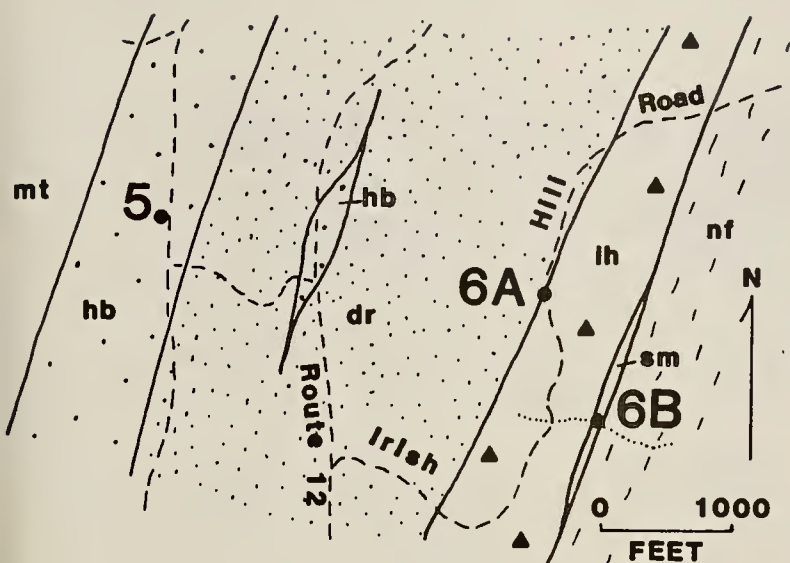
At the southeast corner of the field is a woodsroad heading off to the south and roughly paralleling the eastern contact of the somewhat out-of-place slice of West Berlin Formation. East of the road, where the elevation begins to rise sharply, are outcrops of the Crosstown Road Formation. These include greenschist (non-ankeritic) and green metatuffs, as well as a small plug of metagabbro (about 600 feet from the field). Depending on the time available, a walk down the woodsroad to a dry stream valley just before the next field (1500 feet) will allow examination of an enigmatic rock which appears as a sheared conglomerate with a chloritic matrix. Conversely, it may represent a brecciated, sheared, and recrystallized quartz vein.

Further to the east are exposures of the Northfield Formation. A common characteristic of this contact is an abundance of massive vein quartz, present both as isolated outcrops and as boulders.

Return to the cars and retrace your route back to Route 12.

20.1+ Turn right (north) onto Route 12.

20.4 Turn left, uphill on a gravel road.



Key to Letter Symbols

nf	- Northfield Formation
sm	- Shaw Mountain Formation
ih	- Irish Hill Formation
dr	- Dog River Formation
hb	- Harlow Bridge Quartzite
mt	- Moretown Formation

21.4+ Stop 5: The outcrop is in a triangle of the road on the right side and is partially painted blue. It consists of thickly bedded metaquartzites of the Harlow Bridge Quartzite with metapelitic interbeds. An argument can take place here over the facing direction of the beds.

Continue (south) on the gravel road.

21.5 Turn left (east) on a gravel road.

21.7- Cross the bridge. Large outcrops of thinly bedded metapelites and metasiltstones are exposed below with abundant well-developed, asymmetric folds.

21.7+ Cross the railroad tracks. The first outcrop exposed to the north along the tracks is an isolated slice of Harlow Bridge Quartzite. Turn south on Route 12.

22.0 After passing Ellie's Market on your left, turn left (east) on the gravel road known locally as Irish Hill Road (also Darling Hill Road).

22.6 Stop 6a: The western contact of the Irish Hill Road Formation is exposed on the east side of the road where the road begins to flatten. There is a small turnaround on the left side just above the outcrop. At this contact the Irish Hill Road Formation is strongly sheared with the most recent movement in a plane oriented N36E, 84W. A well-developed lineation in that plane comes from the streaking of blue quartz crystals, now oriented 61, N3E. Plagioclase crystals show the development of pressure shadow tails of mosaic quartz, oriented such that a west over east sense of rotation is indicated. This development of northward plunging lineations and slip with a reverse sense of motion is consistent throughout the fault zone. It is interpreted to indicate that the faulting included a component of left-lateral motion.

Drive back downhill. Note the saprolitic outcrops on your left (east).

22.8- Stop 6b: Park on the left (east) side of the road. Walk east up the path to the power line. In this immediate area there are two major fault traces. One separates Northfield phyllites from Shaw Mountain lithologies (east edge of the power line), and the other separates the Shaw Mountain rocks from sheared pyroclastics of the Irish Hill Road Formation (west edge of the power line). In the small brook at the east edge of the power line are dark gray, mylonites derived from Northfield lithologies but containing late, rhombic carbonate crystals which are filled with oriented inclusions of the minerals of the mylonitic fabric.

On the west side of the power line and south of the brook are outcrops of the Shaw Mountain Formation. These include muscovite-ankerite-quartz schist exposed in the power line clearing. The carbonate crystals in the schist postdate the development of the mylonitic fabric since they are undeformed and contain fine-grained, oriented inclusions. Conglomerate is exposed to the west in the edge of the woods. Deformation there has been primarily brittle with extensive recrystallization. Fabrics seen in thin section indicate a reverse sense of motion on the westerly dipping surfaces.

Back to the north along the edge of the power line and downstream in the brook valley are various rocks of the Irish Hill Road Formation. Many contain the blue quartz and white plagioclase crystals that serve as the diagnostic characteristic of the formation. Dismembered folds in these rocks are well preserved and strain indicators confirm the west over east sense of motion recorded elsewhere.


Return to cars, retrace your route to Route 12.

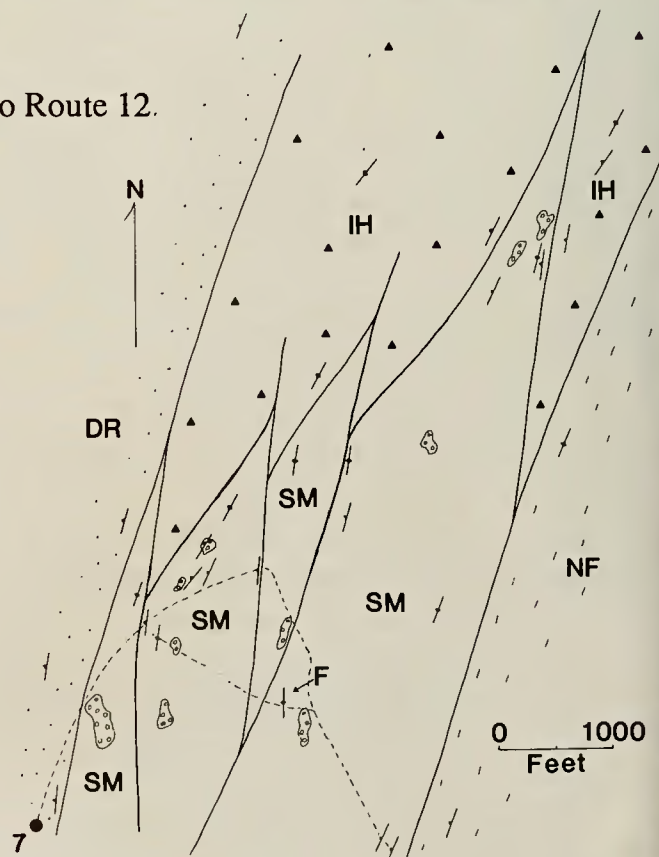
23.3 Turn left on Route 12.

24.5 Turn left (east) on Davis Ave.
The road bends south.

24.7+ Bear left at the fork.

Key to Symbols

- nf - Northfield Formation
- sm - Shaw Mountain Formation
conglomerate
- ih - Irish Hill Road Formation
- dr - Dog River Formation
- F - fossil locality
-  - foliation trend
-  - woodsroad



24.9 Stop 7: Park on the left (north) side just before the power line. Walk uphill under the power line, following the woodsroad. The first cliff on the east side is an exposure of coarse, quartz-pebble conglomerate of the Shaw Mountain Formation. Sheared phyllitic rocks of the Dog River Formation are exposed in the fields to the west. Continue up the woodsroad to where it leaves the power line and forks. Go straight and follow a clockwise loop to examine the various lithologies and structures exposed in this area.

Calcareous metaquartzites are exposed in the woodsroad on the way up the hill. They weather to a characteristic rusty-orange color and often have a strongly developed foliation. At the north end of the loop, just before the road make a sharp corner back to the southeast, strongly sheared rocks characteristic of the Irish Hill Road Formation are exposed in the road. They appear to have been translated to that location in a fault zone. Note the Blue quartz grains in these rocks. Continue, noting the large outcrop of conglomerate on the south side. When you will reach a junction with a branch of the woodsroad coming in from the northwest, take it and locate the small quarry of rusty conglomerate on the south side of the road. Varying degrees of strain are exhibited in this quarry, and the foliation of the muscovite-rich matrix of the conglomerate is well developed. Sulfide mineralization is predominantly pyrite here. Also noted is a emerald-green mineral occurring in very small crystals (fuchsite?).

Return to the woodsroad, continue downhill a few paces, and then move northward to the white outcrops of "limestone". Crinoid fossils are present in here in calcite-rich metasandstones.

Return to the cars and retrace your path to Route 12.

25.5+ Turn left (south) on Route 12.

26.9+ Turn right (west) at the blinker. Drive west around the commons to the southwest corner and head west on Wall Street.

27.1 Stop 8: Park on the right, opposite the small parking lot of the TDS Company. Walk south across the road to the outcrops exposed behind the parking lot. These rocks are located within a broad zone of shearing in the Dog River Formation. Dismembered folds of quartzite are common here, as are braided fabrics. Note the characteristic relationship between the pervasive cleavage oriented N19E, 72W and the spaced cleavage oriented N42E, 57W.

Continue west on Wall Street.

27.2 Turn left (south) on Water Street. The road turns to gravel.

28.1 Bend right (west) and up hill.

28.4 Turn left (south) on a dirt/gravel road.

28.8 Stop 9: Park here at Dole Farm. The outcrop to examine is located 400 feet south, just east of the edge of the field. Exposed here are pale green quartzites and associated phyllites characteristic of much of the eastern portion of the Moretown Formation. The variation in thickness of a massive quartzite bed here may represent either a channel fill or a tectonic feature. Don't forget to look up and enjoy the view.

Return to the cars and retrace your route toward town.

30.4 Turn right (east) on Wall Street. Cross the Dog River, the railroad tracks, and drive east along the south edge of the commons.

30.6 Turn right (south) on Route 12.

31.3 Pass the main campus of Norwich University.

31.6 Bear right (southwest) on Route 12A.

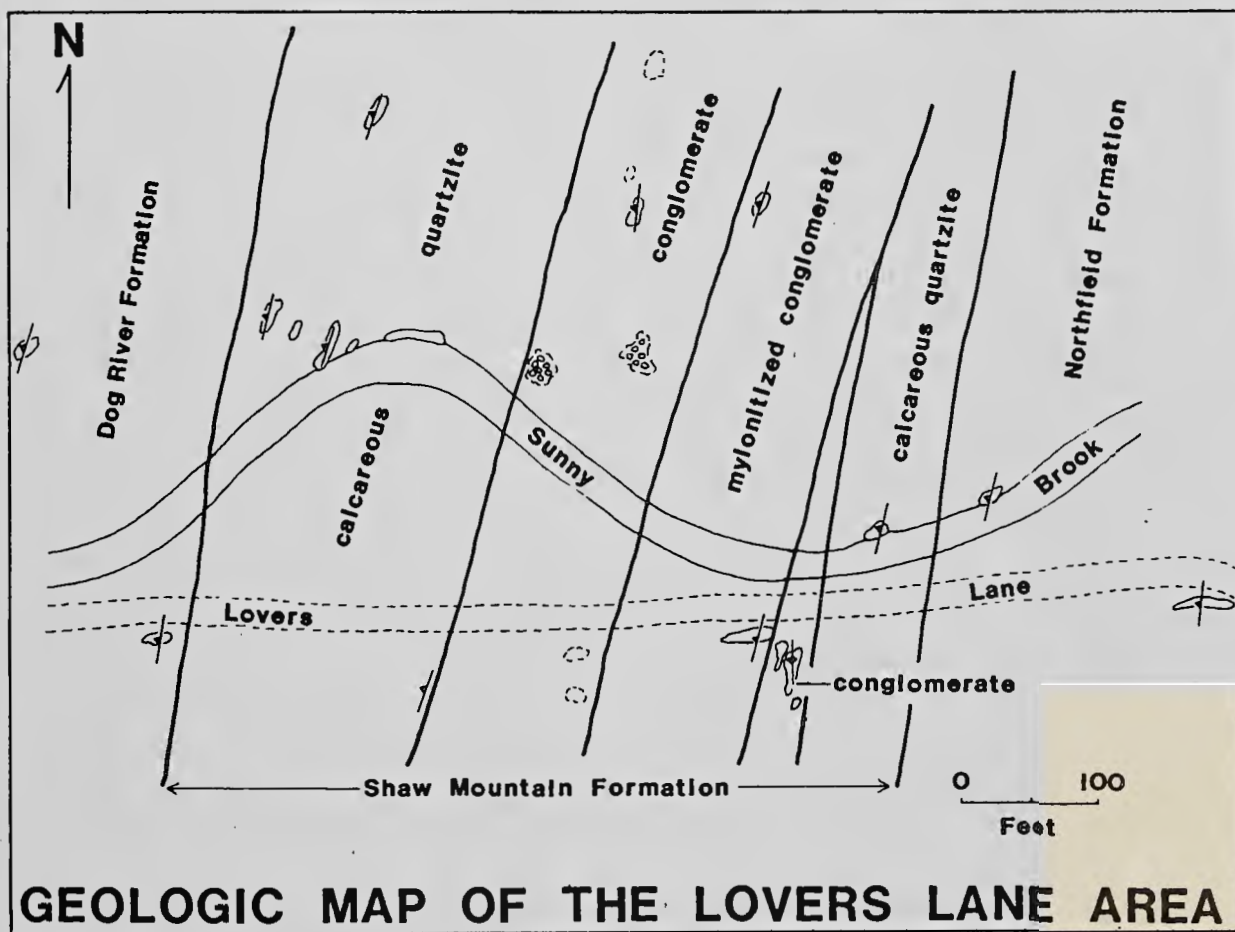
32.4+ Turn left (east) on Lover's Lane.

32.6 Stop 10: Park on the right in the turnout area. This traverse starts in the rusty-weathering, dirty quartzites and phyllites of the Dog River Formation exposed at the roadside just east of the turnout. Continuing east you cross a belt of calcareous quartzites. These rocks are not exposed along the road, but study of outcrops north of Sunny Brook shows general homogeneity across it. The exception is an increased degree of deformation along the western margin. Fossiliferous rocks found by Currier and Jahns (1941) are located on strike to the south on Winch Hill Road. That is also the location of the sample from which Denkler and Harris have recovered and identified conodonts (see page 111).

At the east edge of the calcareous quartzite belt is a fault zone, partially exposed in a small driveway off the south side of Lover's Lane. Some of the rocks here in the fault zone appear out of place, being most similar to volcanics found further to the north.

Continuing east along Lover's Lane, note the large boulders of conglomerate on the south side of the road. These boulders are typical textural representatives of moderate deformation of this lithology. The next outcrop at roadside is severely deformed conglomerate. Here the rocks are very rusty on weathered surfaces but nearly white on fresh surfaces. They have a splintery and papery habit, representing what Currier and Jahns (1941) mapped as papery, felsic tuffs. Careful examination shows unmilled eyes of quartz preserved in these mylonites. Oriented thin sections show imbrication of wedges with fabrics indicating a west over east and left-lateral motion. The results of partial recrystallization of quartz clasts by pressure solution are commonly preserved. Less deformed conglomerates are exposed in the woods just east of the fine-grained mylonite zone.

Walk east on Lover's Lane across the notch in the hillside. Note the half dozen or so spring houses located in this zone. Calcareous quartzite, perhaps connected to one of Currier and Jahns' fossil localities, is exposed on the north side of Sunny Brook. The first outcrops of the Northfield Formation phyllites exposed along the road are located just before the corner, and they show little or no evidence of the faulting nearby. Bedding is preserved with silty and calcareous sandy beds present. Other outcrops of Northfield occur back to the west in the stream and constrain the location of the contact.



Return to the cars and.....

END OF TRIP

If you have left cars at the commuter lot up at Exit 5, drive east on Lover's Lane to Route 12, cross Route 12 and drive east on the Gen. Harmon access road up toward I-89. You will have driven through the Northfield Formation, the Turkey Hill Formation, and into the Waits River Formation. Several post-metamorphic mafic dikes are exposed in the new road cuts of this traverse.

METAMORPHIC VEINS IN THE PALEOZOIC ROCKS OF CENTRAL AND NORTHERN VERMONT

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INTRODUCTION

The trip will examine the petrology and structural setting of metamorphic veins formed during the Taconic and Acadian Orogenies in the Paleozoic rocks of central and northern Vermont. In the Cambro-Ordovician rocks on the east limb of the Green Mountain anticlinorium in central Vermont, four generations of veins with primary metamorphic assemblages are found. In the Siluro-Devonian rocks of the Brownington syncline and the Strafford-Willoughby arch, two prominent generations of metamorphic veins are present. For each vein generation, much of the mineral growth apparently occurred as metamorphic grade was decreasing from the peak conditions of the metamorphic event with which it was associated. Due to time constraints and the lack of access to many of the best studied outcrops in the younger rocks (roadcuts on I-89 and I-91), the trip will primarily focus on the metamorphic veins in the Cambro-Ordovician rocks.

The two major Paleozoic orogenic episodes each consisted of a sequence of tectonic events that involved both deformation and metamorphic mineral growth. Various approaches have been used to decipher the tectonic history of the region. Many studies have emphasized structural aspects, others have been concerned with petrologic aspects, and some have tried to combine both. The approach used here combines petrologic and structural evidence and attempts to make a direct correlation of deformational and mineral growth features by using several generations of prominently developed metamorphic veins. Use of metamorphic veins in northern Vermont as structural and petrologic markers was initially outlined by Anderson and Albee (1975).

The study of the polymetamorphic history of central Vermont by White and Jahns (1950) was the first detailed work of its kind done in the study area. They and most others that followed relied primarily on the sequence of superimposed deformational elements, in particular the sequence of folds and *s*-surfaces. Cady (1969) summarized most of the other previous work in the area. Modifications to the earlier views of the deformational history of northern Vermont have been discussed by Albee (1972) and Anderson (1977a).

The presence of more than one distinctly separate generation of metamorphic mineral growth has long been recognized in the study area. Albee (1968) and Lanphere and Albee (1974) established the presence of at least two generations, one or more generations of Ordovician age and one or more of Devonian age. Laird (1977) and Laird and Albee (1981a, 1981b) have shown the presence of multiple generations of amphiboles in the Paleozoic mafic schists of Vermont and have used amphibole compositions to show differences in baric types of the different generations.

The interpretation of systematic compositional variation in chemically zoned mineral grains is a central concern of this study. Zoning could have been produced by growth during changing physical and/or chemical conditions or by solid-state diffusion after growth or by some combination of the two. Systematic variation that is complex was most likely produced by growth over a period of changing conditions (Anderson, 1977b).

All compositional data discussed were acquired on automated electron microprobes using wavelength-dispersive spectrometers. Analytical and data treatment methods of this study have been discussed by Anderson (1983).

Structural setting of the veins

The structural setting of the metamorphic veins area can be described by considering orientation, form, size, and position within the sequence of superimposed deformational elements. The younger vein generations, the least distorted by superimposed deformation, exhibit a variety of forms including lens-shaped veins parallel to an axial-plane *s*-surface, en echelon gash veins, veins which fill irregular pull-apart structures of several sorts, and sheet-like fracture-filling veins. One vein generation may have several types of forms that vary in predominance from one part of the area to another or vary locally with rock type. Most of the major vein generations are

geometrically related to s-surfaces, but the minor vein generations and one major generation are not. This latter major generation of relatively undeformed veins is related to late mineral growth in the host rocks. The minor vein generations have no apparent relationship to mineral growth in the host rock and are almost invariably of very simple mineralogy, such as quartz only or quartz and calcite.

Even for a single major vein generation considerable variability is observed in the proportional relationships between the smallest dimension (the "width"), the intermediate dimension (not appropriate for sheet-like veins), and the largest dimension. For veins parallel to an s-surface, the smaller dimension is generally perpendicular to the s-surface. The largest dimension in many cases is roughly parallel to hinges of folds related to the s-surface, but shear folding appears to be the dominant fold mechanism and there is not always a clearcut relationship between one of the larger vein dimensions and the fold hinges. Vein widths may be as small as 1-2 mm or as large as 1 m or more. The other two dimensions have similar large variability and for the larger veins both may extend beyond the limits of the exposed rock. The older vein generations are not uncommonly highly deformed and consequently their original size, form, and orientation are difficult to determine.

Some veins of each recognized generation can be shown to be discontinuous in extent on the scale of a few tens of meters or less. Discontinuous veins are petrologically and texturally the same as veins of the same generation that cannot be shown to be discontinuous because the outcrops are small or the veins very large. Veins of the major generations appear to be localized, isolated features that grew in discontinuous tension fractures rather than elements in a large-scale hydrothermal network.

The mechanics of the formation of metamorphic veins and tension fractures in general under high confining pressures have been discussed by a number of authors (for example, Secor, 1969; Beach 1975, 1977; Etheridge, 1983). The tension fractures in which the veins grow cannot form when the differential stress exceeds some upper limit. Etheridge (1983) has estimated the upper limit to be not more than 400 bars and probably closer to 100 bars for typical regional metamorphic rocks. Therefore high relative fluid pressure, close to the confining pressure, was required for vein initiation and growth.

The designation of the structural elements used here is not the standard system of S_1 , S_2 , F_1 , and so on, in part because of the complications presented by the major unconformity in the middle of the study area between Ordovician and Silurian units. S_1 on one side of the unconformity is not equivalent to S_1 on the other side. Also, in the standard system different types of structural elements with the same subscript have no implied genetic relationship so that in dealing with elements interpreted to be cogenetic, the discussion can get quite complex. For the description that follows, letters are used as subscripts and those elements with the same subscript are, based on geometric relationships, interpreted to be cogenetic or at least to all occupy the same position in relative time as determined by their position in the structural sequence of superimposed elements. The oldest group of elements are subscripted "a", the next oldest "b", and so on. This usage will differ from the standard system if any group of cogenetic elements does not contain all types of structural elements.

Examples from Stops in Cambrian and Ordovician rocks

Stops 1 through 4 (Fig. 1 and Itinerary) are in pelitic schists of the Hazens Notch and Underhill formations, Cambrian in age, on the east limb of the Green Mountain anticlinorium. All lie on the high-grade side of the garnet isograd mapped by Christman and Secor (1961). Stops 5 and 9 are outcrops of amphibolite of the Stowe formation, Ordovician in age, from the N-S trending area of garnet-grade through sillimanite-muscovite-grade rocks that coincides with the Worcester Mountains (Cady, 1956; Albee, 1957, 1968; Anderson, unpub. data).

Stops 1-4: Structural Sequence

The structural sequence at Stop 1 is a good starting point for discussion. The oldest secondary s-surface is a highly deformed schistosity, S_a , that at some locations is parallel to original compositional layering (a good example is at Stop 7). At Stop 1 the primary layering has been obscured by metamorphism. S_a is not related to any small folds observed in the area. Parallel to S_a are metamorphic veins that are generally not more than about 10 cm wide and of indeterminate larger dimensions because of strong deformation. These veins, V_a , are isoclinally folded by large

and small tight folds to which the predominant schistosity, S_b , in the pelitic schists is axial-planar. The isoclinal folds, F_b , have hinges that generally trend E-W within S_b , but S_b and F_b are deformed by younger folding so that the orientations of S_b and hinges of F_b are highly variable. Parallel to S_b and therefore axial planar to F_b are metamorphic veins, V_b . Superimposed on S_b , F_b , and V_b are asymmetric folds with N-S axial trends. These are F_c folds and have associated with them an axial-plane slip cleavage, S_c , for which the degree of development varies considerably in the outcrop. In this part of the study area, S_c slip cleavage is generally better developed on the short limbs of the asymmetric F_c folds than on the long limbs, but the rock type is also a factor. Parallel to S_c slip cleavage and commonly within the axial regions of small F_c folds are lens-shaped veins, V_c . The small F_c folds appear to be related to the larger N-S fold structures of the Green Mountain anticlinorium in this area (I avoid the unfortunate term "Green Mountain folds").

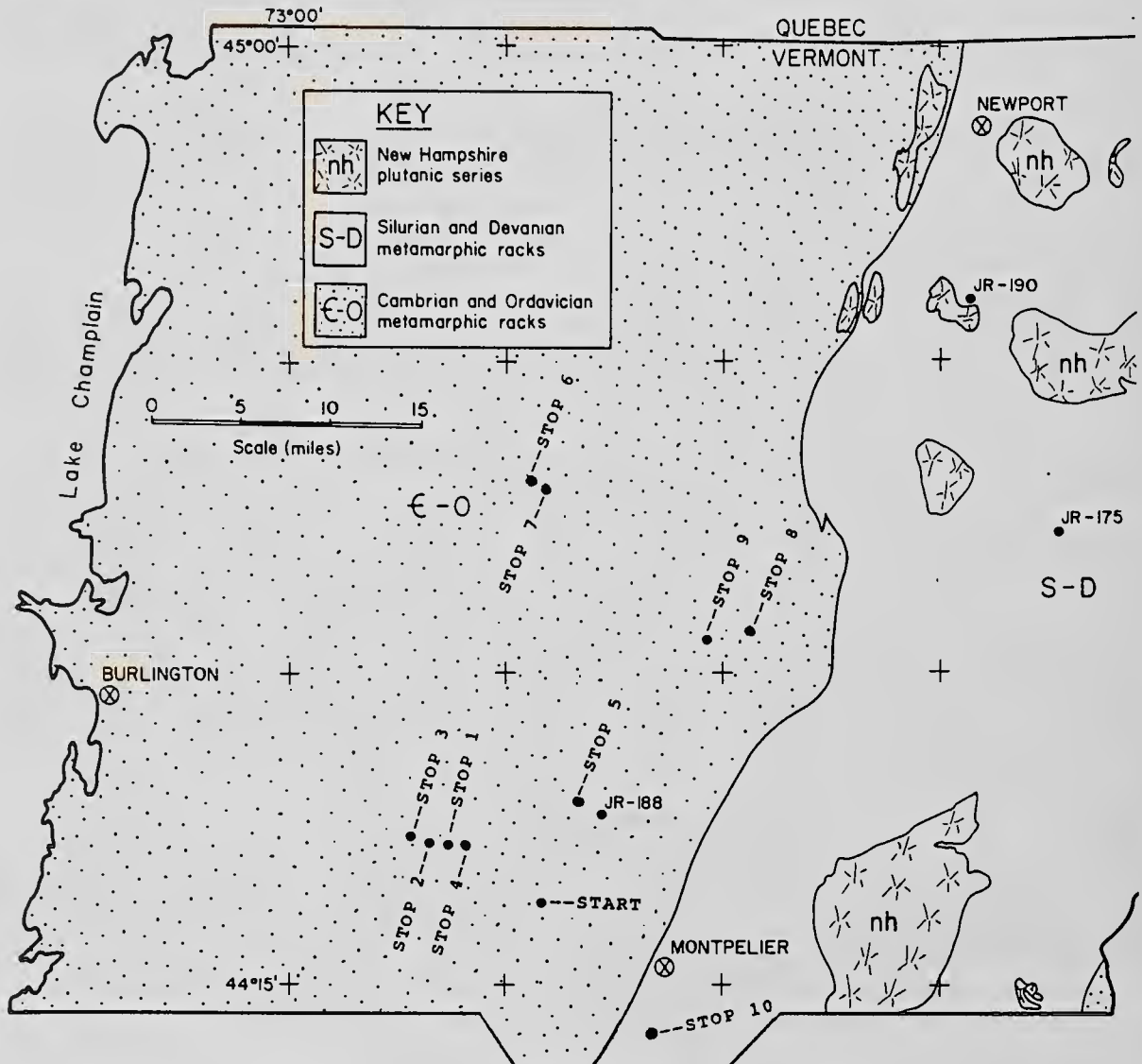


Figure 1. Sketch map of the study area with stops indicated.

S_c cleavage and associated V_c veins are gently folded by open, asymmetric folds (F_d) with N-S to NNE-SSW axial trends. F_d folds are not visibly developed in all parts of the outcrop and vary in development at nearby outcrops. At outcrops where the folds are better developed than at Stop 1, a slip cleavage, S_d , is parallel to F_d axial planes. Where present, S_d is superimposed on the nearly

ubiquitous S_c slip cleavage. Unlike the other s-surfaces, S_d does not have any associated parallel metamorphic veins. The absence of a vein set in this position of the structural sequence is consistent throughout the studied area of Cambrian and Ordovician rocks in northern Vermont.

A fourth major vein generation, designated V_e , is found at Stop 3 and other nearby outcrops. V_e veins crosscut all of the structural elements described above. V_e veins are not associated with any s-surface or fold set but have mineral assemblages within them that are subsets of assemblages of late retrograde minerals in the host rocks. The structural sequence for Stops 1-4 (also applicable to 5-9) is summarized in part A of Table 1.

The superposition relationships among major vein generations is as clear as those among other structural elements. The possibility that a vein parallel to an older s-surface, say V_b parallel to S_b , could have formed at the same time as or later than a superimposed s-surface, say S_c , is ruled out by the way in which the superimposed s-surfaces and folds affect the veins. All the fold generations appear to have formed by mechanisms that mainly involved shear folding, so the presence of tension fractures parallel to folded, pre-existing s-surfaces at high angles to the axial surfaces is much less likely than would be the case if flexure folding were important.

Vein and host rock mineral growth at Stop 1 (location JR-5)

Hazens Notch schist at Stop 1 has the assemblage quartz, plagioclase, muscovite, chlorite, epidote, apatite, ilmenite, pyrrhotite, and zircon with or without garnet, calcite, chalcopryrite, tourmaline, and/or pyrite. Garnet is replaced to varying extent by randomly oriented grains of chlorite with minor muscovite, most extensively in samples with best-developed S_c slip cleavage. Chlorite and muscovite grains with preferred orientation parallel to S_c are in several samples. Some of the grains have been rotated into this position by microfolding of the S_b schistosity and some appear to have grown in this position, crosscutting the microfolded schistosity.

Veins of the three different generations have assemblages that are subsets of the assemblages of adjacent host rocks. Important minerals in all three are quartz, plagioclase, chlorite, and minor muscovite. One or more of the following may also be present in generally minor amounts: ilmenite, apatite, pyrrhotite, calcite, and chalcopryrite. In V_b and V_c veins, secondary pyrite after pyrrhotite and secondary muscovite and paragonite after plagioclase occur in minor but varying amounts. The less deformed veins of the V_b and V_c generations generally have very coarse mineral grains, up to 10 cm or more in greatest dimension, and lack the strong directional fabric that typifies the schist. Many of the coarse grains of vein plagioclase are euhedral to subhedral, in contrast to anhedral porphyroblasts in the schist. Vein chlorite grains are larger than grains in the schist and lack preferred orientation; many are fan-like or accordian-like in form. By comparison to the younger veins, the highly deformed V_a veins appear to have undergone substantial recrystallization and grain-size reduction. Grain sizes and forms in V_a veins are more comparable to those in the schist and also the platy minerals not uncommonly have preferred orientation parallel to S_b .

Samples from a V_b vein - JR-5-A & JR-5-B

Sample JR-5-A includes the edge of the V_b vein and adjacent schist, whereas JR-5-B is from the center of the vein. The vein assemblage is quartz, plagioclase, chlorite, muscovite, calcite, ilmenite, apatite, pyrrhotite partly replaced by pyrite, and chalcopryrite. The contact between the vein and schist is relatively sharp and is grossly parallel to S_b schistosity. The adjacent schist has well-developed S_b schistosity but lacks S_c slip cleavage. The schist has an assemblage that includes the vein minerals plus zircon and tourmaline. Modal abundances of the major minerals are very different in the vein and adjacent schist.

Typical plagioclase porphyroblasts in the host schist in JR-5-A are anhedral and elongate parallel to S_b in contrast to the euhedral and subhedral porphyroblasts in the adjacent V_b vein. Schist porphyroblasts generally have three compositionally and texturally distinct zones; an example of the compositional variation in one grain is shown Fig. 2a. The innermost core is nearly pure albite ($An_{0.2}$ to $An_{1.1}$), next is an outer core of albite with compositions between $An_{1.5}$ and $An_{3.9}$, surrounded by a sharp compositional and optical discontinuity, followed by outer rim with compositions between $An_{21.5}$ and $An_{30.7}$. In some grains the rims and outer cores contain inclusion trains that are sigmoidal in pattern, suggesting possible rotation during growth. The overall compositional zoning of the grains is concentric and the anorthite content increases from core to rim with a reversal in trend near the edges. The optical and compositional discontinuity present in all porphyroblasts is parallel to the concentric zoning pattern.

In the V_b vein, plagioclase has calcic cores with zoning toward less calcic rims (Fig. 2b). Several grains in JR-5-B show increasing anorthite content from cores of about An_{27} to about An_{34} a few tenths of a millimeter away, followed by a reversal and a trend of decreasing anorthite to rims of about An_{15} . Some grains have thin albite rims, $An_{1.6}$ to $An_{2.1}$, separated from grain interiors by optical discontinuities. Such albite rims are found on grains from the center of the vein, but not on grains from the vein edge.

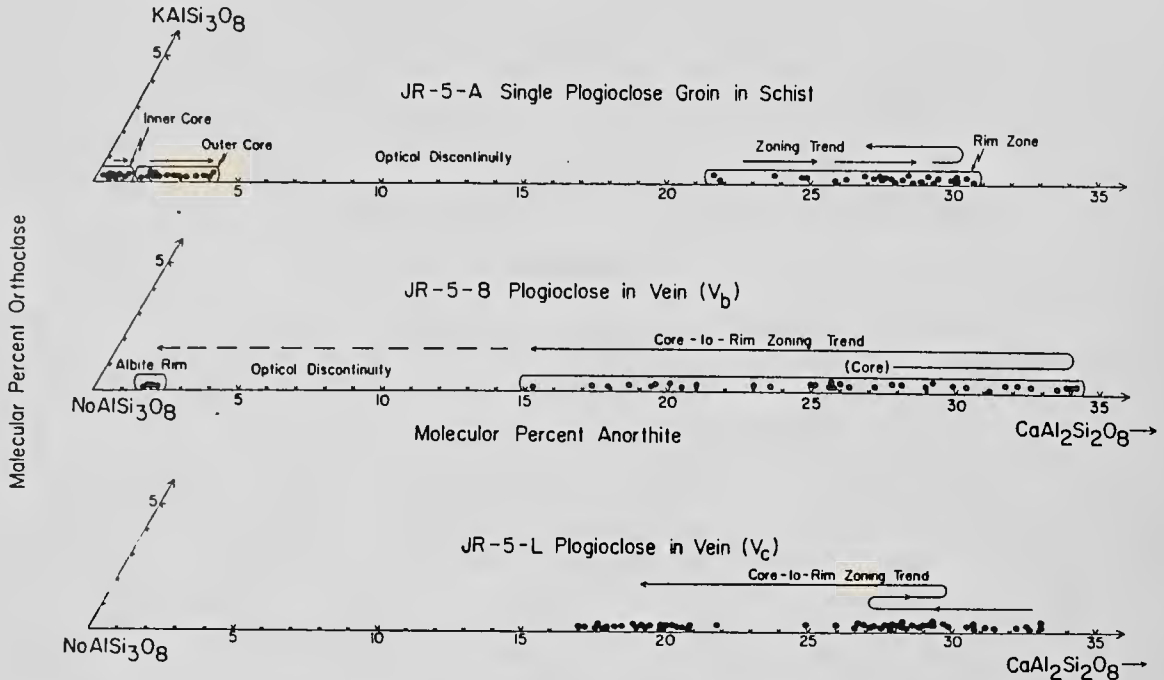


Figure 2. Compositional plots of plagioclase of (a) a single schist porphyroblast in JR-5-A, (b) all analyzed grains in JR-5-B from a V_b vein, and (c) all analyzed grains in JR-5-L from a V_c vein.

Chlorite in both the schist and vein of JR-5-A has a wide compositional range. The host rock chlorite points fall into two compositional groups (Fig. 3a) that correspond to two different textural settings. Points designated as M_a are from small chlorite inclusions within the inner albite cores of the plagioclase porphyroblasts. The compositions shown as M_b come from chlorite grains in the matrix around plagioclase porphyroblasts and from included chlorite in the outer albite cores and oligoclase-andesine rims. Analyzed points in V_b vein chlorite (Fig. 3b) have extensive compositional overlap with M_b chlorite in the schist.

Relatively coarse grains of major minerals are present in many samples along the host-vein interface. Coarse mineral growth or recrystallization that occurred in the original host rock immediately adjacent to the tension fracture is termed here to be in the "vein margin". Some of the vein margin grains also partially extend into the tension fracture. Vein margin growth is texturally different than growth in what is termed here a vein "border zone". Border zones are concentrations of certain minerals (variable from case to case) that have grown within the tension fracture and abut the host-vein interface. It is common for Al-rich minerals to be concentrated in border zones. A moderately well-developed border zone is present in the V_c vein discussed at Stop 1. Coarse euhedral to subhedral plagioclase grains, with minor chlorite, are concentrated in a relatively narrow zone along the host-vein interface. Many veins do not have any border zone developed and in those that do the zone is not necessarily developed everywhere along the host-vein interface. The presence of minor minerals like zircon, graphite, and others in the host rock and their absence in the vein help to locate the host-vein interface.

The most likely mechanism for the zoning in the schist plagioclase was the progressive reaction of epidote with plagioclase to make more Ca-rich plagioclase over a period of slowly increasing metamorphic grade. The apparent change in grade was probably due to increasing T (with presumably increasing P_{load}), but the same effect could have been produced by decreasing

TABLE 1. Structural elements in the study area
(equivalent elements line up horizontally)

A. Ordovician and Cambrian rocks (north-central Vermont)				B. Silurian and Devonian rocks (northeastern Vermont)				C. Structural element designation of Anderson (1977a)				
S-surfaces	Folds	Veins	Mineral growth	S-surfaces	Folds	Veins	Mineral growth	S-surfaces	Folds	Veins	Mineral growth	Approx. age, m.y.
S _a	--	V _a	M _a	--	--	--	--	OS _a	OF _a	OV _a	OM _a	
S _b	F _b (E-W)	V _b	M _b	--	--	--	--	OS _b	OF _b	OV _b	OM _b	440-460 *
S _c	F _c (N-S)	V _c	M _c	S _a	F _a (N-S to NNE-SSW)	V _a	M _a	DS _a	DF _a	DV _a	DM _a	380 **
S _d	F _d (N-S to NNE-SSW)	--	M _d (minor)	S _b	F _b (NNE-SSW)	--	M _b	DS _b	DF _b	--	DM _b	
--	--	V _e	M _e	--	--	V _c	M _c	--	--	DV _c	DM _c	350-360 ***
--	--	--	--	--	F _d (NNE-SSW)	--	--	--	DF _d	--	--	

* Lanphere and Albee (1974) and unpubl. data of M. Lanphere

** Estimate of Laird and Albee (1981b) for D1 using data of Naylor (1971) and Harper (1968) and unpubl. data of M. Lanphere

*** Unpubl. data of M. Lanphere on V_c veins

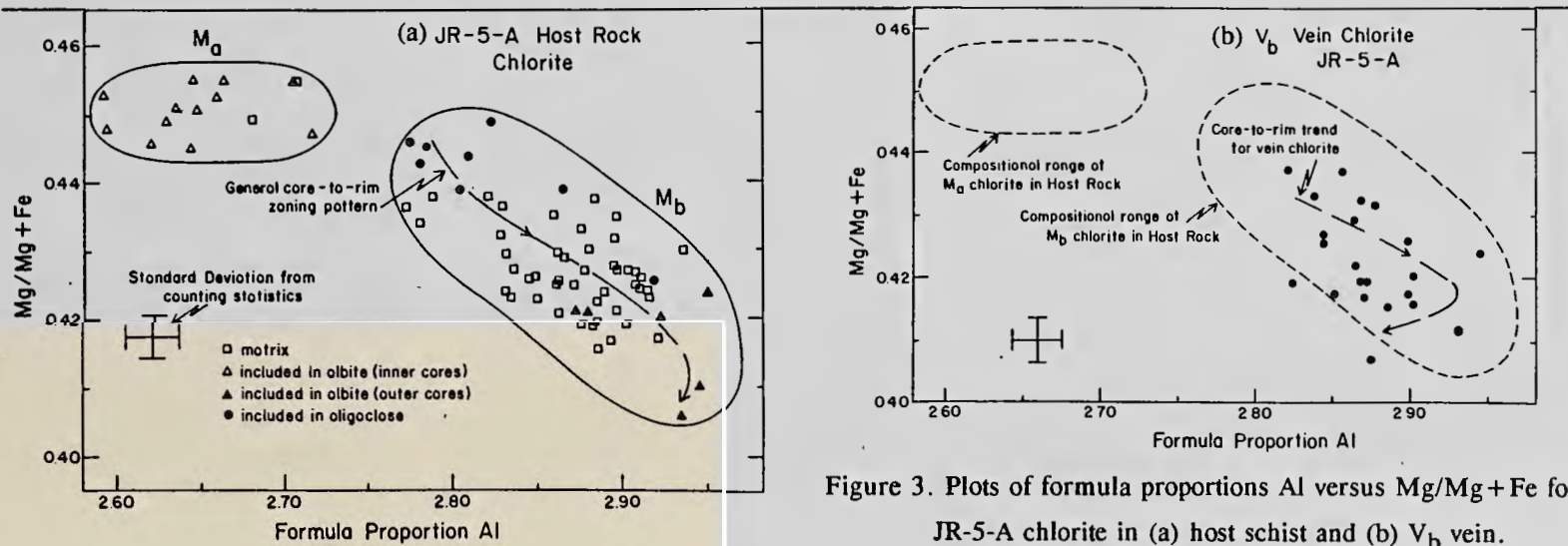


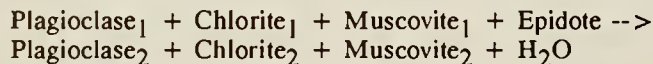
Figure 3. Plots of formula proportions Al versus Mg/Mg+Fe for JR-5-A chlorite in (a) host schist and (b) V_b vein.

P_{H_2O} either by decreasing the total fluid pressure or by changing the fluid composition. Calcite might also have been a reactant, in which case P_{CO_2} would have been important. Several reactions generally similar to those that may have occurred in JR-5-A have been discussed by Crawford (1966).

The reversed pattern in the vein plagioclase indicates that grade was decreasing as the V_b vein formed. Because the formation of V_b veins in tension fractures required high relative fluid pressure, decreasing the relative fluid pressure is not a possible mechanism for the plagioclase compositional zoning. Also the veins have abundant small fluid inclusions (diameters generally less than 15 microns). If the plagioclase zoning was due to changing fluid composition, then the exact opposite change in fluid composition from that for the schist would have been needed for the vein. Changing T and P_{total} seem most likely. The optical and compositional discontinuities in plagioclase porphyroblasts seem to represent the peristerite gap, encountered as changing physical conditions shifted the stable composition of reaction products.

The plagioclase rims in the host rock are compositionally similar to plagioclase cores in the V_b vein. Considering the structural setting of the mineral growth in the schist and in the vein, schist mineral growth associated with the S_b schistosity probably began to form before the mineral growth in the vein parallel to S_b . The vein has no evidence of post-formation recrystallization or strain due to the deformation that produced S_b ; S_b began to form before the V_b vein. The schist plagioclase zoning indicates growth during increasing grade up to the "peak" of the metamorphic event. The reversal in trend and slight decrease in anorthite content as the rims are approached may indicate some late growth as grade decreased from the metamorphic peak. When the system approached the highest grade conditions for this metamorphic event, the V_b vein began to form in a tension fracture parallel to S_b . Plagioclase in the vein grew as the grade conditions reached and then declined from the metamorphic peak. Growth of vein and schist plagioclase probably partially overlapped in time near the peak.

In simplified form, the reaction that accompanied mineral growth in JR-5-A schist might have been:



Quartz and minor amounts of other minerals needed to balance the reaction could be added to either side. Compared to their respective reactant compositions, product plagioclase was more calcic, product chlorite lower in $Mg/Mg+Fe$ and higher in $Al/Al+Si$, and product muscovite lower in phengite component and $Mg/Mg+Fe$ for the period of prograde growth.

The schist chlorite denoted as M_a is interpreted as having been formed with the S_a schistosity, predating the isoclinal F_b folds, although evidence available from other outcrops would have to be discussed to substantiate it. M_a chlorite is included in the inner zones of plagioclase porphyroblasts where inclusion trains, if present, are straight, whereas the inclusions of M_b chlorite are in portions of the porphyroblasts that in some cases have sigmoidal trains. This is compatible with the M_a chlorite forming prior to the isoclinal F_b folding and the M_b chlorite forming during F_b folding with the deformation causing the plagioclase porphyroblasts to rotate as they grew.

Samples from a V_c vein - JR-5-G and JR-5-L

Samples JR-5-G and JR-5-L are from the contact between schist and a V_c vein. The vein is parallel to S_c slip cleavage in the schist and lies in the axial region of a small F_c synform. The assemblage in the schist is essentially the same as that for JR-5-A except that JR-5-G and JR-5-L have minor garnet that is partially replaced by chlorite and muscovite. The assemblage in the V_c vein includes quartz, plagioclase, muscovite, chlorite, calcite, apatite, and pyrrhotite.

Plagioclase grains in JR-5-G schist have more complex compositional zoning than those in JR-5-A schist. The same type of zoning is found in sample JR-5-I, described later. In the V_c vein, the subhedral to euhedral plagioclase grains are concentrically zoned from calcic cores to more sodic rims, with compositions in the range $An_{33.1}$ to $An_{14.8}$ for JR-5-G and $An_{33.0}$ to $An_{16.9}$ for JR-5-L (Fig. 2c). The core-to-rim zoning trends in plagioclase are the same in both samples. The overall trend is of decreasing anorthite content from the calcic cores, but with two small reversals in trend between An_{27} and about An_{30} . Albite has not been found in the analyzed V_c vein samples except as a late alteration product.

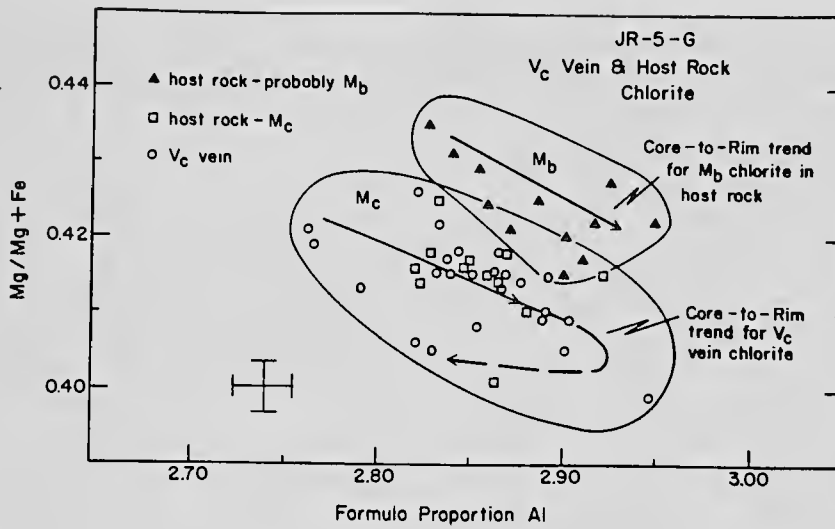


Figure 4. Plot of formula proportions Al versus Mg/Mg+Fe for chlorite in V_c vein and schist of JR-5-G.

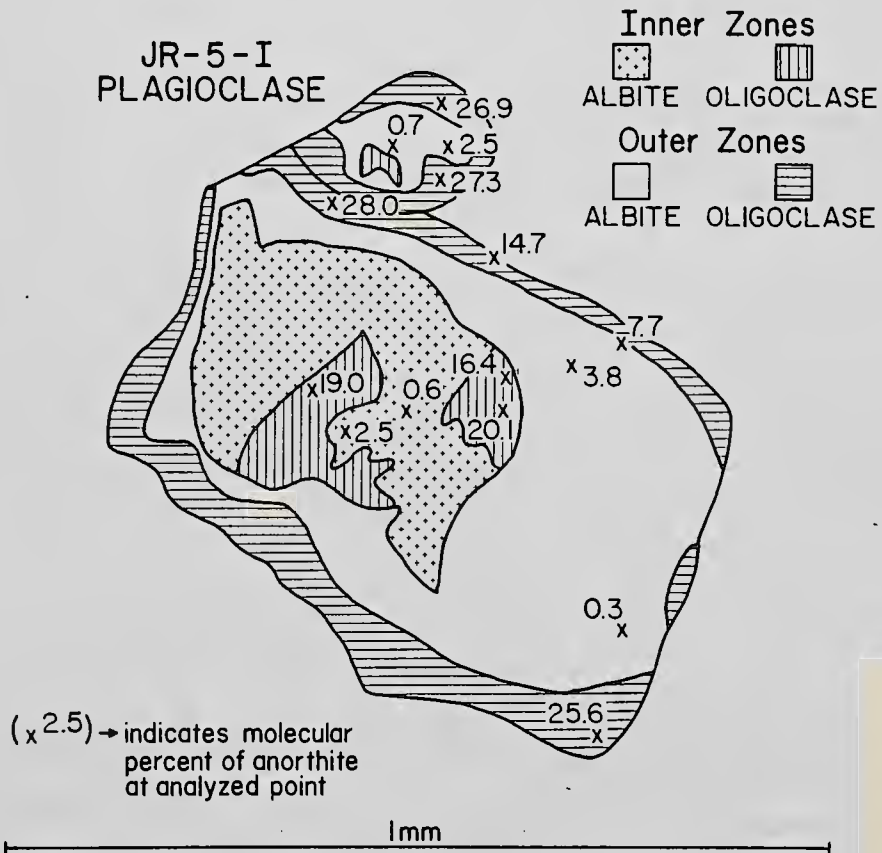


Figure 5. Compositional map of plagioclase grain in JR-5-I schist. Grain map taken from electron beam scanning image of Ca distribution, augmented with conventional quantitative analyses at points indicated with "x".

Chlorite in the schist of JR-5-G can be texturally and compositionally separated into two types (Fig. 4). One type includes chlorite grains with preferred orientation parallel to S_C slip cleavage plus some otherwise similar grains that lack preferred orientation, including chlorite that has partially replaced garnet. These chlorite grains all have similar compositions and are interpreted as representing the same generation of mineral growth in the schist. A second type consists of grains with preferred orientation parallel to S_B schistosity and with a compositional range that is different from the first group. Also plotted on the figure are points from V_C vein chlorite.

The V_C vein plagioclase is similar in compositional range and overall zoning trend to V_B vein plagioclase in JR-5-A and JR-5-B, but the V_C vein is younger than the V_B vein by structural superposition. If the zoning pattern of decreasing anorthite content in the assemblage present indicates decreasing grade, as it is interpreted, then V_C formed in a period of decreasing grade following the peak of a metamorphic event that occurred after the formation of V_B veins. Presence of oligoclase in two generations of veins suggests that two generations of oligoclase might be in the schist.

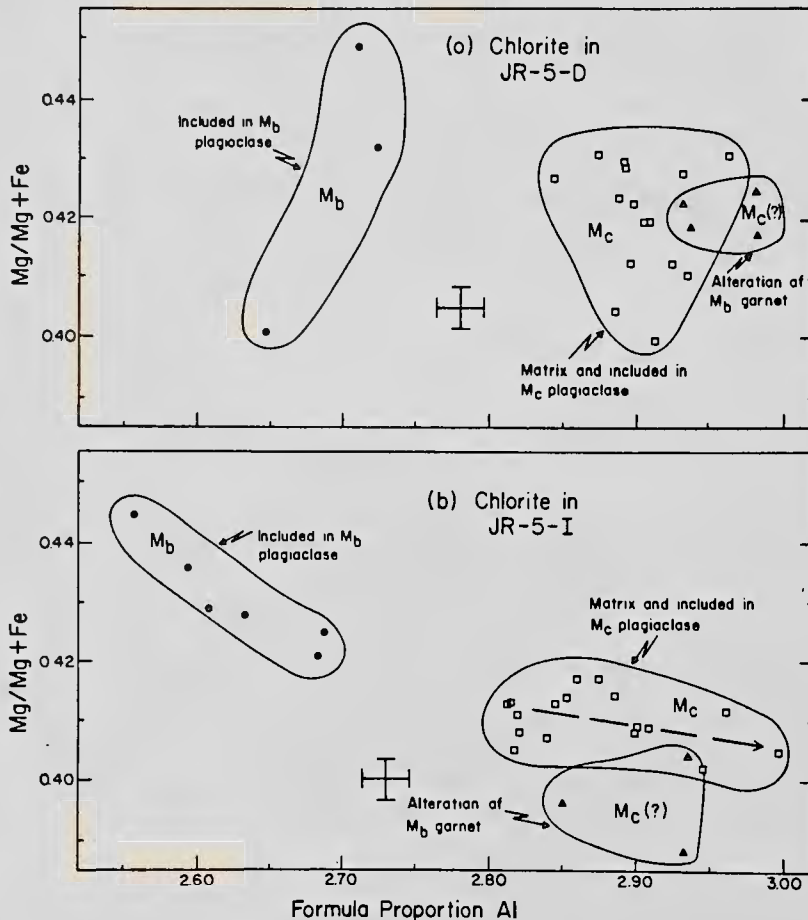


Figure 6. Compositional plots for schist chlorite in (a) JR-5-D and (b) JR-5-I.

Schist samples away from veins - JR-5-I and JR-5-D

JR-5-I and JR-5-D are samples of schist not adjacent to any veins and in which S_C slip cleavage is well developed. Assemblages in these two samples are the same as that in JR-5-G schist.

A compositional map of a plagioclase grain in JR-5-I (Fig. 5) shows the typical compositional zonation of plagioclase in this and similar samples. The mapped grain has a core, cloudy with inclusions of very fine-grained graphite and other minerals, that consists of an area of albite and two areas of oligoclase. The cloudy zone is completely surrounded by a zone of clear, graphite inclusion-free albite, in turn surrounded by an outer zone of clear oligoclase. The zoning pattern in the outer, clear zones is increasing anorthite content from core to rim. The clear albite zone is

separated from the clear oligoclase zone by an optical and compositional discontinuity that is parallel to the concentric zoning pattern in the clear zones. The boundary between the clear albite and cloudy inner zones is not parallel to any compositional zoning patterns of the cloudy interior.

Figures 6a and 6b show analyzed points in chlorite from JR-5-I and JR-5-D. Chlorite grains included in the cloudy cores of the complex plagioclase grains are compositionally distinct from chlorite grains included in the clear outer zones. Very few grains in the matrix of the schist of either sample have preferred orientation parallel to S_b schistosity, but analyzed grains that do are compositionally identical to matrix chlorite grains with preferred orientation parallel to S_c slip cleavage or no preferred orientation. The grains parallel to S_c or without preferred orientation compositionally overlap with the grains included in the clear outer zones of the plagioclase grains. Also in both samples is chlorite without preferred orientation that has partially replaced garnet; analyses from this chlorite, shown with a separate symbol, plot very close to the chlorite included in the clear zones of plagioclase.

Using the evidence all the JR-5 samples, an interpretation of the plagioclase grain in Figure 5 is that the cloudy cores represent M_b growth, or perhaps a combination of M_a albite growth and M_b albite through oligoclase (sodic andesine in some) growth. The clear rims are M_c growth and have the same relationship to the V_c plagioclase that the M_b plagioclase in the simple plagioclase grains of JR-5-A have to M_b growth in the V_b vein. Two generations of oligoclase to sodic andesine are present in the schist and similar plagioclase with reversed zoning patterns is found in two distinctly different generations of veins.

Stop 2 (location JR-73)

Discussed below are analyzed garnet in two samples from Stop 2, a V_b vein (JR-73-B) and schist (JR-73-A) about 1 meter from the V_b vein. Typical pelitic schist at JR-73 has an assemblage that includes quartz, plagioclase, muscovite, chlorite, garnet, magnetite, ilmenite, epidote, pyrite, apatite, tourmaline, and zircon with or without calcite and/or chalcopyrite. Included within garnet porphyroblasts of JR-73-A are grains of chloritoid, although none of the studied samples have chloritoid grains in the matrix. Schist in JR-73-B adjacent to the V_b vein lacks chloritoid inclusions in garnet. The V_b vein assemblage includes quartz, plagioclase, chlorite, muscovite, calcite, ilmenite, and garnet. In this part of the study area, only veins of the the V_b generation have garnet within them. Garnet in the schist is probably M_b in generation based on textural, assemblage, chemical evidence. As at Stop 1, retrogradation of garnet is most extensive in samples with well-developed S_c slip cleavage.

Rim-to-rim profiles of vein garnet grains in JR-73-B (Fig. 7a) are different than rim-to-rim profiles of garnet grains in JR-73-A schist (Fig. 7b). All of the grains are concentrically zoned with respect Fe, Mn, Mg, and Ca) as illustrated by the map of Mn/Mn+Fe+Mg+Ca for grain A (Fig. 8).

The inner parts of the vein garnet grains are similar to typical garnet porphyroblasts in host pelitic schist in this part of the study area. Mn is high in the center and decreases toward the rim in this center zone, whereas Fe increases core-to-rim. Mg is generally in low concentration and shows a gradual increase as Mn decreases and Fe increases. However, only garnet grains in or adjacent to V_b veins have the sharp, inner reversal of Mn in which the proportion of spessartine increases by as much as 10% of the total solid solution over distances of only a few tens of microns. The increase in Mn is accompanied by a sharp decrease in Fe and a smaller decrease in Mg. The outer reversal involves a return to the "normal" zoning trend for Mn, Fe, and Mg, but the values do not generally approach the maximum Fe and Mg and minimum Mn values found at the inner reversal. How the "average" trend of Ca might be correlated with the overall trends for the other elements is obscured by the small scale Ca variations.

Grain D (Fig. 7b) is typical of garnet in Underhill and Hazens Notch schist in this area. Mn is highest in the center and decreases toward the rim, whereas Fe is the opposite. Mg is low in concentration and gradually increases as Mn decreases. Ca zoning has an erratic pattern, with sharp changes in Ca mirrored by opposite changes in Fe and Mn, but no change in Mg.

Other minerals such as plagioclase, chlorite, muscovite, ilmenite, calcite, and chloritoid are also compositionally zoned in these two samples. Besides garnet, the only other major mineral with significant variation in Ca is plagioclase, which has zoning patterns similar to those for M_b schist and vein plagioclase at Stop 1. Calcite grains are zoned with respect to Fe-Mg-Mn substitution for Ca, but the range of Ca variation is limited. Epidote grains are zoned with respect to Fe and Al but not Ca, except in small REE-rich cores in some grains where Ca varies due to the REE substitution

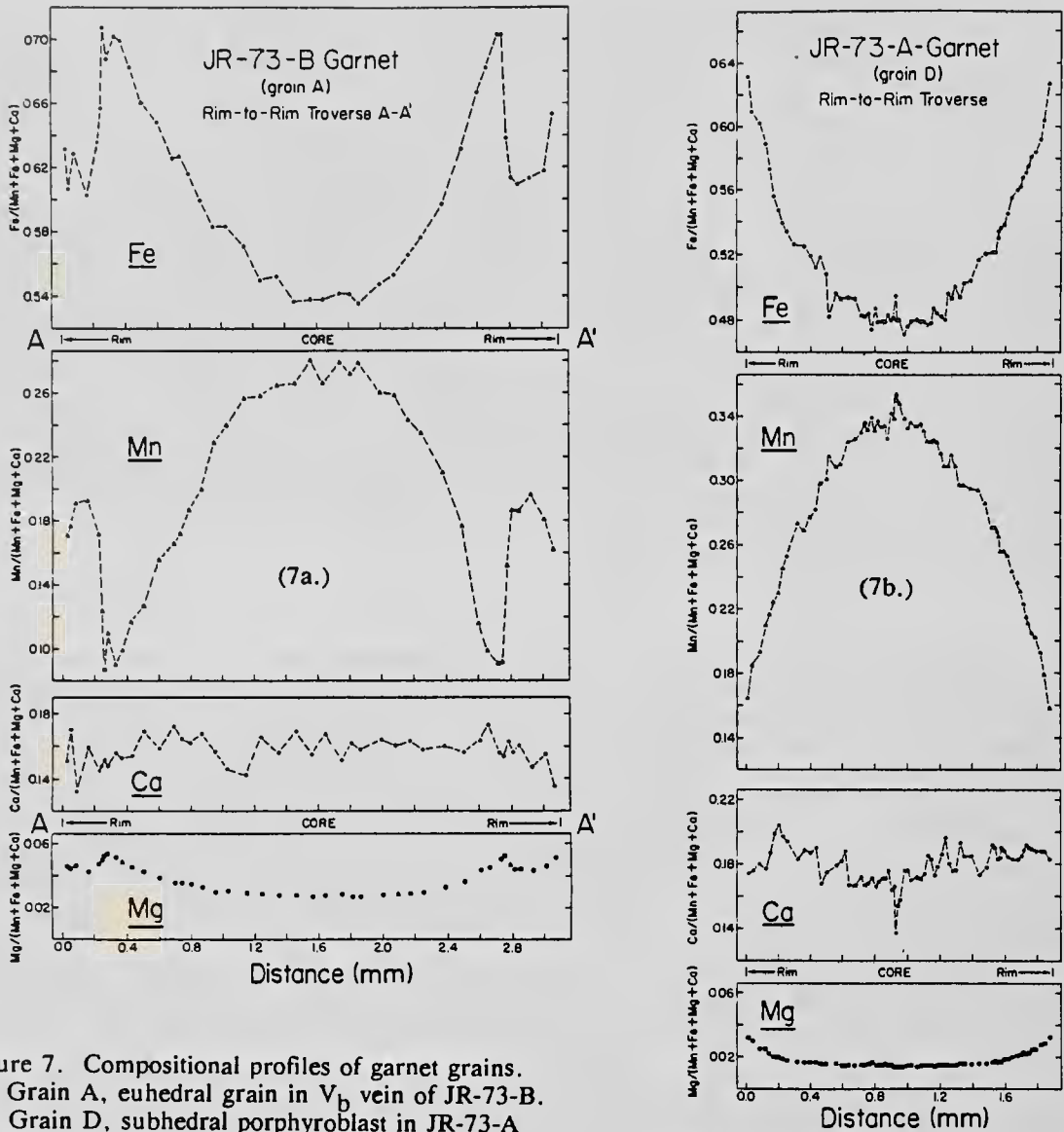
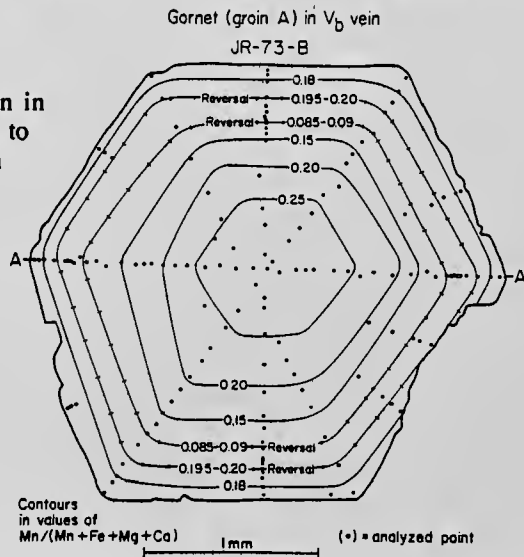


Figure 7. Compositional profiles of garnet grains.
 (a) Grain A, euhedral grain in V_b vein of JR-73-B.
 (b) Grain D, subhedral porphyroblast in JR-73-A schist.

Figure 8. Compositional map of garnet grain in V_b vein of JR-73-B, contoured with respect to $Mn/(Mn+Fe+Mg+Ca)$. Traverse in Fig. 7a is shown as A-A'.



(REE-rich cores have up to 7 wt. % $Y_2O_3 + Ce_2O_3 + La_2O_3 + Nd_2O_3$). Mn-bearing phases other than garnet are ilmenite, chlorite, calcite, and chloritoid.

Textural evidence indicates chloritoid was once present in the matrix of some of the schist but was removed by reaction. Chloritoid grains included in garnet have Mg/Mg+Fe values in the range 0.08 to 0.11 and MnO contents in the range 1.47 to 1.87 wt. %. Ilmenite grains in JR-73-A have MnO contents in the range 1.69 to 2.59%, but there is no textural evidence of substantial reaction of ilmenite. V_b vein calcite have MnO contents in the range 0.24-1.67%.

JR-73-A schist and JR-73-B vein chlorite grains have MnO contents of 0.27-0.51 and 0.24-0.45%, respectively, along with wide compositional ranges with regard to Mg, Fe, Si, and Al. Mn variation in vein chlorite is grossly similar to that in the vein garnet, except much lower concentrations are involved. Cores tend to be high in Mn, which decreases toward the rims, then increases and, on the rims of some grains, decreases again.

Neglecting the fine-scale details, the best explanation for the zoning trends of garnet and other M_b minerals in the schist and V_b vein of JR-73 is shifting equilibrium partitioning among phases as conditions first increased in grade, then decreased. The vein and schist M_b growth probably overlapped in time, but the schist growth began first and the last growth was in the V_b vein. The relatively late breakdown of Mn-bearing chloritoid in the schist may help explain the Mn "spike" in vein garnet and possible smaller counterpart in vein chlorite.

Stowe amphibolite in the Worcester Mountains - Stops 5 & 9

Exposed at both locations are amphibolites metamorphosed to at least garnet grade and then later subjected to retrogradation at chlorite to biotite grade. This is typical of rocks of the Worcester Mountains and in much of the area the higher-grade assemblages have been badly obscured or completely eliminated by retrograde events. The extent of preservation of the higher-grade assemblages and the character of the deformation elements in the Worcester Mountains tend to vary with rock types. Equivalent deformational elements in amphibolite and pelitic schist can have different styles or degrees of development. In general, the amphibolite is less deformed and less affected by the late retrograde metamorphism than pelitic schist.

The age of coarse muscovite from a V_b vein in the Worcester Mountains has been determined by Lanphere and Albee (1974) to be 439 m. y. using $^{40}Ar-^{39}Ar$ methods (the vein was shown to the author in the field by A. Albee). Other primary minerals in the vein are kyanite, garnet, and biotite, now mostly altered. Fine-grained muscovite, pseudomorphous after kyanite, in schist adjacent to the V_b vein was determined by Lanphere and Albee to be 358 m. y. in age. They also analyzed hornblende from Stowe amphibolite at another location in the Worcester Mountains and it had an age of 457 m. y.

Stop 5 (similar to JR-188, 2 miles SE on Mt. Hunger)

Sample JR-188-A is amphibolite with a small V_b vein, parallel to S_b schistosity, and a several small V_c veins parallel to S_c fracture cleavage. The amphibolite assemblage includes hornblende, plagioclase, epidote, quartz, chlorite, biotite, rutile, sphene, muscovite, and apatite. Also present is actinolite after hornblende and some of the actinolite is in bands parallel to S_c that cut across hornblende grains. Some of the chlorite, biotite, epidote, and perhaps all of the sphene are retrograde minerals formed by reaction of hornblende that has preferred orientation parallel to S_b . The V_c veins have actinolite, plagioclase, quartz, epidote, sphene, and chlorite. The V_c veins are similar to those at Stop 9. The V_b vein has amphibole grains with euhedral hornblende cores overgrown by euhedral actinolite rims. This actinolite is texturally and compositionally different (with lower Mg/Mg+Fe values) from that in V_c veins and from the actinolite formed at the expense of hornblende in the host rock. The hornblende interiors of V_b vein amphibole grains are separated from the actinolite rims by sharp optical and compositional discontinuities not unlike the discontinuities described for plagioclase.

Compositional profiles from several core-to-rim traverses across V_b vein amphibole and host rock hornblende (Fig. 9) illustrate the zoning patterns in these two amphibole types. Hornblende interiors of vein amphibole are compositionally similar to host rock amphibole, but the actinolite rims of the vein amphibole are very different.

V_b vein amphibole is similar in a general way to V_b vein plagioclase from JR-5 and JR-73. It grew as grade conditions were changing and the last growth occurred as grade was decreasing. (The compositional variation of amphibole with grade in amphibolites with the same assemblage as

JR-188-A has been discussed by Laird and Albee, 1981b.) Interiors of vein amphibole grains formed at the highest grade and their growth may have overlapped in time with the last growth of M_b host rock hornblende. Amphibole in both the V_b vein and host rock was growing when the peak of the M_b -forming event was attained. As grade conditions declined from the peak, the host rock amphibole ceased to grow but the vein amphibole continued to form in an environment suitable for maintaining euhedral grain shapes. The grade decreased enough for actinolite to become the stable amphibole composition. The discontinuity present in V_b vein amphibole grains could represent either a hiatus in growth or a miscibility gap encountered as conditions changed over time.

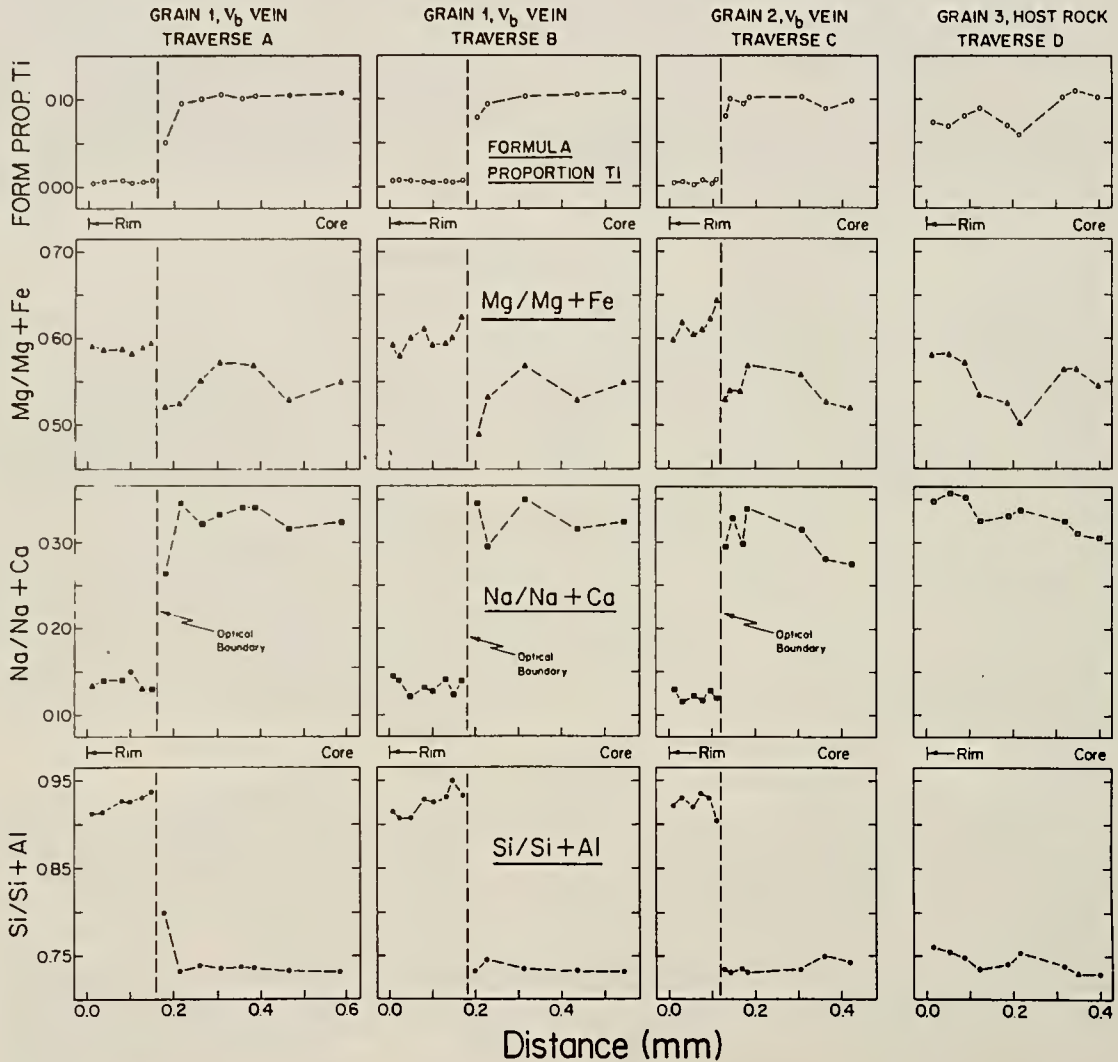


Figure 9. Core-to-rim compositional profiles with respect to $Si/Si + Al$, $Na/Na + Ca$, $Mg/Mg + Fe$, and formula proportions Ti for two amphibole grains in V_b vein and one amphibole grain in host amphibolite, sample JR-188-A.

Stop 9 (location JR-66)

Sample JR-66-F is from a V_c vein parallel to S_c fracture cleavage in Stowe amphibolite. The exposed vein is lens-shaped, 30 cm long and 10 cm at its maximum width. The vein assemblage includes quartz, plagioclase, actinolitic amphibole, chlorite, biotite, microcline, epidote, hematite, magnetite, sphene, and calcite. The host amphibolite has a higher-grade assemblage that was

strongly overprinted by at least one retrograde event. The pre-retrograde assemblage of the host rock includes hornblende, plagioclase, epidote, chlorite, quartz, magnetite, and ilmenite. The host rock retrograde assemblage is the same as the vein primary assemblage, except that in the former microcline is lacking and none of the magnetite is of obvious retrograde origin.

Developed along the host-vein interface of the V_C vein is a 1-2 mm wide border zone, present along most but not all of the examined area of the interface. A zone of chlorite with preferred orientation parallel to the host-vein interface is developed adjacent to the interface. Toward the vein interior, the next zone is one of intergrown chlorite and biotite with the same preferred orientation as the chlorite zone. An innermost zone of biotite with lens-shaped grains of microcline is sporadically developed. Locally developed between the chlorite-biotite or biotite zone and the quartz-rich vein interior is a monomineralic zone of hematite.

Euhedral amphibole grains are distributed throughout all parts of the border zone except the hematite zone. Similar euhedral amphibole grains are in the vein interior. More than half of the amphibole grains in the border zone have preferred orientation with their direction of elongation perpendicular to the host-vein interface. This suggests that some mechanism other than directed stress was responsible for the preferred orientation in the vein border zone. None of the minerals in the vein interior have preferred orientation.

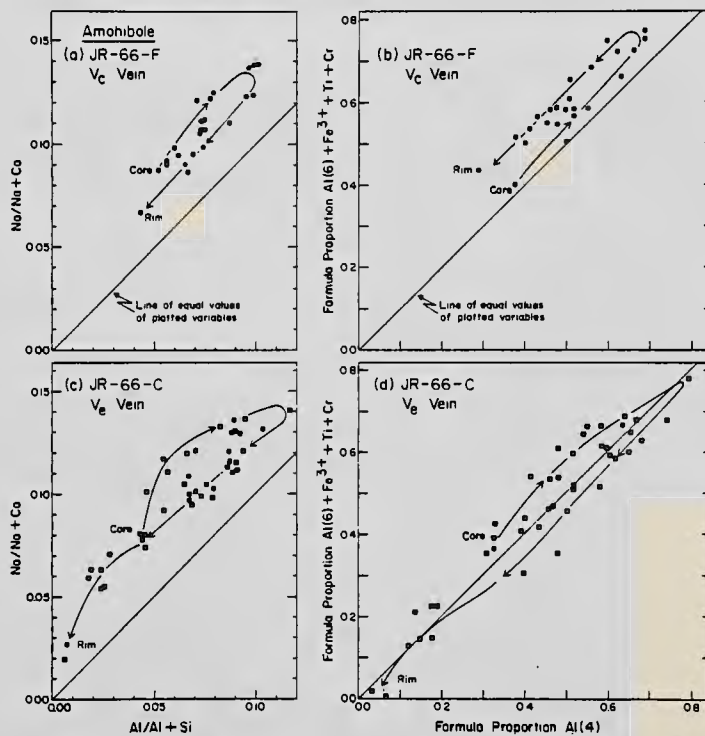


Figure 10. Plots of $Al/Al+Si$ versus $Na/Na+Ca$ and $Al(4)$ versus $Al(6)+Fe^{3+}+Ti+Cr$ for Ca-amphibole in JR-66 samples. (a, b) JR-66-F V_C vein. (c, d) JR-66-C V_e vein.

Typical amphibole grains in the border zone are concentrically zoned with respect to composition. The core-to-rim variation trend (Fig. 10a, b) starts with actinolitic cores, progressively becomes more hornblende-rich, reverses in trend and finally zones out to actinolitic rims with compositions very much like the core compositions. No discontinuities, optical or compositional, are found in the vein amphibole grains. Comparing these compositions to amphibole compositions discussed by Laird and Albee (1981b), the V_C amphibole formed at biotite-grade conditions under an intermediate relative pressure.

Within the host amphibolite of JR-66-F are three compositional ranges of amphibole. Cores of dark amphibole grains have compositions similar to the host rock amphibole of JR-188-A and V_b vein hornblende cores. The hornblende cores are overgrown by more actinolitic amphibole that is similar to the actinolitic rims of the V_b vein amphibole in JR-188-A. A third type of amphibole is present as irregular patches of actinolite that replaces the first two types. This third type compositionally overlaps with the V_C amphibole of JR-66-F. Other retrograde products such as

fine-grained chlorite and biotite are closely associated with the patchy actinolite and they overlap compositionally with their counterparts in the V_c vein as well.

Sample JR-66-C is from a V_c vein that crosscuts S_c fracture cleavage in Stowe amphibolite. The vein assemblage includes albite, calcite, actinolite-tremolite, chlorite, biotite, epidote, magnetite, hematite, sphene, apatite, and minor quartz. This vein does not have a well-developed border zone. Amphibole grains are euhedral with core-to-rim compositional trends for concentrically zoned grains as shown in Fig. 10c, d. The cores are similar to cores of V_c vein amphibole in JR-66-F. The initial trend is increasing Al(4), A-site Na, and Al(6) + Fe³⁺ + Ti + Cr toward more hornblende-rich compositions. Then a reversal in trend occurs and the grains are zoned out to tremolitic rims with very low values of Al and Na.

Although the V_c and V_e veins at Stop 9 both contain assemblages that are the same as the retrograde assemblage found in the host assemblage, the two vein types are structurally distinct in age. Therefore, two generations of retrograde mineral growth (M_c and M_e) can be postulated. Both veins began to grow as grade conditions were increasing in their respective metamorphic events and continued to form as the peak was attained and as conditions declined to lower grade. Growth in the V_e vein continued to lower grade than that attained by the V_c vein.

Stop 10 (a brief look into the Siluro-Devonian units)

The oldest prominent secondary s-surface in the Devonian rocks of this area is a schistosity, S_a , that is axial planar to rarely observed asymmetric folds, F_a , with N-S to NNE-SSW axial trends. In many cases the smallest order F_a folds have wavelengths and amplitudes of sufficiently large size that they can only be observed in the largest roadcuts. F_a folds can be seen at Stop 10 on careful observation. Typically the original layering S_0 is parallel to F_a axial planes on one limb of the asymmetric folds and non-parallel on the other limb. Parallel to S_a schistosity is a generally well-developed set of metamorphic veins, V_a .

At some locations, thin vein-like layers of quartz or quartz-calcite are present that appear to be older than S_a . Generally these are parallel to S_0 but do not seem to be related to a prominent secondary s-surface. Because they have invariably simple assemblages and have no clear relationship to other structural elements produced by deformation or to major metamorphic mineral growth, these veins are treated as a minor generation.

S_a schistosity is folded by a set of prominent small asymmetric folds, with NNE-SSW axial trends. These are F_b folds and have associated with them an axial-planar s-surface, S_b . S_b is particularly well developed in the axial region of the Willoughby arch, where it is a schistosity, but it is either a prominent slip cleavage or a schistosity in other parts of the post-Ordovician sequence of northeastern Vermont. F_b and S_b are equivalent to the minor elements interpreted by Dennis (1956) and Hall (1959) to be related to formation of the Willoughby arch.

No veins parallel to S_b or with any apparent genetic relationship to F_b have been found by the author in northeastern Vermont. The lack of veins seems to be a fundamental feature of this generation of structures in the area. S_b and F_b are crosscut by a later generation of veins, V_c . V_c veins have no obvious relationship to other small-scale structural elements but do have mineral assemblages within them that are subsets of assemblages of a post- S_b mineral growth generation in the host rocks. In areas close to plutons of the New Hampshire series in northeastern Vermont, this late mineral growth generation (M_c) attained grade as high as sillimanite-muscovite. Characteristic of M_c is the random orientation of minerals that in an environment of directed stress would typically have preferred orientation. A few V_c veins have been found that crosscut semiconcordant sheets of granitic rock associated with the New Hampshire series. However, the M_c mineral growth generation has such a close spatial relationship with the plutons with regard to metamorphic grade that the veins were probably formed about the same time as the intrusion and solidification of some of the plutons.

A late generation of open, asymmetric folds with NNE-SSW axial trends and upright axial planes deform the V_c veins. These folds, F_d , are difficult to recognize without the V_c veins as markers, although they also gently deform S_c schistosity or slip cleavage. No s-surface or veins related to F_d have been found. The only other significant generation of structural features found in this part of the study area, except for minor veins, is a locally developed set of open, asymmetric folds with E-W axial trends, upright axial planes, and, where present at all, a poorly-developed axial-plane slip cleavage. These E-W folds are younger than F_a , older than F_d , and may be younger than F_b , although the interference structures (basin-and-dome structures formed on S_a surfaces) produced by the N-S F_b folds and these E-W folds are such that the superposition

relationships are ambiguous; an excellent example is at Stop 10. The E-W folds are designated F_{b+1} , have no associated veins, and are not apparently associated with any substantial mineral growth in the host rock. The structural sequence (excluding F_{b+1}) in Siluro-Devonian units is summarized in part B of Table 1. The interpreted correlation (one which agrees with the scarce age data) across the unconformity between the Cambro-Ordovician and Siluro-Devonian is shown in Table 1, Part C.

Conclusions

Metamorphic veins were formed during four of the five major metamorphic events that occurred during the Taconic and Acadian Orogenies in northern Vermont. The compositional zoning patterns of key minerals in the veins and surrounding host rocks recorded the effects of changing physical and/or chemical conditions during each event. Most of the host rock mineral growth occurred as grade was increasing to the metamorphic peak of each event, in some cases with minor continued growth as grade declined from the peak. Substantial mineral growth in the veins occurred as grade was decreasing from the peak, with varying amounts of early growth overlapping with the prograde host rock mineral growth. By the methods used, each event with associated veins is shown to be a distinct metamorphic "pulse", separated from other events by periods of low-grade conditions.

The timing of vein initiation varied from case to case as indicated by the variation in amount of overlap of host and vein mineral growth of a given generation. The distribution of stress and the relative fluid pressure in a rock volume probably were principal controlling factors that determined where and when a vein would form. The veins grew by precipitation of material in tension fractures and replacement of host rock was not an important process.

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ITINERARY

Assembly point is a picnic area on Route 2 between Middlesex and Waterbury, 2.7 miles east of the junction of Routes 100N and 2 in Waterbury and 2.6 miles west of the Middlesex exit off I-89. Picnic area is on north side of road, with parking on both sides. Meet at 8:30 A.M. Topographic maps: Waterbury, Stowe, and Middlesex 7.5" quadrangles, Hyde Park, Hardwick, Montpelier, and Barre 15" quadrangles.

PLEASE: refrain from sampling the veins from which the compositional data come, especially the smaller ones that could be whacked to oblivion.

Mileage

- 0.0 From assembly point, take Route 2 west (Middlesex quadrangle). Pass through town of Waterbury (Waterbury quadrangle) and continue west.
- 5.3 Underpass below I-89.
- 5.8 STOP 1: Large roadcut in pelitic schist of the Hazens Notch formation on right (north) side of road. Park along right side of road, but be careful as room is tight. Pull completely to the right of the solid white line that marks the edge of the travelled surface. Refer to text for detailed discussion of V_b and V_c veins here. (Sample location JR-5.)
- 6.4 Bolton - Waterbury line.
- 6.7 STOP 2: Small roadcut on north side of Route 2. Use same care in parking here and at next two stops as at Stop 1. A V_b vein with garnet in Underhill schist is located on the top of the outcrop. (Sample location JR-73.)
- 7.3 STOP 3: Fairly large roadcut in Underhill schist on north side of Route 2. A V_e vein is found here, as are good examples of V_c veins with irregular forms. (Sample location JR-163.)
- Turn around and proceed east on Route 2, back toward Waterbury.

- 8.2 Bolton - Waterbury line.
- 9.1 **STOP 4:** Park carefully along the right (south) side of Route 2 near the west end of a long, curving set of roadcuts on both sides of road. These are schists of the Hazens Notch formation just above the garnet isograd as mapped by Christman and Secor (1961). A large V_c vein is exposed on both sides of the road. (Location JR-4.)
Continue east on Route 2.
- 10.9 Junction of Routes 2 and 100N. Turn left (north) on to Route 100 toward Stowe.
- 11.0 (Middlesex quadrangle.)
- 12.0 (Stowe quadrangle.)
- 12.2 Turn right (east) on to Howard Avenue in Waterbury Center. Road is not well marked on Route 100 except for sign that points to Waterbury Center P. O. and Loomis Hill.
- 12.6 Turn left on to Maple Street after passing a crossroad at 12.5.
- 12.7 Turn right on to Loomis Hill Road.
- 13.1 Road bends to right, continue to stay on main road (paved).
- 14.6 Pavement ends, road continues as two-lane dirt road. Pass Ripley road to right and follow main road as it now travels north at the base of the Worcester Range.
- 16.0 Unmarked entrance to right (east) into an old amphibolite quarry.
- 16.1 **STOP 5:** Turn right into a parking lot marked with sign for the Waterbury trail to Mt. Hunger. Additional parking can be accessed by the unmarked entrance at mile 14.6. Follow old quarry road (and Mt. Hunger trail) northeast for about 100 feet, then veer off on one of several small unmarked paths that lead north about 50 feet to the top of the quarry. The walls of the quarry have many large loose blocks that look unstable and are best avoided. Exposed at the top is amphibolite of the Stowe formation and several generations of metamorphic veins. One mile to the east is pelitic schist with the assemblage sillimanite-kyanite-staurolite-garnet-biotite-muscovite-quartz. (Stop 5 is location JR-89, but data from JR-188 will be discussed.)
Return to parking area, turn right (north) back on to dirt road (which immediately narrows to one lane with only infrequent turnouts).
- 17.2 Barnes Hill Road comes in from left. Continue straight (north) on what has now become Stowe Hollow Road and widened back to two lanes. Follow this until it meets Gold Brook Road in Stowe Hollow.
- 19.0 Turn left (west) on to Gold Brook Road and follow it to Route 100.
- 20.2 Stop sign at Route 100. Turn right (north) on to Route 100 and follow it through Stowe.
- 23.7 Stagecoach Road forks off to left from Route 100. Snow's snack bar is just beyond fork.
Make left turn on to Stagecoach Road and follow it north toward Morristown.
- 25.0 (Hyde Park 15" quadrangle.)
- 29.2 Stop sign - continue straight.
- 30.5 Stagecoach Road ends at stop sign, near Lake Lamoille. Turn left (north) toward Hyde Park.
- 31.7 One-lane bridge as you approach Hyde Park. Proceed with caution.
- 31.8 Stop sign. Turn left on to Main Street in Hyde Park and proceed to Route 15.
- 32.3 Stop sign. Turn left on to Route 15 and go northwest past Johnson.
- 38.3 Road to Ithiel Falls forks to right off of Route 15 just before a green iron bridge over the Lamoille River. Take this right fork on to unmarked road (only sign shows direction of Long Trail).
- 39.4 **STOP 6:** Turn left into parking area on left (southwest) side of road, across from roadcut. The road is narrow and visibility is poor, so be careful crossing road to outcrop. Good examples of V_b and V_c veins in biotite-grade Hazens Notch schist are found here. (Location JR-114.)
Reverse direction of travel, go back toward Route 15.
- 39.8 **STOP 7:** Pull off into parking area on right (southwest) side of road. Walk 0.15 mile southeast and downhill along road to outcrops of the Hazens Notch formation on left (northeast) side. Present here are excellent examples of F_b isoclinal folds of S_0 , with superimposed F_c folds and V_c veins. (Location JR-113.)
Return to cars, continue back to Route 15.
- 40.5 Junction with Route 15. Go left (east) on Route 15, toward Johnson.
Continue east on Route 15 past Johnson and Hyde Park.
- 48.6 Junction of Routes 15 and 100S. Continue east on Route 15 toward Wolcott.
- 53.7 (Hardwick 15" quadrangle.)

- 55.3 To the right is a road to the town of Elmore - just before the road is a sign for "Hilltop X-country Ctr". Just beyond this road are roadcuts on both sides of Route 15.
- 55.35 STOP 8 (time allowing): Turn left into the parking area on the north side of Route 15, next to a large roadcut in the Moretown member of the Missisquoi formation. Seen here are some pull-apart structures within which have grown V_c veins. (Location JR-67.)
Reverse direction of travel on Route 15 and go short distance to road to Elmore.
- 55.4 Turn left (south) onto road to Elmore. Go across bridge over Lamoille River and follow the "main" travelled way (which is dirt most of the way and is a washboard on the steeper parts).
- 55.9 Road bends sharply to right.
- 56.8 (Hyde Park 15" quadrangle.)
- 57.4 Road bends sharply to left.
- 58.2 Stop sign at Route 12 in Elmore, Lake Elmore directly across road. Turn left (south) on to Route 12.
- 59.3 STOP 9: Turn sharply right on to small dirt road and park on either side (we will be continuing south on Route 12, so you will eventually have to turn around). Walk uphill (north) about 0.2 mile on Route 12 to a low outcrop of Stowe amphibolite partially hidden in overgrowth on right (east) side. Samples of V_c and V_e veins from here are discussed in the text. (Location JR-66.)
Go back to Route 12 and turn right to continue south toward Montpelier, passing through Worcester and Putnamville.
- 60.3 (Montpelier 15" quadrangle.)
- 71.0 Worcester.
- 73.5 Putnamville.
- 78.2 Coming into Montpelier. Route 12 is called Elm Street here. Continue straight on Elm Street until it runs into State Street.
- 79.4 Route 12 turns to left, but continue straight to avoid downtown traffic.
- 79.7 Turn right (west) on to State Street (Business Route 2), go past Capital Building to Bailey Ave. and follow signs to I-89.
- 80.1 Traffic light at Bailey Ave. Turn left on to Bailey and follow signs to I-89.
- 70.2 Traffic light at Memorial Drive. Turn right on to Memorial and work your way over into left lane so you can get onto I-89 South.
- 81.0 Get on I-89 South (toward White River Junction).
- 81.8 Roadcut with exposure of RMC on left side. (Barre 15" quadrangle.)
- 83.7 Take Exit 7 to right (Barre exit). Keep to right.
- 84.7 Traffic light at Berlin Corner. Turn right on to Paine Tpk.
- 84.95 Turn right at road junction to follow Paine Tpk.
- 85.3 Underpass below I-89.
- 85.35 Turn left on to dirt road (still marked as Paine Tpk.)
- 85.4 STOP 10: Park on right side of dirt road, next to large roadcuts on both sides. F_a and F_b folds in the Barton River member of the Waits River formation can be seen here, as well as S_a , V_a , and S_b . At the south end of the roadcut, basin-and-dome interference structures between F_b and F_{b+1} are developed on an exposed S_a surface. (Location JR-106.)
- SUMMARY - end of trip
Return in the direction from which you came.
- 85.45 Turn right on paved road.
- 85.85 Turn left at stop sign.
- 86.1 Traffic light. You have a choice of going right to get to Barre or the Barre-Montpelier Road or going left to get back on to I-89.

**STRATIGRAPHY AND STRUCTURE OF THE CAMELS HUMP GROUP
ALONG THE LAMOILLE RIVER TRANSECT, NORTHERN VERMONT**

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INTRODUCTION

The purpose of this excursion is to present the stratigraphic and structural relationships of the Camels Hump Group in northern Vermont. The excursion crosses the Georgia Mountain anticline and Hinesburg Thrust in the Milton 7 1/2 minute quadrangle, the Enosburg Falls-Fletcher anticline of the Gilson Mountain quadrangle and the Richford syncline of the Jeffersonville 7 1/2 quadrangle. The eastern part of the Jeffersonville quadrangle lies just to the west of the Green Mountain anticlinorium (figure 1).

Significant problems to be addressed on this trip are the correlation of stratigraphy and comparisons of structure west and east of the Enosburg Falls - Fletcher anticline, correlation of the Underhill formation with the established western stratigraphy, and the origin of the Richford syncline exposed in the adjacent Jeffersonville 7 1/2 minute quadrangle to the east.

MAPPING AND ACKNOWLEDGEMENTS

The mapping conducted by the author in the Gilson Mountain and Jeffersonville quadrangles during the period of 1985 to the present has been funded by the Vermont Geological Survey under the direction of Charles Ratte. Mapping was done directly on 1:5000 orthomap aerial photographs and compiled at 1:12000. The author has benefitted from able assistance in the field by students completing field camp projects in the Gilson Mountain and Jeffersonville quadrangles. The early projects were conducted on 1:12000 enlargements of the topographic base and include the following students: Dave Marshall, Chris Miksic, and Todd Worsfold (1982); Debra Merrill, Steve Schope, Scott Schulein, Dan Dowling, (1983); Dave Iseri, Doug Friant, Doug Graham, Robert Myers and Jeffrey Slade (1984). Later projects were completed on

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the 1:5000 base by Dave Greenwalt, Greg Koop, Hugh Rose (1985) and Michael Landsman, (1986). Peter Thompson's (1975) work in the Enosburg Falls and Jeffersonville quadrangle is shown in part in the northeastern corner of Figure 3B. The author has also benefitted greatly by ongoing studies by his graduate students Maurice Colpron and Bill Dowling whose work in the Oak Hill Group in Quebec have helped to focus on the regional relationships discussed in this report.

REGIONAL SETTING OF THE CAMELS HUMP GROUP

The Camels Hump Group was first defined by Cady (1956) in the Camels Hump quadrangle of Vermont and subsequently subdivided into formations by Doll et al. (1961). The Camels Hump Group are almost entirely rift related rocks which predate the development of the passive continental margin along the ancient "western" boundary (present coordinates) of the Iapetus Ocean (Williams and Stevens, 1974). The Group thus includes most of the rocks stratigraphically overlying the Grenville basement and underlying platform slope or rise sequence rocks of the ancient margin.

In the western outcrop belts of the Camels Hump Group and correlative lower Oak Hill Group rocks of Quebec (Clark, 1934) the sequence is overlain by rift-drift transition "basal" quartzites of the Cheshire formation and drift stage dolomites quartzites and carbonate breccias of the shelf and slope facies. To the east the rocks of the Camels Hump Group are overlain by slope and rise facies rocks of the Sweetsburg and Ottauquechee formations. Significant facies differences on both local and regional scales are noted by previous workers and should be expected in rift environments both parallel and perpendicular to the basin. Detailed stratigraphic comparisons of rocks in the rift basin have been hampered by a lack of detailed mapping within the Camels Hump group and deformation which postdates the formation of the basin. This excursion attempts to detail the stratigraphic relationships of the Camels Hump Group in the vicinity of the Lamoille River transect.

The Lamoille River transect from the Milton to the Jeffersonville quadrangles, provides an excellent array of exposures which help to define the across strike nature of the ancient rifted basin. The western sequence of the Camels Hump Group displays a thin stratigraphic sequence with at least one erosional unconformity and rift-drift to platform cover rocks. The eastern sequence involves a greater percentage of volcanic rocks, thicker rift -clastic sequences and slope to rise cover rocks (Figure 2). The increased rift related subsidence to the east supports the model of formation of "instantaneous" eastward facing rift basins by lithospheric stretching and Airy -type subsidence synchronous with rifting (e.g. McKenzie, 1978).

Significantly, the Lamoille river section differs in detail from that observed to the north in Quebec (e.g. Clark, 1934; Dowling et al., 1987; Colpron et al., this guidebook) and in the Enosburg Falls quadrangle (Dennis, 1964). To the south, in the

Figure 1
Regional Setting
of the
Lamoille River Transect

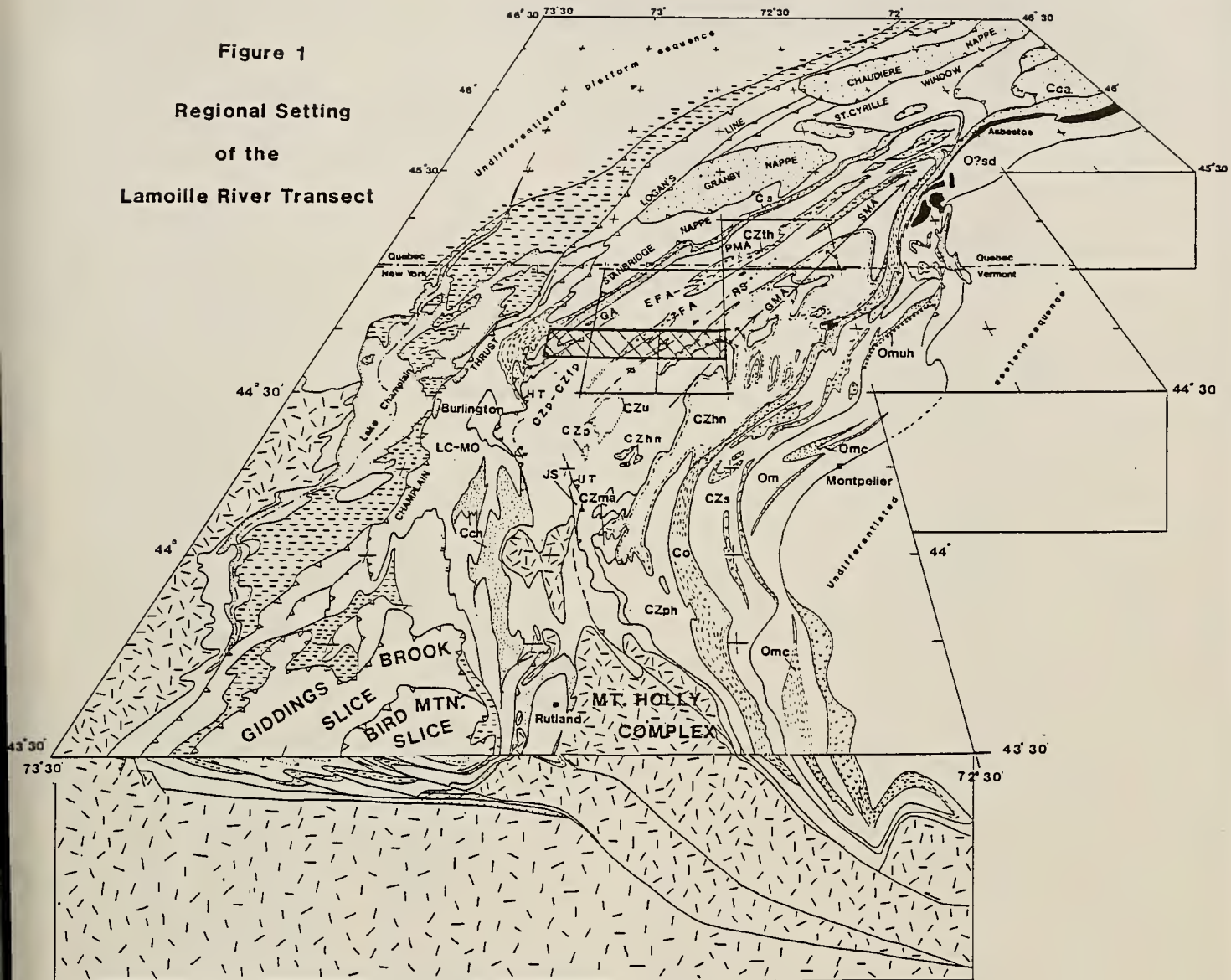
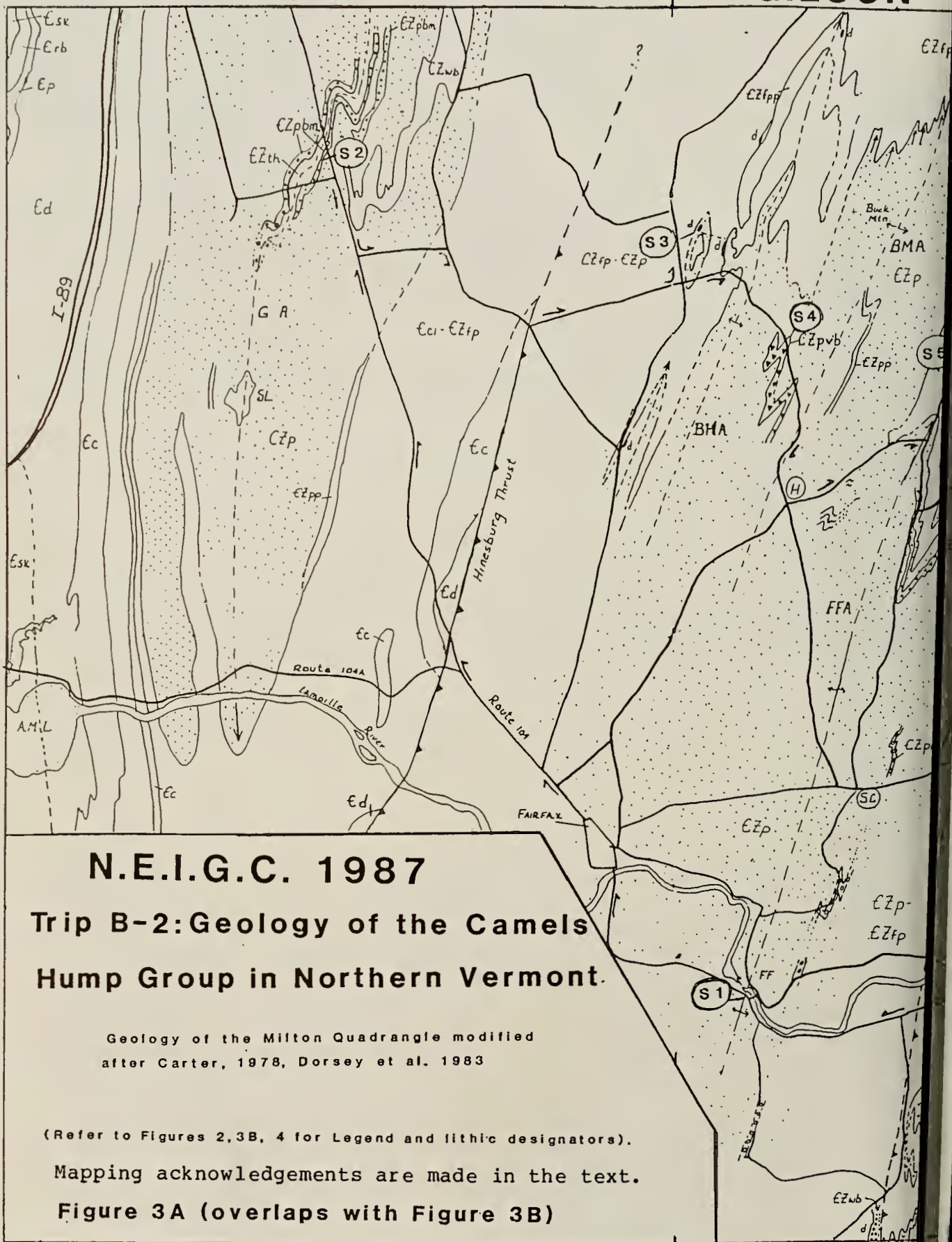


Figure 1: The location of the Lamoille River transect is shown on this block diagram of the Quebec reentrant constructed with a single due north vanishing point. 15' quadrangle boundaries are shown with plus symbols throughout. The diagonally ruled area includes from west to east: part of the Milton, the complete Gilson Mountain and part of the Jeffersonville 7 1/2 minute quadrangles (see figure 3a,3b). The latter two quadrangles are the northern half of the Mt. Mansfield 15' quadrangle (Christman, 1959). The Enosburg Falls 15' quadrangle (Dennis, 1964) is shown immediately north. The Sutton Falls quadrangle (Eakins, 1964) is outlined to the northeast in Quebec. The Camels Hump quadrangle (Christman and Secor, 1961; Thompson and Thompson, this guidebook) is directly south of the Mt. Mansfield quadrangle. **Abbreviations:** EFA: Enosburg Falls anticline; FA: Fletcher anticline; GA: Georgia Mountain anticline; GMA-SMA: Green Mountain-Sutton Mountain anticlinorium; HT: Hinesburg Thrust; JS: Jerusalem slice; PMA: Pinnacle Mountain anticline; RS: Richford Syncline; UT: Underhill thrust. Lithic designators shown on figures 2, 3, and 4 except as follows: Cca and broad stipple pattern: Caldwell, Armagh, Granby and related rocks mostly in Quebec; Cch and fine stipple pattern: Cheshire quartzite; CZma: Mount Abraham Formation; CZph: Pinney Hollow Fm; CZs: Stowe Fm; Om: Moretown Fm; Omc: Cram Hill Fm.; Omuh: Umbrella Hill conglomerate; O?sd: St Daniel Fm.; horizontal dashes: middle to late Ordovician flysch; random dash: Grenville basement of the Adirondack Mtns., Green Mountain Massif and Lincoln Mountain massif; black symbols: Thetford-Asbestos-Orford ophiolites of Quebec; Eden Mills ultramafic body of Vermont. Geology after Doll et al., 1961; Williams et al., 1978; Osberg, 1965; Doolan et al. 1982.

MILTON QUADRANGLE

73° 00'

GILSON



N.E.I.G.C. 1987

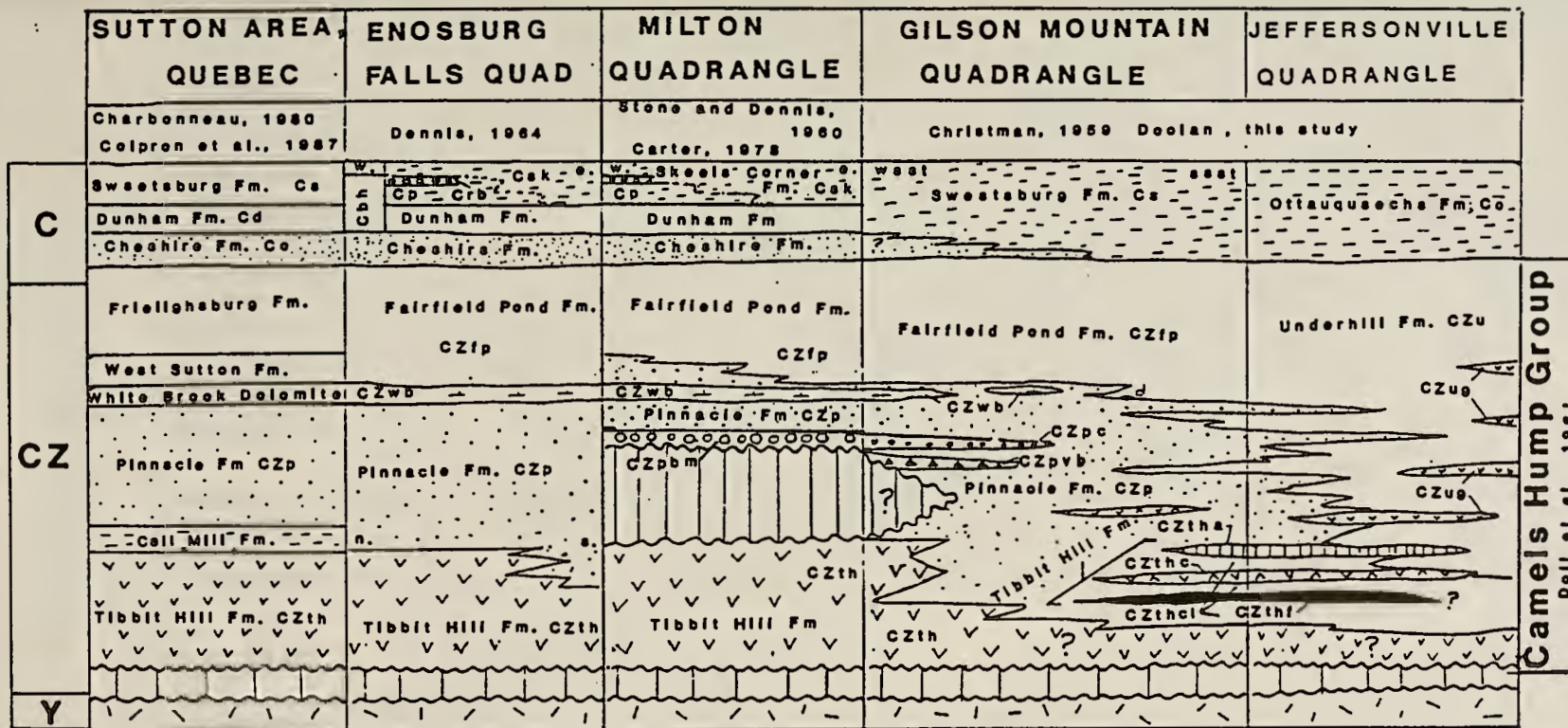
Trip B-2: Geology of the Camels Hump Group in Northern Vermont.

Geology of the Milton Quadrangle modified after Carter, 1978, Dorsey et al. 1983

(Refer to Figures 2, 3B, 4 for Legend and lithic designators).

Mapping acknowledgements are made in the text.

Figure 3A (overlaps with Figure 3B)

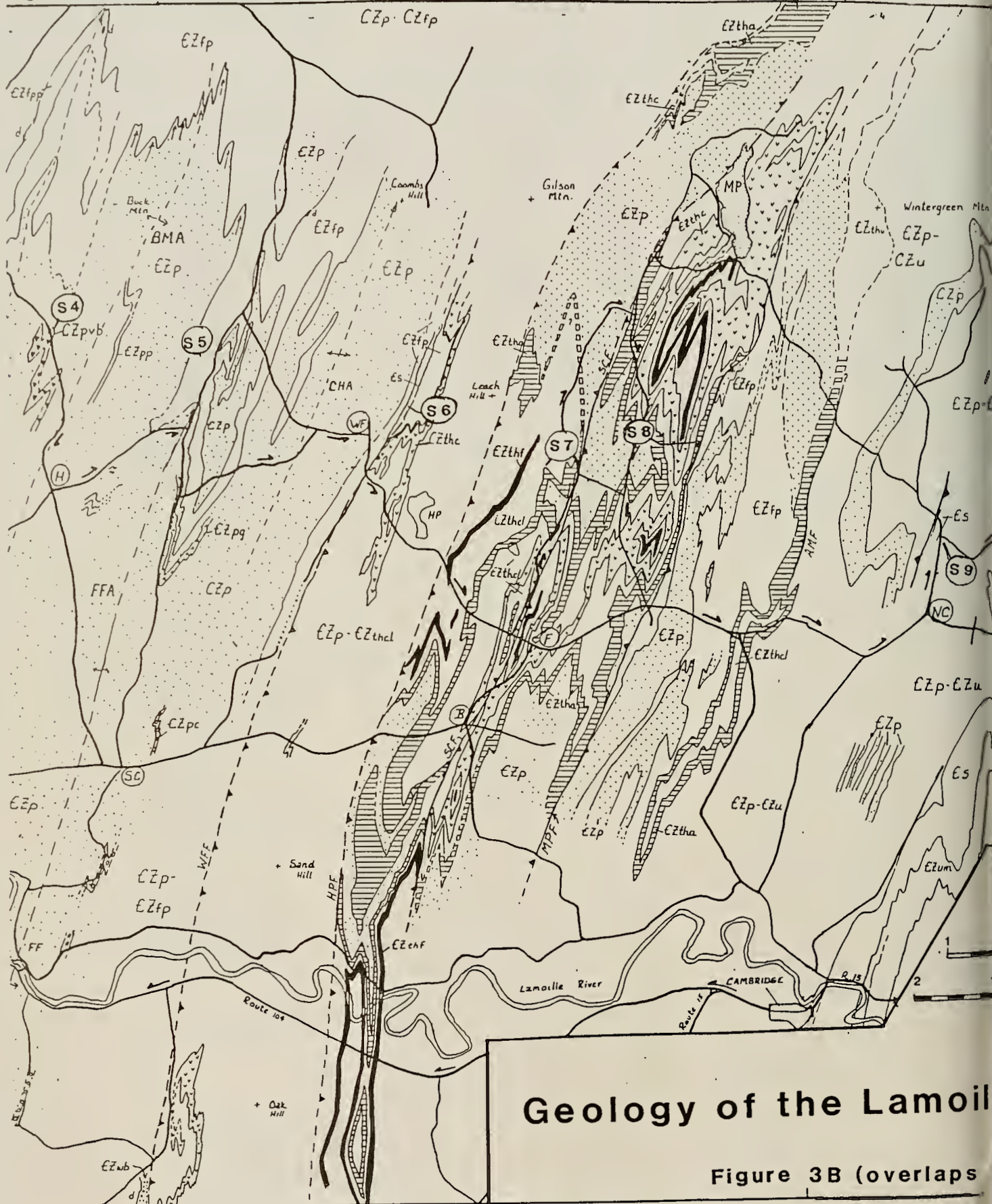


**Correlation of the Camels Hump Group and Cover Rocks,
Northern Vermont and Quebec.**

Figure 2: Stratigraphic correlation of the Camels Hump Group between Quebec and Vermont according to previous workers and this study. The stratigraphic column for the Milton quadrangle is modified from the interpretation of Carter (1979) by proposing an unconformity above the Tibbit Hill volcanics (Stop 2). The Pinnacle exposed in the Georgia Mountain anticline is stratigraphically and lithically correlative with the upper Pinnacle of the Gilson Mountain quadrangle.

GILSON MOUNTAIN QUADRANGLE

72° 52.5'

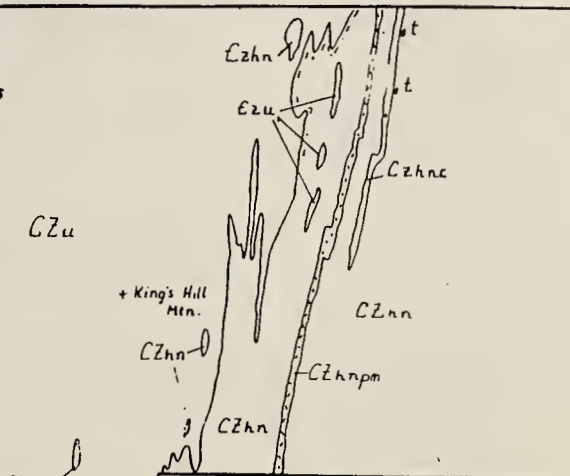
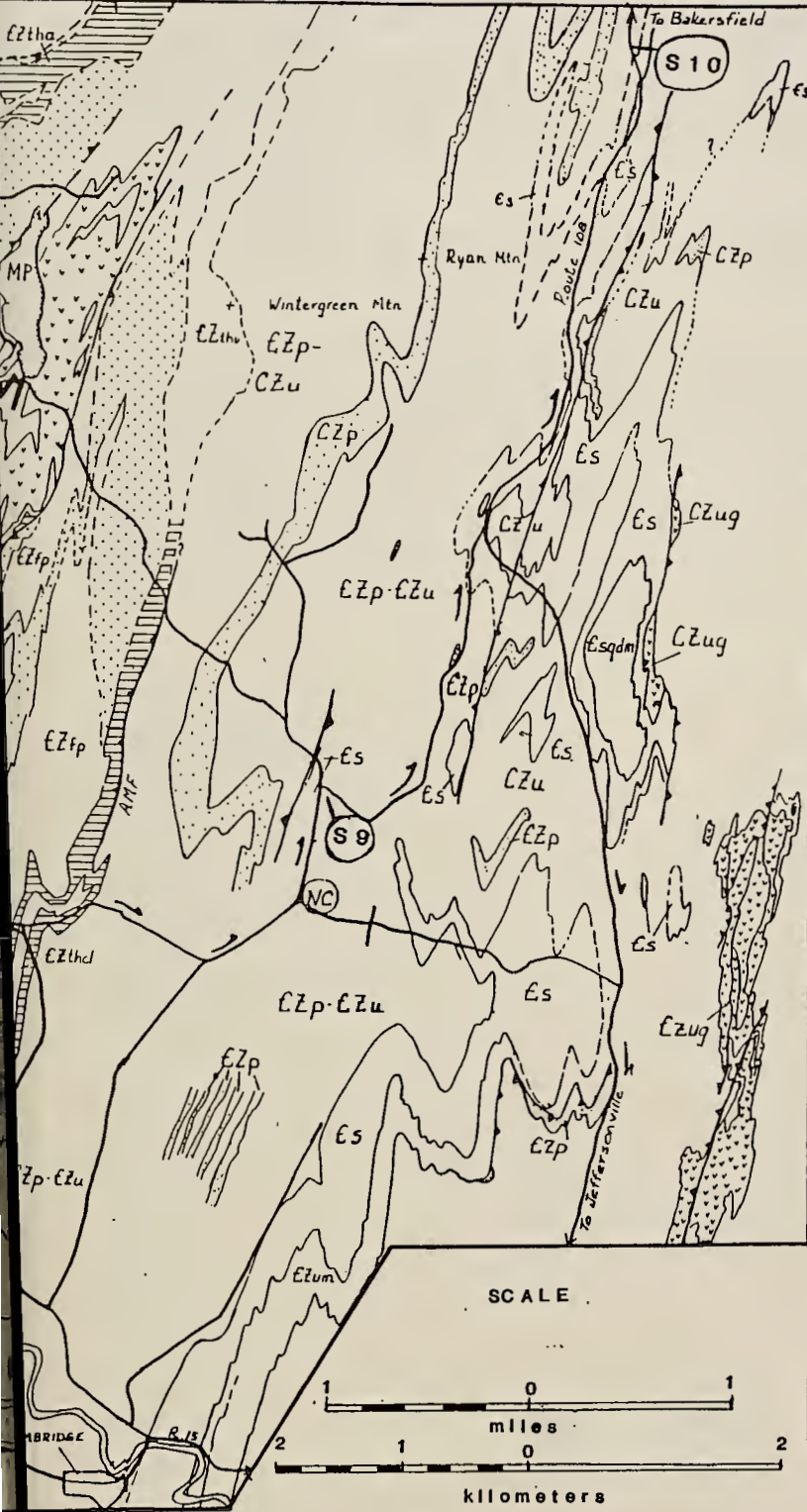


Geology of the Lamoiil

Figure 3B (overlaps

72° 52.5'

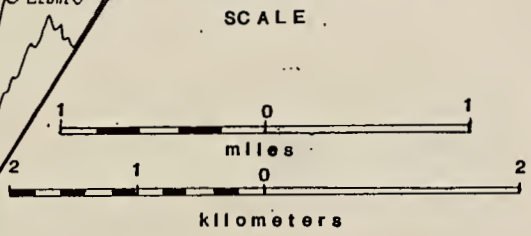
JEFFERSONVILLE QUADRANGLE



Abbreviations

- AMF: Armstrong Mountain fault
- AML: Arrowhead Mountain Lake
- B: Binghamville
- BHA: Buck Hollow anticline
- BMA: Buck Mountain anticline
- F: Fletcher FF: Fairfax Falls
- FFA: Fairfax Falls anticline
- GA: Georgia Mountain anticline
- H: Huntville; HP: Halfmoon Pond
- HPF: Halfmoon Pond fault
- NC: North Cambridge
- MP: Metcalf Pond
- MPF: Metcalf Pond Fault
- SC: Sanderson Corners
- SCF: Stones Creek Fault
- SL: Silver Lake
- CZu: magnetite schist
- CZhn: chloritoid-bearing unit
- t: talc body
- CZhnpm: Peaked Mountain greenstone

(Refer to Figures 2 and 4 for other lithic designators)



← TRIP ROUTE

Ⓢ STOP NUMBER

the Lamoille River Transect

3B (overlaps with Figure 3A)

Lincoln Mountain quadrangle the Camels Hump Group is considerably shortened and only the westernmost and easternmost parts of the Lamoille River section are lithically comparable. The missing section to the south lies mostly within the Gilson Mountain quadrangle and the western part of the Jeffersonville quadrangle. These rocks and their cover presumably have been removed from the Lincoln Mountain area during allochthon emplacement along thrust faults rooted below and within the Underhill Slice of Stanley and Ratcliffe (1985; Figure 1).

STRATIGRAPHY OF THE CAMELS HUMP GROUP ALONG THE LAMOILLE RIVER TRAVERSE

In this section the stratigraphy of the Camels Hump Group as exposed along the Lamoille River transect is described. The formation names as originally proposed are retained wherever possible. However, in light of the facies relationships observed across the strike belt some modification of the stratigraphic nomenclature is recommended. It is not the purpose to propose such nomenclature here since much new geologic mapping involving stratigraphic equivalent rocks described here is ongoing throughout the state. Therefore, the stratigraphic names used in this report should be considered informal.

The stratigraphy of the Gilson Mountain quadrangle is described first. Correlation with the adjacent Milton quadrangle to the west are suggested where appropriate. Secondly the stratigraphy east of the Enosburg Falls - Fletcher anticline is discussed with the emphasis placed on correlation with the western quadrangles. A complete historical development of the stratigraphy is not outlined here, but the interested reader is referred to the excellent synopsis of the regional stratigraphy in the references under the relevant stratigraphic columns shown in Figure 2. A more formal presentation of the stratigraphy for the Gilson Mountain and Jeffersonville quadrangles is forthcoming (Doolan, in preparation; Vermont Geological Survey reports).

As shown on Figure 2, lithic correlatives of the Oak Hill Group of Clark (1934) have been traced southward into Vermont with varying degrees of success. Although Dennis (1964) and Booth (1950) recognize the general sequence from base to top of Tibbit Hill volcanics - Pinnacle Fm - White Brook - Fairfield Pond (equals the West Sutton and Frielighsburg formations of Charbonneau, 1980) - Cheshire Fm, the following changes in the stratigraphy are noted:

1. The Tibbit Hill is interbedded with clastic rocks in the Enosburg Falls quadrangle (Pinnacle facies of the Tibbit Hill of Dennis, 1964);

2. The White Brook Dolomite is a discontinuous horizon in Vermont and when found does not precisely define the boundary between the equivalents of the Pinnacle and Fairfield Pond Formations in Quebec (Booth, 1950; Dennis, 1964);

3. More coarse grained conglomeratic facies are found in the Pinnacle formation of Vermont and an overall increase in

thickness of the Pinnacle is noted southward (Dowling et al., 1987).

4. The Call Mill slate, and West Sutton Formations as mapped in Quebec are only locally definable in Vermont (Booth, 1950; Dennis, 1964).

These differences in the stratigraphic sequence noted in the Enosburg Falls quadrangle become even more apparent in the Gilson Mountain quadrangle.

THE STRATIGRAPHY ALONG AND WEST OF THE ENOSBURG FALLS - FLETCHER ANTICLINE

In this section the stratigraphy of the Camels Hump Group in the Gilson Mountain quadrangle is summarized for the regions along and west of the Enosburg Falls-Pinnacle Mountain anticline. The presence of faults and complex folding of the entire sequence precludes an accurate rendition of the unbroken stratigraphic succession; however, numerous topping criteria in the metagreywackes and excellent rock exposure has enabled a newly defined stratigraphic sequence to be defined.

TIBBIT HILL FORMATION (CZth)

The Tibbit Hill formation was first defined by Clark (1934) for a thick sequence of volcanic rocks which define the base of the Oak Hill Group in Quebec. It is separated from the overlying Pinnacle formation in Quebec by a distinctive argillite called the Call Mill Slate. Southward, the Tibbit Hill volcanics are interbedded with metasediments and the Call Mill marker horizon is commonly not present (Dennis, 1964). Dennis (1964) notes the increase in metasedimentary material in the Tibbit Hill toward the south in the Enosburg Falls 15' quadrangle, and included the metasediments with the Tibbit Hill formation.

Christman (1959), confined the Tibbit Hill formation to metavolcanic rocks interbedded with metasediments in which amphibole and/or feldspar is identifiable in the field. This practice in effect redefined the Tibbit Hill formation by excluding some greenstones which are interbedded with Pinnacle formation. The problem is especially difficult in the Gilson mountain quadrangle where several greenstone occurrences appear to lie close to the stratigraphic top of the Pinnacle Formation. To define the top of the Tibbit Hill by the position of the stratigraphically highest greenstone horizon may exclude from practical useage the Pinnacle Formation.

The problem is partially resolved in this report by restricting the Tibbit Hill formation to three volcanic horizons along with their intervening metasedimentary units which occur in a 6 km. wide fault bounded belt defining the axis of the Fletcher anticline (Christman, 1959). Only three other occurrences making up less than 5% volumetrically of the volcanic rocks of the quadrangle are thereby excluded from the definition of the Tibbit Hill Formation.

The Tibbit Hill formation in the Gilson Mountain quadrangle consists of a bedded sequence of lava flows and metasediments. The formation is restricted to four fault bounded NNE trending sequences each approximately 1-1.5 kilometers in width (Figure 3). The two central fault block sequences exposes the complete Tibbit Hill formation, and provide the basis for correlation with the adjacent sequences to the west and east (Figure 3).

The metasedimentary units are referred to the clastic member of the Tibbit Hill formation (CZthcl). This member is represented by a variety of rock types including rusty weathering phyllitic grits and feldspathic quartzites and wackes, brown to grey punky weathering chloritic wackes, fine grained quartz chlorite albite granulite, calcareous chloritic wackes, and finely laminated chloritic phyllites, grits and volcanogenic tuffaceous metasediments which are locally brecciated. Attempts to subdivide the metasedimentary units even with the help of the distinctive volcanic stratigraphy has been locally successful, but continuity along strike for distances greater than 1 kilometer is rare. The high variability of the metasedimentary units and the lack of mappable marker beds within these units in part explains the failure to document the succession. Although many of the clastic units are lithically similar to the metasediments of the lower Pinnacle formation, they are included with the Tibbit Hill because of their apparant conformity on the outcrop scale with the volcanic stratigraphy.

The volcanic units of the Tibbit Hill Formation are subdivided into three members. These members from bottom to top are given the following names: feldspathic greenstone member, (CZthf); calcareous greenstone/chloritic wacke member (CZthc) and amphibolitic greenstone member (CZtha). The stratigraphic order of these members is documented by graded bedding and/or cross-bedding in the subjacent metasedimentary units.

Feldspathic Greenstone member (CZthf; Stop 8)

This member is described by Christman (1959, p. 18) as follows:

"The feldspathic greenstone is so-named because it contains lath-like remnant phenocrysts of plagioclase feldspar as much as one and one fourth inches long. These phenocrysts are seen most easily on the weathered surfaces as they weather white to light green in contrast to the fine-grained dark green groundmass. The phenocrysts have no apparent preferred orientation."

To this, the author adds that the matrix of the feldspathic greenstone member often weathers to a distinctive olive drab to brown color. The very fine grain size and color of this rock in itself makes the unit quite easy to recognize in the field, even where the phenocrysts are not obvious. Locally, a fine grained resistant and homogeneous brown weathering chloritic wacke underlies the feldspathic greenstone. The similarity of the

matrix of the volcanic rock with the wacke suggests that volcanogenic mixing with clastic rocks preceded the extrusion of the lava.

Excellent exposures of the feldspathic greenstone are found southwest of Metcalf Pond, south and southeast of Leach Hill, along the southeast flank of Hedgehog Hill southward to the Lamaille river, and west of Beaver Brook. The feldspathic greenstone member is approximately 25 to 40 meters thick and surprisingly continuous along strike except where folded out in dome and basin or hook fold structures or truncated by faults.

Calcareous greenstone/chloritic wacke member (CZthc; Stops 6,8)

Calcareous greenstone is characterized in the field by its pitted, somewhat rusty weathered surface and dark green matrix with thin laminae and splotches of calcareous material on fresh exposures. The calcareous layers appear to follow bedding in adjacent metasediments. This member is found stratigraphically above the feldspathic greenstone. The member is best defined as a mappable horizon in the east central block of Tibbit Hill (that block between the Stones Creek and Metcalf Pond faults, Figure 3). West of the Stones Creek fault and east of the Halfmoon Pond fault, the calcareous greenstone occurs in similar stratigraphic position as thin laminae interbedded with chloritic and quartzo-feldspathic meta-sediments. In this region the calcareous greenstone is best described as a volcanogenic metasediment. Even in the east central block of Tibbit Hill east of the Stones Creek Fault the calcareous greenstone is locally thin and interbedded with chloritic and calcareous wackes with detrital feldspar and quartz visible on the weathered outcrop surface. The calcareous greenstone is best developed in the region surrounding Metcalf Pond where the section is repeated by numerous folds around a major hinge. The thickness of the calcareous greenstone is conservatively estimated to be a maximum of about 110 meters. The transitional nature of the contact between metasediments (wackes and chloritic grits) with the calcareous greenstone in all areas and local interbedding of the greenstone with metasediments is the basis for not restricting this member to strictly a volcanic origin.

Calcareous greenstones identical to those described above are found in three areas west of the Halfmoon Pond fault: 2.8 km SE of Fairfax Falls; in the West Fletcher area immediately east of the West Fletcher fault (Stop 6) and in the region from Gilson Mountain to the Gore. The lack of other marker horizons of the Tibbit Hill formation between the West Fletcher and Halfmoon Pond faults does not allow a direct comparison of these volcanic horizons with that of the Tibbit Hill stratigraphy to be made. Ongoing mapping will hopefully resolve the problem.

Amphibolitic greenstone member (CZtha; Stops 7,8)

The stratigraphic top of the Tibbit Hill formation is defined here to be the amphibolitic greenstone member. This unit

outcrops in massive ledges and whalebacks throughout the belt of Tibbit Hill exposures. Amphibole (actinolitic hornblende) occurs as dark green porphyroblasts in random to strongly lineated orientations where it defines an L-S fabric parallel to the dominant foliation. Locally the amphibole is retrograded to chlorite and/or chlorite/actinolite clumps pseudomorphing the earlier formed amphibole grains. Outcrops have the appearance of massive wacke and grades into calcareous chloritic wacke in some localities. Although the coarse grained nature of this rock suggests a diabasic or gabbroic protolith the continuity of this member throughout the Tibbit Hill Formation in the Gilson Mountain quadrangle suggests that the rock originated as a basic lava flow. Thin sections also support the view that the primary amphiboles in the rock were porphyroblasts (see also Christman and Secor, 1961)

The amphibolitic greenstone is only about 15-20 meters thick; however in the excellent exposures in the vicinity of Hedgehog Hill the amphibolite outcrops over a wide area due to a complex fold pattern involving several refolded fold hinges.

At Stop 7 the amphibolitic greenstone is exposed in the hinge of a fold where it appears to display columnar jointing. Dennis describes volcanic horizons in the Enosburg Falls quadrangle with pillow lava structures. Throughout the Tibbit Hill Formation such primary volcanic textures are not common.

The reader is referred to Christman (1959) for excellent descriptions of amphibolitic greenstone and other members of the Tibbit Hill Formation.

Other occurrences of greenstone

Three other occurrence of greenstone occur west of the fault bounded sequence of Tibbit Hill Formation described above. A calcareous greenstone with gradational contacts with chloritic and locally calcareous wackes within the mapped syncline east of the Fairfax Falls anticline and north of Sanderson Corner (Figure 3); a volcanic breccia unit (CZpvb) west of the Buck Mountain anticline (Stop 4) and a thin belt of greenstone mapped as Tibbit Hill by Booth (1950), and Carter (1979) along the axis of the Georgia Mountain anticlinorium (Stop 2). The first two occurrences appear to be close to the stratigraphic top of the Pinnacle Formation. They are both thin (< 15 meters) and could not be traced along strike.

Outcrops of the Tibbit Hill formation in the Georgia Mountain anticlinorium are not abundant and have been previously interpreted (Carter, 1979; McBean, 1979) to occur as several horizons of thin volcanic flows interbedded with coarse boulder conglomerates containing numerous lithologies including Precambrian granitic gneisses, slates and fine grained wackes. McBean referred to the spectacular boulder horizon as the Beaver Meadow member of the Pinnacle Formation (CZpbm of Figures 2, 3, 4). Several visits to the outcrops by the writer and Bill Dowling

in June, 1987 have uncovered consistent topping directions away from the Tibbit Hill volcanic outcrops toward the sharply truncating basal boulder conglomerate horizon. Based on this data and the fact that slate fragments identical to the Call Mill in the Quebec sequence are locally abundant toward the base of the conglomerate, we interpret the Beaver Meadow to be a basal conglomerate unconformably overlying the Tibbit Hill formation. Pinnacle wackes interbedded with and overlying the Beaver Meadow boulder conglomerate are characteristic of the upper Pinnacle formation in the Gilson Mountain quadrangle and locally contain isolated pods of dolomite. Similar occurrences of dolomite are present in the upper Pinnacle of the Sutton area of Quebec as well (W. Dowling, personal communication, 1987).

The Tibbit Hill formation in the Georgia Mountain anticline is similar to the very fine grained compact matrix of the feldspathic greenstone which marks the lowest stratigraphic horizon of the Tibbit Hill sequence in the Fletcher area.

The Beaver Meadow Conglomerate (CZpbm) occurs at about the same stratigraphic position as slate pebble conglomerates and the volcanic breccia unit in the Gilson Mountain quadrangle.

PINNACLE FORMATION (Stops 1, 2, 3)

The Pinnacle Formation in the Gilson Mountain quadrangle consists predominately of greywacke interbedded with rusty phyllitic grits with quartz veins. The wacke is variable even along strike and the stratigraphic succession of the Pinnacle is not known in detail. The base of the Pinnacle is placed at the top of the amphibolitic greenstone in the area of the Fletcher anticline. The metasediments overlying the amphibolitic greenstone contain a greater proportion of massive wacke units and lesser amounts of volcanoclastic material than metasediments within the Tibbit Hill. Without the amphibolitic greenstone bed however the contact would not be easy to define.

The major outcrop belt of Pinnacle formation occurs to the west of the West Fletcher fault. East of the Hinesburg thrust, the Pinnacle is exposed in a repeated sequence across 4 anticlinal fold axes with intervening synclines. The anticlinal axes expose massive chloritic feldspathic greywackes which form resistant ridges. The rock is dark green on fresh surfaces and speckled with numerous white weathering plagioclase detritus and angular grains of quartz. Locally, this unit contains isolated angular clasts of tan weathering phyllite and phyllitic grit. Excellent examples of this rock type occur on Buck Mountain and at the anticlinal hinge at Fairfax Falls (Stop 1).

The upper part of the Pinnacle contains "cleaner" less chloritic quartz feldspar wacke and massive flaggy feldspathic quartzites. West and north of Buck Mountain, the upper Pinnacle contains rather continuous horizons of quartz pebble conglomerate. These resistant rocks contain calcareous matrices especially in proximity with horizons of White Brook dolomite

(Stops 2, 3). Interbeds of grey to green phyllite and grit are found throughout the Pinnacle but are more abundant near the top. The finer grained metasediments display thin vien quartz stringers and are rusty to tan on weathered surfaces. The interbeds range in size from thin phyllitic tops on graded beds to discontinuous horizons 10 meters or so thick.

Few marker horizons occur within the Pinnacle sequence; however, several are potentially important horizons to further refine the Pinnacle stratigraphy in the area. These include a volcanic breccia horizon (CZpvb; see Stop 4), a phyllitic conglomerate (CZpc; Figure 3), and a single horizon of calcareous greenstone (see above). These units all appear to be towards the top of the Pinnacle formation (Figures 2, 3).

WHITE BROOK FORMATION (Stop 3)

The White Brook Formation of Clark (1934, 1936) only sporadically occurs in the Gilson Mountain quadrangle. Where present it occurs as discontinuous buff to cream colored massive dolomite and dolomitic sandstone. The quartz content contributes to a resistance from erosion and the common occurrence of quartz vien stringers. Dolomite is not restricted to a single stratigraphic horizon. The lowest dolomite is found at the top of the coarse grained greywacke sequence which is taken to be the top of the Pinnacle when dolomite is absent. In the Milton quadrangle where the Pinnacle formation overlies a coarse boulder conglomerate, dolomite occurs below the top of the Pinnacle section. In the region north of Buck Hollow school, dolomite also occurs sporadically above laminated grits and phyllites which stratigraphically overly the defined top of the Pinnacle. These stratigraphically higher dolomite occurrences are in contact with homogeneous grey to black slate. It is uncertain if these homogeneous slates correspond to West Sutton Formation of Clark. The interval of dolomitic horizons is about 25 to 30 meters above the top of the Pinnacle formation. Because of the sporadic, thin and discontinuous occurrences of dolomite in the Gilson Mountain quadrangle, the White Brook is not separated on the accompanying map; however, dolomite occurrences are located by "d" on Figure 3.

FAIRFIELD POND FORMATION (STOPS 3, 5)

The Fairfield Pond was introduced to Vermont stratigraphy by Dennis (1964) as a member of the Underhill Formation. As defined by Dennis it includes the thinly laminated quartzites, grits and phyllites of the lower Cheshire Formation and the more homogeneous slates and phyllites of the West Sutton Formation. The Fairfield Pond is directly correlated with the West Sutton and Frelighsburg formations of Charbonneau (1980) in the Sutton quadrangle of Quebec. It may prove useful to future detailed mapping in pre-Cheshire rocks to subdivide the Fairfield Pond Formation into Frelighsburg and West Sutton facies. Examples of these facies will be shown at Stop 3.

As defined all of the rocks stratigraphically above the Pinnacle and below the massive quartzite beds of the Cheshire are included with the Fairfield Pond formation except where the White Brook dolomite occurs. The contact between the Fairfield Pond and the Pinnacle closely corresponds to the lowest position of dolomite associated with quartz pebble conglomerate and wacke of the Pinnacle in the Gilson Mountain quadrangle. Argillaceous wackes, phyllites and grits below the contact are thin and discontinuous as are wacke and conglomeratic facies which sporadically occur above the contact. Grey to black phyllites mapped around the Buck Hollow anticline (Figure 3) within the Fairfield Pond Formation are lithically similar to the West Sutton of Clark (1934) but occur at a higher stratigraphic position relative the Pinnacle. Mapping along the expected strike path of these phyllites has not been completed.

A second outcrop belt of thinly laminated argillaceous quartzites and grits which stratigraphically overly Pinnacle wacke are found between the Armstrong Mountain and Metcalf Pond Faults within the Fletcher anticline to the east. These rocks are identical to the Fairfield Pond lithologies mapped to the west but appear to have more phyllitic to schistose foliation surfaces. These rocks have been previously referred to as Underhill formation but because of their close association with Pinnacle and their lithic similarity with the rocks to the west they are included here with the Fairfield Pond. Contacts between the Fairfield Pond lithologies and the amphibolitic greenstone member of the Tibbit Hill to the west of the Armstrong Mountain fault (Figure 3) have not been observed, but a fault contact is suspected on stratigraphic grounds.

The Fairfield Pond formation in the Milton Quadrangle was restricted by Carter (1979) and Dorsey et al. (1983) to argillaceous rocks east of the Hinesburg Thrust. Rocks surrounding the Pinnacle in the Georgia Mountain anticline west of the thrust were referred to by these authors as the lower Cheshire. The lower and upper Cheshire of these authors corresponds to the Gilman Formation of Booth (1950). Stone and Dennis (1964) defined the Fairfield Pond to include the lower argillaceous quartzite of the Gilman formation as well as the West Sutton formation of Clark (1934). Consequently Dennis's map for the region defines the argillaceous rocks surrounding the Pinnacle in the Georgia Mountain anticline as the Fairfield Pond formation. The use of both lower Cheshire and Fairfield Pond appears to be in violation of the existing stratigraphic nomenclature. On Figure 3, the lower Cheshire of Dorsey et al. (1983) is designated as the Fairfield Pond Formation in keeping with the previous mapping of Stone and Dennis (1964). Lithically the rocks are similar on both sides of the Hinesburg Thrust and differ only in the abundance of thin quartzite laminae in the argillaceous matrix. As noted by previous workers the transition from the Fairfield Pond to the Cheshire is gradual. The apparent occurrence of "upper" Fairfield Pond close to the Pinnacle suggests that the Fairfield Pond Formation is considerably

thinner on the west side of the Hinesburg thrust compared with the east side.

STRATIGRAPHY EAST OF THE ENOSBURG FALLS -FLETCHER ANTICLINE

As noted above, Fairfield Pond and Pinnacle lithologies occur east of the main belt of Tibbit Hill volcanic members which define the Enosburg Falls -Fletcher anticline (Figure 3). East of this sequence of Fairfield Pond-Pinnacle sequence is an elongate belt of upper Tibbit Hill Formation (amphibolitic greenstone member and associated metasediments). The Armstrong Mountain fault at least locally separates the Tibbit Hill from the metasediments to the east and it is possible that the entire eastern amphibolitic greenstone west of the Armstrong Mountain fault is fault bounded.

From the amphibolitic greenstone eastward to the black slates and phyllites of the Sweetsburg Formation, no Tibbit Hill units are found and the entire sequence consists of quartz rich argillites and phyllites and schistose phyllites interbedded with wackes identical to those of the Pinnacle Formation. The tan weathering argillaceous metasediments are variable in quartz, chlorite, and white mica content but locally display excellent laminae characteristic of the Fairfield Pond formation. These rocks have previously been referred to as Underhill formation and that usage is followed in Figure 3. On the basis of lithic similarity and association this belt of rocks is correlated with the Fairfield Pond- Pinnacle sequence to the west. Locally, dolomite horizons associated with slightly calcareous quartz wackes are present west of the Sweetsburg lithologies This supports the view that White Brook lithologies are only sporadically present in this part of the Camels Hump Group. Importantly, the White Brook horizons along the east side of the Enosburg Falls-Fletcher anticline are associated with wacke characteristic of the Pinnacle formation - a situation similar to the the White Brook occurrences on the west side of the anticlinorium. Dennis (1964) interpreted black slate horizons near the Underhill -Pinnacle contact on the east side of the EFF antinclinorium as White Brook. These black slate occurrences are similar to rocks mapped as Sweetsburg in this report.

East of the outcrop belt of Sweetsburg formation the Underhill consists of a heterogeneous sequence of qtz-albite-chlorite-musc/ser-schists and phyllites. Magnetite is locally abundant and modal proportions of the minerals are variable. Greenstone lithically and chemically similar to the amphibolitic and calcareous greenstones of the Tibbit Hill are locally interbedded. Pinnacle lithologies are not common but where present appear as marker horizons which are discontinuous along strike. Garnet is not recognized in the field in these rocks but several localities of abundant chloritic pseudomorphs after garnet are present.

The stratigraphy and structural setting of the rocks lying between the Enosburg Falls -Fletcher anticline and the Richford

syncline is similar to the Mansville phase of the Oak Hill sequence first observed by Clark (1934, 1936) east of the Pinnacle Mountain anticline in the Sutton quadrangle of Quebec (Eakins, 1964; Clark and Eakins, 1968; Rickard, 1965). The Mansville phase includes rocks representative of the complete but structurally shortened or thinned Oak Hill sequence. East of the Sweetsburg outcrops defining the Richford syncline in Quebec the Bonsecours Formation occupies a similar position as the correlative Underhill Formation rocks east of the Richford syncline in Vermont. M. Colpron (personal communication, 1987) reports that the Mansville phase involves a complex fold and fault history which juxtaposes stratigraphically older sequences against younger rocks from west to east. It appears that the geology east of the Enosburg Falls - Fletcher anticline in northern Vermont may have a similar origin to the Mansville phase rocks in Quebec.

COVER ROCKS TO THE CAMELS HUMP GROUP

Cover rocks to the Camels Hump Group undergo significant facies changes across the Lamoille River transect. These changes are consistent with the different drift stage environments recorded in the cover rocks of the Camels Hump Group. In the Milton quadrangle the cover is "basal" quartzite of the Cheshire overlain by the Dunham formation representative of the Lower Cambrian platform. The Dunham is either overlain by Skeels Corner Slate or the Parker -Rugg Brook -Skeels Corner sequence (Stone and Dennis 1964; Dorsey et al. 1983; Figure 3). The grey to black laminated and carbonate-bearing slates are representative of shelf edge to slope deposits which further west correlate with the Monkton, Winooski, Danby and Clarendon Springs formations of the middle to upper Cambrian platform (Rogers, 1968, Dorsey et al. 1983; Figure 4).

Post-Fairfield Pond cover rocks are again found west of the West Fletcher fault (stop 6) and in the Richford syncline where black to grey pyritiferous slates correlated with the Sweetsburg Formation are found (Figure 3). These rocks are described below.

SWEETSBURG FORMATION (Cs; Stops 6, 9, 10)

Two separate outcrop belts of Sweetsburg formation rocks are found in the transect. The western belt occurs west of the West Fletcher fault (Stop 6). Here, the Sweetsburg rocks display gradational contacts along the eastern side of the outcrop belt with grey and rusty phyllites and laminated chloritic quartzose grits and phyllites of the Fairfield Pond formation. The Sweetsburg formation in this locality consists of interbedded rusty and non rusty carbonaceous to grey pyritiferous black slates and phyllite interbedded with quartzite and thin beds of marble. Numerous fold hinges suggest that the isolated occurrence may occupy an infolded inlier overthrust by the Tibbit Hill sequence along the West Fletcher fault.

A second belt of Sweetsburg formation rocks first mapped by

Christman (1959) outcrops in a discontinuous northeasterly trend across the Jeffersonville quadrangle from Cambridge to the northern limit of the quadrangle east of Route 108 (Figure 3). Dennis (1964) mapped similar rocks along the strike belt in the Enosburg Falls quadrangle to Richford, Vermont which correlates with the Sweetsburg formation mapped by Osberg, (1965) and earlier by Clark, (1934, 1936) in the Sutton quadrangle of Quebec. Osberg (1965) traced the Sweetsburg lithologies through a series of infolds into the serpentine belt of the Eastern Townships where it is referred to as the Ottauquechee Formation (Figure 1). Although Canadian workers prefer to restrict the useage of "Sweetsburg" to rocks outcropping above the Dunham formation along the west flank of the Pinnacle Mountain anticline in Quebec, the writer prefers to extend the Sweetsburg to the rocks in the Richford syncline for the following reasons: 1. the black carbonaceous phyllites in the Richford syncline are interbedded with carbonate bearing horizons as well as white to grey quartzites. Carbonate layers in the Ottauquechee are not common in central and southern Vermont where the Ottauquechee was first described (Perry, 1929); 2. The Ottauquechee Formation may undergo revision in definition with ongoing mapping underway in the Lincoln Mountain quadrangle (see Stanley and others, this guidebook).

The Sweetsburg lithologies mapped in the Richford syncline do however bear strong resemblance to the Ottauquechee as mapped in the serpentine belt of Quebec and northern Vermont (e.g. Doolan and others, 1982) and the correlation of Osberg, (1965) between Sweetsburg and Ottauquechee is likely correct. Ongoing studies of these lithologies on both flanks of the Green Mountain anticlinorium will hopefully define a more detailed stratigraphy within these cover rocks. If the stratigraphic relationships of the Cambrian platform are correct (Mehrtens, 1987; Dorsey et al. 1983) the continental slope rise sequence represented by the Sweetsburg and Ottauquechee Formations could span the entire Cambrian period.

The map pattern evolving for the Sweetsburg formation in the Jeffersonville quadrangle is apparently controlled by early west north westerly trending folds refolded by folds associated with the dominant foliation. Faults parallel to the dominant foliation are suspected by map pattern offsets and locally intense shear zones. Earlier faults may be present at least on the east side of the Sweetsburg outcrop belt as evidenced by isolated blocks of amphibolitic and calcareous greenstone along the contact (Rose, in progress). Contact relationships between the Sweetsburg formation and the adjacent Underhill is at least locally conformable. For example in the East Fletcher area, the sequence of black slate and phyllites, rusty quartzose grits and phyllites, albitic schists and magnetite schists can be followed across a dome and basin geometry. Lack of adequate outcrop control combined with a very complex fold and fault deformation history and the heterogeneous nature of the Underhill formation does not allow further elaboration of the original stratigraphic relationships to be made at this stage.

HAZENS NOTCH FORMATION

Along the west flank of the Green Mountain anticlinorium in the eastern part of the Jeffersonville quadrangle the Underhill is structurally intermixed with the rocks of the Hazens Notch formation. The Hazens Notch which presently is included within the Camels Hump Group by Doll et al. (1961), is not visited on this excursion. The palinspastic restoration of the position of the Hazens Notch formation with respect to the other formations of the Camels Hump Group and associated cover is a problem of great importance to Vermont geology. The Hazens Notch is host to Vermont's only blueschist occurrences (Laird and Albee, 1975; 1981), as well as numerous serpentinite and talc bodies which record all the observed deformations of the host rocks. The Hazens Notch as presently exposed is likely not part of the stratigraphy of the passive margin formation involving rift to drift stage sedimentation. The blueschist/serpentine/greenstone associations with the Hazens Notch metasediments argues in favor of formation within a subduction melange formed during the closing stages of Iapetus. The lithologies present in the Hazens Notch suggest that most of the "North American" protolith for the melange involved Sweetsburg and Ottauquechee cover rocks as well as the underlying Camels Hump group.

For further discussion of the Hazens Notch Formation the reader is referred to Thompson (1975; Thompson and Thompson, this guidebook).

DISCUSSION OF THE STRATIGRAPHY

The stratigraphy of the Camels Hump Group in northern Vermont is discussed in terms of the palinspastic reconstruction shown in Figure 4. The Camels Hump Group is interpreted in this reconstruction to represent clastic sedimentation and volcanism occurring synchronous with rifting of the Grenville basement prior to the development of a passive margin bordering the Iapetus Ocean. The age of rifting is unknown but based on regional relationships in the maritime and central Appalachians, a 600 - 620 m.y.b.p. age is inferred.

The oldest rocks known in the Camels Hump Group are the Tibbit Hill volcanic rocks as seen in the Pinnacle Mountain anticlinorium of Quebec and the Enosburg Falls anticlinorium of northernmost Vermont (Figure 1). In the vicinity of the Lamoille River transect, these rocks are only exposed in the core of the Georgia Mountain anticline in the Milton quadrangle. The thickness of the basal flow of the Tibbit Hill is not known but could be considerable if they represent continental flood basalts formed on rapidly extending and subsiding continental crust.

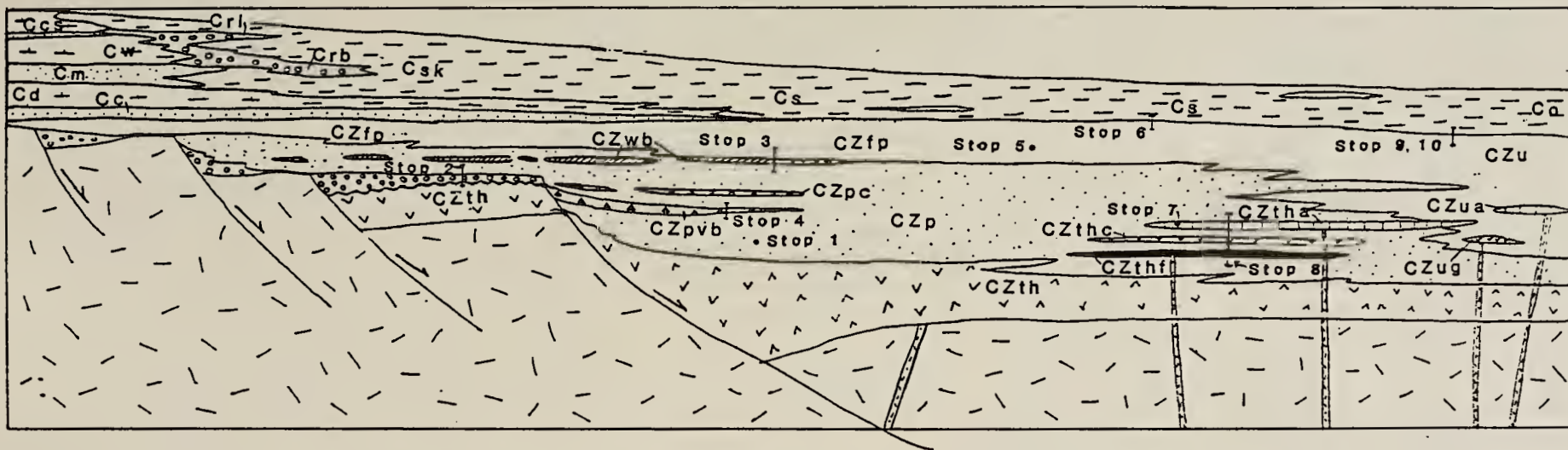
The lowest member of the Tibbit Hill Group (CZthf) in the Fletcher anticline is considered to be somewhat younger or at best only slightly older than the youngest Tibbit Hill exposed in

the Pinnacle Mountain-Enosburg falls anticline to the north. This is based on the observation of interbedding of Pinnacle wacke and related clastic rocks with the Tibbit Hill of the Fletcher area which is not observed in Quebec, and the similarity of the volcanic rocks at the eroded top of the Tibbit Hill in the Georgia Mountain anticline with the lowest volcanic horizon of the Tibbit Hill in the Fletcher anticline.

Erosion of the flood basalts occurred in regions not extensively thinned or loaded during the early rift stage as evidenced by the Beaver Meadow conglomerate unconformably overlying Tibbit Hill in the Georgia Mountain anticline (Stop 2). This unconformity is considered to have developed during a second rifting stage which supplied a variety of rounded clasts from Precambrian sources subsequent to Call Mill time in Quebec. This second rifting resulted in more rapid subsidence and extension to the east and perhaps coeval with flows seen today in the Tibbit Hill of the Fletcher anticline. Breccia units near the top of the Pinnacle formation in the western part of the Gilson Mountain quadrangle may be coeval with the second rifting.

The transition from the Pinnacle to the Fairfield Pond is interpreted to result from more distal source areas of clastics and a widening estuary capable of dolomite production (White Brook Formation). Locally, such as in areas located near clastic source terrains, dolomite production was impeded by continued influx of coarse grained wacke horizons. The Fairfield Pond and Underhill Formation are considered to be correlative but the Pinnacle Horizons found within the Underhill suggest that some of the finer grained protoliths of the Underhill may be coeval with coarse grained Pinnacle wacke to the west. The age of the volcanic horizons in the Underhill are considered to be slightly younger than those of the Tibbit Hill assuming that the axis of volcanism continues to migrate eastward with time in the axis of greatest lithospheric thinning.

The distribution of rock seen the Camels Hump Group across the Lamoille River transect are interpreted to reflect the evolution of an extending and subsiding Grenvillian lithosphere. The thickness of the lithosphere continued to provide sedimentologic controls on the distribution of cover rocks during the drift stage of passive margin development. Cheshire quartzite and Dunham dolomite were deposited on the filled rift basins located above slowly subsided and relatively thick lithosphere. These platform sediments may not have extended much across the palinspastic position of the Gilson Mountain quadrangle where subsidence due to loading and extension was presumably higher. During the drift stage, thermal subsidence controls dominated over the earlier rift related subsidence causing the platform to slope transition to migrate considerably westward from the eastern limit of Dunham dolomite (Figure 4).



Interpretive Reconstruction of the Stratigraphy along the Lamolle River Transect

Figure 4: Interpretive restoration of the ancient rift to drift stage passive margin of North America in the vicinity of the Lamolle River transect. The Camels Hump Group rocks (Tibbit Hill, CZth; Pinnacle CZp; White Brook, CZwb; Fairfield Pond, CZfp; and Underhill, CZu) are considered to be wholly developed in a rift environment. Variations in the stratigraphy and thickness of the various units reflect the effects of syn-rift volcanism, erosion and deposition. These effects are profound near regions of exposed basement such as proposed for the Milton quadrangle and western part of the Gilson Mountain quadrangle. Cover rocks to the Camels Hump Group are either rift - drift transition rocks (Cheshire to Dunham) or drift stage platform, slope, or rise deposits. The reconstruction suggests that the shelf - slope break, formed during the drift stage, occurred close to the basinward limit of rift stage erosion.

STRUCTURAL GEOLOGY

Introduction

The Camels Hump Group is polydeformed and metamorphosed. The deformation involves three periods of foliation development each of which is associated with folding. The map pattern structures are controlled largely by the second deformation structures including upright to overturned folds and steep faults subparallel to the second foliation which is usually the dominant foliation observed in outcrop. Kink folds post date the development of the dominant foliation and appear more common where superposed on fold hinges related to the second generation folds.

Major structures along the transect as defined by previous workers include the Georgia Mountain anticline, the Dead Creek syncline/Hinesburg Thrust, the Fletcher anticline, the Richford syncline and the Green Mountain anticlinorium. This study has resulted in refining the structural detail to the regions east of the Hinesburg thrust and especially within the Fletcher anticline. Work to the east of Fletcher anticline is still in early stages of mapping so structures in this area will not be discussed in detail.

As will be discussed more fully below the writer wishes to retain the use of the Fletcher anticline as first proposed by Christman (1959) for the region exposing the Tibbit Hill formation in the Gilson Mountain quadrangle. The lack of documented continuity between the Fletcher anticline with the Pinnacle Mountain-Enosburg Falls anticlinorium to the north combined with the unique Tibbit Hill stratigraphy recorded in rocks of the Fletcher anticline justifies that a distinction be made at this time.

The structural geology of the Camels Hump Group and cover rocks along the Lamoille River transect is discussed here in three parts. Part 1 discusses the geology west of the Fletcher anticline; part 2 discusses the fold and fault structures within the Fletcher anticline and Part 3 discusses the structural geology of the rocks within the region of the Richford syncline. A discussion of the structural evolution of the area in the context of regional geology concludes this section.

STRUCTURAL GEOLOGY EAST OF THE HINESBURG THRUST

Major structures in the Gilson Mountain quadrangle between the Hinesburg thrust and the Fletcher anticline include axial traces of anticlines and synclines which trend between N10E and N30E parallel to the dominant schistosity in the rocks. The Pinnacle Formation and overlying White Brook and Fairfield Pond formations in the northwestern quadrant of the quadrangle define four upright to overturned anticlines and associated synclines which generally plunge northerly. The axial traces of four anticlines are shown in Figure 3 based on observed bedding -

cleavage relationships and fold hinges. From west to east, these anticlines are referred to as follows: Buck Hollow anticline; Buck Mountain anticline; Fairfax Falls anticline and Coombs Hill anticline. The Coombs Hill anticline is not well defined but is inferred by the mapped synclinal trace to the west and several east facing topping directions to the east of the axis. The lack of distinctive marker horizons in areas of good exposure make it difficult to further define the fold geometry in this region.

The fold pattern is predominately controlled by F2 fold event whose axial plane cleavage is the dominant schistosity (referred to as S_n) in the rocks. An early F1 fold event is inferred on the basis of a ubiquitous S_{n-1} cleavage, numerous quartz veins folded by the dominant schistosity and a dome and basin map pattern believed to be derived by interference by the early fold events. A late fracture cleavage (S_{n+1}) is common in all areas of the quadrangle.

Fold geometry appears in the field and in map pattern to be asymmetric and west verging for the Buck Hollow and Buck Mountain anticlines. The Fairfax Falls anticline and Coombs Hill anticline appear to be more upright but data is not as abundant compared with the more westerly situated anticlines. Wavelength from anticlinal crest to anticlinal crest is approximately 1200 meters however, numerous smaller anticlines and synclines are noted on the limbs of the larger structures. Some of these smaller fold structures mapped around the Buck Mountain anticline have wavelengths of about 100 meters. Synclinal axes are not commonly observed because of the propensity of the overturned anticlinal limbs and hinges to form west facing cliff faces; shear zones observed along the overturned limbs parallel the axial surfaces of minor structure and the S_n foliation. Shearing thus appears to be a consequence of the F2 folding.

Structural relief does not appear to be great between the Hinesburg thrust and the Coombs Hill anticline; the stratigraphic units are repeated within the limits of lower Pinnacle to the upper ? Fairfield Pond formation. From the Coombs Hill anticline to the West Fletcher fault the structural relief increases bringing rocks as young as Sweetsburg in close proximity to Tibbit Hill volcanic rocks. (Figure 3). In the vicinity of the West Fletcher fault, the Sweetsburg - Fairfield Pond contact is characterized by a zone of rusty weathering grey to light green phyllites with intercalations of black slate. Shear zones are noted in several localities where the Sweetsburg -rusty phyllite contact is exposed. Bedding -cleavage relationships along the eastern contact of the Sweetsburg suggest that the overturned limb an anticline is sheared out along the contact. The Sweetsburg outcrop belt along the west side of the fault is interpreted to be a synclinal infold modified by subsequent faulting contemporaneous with the development of the dominant foliation.

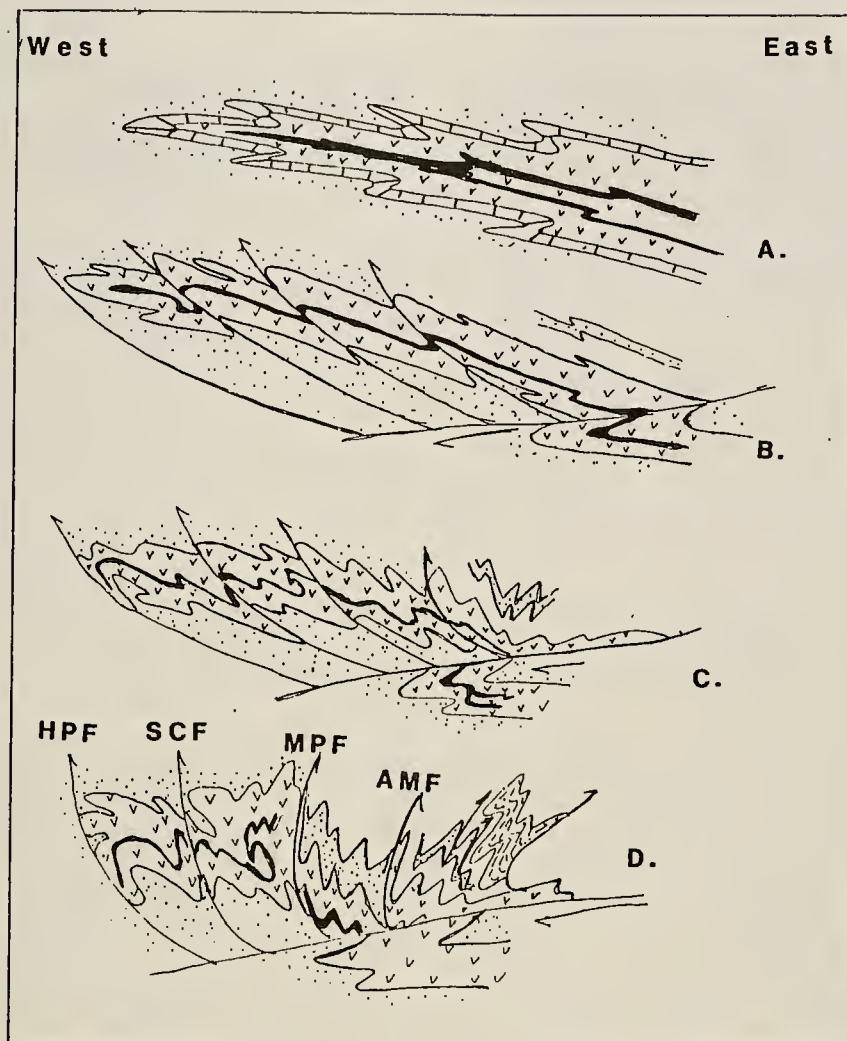
STRUCTURAL GEOLOGY IN THE FLETCHER ANTICLINE

The Fletcher anticline is an elongate NE trending structure underlain by the Tibbit Hill Formation along the Lamoille River transect. It extends across the full north south extent of the Gilson Mountain quadrangle northward into the southern end of the Enosburg Falls quadrangle mapped by Dennis (1964). The southern limit is not known but mapping by Christman (1959) suggests that the structure continues to at least to Jericho, Vermont in the southwestern corner of the Mount Mansfield 15' quadrangle. Christman and Secor (1961) report exposures of feldspathic, calcareous and amphibolitic greenstone in the Richmond and Huntington area in the southwest corner of the Camels Hump quadrangle. These occurrences suggest that the rocks found within the Fletcher anticline in the Gilson Mountain quadrangle may extend as far south as the Linclon Mountain quadrangle (Figure 1).

The "anticline" nature is based on the supposition that the rocks are stratigraphically older than the Pinnacle and Fairfield Pond formations observed on either side of the Tibbit Hill formation. The structure of the Fletcher anticline is far more complicated than the name implies and is more properly referred to as a nappe involving at least three periods of deformation. The pre-Sn deformational history associated with the Fletcher anticline appears to be more complicated and pervasive than the deformation in rocks to the west. This suggests that the Tibbit Hill rocks of the Fletcher anticline have been transported westward during pre-Sn deformation time relative to the Camels Hump Group rocks and cover situated in the western part of the Gilson Mountain quadrangle.

The structural complexities of the nappe can in part be resolved because of excellent exposure and a detailed volcanic stratigraphy discussed previously. Restoration of the folded stratigraphy is hampered however by a lack of fabric data (especially pre-Sn fold data) associated with the early deformation and the intense transposition of the earlier structures by the folding and faulting during the development of the Sn cleavage.

A fold generation model is schematically shown in Figure 5 based on the present map pattern. The model is based on the assumption that the amphibolitic greenstone constitutes a single horizon and that the syn to post Sn faults which transect the structure are relatively minor compared to the pre-Sn structure associated with the nappe formation and emplacement. The actual fold pattern is probably more complex than the model which is simplified by showing all fold generations as coaxial. The model however, shows that the dominant foliation imposed on the Fletcher anticline is superposed on an already tightly folded nappe structure. As shown on Figure 5 C,D, the Fletcher nappe/anticline is interpreted as a "pop-up" structure removed from its root during the backfolding backthrusting stage. This interpretation would resolve the problem of tracing the Tibbit



Structural Evolution of the Fletcher Anticline.

Figure 5: The evolution of the Fletcher anticline involves an early stage of tight to isoclinal folding not recognized in the map pattern of more westerly rocks. This is represented in A. by the emplacement of a nappe during S_{n-1} cleavage development. The refolding of the nappe is interpreted as a combination west verging folds and faults which steepen during a backfolding stage (B, C, D). The lack of volcanic rocks east of the Fletcher anticline to the Richford syncline suggests that the rocks in this vicinity represent the upper limb cover rocks to the Tibbit Hill formation. Unpublished vibroseis profiling along the western part of this schematic cross section support the flat lying nature of the Fletcher anticline and the west dipping fault "delaminating" the previously emplaced nappe. The folded dash pattern in Figure 5D represents the Sweetsburg lithologies of the Richford syncline (Figure 3B). Figure 5 B,C,D are all considered to be synchronous to the development of the dominant foliation S_n . HPF: Halfmoon Pond fault; SCF: Stones Creek fault; MPF: Metcalf Pond fault; AMF: Armstrong Mountain fault. See text and Figure 6 for further elaboration.

Hill rocks exposed in the Fletcher anticline to the north or south. Critical areas to further test the model are in the western part of the Jeffersonville quadrangle and the southern part of the Enosburg Falls 15' quadrangle. Neither area has yet been subjected to detailed mapping.

The Fletcher anticline appears to be fault bounded and imbricated. Five faults are shown in Figure 3 with a spacing of about 1300 meters. From west to east these faults are designated: the West Fletcher fault (WFF); the Halfmoon Pond fault (HPF); the Stones Creek fault (SCF); the Metcalf Pond fault (MPF); and the Armstrong Mountain fault (AMF). The WFF is suspected on the basis of noted shearing and intense shortening of the Fairfield Pond Formation to the west of the fault (see section on stratigraphy) and the abrupt stratigraphic juxtaposition of Tibbit Hill lithologies to the east and Fairfield Pond -Sweetsburg lithologies on the west. In the southern limit of the area, greenstone similar to the calcareous greenstone at West Fletcher occurs in close proximity to White Brook Dolomite breccia; this is the only basis for the extension of the West Fletcher fault southward.

The HPF is defined by sharp map pattern truncation of all the members of the Tibbit Hill formation along the projected trace of the fault. The fault is poorly constrained in the region east of Gilson Mountain. The fault can be directly observed just north of the Lamoille River 800 meters ESE of Sand Hill by truncation of the amphibolitic greenstone member of the Tibbit Hill formation (Figure 3).

The MPF is defined by zones of intense shear along its trace, isolated pods or slivers of amphibolite along the contact and by truncation of the Tibbit Hill formation rocks. Isolated occurrences of black slate of unknown origin are found east of the fault east of Metcalf Pond.

The AMF is only suspected to be a fault of any significance. Intense shear zones which separate the amphibolitic greenstone and magnetite bearing schists of the Underhill Formation are observed on the northeast side of Armstrong Mountain in the westernmost part of the Jeffersonville quadrangle. The same boundary cannot be documented as a fault surface along the full extent of the Tibbit Hill formation in this area.

The SCF is a suspected fault based on the map pattern of the Tibbit Hill on each side (Figure 3). Along most of its length in the Stones Creek valley, outcrop control is poor. The best evidence for the fault is an isolated occurrence of amphibolitic greenstone west of Metcalf Pond and shearing out of tight isoclinal folds east of Hedgehog Hill north of the Lamoille River (Figure 3).

STRUCTURAL GEOLOGY EAST OF THE FLETCHER ANTICLINE

Our knowledge of the geology east of the Fletcher anticline is based on the work presently underway by Mock in the East Fletcher area of the Richford syncline, the work of Thompson (1975) in northeastern part of the Jeffersonville quadrangle and contiguous areas to the north, and several field camp projects completed by University of Vermont undergraduates. In addition, Hugh Rose and Greg Koop conducted Senior Research studies at the University of Vermont in the areas north of Jeffersonville. Rose mapped the relationships of the metasediments and greenstone of the Underhill formation and the Sweetsburg Formation east of Route 108 and Koop mapped the Sweetsburg-Underhill - Pinnacle sequence west of Route 108 and ESE of North Cambridge (Figure 3)

These studies have concluded the following with regard to the structure of this area:

1. The orientation of the dominant cleavage (Sn) systematically changes from steep easterly dips east of the Fletcher anticline, to vertical dips in the Richford syncline to west dips east of the Richford syncline (Table 1). Thompson (1975) reports an average dip of 45 degrees to the east for Sn with progressive shallowing (83 to 28 degrees) in approaching the Green Mountain axis.

TABLE 1

REPRESENTATIVE Sn DATA ACROSS THE LAMOILLE RIVER TRANSECT					
Georgia Mtn	Gilson Mtn quadrangle		Jeffersonville quadrangle		
	west	east	west	central	east
N10E,80E	N26E,80E	N19E,85E	N24E,90	N9E,64W	N18E,45W
Carter (1979)	This paper				Thompson (1975)

2. Rose has documented several faults with west over east sense of motion and orientation parallel with the dominant schistosity of the area. Mock also has mapped steep shear zones within the Richford syncline parallel to the dominant foliation of the area.

3. Slivers of amphibolite and wacke identical to the rocks mapped to the east as Tibbit Hill and Pinnacle, respectively, are found by both Koop and Rose which predate the formation of the dominant foliation. The faults are tentatively included with the first deformation with east over west sense of motion based on the location of the stratigraphically older slivers east of stratigraphically younger rocks along the fault (Figure 3).

4. The Hazens Notch formation structurally underlies the Underhill Formation as a result of eastward vergent folding related to the west dipping dominant foliation (Thompson, 1975).

5. The Sweetsburg Formation map pattern mapped by Rose and Mock includes dome and basin and hook structures supporting the view that the structure involves the interference of at least two periods of deformation involving folding.

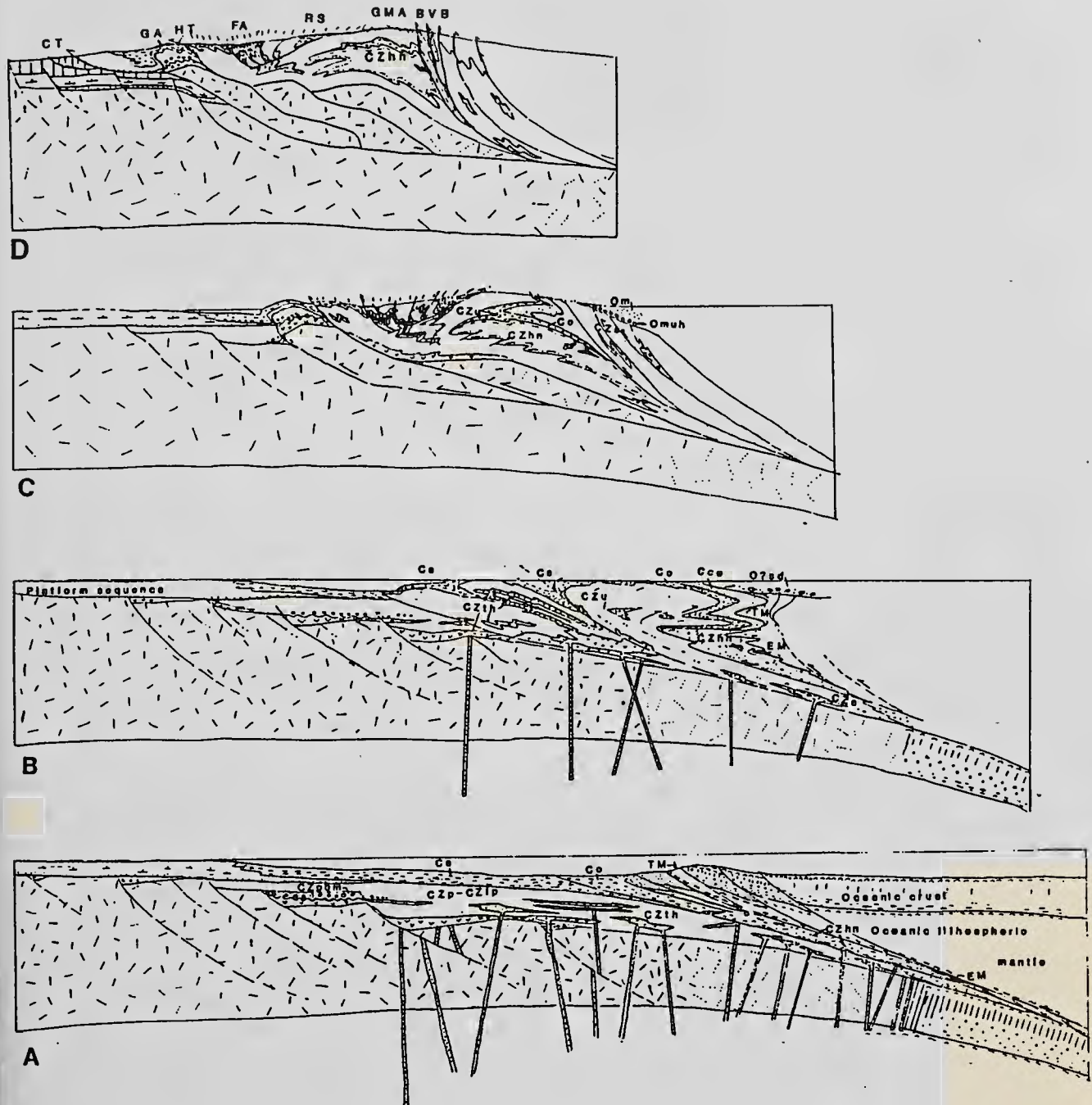
6. The S_{n+1} fracture cleavage, referred to as the Green Mountain cleavage because of its regional association with the Green Mountain anticlinorium, is present throughout the region east of the Fletcher anticline but appears not to play a strong control in the map pattern.

DISCUSSION

The stratigraphy and structure of the Lamoille River transect is briefly discussed here with the use of a model shown in Figure 6 which summarizes our present understanding of the geology. The metamorphic history of the area while not discussed in this report is not incompatible with the model and indeed may provide a quantitative test in the future for the P,T trajectories suggested for various segments of the transect. The model for the hinterland evolution of the Camels Hump Group and cover rocks is similar to the delamination model recently proposed for the southeastern Canadian Cordillera by Price (1986).

The palinspastic reconstruction for the Camels Hump Group (Figure 4) is incorporated into the reconstructed passive margin undergoing subduction in the late Cambrian to early Ordovician time (Figure 6A). The obduction of oceanic lithosphere occurred at this stage. The Thetford - Asbestos ophiolite belt of Quebec represents the emplacement at high structural levels and the Eden Mills ophiolite aureole in Vermont was emplaced at mid to deep crustal levels (Doolan et al. 1982). The high pressure blueschist assemblages reported by Laird and Albee (1981) also originated at this stage. Cover rocks to the Camels Hump Group (Ottawaquechee Fm) are subducted beneath the ophiolite and mix with underlying Camels Hump group and ophiolitic slivers to form the Hazens Notch formation.

Continued subduction - related compression in the early to mid Ordovician results in tectonic thickening of the supracrustal rocks (Figure 6b). Tectonic wedging of deep crustal oceanic lithosphere aids in delaminating the supracrustal rocks causing North American margin rocks to be directed eastward over the oceanic lithosphere. The Ottawaquechee formation is shown as an example of a high crustal level cover rock to the Camels Hump Group being directed eastward over Hazens Notch and earlier emplaced oceanic lithosphere. At deeper and more cratonward positions in the orogen east over west directed faults and nappes represent the first phase deformations observed in the Lamoille



Evolution of the Camels Hump Group in the Context of Regional Geology

Figure 6: The cartoon reconstruction of deformational events of the Taconic Orogen is schematic and not to scale. The degree of shortening shown and the relative thickness of the units involved is drawn within the constraints of clarity and space. Lithic designators are previously defined and/or shown in Figure 1. Although the reader should refer to the text for the discussion of this figure, the diachronous nature of deformation and the important influence of west over east shortening by folds and faults in the upper part of the crust should be noted. Abbreviations in figure 6D: CT = Champlain thrust; GA = Georgia Mountain anticline; HT = Hinesburg thrust; FA = Fletcher anticline; RS = Richford syncline; GMA = Green Mountain anticlinorium; BVB = Baie Verte - Brompton line.

River transect. Tectonic thickening in the outboard regions of the Camels Hump group results in pervasive and maximum metamorphic recrystallization of Taconic age. The Richford syncline is schematically shown as the region representing the cover rocks for the westerly directed Fletcher nappe overridden to the east by the more thoroughly recrystallized rocks of the Underhill formation. The Fletcher nappe overrides the cover rocks of the Camels Hump Group as evidenced by the Sweetsburg formation rocks observed west of the West Fletcher fault (Stop 6).

Basement slivers detached from the craton shown in Figure 6c are a consequence of continued subduction of the North American margin below the developing accretionary prism. These slivers aid in bringing deep crustal rocks to more shallow levels and to serve as tectonic wedges for further delamination and backfolding of rocks situated at higher structural levels in the crust. The second and most pervasive deformation of the rocks west of the Richford syncline occurred at this stage. The Fletcher nappe is backfolded along with cover rocks resulting in a steep structurally thinned "Mansville Phase" fold/fault sequence to the east of the Fletcher anticline. The Stowe formation considered to be the most distal equivalent of the Camels Hump Group (Coish et al. 1985), and metamorphosed at deep crustal levels is thrust westwards onto the backfolded Ottauquechee- Hazens Notch - and oceanic remnants. The uplifted Stowe consequently served as a source area for at least part of the unconformably overlying Moretown Formation locally marked by the Umbrella Hill Conglomerate (Figure 1, 6c; Doolan et al. 1982).

The stages represented by Figure 6 a, b, and c are subduction related processes associated with accretion tectonics in a subduction complex. Allochthon emplacement, in the Taconics and Quebec occurred during these stages. In Quebec the allochthons involved rocks which were predominately cover rocks to the Camels Hump Group (e.g. Stanbridge Nappe, Levis Nappe). The Charney/Granby/Armagh/Caldwell rocks may also be equivalent to the cover rocks in age (post lower Cambrian) but their original stratigraphic position with respect to the Camels Hump Group and the correlative lower Oak Hill Group of Quebec is not established. The lower allochthon slices of the Taconic mountains include stratigraphic correlative of both the Camels Hump Group and cover rocks in the Lamoille River transect. The higher Taconic slices are composed of correlative of the lower parts of the Camels Hump Group and basement. The accretion tectonics of the Quebec Appalachians in terms of both the obduction phase and allochthon development represents considerably higher structural levels than is observed in the central and southern part of the New England Appalachians. The Lamoille River transect represents a transitional level of crustal involvement between the Quebec and southern Vermont Appalachians.

The Taconic orogeny in the northern Vermont section can be explained by subduction related accretionary tectonics. The Moretown and St. Daniel basins formed outboard of the forearc

region created by the subduction zone complex. Collision stage tectonics involving active island arcs and/or already accreted terrains outboard of the interarc basin strongly modified the Taconian structures especially in promontory regions of the Grenvillian margin. Figure 6 D is a schematic cross section of the northern Vermont orogen at the conclusion of the collision stage. With regard to features seen on this excursion this stage was responsible for the formation of the Sn+1 cleavage, the Green Mountain anticlinorium and further westward imbrication of the foreland region. The possibility exists that much of the post Moretown deformation in northern Vermont is associated with the Acadian deformation which strongly modified Taconian geology along the Vermont -Quebec serpentine belt (Baie Verte Brompton zone of Figure 6D).

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ROAD LOG FOR TRIP B-2

Meeting point is the Cambridge Pharmacy, Route 15, Cambridge, Vermont. Two village markets are nearby to purchase lunch and refreshments. We will not be near a store at lunchtime. Since the trip will end at the same place all efforts will be made to consolidate participants into a smaller number of vehicles. Please cooperate in this matter since several of the stops require that we park on private property. The trip log begins at a commuter parking lot on the south side of Route 15, on the excursion.

PLEASE NOTE; IF YOU PLAN TO VISIT THESE STOPS ON YOUR OWN BE SURE TO OBTAIN PERMISSION FROM LANDOWNERS TO TRESPASS ON THEIR LAND.

- 0.0 Leave commuter parking lot and proceed west (left) on Route 15.
- 0.3 Intersection of Route 15 and Route 104. Bear right on Route 104.
- 0.6 Hills to the north (right) are underlain in part by massive amphibolitic greenstone of the Tibbit Hill formation. They mark the eastern limit of the Tibbit Hill formation in this area (Figure 3).
- 2.4 Road intersection. Proceed straight on Route 104 west.
- 3.0 Hills to left underlain by feldspathic and amphibolitic greenstone members of the Tibbit Hill formation. The Fletcher anticline crosses the river near this point (Figure 3).
- 4.5 Road intersection on left; proceed straight on Route 104 west.
- 5.5 Road intersection on left; continue west on Route 15.
- 5.8 Pull off on the northside of Route 104 just past the intersection with road leading over the Lamoille River.
- STOP 1. Fairfax Falls power station. Excellent examples of the lower part of the Pinnacle Formation as seen in the Gilson Mountain quadrangle outcrop on both sides of the road and on both sides of the river below the dam. We will confine our stop to the roadcuts. Topping directions and bedding/cleavage relationships suggest this stop is near or at the axis of an upright anticlinal hinge referred to as the Fairfax Falls anticline on Figure 3. The massive chloritic wacke contains abundant quartz - feldspar detritus. Thin sections show the development of stilpnomelane. Numerous north dipping kink bands strike both in east northeasterly and west northwesterly directions. The dominant foliation strikes about N20E and dips steeply to the east. S_n+1 have similar strikes but dips steeply to the west. Large clasts are rarely found; however coarse conglomeratic facies occurs as a rather consistent horizon to the east of the mapped trace of the Fairfax Falls anticline (see Figure 3).
- 5.8 Continue west on Route 104.
- 6.4 Entering the Milton Quadrangle

- 6.7 Junction with Route 128; bear to right on Route 104.
- 7.4 Cross the Lamoille River and enter the village of Fairfax.
- 7.55 Blinking yellow light; bear left on Route 104 west.
- 9.1 Outcrop on the north side of the road, just past road inter section is typical example of the Fairfield Pond formation.
- 9.15 Intersection of Route 104A on left; bear right and continue on Route 104.

Outcrops of quartz - rich dolomite of the upper Dunham and the Cheshire quartzite on the east side of Route 104 as you climb the hill. The Hinesburg thrust juxtaposes rocks of the Fairfield Pond fm. against these lithologies (Carter, 1978). The Dunham occurrences along the fault mark the axis of the Dead Creek Syncline of Booth (1950) and was interpreted as a syncline on the Centennial Map (Doll et al. 1961)

- 10.1 To the west are hills underlain by the Pinnacle formation coring the Georgia Mountain anticline
- 12.0 Road intersection on west; continue straight.
- 12.1 Road intersection on east; continue straight.
- 12.6 Road intersection on west.. this road leads to Beaver Meadow, the type locality of the conglomerate to be visited at stop 2; continue straight.
- 13.0 Farm on west side of road; pull off the road and await instructions for parking for Stop 2.

STOP 2. Park the cars and walk south to the outcrops on the west side of Route 104. Allen McBean will lead the discussion of these outcrops based on his Senior Research on the Beaver Meadow conglomerate. Exposures of medium to coarse grained wacke of the "upper" Pinnacle are exposed adjacent to the road to the south. These rocks contain pods of buff weathering dolomite which characterize the upper Pinnacle in the Oak Hill Group of Quebec (W. Dowling, personal communication, 1987). To the west the wacke is interbedded with several horizons of conglomerate composed of slate and previously deformed granitic gneiss and perthite. The igneous boulders are well rounded and range in size from 5 to 40 cm. The gneissic foliation of the clasts is randomly oriented with respect to the schistosity of the matrix. Besides the slate, sedimentary clasts include wacke and arkosic sandstone. The slate fragments are more abundant to the west where the boulder conglomerate is in sharp contact with the fine grained dark green rocks of the Tibbit Hill fm. The hematiferous slate fragments are lithically similar to the Call Mill slate which stratigraphically overlies the Tibbit Hill in Quebec (W. Dowling, personal communication). This and other evidence is discussed in proposing that the Beaver Meadow conglomerate marks and unconformity over the Tibbit Hill formation.

- 13.0 Proceed southerly on Route 104 to the unnamed dirt road which is the first left turn.

- 13.8 Intersection of road to west; turn left onto this unnamed dirt road.
- 14.4 T-intersection with another unnamed dirt road; excellent exposures of the White Brook dolomite outcrop in low ridges on the west side of the road to the north; we will proceed to the south (right turn).
- 15.3 Y intersection close to the contact of the Hinesburg Thrust as mapped by Carter, (1978); bear LEFT at the intersection.
- 15.6 Hill in front of you is Buck Hill underlain by Pinnacle Fm rocks in the Gilson Mountain quadrangle.
- 16.4 CROSSROADS. Proceed north (LEFT) on paved road known as Buck Hollow Road (unmarked). We have now entered the Gilson Mountain quadrangle.
- 16.7 Collins Farm on the east side of the road. Pull off the road and await parking instructions. Proceed east along path in the back of the cow barn to the outcrops in the pasture.
- STOP 3. This stop examines the contact relationships between the upper Pinnacle and the lower Fairfield Pond fm. The contact is not typical in that dolomite occurs along the contact at this locality. Such occurrences are not common in the Gilson Mountain quadrangle. The uppermost Pinnacle is a quartz pebble conglomerate which locally is highly calcareous. Proceed southward, along the ridge to outcrops of the argillaceous quartzites and thinly laminated argillites of the Fairfield Pond fm. Contact relationships with the quartz pebble conglomerate will be seen. The structure and outcrop distribution suggests a basin or dome structure cored by the younger Fairfield Pond Formation. If time permits, cross Polly Brook to the east to a synclinal hinge in the thinly laminated Fairfield Pond formation outcropping east of the brook. The syncline separates the two ridges displaying the Fairfield Pond/ Pinnacle contact. The Fairfield Pond is discussed in terms of its correlatives in Quebec (see Fig.2).
- 16.7 Proceed southward on the Buck Hollow road to the crossroads
- 17.1 At the CROSSROADS. Turn west (LEFT) onto unmarked dirt road. Cliffaces on the west side of Buck Mountain ahead are of coarse facies of the upper Pinnacle formation.
- 17.8 Road intersections; continue straight.
- 18.1 Pull off road and wait for parking instructions. At mobile home along east side of the road. Park and observe glacially polished pavement outcrop kindly stripped of cover by the landowner. NO HAMMERS ON THIS OUTCROP. Excellent hammer exposures are located just off the west side of the road.

STOP 4. Volcanic breccia of the Pinnacle formation (CZpvb on Figure 2,3,4). This unit is only found in a single horizon approximately 60 meters thick which is folded into a series of north plunging synclines and anticlines on the east side of the Buck Hollow anticline. The dominant foliation dips steeply eastward with a strike of N35-40E. The rock breccia appears homogeneous both in composition and size of the clasts; however some variation is seen in the larger exposures on the

west side of the road. The matrix also appears to be volcano genic in origin and is tentatively interpreted as a tuff breccia. On the west side of the road a thin horizon of tuff without clasts is seen on the western contact of the breccia but this horizon is not continuous along strike. The Pinnacle wacke exposed near the contact is more chloritic than the wackes found near the top of the formation. The position of the breccia in map pattern suggests that it does not occur far from the top of the Pinnacle formation. The relationship between this volcanic breccia with the unconformity overlying the Tibbit Hill formation at Stop 2 is discussed.

- 18.1 Continue south on the dirt road to Huntville.
- Outcrops on the east side of the road are massive wackes and interbedded argillite of the Pinnacle formation
- 19.3 CROSSROADS marks the location of Huntville. Turn onto the dirt road heading to the east (LEFT).
- 20.4 T-INTERSECTION. Proceed north (LEFT) on dirt road.
- 20.9 Mobile home on the west side of the road. Again, a beautifully exposed and clean outcrop provided by the landowner. Park in the driveway and proceed to the outcrop north of the mobile home.
- STOP 5. This stop is of the Fairfield Pond formation as mapped in most of the Gilson Mountain quadrangle. Numerous minor structures are preserved by the well bedded nature of this formation. Fold axes of several generations plunge predominantly northward with west over east rotation. This suggests the outcrop lies west of the synclinal axis separating the Buck Mountain and Fairfax Falls anticline. Fold axes vary between N5W and N45E and plunge shallowly northward. The dominant foliation dips moderately to the east and strikes N30-45E. Sn+1 is steeper and more northerly. NNW trending kink bands dip northerly.
- 21.0 Proceed north on the dirt road
- 21.4 Road intersection. Proceed east (RIGHT) onto paved road.
- 22.3 Approximate location of the trace of the Coombs Hill anticline
- 22.5 Road intersection; proceed straight ahead on paved road.
- 22.8 Road intersection. This is West Fletcher! Bear right on the paved road.
- 23.1 Road intersection; bear left on the paved road.
- 23.3 Private road which is easily missed on the north (left) side of road. TURN LEFT just past the tennis court.
- 23.5 Stop just short of metal gate on farm road back up and park in clearing along the side of road and walk to the gate.

STOP 6. Purpose of this stop is to examine the rocks on either side of the West Fletcher fault. Just to the north of the iron gate on the east side of the road are excellent exposures of argillite and wacke exposed in a fold hinge. The rocks are interbedded with a horizon of calcareous greenstone which is found about 200 feet north. Continue up the road about 1300 feet to Y intersection with smaller woods road. Bear left and walk about 450 feet to a small brook. Proceed downstream to excellent rock exposures. The 500' traverse crosses rocks of the Fairfield Pond fm. (85') into tan weathering rusty phyllites (100') which have a gradational contact with black carbonaceous phyllites and quartzites typical of the Sweetsburg fm. to be visited in stops 9 and 10 (280'). To the west rusty phyllites (80') grade into the black phyllites. Along strike the rusty and tan phyllites appear to be in bedded contact with the coarse pebbly wackes of the Pinnacle Formation to the west. Outcrop maps of the area available to participants at this stop will be used to discuss the alternative interpretations of the black slate lithologies and their contact relationships with adjacent rocks.

- 23.5 Return to the paved road.
- 23.8 At the intersection with paved road; proceed east (LEFT).
- 25.3 INTERSECTION with dirt road on north; proceed north (LEFT) on the dirt road.
- 26.6 Lancaster's farm on the east side of the road south of the road intersection. Pull to side of the road and await parking instructions.

STOP 7. Pavement outcrops on the west side of the road of amphibolitic greenstone member of the Tibbit Hill formation (CZtha). Walk about 1400' west across the pasture to the sugar house on the hill. Large ledge to the south of the sugar house displays three dimensional exposures of structures interpreted to be columnar jointing. Such features have not been reported in the rift volcanics of the northern Appalachians. Alternative explanations welcome!

- 26.6 Proceed north on the dirt road.
- 27.9 Road bends sharply to right (east).
- 28.1 T-INTERSECTION. Turn south (RIGHT) on dirt road.
Large outcrops on east side of road are typical massive exposures of the amphibolitic greenstone member of the Tibbit Hill formation.
- 28.7 Cross small brook.
- 28.8 Turn LEFT at mailbox (Brodericks' residence)
Proceed up the driveway and await parking instructions.

STOP 8. This stop will entail a 1/2 mile walk along a marked trail which crosses an anticlinal limb of the stratigraphy of the Tibbit Hill formation. Comparisons of the stratigraphic section on adjacent flanks of the anticline will be made. Of particular interest is the interbedded clastic rocks of the Tibbit Hill formation which clearly differentiates the Tibbit Hill of the Fletcher anticline from the Tibbit Hill along the Pinnacle Mountain-Enosburg Falls anticline to the north. 1:5000 outcrop maps of the traverse will be available to participants at this stop to aid in location and facilitate discussion.

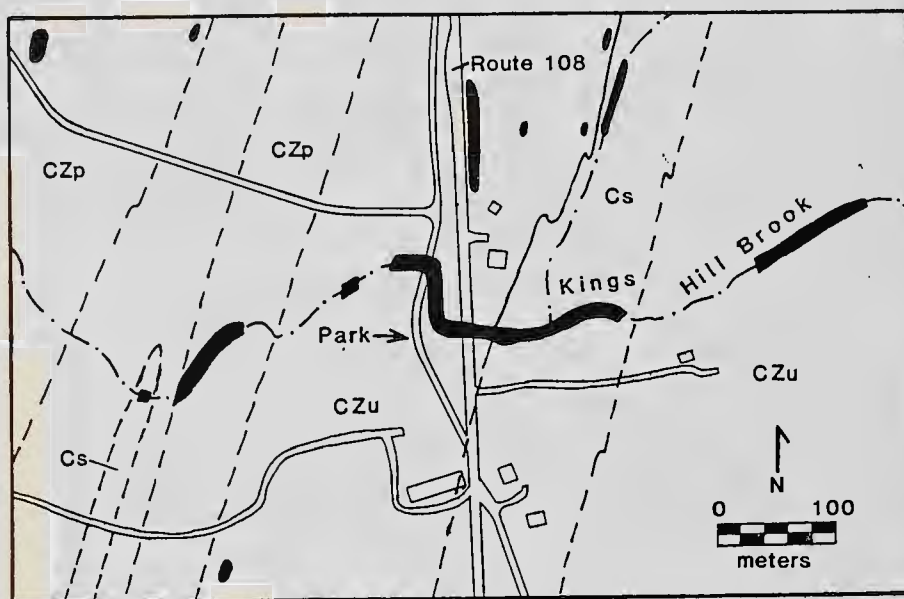
Return to cars and proceed back to the dirt road.

- 28.9 Intersection with dirt road; proceed south (LEFT).
- 29.5 Road intersection on right; continue straight.
- 30.3 INTERSECTION with paved road (Cambridge-Fletcher road)
Turn east (LEFT).
- 30.9 Y-INTERSECTION; BEAR LEFT onto dirt road.
- 31.2 Outcrops of amphibolitic greenstone mark the eastern limit of the Tibbit Hill fm.'
- 31.9 T-INTERSECTION; proceed north (LEFT) on dirt road toward North Cambridge.
- 32.5 Road intersects on east side; This is North Cambridge! Proceed straight ahead.
- 33.0 Intersection on right after crossing bridge. Pull off to side and park. Proceed to outcrops on the east side of the stream.

STOP 9. The large outcrop of lithologies of the Underhill formation (Czu) display excellent minor structures and well developed schistosity surfaces. This outcrop is compared with Sweetsburg formation rocks located on the west side of the road just northwest of the intersection. If the group is of reasonable size and/or willing, proceed down steep embankment to the northwest of the Sweetsburg outcrop to fault melange of Sweetsburg Formation lithologies. As of this writing, the continuation of the fault has not been mapped. These Sweetsburg lithologies mark the western limit of the Richford syncline in this area; outcrops of typical Pinnacle wacke containing detrital blue quartz grains outcrop to the west.

- 33.0 Turn east (RIGHT) at the road intersection.
- 34.8 Cross bridge followed by railroad tracks.
- 35.0 T-INTERSECTION with Route 108. Turn north (LEFT) onto 108.
- 37.7 Y-INTERSECTION with dirt road on west side of Route 108; Bear LEFT on the dirt road and park on side of road south of the bridge over Kings Hill Brook.

STOP 10. Tim Mock will lead the discussion of the rocks observed at this stop. The rocks on either side of Route 108 are examined and include rocks identical to those of the Pinnacle formation, the Underhill Formation and the Sweetsburg formation. The Sweetsburg formation at this locality displays interbedded marble and thin bedded quartzite along with the more typical carbonaceous phyllite.



Return to cars and proceed north to the intersection with Route 108.

- 37.8 Intersection with Route 108. Turn south (RIGHT) and head toward Jeffersonville.
- 44.9 Intersection with Route 109 on west side of Route 108; Proceed straight on 108.
- 45.2 Cross the Lamoille River.
- 45.4 INTERSECTION with Route 15; turn south (RIGHT) on Route 15.
- 45.7 Bear right at the blinking yellow light
Cross the Lamoille River
- 47.7 Flashing yellow light at the "wrong way bridge", proceed left at the light crossing the Lamoille River and staying on Route 15.
- 47.9 Return to the Cambridge Pharmacy in Cambridge.

END OF TRIP

LITHOFACIES, STRATIGRAPHY, AND STRUCTURE IN THE ROCKS OF THE
CONNECTICUT VALLEY TROUGH, EASTERN VERMONT

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Cady (1960) referred to the large area in eastern Vermont of primarily gray metasedimentary rocks bounded by the Taconic unconformity on the west and the Monroe line on the east as the Connecticut Valley-Gaspé synclinorium. This belt of rocks extends northeast from Vermont across southeastern Quebec and out to the end of the Gaspé Peninsula (Williams, 1978). To the south, it extends across western Massachusetts (Zen and others, 1983) and western Connecticut (Hatch and Stanley, 1973; Rodgers, 1985) to Long Island Sound. The emphasis on this trip is on these rocks as a sedimentary sequence. The stratigraphic data that I will show suggest that, at least in this area, the structure of the rocks in this belt is better interpreted as anticlinal. The belt herein will be referred to as the Connecticut Valley trough.

The rocks of the trough are gray slates, phyllites, schists, micaceous quartzites, and punky brown-weathering quartzose marbles. Metavolcanic rocks, primarily the Standing Pond Volcanics, constitute no more than a few percent of the total section. Because the formations that have been mapped in this package of rocks consist of different proportions of the same metasedimentary lithologies, the Connecticut Valley trough has long been considered to consist of one closely related sedimentary sequence. Since publication of the Centennial Geologic Map of Vermont (Doll and others, 1961), this sequence has been widely accepted as Late Silurian to Early Devonian in age.

Recent discoveries in rocks of the sequence of Middle to Late Ordovician graptolites in southern Quebec (Bothner and Berry, 1985) and southeast of Montpelier, Vermont, (Bothner and Finney, 1986), and of Early Devonian plants (Francis Hueber, Smithsonian Institution, written commun., October 5, 1985) in southern Quebec have raised questions about the age of the rocks and whether they do indeed represent one coherent stratigraphic sequence. This field trip will not resolve the question of the age(s) of the rocks, but it should shed some light on whether or not, or to what extent, they can reasonably be considered to form one continuous sedimentary sequence.

The most recent published compilation of the rocks of the Connecticut Valley trough in Vermont is that of Doll and others (1961). On their map, the rocks of the trough are divided into three formations--the Northfield, the Waits River, and the Gile Mountain. A slightly simplified version of their map of these rocks is shown here as figure 1. The present field trip is one outcome of a project to restudy the depositional history of the rocks in the area of the figure. These rocks probably were deposited at or somewhere near the eastern margin of North America during the time interval between the Taconian and Acadian orogenies. Tentative conclusions of this restudy, which will be pointed out and discussed on the trip, include (1) reassignment of the

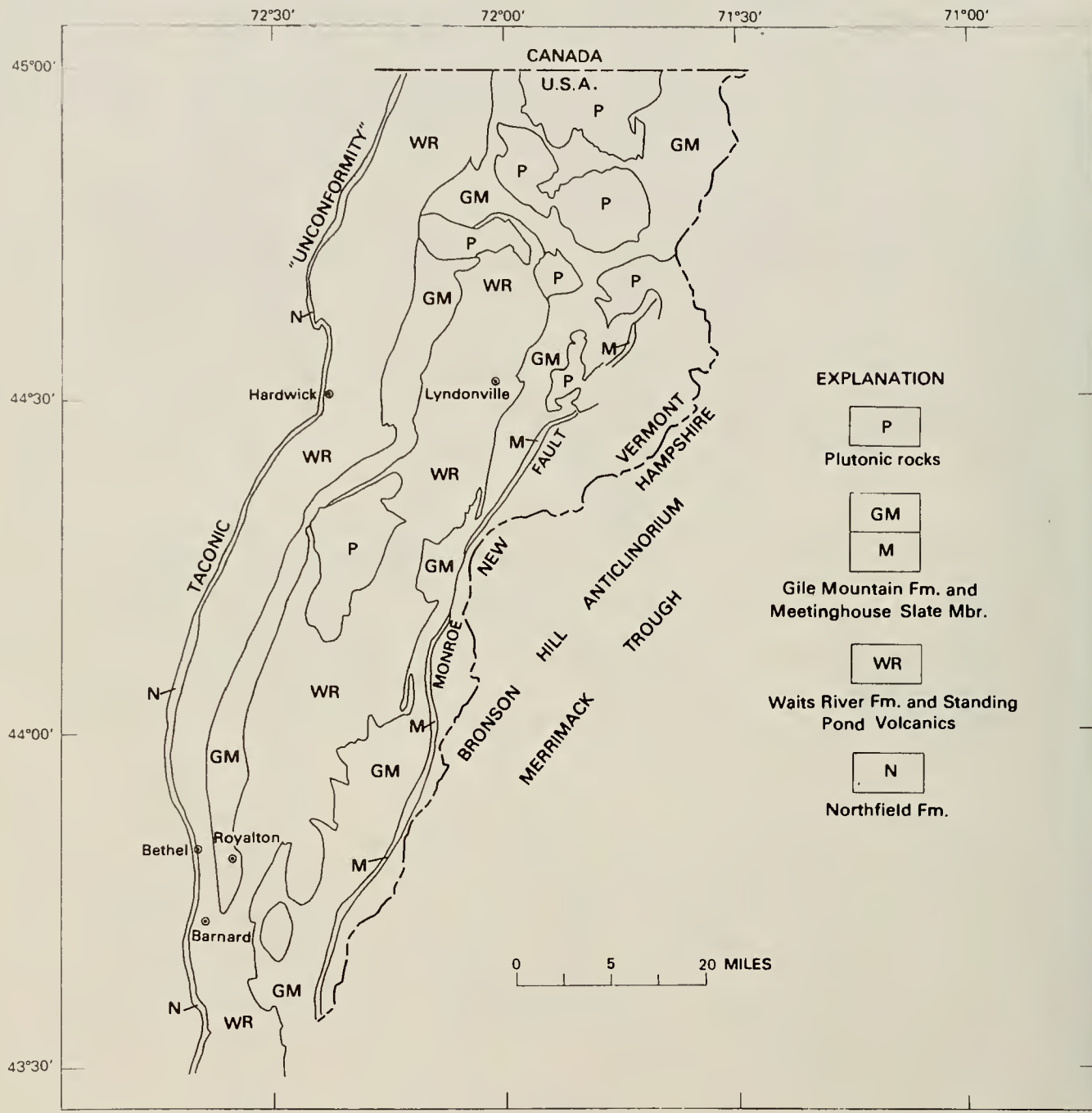


Figure 1. Geologic map of the metasedimentary rocks of the Connecticut Valley trough of eastern Vermont north of about 43°30', modified slightly from Doll and others (1961).

Northfield Formation to a position above, rather than below, the Waits River Formation and consequent correlation of the Northfield with the Gile Mountain, (2) recognition of at least three mappable lithofacies within rocks previously mapped as undifferentiated Gile Mountain Formation, (3) reinterpretation of the overall structure of the sequence as something more closely approximating an anticlinorium than a synclinorium, (4) a possible sedimentary model for the observed lithofacies that suggests an easterly source for the sediments of the sequence, and (5) some structural complexities that result from that sedimentary model.

As noted above, the commonly accepted concept of the stratigraphy of the Connecticut Valley trough (Doll and others, 1961) divides the rocks into three formations. The Northfield Formation forms a narrow belt about 1/2 mile wide along the western margin of the trough (fig. 1) and is interpreted by Doll and others (1961) as forming the basal unit of the sequence. They described the Northfield as consisting of "dark gray to black quartz-sericite slate or phyllite with fairly widely spaced interbeds a few inches thick of siltstone and silty crystalline limestone...". They showed the Waits River Formation on their map explanation as being stratigraphically equivalent to the Gile Mountain, but, on their cross sections, it appears as stratigraphically below the Gile Mountain. The Waits River consists of dark-gray slate, phyllite, or schist similar to that of the Northfield, interbedded with "gray quartzose and micaceous crystalline limestone weathered to a distinctive brown, earthy crust" (Doll and others, 1961). This latter rock is widely known to Vermont geologists as "the punky brown". At or near the top of the Waits River is the Standing Pond Volcanic Member which consists of a few hundred feet of amphibolite and greenstone. The third formation of Doll and others (1961) in the trough sequence is the Gile Mountain Formation. It, too, contains dark-gray phyllite or schist, characteristically interbedded with light-gray micaceous quartzite. Minor beds of brown-weathering crystalline limestone (marble) may be present. Approximately the easternmost 1/2 to 1 mile of the Gile Mountain was distinguished by Doll and others (1961) as the Meetinghouse Slate Member, consisting of "gray slate or phyllite characterized by beds of gray schistose quartzite 1/8 inch to 3 inches thick". The Meetinghouse was interpreted by them as forming the basal unit of the Connecticut Valley sequence along the eastern margin of the trough (the Monroe line) and thus to be approximately correlative with the lithically similar Northfield Formation along the western margin of the trough. The marginal surfaces of the trough, the Taconic unconformity on the west and the Monroe line on the east, were interpreted by Doll and others (1961) as unconformities. Note, however, that all of the quadrangle reports (Eric and Dennis, 1958; Hall, 1959; White and Billings, 1951; Hadley, 1950; Doll, 1944; Lyons, 1955) along the Monroe line from which Doll and others (1961) compiled their map interpreted the "Monroe line" as a fault and placed the Meetinghouse above, rather than below, the main body of the Gile Mountain.

Figure 1 shows that Doll and others (1961) have mapped two north-south belts of both the Waits River and the Gile Mountain Formations. The two belts of the Waits River join in the southern part of figure 1 at the south end of what Doll and others (1961, cross section C-C') show

as the Townshend-Brownington syncline. Fisher and Karabinos (1980) and I (fig. 2) have since reconfirmed from graded beds the correlation of the two belts of the Waits River and the structural nature of this syncline, which plunges gently north in this area. Figure 1 also shows that the two belts of Gile Mountain Formation come together at the north end of the figure. The abundance of intrusive rocks in this area tends to obscure stratigraphic relations, but previous detailed mapping and my own reconnaissance indicate that the eastern belt of Waits River does terminate northward, approximately as shown. This asymmetric distribution of formations east to west across the trough required Doll and others (1961, cross sections) to pinch out the Waits River Formation eastward between its eastern outcrop belt and the Monroe line.

Reconnaissance mapping in part of this area has shown that these rocks can be subdivided further. One of the major objectives of the trip will be to demonstrate the newly differentiated lithofacies of the Gile Mountain and to discuss what they may mean in terms of depositional environment, stratigraphic sequence, and the implications to structural interpretations. The following discussion, as well as the field trip, will progress from west to east across the trough.

The Northfield Formation, to be seen at STOP 1, forms a narrow belt along the western margin of the mapped area (figs. 1, 2). Although primarily dark-gray, aluminous, graphitic phyllite (or slate or schist, depending on metamorphic grade) and minor brown-weathering marble beds, the Northfield locally contains beds of light-gray, fine-grained micaceous quartzite as much as a few inches thick that generally are sharply bounded on one side but grade on the other side into dark-gray phyllite. The thickness and grading style of these graded beds markedly resemble those of the graded beds of the western belt of the Gile Mountain (described below). Furthermore, at the localities where graded beds were seen in the Northfield near the contact with the Waits River, those graded beds all indicate that the Northfield overlies the Waits River, contrary to the traditional view that the Northfield is stratigraphically beneath the Waits River. Some of those graded beds will be seen at STOP 1. If the Northfield is indeed stratigraphically above the adjoining Waits River, then it is in the same stratigraphic position as the Gile Mountain Formation as discussed below. It is hoped that STOPS 1 and 2 will demonstrate the similarities between these two units.

East of the Northfield is the westernmost of the two belts of Waits River Formation. The contact between the two formations is gradational through an interval of as much as a few hundred yards by eastward increase in the number and thickness of brown-weathering quartzose micaceous marble beds in a matrix of dark-gray aluminous phyllite. This contact will be seen at STOP 1. In this western belt of Waits River, these marble beds are 1 to 30 feet thick and constitute about 15 to 30 percent of the formation. Graded or ungraded beds of quartzite are very rare in this unit.

In the middle of this western belt of Waits River is a narrow belt of dark-gray phyllite with only minor beds of marble but with scattered beds of graded micaceous quartzite. It is shown on figure 2 as a belt

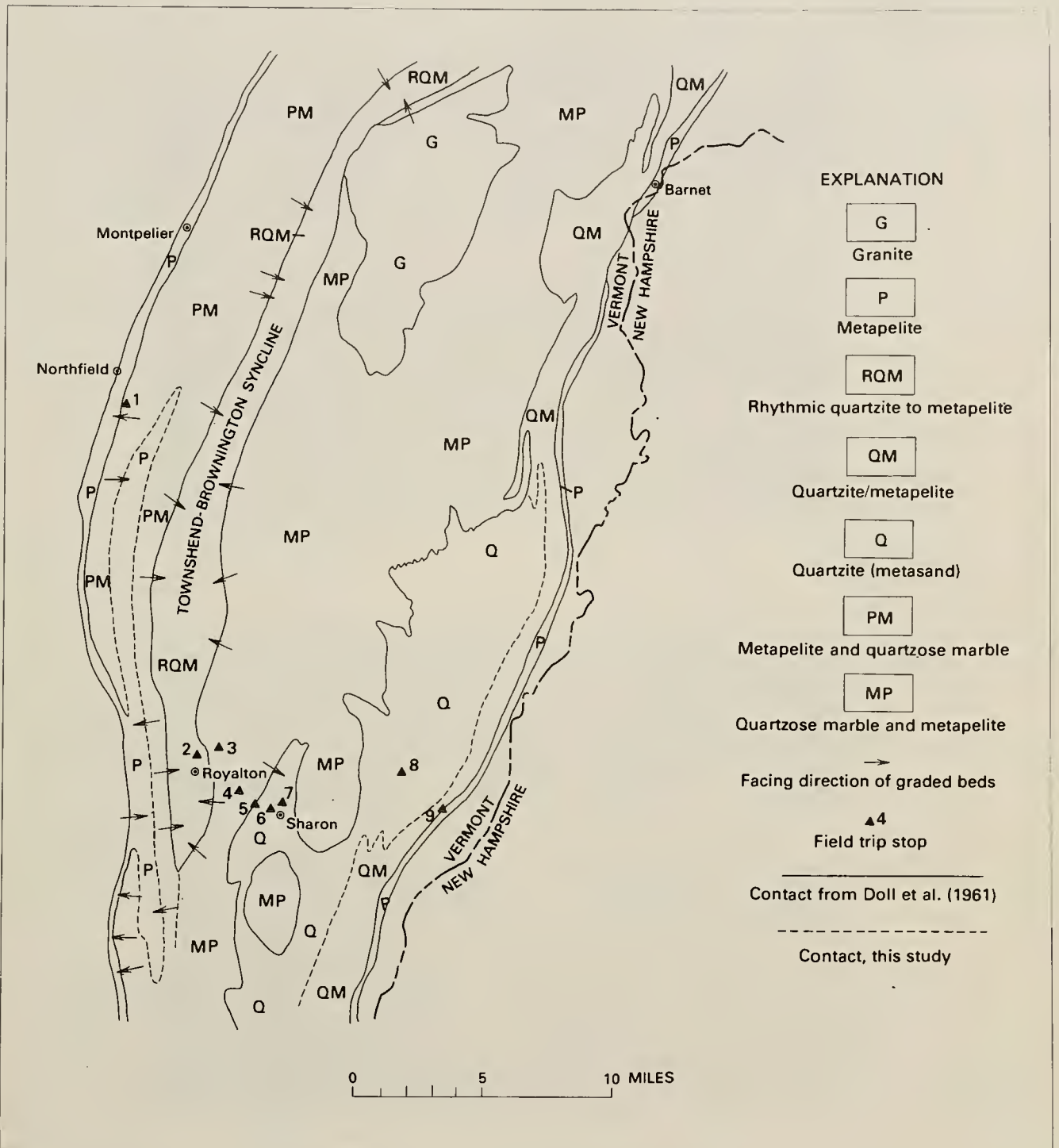


Figure 2. Geologic map of part of the area of figure 1 showing the lithic subdivisions of the present study.

of "metapelite" facies 2 miles west of Royalton. This belt does not appear on any published maps, although what may be the same belt is indicated on an unpublished manuscript map of the Barre 15-minute quadrangle by Richard Jahns and Walter White cited by Doll and others (1961). Available graded beds suggest that these rocks form a syncline within the broader belt of Waits River Formation (fig. 2). If so, they may be another belt of Gile Mountain correlative rocks. The abundance of graded beds intermediate between that of the Northfield and that of the western belt of Gile Mountain is at least compatible with that correlation, although the fact that brown-weathering marble beds are more abundant here than in either the Northfield or the western Gile Mountain is not as readily explained.

East of the western belt of Waits River is the western belt of the Gile Mountain, which will be seen at STOP 2. The rocks of this belt form a very distinctive unit, the structural potential of which was recognized by Fisher and Karabinos (1980). It consists of thin (generally 2-6 inches), almost universally graded beds of fine-grained micaceous quartzite and gray phyllite or schist (fig. 3). In an excellent detailed study of a small area around Royalton (figs. 1, 2), Fisher and Karabinos (1980) demonstrated from graded beds that the Gile Mountain Formation (at least in that area) stratigraphically overlies the two adjoining belts of Waits River. It had been so interpreted by Doll and others (1961) but without the benefit of documenting sedimentary structures. My subsequent work with graded bedding north and south of Royalton completely confirms those conclusions and further demonstrates that the Gile Mountain is structurally as well as stratigraphically above the Waits River (fig. 2) --the Townshend-Brownington syncline is alive, well, and real.

Next to the east from the western Gile Mountain belt is the eastern belt of Waits River Formation. As with the western Waits River belt, this belt consists of gray phyllite or schist interbedded with gray, brown-weathering, quartzose, micaceous marble. It differs from the western belt primarily in appearing to have a slightly higher percentage of marble beds (commonly 50-80 percent). The interbedded schists also appear to contain significantly more quartz veins and to be somewhat less aluminous. These rocks will be seen at STOPS 3 and 4.

At the east contact of the eastern belt of Waits River is the Standing Pond Volcanics. In this area of gray metasedimentary rocks, it forms a distinctive unit, generally only a few hundred feet thick, at or very close to the contact between the eastern belt of Waits River and the eastern belt of Gile Mountain. The Standing Pond will be seen at STOP 5. It has served as a critical marker horizon for working out the structure of this part of Vermont (Doll, 1944; White and Jahns, 1950; Lyons, 1955).

To the east of the Standing Pond is the eastern belt of the Gile Mountain Formation (fig. 1). Although the lithologies of the eastern and western belts of the Gile Mountain have only recently been differentiated (Hatch, 1986), they differ quite markedly, particularly south of about 44° latitude. In this part of the State, I have subdivided the eastern Gile Mountain into two distinct lithofacies (fig.

A



B



Figure 3. Photographs of the rhythmically graded facies of the western belt of the Gile Mountain Formation. A, View looking north at exposure near the south end of the Brownington syncline. Graded beds top to the east (right). B, Exposure at STOP 2. Graded beds top to the left.

2), which will be seen at STOPS 6,7,8,and 9. The western part of the eastern Gile Mountain belt consists of thick (up to 30 feet) beds of micaceous quartzite and quartz-feldspar-mica schist (fig. 4). These rocks, particularly the schists, have a brownish cast in outcrop or hand specimen and contain no visible graphite, in contrast to the gray graphite-bearing schist beds of the western Gile Mountain. Furthermore, graded bedding is very rare in these rocks, which I interpret to be metamorphosed, relatively rapidly deposited sands and muddy sands. They contrast to the more slowly deposited rhythmically graded rocks of the western belt, which I interpret to be metamorphosed turbidites.

The eastern part of the eastern belt of Gile Mountain is underlain by what is designated the quartzite/metapelite facies in figure 2. These rocks are distinctly different from the rocks of the quartzite facies to the west (fig. 2) and, in some, but not all, ways, resemble the rocks of the rhythmically graded facies of the western Gile Mountain. The quartzite-metapelite facies rocks are fine-grained, light-gray micaceous quartzite and dark-gray, aluminous, graphitic phyllite. Beds generally range in thickness from a fraction of an inch to about a foot (fig. 5), and show much local variation in contrast to the more consistent bedding thickness of the rhythmically graded facies to the west (compare figs. 5 and 3). The quartzite/metapelite facies rocks are believed to have formed as slowly deposited fine to very fine grained silt and clay-sized sediment. Approximately the easternmost 1/2 to 1 mile of this facies is predominantly, and in some exposures exclusively, dark-gray phyllite or slate. These rocks were mapped by Doll and others (1961) and by the quadrangle mappers from whom they compiled as the Meetinghouse Slate Member of the Gile Mountain Formation. Although all of the quadrangle reports from which Doll and others (1961) compiled their map placed the Meetinghouse at the top of the Gile Mountain, they placed it at the bottom. Neither group published any hard data on which to base their interpretation. My own work has turned up abundant evidence for isoclinal folding in these rocks, but a slight majority of graded beds near the gradational western contact of the Meetinghouse suggests that the Meetinghouse may lie stratigraphically at the top, rather than at the bottom, of the Gile Mountain. The rocks of this eastern part of the eastern belt of the Gile Mountain and the Meetinghouse will be seen at STOP 9.

Figure 6 is a reinterpreted cross section across the trough. It differs from the cross sections of Doll and others (1961) in a number of ways. First, for the reasons cited by Hatch (1985) and by Westerman (1985), as well as the stratigraphic topping data cited here, both boundaries of the trough are shown as faults. Second, not only the Northfield, but also the narrow belt of Northfield-like rocks in the middle of the western belt of Waits River are shown on figure 6 as stratigraphically above the Waits River and stratigraphically equivalent to the Gile Mountain. Third, the resulting configuration (fig. 6), as noted above, is more that of an anticlinorium than a synclinorium.

Figure 7 is an attempt to arrange the facies of the rocks in figure 6 into a depositional model. The distribution of coarse- and fine-grained rocks in figure 7 suggests an easterly source region, particularly for the Gile Mountain and correlative rocks. I further suggest that the apparent higher quartz content in the carbonate-quartz



Figure 4. Photograph of the thick-bedded micaceous quartzite facies of the Gile Mountain in the western part of the eastern belt. Pencil is about 6 inches long.



Figure 5. Photograph of the quartzite/metapelite facies of the eastern part of the eastern belt of the Gile Mountain. From cut on Route I-93 about 5 miles northeast of Barnet, Vermont. Coin is about one inch across.

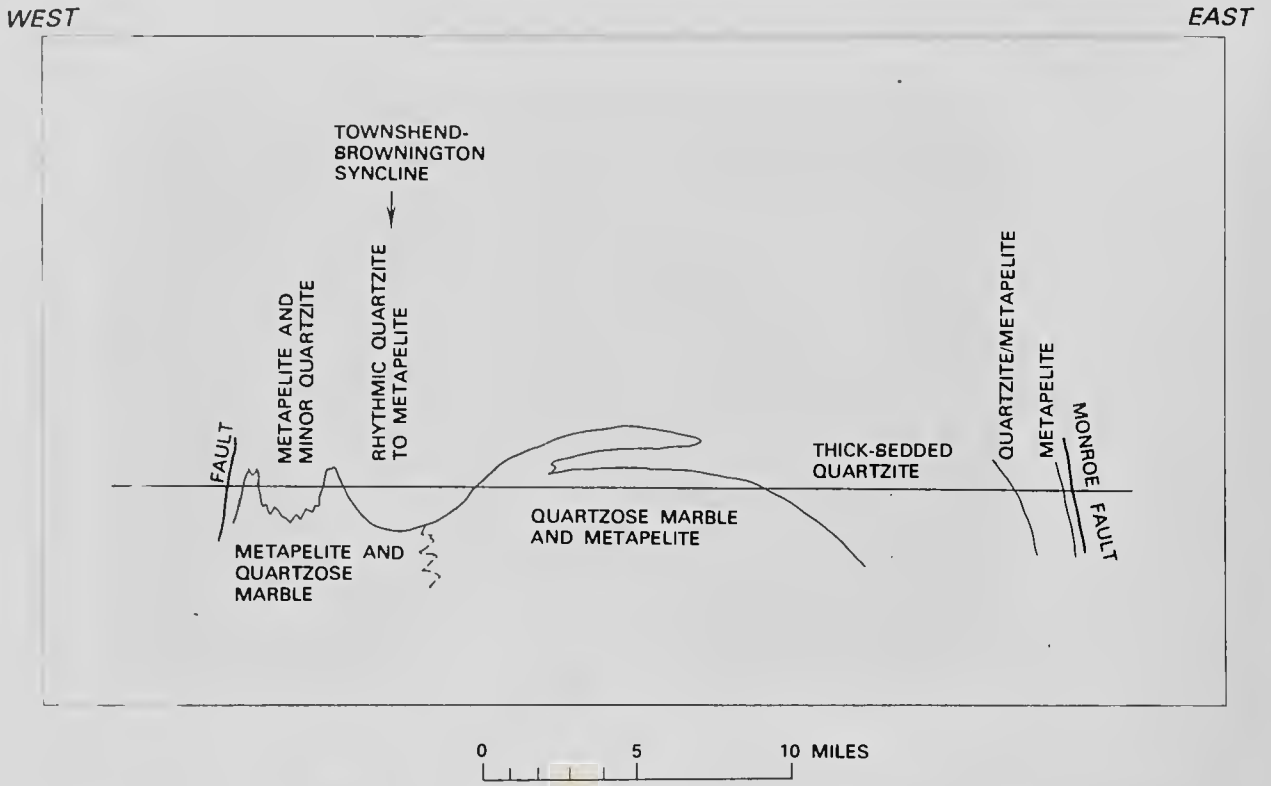


Figure 6. Cross section across the Connecticut Valley trough at about 44° showing proposed revised stratigraphic-structural model.

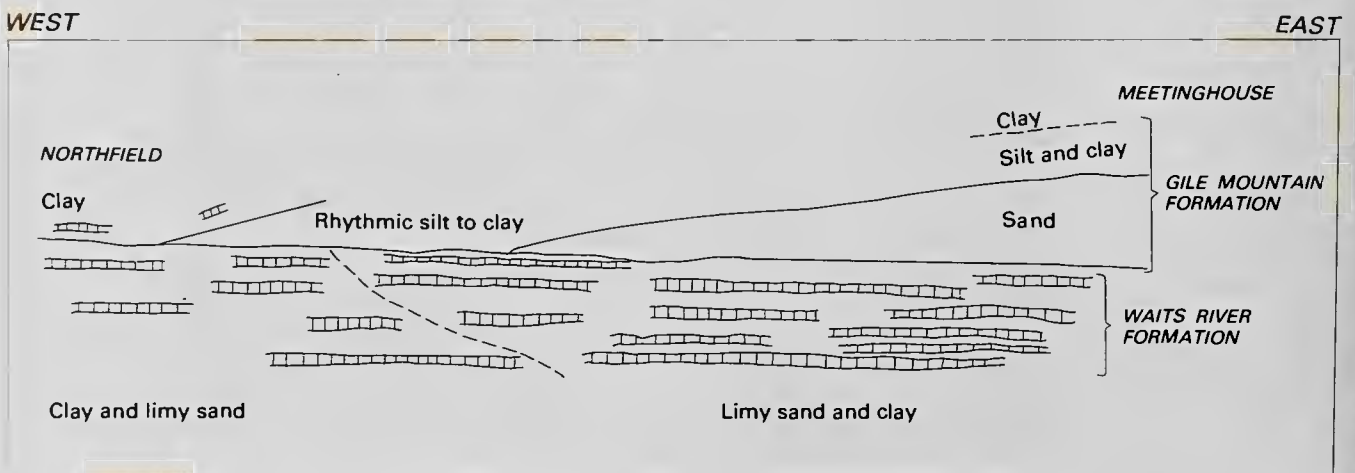


Figure 7. Unfolded and unmetamorphosed restoration of the cross section shown in figure 6.

sands of the eastern belt of Waits River and the lower alumina content of the shales of the eastern Waits River relative to the western Waits River rocks also could be interpreted to indicate an eastern source. The significance of the higher percentage of carbonate sands in the eastern Waits River is questionable, but the observed relations seem at least compatible with an eastern derivation.

Lengthy speculation about the exact location of this easterly source region for the Connecticut Valley trough sediments is beyond the realm of this guidebook article. Suffice it to say, however, that regardless of the exact age(s) of the trough rocks, if they are indeed post-Taconian and pre-Acadian, the presence of Early and Late Silurian and Early Devonian strata on the Bronson Hill anticlinorium severely constrains the times during which the Bronson Hill could be a source. Furthermore, the Bronson Hill contains no presently exposed or known source rock for the carbonate sands of the Waits River. And finally, present ideas on the Merrimack trough east of the Bronson Hill indicate that it was receiving sediment from a western source from the Late Ordovician on into the Silurian (Moench, 1969; Hatch, Moench, and Lyons, 1983). Whatever the answer to the question of the source area of the Connecticut Valley trough rocks, it is not simple.

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ROAD LOG FOR TRIP B-3

Pertinent maps:

Topo maps, 7.5-minute scale, in order of the trip:

Northfield
 Roxbury
 Randolph
 Bethel
 South Royalton
 Sharon
 South Strafford

Geologic maps:

Doll and others (1961) (State map)
 Ern (1963) (Randolph 15-minute quadrangle)
 Doll (1944) (Strafford 15-minute quadrangle)

The trip will assemble in the Norwich University (Northfield) parking lot immediately south of the Cabot Science Building. Because parking is limited at most of the trip stops, PLEASE consolidate into as few vehicles as possible. The trip will return to the parking lot at the end of the day.

LUNCHES: Bring a lunch or makings therefor. We do not plan to stop at any eatery or store.

A few stops will involve moderate walks along roads, but no strenuous traverses are planned.

The total trip will involve about 100 miles round trip, so drivers should please be sure to have sufficient gasoline before starting out.

Mileage

- 00.0 Norwich University parking lot.
 From parking lot turn right (south) onto Route 12.
- 00.1 Bear left on Route 12 at junction with Route 12a.
- 01.3 Intersection of Route 12 with brand new (post map) section of Route 64. Park on right shoulder of Route 12.

STOP 1 (Roxbury 7.5-minute quadrangle)

We will examine the new cuts at the intersection and east along new Route 64. At the intersection, the exposed rocks are typical of the Northfield Formation. The rock is medium-dark-gray, slaty, graphitic, aluminous phyllite, here at biotite grade. Although bedding is difficult to see on the fresh surfaces of the cut, the glacially polished surface on the top of the exposure shows a few beds an inch or so thick of lighter gray rock. This lighter gray rock is richer in quartz and poorer in micas and graphite than the darker phyllite and is interpreted to represent original silt-rich beds within the sequence of clay-sized sediment. The gradational and sharp boundaries of these beds of originally coarser (relative to clay) silt are interpreted to reflect graded bedding from which primary stratigraphic tops can be determined. A few thin (2- to 6-inch) beds of brown-weathering quartzose marble ("punky brown" of local jargon) are present.

To the east, up new Route 64, the rock is predominantly the same medium-dark-gray phyllite with local thin beds of lighter gray metasilt. Approximately 3,000 feet east of Rte. 12 the punky brown-weathering quartzose marble beds have increased gradually in number and thickness. Here they are as much as 3 feet thick and form about 10 percent or more of the section. Previous mappers (Richard Jahns and Walter White, unpublished manuscript map) and I call these rocks the Waits River Formation. The boundary between the Waits River and the Northfield is somewhere in the long cut through which we have just walked. It is clearly gradational by progressive eastward increase in quartzose marble in a sequence dominated by Northfield-type phyllite. The bedding style and the characteristic punky brown weathering of the quartzose marble beds can be seen best on the top surface of the very eastern end of the cut. These rocks characterize the western belt of the Waits River Formation, at least between Barnard and Hardwick. This will be our only formal stop in this belt. Although faint graded beds face both east and west through the long cut containing the Northfield-Waits River contact, those observed closest to the contact face west. If correctly interpreted, these tops suggest that the Northfield is stratigraphically above the Waits River, the opposite relation to that presented by Doll and others (1961) and all previous reports. Graded beds near the Northfield-Waits River contact to the south near Bethel (fig. 1) also suggest that the Northfield stratigraphically overlies the Waits River (fig. 2) (Hatch, 1986). Keep these relations and the faint graded beds in the Northfield in mind when we look at the rocks of the Gile Mountain Formation at Stop 2.

A second cleavage that strikes about N25°E and dips about 45°NW is locally well developed and is axial planar to relatively open folds that fold both bedding and earlier schistosity. The earlier schistosity can be locally demonstrated to be axial planar to isoclinal folds in bedding.

Cubes of pyrite about 1/2 inch across are abundant in a narrow zone about 3,900 feet east of Route 12. I have noted them elsewhere at about the same distance from the western margin of the trough. Their unfractured and undeformed character suggests that they formed relatively late, after development of the folds and cleavages. One interesting possibility is that they might be related to Mesozoic faulting.

Return to the cars and continue south on Route 12.

- 05.5 Parking area for Baker Pond boat access on the right.
- 05.7 Route 65 enters from left. Continue straight on Route 12.
- 11.0 Village of East Braintree.
- 16.3 Junction with Route 12A in Randolph Village. Bear left on Route 12 through village.
- 23.8 Follow Route 12 through Bethel Village.
- 24.4 At east end of Bethel Village, leave Route 12, which turns right to cross the White River, and continue straight ahead (southeast) on Route 7 EAST which came in from the right.
- 25.9 Troop E of Vermont State Police on right.
- 27.2 Pass under Route I-89.
- 27.9 Junction with Route 14. Turn right on Route 14.
- 28.6 Pass under railroad bridge.
- 29.4 Village of Royalton.
- 29.9 Pass under railroad bridge again.
- 30.4 Park in parking areas on right.

STOP 2 (South Royalton 7.5-minute quadrangle)

Outcrops are in the White River just south of the road. The rocks are very well bedded interbedded gray graphitic phyllite or schist and light-gray, fine-grained micaceous quartzite in roughly equal proportions (fig. 3). Most beds are graded in the same manner as those at Stop 1. These rocks were mapped by Ern (1963) and by Doll and others (1961) as the westernmost of the two belts of Gile Mountain Formation (fig. 1). This exposure and others in the immediate area provided the evidence upon which Fisher and Karabinos (1980) based their conclusion that this belt of Gile Mountain Formation stratigraphically overlies the Waits River Formation that bounds it to the east and west. As a result of two periods of folding, beds face both east and west in the exposures at this stop, but numerous excellent exposures along both contacts of this belt to the north and south between Barnard and southeast of Montpelier (fig. 2) all support the interpretation of Fisher and Karabinos (1980). Abundant structural data further indicate that the belt forms both a structural and a stratigraphic syncline.

If my reading of the graded beds near the Northfield-Waits River contact at Stop 1 and other locations is correct, the Northfield is correlative with the rocks here at this stop. I suggest that the faintly graded phyllite and quartzite at Stop 1 are a (slightly lower grade) more distal facies equivalent of the rocks we are standing on here.

Return to the cars and continue east on Route 14.

31.4 Turn left (north) onto Route 110 toward Tunbridge and Chelsea.

31.9 Park on right shoulder of road.

STOP 3 (South Royalton 7.5-minute quadrangle)

Cuts on both sides of the road are in the eastern belt of the Waits River Formation, east of the Gile Mountain rocks at Stop 2.

The rocks here are punky-brown-weathering quartzose marble, in beds as much as 10 feet thick, and graphitic quartz-mica phyllitic schist. Thick beds of schist commonly contain beds of marble 2 to 4 inches thick. Quartz veins averaging 2 inches in width and 1 or 2 feet in length are common and characteristic of the eastern belt of Waits River. Compare the rocks here with the rocks at the eastern end of Stop 1, particularly in terms of protolith and possible facies relations. Also note that these rocks, as is almost universally true throughout this eastern belt of Waits River, appear to be much more complexly deformed than the rocks to the west. To what extent is this apparent greater (more complex) deformation the result of more intense or complex folding and metamorphism and to what extent does it result from the greater ability, or propensity, of the carbonate rocks here to flow during deformation? I favor some of both.

Return to the cars and continue north on Route 110.

32.1 Turn left onto a narrow paved road that goes back south parallel to Route 110.

32.4 Rejoin Route 110 and go south.

32.8 Turn left (southeast) onto Route 14 South.

34.3 Park in parking area on right.

STOP 4 (South Royalton 7.5-minute quadrangle)

Outcrops of the eastern belt of the Waits River Formation are in the White River on the right.

The rocks here are gray, graphitic, medium-grained phyllitic mica schist and gray quartzose marble in beds 3 to 10 feet thick. The schist contains many quartz veins. The marble beds are not only complexly folded, but individual beds change in thickness along strike, suggesting flowage during deformation. Although the Waits River rocks at Stop 1 showed two periods of folding and cleavage, the deformation here seems more complex and intense than in the western belt. Note also the rusty coating on the schists, which seems to characterize the Waits River schists (particularly of the eastern belt) in contrast to the much less rusty, lithically similar schists of the Gile Mountain.

Return to the cars and continue southeast on Route 14.

35.2 Town line, enter Sharon.

35.9 Park in parking area on the right.

STOP 5 (Sharon 7.5-minute quadrangle)

Walk ahead (southeast) down Route 14 about 1,300 feet to a point just east of a small brook crossing under the road and go right to outcrops in the White River.

The rock here is dark-gray-green amphibolite assigned to the Standing Pond Volcanics by Doll (1944) and Doll and others (1961). This unit, which is generally only a few hundred feet thick, has played a key role in working out the structure of this part of Vermont (Doll, 1944; White and Jahns, 1950; Lyons, 1955). It is interpreted to be a metamorphosed mafic tuff and, thus, should represent a time surface in the midst of the metasedimentary pile. In this area, it occurs at or very close to the contact between the eastern Waits River and the eastern Gile Mountain, although it has been mapped locally at a significant distance from it (see, for example, Doll, 1944; Lyons, 1955), suggesting that the metasedimentary unit boundaries are at least locally time transgressive.

Return to the cars and continue southeast on Route 14.

36.8 Park on the right shoulder on the grass under or immediately east of Route I-89.

STOP 6 (Sharon 7.5-minute quadrangle)

Walk back to outcrops in the White River about 600 feet northwest of Route I-89.

The rocks here are distinctly brownish-gray quartz-feldspar-biotite-garnet micaceous quartzites and schists. A few beds of punky-brown-weathering quartzose marble are as much as 3 feet thick. These rocks were mapped by Doll (1944) and Doll and others (1961) as Gile Mountain Formation. Although definitive graded beds were not seen in the exposures along the White River, graded beds a few miles to the north (fig. 2) do suggest that these rocks are stratigraphically above the Standing Pond, in agreement with the traditional view.

Note, however, the differences between these rocks and the rocks of the western belt of Gile Mountain at Stop 2. Here, the sand to shale (quartzite to schist) ratio is significantly higher; the quartzite beds are much thicker; the rocks are more feldspathic; the schists contain much less, if any, graphite; the biotite content is much higher; and the overall aspect of the rocks is brown rather than gray. Yet the available structural and graded bed evidence indicates that both the rocks at Stop 2 and the rocks here stratigraphically and structurally overlie the intervening belt of Waits River. Note also that punky brown quartzose marble beds are more abundant and thicker here than at Stop 2. Finally, note that the Standing Pond Volcanics that separate the rocks here from the Waits River to the west have not been observed anywhere along either contact of the western belt of Gile Mountain with the bounding Waits River (Doll and others, 1961). Regardless of the restored original distance between these two belts of exposure (shown as about 25 miles by Doll and others, 1961, cross section C-C'), it seems unusual that none of the Standing Pond volcanic tuff is preserved along or near the contacts of the western Gile Mountain belt. These relations

are presumably trying to tell us something about the history of these rocks, but, to date, they have not spoken clearly enough. In any event, I suggest that, if the rocks here are stratigraphically correlative with the western Gile Mountain, they are a more proximal facies.

Return to the cars and continue east on Route 14.

37.7 Village of Sharon. Turn left (northeast) on Route 132.

37.9 Pass under Route I-89.

38.0 Park where you can near, but not obstructing, the Half Acre Motel.

STOP 7 (Sharon 7.5- minute quadrangle)

Walk up (west) the ramp toward Route I-89 North about 450 feet to the outcrops on the right. KEEP OFF THE RAMP ROADWAY

The rocks here are in the same belt, or tongue, west of the Strafford dome of Gile Mountain Formation as those seen at Stop 6. Here, beds of brown micaceous quartzite are as much as 3 feet thick and the intervening schists are very quartzofeldspathic and graphite free. Quartzose marble beds, although present here, are rarely more than a few inches thick. I interpret the protoliths of these rocks to have been coarser grained and more quartzofeldspathic than the protoliths of the rocks at Stop 2. They are mapped in figure 2 as the quartzite facies.

Return to the cars and continue northeast on Route 132.

40.8 Note gravel road to left to "High Lake", formerly known as Standing Pond. Would you believe the High Lake Volcanics? Continue northeast on Route 132.

42.2 Cuts on both sides of the road at the top of the rise are in the Waits River Formation in the core of the Strafford dome.

44.2 T intersection. Turn right (east) on Route 132 and proceed through village of South Strafford.

44.6 Route 132 turns left at the far end of the village--follow it.

46.5 The old Elizabeth Copper Mine is about 1/2 mile upslope to your right (south). Despite the mine having been closed for about 30 years, note the iron stain in the West Branch of the Ompompanoosuc River which you have been following.

48.9 Park close up on the right shoulder.

STOP 8 (South Strafford 7.5-minute quadrangle)

Outcrops are in the West Branch of the Ompompanoosuc River to your right. This locality is called Rices Mills on the South Strafford 7.5-minute and the old Strafford 15 minute quads.

The rocks here are brown to gray-brown, medium-grained micaceous quartzite and quartz-mica schist, with very minor quartzose marble. They have been mapped as Gile Mountain Formation, and I include them in

the same "quartzite" facies as Stops 6 and 7. Beds are relatively faint and are a few inches to a few feet in thickness. Graphitic pelites are rare to absent. Again, I suggest that these rocks are a coarser grained, more rapidly deposited, more proximal facies than the rocks of the rhythmically graded western Gile Mountain facies seen at Stop 2.

Return to the cars and continue southeast on Route 132.

52.3 Make two left turns within about 50 feet. Follow signs to Union Village and Union Village Dam.

52.8 Bear slightly left. Do not turn right through covered bridge.

52.9 Enter grounds of Union Village Dam (U.S. Army Corps of Engineers).

Follow paved road up to top of dam and park around the flag pole.

STOP 9. (South Strafford 7.5-minute quadrangle).

This impressive structure is the U.S. Army Corps of Engineers Union Village flood control dam on the Ompompanoosuc River.

The rocks to be seen here have been assigned by Doll (1944) and by Doll and others (1961) to the Gile Mountain Formation and to the Meetinghouse Slate Member thereof. They are at the extreme eastern edge of the Connecticut Valley trough; the Monroe "line" or fault (Eric and others, 1941; Hatch, 1985, in press) that bounds the trough on the east passes within a few feet of the covered bridge that you just passed in the village. In this study, these rocks are mapped within the quartzite/metapelite facies and the metapelite facies (fig. 2).

From the flagpole, walk about 400 feet northwest to a set of pavement outcrops of interbedded dark-gray slaty phyllite and light-gray micaceous quartzite. The beds of micaceous quartzite are mostly 2 to 4 inches thick, and some are graded. The rocks and their bedding style here are somewhat similar to the rocks at Stop 2, but are clearly very different from the rocks at Stops 6, 7, and 8. Yet the rocks at Stops 2, 6, 7, 8, and 9 have been mapped previously as undifferentiated Gile Mountain Formation (excepting the separation of the Meetinghouse). I interpret the rocks here to have been deposited originally as fine silt and mud, in contrast to the rocks at Stops 6, 7, and 8 that I interpret to have formed as sand and muddy sand. Differences between the rocks here and those at Stop 2 are primarily that the rocks at Stop 2 are more regularly bedded (more consistent bedding thickness) and more consistently graded.

Structures readily seen in these outcrops include isoclinal folds, minor faults, and cleavages. The dominant schistosity in the phyllites is parallel to the axial planes of the isoclinal folds that are believed to be the earliest folds. The axes of these folds are subhorizontal. Many beds near fold hinges have been offset a few inches or more along planes that parallel the axial planar schistosity. Similar faulting has been observed elsewhere near the eastern edge of the trough. I suggest that the faulting and isoclinal folding were roughly synchronous and that both formed in a compressional regime during early stages of the Acadian orogeny. I further suggest that the minor faults seen here may be indicative of larger scale Acadian thrusting along the Monroe fault

zone. Note the quartz veins that have been folded isoclinally with the beds, indicating that at least some of the quartz veins in these rocks formed early in the deformational history.

From these exposures, walk south about 500 feet to a vertical cut immediately north (upstream) of the concrete spillway dam. PLEASE do not cross over the spillway dam despite the tempting outcrops on the other side. Pertinent features can be seen on the north side of the dam. The rocks here are the same as those just seen, with the addition of some late structures. Conspicuous are kink folds and kink bands that deform both bedding and the dominant schistosity and appear to be the youngest structures in the rocks. The kink bands strike northeast, parallel to bedding and schistosity, but dip only about 40° west in contrast to the essentially vertical bedding and schistosity. Also present here is a 3-inch-wide zone of brittle crushing that parallels the early schistosity. Similar kink bands are present along the trend of the Monroe fault to the north and south, as well as along the Ammonoosuc fault to the northeast. Similar crushed zones are displayed very well to the north at the point where the Monroe fault crosses the new cuts for Route I-93, a few miles west of Concord, Vermont. All of these structures are interpreted to have formed as a result of Mesozoic extensional normal faulting, extending north from the Mesozoic basins along the Connecticut River valley in Connecticut and Massachusetts.

Walk up the slope, past the cars, and continue along the road across the top of the dam to outcrops at the east end of the dam. In this vertical cut, you can see more isoclinal fold hinges and more truncations along vertical planes parallel to the early schistosity. The hinges of the isoclines appear to plunge very gently north to horizontal. Note the presence here of some quartz veins that are significantly thicker (6 to 12 inches) than those at the west end of the dam. They cut both beds and schistosity and, thus, are later than other thinner veins here that parallel schistosity and may be of the earlier generation. Perhaps these later thicker quartz veins are Mesozoic.

A few hundred feet southeast down the now blocked-off road is an exposure in which two sets of kink bands can be seen dipping moderately northwest and southeast. This conjugate set suggests vertical compression, compatible with the vertical crushed zone at the west end of the dam and with Mesozoic extension. Continue about 500 feet southeast down the blocked-off road to a small exposure on the left of dark-gray, graphitic slate typical of the Meetinghouse Slate.

Continue south down the blocked-off road. Turn right on the paved road and follow it about 1,000 feet south. About 90 feet before the covered bridge (yes, the same one you saw driving in) is an outcrop on the right of greenstone with excellent slickenlines plunging directly down the subvertical slickenside surface that parallels the schistosity. This rock was mapped by Doll (1944) and by me as being immediately southeast of the Monroe fault. I interpret the slickensiding to be Mesozoic and related to the crush zone and the kink folds and kink bands that you have just seen. Continue through the covered bridge to the T intersection. Meetinghouse Slate is exposed at a number of small outcrops here. Turn right at the T and walk back up to the cars at the top of the dam.

This is the end of the trip.

For the fastest route back to Northfield, retrace the roadlog back to Stop 7 (Mile 38.0 on the roadlog) and turn right up the ramp onto Route I-89 North. Follow Route I-89 North to Exit 5 and take Route 64 west to Route 12 to Northfield Center and Norwich University.

For the fastest route south, go back down to Union Village and follow Route 132 South to Route 5 south to Norwich where you can pick up Route I-91 South which in turn leads to Route I-89 South.

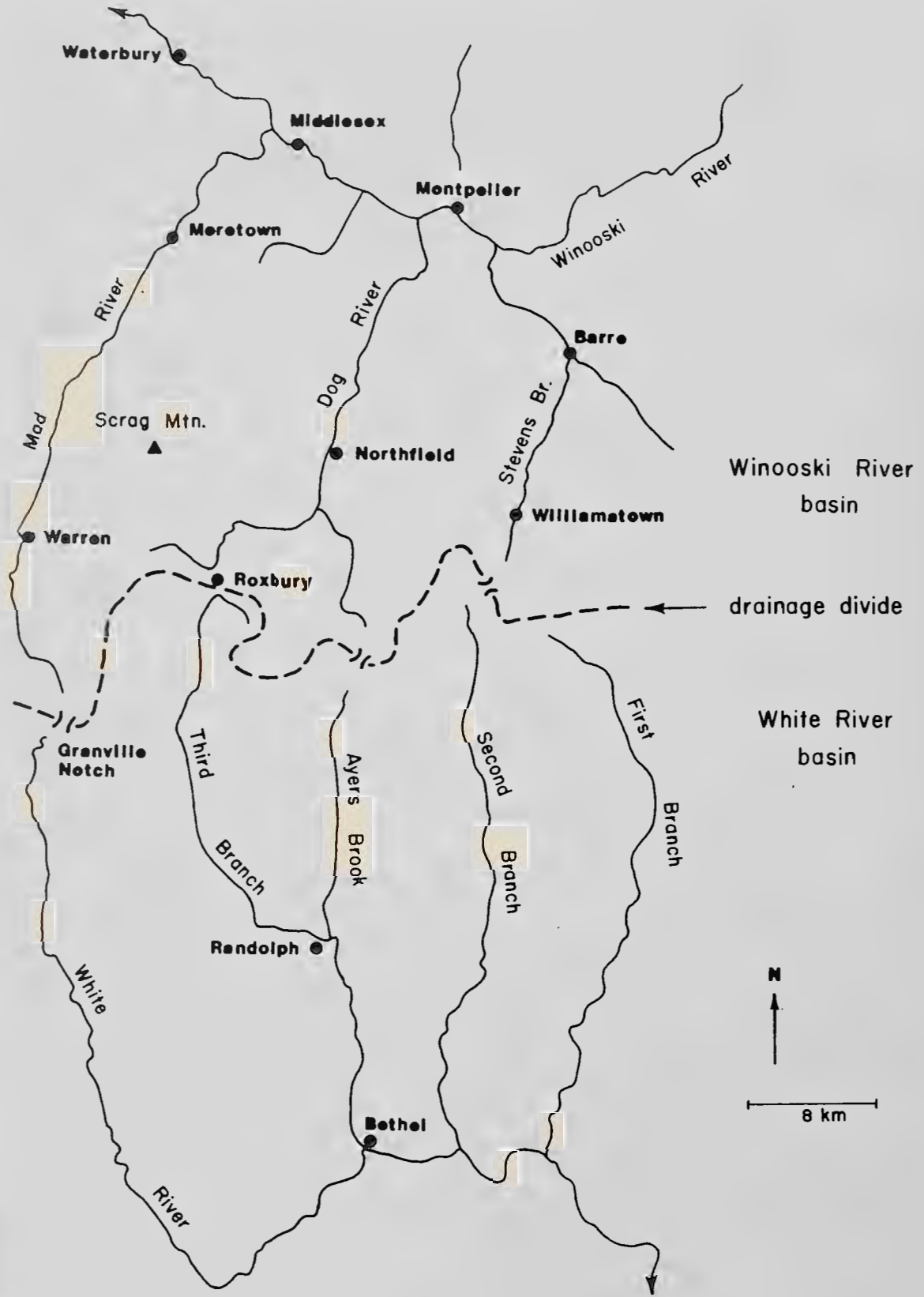


Figure 1. Stream drainage patterns in central Vermont.

HISTORY OF GLACIAL LAKES IN THE DOG RIVER VALLEY, CENTRAL VERMONT

by

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INTRODUCTION

The area traversed on this field trip lies on the Montpelier, Barre West, Northfield, and Roxbury 7.5-minute U.S. Geological Survey topographic maps in central Vermont. The terrain is underlain by metamorphosed eugeosynclinal rocks that are part of a complex zone of imbricate slices bounded by thrust faults. The age of rocks west of the Dog River is Cambrian and Ordovician. To the east the age of the rocks is under debate, but they are in the range of Ordovician to Devonian. The rocks in central Vermont were deformed during the Acadian orogeny and were intruded by Barre-type granites about 380 million years ago.

Erosion has produced a crude trellis drainage pattern characterized by alternating linear ridges and subsequent valleys that trend north-northeast/south-southwest (Fig. 1). Local relief is on the order of 300 meters (1,000 feet) with Scrag Mountain, elevation 887.5 meters (2,911 ft), being the local high point in the Northfield Range.

Critical to an understanding of the deglacial history of the area is the drainage divide separating north-flowing tributaries of the Winooski River from south-flowing tributaries of the White River (Fig. 1). In the larger view this divide separates the St. Lawrence River and Connecticut River drainage basins. In late-glacial time, low spots on the drainage divide acted as thresholds for proglacial lakes that formed after the ice margin had retreated north of the divide. The important thresholds that controlled local ice-marginal lakes are, from west to east: (1) Granville Notch, 430 meters (1,410 ft) ASL, (2) Roxbury, 308 meters (1,010 ft) ASL, and (3) 4.0 kilometers south of Williamstown, 279 meters (915 ft) ASL. A fourth threshold, located at the head of Brookfield Gulf (Ayers Brook) 12 kilometers south of Northfield, has an elevation of 436 meters (1,430 ft) ASL. A small proglacial lake that developed north of this threshold has no major deposits associated with it, and did not play a major role in the deglacial history.

The purposes of this field trip are to: (1) study glacial and postglacial sediments in order to document a sequence of down-dropping proglacial lakes and thereby establish the pattern of deglaciation, and (2) evaluate postglacial rebound as measured by Koteff and Larsen (1985 and in prep.).

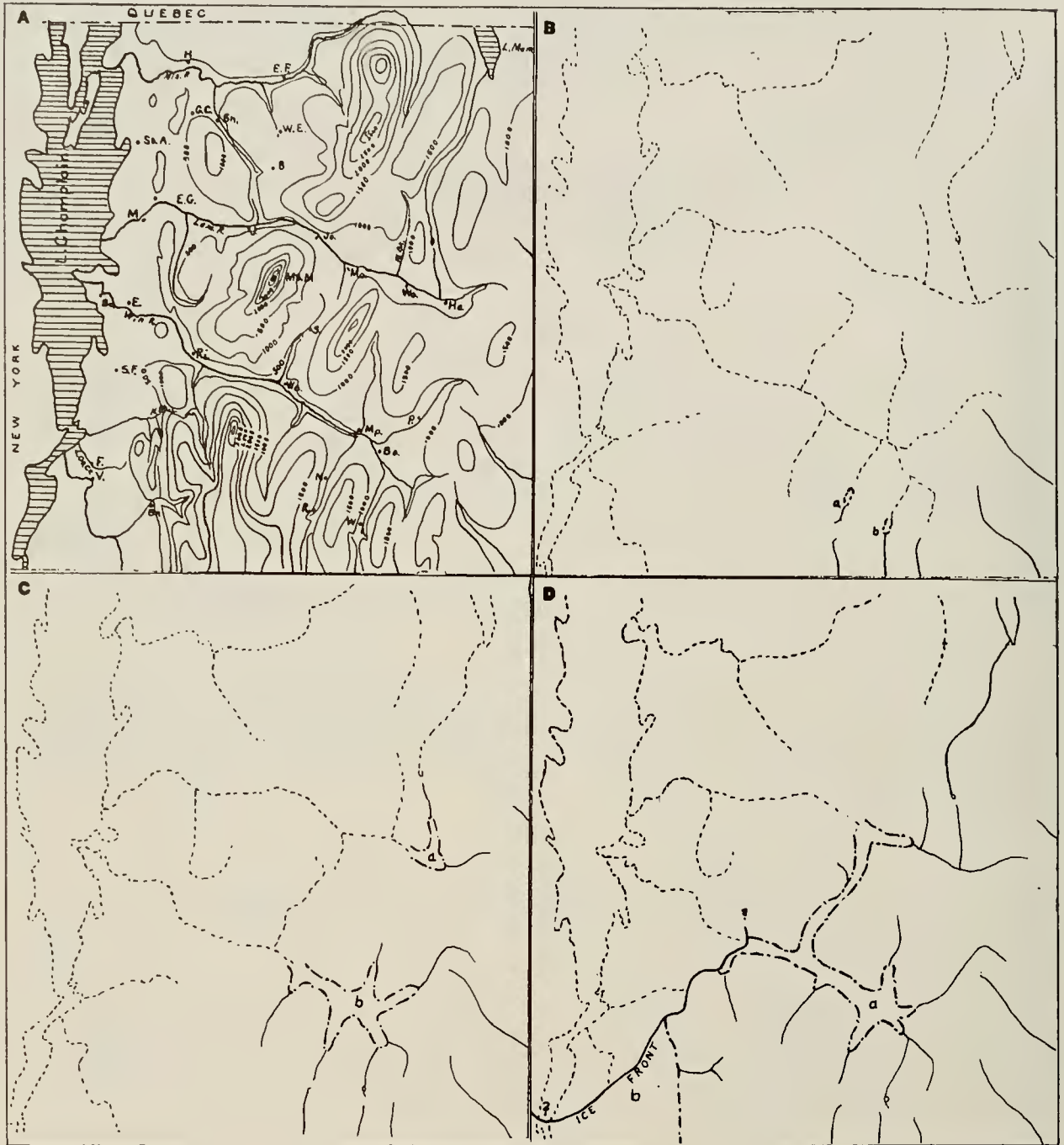


Figure 2. Portion of Plate XXI from Merwin (1908): (A) sketch map of northwestern Vermont; (B)-(D) lake stages in northwestern Vermont; (B) a, marginal lake south of Northfield (945 ft); b, Lake Williamstown (890 ft) discharging into the Connecticut River; (C) a, Second Lake Lamoille, b, First Lake Winooski at a stage represented by an altitude of about 745 feet at Plainfield,; (D) a, Lake Mansfield, b, Lake Vermont or Lake Albany.

HISTORICAL BACKGROUND

Merwin (1908) apparently made the first substantial study of glacial lakes in the Winooski drainage basin. He recognized that the ice margin retreated to the northwest resulting in a sequence of glacial lakes (Fig. 2). In his stage I, Merwin recognized two early ice-marginal lakes in the valleys of the Stevens Branch and Dog River. The lake in the Stevens Branch valley he named "Lake Williamstown", which "discharged into the Connecticut River" (Fig. 2B). He identified the lake in the Dog River valley as "a glacial lake south of Northfield". On the map, Merwin shows this lake draining north into the ice and not south into the Connecticut River basin as the evidence suggests. This latter feature has been named Lake Roxbury because the threshold of the lake is located in the village of Roxbury (Larsen, 1972). Merwin did not recognize a third lake of about the same age in the Mad River valley. That lake has been named Lake Granville because its threshold is located at Granville Notch (Larsen, 1972).

The name "First Lake Winooski" was used to describe an ice-marginal lake that extended from Middlesex to Plainfield and into the lower valleys of the Dog River and the Stevens Branch (Fig. 2C). First Lake Winooski was "represented by an altitude of 745 feet at Plainfield" (Merwin, 1908). It is not clear where the outlet for the lake was located, however, we can surmise that he thought that drainage was to the west on or through the ice. Today, it is apparent that, as long as ice blocked the Winooski drainage on the west, meltwater was forced to flow into the Connecticut basin by way of the Stevens Branch valley. In a third stage, Merwin recognized a "Lake Mansfield" that occupied the Stowe valley north to Morrisville and the Winooski valley from Barre to Jonesville. The outlet for Lake Mansfield was on the west side of the Green Mountains and into a glacial lake in the Champlain Valley.

Merwin worked only one summer (1906) and covered all of northwestern Vermont. His analysis suffered from a lack of modern topographic maps with good altitude control. Because of his early recognition of a logical sequence of glacial lakes in the Winooski basin, I have retained his terminology for three lake stages in central Vermont (Larsen, 1972).

Stewart and MacClintock (1969) did not follow Merwin, and devised a sequence of glacial lakes in central Vermont that, in my view, cannot be defended. Their sequence of diagrams (1969, Figs. 18-22) clearly requires 73 meters (240 ft) of erosion at the Roxbury threshold and 33.5 meters (110 ft) of erosion at the Williamstown threshold following ice retreat. This is in spite of the fact that, in each case, constructional ice-contact landforms located immediately north of each threshold are graded to the present-day threshold, and not to some higher level. The reader is invited to read Stewart and MacClintock (1969, p. 139-158) in order to make a judgement.



Figure 3. Proglacial lakes of Stage I: Lake Granville, Lake Roxbury and Lake Williamstown. Individual shorelines are based on a projection rising 0.90 m/km toward N21.5°W from the threshold of each lake. Solid triangles denote deltas graded to a lake level. In Figures 3-5 the ice margin damming each lake is shown as a short direct line for diagrammatic purposes. Tongues of stagnant ice probably occupied much of the area of each lake shown. (Explanation for Figures 3-5; B, Barre; H, Huntington; J, Jonesville; I, Irasville; M, Montpelier; Mi, Middlesex; Mo, Moretown; N, Northfield; NU, Norwich University; P, Plainfield; Ri, Riverton; Ro, Roxbury; W, Waterbury; Wa, Warren; Wi, Williamstown; Wo, Worcester.)

GLACIAL LAKES IN THE WINOOSKI VALLEY

In 1972, I proposed a three-stage sequence of glacial lakes in the Winooski valley (Larsen, 1972). In simplest terms, Stage I consisted of three separate ice-marginal lakes that drained south over separate thresholds, Stage II saw integration of those smaller lakes into one large lake draining south, and Stage III occurred when meltwater was permitted to drain west through the Green Mountains.

Stage III	Lake Mansfield II Lake Mansfield I			drained west
Stage II	Lake Winooski			drained south
Stage I	Lake Granville	Lake Roxbury	Lake Williamstown	drained south

STAGE I LAKE WILLIAMSTOWN, LAKE ROXBURY, LAKE GRANVILLE

Three proglacial lakes developed when the north-draining tributaries of the Winooski River, the Mad River, the Dog River and the Stevens Branch were dammed on the north by tongues of the retreating ice sheet. These lakes drained southward over thresholds located at: (1) Granville Notch, (2) Roxbury, and (3) 4.0 kilometers south of Williamstown (Fig. 3). From west to east, the lakes are named Lake Granville, Lake Roxbury and Lake Williamstown. Evidence that these lakes existed is shown by ice-contact deltas that are graded to the controlling threshold in each lake basin. A Lake Granville delta can be seen at Stop 3 (Hartshorn pit) on Field Trip C-1 (Ackerley and Larsen, this volume) and a Lake Roxbury delta will be seen on this trip at Stop 6. In this latter feature, the Roxbury delta, foreset beds, dune crossbedding, and ripple-drift cross-lamination all indicate transport to the south during deposition. The surface of the Roxbury delta is pockmarked with kettles on the north and was fed by a subglacial stream as indicated by an esker that extends 2.0 kilometers east-southeast from the head of the delta.

Retreat of the ice margin to the north resulted in north-eastward expansion of Lake Roxbury. At Riverton (Stop 2), south-dipping crossbeds in esker gravel and subaqueous outwash testify that Lake Roxbury existed at least 14.3 kilometers northeast of its threshold. Continued downwasting and retreat of the ice margins bordering Lakes Roxbury and Williamstown finally resulted in the lowering of Lake Roxbury by 24.4 meters (80 ft) to the level of Lake Williamstown. This occurred when the ice withdrew below the 1,040-foot contour on the ridge separating the valleys of the Dog River and the Stevens Branch, or earlier. The locality is on the Barre West quadrangle, 3.7 kilometers north of Berlin Corners.



Figure 4. Stage II. Approximate shoreline of glacial Lake Winooski based on a projection rising 0.90 m/km toward $N21.5^\circ W$ from the 279-meter threshold 4.0 kilometers south of Williams-town. Solid triangles denote meteoric deltas built during Stage II. (See Figure 3 caption for abbreviations of town names.)

STAGE II LAKE WINOOSKI

The draining of Lake Roxbury to the level of Lake Williamstown marked the beginning of Lake Winooski, a major proglacial lake that drained south over the Williamstown threshold. At its maximum extent, Lake Winooski extended throughout the valleys of the Mad River, the Dog River, and the Stevens Branch. It is not clear whether Lake Winooski ever extended northward into south-draining tributaries like the North Branch or the Kingsbury Branch. Careful inspection of the North Branch area north of Montpelier on aerial photographs and topographic maps reveals no delta or terrace on or near the projected shoreline of Lake Winooski. This implies that the North Branch valley was filled with ice when Lake Winooski existed. The upper Winooski valley and the Kingsbury Branch valley also may have been filled with ice when Lake Winooski existed.

Figure 4 is confirmation of the amount of postglacial rebound reported by Koteff and Larsen (1985, and in prep., see Field Trip A-3, this volume). In the Connecticut Valley, the Lake Hitchcock shoreline has been measured to be planar and to rise 0.90 m/km (4.74 ft/mi) toward N21.5°W. The uplift of the land was due to the removal of the weight of the ice sheet. Uplift appears to have been delayed until after the Champlain Sea incursion started (Fig. 8 and text, Field Trip A-3, this volume). Because Lake Winooski is older than the Champlain Sea, we can assume that the amount of postglacial uplift measured for Lake Hitchcock applies to central Vermont, and probably to much of western New England. Therefore, using the value of 0.90 m/km to N21.5°W, I projected a shoreline on the topographic map from the Williamstown threshold at 279 meters into the Winooski basin as an approximation of the shoreline for glacial Lake Winooski (Fig. 4). The projected shoreline falls right on the break-in-slope of 6 deltas in the Dog River valley and 4 deltas in the Mad River valley. The 10 deltas are all meteoric deltas that formed after the ice margin had retreated to the north permitting integration of the three lakes of Stage I into one Lake Winooski. The fact that 10 deltas fall right on the projected shoreline confirms both the existence and position of Lake Winooski and the value of postglacial rebound of Lake Hitchcock as measured by Koteff and Larsen.

Retreat of the ice margin that blocked Lake Winooski to the west-northwest down the Winooski valley can be documented at three localities: (1) on the south bank of the Winooski River 3.4 kilometers southeast of Middlesex, (2) at the Vermont Highway Department pit 2.0 kilometers north of Middlesex, and (3) north of the railroad 5.0 kilometers west-northwest of Waterbury. At these localities crossbedding in esker gravel and/or subaqueous outwash consistently dips to the east-southeast indicating that a subglacial meltwater source lay immediately to the west-northwest in the retreating ice sheet.

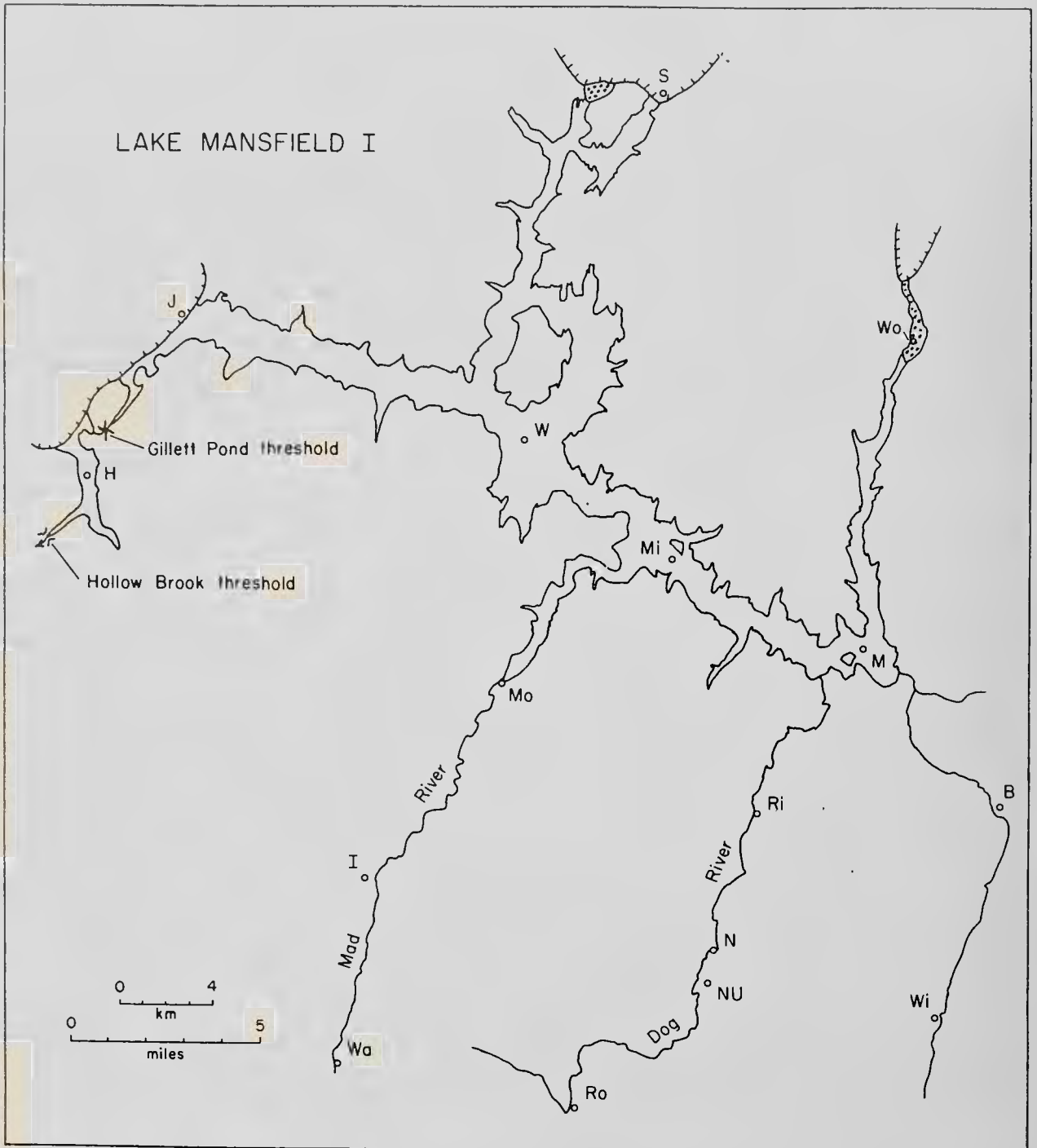


Figure 5. Stage III. Approximate shoreline of glacial Lake Mansfield I based on a projection dropping 0.90 m/km toward $S21.5^{\circ}E$ from the 229-meter threshold just southwest of Gillett Pond. Exact position of shoreline in the Winooski basin east of Middlesex is conjectural because older lake-bottom deposits that once stood higher than the projection probably have been removed by erosion. (See Figure 3 caption for abbreviations of town names.)

STAGE III LAKE MANSFIELD

When the ice margin retreated to Jonesville and Huntington, meltwater was able to escape to the west through the Green Mountains and south in the valley of the Huntington River (Fig. 5). Lake Winooski was lowered to the level of Lake Mansfield I and the 279-meter threshold south of Williamstown was abandoned. Lake Mansfield had more than one phase because there are several possible thresholds in the vicinity of Huntington ranging from 229 meters (750 ft) at Gillett Pond to 204 meters (670 ft) at the Hollow Brook threshold 3.0 kilometers southwest of Huntington. Wagner (1972) has identified three lakes that drained southwest through Huntington and ultimately over the Hollow Brook threshold. For simplicity, as it pertains to the early postglacial history of central Vermont, I recognize Lake Mansfield I (Gillett Pond threshold) and Lake Mansfield II (Hollow Brook threshold).

If we project a shoreline from the 229-meter Gillett Pond threshold with a gradient of 0.90 m/km down toward S21.5°E, that projection extends into the Dog River valley to Northfield Falls. However, Lake Mansfield I did not occupy the Dog River valley because older lake-bottom deposits lie at least 15 meters (50 ft) above the projection. Stream terraces cut into the older lake-bottom deposits fall on the projection, indicating that the early Dog River was flowing north at that time, or later. In the Mad River valley most of the major stream terraces fall on a concave-up profile that appears to join the above projected shoreline at an elevation of 213.4 meters (700 ft) at the village of Moretown (Fig. 6). On the east side of the Mad River valley, 2.7 kilometers northeast of Moretown, a small delta appears to be graded to the Gillett Pond projection (Fig. 6). Other fluvial surfaces that appear to be graded to this projection are 3.2 kilometers north of Montpelier and 4.0 kilometers west-northwest of Montpelier. Important large deltas at Worcester and 2.4 kilometers west of Stowe are about 14 meters (45 ft) above the Gillett Pond projection and probably were graded across ice, or to an unrecognized threshold.

Retreat of ice from the Gillett Pond threshold permitted Lake Mansfield I to drop 20 meters to the Hollow Brook threshold. A shoreline projected southeast from the Hollow Brook threshold, 20 meters below the Gillett Pond projection, intersects older lake-bottom deposits everywhere in the Winooski basin east-southeast of Waterbury. Therefore, Lake Mansfield II did not extend east of Waterbury and the Winooski River and all of its tributaries east of Waterbury were now on a fluvial grade to Lake Mansfield Phase II. The stream terrace to be observed at Stop 2 and other major terraces in the Dog River were part of that fluvial system (Fig. 7). Retreat of the ice margin from the lower Huntington River caused the draining of Lake Mansfield II to a lower lake thus permitting the Winooski River and its tributaries east of Waterbury, like the Dog and Mad Rivers, to erode down to their present levels.

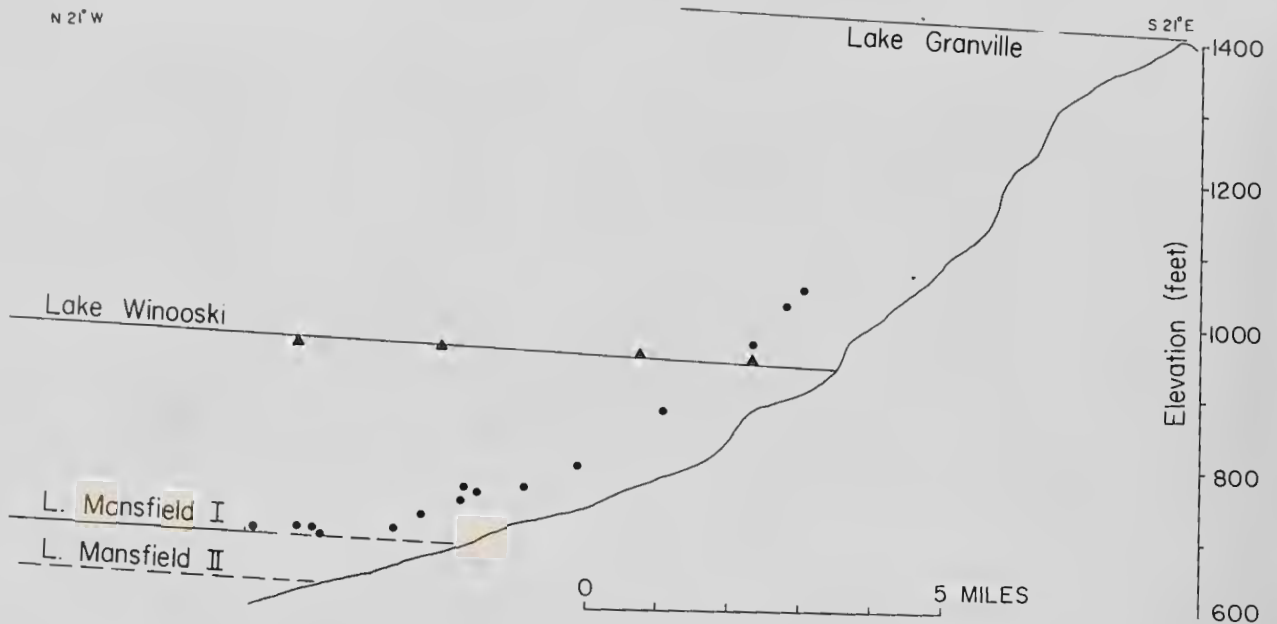


Figure 6. Projected profile showing elevation of major terraces (solid dots) in the Mad River valley. The levels of Lakes Granville, Winooski, Mansfield I, and Mansfield II are shown rising toward N21°W. Deltas built into Lake Winooski are shown by triangles. Terraces below the Lake Winooski level appear to form a single, concave-up, fluvial profile graded to glacial Lake Mansfield I at Moretown. Solid line at bottom denotes projected profile of the Mad River.

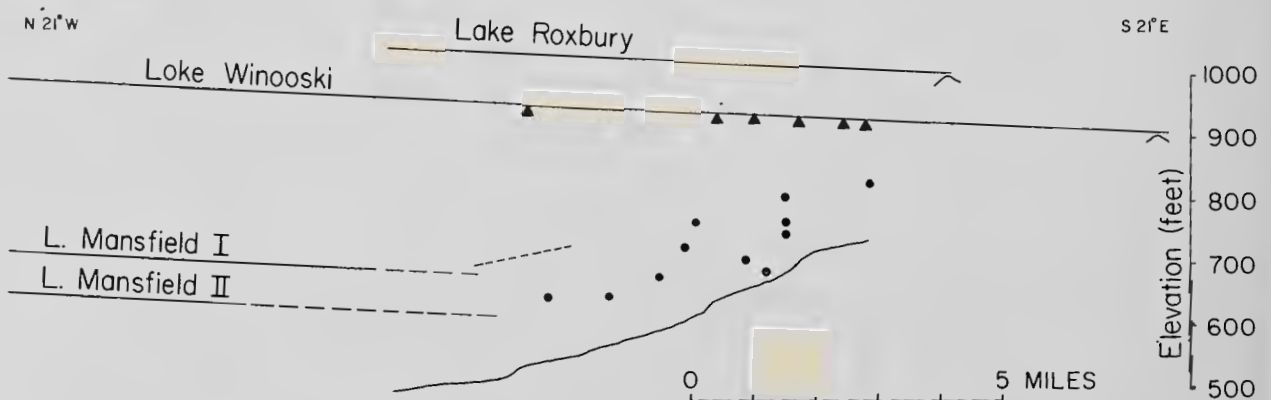


Figure 7. Projected profile showing elevation of major terraces in the Dog River valley (solid dots). Lake levels are shown rising toward N21°W. Lake Winooski deltas in the Dog River valley appear as triangles. Major terraces in the Dog River valley appear to be graded to both Lake Mansfield I and Lake Mansfield II. Solid line at base is projected profile of the Dog River. Line with short dashes is highest level of lake-bottom deposits.

QUATERNARY SEDIMENTS OF THE DOG RIVER VALLEY

Evidence for more than one glaciation has not been found during detailed mapping of the Dog River valley in the Northfield and Roxbury 7.5-minute quadrangles. A small exposure of saprolite represents preglacial material and will be visited at Stop 3. Three facies of till are recognized: (a) a greenish-gray sandy till on the west flank of the Dog River valley, (b) a gray silty till on the east flank, and (c) a gray clayey till in and near the valley bottom. The greenish-gray till is similar in color to underlying bedrock of the Moretown Formation. The gray silty till is similar in color to underlying bedrock of the Northfield and Waits River Formations. The gray clayey till is composed of deformed clay and silt layers resembling varves formed in a glacial lake. This suggests that the last ice sheet dammed a lake in the north-draining Dog River valley just prior to its advance over the region.

The following three sedimentary units are found in a fining-upward sequence at six different pits in the Dog River valley (Fig. 8). The sequence is nearly identical to that described by Rust and Romanelli (1975) for subaqueous outwash in the Ottawa, Ontario, region. The similarity between the sediments that they described in the Ottawa region and those in the Dog River valley is truly remarkable given the physical differences in the two respective environments of deposition. In the Ottawa area, deposition apparently was in a broad, open lake in a region of low relief, whereas in the Dog River valley, deposition was in a very narrow lake in a region of moderate relief.

Coarse-grained ice-contact deposits are common along the axis of the valley between Riverton and Roxbury. Poorly sorted pebble gravel with cobbles in south-dipping crossbed sets 2.0 meters thick commonly is interbedded with pebbly coarse to fine sand and silt. The deposits are confined to narrow elongate areas that are interpreted to be segments of eskers formed by subglacial meltwater streams. Till is rarely observed under the esker sediments and bedrock scoured by meltwater is common. The sediments are rarely collapsed except on their outside margins. The situation requires a subglacial meltwater stream capable of eroding and reworking till. The model of subglacial fluvial erosion described by Gustavson and Boothroyd (1982) for the Malaspina Glacier, Alaska, appears to be applicable here. In their model a groundwater table (potentiometric surface) rises upglacier from a proglacial lake and supplies the head required to drive meltwater and meteoric precipitation through a subglacial stream system. Such a system is required in order to explain the observations made, and to supply the large quantities of subaqueous outwash observed in the Dog River valley.

Overlying the coarse-grained ice-contact deposits are 1.0 to 5.0 meters, or more, of medium to very coarse sand and pebble gravel in trough-crossbed sets 10 to 50 centimeters thick. The dip direction of the crossbeds is highly variable, although gen-

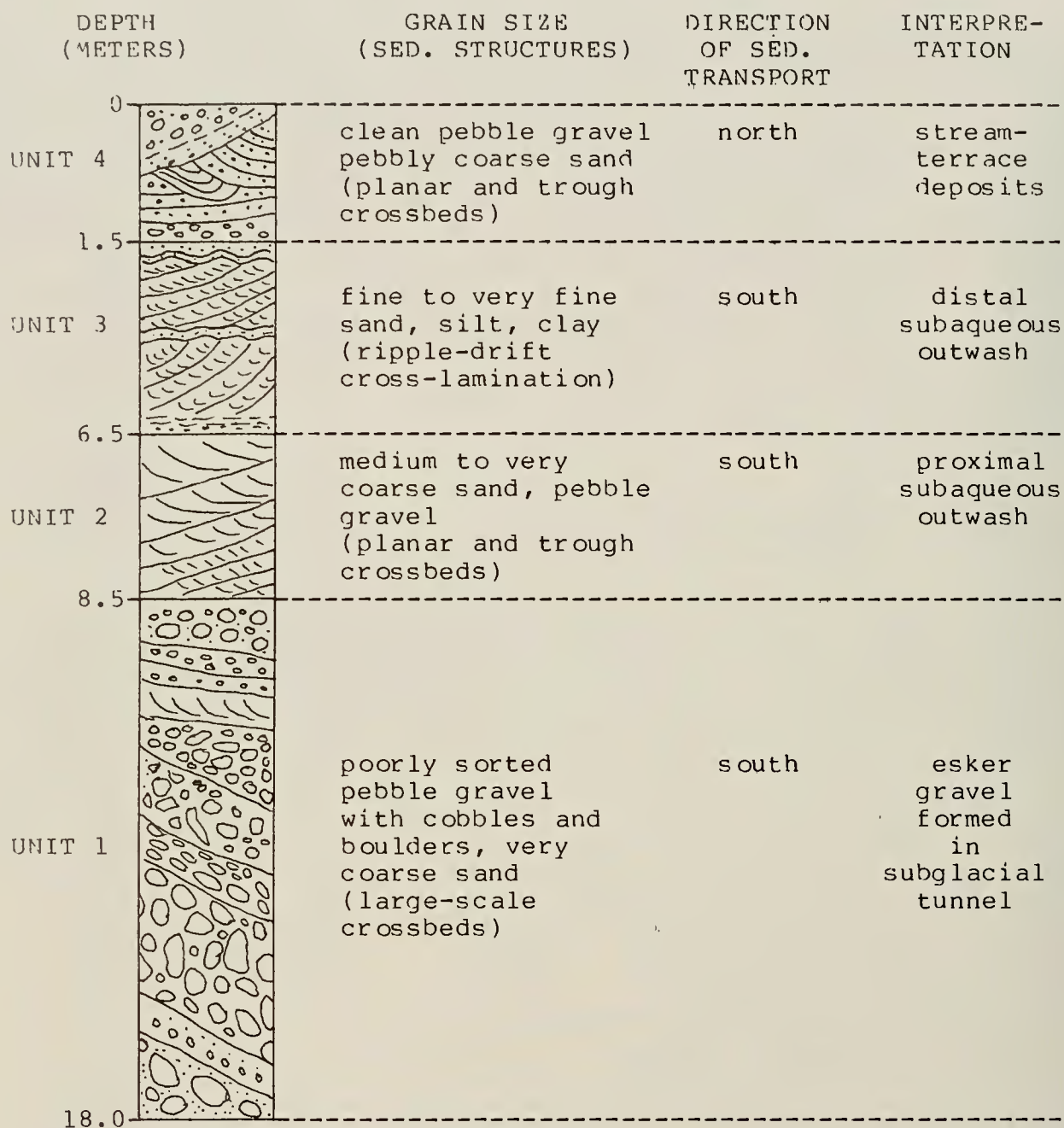


Figure 8. Generalized stratigraphic section for gravel pits in the Dog River valley. Units 1, 2 and 3 constitute a fining-upward sequence produced by northward retreat of an ice margin in a lake. Sediment is delivered to the lake floor by south-flowing turbidity currents issuing from a subglacial tunnel. Unit 4 was formed after the lake drained (Compare with Rust and Romanelli, 1975; depth scale is not linear).

erally it has a southern component. At some localities the dip direction is at right angles to the trend of the esker deposits. At Stops 2 and 8 large channels truncate well-bedded medium and fine sand. The channels are filled with structureless medium sand and contain clasts up to 20 centimeters long of very fine sand and silt. A broad U-shaped channel 16.5 meters wide and 2.0 meters high was formerly exposed at Stop 8 about 15 meters north of the esker gravel. The observations indicate that deposition of Unit 2 was by density currents close to the mouth of a subglacial tunnel. I characterize Unit 2 as proximal subaqueous outwash (Fig. 8).

Second to till, Unit 3 is the most widespread surficial unit in the Northfield 7.5-minute quadrangle. It underlies an area about 1.0 kilometer wide along the bottom of the Dog River valley. The deposits consist of fine to very fine sand, silt and clay that are often rhythmically bedded. In places, angular ice-rafted clasts are common. Ripple-drift cross-lamination formed by south-flowing turbidity currents indicates formation of Unit 3 as distal subaqueous outwash in south-draining Lake Roxbury. Fine-grained lake-bottom deposits, for example varved silt and clay, that might be associated with north-draining Lake Winooski are not common, or are difficult to recognize. A large portion of Lake Winooski bottom deposits probably were removed by post-lake stream erosion, because in many exposures stream-terrace gravels rest directly on fine sand with south-dipping crossbeds.

Unlike Units 1 and 2, which are relatively undeformed except for minor faults, Unit 3 displays various degrees of deformation. At two construction sites studied in 1979 and 1981 at Norwich University highly deformed (collapsed) ice-contact lake deposits were observed to grade upward through lesser deformed sediments into undeformed lake-bottom deposits. This indicates that lake-bottom sedimentation was contemporaneous with melting of buried ice, and suggests a somewhat different configuration of the ice margin than that shown by Rust and Romanelli (1975, Fig. 14) (Fig. 9).

Deltaic deposits of Lake Roxbury are exposed only at Stop 6 (Roxbury delta) where sediments range from pebble gravel with cobbles to very fine sand and silt. Crossbedding at all scales indicates transport to the south during deposition. Compared to the Roxbury delta, the younger Lake Winooski deltas generally are finer grained with clean pebble gravel in the topset beds and pebbly medium to coarse sand on the sides of the deltas.

Stream-terrace deposits, alluvial fan deposits, and alluvium are all similar in that they consist mainly of pebble gravel and pebbly coarse sand. They each display medium-scale crossbeds and imbricated flat pebbles that indicate down-valley transport of sediment after glacial Lake Winooski had drained.

SHAPE OF THE ICE MARGIN DURING RETREAT

A sequence of three fining-upward sedimentary units in subglacial and subaqueous outwash in the Dog River valley is nearly identical to that described by Rust and Romanelli (1975) in the Ottawa, Ontario, region. In two of their "Depositional models for subaqueous outwash in ridges at the ice front", Rust and Romanelli (1975, Fig. 14) show a retreating ice margin that is both straight and steep. In a third model they show a subglacial stream entering a lake at a reentrant formed by two straight and steep ice fronts.

There is an important distinction between Unit 3, distal subaqueous outwash, in the Dog River valley and fine-grained subaqueous outwash in the Ottawa area. In the Dog River valley, Unit 3 commonly is highly collapsed, whereas Units 1 and 2, esker gravel and proximal subaqueous outwash, show only minor faults. This suggests that Unit 1 was deposited on ice, whereas Units 1 and 2 were not. I believe that these observations require both a deep reentrant in, and a low surface gradient of, the subaqueous ice margin during retreat (Fig. 9). A crude and subaerial analog for a deep reentrant in the ice margin occurs today along the stagnant outer margin of the Malaspina Glacier where subglacial streams emerge as fountains that feed narrow ice-bounded outwash plains. If the edge of the Malaspina Glacier were to be submerged, and basal ice contained enough debris so that it would not float, then a subglacial stream would emerge at the head of a deep subaqueous reentrant in the ice margin. Coarse-grained material in traction (Unit 2) would be confined to the reentrant and fine-grained material in suspension (Unit 3) would be carried down-lake and spill out of the reentrant onto stagnant ice. Continuous melting of the stagnant ice on the sides of the reentrant during deposition of the fine-grained material would account for the progressive upward change from highly deformed to undeformed lake-bottom deposits found in the Dog River valley.

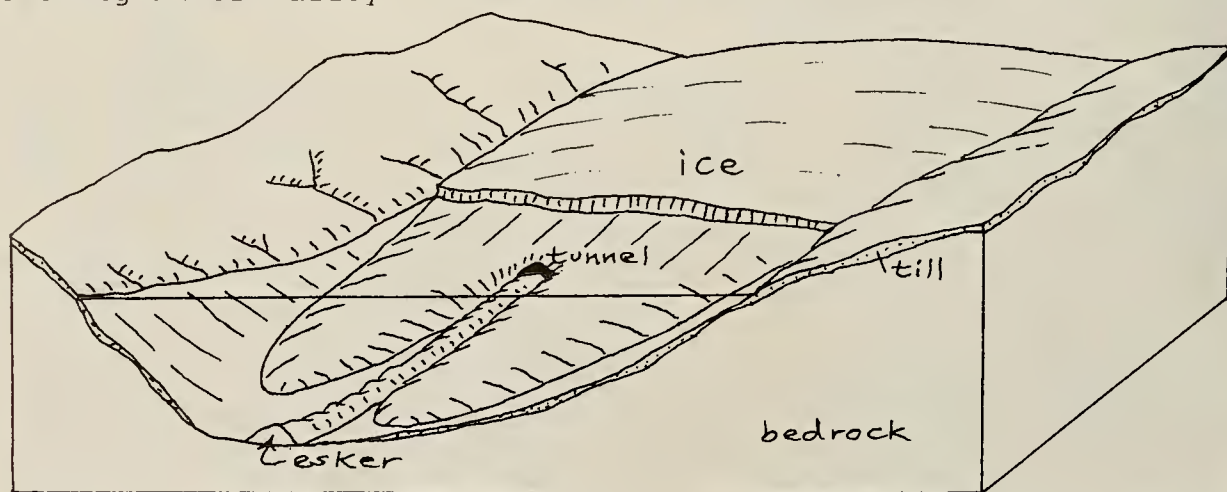


Figure 9. Proglacial environment of deposition for Unit 3, distal subaqueous outwash.

ACKNOWLEDGEMENTS

Mapping in the Northfield quadrangle was supported by the Vermont Agency of Environmental Conservation and mapping in the northern part of the Roxbury quadrangle was supported by the U.S. Geological Survey. Norwich University supplied logistical support, Brad Jordan drafted most of the figures, and dozens of Norwich University students have written reports about sediments in the area. Robert E. Schmidt, PLS, measured the topset/fore-set contact of the Lower Branch delta. Joseph H. Hartshorn, Byron D. Stone, Carl Koteff, and Doug Selden have all contributed ideas to this study. Maureen Larsen reviewed the paper and made suggestions for improvements. To all of the above I extend a hearty thank you.

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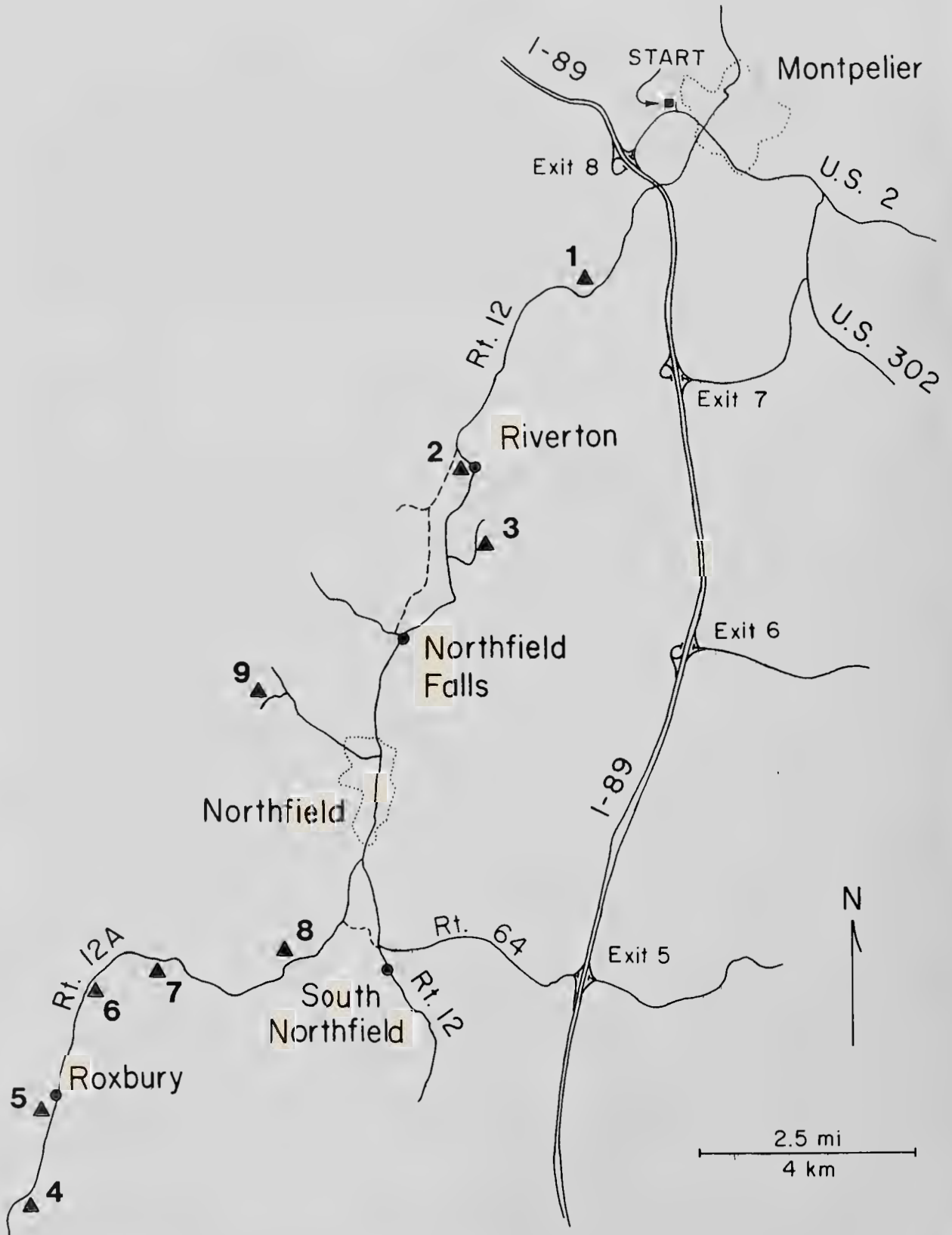


Figure 10. Route for field trip B-4. Solid triangles denote field trip stops.

ROAD LOG

START AT PARKING LOT AT MONTPELIER HIGH SCHOOL JUST NORTH OF U.S. ROUTE 2, AND 0.75 OF A MILE NORTHEAST OF EXIT 8 OF INTER-STATE I-89.

Mileage

- 0.0 Leave Montpelier High School, mileage starts at stop sign, turn right
- 0.02 At traffic light turn left (east) on U.S. Route 2
- 0.45 At traffic light turn right (south) on Route 12
- 2.25 Steel girder bridge over Dog River
- 2.90 Park on left at junction of Route 12 with Browns Mill Road. Walk southwest 125 feet, cross Route 12 and walk northwest up small dirt road, turn left (southwest) and walk about 750 feet along railroad to first outcrops.

STOP 1. ABANDONED WATERFALL

Three half potholes with depressions at their bases plus a fourth water-abraded surface, 0.6 meter in diameter, probably were formed by headward erosion by the Dog River. The largest pothole lies south of the railroad and faces north. The abraded surface is more than 2.0 meters high and measures 2.1 by 1.3 meters at the lip of a filled depression. The other features lie north of the railroad and face south. Twelve meters due north of the largest pothole is a depression 1.0 meter in diameter with an abraded surface that rises 4.4 meters up and to the right. Six meters to the east of the latter is a small depression with an abraded surface 3.4 by 1.3 meters in size. Because the half potholes are located at the east end of the railroad cut and because there are no potholes or abraded surfaces associated with the higher man-made cut to the west, one can presume that a waterfall migrated from northeast to southwest along the trend of the present railroad track. Further, the presence of loose, clean pebble gravel in the 1.0-meter pothole suggests that the pothole cutting took place in postglacial time after Lake Mansfield had drained. It appears that following the draining of Lake Mansfield the Dog River cut down through glaciolacustrine fine sand and was superimposed across the bedrock ridge for a short period of time before shifting position to its present course.

Proceed south on Route 12

- 4.75 Outcrop on the right of quartzite, phyllite and greenstone in the Ordovician Moretown Formation
- 5.85 Leave Route 12 and proceed straight on dirt road where Route 12 turns sharply to the left.

6.00 Turn left into pit area

STOP 2. BILL LYONS PIT.

The minimum depth of Lake Roxbury above the terrace level was about 117 meters (384 ft). The plan is to start at a small face on the west (A), move to the north pit (B), and then to the middle pit (C), and finally to the large south face (D).

(A) WEST FACE: A small curved face 8 meters wide and 3.0 meters high has, at the base, 2.7 meters of bedded fine to very fine sand with ripple crossbedding dipping to the south overlain by 0.3 meter of pebble gravel. Crude imbrication indicates transport to the north during time of deposition. This is a familiar scene repeated many times throughout the Dog River valley, that is, lake-bottom sediments formed by density currents moving south in glacial Lake Roxbury capped by stream-terrace gravels formed by a north-flowing stream.

(B) NORTH PIT: The exposure is 2 to 3 meters high in a curved face over 20 meters long. Fine to medium sand and pebbly sand occurs in west-dipping foresets. Within the foreset beds ripple crossbedding and trough crossbeds dip south. To the west the foreset beds dip to the southwest into the face. Above pebbly coarse sand in the foresets is fine to very fine sand which, in turn, is overlain by pebbly coarse sand in dune beds dipping south.

(C) MIDDLE PIT: The curved exposure faces north-northeast and is 9.3 meters high and more than 30 meters long. The face is the mirror image of Figure 12 in Rust and Romanelli (1975), and displays half of a collapsed syncline with laminated very fine sand and silt. Adjacent on the west is an 8-meter wide zone with nearly vertical normal and reverse faults. The faults are 0.1 to 3.5 centimeters wide and consist of fine sand packed with silt. West of the fault zone is uncollapsed fine to coarse sand with south-dipping crossbeds 0.5 meter thick. On the west in 1986, was a channel filled with structureless medium to coarse sand containing clasts of very fine sand and silt up to 20 centimeters long. The aspect of the channel, which was formed as a grain flow, was similar to that in Figure 6 in Rust and Romanelli (1975). Exposed at the top are remnants of stream channels filled with pebble gravel.

(D) SOUTH FACE: The exposure is in the shape of a U that opens to the northeast. At the southwest corner the face is 16 meters high with 10 meters of section exposed. The four main units, from bottom to top, are: (1) 2.8 meters of poorly sorted pebble gravel with cobbles and boulders, crossbeds up to 2.0 meters thick dip to the southwest; (2) 1.0 meter of pebbly very coarse sand and pebble gravel in trough crossbeds that dip to the west and southwest; (3) 5.0 meters of fine sand with ripple-drift cross-lamination alternating with beds of very fine sand and silt, a west-dipping curved surface truncates the bedding

above which similar fine-grained sediments are conformable. The surface is a slump scar or fault because small slump blocks of the lower sediment have been dropped down along the surface; (4) 1.2 meters of clean pebble gravel interpreted to be stream-terrace deposits. Unit 1 is interpreted to have been formed by a high energy stream flowing south in a subglacial tunnel. Units 2 and 3 constitute proglacial subaqueous outwash formed by density currents flowing south in glacial Lake Roxbury. Unit 4 was formed by the Dog River flowing north on a fluvial surface graded to the Hollow Brook threshold of Lake Mansfield.

Proceed out of pit, turn right (north) on Chandler Road

6.6 Turn right (east) on Route 12

6.85 Steel girder bridge over Dog River at Riverton

8.00 Ellie's Farm Market on left

8.10 CAUTION, turn left across traffic on Darling Road, proceed east 0.5 to 0.6 mile on narrow dirt road

8.6 or 8.8 Cars must turn around individually at one of two spots. Please do not block Darling Road entirely

STOP 3. DARLING ROAD SAPROLITE LOCALITY

This may be the first bona fide saprolite found in central Vermont outside of areas of the calcareous Waits River Formation. The exposure, 75 meters long and up to 3 meters high, was exposed during road improvements in 1983. The thickness of the saprolite is not known at this time. The material is a structured saprolite that can easily be dug with a shovel. Foliation in the saprolite is vertical and strikes N25°E. Fresh bedrock nearby to the north is a greenish-gray schist with crystals of albite and blue quartz in a matrix of quartz, chlorite, and muscovite. It has a strike of N20°E and a dip of 84°SE. The locality is anomalous in that it is on the east side of the Dog River valley and was exposed directly to glacial ice flowing from the north-northwest. Saprolite in glaciated terrain in Precambrian gneiss near Quebec City has been attributed to deep weathering along a fault (LaSalle and others, 1983). The Darling Road saprolite probably is associated with a fault (David S. Westerman, 1983, pers. commun.).

Return to Route 12

9.4 Turn left on Route 12, proceed south

10.6 Northfield Falls

11.4 During construction of Grand Union on left a section was exposed with stream-terrace gravels over collapsed and

truncated ice-contact deposits with south-dipping crossbeds

Cross Dog River, enter Village of Northfield

- 12.4 Cross Dog River near center of Northfield
- 13.4 Turn right off Route 12 at small park, possible rest stop
Proceed south on Route 12
- 13.7 Turn right on Route 12A toward Roxbury, proceed south and west up the valley of the Dog River
- 15.1 Pass under Harlow Railroad Bridge
- 16.0 Cross Dog River at Northfield Country Club
- 16.4 Segment of Northfield esker on left
- 17.2 Pit on left in Northfield esker (Stop 7)
- 17.8 Cross Dog River and pass under railroad
- 18.4 Pit on left in Roxbury delta (Stop 6)
- 18.6 Frontal slope of Roxbury delta on left
- 19.9 Threshold of Lake Roxbury in village of Roxbury (Stop 5)
- 20.2 In the outlet stream channel from glacial Lake Roxbury
- 21.1 Turn left, enter Vermont State pit, circle the wagons

STOP 4. MORPHOSEQUENCE AT VERMONT STATE PIT

This is a depleted pit that had at one time 2 meters of very poorly sorted pebble-cobble gravel with imbrication showing transport to the west overlying 3 meters of coarse sand with trough crossbeds climbing to the south. In the bedrock ridge just southwest of the pit is a small V-shaped notch that acted as the threshold for a small glaciofluvial morphosequence that was formed when the ice margin retreated north of the notch. The initial deposits, the coarse sand with trough crossbeds, were deposited in a stream that flowed through the notch. When the ice margin moved to the northwest and away from the rock ridge in which the notch is located, a lower threshold was established and the pebble-cobble gravel was graded to this lower outlet. At that time, the ice margin was located along the present site of Route 12A. With continued northward retreat of the ice margin, meltwater at this site was able to flow directly into what is now the Third Branch of the White River.

Bedrock in the pit is quartz-albite-sericite schist with a pinstripe texture that has been dismembered by tectonism. It belongs to the Ordovician Moretown Formation and has been neatly polished by meltwater scour.

Proceed north on Route 12A, at first low spot in road look northwest across Third Branch of the White River to abandoned serpentine quarry

22.3 Turn left into driveway of former railroad depot, now Roxbury Town Offices

STOP 5. THRESHOLD OF GLACIAL LAKE ROXBURY

The drainage divide between the north-flowing Dog River and the south-flowing Third Branch of the White River is located just east of Route 12A opposite the Town Office. When the margin of the Laurentide ice sheet retreated north of this divide and blocked the Dog River, glacial Lake Roxbury was formed. The elevation of the lake at the threshold was about 307.9 meters (1,010 ft).

Proceed north on Route 12A on the floor of Lake Roxbury

23.6 Cross the Dog River and ascend the frontal slope of the Roxbury delta

23.9 Park on right at entrance to pit owned by John Gross

STOP 6. ROXBURY DELTA

Foreset beds, dune crossbeds, ripple-drift cross-lamination, and imbricate structure all indicate southward transport of sediment during construction of the delta. Collapsed bedding, kettles, and an esker, which extends 1.9 kilometers down the Dog River valley give evidence for an ice-contact origin for the delta, which was built by meltwater streams flowing into glacial Lake Roxbury. The topset/foreset contact has not been accurately surveyed but it obviously was controlled by the elevation of the drainage divide at Stop 5.

Proceed north and east on Route 12A

25.1 Park on right at entrance to pit owned by Edward Allard

STOP 7. NORTHFIELD ESKER

Large-scale crossbeds in pebble-cobble gravel at the base dip to the northwest and suggest formation in a subglacial stream that flowed in a tunnel to the west up the valley and to the Roxbury delta. The section is a fining-upward sequence, similar to that at Stop 2, with pebbly coarse sand with dune crossbedding in the middle and with collapsed fine sand with minor ripple-drift cross-lamination at the top.

Proceed east on Route 12A

- 25.2 Terraces on the left are cut into a delta that falls on the projected shoreline of Lake Winooski (Fig. 4)
- 25.9 Pit on the right and south of the Dog River formerly displayed west-dipping dune crossbeds in coarse sand overlain by stream-terrace gravels with imbrication indicating flow to the east
- 26.3 Cross Dog River at Northfield Country Club
- 26.8 CAUTION, turn left across traffic into DuBois pit

STOP 8 (optional). DUBOIS PIT IN NORTHFIELD ESKER

Excellent stratigraphic sections have been available for study for the past year because of recent excavations. The section consists of the 3 major units described in Figure 8. At the bottom, Unit 1 is 16 meters thick and has interbedded pebble-cobble gravel and pebbly very coarse sand in large-scale crossbeds that dip to the west and southwest. Unit 2 has 2.0 meters of coarse sand in crossbeds and massive grain flows that formed as proximal subaqueous outwash. On the top, Unit 3 consists of 7.5 meters of very fine sand and silt that is faulted and collapsed. Bedding near the top of the section dips 40 degrees to the northwest. Unit 3 displays a much higher degree of deformation than that observed in Units 1 and 2.

Turn left on Route 12A, proceed north

- 29.1 Turn left (north) on Route 12
- 29.25 A small Lake Winooski delta is located 0.3 of a kilometer east-southeast of the junction of Route 12 and Stage Coach Road
- 30.1 A local landmark, the Pinnacle, is a Lake Winooski delta and is located 0.3 of a kilometer east of Route 12 at the Northfield Savings Bank
- 30.2 Cross Dog River
- 30.25 Turn left (west) on Water Street just beyond P C Market
- 30.5 Turn right on Union Street (becomes Union Brook Road), proceed west-northwest up valley of Union Brook
- 31.7 Turn left (southwest) on dead end road (Fel Fernandez)
- 31.8 Dam on right
- 31.9 Sheldon pit on right is Stop 9. Easiest procedure is to drive to end of road and turn around there

32.15 Turn around at Fernandez residence

32.4 Park on right

STOP 9. SHELDON PIT

Pit is in topset beds of a meteoric delta graded to Lake Winooski (Fig. 4). Walk about 120 meters due west on delta surface to exposure located down over bank. Deformed clay-silt varves are believed to be overlain by till, and to represent deposits in a proglacial lake that was overridden by the last ice sheet.

32.6 Turn right (east) on Union Brook Road

33.8 Turn left (north) on Water Street

34.05 Route 12, end of trip. Montpelier is 9 miles north on Route 12. The shortest route to I-89 is to proceed 2.0 miles south on Route 12, turn left on Route 64 and proceed east to Exit 5 of I-89

CRETACEOUS INTRUSIONS AND RIFT FEATURES IN THE CHAMPLAIN VALLEY OF VERMONT

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INTRODUCTION

On this field trip we will examine sites near Burlington, Vermont where alkalic dikes and fractures are exposed that could be related to Cretaceous rifting. The setting of these and many similar features is the structural (as opposed to sedimentary) basin of the Lake Champlain Valley, which invites comparison with younger and better-studied continental rifts, such as the Rio Grande Rift of eastern New Mexico or the Gregory Rift of eastern Africa. The validity of such an interpretation depends upon careful study of the tectonic history of the Valley, especially the timing of faulting and its relation to magmatism.

The Lake Champlain Valley between Vermont and New York is from 20 to 50 km wide and 140 km long between the northern Taconic Mountains and the Canadian border. Topographically, the Taconic klippe interrupts the southern Champlain Valley, but structurally the same valley terrane widens to connect with the northern Hudson Valley southeast of the Adirondack Mountains. The surface of the lake is only 29 m above sea level while the deepest part of the lake, near the western side, approaches 120 m below sea level. Many peaks of the Adirondacks to the west and the Green Mountains to the east rise above 1000 m, providing considerable relief to the Valley margins. The best exposures exist along the lake shores, some river and stream banks, highway cuts, and quarries. The glacial soils on many hillsides are thin enough to reveal bedrock rubble (including dike float), but much of the area is also covered by thick, glacial lake and marine sediments that make good farmland but effectively hide the bedrock and structures.

The six stops of this field log include outcrops of the three major igneous rock types - monchiquite and camptonite (varieties of volatile-rich alkali basalt), and bostonite (hypabyssal trachyte) - a bimodal association characteristic of many regions of intra-plate or rift volcanism. The timing of the faulting that produced the Champlain Valley may be constrained by physical or geographic associations with the Early Cretaceous (115-135 Ma) magmas. Although no intersections of major-displacement faults with dikes are clearly exposed in the Champlain Valley, there are good outcrops at these six stops that show minor faults and fractures believed (or hoped) to be related to the major tectonic events.

Dikes and other intrusive rocks

Figure 1 (modified from McHone and Corneille, 1980) shows locations for many of the igneous intrusions of the central Lake Champlain Valley. As shown by other regional maps (McHone, 1984), dikes become scarcer southward from Vergennes, but are fairly abundant across the Taconics and Green Mountains to the west and southeast of Rutland, Vermont. The northern Champlain Valley is

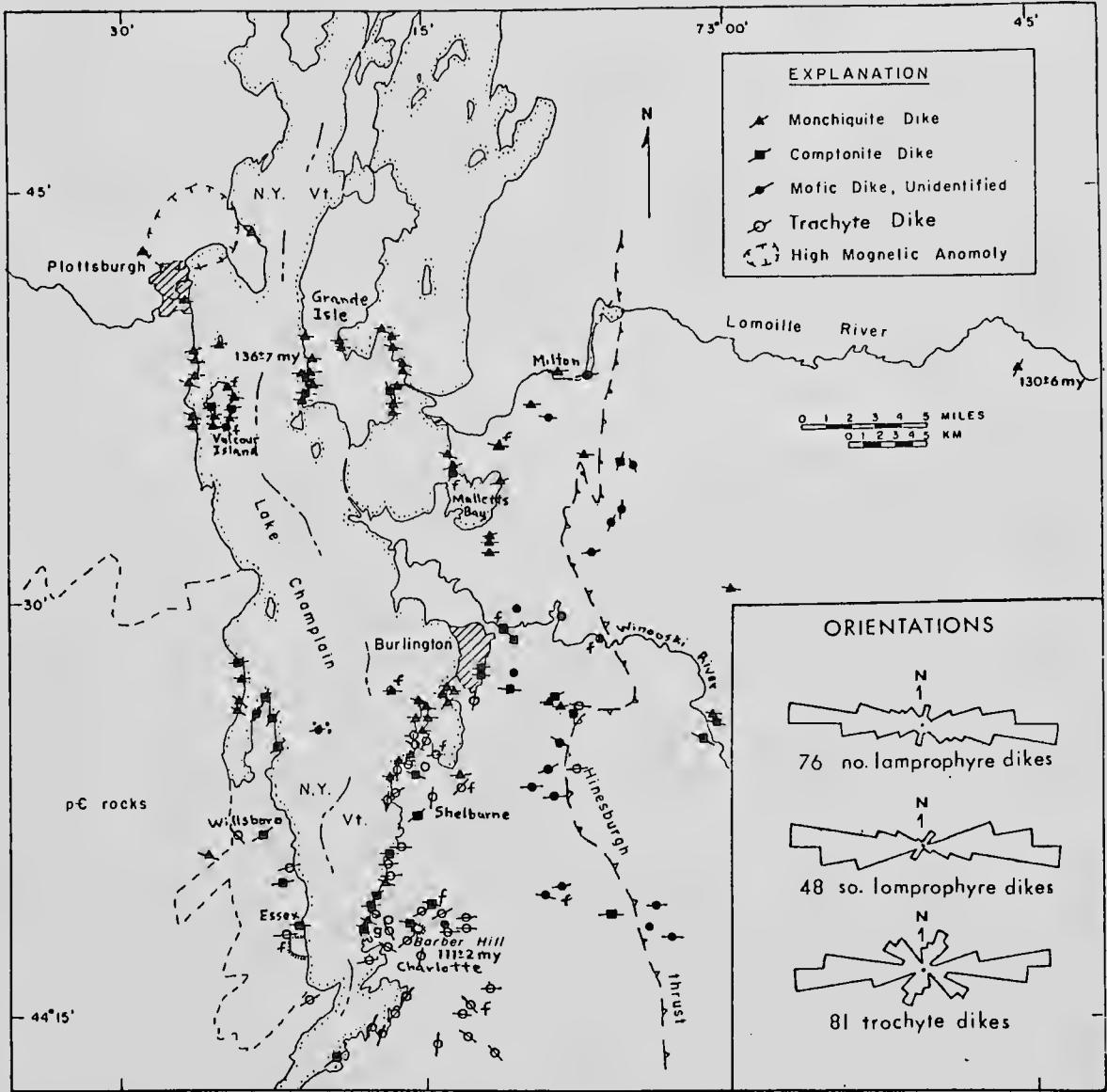


Figure 1. Locations and orientations of alkalic dikes in the central Lake Champlain Valley of Vermont and New York (after McHone and Corneille, 1980). F = fault-related dike (Table 1).

curiously devoid of dikes from North Hero well up into Quebec, but similar dikes are again abundant in the Monteregian Hills province ESE of Montreal. Lamprophyre dikes occur with lesser frequency westward as far as the east-central Adirondack Mountains, and are scattered but continuously present eastward across Vermont, New Hampshire, and the southern half of Maine.

Special studies of Champlain Valley igneous rocks start with early publications by Thompson (1860) and Hitchcock (1860), followed by petrographical and theoretical work by Kemp and Marsters (1893), Shimer (1903), Alling (1928), Hudson and Cushing (1931), Laurent and Pierson (1973), and McHone and Corneille (1980). Other geologists who mention or describe dikes as part of regional mapping studies are listed in the references section of this paper. The Champlain Valley intrusions are now well located and studied petrographically, and a small number have been chemically analyzed by Kemp and Marsters (1893), Laurent and Pierson (1973), and McHone and Corneille (1980). According to Ratte' and others (1983), several unpublished thesis studies of the dikes are known, although they vary in availability and therefore usefulness.

As indicated on Figure 1, the dikes apparently separate into two swarms across the lake into New York. At least 80 dikes of monchiquite, a few camptonite dikes, and no trachyte types are found in the northern swarm across Milton, Malletts Bay, southern Grande Isle, and the Plattsburgh area (Shimer, 1903; Fisher, 1968; McHone and Corneille, 1980). The monchiquite dikes lack significant feldspar, and are commonly rich in Ti-augite, calcite, kaersutite, and/or olivine or phlogopite, with analcime in the matrix. Some of the dikes approach alnoite or carbonatite in composition. Although we will not visit the northern swarm because of time constraints, monchiquite dikes in this area are exposed along Route 2 west of I-89 (Table 1) and in several roadcuts along I-89 east of Malletts Bay. Diment (1968) has outlined a strong geophysical anomaly east of Plattsburgh that probably is caused by an unexposed gabbroic pluton, similar to some plutons of the Monteregian Hills in adjacent Quebec.

Monchiquite, camptonite, and all of the trachytic dikes occur in the southern swarm (over 150 dikes) across from Burlington and Charlotte, Vermont to Willsboro and Essex, New York (Fig. 1). The trachytic dikes are commonly called bostonite (fine-grained, alkali-feldspars in clusters), despite the wide variations of beige, brown, and red colors with anorthoclase, albite, and quartz phenocrysts in an altered felsic matrix. Some show corroded oxybiotite grains. Camptonite has more plagioclase (restricted to groundmass) than analcime, much augite and rarely olivine, and variable amounts of kaersutite (a Ti-rich variety of brown hornblende). Amygdules and ocelli (formed as immiscible felsic-magma blebs) are common in camptonite. Phenocrysts of augite and kersutite are visible.

For most of the Champlain Valley, east-west to N80W dike trends are the rule (Fig. 1). Northeasterly trends are more common in New York and also to the east in Vermont. The trachytes show much more variation, especially near the Barber Hill stock in Charlotte, where Gillespie (1970) observed a radial pattern. A massive trachyte sill, covering a square mile or more, is exposed at Cannon point and inland south of Essex, New York (Buddington and Whitcomb, 1941). Trachyte dikes, sills, and intrusive breccias are abundant in southern Shelburne Point and probably indicate another syenitic pluton at shallow depth.

Intrusion ages

The few radiometric dates for Champlain Valley igneous rocks compare well with Early Cretaceous dates of the Monteregian Hills of adjacent Quebec, and for other intrusions of the New England-Quebec igneous province of McHone and Butler (1984). McHone (1984) summarized radiometric ages of northern New England dikes, including two for local lamprophyre dikes. Zartman and others (1967) found a Rb-Sr age of 136 \pm 7 Ma, using phlogopite from a dike of lamprophyre (ouachitite or monchiquite) on the western shore of Grande Isle. Using kaersutite separated from a monchiquite dike located about 35 km to the east (in the Green Mountains), McHone (1978) obtained a K-Ar age of 130 \pm 6 Ma. To the south, in the northern Taconics west of Rutland, Vermont, camptonite dikes have dates of 105 \pm 4 Ma and 110 \pm 4 Ma (McHone, 1984). In the eastern Adirondacks, Isachsen (1985) verbally reported K-Ar dates of 113, 123, and 127 Ma on camptonite dikes, and 137 and 146 Ma on dikes that apparently are monchiquite.

Armstrong and Stump (1971) reported a K-Ar date of 111 \pm 2 Ma for the syenitic Barber Hill stock at Charlotte, using a mis-acknowledged sample provided by Gillespie (1970) of "slightly altered" biotite. The Barber Hill stock is considered to be cogenetic with the bostonite (trachyte) dikes of the area. Seven bostonite dike samples fall along a whole-rock Rb-Sr isochron of 125 \pm 5 Ma (McHone and Corneille, 1980). Partial Rb-Sr data collected by Fisher (1968) for the Cannon Point trachyte sill, across the lake at Essex, New York, indicated an age of "less than 140" Ma, but also fits onto the 125 Ma isochron. Isachsen (1983, pers. comm.) has found a K-Ar age of 120 Ma for a trachyte dike in Willsboro, New York. The 111 Ma date for Barber Hill could be reinterpreted as a cooling date, about 10 Ma later than intrusion.

Two monchiquite dikes are crosscut by a bostonite dike and a bostonite sill along the shoreline SW of Shelburne Point (Kemp, 1893; Welby, 1961). Welby (1961, p. 188) reported that a camptonite dike crosscuts the Barber Hill syenite "near the crest of the hill at its northwest corner". In combination with the radiometric data, these crosscutting relationships are consistent with ages generally about 135 Ma for monchiquite, 125 Ma for trachyte/bostonite/syenite, and 115 Ma for camptonite, plus or minus 5 to 10 Ma for each type. These ages use old radiometric constants, and dates recalculated to new constants are 3 to 5 Ma older but do not change the age relationships.

Faults and faulting

Published maps of the Champlain Valley by Hudson (1931), Quinn (1933), Welby (1961), Doll and others (1961), Fisher (1968), and Isachsen and Fisher (1971) all seem to show different high-angle faults. Figure 2 is a somewhat generalized summary map, omitting some of the more imaginative faults suggested by Hudson (1931). Quinn's 1931 dissertation work included calculations of the percentage and directions of crustal extension caused by normal faulting in the region, and are reproduced in Figure 2. The north-south St. George fault system along the eastern side of the Valley is described by Stanley (1980), partly based upon his highly-valued student mapping projects. Stanley and Sarkesian (1972) and Stanley (1974) have made careful analyses of joint and fault strains, quartz lamellae orientations, and other structures to interpret stress patterns for these features.

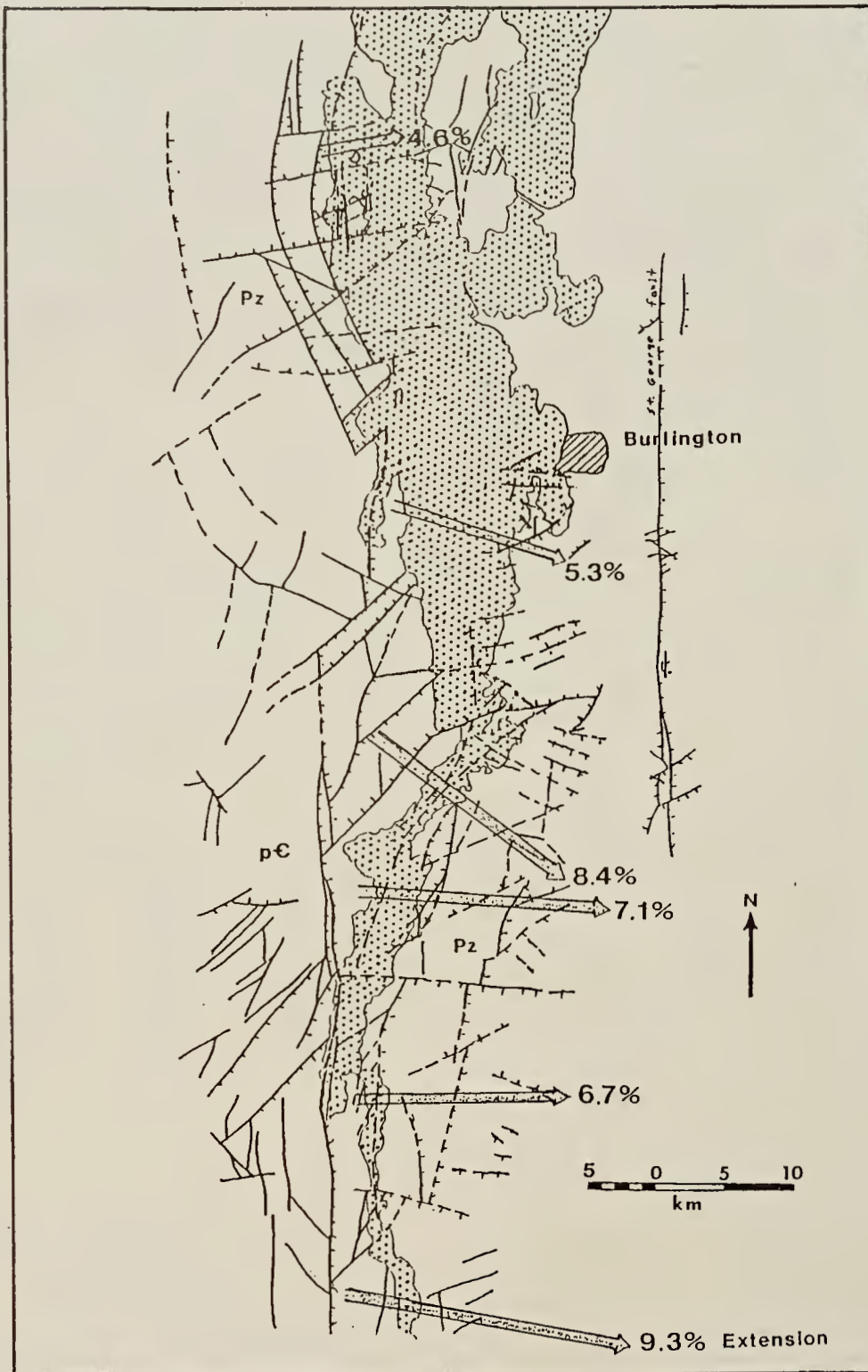


Figure 2. High-angle faults of the central Lake Champlain Valley, adapted from Quinn (1933), Welby (1961), Fisher (1963), Isachsen and Fisher (1970), and Stanley (1980). Extension vectors from Quinn (1933).

Most of the high-angle faults can be grouped as "longitudinal faults" (roughly N-S) or "cross faults" with both E-W and NE-SW trends. Geologists who originally mapped these structures related the faults to Paleozoic tectonic events, perhaps associated with westward movements of the Champlain and Hinesburgh thrust faults (Taconic/Acadian). Welby (1961) makes it clear that many cross faults have offset the Champlain thrust, and that cross faults also offset longitudinal faults (e.g. Stanley, 1980). The fault pattern in New York (Fig. 2) appears to show that several of the major longitudinal Adirondack border faults (Isachsen and Fisher, 1971) crosscut NE-trending Adirondack faults. Several longitudinal and cross faults of the eastern Adirondack border are well exposed, such as at Port Henry (McIlhene, 1987).

The high-angle faults are usually described as having dip-slip or normal offsets, but Stanley and Sarkesian (1972) and Stanley (1974) cite evidence for strike-slip or wrench movements of cross faults in the Shelburne Bay area. The faults have brittle features, and apparently moved at shallow (less than 2 km) depth (Stanley, 1974). Because vertical offsets of at least 850 m are preserved along some of the faults, roughly 2 to 3 km of post-faulting erosion of overlying rock is indicated. If the stratigraphy of the Champlain Valley was once complete with Silurian and Devonian or younger units, much uplift and erosion must have preceded faulting.

Isachsen (1975) and Isachsen and others (1983) argue that the Adirondack dome could be young, perhaps with Holocene uplift. The relief of the Adirondacks relative to the Champlain Valley is clearly based on movements along faults that are part of the Champlain Valley system, and so Burke (1977) proposed that the Champlain and adjacent Lake George valleys are grabens developed by Neogene continental rifting. Crough (1981) proposed uplift during Cretaceous-early Tertiary time for the Adirondacks and New England, based on fission tracks, stratigraphic arguments, and the presence of the Cretaceous intrusions, and he promoted a "hotspot track" model for the events. The absence of Triassic and Jurassic sediments in the Champlain Valley could indicate a higher elevation for the area during the Early Mesozoic, when Atlantic rifting produced large, deep sedimentary basins in southeastern New England and offshore (McHone, 1982). Finally, the present Champlain topography must predate the Miocene Brandon lignite and kaolin deposits, preserved along the eastern margin of the Valley by the Green Mountain front.

With the ages of the Champlain intrusions fairly well known, crosscutting relationships with the faults are critical to the tectonic model. Unfortunately, no intersections of dikes with major faults are clearly exposed, although a few are close! Locations where dikes intersect Champlain faults are listed in Table 1, and both pre- and post-dike fault movements are indicated. A major problem is estimating true offset, because at several of the exposures only minor apparent lateral offset may result from a great deal of mainly dip-slip motion. Future work with a portable magnetometer could help to show the nature of fault intersections with mafic dikes under shallow cover. Faulting and dike intrusion may be part of the same rifting event, as originally envisioned by Kumarapeli and Saull (1966) for a larger region that includes the Champlain Valley.

TABLE 1. FAULTED (?) AND FAULT-CUTTING DIKES OF THE LAKE CHAMPLAIN VALLEY

Dike Data	Location	Description and Reference (number)
Monchiquite N65W 15" wide	Crosses northern Juniper Island	Dike shows 15" offset on NE side of island (1)
Lamprophyre 10" wide	End of Clay Point, Colchester	Right-lateral offset of 3' along a N-S fault. Shale is also rotated (1,4)
Lamprophyre 3' wide	Hubbell's Falls, Winooski River, Essex	2 left-lateral offsets reported (1), but no offsets seen by Perkins (4)
Camptonite N83W, 78N 148 cm wide	E. side Rte. 7, 0.5 miles N. of Charlotte intersection	Fault about N30E, 69SE, subparallel to shale cleavage. Apparent offset is 106 cm left-lateral (not true offset) (2)
Trachyte N55E, 81SE 6-11 cm	E. side Rte. 7, 0.8 miles N. of Shelburne. 25 m N. of major fault	Silicified, fills fault plane (offset unknown) in Winooski dolostone. Green color may relate to Mg reaction (?)
Camptonite N24W, 79 SW 0-79 cm wide	Winooski River below Woolen Mill, Winooski	2 right- and 1 left-lateral offsets of 19 to 64 cm, along N5W to N28E joints and syn-intrusional? fault (3)
Camptonite (no data)	North end Monkton Ridge c.2 miles SW Hinesburgh	Cuts minor N-S normal fault associated with major N-S St. George fault (3)
Trachyte E-W (?)	Near bridge over Lewis Creek, North Ferrisburg	Possibly offset by N80W North Ferrisburg fault (5)
Monchiquite E-W, 2' wide	Shoreline at Orchard Point, Shelburne	One of 2 dikes cut by trachyte, offset 1-2', no other data (1)
Trachyte c. N85E, 75NW c. 22' wide	Crosses Reber Road 1.2 miles SW of Rte. 22 in Willsboro	Silicified "keratophyre" (6). Aligns (no offset) on both sides of Adirondack border fault (disputed by Isachsen)
Trachyte sill (no data)	1 mile SW of Essex Village, SE from road	Abruptly terminates at N50E normal fault, but not well exposed (6)
Monchiquite N84W 34" wide	Just north of Beauty Bay, Valcour Island	Walls of dike are offset 1'11", down to north (7)
Monchiquite (2 dikes)	SE corner of Valcour Island	SW dip due to drag rotation along the adjacent Valcour Cove fault (7)
Monchiquite N83W, 75NE 78 cm wide	North side Rte. 2, c.1.2 miles W. of I-89 exit for Grande Isle	70 cm offset, upper side to North along sub-horizontal fault. Dike brecciated at fault. True offset unknown.

References: (1) Thompson, 1861; (2) McHone, 1978; (3) Stanley, 1980; (4)
Perkins, 1908; (5) Welby, 1961; (6) Buddington and Whitcomb, 1941;
(7) Hudson and Cushing, 1931

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ITINERARY

The starting point is the parking lot behind Perkins Geology Hall on the western side of the University of Vermont, off Colchester Avenue (Fig. 3). Cars can usually be left here on weekend days or when the University is not in session. Otherwise, get a visitors permit. Refer to Figure 3 for map locations of stops 1 - 6. Traffic is commonly heavy and fast along Route 7, so please take your time and drive carefully (you can always catch up later). At stop 4, parking and access through private property is by special permission. Food can be purchased at our lunch stop.

The stops are within the Burlington and Mt. Philo 7 1/2' USGS quadrangles, and are along or within a few miles of U.S. Route 7. The Vermont Atlas and Gazetteer (David DeLorme and Co.), available in local stores, has a convenient scale and is recommended as well. The local geology is mapped by Cady (1945), Welby (1961), and in summary by Doll and others (1961). Other areas of the Champlain Valley that contain dikes are mapped by White (1894), Buddington and Whitcomb (1941), Stone and Dennis (1964), and Fisher (1968).

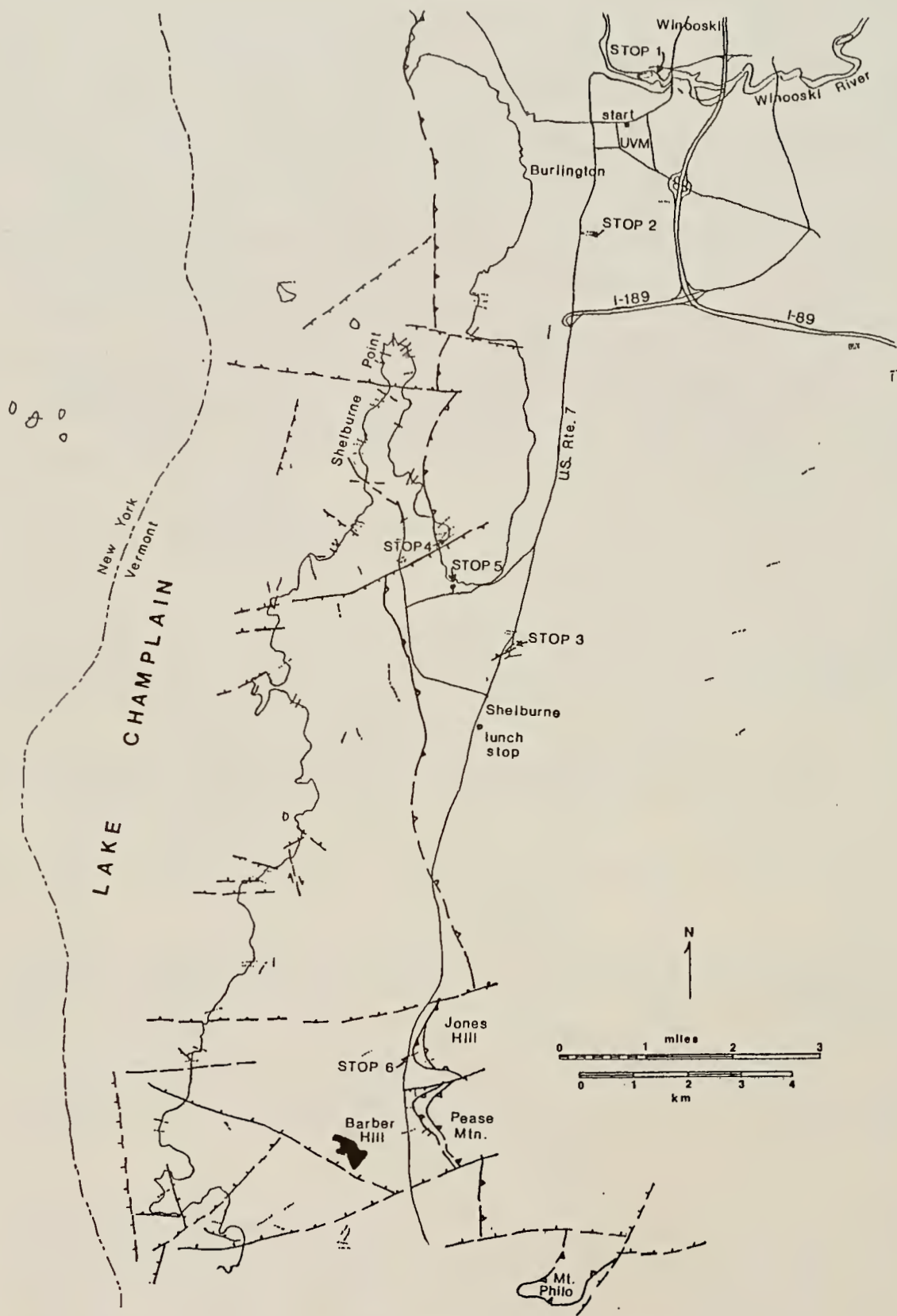


Figure 3. Location map of field stops and dikes in the Burlington - Shelburne area, Vermont (adapted from Welby's 1961 geologic map).

Mileage		Description
Total	by Point	
0.0	0.0	Starting from parking lot behind Perkins Hall (Department of Geology) at UVM, turn right (east) onto Colchester Avenue. Street will curve to the north and descend a hill.
0.8	0.8	Bridge over Winooski River. Colchester Avenue becomes Main Street in Winooski.
0.9	0.1	Turn left (west) at light onto West Canal Street, just past bridge.
1.1	0.2	Park along street near the Woolen Mill. Walk west through the parking lot, around the fence and follow path SE down to the river (about 300 m).

STOP 1. The Cambrian Winooski dolostone at this location shows well-developed fractures that intersect a camptonite dike, near the end of the Winooski River gorge called Salmon Hole. The cliff face follows joints trending between N15W and N25W, about the same trend as the vertical dike present at the cliff base. Three small-displacement, N-S faults are mapped in the area (Stanley, 1980, fig. 7). A N50E joint set crosses both the dike and dolostone. N30E and N5W fractures are more common in the dolostone than in the dike, and many of these joints appear only on the east wall above and up to the dike, but not crossing it.

The dike is a dense and fine-grained augite camptonite, with many small blue (chalcedony?) amygdules. The dike pinches out or is truncated under sand near the river to the south, although a thin dike stringer is present into the water. The main dike extends with some pinch and swell for over 75 m NNW to its cover near some mill turbine ruins, and is 74 to 79 cm wide. A similar dike appears on line with the trend at Schamska Park about 1/2 mile to the SSE and may be connected.

At least three offsets are exposed. The southernmost shows right-lateral offset of 64 cm where a N28E fracture is exposed in the dolostone wall. The dike is 48 cm wide in the offset, and the rock is foliated or sheared parallel to the fracture. The fracture does not appear to extend into the dolostone west of the dike, and is interpreted as a syn-intrusional feature in which slip occurred only on the eastern side shortly after or while the magma conduit opened.

The middle offset is along a fault oriented N2W, 55W that Stanley (1980, p.26) calls a normal fault of minor displacement, possibly related to the longitudinal St. George fault system along the eastern Champlain Valley. Stanley (1980, p.26) states that the fault predates the dike, but about 22 cm of poorly-exposed left-lateral offset can be observed, with the fault continuing in the dolostone south of the dike. In addition, the dike is strongly and closely fractured at this offset.

A third small, right-lateral offset farther north appears to be another syn-intrusional feature, with a small dike stringer extending through the fracture opposite the main dike. The fracture is about N70E, 87SE and does not extend past the dike to the west.

Return on the same streets past the Geology Department and UVM (do not follow Route 7 along the river road).

- | | | |
|-----|-----|--|
| 2.3 | 1.2 | Colchester becomes Pearl Street. |
| 2.5 | 0.2 | Turn left (south) onto South Willard Street (reconnect with Route 7). |
| 3.8 | 1.3 | Roundabout at foot of hill past Dunkin Donuts. Continue south (Route 7). |
| 4.0 | 0.2 | Turn left (east) onto Hoover Street. Up short hill, park in quarry (not on grass). |

STOP 2. Redstone Quarry. This quarry is owned by UVM and has been used for many introductory geology field trips. The red Cambrian Monkton quartzite also contains pale-yellow dolostone beds at this quarry, and many 19th century Burlington houses have foundations or walls made of these rocks. Excellent soft-sediment features are exposed, including ripple marks and hailstone (?) impressions. Please do not climb the walls, and stay out of the adjacent private property and gardens.

Three camptonite dikes are exposed in the northern part of the quarry. The southernmost is 110 cm wide, N85E, 85S, and shows its vertical dimension well along the wall of the quarry. This dike has small nodules of granite, metagabbro and gneiss carried up from Grenvillian basement some distance (several thousand feet?) below, plus several larger dolostone slabs. This is the "Willard's Ledge" dike mentioned by Thompson (1860, p. 580), which he believed to be exposed again "a few rods to the east". Kemp and Marsters (1893) referred to this dike as an example of "augite camptonite", in which Ti-augite phenocrysts predominate rather than the brown hornblende that is an essential part of the camptonite definition. This variety of camptonite has since been shown to be common.

The northern dike in the quarry is also augite camptonite, N86W, 81N, and 66 cm wide. The middle dike is a narrow stringer of glassy augite camptonite about 10 m to the south. It is only about 10 m long, pinching out at both ends with a maximum width of 12 cm. It curves from N80W (thicker part) to N55W (thinner).

Joint patterns in the quarry have not been carefully examined for this report, but an E-W set can be seen, as well as some curving joints like the fracture filled by the stringer dike. We will note a few of the orientations, but watch out for radiating fractures around blasting holes. The continuation of the southern dike eastward into the quarry wall is not clear, and needs examination.

Return to Route 7 down Hoover Street.

- | | | |
|-----|-----|---|
| 4.5 | 0.5 | Turn left (south) onto Route 7. Be careful! If traffic is too heavy, turn right and go around the roundabout. |
| 5.4 | 0.9 | Pass under I-189 exchange, continuing S. |
| 8.9 | 4.4 | Through light at Jelly Mill Common. |

- 9.3 0.4 Turn left (southeast) at intersection curve past Dutch Mill. Watch traffic!
- 9.4 0.1 Turn left, park in First Baptist Church lot. Walk across road, through parking lot adjacent to hill, right along Route 7 highway cut. Stay off the pavement (no need to cross the highway).

STOP 3. S9 roadcut. This cut is dominated by one large and several smaller cross faults, and was described by Stanley and Sarkesian (1972). The exposed surface of the footwall on the major fault is oriented N70E, 75NW, along which Winooski dolostone has dropped against Monkton quartzite. At least eight smaller NE and N-S-trending faults are exposed as well, and slickensides indicate mostly dip movement (near-vertical maximum compression). Gouge is common, and with the slickensides indicate a brittle environment of faulting. Stanley and Sarkesian (1972, p. 132) reported that quartz lamellae at this site indicate NE-SW compression, interpreted as preceding the final fault movement.

About 25 m north of the major fault, a narrow (4 to 12 cm wide) green-colored trachyte dike has intruded a small fault, oriented N54E, 85 SE in the Winooski dolostone. Please do not sample any of this dike...it is unique! The pale grass-green color has not been observed elsewhere, but is believed to be caused by an unknown reaction of the magma with the dolostone. The dike is silicified but preserves altered alkali feldspar phenocrysts. A strange texture of frothy brown bubbles is observed in thin section, along with much dolostone microbreccia incorporated by the dike. An x-ray diffraction pattern of the rock identified alpha quartz, clay minerals, and smaller peaks of unknown cause.

In times of light traffic, two N85E monchiquite dikes can be visited just north of the driveway across the highway at the northwestern end of the roadcut. The southernmost is very weathered to a light-brown color close to that of some trachyte dikes. A large group should not try to cross this busy road.

Return to Route 7.

- 9.6 0.2 Turn left (south) CAREFULLY onto Route 7.
- 10.4 0.8 Through stop light, Shelburne Village.
- 10.7 0.3 Turn left into parking lot at Harrington's, home of the "world's best ham sandwich", for lunch stop. A rest room is available upon polite request. A gas station is nearby. Also, Cafe' Shelburne for the elite eater. Please limit lunchtime to 45 minutes or less.

Return north on Route 7 to Village.

- 11.0 0.3 Turn left (west) at stoplight in Shelburne Village.
- 11.1 0.1 Cross railroad track.
- 11.7 0.6 Road curves right (north).

- 12.6 0.9 Stop sign. Continue straight north toward Shelburne Point.
- 13.4 0.8 Turn right (east) into parking lot south of large brown barn. Parking by permission only, do not block driveway (boat repair shop).

Walk east across field toward southern side of hill on the lake above Shelburne Bay. This is private property of Mr. Thomas Cabot, and access permission is for this trip only. Walk though the woods, down the hill toward the lake. Assuming normal low Fall lake levels, we will walk along the shore northward along the hillside. If high water, stay well above the lake, and travel around the eastern hillside through the woods, where thick float can be seen.

STOP 4. Shelburne Point intrusive breccia. At least four trachyte (bostonite) dikes are found around the eastern side of this point, three of which contain abundant xenoliths of many Paleozoic and Proterozoic rocks that underlie the region. The intrusive breccias have received attention from Hitchcock (1860), Kemp and Marsters (1893), Perkins (1908), Powers (1915), Hawley (1956), and Welby (1961). The most southerly breccia dike is about 4 feet wide and is more than half xenoliths by volume, including many Grenvillian basement rocks as well as shale, quartzite, limestone, and porphyritic syenite cobbles. A great deal of similar material occurs as float along the hillside above and southwest of this dike. At least one other breccia dike farther north is very narrow (a foot or so), and has been eroded into a "chasm squeeze" into which you must fit sidewise for sampling.

Kemp and Marsters (1893) suggested that the abundant xenoliths are derived from breccia along an older fault that has been intruded by bostonite magma. Many of the xenoliths are remarkably rounded, almost like stream cobbles. Similar breccia dikes, perhaps the same ones, also appear on the southwest shore of Shelburne Point. Welby (1961) has mapped a high-angle fault that displaces the Champlain thrust nearby to these breccias, lending support to Kemp's idea. The concentration of trachytes and their syenite xenoliths (autoliths?) indicates the presence of a syenitic pluton directly beneath southern Shelburne Point. Other xenolith-rich dikes across the region are also associated with faults (McHone and Williams, 1985).

Return the same route back to cars, and turn left (south).

- 14.2 0.8 Intersection near Shelburne Farms. Turn left (east).
- 14.9 0.7 Turn left (north) into large parking area of Shelburne Bay Access. Walk out onto rocky peninsula.

STOP 5. Shelburne Bay Access. Joints and small-displacement faults are well exposed at this classic teaching site, described by Stanley (1974). North-south and east-west faults show both dip and strike-slip offsets of less than 30 cm. Some E-W fractures are extensional and are filled with quartz. Careful analysis of the fractures by Stanley (1974) and his students has shown that an early set developed with generally E-W principle compressive stress, followed by a second set with roughly N-S compression. Low confining stress

conditions indicate shallow depths (less than 2 km) during deformation. Although Stanley (1974) originally related the Shelburne Access fracturing to Acadian (Devonian) tectonism, a Mesozoic time of deformation is also reasonable for at least some of these features. Stanley (1974) especially pointed out the proximity to the major Shelburne Point cross fault near to the previous stop (and almost visible from the access rocks).

Return back the same way (turn right from the parking lot)

- | | | |
|------|-----|---|
| 15.6 | 0.7 | Back at intersection near Shelburne Farms, turn left (south). Continue back (SE) to Shelburne. |
| 17.2 | 1.6 | Turn right (south) onto Route 7 at stop light, Shelburne Village. |
| 18.0 | 0.8 | Through blinking yellow light. |
| 18.2 | 0.2 | Gecewicz Farms fruit stand on left. |
| 20.9 | 2.7 | Good view of Champlain Valley. |
| 21.7 | 0.8 | Jones Hill on left, Barber Hill is low hill in valley about one mile directly ahead. |
| 22.2 | 0.5 | Intersection with F-5 (Charlotte). Turn around by turning left into gas station, return north on Route 7 so that we can park on the east side of the highway. |
| 22.7 | 0.5 | Park off pavement along east side of Route 7. |

STOP 6. Jones Hill dike. This dike was exposed by highway construction in the late 1960's. The Champlain thrust fault has capped this hill and Pease Mountain to the south, as well as Mt. Philo and others of the "Red Sandrock Range", with durable Monkton Quartzite. According to Welby (1961), younger cross faults cut the Champlain thrust, contributing to erosion that has isolated these hills from one another. As at the Pease Mountain roadcut farther south on Route 7, the Jberville shale is highly contorted and folded. The offset visible in this camptonite dike partly follows shaley cleavage, but it is clear that the dike crosscuts most of the deformation. Although generally covered by rubble, the southern end of the offset is a sharp break. However, the dike along the offset is fine-grained like the chill margin, so it may an intrusional feature. As listed in Table 2, the petrography of this dike shows it to be a normal camptonite, with modal variations that might be attributed to cooling rates. Chemical analyses of two of the five samples shows that the dike magma was typical basanite, except for its high volatile content.

Trachyte dikes are exposed in the Pease Mountain hillside and roadcut to the south, and they trend toward Barber Hill, the top of a small (?) syenitic stock nearby to the west of Pease Mountain. Ambitious geologists might also attempt to visit other offset dikes listed in Table 1, some of which lack serious study.

End of trip. Return north to I-189 for easy access to I-89 and Montpelier.

TABLE 2. MODAL AND CHEMICAL ANALYSES ACROSS THE JONES HILL DIKE, CHARLOTTE

SAMPLE	A	B	C	D	E		B	D
Ti-augite	18.9	23.0	21.6	25.8	17.1	SiO ₂	43.9	44.57
Alt. cpx	2.3	tr.	tr.	tr.	tr.	TiO ₂	2.31	2.16
Kaersutite	8.7	27.1	25.7	31.2	17.2	Al ₂ O ₃	14.6	13.71
Plagioclase	7.5	25.0	25.1	22.2	15.8	Fe ₂ O ₃	2.24	n.a.
Calcite*	2.6	2.7	3.9	1.1	4.2	FeO	7.07	9.35**
Selvage***	51.4	0.0	0.0	0.0	36.7	MnO	0.17	0.18
Apatite	1.2	2.3	1.9	1.2	0.8	MgO	6.71	8.39
Mag.+Pyr.	7.4	11.4	5.0	7.2	8.2	CaO	9.71	10.44
Analcime	tr.	8.5	14.7	10.3	tr.	Na ₂ O	3.68	2.94
Serpentine*	tr.	tr.	2.1	1.0	tr.	K ₂ O	2.30	2.67
						P ₂ O ₅	0.85	0.55
						H ₂ O+	1.81	0.90
						CO ₂	3.49	3.41
						H ₂ O-	0.63	0.25
						Total	99.47	99.52
						ppm Rb	49	79.6
						Sr	1311	1130
						Y	n.a.	25.2
						Zr	336	230
						V	n.a.	189
						Cr	n.a.	444
						Ni	n.a.	168
						Ba	1140	1150
						Sr ⁸⁷ /Sr ⁸⁶	0.7046	n.a.

*mostly replacing oliving phenocrysts

** total iron as FeO

***mostly devitrified glass, incl. microlites of
plag., kaers., opaques, analcime, & apatite

n.a. = not analyzed

Note: A - E are fist-size samples taken across
the 148-cm dike from north to south. Approximate
locations: A = 0 cm (north contact); B = 17 cm;
C = 53 cm; D = 95 cm; E = 148 cm (south contact).
Modes are by 1000-point counts of single sections.
Source: unpub. 1978 PhD dissertation by McIlhene.

THE CAMBRIAN PLATFORM IN NORTHWESTERN VERMONT

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INTRODUCTION

Cambrian and Lower Ordovician clastic and carbonate sediments in the Vermont portion of the northern Appalachians were deposited on a passively subsiding shelf following late Precambrian rifting of the Iapetus Ocean (Rodgers, 1968). These sediments must have accreted at a rate which kept pace with subsidence as the shelf assumed the morphology of an accretionary rimmed platform (Read, 1985) during the Lower Cambrian. Distribution of facies indicates that the shallow water interior of this platform was affected by tidal fluctuations, and it passed laterally into open shelf regions and ultimately into deeper water basins (Mazzullo and Friedman, 1975). This general sequence of facies has been summarized for the Cambrian and Lower Ordovician of the Appalachians by several authors (Rodgers, 1968; Palmer, 1971). The facies associated with the pericontinental portion of the sequence have been described by Myrow (1983), Rahmanian (1981), Chisick and Friedman (1982), Braun and Friedman (1967) and Speyer (1982). Outer shelf facies have been described by Keith and Friedman (1977). It is important to note that none of these works document the complete sequence from epicontinental seas through to shelf edge to slope environments; most concentrate on the description of a portion of the platform sequence. In fact, looking at Cambro-Ordovician sequences throughout the entire Appalachians, only Pfeil and Read (1980) describe a platform to basin sequence, but it has been dismembered by faults and cannot provide information on the original geometric relations on the platform.

This field trip guide describes the facies and evolution of a portion of the Cambro-Ordovician carbonate platform in northern Vermont (Figure 1). The Dunham Dolomite represents the first carbonate sediment deposited on the newly formed shelf, and we see that the facies distribution and paleogeography of the Dunham platform controls what develops in subsequent Cambrian units. Study of the Dunham Dolomite can therefore tell us much about the morphologic evolution of the entire Cambro-Ordovician platform. The sediments underlying the Dunham Dolomite (Figure 2) have been described by Myrow (1983) as tidally-influenced shallow shelf clastics (Cheshire Quartzite) which overlie the clastic rift basin fill sediments of the late Precambrian Pinnacle and Fairfield Pond Formations (Tauvers, 1982). The Cheshire Quartzite will not be seen on this field trip.

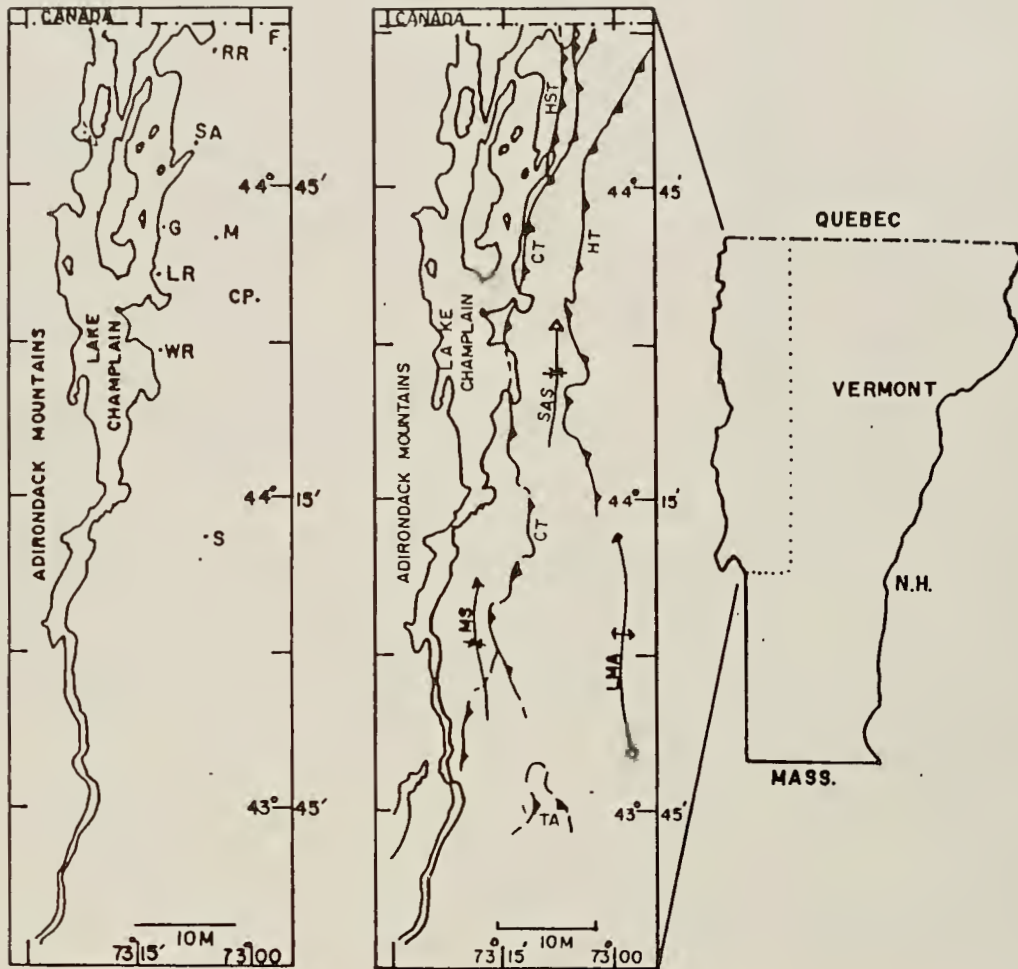


Figure 1. Locality map for some of the important outcrops of Cambrian units, and major structural features in the region.

LMA- Lincoln Mountain Anticlinorium; MS- Middlebury Synclinorium; SAS- St. Albans Synclinorium; TA- Taconic Allochthon; CT- Champlain Thrust; HT- Hinesburg Thrust; HST- Highgate Springs Thrust.
 S- Shelburn; WR- Winooski River; CP- Colchester Pond; LR- Lamoyille River; M- Milton; G- Georgia; SA- St. Albans; RR- Rock River; F- Franklin

Overlying the Dunham Dolomite lies the lower Middle Cambrian Monkton Quartzite (Figure 2), a mixed carbonate and siliciclastic unit which records tidal flat to platform margin sedimentation. The Monkton Quartzite is overlain by the Middle Cambrian Winooski Dolomite, and detailed lithofacies analysis has not been completed on this unit, but the few remaining sedimentary and biogenic structures suggest it also records peritidal to platform margin environments of deposition. The Upper Cambrian Danby Quartzite gradationally and conformably overlies the Winooski Dolomite and Butler (1986) has documented that the Danby records tidal flat to platform margin sedimentation with significant storm overprinting.

This field trip will look at each of these units and examine some of the evidence for the environmental interpretations. The field trip will examine the Dunham, Monkton and Winooski Formations in the Milton area and the Monkton, Winooski and Danby Formations in the Burlington/Winooski regions.

GEOLOGIC SETTING AND STRATIGRAPHY

The Cambrian to Lower Ordovician stratigraphic sequence in western Vermont outcrops in a north-south trending belt (Figure 1), a region bordered on the east by the Green Mountain Anticlinorium, a belt of Precambrian rocks thought to represent the easternmost occurrence of the North American craton in the Lower Paleozoic (Rodgers, 1968). The north-south trending outcrop belt consists of several major fold belts (St. Albans Synclinorium, Middlebury Synclinorium) and thrust sheets (Champlain, Hinesburg, Pinnacle, Highgate Springs). The north-western portion of the outcrop belt is ideally suited for sedimentologic studies of the Cambrian to Lower Ordovician stratigraphic sequence because it lies within the Quebec Reentrant (Williams, 1978, Thomas, 1978), which kept deformation and metamorphism associated with the Taconic and Acadian Orogenies to a minimum. The most complete exposures of the Lower Paleozoic are contained within thrust sheets in this region. Stratigraphy within the thrust sheets is coherent, which enables us to reconstruct original geometric relationships.

Cambrian and Lower Ordovician clastics and carbonates in the northern Appalachians were deposited on a tectonically stable shelf which developed following late Precambrian rifting of the Iapetus Ocean. This shelf was undergoing thermal subsidence throughout the interval when Cambro-Ordovician sediments were being deposited.

STRATIGRAPHIC TERMINOLOGY

The Cambro-Ordovician stratigraphic sequence in northwestern Vermont was divided into two sequences by Dorsey (1983): a western shelf and an eastern basinal sequence (Figure 2). The western shelf sequence is composed of alternating siliciclastic (Cheshire, Monkton, Danby Formations) and carbonate (Dunham Winooski and Clarendon Springs Formations) units. The stratigraphy of the basinal sequence which corresponds to the shelf sequence but deposited in deeper water, consists of shale units (Parker, Skeels Corners, Morses Line) with isolated breccia units (Rugg Brook, Rockledge Formations). Unlike the western shelf sequence, correlations are relatively well developed within the basinal deposits. The stratigraphic nomenclature for these units was developed by Shaw (1958), revised by Palmer (1970) and Palmer and James (1980), and most recently revised by Mehrtens and Dorsey (1987) and Mehrtens and Borre (in press) and is presented in Figure 2.

DEPOSITIONAL ENVIRONMENTS OF THE WESTERN SHELF SEQUENCE

Pre-Cheshire Units

The Pinnacle and Fairfield Pond Formations underly the Cheshire Quartzite in central Vermont. The stratigraphy and structure of these units was studied by Tauvers (1982) and their depo-tectonic setting described by Dorsey and others (1983). The Pinnacle and Fairfield Pond Formations are interpreted as representing rift basin fill sediments deposited following initial rifting in the Eocambrian. Doolan and others (1982) has suggested that this rifting may have occurred at approximately 560my before present. The topography of the rift basin resulted in a basal unit of coarse-grained clastics, possibly alluvial fan in origin (Tauvers, 1982) overlain by finer-grained siliciclastic sediments of the Fairfield Pond Formation, interpreted as forming in marginal marine basins. The contact of the Fairfield Pond Formation with the overlying Cheshire Quartzite was shown by Tauvers to be conformable. These units will not be seen on this field trip.

Cheshire Quartzite

The Cheshire Quartzite will also not be seen on this field trip as the best exposures of this unit occur in west-central Vermont. A detailed field and petrographic study of the lithofacies of the Cheshire Quartzite was completed by Myrow (1983). The Cheshire is an important unit because it represents the transition from the siliciclastic rift basin fill sediments to those of the newly developing platform (rift-drift transition).

Myrow recognized eight distinct lithofacies within the Cheshire. The lower unit of the Cheshire is arkosic to subarkosic in composition; it is similar to the Gilman Quartzite

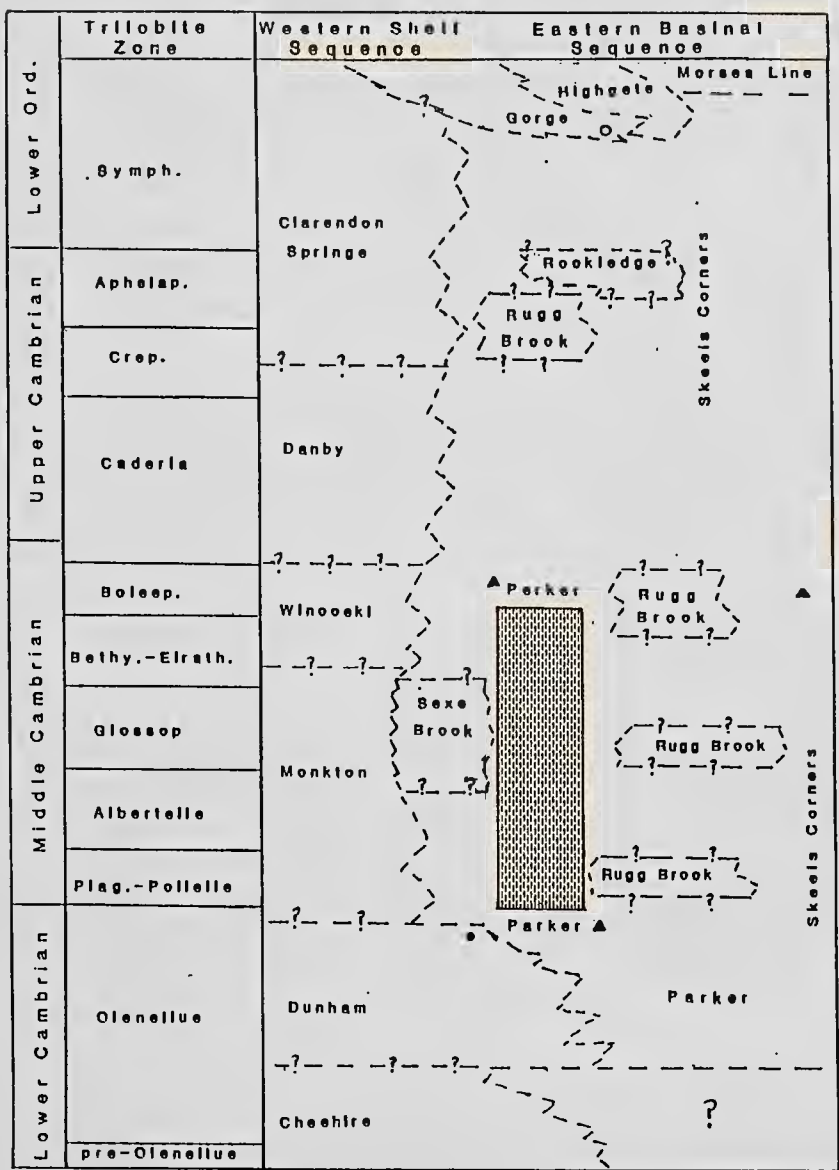


Figure 2. Correlation chart for the Cambro-Ordovician strata of northwestern Vermont. Chart is based on biostratigraphic data by Palmer (1970), Palmer and James (1980), Landing (1983) and Mehrtens and Gregory (1983), and physical stratigraphic relationships by Mehrtens and Dorsey (1987) and Mehrtens and Borre (in press). Age relationships on the platform are approximate because they are based on intertonguing relationships with the basinal units. The Hawke Bay event is illustrated by the shaded bar within the Parker Slate. Pods of Rugg Brook Dolomite extend from post-Dunham to Rockledge time. The Skeels Corners Slate has been dated as Bolaspidella zone in age but mapping relationships suggest it has a much broader age range.

exposed in nearby southern Quebec. The upper unit of the Cheshire is a quartz arenite and is similar to the Cheshire at its type section in Massachusetts.

The lower Cheshire is composed of five lithofacies: 1) fine-grained mottled grey, argillaceous arkose. Distinctive characteristics include extensive bioturbation; thin, white, rippled beds and disseminated shale partings. 2) fine-grained, white subarkosic and fine-grained grey arkosic beds. Clay drapes commonly overlie the arkosic horizons. Distinctive features include: ripple bedding, wavy and lenticular bedding, thick and thinly interlayered bedding, horizontally stratified and cross laminated beds, and U-shaped vertical burrows. 3) fine-grained, white subarkosic beds with thin clay drapes. Distinctive characteristics include: medium to thick massive and horizontally stratified beds, occasionally exhibiting tabular cross stratification, massive lenticular beds, lenticular low angle trough cross stratification, rippled beds, reactivation surfaces and erosional surfaces. 4) thin, lenticular, structureless sand bodies with erosional bases and flat upper surfaces. Low angle cross stratification can be present. 5) tabular sand beds characterized by planar, non-erosive bases and reworked tops, and a notable lack of internal structures.

The upper Cheshire is composed of three lithofacies: 1) a pink to white weathering, moderately to poorly sorted, massive, structureless, fine-grained quartzite whose composition ranges from quartz arenite to arkose. Minor amounts of carbonate cement can be present. 2) a shale clast conglomerate composed of interbedded quartzite with shale clasts or chips. 3) massive quartzite beds, lenticular in shape with large scale erosional surfaces at their bases. Beds exhibit large scale trough cross stratification.

Interpretations:

These eight lithofacies can be interpreted to represent sediments deposited on a newly formed shelf, at least in part within wave base, and partially tidally influenced. The Cheshire Quartzite is thought to represent the marine shelf sand blanketing the underlying rift basin topography. Shelf sediments of the lower Cheshire exhibit periodic storm sedimentation, and are capped by the prograding strandline sands of the upper Cheshire. This interpretation is based on: 1) position within the Cambrian stratigraphic sequence; 2) absence of any lithofacies characteristic of the supratidal environment; 3) comparison to stratigraphic sequences of similar rock units elsewhere in the Appalachians/Caledonides (Swett and Smit, 1972).

Dunham Dolomite

The Dunham Dolomite will be the first unit seen on this field trip.

The lithofacies and depositional environments of the Dunham

Dolomite were studied by Gregory (1982) and Mehrtens and Gregory (in review). These authors recognized that the Dunham Dolomite is a 400 meter thick unit composed of four major lithofacies representing peritidal, subtidal/open shelf, channel and platform margin environments. The base of the Dunham Dolomite is in gradational and conformable contact with the underlying Cheshire Quartzite, and the Dunham represents the first carbonate deposit on the newly formed shelf.

The peritidal facies of the Dunham is characterized by a bedding style termed "sedimentary boudinage", which describes the rhythmic interbedding of lithologies and subsequent differential compaction to produce beds which contain pods, or boudins. In the Dunham the interbedding consists of beds of pure dolomite (white) and silt-rich dolomite (red). This rhythmic interbedding is probably the result of deposition in a tidally-influenced regime. Bioturbation has disrupted burrowing, and early cementation has compacted some horizons sufficiently enough to form rip-up clasts and local intraformational conglomerates. Cryptalgalamintes also occur in this facies.

The subtidal/open shelf facies is characterized by shallowing-up cycles 6 to 10 meters in thickness which have at their bases massive beds of bioturbated dolomites which pass up into the rhythmically interbedded dolomite and silt-rich dolomite of the peritidal facies. The bulk of the Dunham Dolomite is composed of these shallowing-up cycles, indicating that the tidal flats prograded into the adjacent platform almost to the shelf margin.

The third lithofacies, the channel deposits, are interbedded with both the peritidal and subtidal/open shelf facies. The channels exhibit lenticular beds with downcutting bases, abundant quartz sand and both intraformational and "exotic" clasts. Trough cross stratification is also common.

Rocks characteristic of the platform margin lithofacies exhibit horizons of polymictic breccias within a quartz sand-rich dolomite matrix, interpreted as talus deposits and debris flows accumulating off the Dunham carbonate bank. In the Burlington and Winooski regions the Dunham is gradationally overlain by the Monkton Quartzite, but in the Georgia area the breccias grade conformably into clast-rich horizons of the Parker Slate, preserving the platform-to-basin transition (Mehrtens and Borre, in press).

Analysis of the distribution of the platform margin lithofacies is important in developing a model for the geometry of the Lower Cambrian carbonate platform, since these deposits very accurately place the position of the platform-to-basin transition. The distribution of the platform margin deposits indicate that the Dunham passed eastward, down-dip into the Parker Slate, (for example, in the vicinity of Arrowhead Mountain). This down-dip facies change is related to the passage into the shale basin and deeper water sediments of the Iapetus Ocean. Platform margin

deposits also indicate that the shallow water deposits of the Dunham pass northward into a basin termed the St. Albans Reentrant (Mehrtens and Dorsey, 1987), which represents a foundered graben within the shelf. Based on outcrop patterns of the platform margin facies, shale units, and breccia deposits within the shales, Mehrstens and Dorsey (1987) and Mehrstens and Borre (in press) defined the margins of the St. Albans Reentrant and suggested that it was a major intrashelf basin accumulating basinal shales which influenced the distribution of the shallow water platform deposits of the Dunham and post-Dunham facies.

Monkton Quartzite

The lithofacies and environments of deposition of the Monkton Quartzite were studied and summarized by Rahmanian (1981), who recognized seven lithofacies, three of which consist of mixed siliciclastic and carbonate sediments, three of which are pure siliciclastic deposits and one is an oolitic dolomite facies. The 300 meter thick Monkton Quartzite is composed of cyclic shallowing-up cycles characterized by repetitive packages of: 1) basal subtidal siliciclastic sand shoals and channels overlain by, 2) interbedded siliciclastic sand, silt, and carbonate intertidal flat sediments, capped by, 3) carbonate muds of the high intertidal and supratidal flat. These cycles are interpreted to represent prograding tidal flat deposits. Two siliciclastic lithofacies have been recognized: 1) sand bars and tidal channels and, 2) mixed rippled sands with mud drapes of the intertidal. These supra-, inter-, and shallow subtidal sediments pass downdip to the east and north (into the St. Albans Reentrant) into subtidal oolitic dolomites and platform margin breccias.

The high degree of similarity between the environments of deposition and facies distribution between the Dunham Dolomite and Monkton Quartzite (Mehrtens, 1985) suggests that the morphology of the Cambrian platform was established in Dunham time and maintained through Monkton deposition. Although the composition of the platform sediments changed from dominantly carbonate (Dunham) to mixed siliciclastic/carbonate (Monkton), the environments of deposition in which these sediments were deposited, and the distribution of these environments, remained the same. Whatever generated the source for the Monkton sands did not effect the geometry of the platform on which they were deposited.

On the shallow water platform the Monkton Quartzite is gradationally overlain by the Winooski Dolomite. This can be seen, for example, in the Winooski region. In the Milton and Georgia areas (eastward and northward) the Monkton is overlain by undifferentiated Parker and Skeels Corners Slates (Mehrtens and Borre, in press).

Winooski Dolomite

The environments of deposition and lithofacies of the

Winooski Dolomite have not yet been studied in detail, but initial studies suggest that it is approximately 300 meters thick and is composed of the following lithofacies: 1) interbedded rippled fine-grained sand and silt with minor clay, 2) dolomite with planar cryptalgalaminite structures; 3) dolomite with LLH stromatolites reaching a height of 50cm; 4) dolomite with disseminated quartz sand; 5) quartz arenite beds with a dolomite matrix, and 6) polymictic breccia beds with a matrix of dolomite and quartz sand-rich dolomite. No shallowing-up cycles have yet been recognized within the Winooski.

Lithofacies 1 through 5 are arranged in a vertical stratigraphic sequence in an active quarry in Winooski, and are characteristic of the base of the unit. Lithofacies (1) and (2) are interbedded with the underlying Monkton Quartzite and are interpreted to represent peritidal deposits. Lithofacies (3), (4), and (5) overlies facies (1) and (2) and they make up the bulk of the stratigraphic sequence seen along the Winooski River (Stop 6). Due to an absence of any obvious sedimentary structures, and stratigraphic position overlying the peritidal facies, lithofacies (3), (4), and (5) are interpreted as shallow subtidal in origin. Lithofacies (4), (5) and (6) are recognized as composing the uppermost horizons of the Winooski Dolomite, and are interpreted to represent subtidal and platform margin deposits, respectively. As seen in the Dunham and Monkton Formations, the Winooski Dolomite also exhibits significant facies changes parallel to depositional strike. The platform margin breccias of the Winooski pass northward into the shales and breccia horizons of the Rugg Brook Formation (Mehrtens and Borre, in press) in the St. Albans Reentrant. To the south, well up onto the adjacent shallow water platform, the Winooski Dolomite passes gradationally upward into the Danby Quartzite.

Danby Quartzite

The Danby Quartzite is a 35-80 meter thick mixed siliciclastic-carbonate unit. Near its type section in southern Vermont the Danby is characterized by a siliciclastic basal unit and an upper carbonate unit termed the Wallingford Member. To the north, in this study area, the Danby thins and becomes a mixed siliciclastic-carbonate unit composed of interbedded quartz arenite, pure dolomite, dolomitic sandstone and sandy dolomite. Four lithofacies have been identified by Butler (1986): 1) intertidal to shallow subtidal, 2) subtidal, 3) open shelf and 4) platform margin. The inter- to shallow subtidal facies is characterized by interbedded sandy dolostone, quartzose sandstone and shales with mudcracks, vertical burrows, wave and current ripples, cryptalgalaminites and oncolites. The subtidal sediments are composed of thick bedded sandy dolostones and pure dolostones with herringbone cross stratification. The open shelf facies is characterized by thick bedded, coarse-grained dolomitic sandstones and quartzose sandstones with large scale tabular cross stratification. Platform margin deposits include polymictic breccias in a dolomite matrix, and cross bedded sands interpreted as platform margin sand bodies.

The Danby Formation is interpreted to represent sediments deposited on the Cambrian platform in a complex mosaic of deposits recording both fairweather and storm processes. Evidence for storm deposition is best exhibited in the inter-to shallow subtidal facies, where laterally discontinuous bedding, erosional downcutting surfaces, hummocky cross stratification, and graded bedding are common.

PLATFORM GEOMETRY

Figure 3, taken from Dorsey and others (1983) summarizes the geometry of the Cambrian platform in northwestern Vermont. Several important features are shown on this diagram, constructed from a view looking southeast. The St. Albans Reentrant, the shale basin lying along depositional strike in the shelf, is shown. Note also the north-to-south, and west-to-east facies changes present within every Cambrian platform deposit. The shallow water facies are present in the south, and they pass northward and eastward into subtidal and platform margin deposits, and ultimately into the shale basin. The diagram also shows the localization of the platform margin from Dunham through Danby time. This is important because it indicates that the platform was characterized by vertical aggradation, or upward building, throughout the Cambrian. Platform facies did not build out into the basin, nor did the platform founder and experience significant onlap of basinal shales. What could have caused the pronounced localization of the platform margin? If the St. Albans Reentrant is indeed a graben within the shelf which foundered as a result of movement on underlying Eocambrian rift block terrain (Mehrtens and Dorsey, 1987), in other words, a lystric fault, then these localized deposits accumulated along the fault scarp. Sedimentation on the platform itself was able to keep pace with thermal subsidence on the young, hot, recently-rifted margin, and the sediment built vertically, with an abrupt pinchout into the adjacent basin. The timing of initial movement of the graben which formed the St. Albans Reentrant is thought to be late-to-post-Dunham time, based on the fact that the Dunham Dolomite is the only shelf unit which continues across what becomes the shale basin. Following deposition of the Dunham, the facies on the northern rim of the Reentrant are different than those to the south (Mehrtens and Dorsey, 1987).

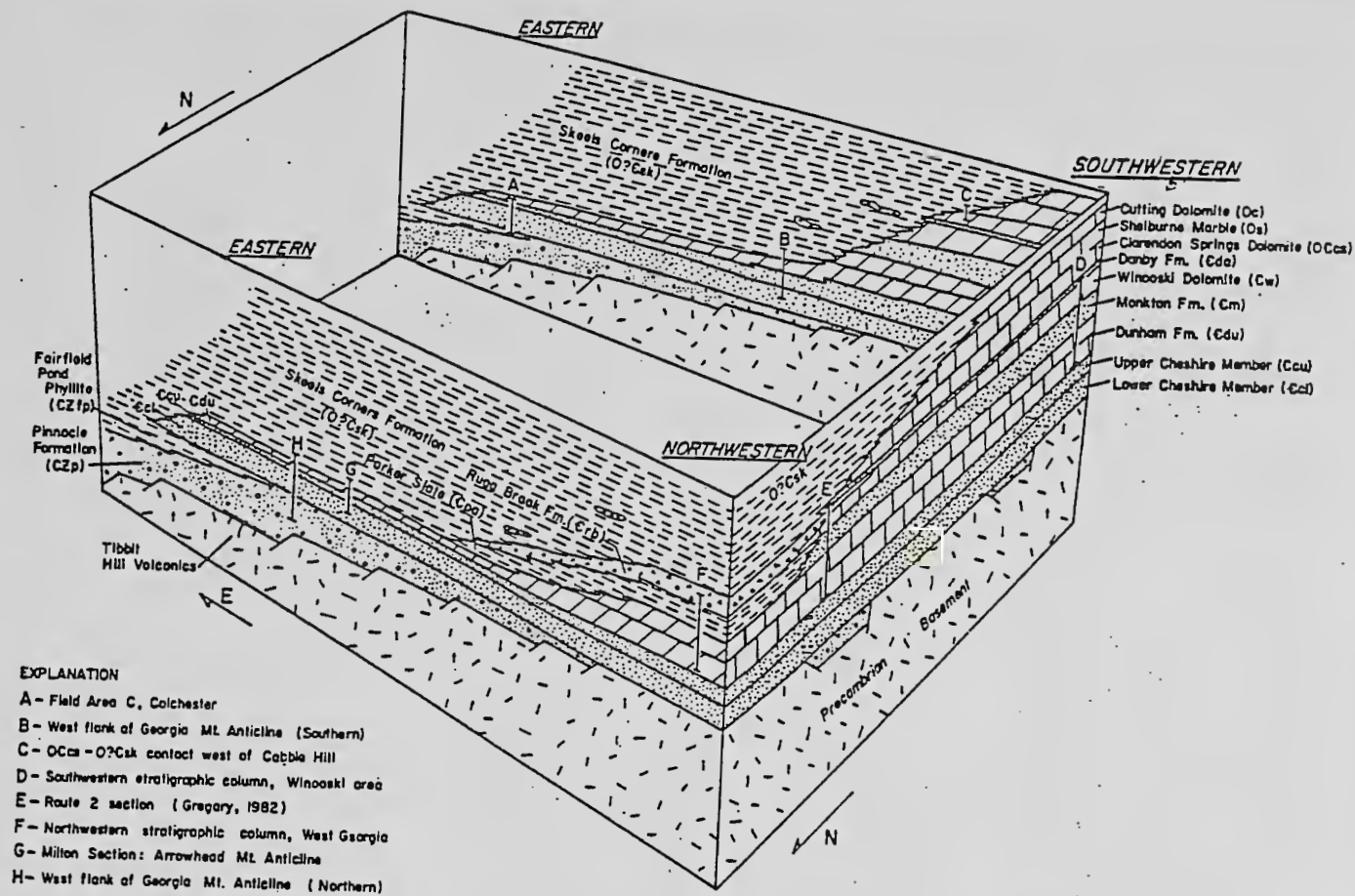


Figure 3. Block diagram from Dorsey and others (1983) illustrating the proposed paleogeography of the southern margin of the St. Albans Reentrant. Diagram shows the platform deposits pinching out to the east, into the marginal Iapetus Ocean, and to the north, into the St. Albans Reentrant. This field trip will be viewing the sections E (Route 2) and D (Winooski River).

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FIELD TRIP ITINERARY

This field trip will stop at exposures of all of the Cambrian units with the exception of the Cheshire Quartzite. All outcrops lie on the upper plate of the Champlain Thrust, and the geographic distribution of facies represents original relationships. Bedding dips gently to the east, north-east. For convenience the trip starts in the north, where the subtidal and platform margin facies occur and ends in the Burlington/Winooski region where tidal flat and shallow subtidal deposits are exposed. Hopefully at the end of the trip you will be able to see the south-to-north facies changes which occur, the localized platform margin facies, and the general features of the Dunham through Danby Formations.

Assembly point: Sand Bar State Park on Route 2 (mileage 0.0)

Mileage

1.6 Stop 1- Abandoned quarry on north side of Route 2

Park on the southwest (right) shoulder, cross Route 2, and ascend overgrown driveway into quarry. The Dunham Dolomite exposed here is the basal facies of the Dunham, characterized by the rhythmic interbedding of dolomite (white) and silt-rich dolomite (pink). This bedding style is thought to be the product of alternating deposition of carbonate and clastic laminae in a tidal flat setting, producing "ribbon bedding". Subsequent differential compaction produces "sedimentary boudinage". There is local imbrication of clasts. Irregular-shaped pods of dolomite are interpreted to be burrows. The white dolomite pods are often cored by calcite. Greiner (1982) has documented occurrences of gypsum. Cryptalgalaminites

resembling LLH stromatolites are present within the red silty-dolomite horizons.

Return to cars and continue east on Route 2.

2.4 Stop 2 Shallowing-up cycles

This roadcut on the left (north shoulder) exposes 3 shallowing-up cycles within the Dunham Dolomite. One SUC can be studied in detail on the northwest corner of the outcrop. The SUC is composed of a basal subtidal dolomite overlain by sedimentary-boudinaged beds of the peritidal facies seen at Stop 1. These cycles are similar to "muddy shallowing-up cycles" of James (1983), and they consist of 6-10 meters of bioturbated, sandy dolomite passing up into ribbon-bedded dolomite and silt-rich dolomite, local intraformational conglomerates and cryptalgalaminites. Note the color change (white to pink) which results from increasing amounts of silt accompanying the change from sub- to peritidal sediments. The contact of the peritidal with the next overlying cycle is sharp and scoured, but there is no sedimentologic evidence of subaerial exposure.

Return to cars and continue driving east on Route 2.

3.1 Stop 3- Dunham subtidal and platform margin facies

Pull off about 100 yards beyond the speed limit sign on this long roadcut. At the base of this outcrop (west end) there are good exposures of the subtidal facies of the Dunham with the characteristic mottled texture produced by burrowing. Burrow mottles are irregular in shape and devoid of quartz and feldspar sand. Burrows are generally about 1 cm across but may be as large as 8cm. Between the white burrows the matrix is very clay-rich and variable in color. Stone and Dennis (1964) attribute this color variation to differing concentrations of trace metals. Specimens of Salterella conulata (Mehrtens and Gregory, 1983) were found in this facies.

The platform margin facies is exposed on the east end of the same outcrop. This facies is composed of chaotically-bedded, laterally discontinuous horizons of breccia in a sand-rich dolomite matrix. Clast composition is highly variable, and includes chert pebbles, sandstones, sandy-dolostones and dolomitic sandstones. Breccia beds are structureless and poorly sorted by graded beds are present in dolomitic sandstone horizons.

Return to cars and continue east on Route 2.

3.8 Stop 4-Monkton Quartzite, subtidal and shelf edge facies

This roadcut exhibits the subtidal and shelf edge facies

of the Monkton, as evidenced by the overall thickness of individual beds, increasing amounts of shale between sandstone beds, presence of relict oolites in some dolomite horizons, and occurrences of large scale tabular cross stratification. Many of these beds probably represent shelf edge sand shoals. Features at this outcrop can be compared to those seen at Stop 6.

Walk east 0.3mi to the small knoll beyond the road sign. Here the polymictic breccia of the platform margin facies is exposed. The clasts are floating in a matrix of sandy dolomite and are interbedded with cross bedded sandstones.

Return to cars and continue east on Route 2.

- 4.7 T-insection with Routes 2 and 7 at Chimney Corners. Turn left.
- 4.8 Pull off into commutor parking lot.

Stop 5- Winooski platform margin facies

Walk from the parking area back to the intersection, looking first at the outcrop on the southwest side. This is an exposure of recrystallized dolomite, cross bedded, and in places oolitic, of the uppermost Winooski. Cross the road to the low-lying outcrop on the east side of Route 7. Note the variable clast composition and abundance of sand in the dolomite matrix in this platform margin breccia. These outcrops of Winooski at Chimney Corners and the I-89 exit ramps to the west are the northernmost outcrops of Winooski in northwestern Vermont. Along routes I-89 and 7 to the north are exposures of the Rugg Brook Dolomite (a basinal breccia deposit).

Return to cars and head for Burlington.

- 5.1 Southbound entrance ramp on I-89
Outcrops of the Monkton Quartzite occur as roadcuts all along Route 89
- 11.1 First outcrop of Winooski Dolomite on the median of I-89
- 11.3 Exit of i-89, southbound onto Routes 2 & 7.
- 12.5 Intersection in Winooski with Routes 2, 7, & 15. Continue straight ahead on Routes 2 & 7.
- 12.7 Bridge over the Winooski River (Stop 6 is below). Bear right at "Y".
- 12.9 Park in the small pull-off on the right or across the street in store parking lots.

Stop 6 Salmon Hole- Tidal flat facies of the Monkton,

Winooski and Danby Formations

Descend the pathway down to the broad bedding planes of the Monkton in the south bank of the river. This beautiful outcrop of Monkton is in danger of going underwater from a soon-to-be-constructed dam, in which case these excellent examples of tidally-bedded sands, shales and dolomite of the supra- and intertidal Monkton would go under! Examine the multitude of rippled surfaces, note that flow directions are highly variable. Look also at the numerous bedding traces, both vertical and horizontal. Mudcracks can also be found. Buff colored dolomite beds here are interpreted to be supratidal in origin, so the uppermost bedding planes are an example of a shallowing-up cycle in the Monkton.

The Monkton/Winooski contact is under water here but can be seen at a quarry near the I-89 exit. It is gradational over about 10 meters, with progressively decreasing amounts of sand up section into the Winooski. To view the Winooski climb up out of the river, walk north across the bridge to the north bank and descend. The Winooski Dolomite does not exhibit many features but cryptogalaminites are present as thin wisps of carbonaceous material with sand grains concentrated along the laminae. Quartz sand is disbursed throughout the unit (eolian?) and becomes progressively more abundant up section as the contact with the Danby is approached. During low water levels the entire section up into the Danby can be walked, but otherwise climb up out of the river again, this time cross Routes 2 & 7, walk past the Champlain Mill shopping arcade and descend to the river on the upstream (east) side of the building. You are now on bedding planes of the shallow subtidal facies of the Danby Quartzite. There are many excellent sedimentary structures exposed at this outcrop, including hummocky cross stratification, complexly-interwoven ripple bundles, bedding planes with interference ripples and graded beds. Biogenic structures include small LLH stromatolites and oncolites. Most of these features suggest that the sediments of the Danby were frequently reworked by storm action, resuspending and reworking the substrate, and rapidly depositing sediment during post-storm surge ebb.

Return to cars. End of trip.

Stratigraphic Definition of the Piermont Allochthon, Sunday Mountain to
Albee Hill, New Hampshire

Robert H. Moench and Katrin Hafner-Douglass
(Trip cancelled 4-15-87, by necessity, with regrets)

The Piermont allochthon, whose existence was not expected prior to our 1985 field season, is a fault-bounded tract composed mainly of Silurian metasedimentary and metavolcanic rocks and probably associated mafic dike swarms that extend at least 100 km from Sunday Mountain, near Orfordville, to a few kilometers north of Groveton, New Hampshire, and possibly an additional 200 km northeast to northern Maine. In an area that was previously mapped almost entirely as the Albee Formation (Upper Cambrian? and Lower Ordovician) south to Groveton, the allochthon contains recognized equivalents of all the Silurian formations of the Rangeley-Phillips area of western Maine, in ascending order: Greenvale Cove, Rangeley (three members recognized), Perry Mountain (plus a volcanic-bearing member), Smalls Falls, and Madrid. Small remnants of the Quimby (Upper Ordovician?) and Littleton (Lower Devonian) are exposed as well. These formations are juxtaposed against an autochthonous sequence in which the Silurian is represented only by discontinuous lenses of Clough Quartzite, and locally by the Fitch Formation. Rocks of the allochthon are interpreted to have originated 15-25 km to the southeast near the Silurian tectonic hinge of western Maine and east-central New Hampshire, and to have been transported to their present position prior to about 400 Ma. This trip will examine the autochthonous and allochthonous sequences, the pertinent localities of isotopically dated rocks, and at least one exposure of the Foster Hill sole fault, which is the inferred base of the Piermont allochthon.

A TRANSECT THROUGH THE PRE-SILURIAN ROCKS OF CENTRAL VERMONT

by

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INTRODUCTION

The central Vermont region is ideally suited to reassess the stratigraphy and structure of the pre-Silurian hinterland because many of the units mapped by previous workers and shown on the Geological Map of Vermont are well exposed in a relatively small region of approximately 200 sq. mi. Furthermore, the transect extends from the carbonate platform through the Middle Proterozoic rocks of the Lincoln massif to the basal Silurian-Devonian unconformity of eastern Vermont (fig. 1). The earlier work by Osberg (1952) and Cady, Albee, and Murphy (1962) is very important because it highlighted many important problems and delineated many of the important formations of eastern Vermont. Subsequent work by Albee (1965, 1968), Laird and Albee (1981), and Laird and others (1984) has placed important constraints on the metamorphic petrology and history of the eastern cover sequence. Isotopic analysis by Laird and Albee (1981) and Laird and others (1984) and discussions by Sutter and others (1985) indicate that much of the metamorphism is associated with the Taconian orogeny although the degree of subsequent cooling and/or Acadian metamorphism is still very much in debate. Important geochemical work by Coish (1985, 1986) and his students at Middlebury College indicates that the mafic schists become more oceanic as they are traced eastward through much of the eastern pre-Silurian cover sequence.

The ultimate goal of our current work is a better understanding of the tectonic evolution of the Taconian orogeny for western New England. Four important questions must be addressed in achieving this goal. They are: 1) What is the dominant mechanism of shortening in the hinterland, folding or faulting?, 2) What is the age of the various structures?, 3) Does the hinterland shorten continuously during orogeny or is it only deformed early in the mountain building process and subsequently acts as a rigid plunger transmitting stress to the imbricating foreland?, and 4) How do the rocks in the pre-Silurian hinterland correlate with those in the Taconic allochthons. Answers to these questions are not only important for the Appalachians in western New England but are important in understanding the processes of compressional tectonics in other mountain belts of the world where small-scale mapping is only available at best.

Recent mapping by Stanley and his graduate students (Tauvers, 1982; DiPietro, 1983; Strehle and Stanley, 1986; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986; DelloRusso and Stanley, 1986;

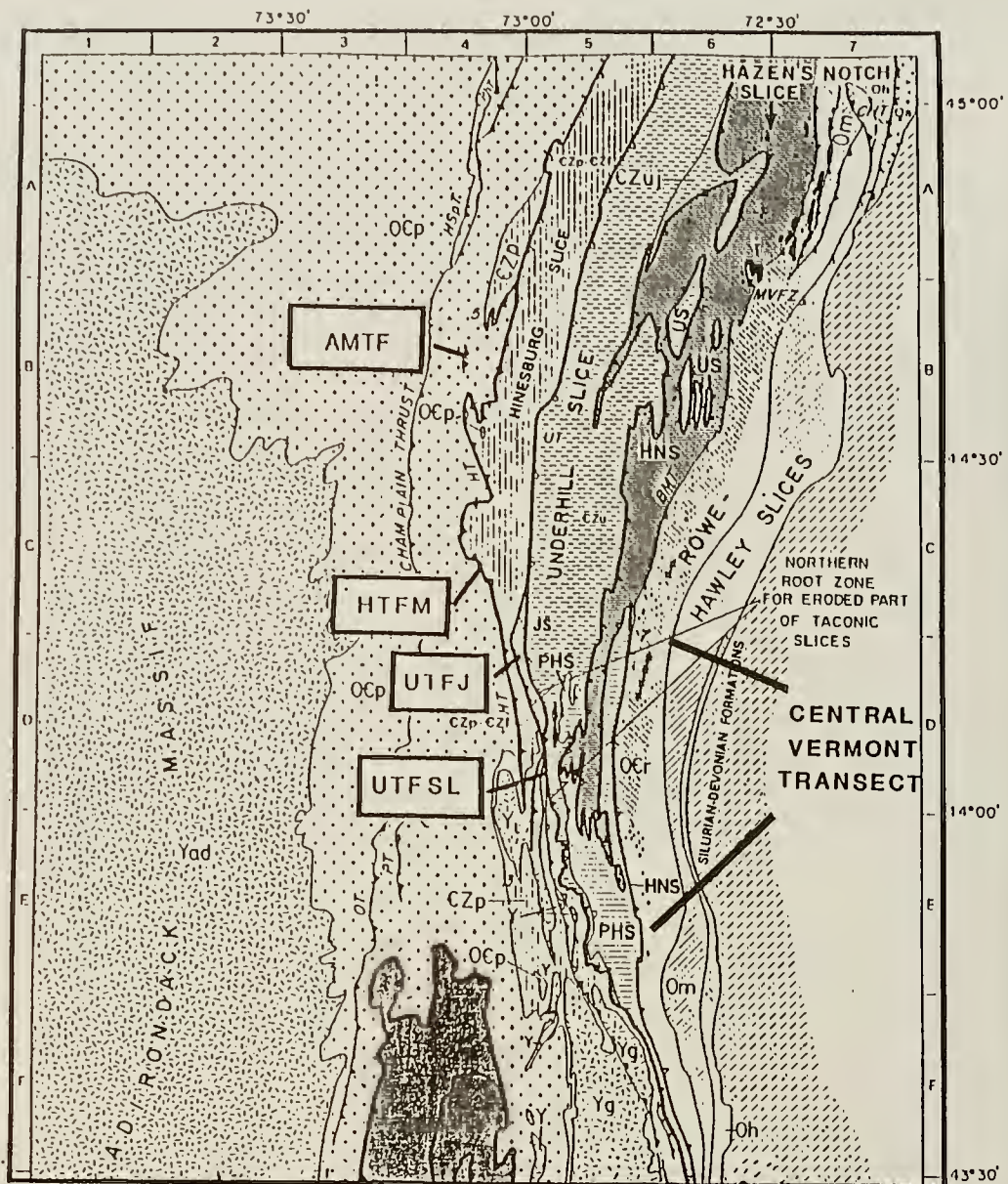


Figure 1 Interpretative Tectonic Map of Vermont and eastern New York showing the general location of the Arrowhead Mountain thrust fault (AMTF), the Hinesburg thrust fault at Mechanicsville (HTFM), and the Underhill thrust fault at Jerusalem (UTFJ), and South Lincoln (UTFSL). The geological map is taken from Stanley and Ratcliffe (1985, Pl. 1, figure 2a). Symbol T in A6 is a glaucophane locality at Tilliston Peak. Short line with x's (Worcester Mountains) and line with rhombs (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones of Albee (1968). Widely-spaced diagonal lines in northcentral Vermont outline the region that contains medium-high pressure amphibolites described by Laird and Albee (1981b). Irregular black marks are ultramafic bodies. Open teeth of thrust fault symbols mark speculative thrust zones. The following symbols are generally listed from west to east. Yad, Middle Proterozoic of the Adirondack massif; Yg, Middle Proterozoic of the Green Mountain massif; YL, Middle Proterozoic of the Lincoln massif; Y, Middle Proterozoic between the Green Mountain massif and the Taconic slices, Vermont; OCp, Cambrian and Ordovician rocks of the carbonate-siliciclastic platform; rift-clastic sequence of the Pinnacle (CZp) and Fairfield Pond Formations (CZf) and their equivalents on the east side of the Lincoln and Green Mountain massifs, PhT, Philipsburg thrust; HSpT, Highgate Springs thrust; PT, Pinnacle thrust; OT, Orwell thrust; UT, Underhill thrust; HT, Hinesburg thrust; U, ultramafic rocks; CZu, Underhill Formation; CZuj, Jay Peak Member of Underhill Formation; OCr, Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation and its equivalents in Vermont; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens Notch slice; MVFZ, Missisquoi Valley fault zone; PHS, Pinney Hollow slice; BMT, Belvidere Mountain thrust; CHT, Coburn Hill thrust; Oa, Ascot-Needon sequence in grid location 7A.

Prewitt, Haydock, Armstrong, Kraus, Walsh, Cua, and Kimball, all in progress) has shown the following important relations: 1) The Middle Proterozoic rocks of the Lincoln massif are progressively deformed to the east by east-over-west folds and synmetamorphic faults of post-Grenvillian age (Prahl, 1985; DelloRusso, 1986). 2. The eastern boundary of the massif is deformed by several synmetamorphic thrust faults (South Lincoln, Jerusalem, and Underhill thrust faults) which contain slivers of Middle Proterozoic rocks (Tauvers, 1982; DelloRusso, 1986; Strehle and Stanley, 1986). 3. The Hoosac (Pinnacle), Underhill, Mt. Abraham, Hazens-Notch, Pinney Hollow, Ottauquechee, and Stowe Formations in the eastern cover sequence are deformed by major pre- or early-metamorphic thrust faults that have been tightly folded and refolded into sheath-like folds and cut by synmetamorphic and late- to post-metamorphic thrust faults (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986; Armstrong, Kraus and Walsh, in progress). 4. Fabrics along synmetamorphic faults indicate a complex history of east-over-west movement followed in many places by later flattening which commonly obscure the earlier direction of fault movement. 5. The rocks within these formations have been divided into at least 7 thrust slices that record a complex history of deformation caused by later folding and renewed faulting. Thrust emplacement in the hinterland does not follow a simple east-to west progression toward the foreland. 6. This complex structural history requires that several of the earlier formations (Hazens Notch, Underhill) in the pre-Silurian hinterland be abandoned or redefined. Furthermore, the Ottauquechee Formation and its equivalents (Granville Formation, Battell Member of the Underhill Formation, and carbonaceous units in the Stowe, Pinney Hollow, Hazens Notch Formations) appear to be far more extensive across strike than previously shown by Doll and others (1961). The resulting story is one of an original, relatively simple stratigraphy that was subsequently involved in a very complex structural history. The result is the distribution of units as we see them today. 7. Correlation of mineral growth and structural sequence indicate that synmetamorphic faulting occurred during peak metamorphism and continued during retrogression to produce strongly elongated chlorite clusters in the Underhill, Hazens Notch, Mt. Abraham, Pinney Hollow and Stowe Formations (Lapp, 1986; O'Loughlin, 1986; Armstrong, in progress). 8. Older thrust faults, which are largely responsible for the distribution of thrust slices, occurred before and during peak metamorphism.

Analysis of mineral assemblages by Albee (1965, 1968) and Laird and others (1984) and amphiboles by Laird and Albee (1981, 1984) suggest that peak metamorphic conditions occurred at temperatures of 500°C - 530°C and confining pressures of 4 to 5 kbars (13 to 16 km). Recent information by Laird (this guidebook) on amphiboles from central Vermont indicates that all the mafic schist west of the Green Mountain crest record medium-facies series metamorphism whereas those to the east in the pre-Silurian rocks show medium-high pressure series metamorphism. Based on a continuous and progressive sequence of structural deformation and metamorphic fabric a Taconian age (Middle Ordovician) is preferred for most of

the pre-Silurian geology of central Vermont. The young 385 m.y. 40Ar/39Ar age reported by Laird and others (1984) for muscovite and biotite at Mt. Grant and the 382 m.y. 40Ar/39Ar age for amphibole just east of the Lincoln massif (Lanphere and others (1983) may suggest prolonged recrystallization and late uplift, perhaps over a deep ramp, for this part of Vermont.

THE LINCOLN MASSIF

The Lincoln massif represents the northern most exposure of Middle Proterozoic rocks in the Appalachian Mountains in the United States (fig. 1). Although small compared to other external massifs of the Appalachians (about 50 sq. miles), it is important because 1) it contains mafic dikes associated with Late Proterozoic rifting, 2) the relations between the Middle Proterozoic and its overlying Late Proterozoic cover are reasonably well displayed, and 3) it preserves a record of the progressive deformation of the massif during the Taconian orogeny. The massif is divided into an eastern and western part by a syncline of the lower part of the Pinnacle Formation.

All of the early work on the northern part of the massif was done by Cady, Albee, and Murphy (1962) for the U. S. G. S. Lincoln Mountain 15 minute quadrangle. The southern part of the massif was mapped by Osberg (1952). Our current understanding of the massif is based on detailed mapping at a scale of 1:24,000 or larger of the northern half by Tauvers (1982a), Prah1 (1985), DelloRusso (1986) and DelloRusso and Stanley (1986).

Middle Proterozoic Basement Rocks

The Lincoln massif is largely made up of massive to well-foliated, granitic orthogneiss that becomes progressively more overprinted by lower Paleozoic deformation and metamorphism from west to east. The gneisses of the Western Lincoln massif consist of a massive, pink to light - gray weathered, medium grained biotite - quartz - feldspar paragneiss; a finer grained, strongly foliated, biotite gneiss with conspicuous magnetite octahedra and large biotite grains; and a relative thin, but continuous coarse-grained, augen gneiss containing plagioclase, quartz and microcline. The gneiss of the Eastern Lincoln massif is largely made up of fine-grained, massive, light-gray to white, quartz-plagioclase gneiss with microcline and perthite common locally. Biotite is noticeably absent or rare compared to its abundance in the gneisses of the Western Lincoln massif. Epidote, calcite and sericite are abundant particularly where the gneiss is severely altered by Paleozoic deformation. This unit is compositionally homogeneous throughout the Eastern Lincoln massif and it is not interbedded with any of the other rock types suggesting that it is of meta-igneous origin compared to the paragneiss of the Western Lincoln massif.

Metasedimentary sequences consisting of quartzite, aluminous schist, mafic schist, or augen gneiss are commonly found in isolated patches along the margin of the Lincoln massif. In the

Western Lincoln massif 1 to 5 m thick beds of massive, gray to white quartzite occur near the microcline augen gneiss that outlines a series of Paleozoic folds which extend across this part of the massif. Isolated patches of quartzite are also found near the basement-cover contact. These quartzites may correlate with similar quartzites mapped along the western part of the Eastern Lincoln massif where they are associated with light-colored tourmaline - bearing, white mica schist, and mafic schist. The quartzite commonly contains minor amounts of sericite, tourmaline, biotite, magnetite and rutile. The white mica schist is distinctive in that it contains large porphyroblasts of chloritoid and abundant needles of tourmaline.

Along the eastern boundary of the Eastern Lincoln massif, a coarse-grained, plagioclase augen gneiss is interlayered with chloritic and conglomeratic quartzite. This unit contains coarse plagioclase augen up to 4cm across. The biotite-rich matrix of the gneiss contains chlorite, epidote, and small clasts of blue quartz. This unit may represent a metamorphosed arkose in which the source area consisted of tonalite and mafic rocks. To the west within this unit, the augen gneiss becomes coarser grained and more biotite-rich near its contact with the granitic orthogneiss whereas, to the east, it becomes finer grained and more feldspathic as the basement-cover contact is approached.

Mafic schist occurs in a variety of settings that range from dikes, to fault zones, and conformable beds in metasedimentary sequences. Amphibolite layers with pyroxene, hornblende, biotite, epidote and plagioclase are found as isolated outcrops where it is interlayered with the medium grained, biotite gneiss of the Western Lincoln massif. Garnet amphibolite is associated with the bedded quartzite and tourmaline sericite schist along the western part of the Eastern Lincoln massif and may well represent mafic flows deposited on that sequence. Tauvers (1982a) mapped biotite-rich amphibolite as interlayered belts within the light-colored, granitic gneiss in the very northern part of the Eastern Lincoln massif. These belts may well represent highly-deformed Grenvillian dikes.

The remaining mafics schists are thin, highly foliated, biotite-rich schists that occurs as linear bodies with sharp contacts within the granitic gneiss of the Eastern Lincoln massif. DelloRusso (1986) has shown that these bodies are spatially related to Paleozoic fault zones, although they may have had such different origins as intrusive dikes or metasomatically altered amphibolites within fault zones.

One of the most important mafic rocks in the Eastern Lincoln massif are found as linear, metamorphosed dikes that truncate the Grenvillian foliation in the orthogneiss and in turn are truncated by the basal contact of the Late Proterozoic Pinnacle Formation. These rocks are composed of quartz, chlorite, epidote, biotite, and plagioclase. Primary amphibole is absent. Large plagioclase phenocrysts are characteristic and locally abundant although the

grains are saussuritized. Dike contacts with the granitic gneiss are sharp and xenoliths of gneiss are common in the dikes. Many of the dike margins are altered to biotite-rich schist that contain larger grains of randomly-oriented biotite. The map and contact relations of these dikes indicate that they formed after the Grenvillian orogeny, but before the deposition of the basal units of the Pinnacle Formation. They probably represent mafic dikes that were associated with the rifting of the North American continental crust during the Late Proterozoic. Current geochemical work by Raymond Coish and his students at Middlebury College suggest that they are chemically similar to the Tibbit Hill Volcanic Member of the Pinnacle Formation, a major rift-clast facies of western New England.

Late Proterozoic Cover Rocks (fig. 2)

The Late Proterozoic cover-sequence north and west of the Lincoln massif consists largely of magnetite-bearing, quartz-feldspar metawacke with matrix-supported conglomerate interbeds. Discontinuous beds of pink to salmon-colored dolostone and associated chloritic metawacke and schist are locally important. Discontinuous quartz cobble conglomerates define the contact between the Pinnacle Formation and the underlying Middle Proterozoic basement of the Lincoln massif. The conglomerates range in thickness from 1 to 20 m. Some of the thicker conglomerates are rich in large cobbles and boulders of plagioclase gneiss which suggests rapid deposition, in contrast to the thinner quartz-rich conglomerates which indicate more prolonged erosion and winnowing. These relations indicate a varied topographic relief to the basement. Based on this evidence Tauvers (1982a) suggested that these rocks were deposited in rift basins which rapidly filled and were replaced by a relatively shallow, broad, featureless basin represented by the dark shales and thin siltstones of the Fairfield Pond Phyllite. This unit in turn grades upward into siliciclastic and carbonate rocks of the stable platform.

In contrast to the depositional contact observed along the western and northern margins of the Lincoln massif, the eastern basement-contact appears to be locally tectonic, although it is poorly exposed (fig. 3). Directly east of the inferred contact, fault zones with slivers of Middle Proterozoic granitic gneiss have been mapped within the basal 500 m of the cover sequence. Conglomerate is rare along the basal part of the eastern cover sequence. One outcrop at South Lincoln, Vermont, exposes a 40 m section of highly-deformed, matrix-supported conglomerate made up of pebbles and cobbles of quartzite and gneiss (Tauvers, 1982; Strehle and Stanley, 1986). The overlying rocks of the Hoosac Formation are made up of quartz-feldspar metawacke interlayered with biotite schist and muscovite-biotite schist some of which contain garnet. Albitic schist, which is typical of the Hoosac to the south along the eastern side of the Green Mountain massif, is present but minor compared to the total section at Lincoln township. These relations suggest that the cover sequence

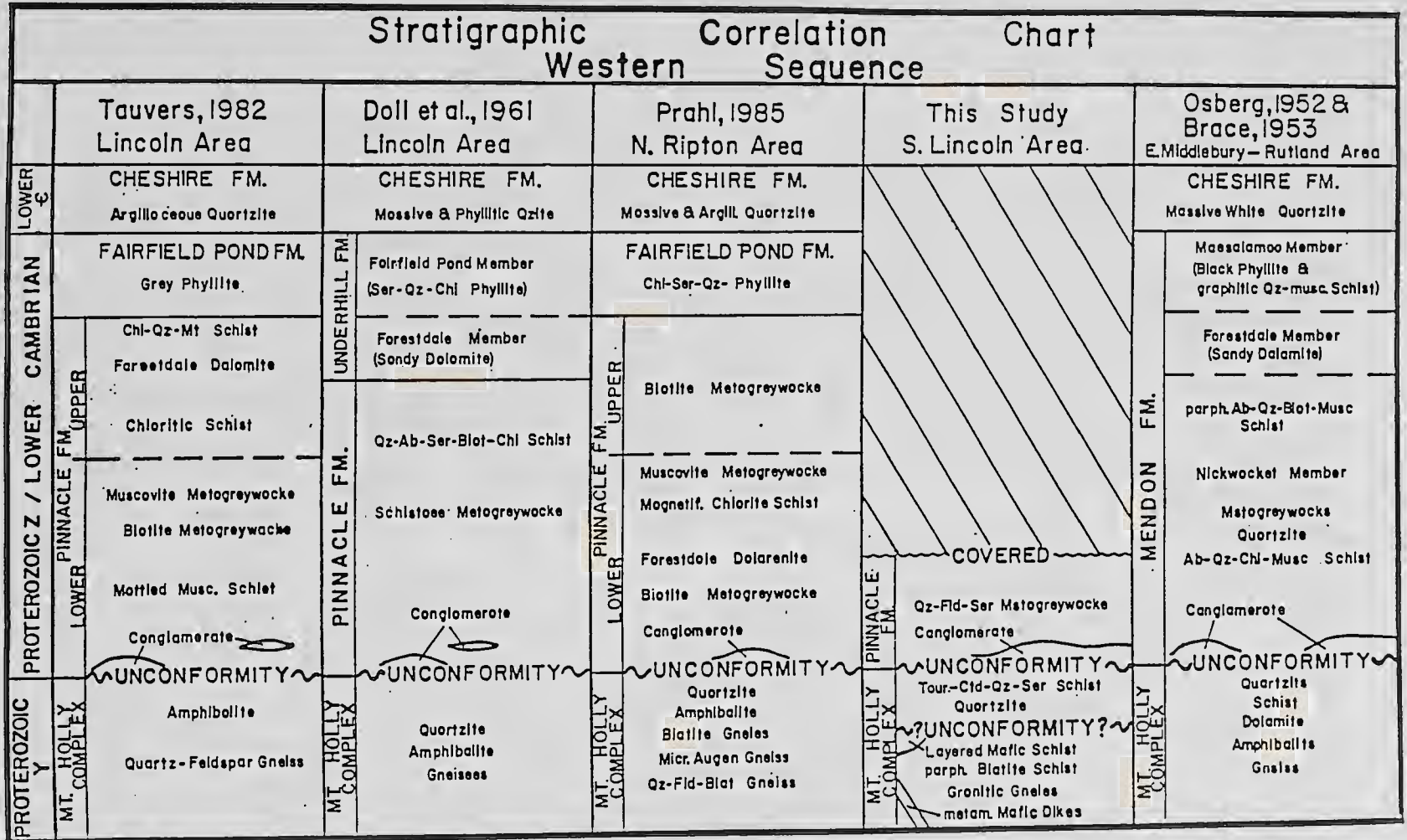


Figure 2

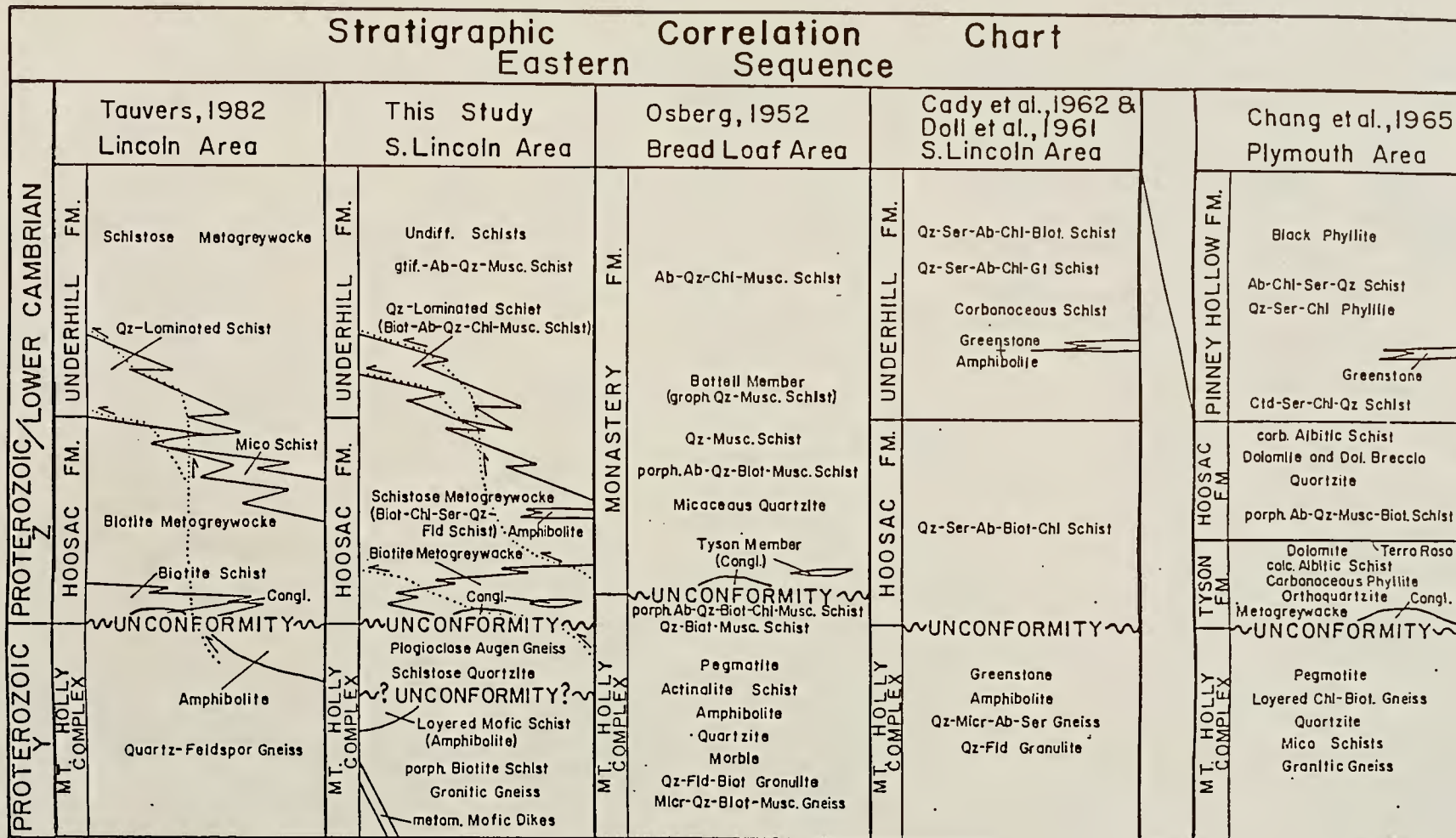


Figure 2

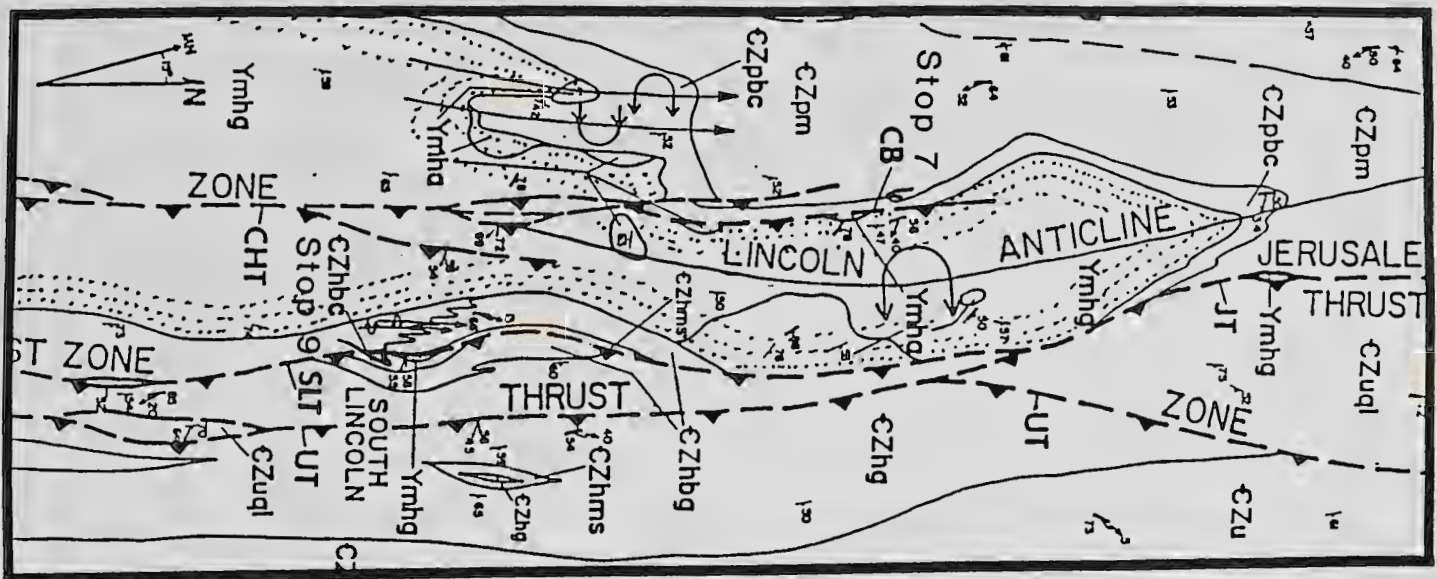


Figure 3

Geologic map of the northern end of the Eastern Lincoln massif illustrating the contrast between the western and eastern border. The western boundary is marked by an angular unconformity that is vertical or overturned to the east. A prominent basal conglomerate is found along this contact and faults are rare. In contrast, major thrust faults mark the eastern boundary. Lithic designators are the following: Ymhg - granitic gneiss of the Mt. Holly Complex, Ymha - amphibolite or amphibolite-rich parts of the Mt. Holly Complex, CZpbc - basal conglomerate in the Pinnacle Formation, CZpm - muscovite-rich metawacke of the Pinnacle Formation, CZu - Underhill Formation, CZuql - quartz laminated schist of the Underhill Formation, CZhg - schistose metawacke of the Hoosac Formation, CZhbg - biotite metawacke of the Hoosac Formation, CZhms - mafic schist in the Hoosac Formation. CB - Crash Bridge, CHT - Cobb Hill thrust zone, JT - Jerusalem thrust fault, UT - Underhill thrust fault, SLT - South Lincoln thrust fault.

immediately overlying the Lincoln massif is largely made up of the rift clastic rocks that were deposited in a site that was located between the basal sections presently exposed along the western and eastern boundaries of the Middle Proterozoic rocks of the Green Mountain massif.

Structural Relations

Grenvillian Structures Few major structures of Grenvillian age can be recognized in the Middle Proterozoic rocks of the Lincoln massif. Perhaps the most important feature is the contacts of the metasedimentary sequences in the massif with the adjacent gneisses. In the Eastern Lincoln massif this contact appears to be an unconformity since the contact is sharp and it is not cut by dikes or sills of the granitic gneiss. Where quartzite overlies the gneiss, however, clasts of the underlying gneiss are absent indicating that mature, quartz-rich sands were deposited on a gentle erosional surface. Similar relations are also observed where quartzite is in contact with gneiss along the margins of the Western Lincoln massif.

The metasedimentary sequence of plagioclase augen gneiss, quartzite, and conglomerate along the eastern margin of the Eastern Lincoln massif suggests quite a different setting. The sedimentary interpretation of the augen gneiss (arkose) suggests that this sequence of rocks represents the rapid deposition of very immature material from an older tonalitic terrane. The interlayered quartzites could represent mature sands deposited during periods of relatively slow sedimentation. Although the contact with the underlying granitic gneiss is not exposed, the increase in grain size towards this contact suggests a fining-upward sequence.

The only major tectonic structure of Grenvillian age that has been mapped in the northern part of the Lincoln massif is an east-plunging fold delineated in a quartzite - tourmaline schist - mafic schist sequence on the western side of the Eastern Lincoln massif.

Although the Grenvillian fabric in the gneisses of the Lincoln massif is overprinted by Paleozoic structures, it is easily identified by a coarse mineral layering defined by alternating quartz-rich and feldspar-rich bands containing well-developed quartz rods. This layering is discordant to the basement-cover contact throughout the Lincoln massif although it is more difficult to distinguish from the younger Paleozoic foliation along the eastern margin of the massif.

Taconian (?) Structures

A Taconian(?) foliation marked by aligned micas in the gneiss becomes progressively more pervasive as it is traced eastward across the Lincoln massif. Chlorite and biotite are present within this foliation where the host rocks are richer in iron and magnesium than in the sericite - bearing granitic rocks which

underlie much of the Eastern Lincoln massif. Available isotopic age information from the surrounding area suggests a Taconian age for this foliation (summarized by Sutter and others, 1985). This foliation was formed when the Lincoln massif was folded into a series of doubly-plunging, north-trending anticlines separated by a intervening syncline. This syncline divides the massif into the Western and Eastern massifs.

The structural features of the Western and Eastern massifs differ in degree rather than style. Broad folds outlined by the microcline augen gneiss and quartzite dominate the Western Lincoln massif. A few faults have been mapped, but they are of minor importance. In contrast, faults dominate the Eastern Lincoln massif where they become more numerous and pervasive toward the eastern-cover sequence. These faults are of two kinds, east-directed reverse faults and normal faults associated with the flexural-slip folding of the western part of the Eastern Lincoln massif and west-directed thrust fault associated with the Cobb Hill thrust zone in the Eastern Lincoln massif and the imbricate slices of the basal part of the eastern-cover sequence. Well preserved asymmetrical fabrics along all of the fault zones record the consistent east - over - west direction of motion across the massif. The structural profile across the Lincoln massif therefore records the progressive failure of a representative part of North American crust as it was incorporated into the collisional zone during the Taconian orogeny. This detailed mapping has led to the interpretation that the Lincoln massif developed as a series of large-scale basement folds, which were progressively imbricated from east to west and subsequently transported westward along the Champlain thrust zone during the Taconian orogeny.

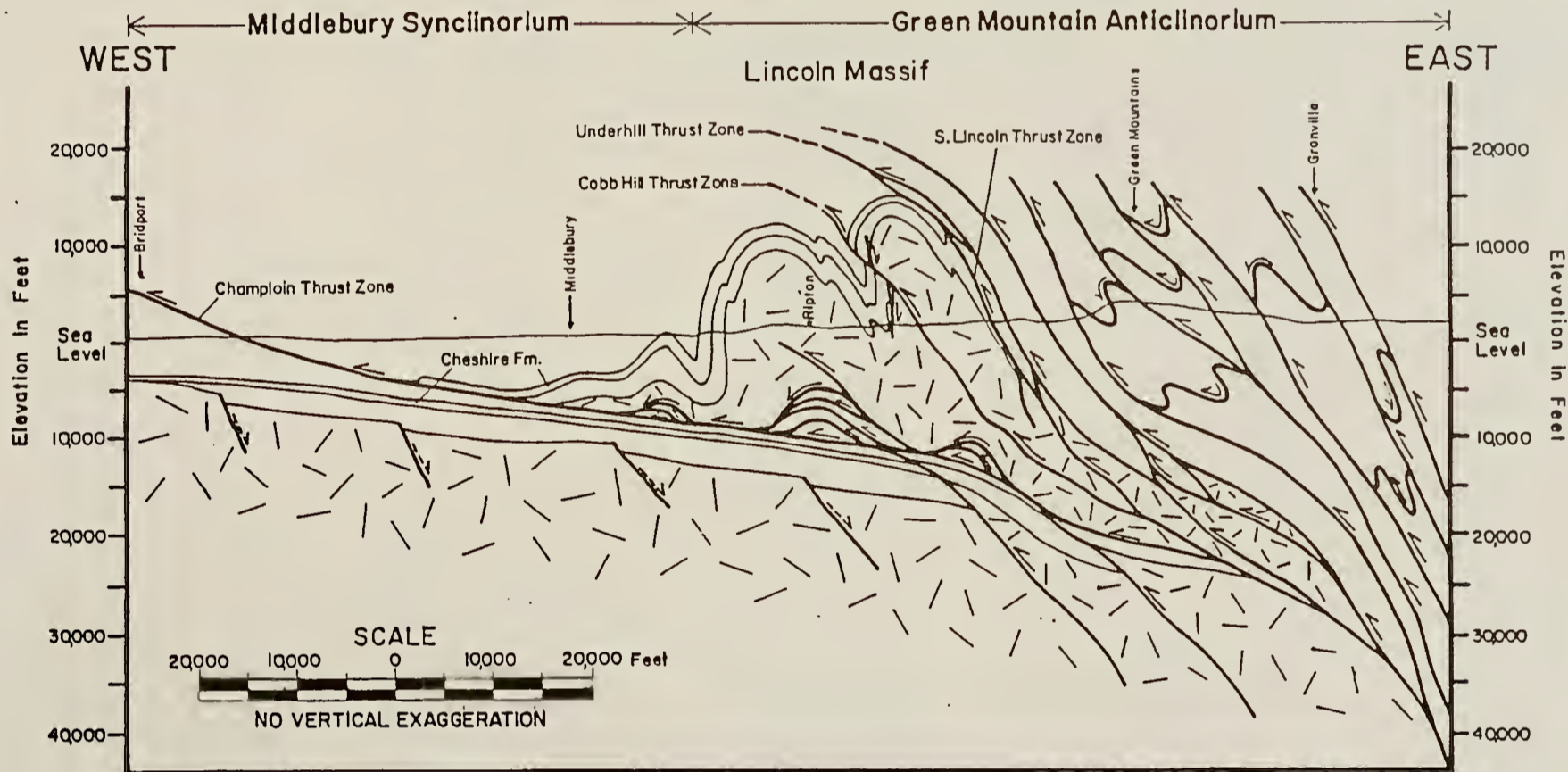
Metamorphism

Few definitive indicators of Grenvillian metamorphism are present in the Lincoln massif because the rocks are largely granitic in composition. Taconian metamorphism has severely altered the original Grenvillian assemblages. The presence of garnet amphibolites and pyroxene-bearing amphibolites suggest that Grenvillian metamorphism was at least at the conditions of the amphibolite facies or perhaps even higher. These assemblages were recrystallized under greenschist to epidote-amphibolite facies conditions during the Taconian orogeny.

EASTERN COVER SEQUENCE

The pre-Silurian eastern cover sequence consists of the Hoosac, Underhill, Mt. Abraham, Hazens Notch, Pinney Hollow, Ottauquechee, Stowe and Missisquoi Formations of Doll and others (1961). These formations are largely made up of gray to green, non rusty-weathering schists; rusty-weathering, brown to dark gray to black carbonaceous schist; aluminous, silvery chloritoid-bearing schist; metawacke; quartzite; granofels; and abundant mafic schist (greenstone). Albite porphyroblasts are common in many of the

INTERPRETIVE CROSS-SECTION ALONG LATITUDE 44°00'N
THROUGH THE LINCOLN MASSIF—CENTRAL VERMONT



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Figure 4

Interpretative east-west cross section along latitude 43°00' N through the Lincoln massif, central Vermont. This cross section shows the Champlain thrust fault projected eastward beneath the Lincoln massif where it roots to the east of the exposed basement. Part of the autochthonous platform (CP) and rift-clastic (RC) sequence is shown to exist beneath the Champlain thrust fault. Anticlinal stacks or duplexes shown above the Champlain thrust fault are speculative, but structures of this type are thought to be an important element in the development of large scale basement folds. This cross section shows that Lincoln massif as a series of basement folds that were progressively imbricated from east to west and subsequently transported westward on the Champlain thrust fault during the latter part of the Taconian orogeny (Stanley and Ratcliffe, 1985). Renewed movement on the Champlain thrust fault or a deeper thrust fault during the Acadian orogeny can not be ruled out and may in fact be suggested by the 385 m.y. isotopic age from muscovite and biotite from the crest of the Green Mountains (Laird and others, 1984). Although the fault geometry in the eastern cover sequence is diagrammatic, it is based on the fact that premetamorphic (represented by the folded lines) and synmetamorphic (gently curved lines) thrust faults dominate this sequence.

units. Serpentinite and related rocks are largely found in the Ottauquechee Formation, other carbonaceous schists, and the Moretown Member of the Missisquoi Formation. Many of these rocks are found along faults (Stanley and others, 1984). Although the ultramafic rocks could have originally been olistoliths from a large slice of oceanic crust comparable to those mapped in Quebec and Newfoundland, subsequent deformation has totally obscured such an origin. Blue quartz is found in the metawackes of the Hoosac, Pinney Hollow, and Ottauquechee Formations.

The mafic schists are important because they are easy to map and provide valuable geochemical, petrologic and isotopic age data that helps to constrain the tectonic evolution for western New England (Coish and others, 1985, 1986; Laird and others, 1984, for example). Those in the Pinney Hollow and Stowe are particularly helpful in locating premetamorphic and synmetamorphic faults. Many of the mafic schists that we have mapped west of the Green Mountain crest in the Hazens Notch and Underhill Formations are slices or slivers on faults.

Stratigraphy

The rocks in the pre-Silurian eastern cover sequence have been viewed as a coherent although highly deformed sequence of units that become younger to the east (Doll and others, 1961). Thus the Hoosac Formation to the west was considered the oldest (Late Precambrian to Lower Cambrian) and the Missisquoi Formation was considered the youngest (Ordovician). Fossils are absent from the sequence and the supposed ages are based on correlation to similar fossiliferous rocks either in Quebec or the Giddings Brook slice of the Taconic allochthons. These correlations are highly speculative since none of the pre-Silurian units in eastern Vermont can be traced uninterrupted to the fossiliferous localities.

Recent detailed mapping in a number of areas in Vermont have clearly demonstrated that the pre-Silurian section is marked by major faults that are largely premetamorphic (or early metamorphic) and synmetamorphic in age (Stanley and others, 1984; Tauvers, 1982; DiPietro, 1983; Loughlin and Stanley, 1986; Lapp and Stanley, 1986; DelloRusso and Stanley, 1986; Prewitt, Armstrong, Kraus, in progress). Similar findings have been reported by Zen and others (1983) in western Massachusetts and Karabinos (1984) for the basal part of the sequence along the eastern border of the Green Mountain massif in southern Vermont. Thus the sequence is not coherent and the age assignments, stratigraphic sequence and geological evolution are all in doubt and must be reinterpreted in light of these findings. Several authors have suggested that these rocks represent the deformed remains of an accretionary wedge that developed during the Taconian orogeny (Rowley and Kidd, 1981; Stanley and others, 1984; Stanley and Ratcliffe, 1985). If true, we might expect a very complex tectonic history superposed upon an original sedimentary sequence that formed along the continental margin of ancient North America.

During the past 4 years our work in central Vermont has shown that many of the mappable units occur in discontinuous lenses and belts which are repeated across strike. The contacts between most of these units truncate smaller units such as greenstone, metawacke, and schist that we map in previously recognized formations. The carbonaceous schists form an excellent example of this relationship. Carbonaceous rocks are found in the Underhill, Granville, Hazens Notch, Pinney Hollow, Stowe and are characteristic of the Ottauquechee Formation. Each of these units are rusty weathering, are marked by a certain amount of sulphidic stain, and contain varying amounts of graphite. Black albite is found in all the units even the Ottauquechee where albite is certainly subordinate compared to the other carbonaceous units. Black, gray, and white quartzite is present in all the units although it is most abundant in the Ottauquechee. Interestingly enough, rocks with all of these characteristics were earlier mapped in the Ottauquechee Formation (Osberg, 1952; Cady and others, 1962). Osberg (1952) recognized the similarity among some of these belts and on his map showed synclines of Ottauquechee throughout the eastern part of the Pinney Hollow. On the Bedrock Geological Map of Vermont (Doll and others, 1961), however, they are shown as carbonaceous schists in the Pinney Hollow. We believe this was a unfortunate choice because the map shows the Ottauquechee as a single linear belt separating the Pinney Hollow from the Stowe rather than several belts which can be recognized across strike. We would go even farther than Osberg did and suggest that the Ottauquechee or rocks that are lithic correlative to it occur as far west as the Underhill (Battell Member) and as far east as the carbonaceous schists in the Stowe. Furthermore, we have been able to demonstrate at the scale of 1:12,000, or larger, that the carbonaceous schists truncate units in the adjacent formations. This relationship, combined with fabric information, indicates that the contacts are either premetamorphic, early metamorphic, or synmetamorphic faults. We propose, therefore, that the Battell Member of the Underhill, the Granville Formation, the carbonaceous schists in the Hazens Notch, Pinney Hollow, and Stowe were originally part of one continuous deposit that extended eastward from the ancient continental crust of North America to oceanic crust. Changes in sedimentary facies did exist but its basic carbonaceous, sulphidic character persisted. Their present distribution is the result of a very complex history of deformation.

Another example is provided by the silvery green to gray green to light gray schists of the Hazens Notch, Pinney Hollow and Stowe. Albite porphyroblasts and deformed quartz veins are characteristic of each of these units. Magnetite is common. Greenstones that look identical in outcrop are common in the Pinney Hollow and Stowe. They are different than those in the Hazens Notch where the greenstones are darker in color and commonly albitic. Osberg (1952) recognized this similarity in the Pinney Hollow and Stowe, but showed them as separate formations because of their differences and the fact that the Ottauquechee consistently separated the two. We suggest, however, that the Pinney Hollow and Stowe were once

part of the same deposit. We further suggest that the light gray, magnetite-bearing, albitic schist of the Hazens Notch represents a western facies of the Pinney Hollow. Similar rocks also are found in large tracks of the region mapped as Underhill (Thompson and Thompson, this guidebook; Walsh, personal communications). This proposal requires therefore that the Ottauquechee overlie the Pinney Hollow and Stowe.

Figure 5 is a stratigraphic correlation chart for the pre-Silurian of eastern Vermont. The position of the respective columns in central Vermont is based on recent mapping and the discussion in the foregoing paragraphs. The correlations with the sequence in the Taconic allochthons and the eastern limb of the Berkshire massif is based on lithic similarities and sequence among the respective areas. The correlation between the rocks of Group 3 slices of Stanley and Ratcliffe (1985) and the eastern Vermont sequence has been proposed by a number of earlier workers (Keith, 1934; Thompson, 1967). The stratigraphic arrangement presented here is therefore tentative and is undergoing modification during each field season.

A number of important aspects of this correlation chart require explanation because they differ from the more traditional view of the pre-Silurian stratigraphy in Vermont. They are the following:

1. Major thrust zones separate each of the columns. Thrust faults are also present within many of the columns. For most of the units shown in this manner it is possible to locate their respective root zone. For example, the Mt. Abraham Schist roots along the western boundary of the Pinney Hollow and the carbonaceous schist of the Hazens Notch roots along the western boundary of the Ottauquechee Formation. The serpentinites and related rocks are presently found along faults. The only exception to this rule are the serpentinites found in the eastern part of the Moretown near Roxbury. These bodies are largely found within greenstone and direct evidence of throughgoing faults is absent. Evidence concerning their original mode of emplacement for all the ultramafic rocks is uncertain.

2. The albitic rocks in the noncarbonaceous rocks of the Hazens Notch, Pinney Hollow and Stowe are considered to be equivalent. Similar rocks also exist in the Underhill. These rocks are correlated to parts of the Greylock Schist (CZga) and Hoosac Formation (CZhab, CZh) of western Massachusetts and the Netop Formation of southwestern Vermont.

3. The Battell Member of the Underhill (not shown) and the carbonaceous schists in the Pinney Hollow are assigned to the carbonaceous schist of the Hazens Notch Formation (O'Loughlin and Stanley 1986; Lapp and Stanley, 1986). These units are a facies of and equivalent to the Ottauquechee Formation. It is identical and traceable into the Granville Formation of Osberg (1952). Although we assigned these carbonaceous schist to the Hazens Notch, we now believe that the term "Granville" would possibly be a better

EXPLANATION FOR FIGURE 5- The lithic symbols shown in each column of figure 5 are explained by geographic locality starting with the oldest unit.

West Central Vermont - Ymh- gneissic gneisses of the Mt. Holly Complex, Y - mylonitic gneiss at South Lincoln, CZpbo - basal conglomerate in the Pinnacle Formation, CZpbg - biotite wacke, CZpfd - Foreatdale Marble, CZpm - muscovite wacke, CZpcl - chlorite wacke and chlorite schist, CZhbo - basal conglomerate in the Hoosac Formation, CZhbg - biotite wacke, CZhg - schistose wacke, CZugl - quartzite laminated schist of the Underhill Formation, CZufg, well foliated wacke, CZu - mixed unit consisting of garnet, chlorite, muscovite schist, thin quartzite, CZuga - garnetiferous schist, CZfp - gray, finely laminated phyllite of the Fairfield Pond Formation, Cca - argillaceous quartzite of the Cheahira Formation, Ccm - massive, bedding quartzite of the Cheshire Formation, Cd - Dunham Dolomite, Cm - Monkton Quartzite, Cw - Winoski Dolomite, Cda - quartzite and dolomite of the Denby Formation, Cce - Clarendon Springs Dolomite, Lower Ordovician Beakmantow Group oolite of limestone, dolomite and minor quartzite and shale of the Sholburne Formation (Oo), Cutting Dolomite (Oc), Baacom Formation (Ob), and Chipman Formation (Ocb), Omm - Middlebury Limestone, Oo - Orwell Limestone, Ogf - Glens Falls Formation, Owl - fossiliferous limestone lenses near the base of the Walloomaac Formation, Oh - black carbonaceous slate and phyllite of the Hortonville Formation, Ow - dark gray, graphitic schist and phyllite of the Walloomaac Formation, Oww - Whipatock Breccia Member of the Walloomaac Formation.

Taconic Allochthon - Group 1 and 2 - CZnr, Reneealner Graywacke Member of the Nassau Formation, CZnv - metabasalt and basaltic tuff, CZnm - lutroite, yellowish-green, purple laminated chloritoid - chlorite phyllite of the Mettawee Member, CZnt - Truthville Slate, CZnb - Bomoseen Graywacke Member, CZnzh - Zion Hill Quartzite, CZnmp Mud Pond Quartzite, Cwc - dark grey to black, pyritiferous and calcareous slate with thin sandy laminae of the West Cauleton Formation, Chh - bluish-gray weathering black sulphidic slate and chert, Opo - white weathering, well laminated gray slate and chert, On - Normanekill Formation, Onmm - black slate and thin bedded chert of the Mount Merino Member, Onir - red and green slate of the Indian River Member, Onag - Aoutin Glen Graywacke Member.

Group 3- CZga, Light green to grey, white albitic schist with some magnetite, chlorite granofels of the Creylock Schist, CZne - Netop Formation in the Dorset Mt. area, CZgb - black of gray chloritoid or stilpnomalane schist, quartz knotted schist, lenses of feldspathic quartzite, conglomerate, and pink dolomite, CZg - light green, lutroite chloritoid phyllite and minor beds of white albitic schist, CZac - St. Catherine Formation in the Dorset Mt. area.

East Limb of the Berkshire massif - CZhb (CZh) - gray to white schist spotted schist of the Hoosac Formation, CZhgeb (CZhge) - green schist magnetite schist, CZhg - green aluminous chloritoid schist. The symbols in parentheses refer to designations of equivalent units on the Geologic Map of Massachusetts (Zen and others, 1983).

Green Mountain anticlinorium - Northfield Mountains (Many of the symbols are repeated from column to column. These symbols are explained in the first column that they appear in as the columns are read from left to right)

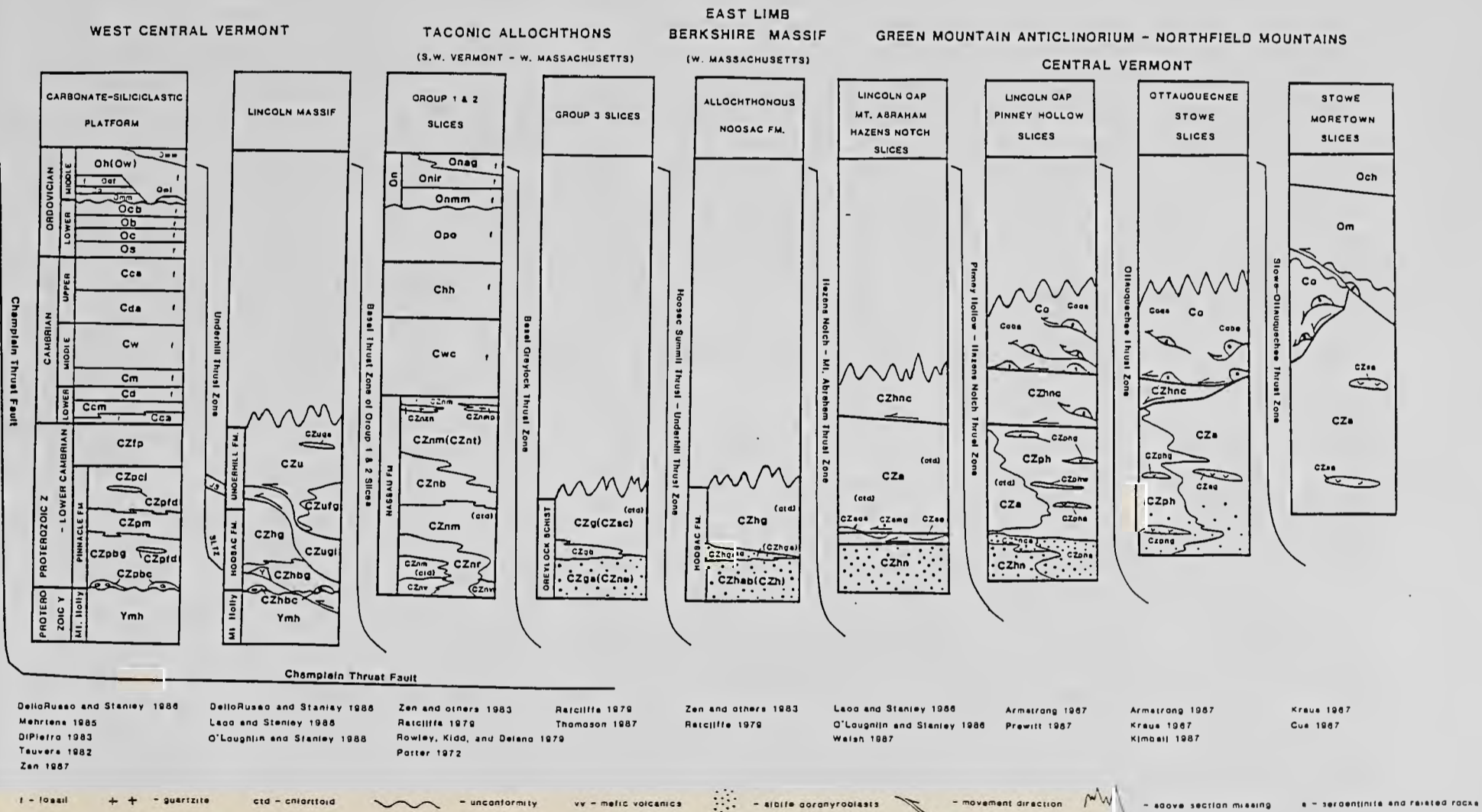
Lincoln Cap-Mt. Abraham-Mezena Notch area - CZhn silvery white to dark green to black schist spotted with white schist, minor white to gray laminated quartzite of the Mezena Notch Formation. Mafic schist abundant in the eastern part, CZhnc - rusty weathering, black albitic schist with widespread graphite and minor thin black or gray quartzite. CZhnc has been called the Battell Member of the Underhill (Doll and others, 1961) and the Grenville Formation (Osberg, 1952). Cza - Mt. Abraham Schist. Silvery colored peragonite - muscovite - chloritoid - chlorite (garnet) schist with a distinctive poorly seen on the schistosity, CZemg - similar to the main body of Mt. Abraham Schist but with abundant magnetite and chloritized garnet, CZaga - muscovite - chlorite schist of the Mt. Abraham Schist with large porphyroblasts of garnet and minor chloritoid,

Lincoln Cap-Pinney Hollow area - Additional symbols are: CZhnca - rusty weathering, dark grey to black schist with discontinuous patches of graphite. Traceable into the Grenville Formation. CZph - silvery green muscovite - chlorite - quartz schist of the Pinney Hollow Formation, CZphe light gray muscovite - chlorite - quartz schist with albite porphyroblasts, CZphg - mafic schist of the Pinney Hollow, CZphw - gray wacke with minor blue quartz, Co - black, pyritiferous and graphitic schist of the Ottauquechee Formation, Cobq - black and gray quartzite of the Ottauquechee Formation, Coqa - sandy quartzose schist of the Ottauquechee Formation, s - serpentinite and talc-carbonate rock.

Ottawaquechee - Stowe area - Additional symbols are: CZe - Stowe Formation - silvery green, muscovite - chlorite - quartz schist identical to CZph, some schists are richer in chlorite, CZeg - mafic schist in the Stowe Formation.

Stowe-Moretown area - Om, pinetriped schist and mafic schist of the Moretown Formation, Och - black, graphitic schist and thin quartzite of the Crem Hill Formation.

LITHOTECTONIC CORRELATION CHART FOR CENTRAL VERMONT



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Figure 5

Stanley and Armstrong 1987

choice. Much of the carbonaceous schist within the Hazens Notch proper is considered to be a highly deformed thrust slice although some of it may well be a facies of the noncarbonaceous Hazens Notch. According to our present thinking the "allochthonous" part of the carbonaceous schist would be reassigned to the Granville Formation. The stratigraphic column for the Lincoln Gap-Pinney Hollow slices does show some carbonaceous schist as part of the Hazens Notch. Resolution of this problem awaits further work.

4. The Ottauquechee Formation is shown to overlies the Pinney Hollow and Stowe. This contact is a major premetamorphic or early metamorphic fault zone. The root zone for the Ottauquechee slice has yet to be found but it is thought to be located somewhere in the eastern part of the belt mapped as Stowe by Doll and others (1961) or along the western part of the Moretown.

5. The Ottauquechee Formation is an enigma. Because it contains ultramafic rocks it is thought to have been deposited far to the east on oceanic crust. It contains, however, much quartzite and quartz-rich schist which would suggest a nearby continental source. The presence of rare blue quartz supports this view. How do rocks of this type end up with ultramafic rocks? Did the quartz come from a continental source to the east or north and not to the west as traditional is thought to be the case for rocks with blue quartz? Could some of the rocks in the Ottauquechee represent trench clastics eroded from an older accretionary wedge?

6. The Mt. Abraham Schist is definitely allochthonous and roots along the western part of the Pinney Hollow where similar rocks are found in the Rochester quadrangle. These rocks are correlated to the chloritoid-bearing schists in the Group 3 slices of Stanley and Ratcliffe (1985) and the Saint Catherine Formation of Thompson (1967).

It should be evident from the foregoing discussion that many of the units in the pre-Silurian sequence west of the Moretown are thought to be lateral equivalents of each other rather than a simple eastward-younging sequence in the earlier view shown on the Geologic Map of Vermont (Doll and others, 1961). We do believe, however, that the sequence originally became younger to the east because the average age of the sediment cover becomes younger toward the ridge in an expanding ocean (fig. 6). This sequence was then deformed into a complex arrangement of units in a setting that is similar to an accretionary wedge between two converging plates. Our mapping only begins to show the complexity of this deformation. Much of the earlier history is destroyed by the process itself.

Structure - One of the important questions posed earlier in this paper was the nature of the shortening mechanism in the metamorphosed pre-Silurian hinterland. Was folding the major mechanism or was faulting? To what degree are both mechanisms important? And finally, how are they related, if evidence for both exist? Answers to these questions depend on scale, the

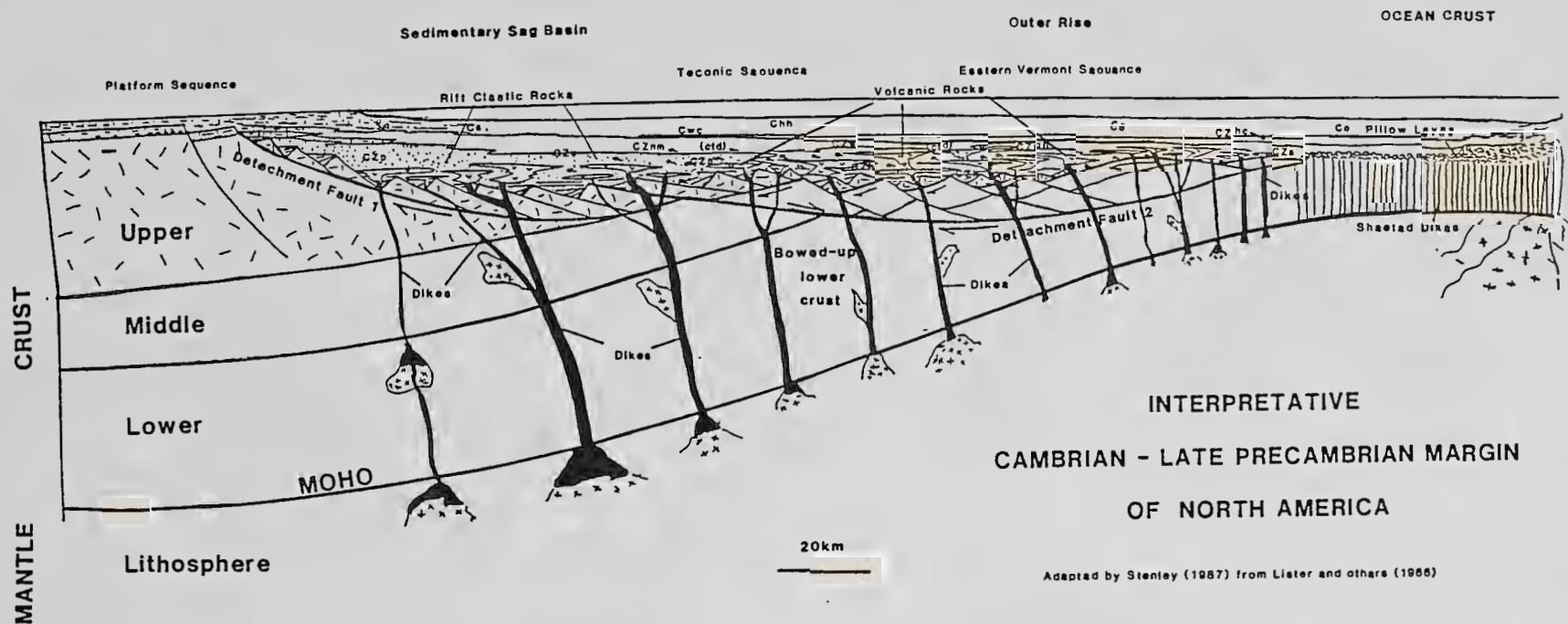


Figure 6

Speculative diagram showing the inferred stratigraphic relations among the Late Precambrian and Cambrian rocks along the ancient North American margin. The reroft model is taken from Lister and others (1986). Evidence is not available to support this particular rift model rather than others for the New England Appalachians. It is used, however, simply as a template to show the inferred stratigraphic relations of the overlying cover rocks. Lithic designators are the following: CZp - Pinnacle Formation, CZu - Underhill Formation, CZn - Nassau Formation, CZnm - Metawes Member of the Nassau Formation, CZs - Mt. Abraham Schist, CZph - Pinsky Hollow Schist, CZhc - Stowe Formation, CZhc, carbonaceous schist of the Hazens Notch Formation and its equivalents (Granville Formation), Cs - Sweetsburg Formation, Cwc - West Castleton Formation, Chh - Hatch Hill Formation, Co - Ottauquechee Formation. Other symbols are: ctd - chloritoid, a - albitic rocks, vv - mafic volcanic rocks, ++ gabbroic rocks. Equivalent carbonate and siliciclastic rocks are shown on the left (western) side of the diagram.

Table 1

Criteria for Pre-, Syn-, and Post- Metamorphic Faults

PRE-METAMORPHIC FAULTS Regional metamorphism occurs after movement. Early mineral assemblages are absent.

Criteria

1. Mappable lithic units are truncated on the hanging wall and foot wall of the fault zone.
2. Silvers of rocks foreign to either the hanging wall or foot wall (i.e. ultramafic rocks, felsic plutonic rocks, Proterozoic Y gneiss, exotic metasedimentary rocks).
3. Preserved fault zone fabrics are rare or vary difficult to recognize.

SYN-METAMORPHIC FAULTS Metamorphic recrystallisation occurs during faulting.

Criteria

1. Criteria 1 and 2 listed for pre-metamorphic faults.
2. Ductile fabrics preserved along fault zones.
 - a. Sharp lithic contacts that truncate internal fabric (i.e. older foliations, compositional layering, quartz veins).
 - b. Prominent anastomosing foliation is parallel to or is at a low angle to the fault surface.
 - c. Prominent lineation formed by elongate minerals or mineral clusters.
 - d. C-S fabrics.
 - e. Asymmetric folds, sheath folds, and/or reclined folds with sheared off limbs.
 - f. Mylonitic fabrics marked by fine grained, recrystallized minerals compared to larger grains in the surrounding rocks. Porphyroclasts may be present. Evidence for ductile flow is best seen in quartz and feldspar where undulose extinction, deformation twine, deformation lamellae, deformation bands, subgrains, polygonization and sutured grain contacts suggest dislocation glide, dislocation creep and grain boundary migration as important deformation mechanisms. Continued strain during recrystallisation is marked by elongate new grains with avoidance of internal deformation.
 - g. Absence of cataclastic fabrics.

POST-METAMORPHIC FAULTS Movement occurs after regional metamorphism, but retrograde metamorphism may occur along the fault zone.

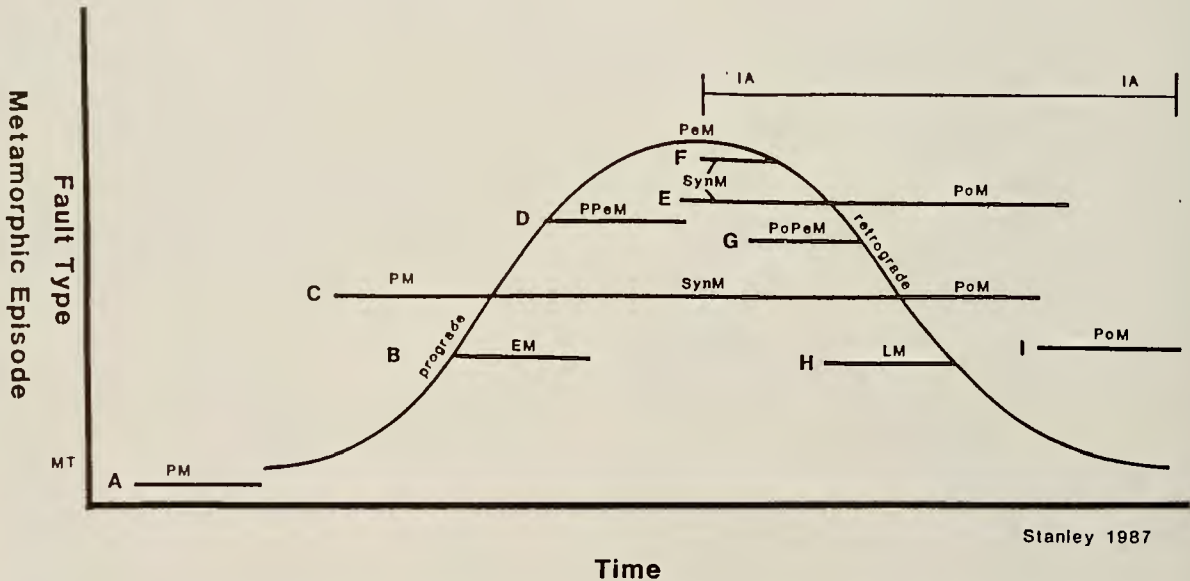
Criteria

1. Criteria 1 and 2 listed for pre-metamorphic faults.
2. Strongly oriented, fault zone fabrics are well preserved and will depend upon the depth at which movement occurred. Brittle fabrics reflect movement within several kilometers of the surface whereas fabrics typical of the brittle-ductile transition reflect deeper levels.
 - a. Brittle-ductile fabrics consist of a mixture of minerals such as quartz and feldspar in which some of the minerals deform by ductile mechanisms (quartz) and others by fracture (feldspar). The overall mechanical behavior of the rock depends on the dominant mineral.
 - b. Retrograde minerals along fault zones. In pelitic rocks this may consist of fine grained mica (phyllonites). In gneisses the mica-rich layers are rich in white mica, chlorite, and/or biotite. Finely crushed feldspar may also be present.
 - c. Brittle fabrics consist of grains which are angular with random size and orientation. Microfractures are abundant.

COMPOUND FAULTS Repeated movement occurs under different metamorphic conditions.

Criteria

1. Evidence for post-metamorphic faults superposed on older syn-metamorphic faults are readily preserved and easily recognized. All other examples of fault-superposition are unlikely to be recognized because of subsequent metamorphism.



A simplified graph showing the schematic relations of eight representative fault types and a single metamorphic episode in the hinterland. PM represents pre-metamorphic faults, EM represents early metamorphic faults, PPeM represents pre-peak metamorphic faults, SynM represents syn-metamorphic faults, PoPeM represents post-peak metamorphic faults, LM represents late metamorphic faults, and PoM represents post-metamorphic faults. Faults C and E represent faults with a composite history. The distinction between "pre-metamorphic" and "early metamorphic" faults is commonly difficult to determine because metamorphism obliterates much, if not all, of the fabric information. The criteria that are listed under "pre-metamorphic faults" would apply therefore to "early metamorphic" faults although some fabric information might be preserved as inclusions in porphyroblasts. The distinction between "early metamorphic" and "pre-peak metamorphic" faults as well as "post-peak metamorphic" and "late metamorphic" faults is a matter of degree and subjective evaluation of the fault-zone fabric and mineralogy. All faults are distinguished by criteria 1 and 2 for pre-metamorphic faults. Oriented, fault-zone fabrics are preserved for those faults that were active during and after the peak of metamorphism and may well apply to "pre-peak metamorphic" faults. Ductile fabrics are formed during the metamorphic episode, while brittle fabrics develop before and after metamorphism. Transitional fabrics (brittle-ductile) are formed in the region where the fault crosses the metamorphic curve after the metamorphic peak has been achieved. PeM represents the peak of metamorphic intensity. MT represents the "metamorphic threshold" IA represents the range of possible isotopic ages associated with the metamorphic episode.

availability of easily - recognized units, and the degree to which critical fabrics are preserved. Metamorphism, despite producing minerals that can be used to determine the physical conditions of deformation, commonly anneals and simplifies preexisting fabrics so that much of the deformational history is erased. Thus, we are left with lithic units whose field characteristics are sufficiently distinctive that they can be recognized and mapped from place to place at a scale suitable to show their mutual relations. For example, if we can show that a series of metasedimentary units can be traced consistently through a complex fold pattern and can be used to predict larger structures, then a fair chance exists that they are part of an original depositional sequence. This conclusion is strengthened if their contacts are gradational, even over a small distance, and show no signs of intense, concentrated shear. On the other hand, if a metasedimentary or metavolcanic unit is in contact with different units then the contact is either an unconformity or a fault. Here again the fabric along the contact may be helpful. If the contact is a synmetamorphic or postmetamorphic fault, then well oriented fabrics with displacement criteria are commonly present (table 1). Premetamorphic or early metamorphic faults are more difficult to recognize because fault-related fabrics are largely destroyed by metamorphism. Here we must depend upon the mutual relations of four or more lithic units. Such faults are recognized when two or more units on both the footwall and hangingwall are truncated along the contact (i.e. the fault, table 1). This criterion applies equally well with synmetamorphic and postmetamorphic faults, but in these cases additional evidence is available in the fault fabric.

Additional problems must be solved when analyzing the simple-shear fabric along a proposed fault. The presence of a well developed schistosity and a prominent lineation is really not sufficient since these types of fabrics can be formed equally well by pure shear (flattening) as they can by simple shear. Such asymmetrical fabrics as drag folds, C-S fabrics, mica fish, porphyroclasts with asymmetrical tails, and rotated porphyroblasts are the type of evidence that indicate simple shear rather than flattening. If a contact has features such as these, it is quite likely that the contact in question is a fault, particularly if it truncates other rock units. On the other hand, the presence of symmetrical fabrics such as overgrowths or "beards" on porphyroclasts and the absence of asymmetrical fabrics with a consistent sense across a contact would indicate that the contact is simply a product of severe flattening (pure shear). An offset of an older marker can occur in situations like this but here it is due to volume loss during flattening rather than simple shear displacement along a fault. In many situations not all the evidence is available to determine if a given contact is a fault or not.

Table 1 lists some of the criteria that have been used to recognize faults in the pre-Silurian hinterland of central Vermont. With the possible exception of some of the old faults that we map as "pre- or early metamorphic" faults, most of the faults that we have mapped in the central Vermont transect were formed under elevated

temperatures (300°C to 530°C) and pressures (2 Kbar to 5 Kbar or approximately a depth of 6 to 16 km) and thus are considered to have formed by ductile or semiductile processes. These faults are divided into eight groups depending on when they occurred relative to the dominant metamorphism represented by the mineral assemblages in the surrounding rocks. The simplified graph in Table 1 shows the schematic relations of these eight groups relative to a single metamorphic episode. The faults are classified as "premetamorphic", "early metamorphic", "pre-peak metamorphic", "peak metamorphic", "post-peak metamorphic", "late metamorphic", or "postmetamorphic" depending on when they occurred relative to the peak of metamorphism when the highest grade assemblage was formed in the rocks. "Pre-peak", "peak", and "post-peak" metamorphic faults are considered to be "synmetamorphic". Compound faults are those where repeated movement occurs under different metamorphic conditions. The age of synmetamorphic and postmetamorphic faults can be determined by isotopic analysis of minerals with radioactive elements that formed during the history of the fault. For all the premetamorphic and early metamorphic faults these ages simply represent an upper limit - the faults are older than the calculated age. The criteria and concepts listed in Table 1 have been used with some success (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986; DelloRusso and Stanley, 1986) but are constantly undergoing revision as a result of current field work and petrographic analyses.

Let us return to the early question of the relative importance of folding and faulting. There is little doubt that folding has been an important component in the total shortening in the pre-Silurian hinterland. Minor folds of at least five generations can be seen in many parts of the area. Major isoclinal folds are outlined by many mappable units. The hinges of these folds plunge to the east and southeast nearly parallel to the dip of their associated axial surface schistosity. Their geometry therefore describes complex sheath folds whose shape is a product of severe flattening and profound elongation parallel to the long dimension of the sheath (Armstrong and Prewitt, in progress). It is this generation of folds (Fn) that accounts for most of the recorded fold - related shortening according to our present studies. The schistosity (Sn) that is parallel to the axial surface of these folds forms the dominant schistosity throughout the region. This generation of structures deforms and severely transposes an older generation of folds (Fn-1) and is, in turn, folded by several younger generations (Fn+1 and Fn+2) as it is traced westward across the axis of the Green Mountain anticlinorium (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986). A even younger fold generation (Fn+3) is found to the east along parts of the Northfield - Braintree Mountains.

The major part of the shortening across the pre-Silurian hinterland, however, is accounted for by the pre- or early metamorphic and synmetamorphic faults. Much of this shortening here appears to be associated with the older pre- or early metamorphic faults because they largely control the distribution of similar map units across the region (for example, the carbonaceous

schist of the Hazens Notch Formation or the Mt. Abraham Schist). It is these faults that are subsequently deformed into large sheath folds. The sheath folds are in turn cut by the more planar synmetamorphic faults which parallel the dominant S_n schistosity. The older F_{n-1} folds may have developed with the older faults in which case the faults would be "early" rather than "pre" - metamorphic.

Metamorphism and Age of Deformation The metamorphism in central Vermont has received considerable attention by Albee (1965, 1968), Laird and Albee (1981), Laird and others (1984). This work and new findings are summarized by Laird in this guidebook. Studies on the relation between structure and mineral growth are reported in O'Loughlin and Stanley (1986) and Lapp and Stanley (1986). The results of this work along the western side of the Green Mountains shows that there is a continuous overlapping relation between mineral growth and the structural sequence F_{n-1} through F_{n+2} . For example S_{n-1} is preserved in isolated hinges between S_n or as inclusion trails of graphite, white mica, quartz, and opaque in porphyroblasts of albite and garnet. Some of these trails can be traced into the S_n schistosity. All the peak metamorphic minerals such as chloritoid, garnet, kyanite, and plagioclase either are oriented parallel to or grow across the S_n schistosity. Chlorite occurs as fine and coarse grains. The fine - grained chlorite is oriented parallel to S_n . The coarser - grained chlorite developed from garnet and is found in pressure shadows that are oriented parallel to S_n . Prominent chlorite streaks, which also grew from garnet, are well developed on the dominant S_n schistosity in the Mt. Abraham Schist and the Pinney Hollow. Chlorite is also deformed by S_{n+1} , and is oriented parallel to S_{n+1} along the west side of the Green Mountains. The only minerals which clearly represent a separate period of growth are large muscovite and biotite grains that cut across S_{n+1} and S_{n+2} with random orientation in the Underhill and Hoosac Formations. These cross micas are not the ones that gives the 385 m.y. age at Mt. Grant. Those grains are associated with S_n and S_{n+1} .

The S_n schistosity on the west side of the Green Mountains can be traced eastward to Granville Gulf where it forms the dominant schistosity defined by chlorite, epidote, and actinolite in greenstones. Barroisitic hornblende surrounding actinolite from this locality gives $^{39}Ar/^{40}Ar$ ages of 448-471 m.y. (sample V145 of Laird and Albee, 1981, Laird and others, 1984). The barroisitic hornblendes may have formed during S_{n-1} although there is no evidence for this relation. Laird (this guidebook) states that these rocks have undergone an earlier medium-high pressure series metamorphism represented by the barroisitic hornblende. This was then followed by a lower grade greenschist facies metamorphism. Similar relations are preserved in the pelitic schists to the west where kyanite and garnet are retrograded to sericite and chlorite. At present the relation between the isotopically - dated minerals and the respective schistositities that have been recognized across transect are still not clear. Are the younger ages simply a result of cooling or were metamorphic reactions still going on west of the

Green Mountain axis until 385 m.y. ?

ACKNOWLEDGEMENTS

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THE LINCOLN MASSIF AND ITS IMMEDIATE COVER

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ITINERARY

The transect across central Vermont is divided into two days. The Saturday trip will begin along the western boundary of the eastern Lincoln massif and progress eastward to the crest of the Green Mountains where we will only see the Hoosac, Underhill, Mt. Abraham and Hazens Notch Formations. The remaining part of the pre-Silurian eastern Vermont sequence will be seen on Sunday when the trip will concentrate on the Hazens Notch, Pinney Hollow, Ottauquechee, and Stowe Formations in the valleys of the White and Mad Rivers. The introductory part of this paper entitled "A Transect through the Pre-Silurian rocks of Central Vermont" applies to both days. The figures are numbered sequentially as if the two days were a single trip. The "References Cited" also apply to both days.

Meeting place for Saturday trip - Lincoln General Store, Lincoln, Vermont. Time 8:30 A.M.

Geological maps, cross sections and texts for all the stops for Part I in the Lincoln massif and the eastern cover sequence are available through the Vermont Geological Survey. They are Tauvers (1982a), DiPietro (1983), Strehle and Stanley (1986), DelloRusso and Stanley (1986), O'Loughlin and Stanley (1986), and Lapp and Stanley (1986). The following stop descriptions, therefore, depend heavily on this information.

Stops 1 through 6 - The western boundary of the Eastern Lincoln massif (fig. 7; plate 2, DelloRusso and Stanley, 1986) (DelloRusso) - These stops show the basement cover contact on the western limb of the massif, a small, newly discovered syncline of cover within the massif, and Late Proterozoic mafic dikes. A discussion of the chemistry of these dikes will be presented by Ray Coish of Middlebury College. These stops will also demonstrate the difficulty in recognizing the precise boundary between the Middle Proterozoic basement and its immediate cover even in areas where the deformation is not intense.

Stop 1 - This is a 12 ft high exposure of coarse cobble conglomerate containing abundant, matrix-supported cobbles of quartzite and gneiss. The cobbles are derived from the nearby Mt. Holly Complex. The composition of the matrix is similar to the finer-grained metawacke found elsewhere in the basal part of the Pinnacle Formation. Sedimentary features such as bedding and

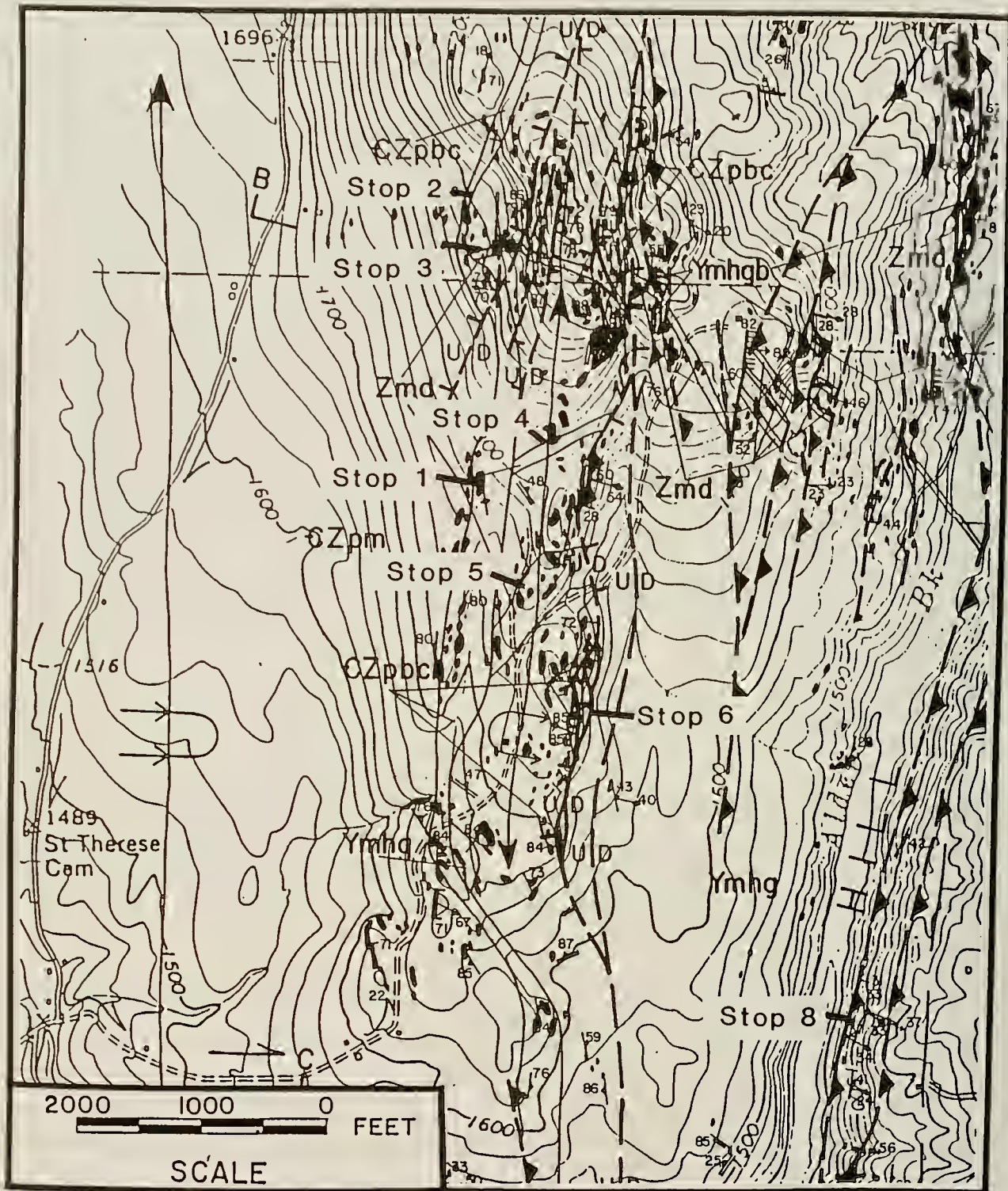


Figure 7

Geological map taken from DelloRusso and Stanley (1986, pl. 2) showing stop locations 1 through 6 for the western limb of the Eastern Lincoln massif. The long arrows parallel to the roads show the route. Lithic symbols are the following: Ymhg - granitic gneiss of the Mt. Holly Complex, Ymhgb - biotite schist zones in the granitic gneiss of the Mt. Holly Complex, Zmd - mafic dikes in the Mt. Holly Complex, CZpbc - basal conglomerate of the Pinnacle Formation, CZpm - schistose metawacke of the Pinnacle Formation. High angle faults are shown by thick dashed lines with "U" and "D" symbols. Thrust faults are shown by thick dashed lines with solid bars.

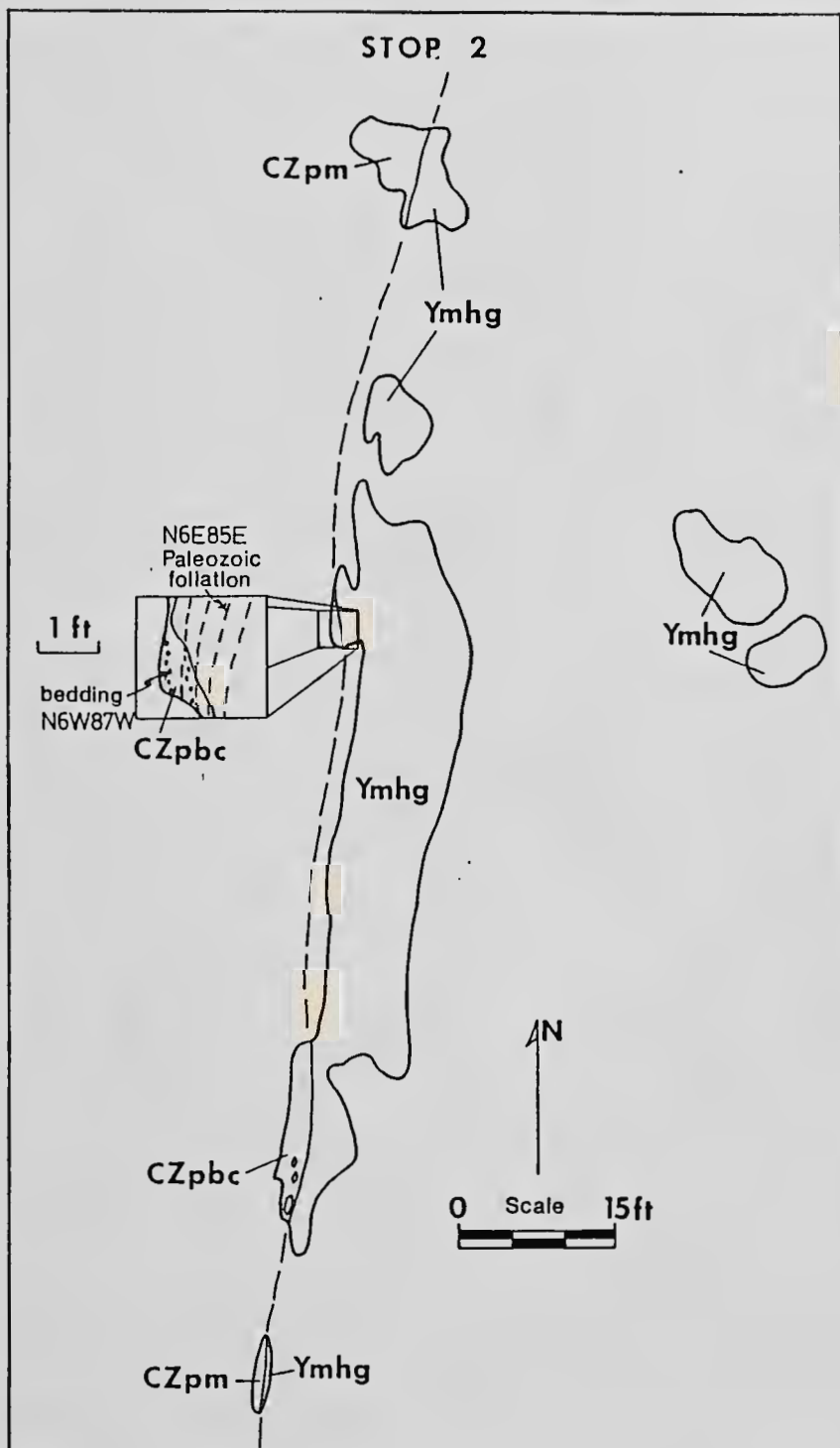


Figure 8

Detailed geological map of Stop 2. Lithic symbols are listed in the caption for figure 7. Geology by DelloRusso.

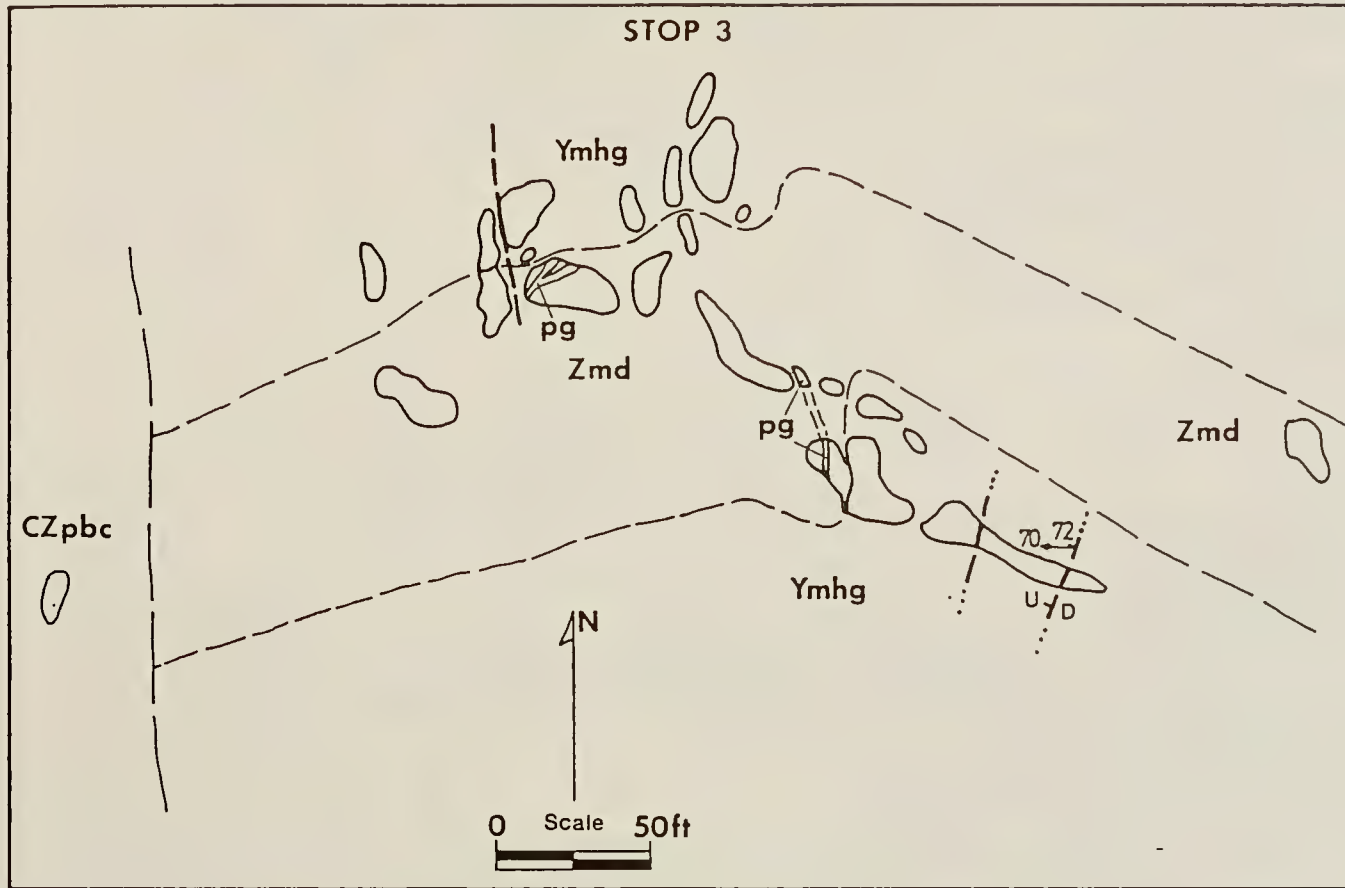


Figure 9

Detailed geological map of Stop 3. Lithic symbols are listed in the caption for figure 7. The symbol "pg" refers to pegmatite. Geology by DelloRusso.

channels are absent. Note that the cobbles show little or no deformation. There is only one Paleozoic foliation here.

Stop 2 - (fig. 8) - Exposure of the basal unconformity. This stop illustrates the difficulty in recognizing the precise unconformity. As a guide try first to recognize the boulders and cobbles. After you have done this successfully, then locate the unconformity. Note the similarity between the metawacke and the deformed gneiss directly beneath the basal conglomerate. The attitude of bedding and the respective foliations are shown on the small inset map on figure 8. Note that the Paleozoic foliation is only well developed in the gneiss near the unconformity. This feature suggests that the gneiss was weathered and hence weakened before deposition of the cover. Magnetite is common in the metawacke. The gneiss is the typical light-gray to white weathering, massive, medium-grained muscovite - perthite - microcline - quartz - plagioclase gneiss that underlies much of the Eastern Lincoln massif.

Stop 3 - (fig. 9) - This is an exposure of a Proterozoic mafic dike that cuts across the Grenvillian foliation in the gneiss. The dike can be traced to the west where it is truncated by the basal conglomerate. The dike contact is exposed in two places in the outcrop. The dike rock is a dark-green to brown weathering, fine- to medium-grained schist containing sericite, epidote, biotite, chlorite, and plagioclase with minor carbonate, opaques, and sphene. Along the contacts the schist is rich in biotite. Large, relict, euhedral white plagioclase phenocrysts are common.

Stop 4 - Walk south from Stop 3. This is another exposure of a Late Precambrian dike. Note that the dike cuts the Grenvillian foliation in the surrounding gneiss. Xenoliths of gneiss are present. Geochemical data indicate that these dikes were originally high-titanium, high phosphorous, alkaline basalts (Filosof, 1986) that are similar in composition to the Tibbit Hill Volcanics of the Pinnacle Formation (Coish and others, 1986). Thus dikes such as these could be feeder dikes to the overlying mafic volcanic rocks in the overlying rift clastic rocks.

Stop 5 - Walk south from Stop 4. This is another Late Proterozoic dike. It contains gneiss xenoliths. Note the contact. The Paleozoic foliation is only poorly developed at best.

Stop 6 - This series of outcrops is in a small, parasitic syncline on the western limb of the massif. High-angle faults developed during the formation of the syncline and cut the basement as well as the overlying cover. As a result it is often difficult to separate deformed basement gneiss along the faults from the overlying basal cover. This is further complicated by the fact that the gneiss directly beneath the cover appears to have been weathered before the basal conglomerate was deposited. The presence of gneiss boulders and quartz cobbles in several places, however, helps delineate the unconformity and resolve some of these problems. In one exposure a pegmatite in the gneiss is truncated by the unconformity. Along the faults, the gneiss and overlying

Pinnacle metawacke are deformed into a tan-colored schist that contains quartz, feldspar, and sericite.

Stop 7 - "Crash Bridge" - (fig. 10 and 11; DelloRusso and Stanley, 1986) (Stanley) - This outcrop is located along the New Haven River at the second bridge to the east of the general store in Lincoln. Here the Middle Proterozoic granitic gneiss and amphibolite are truncated by the basal conglomerate of the Pinnacle Formation which is strongly overturned to the east. Unlike stop 1 the contact here is a thrust fault marked by sheared biotite schist and muscovite schist with gneiss fragments (fig. 10). Again these rocks are difficult to distinguish from some of the overlying metawackes because muscovite schist looks much like the overlying metawacke. In thin section, however, the two differ in the degree of grain orientation and the percentage of biotite. The muscovite schist has no biotite and the muscovite and quartz are strongly oriented compared to the metawacke where the grains are more random and biotite is fairly abundant. It is therefore not surprising that the contact between the fault zone and the basal conglomerate is indistinct.

The basal conglomerate at "Crash Bridge" is quite different than the conglomerate at Stop 1 which is more typical of the basal unit. Here the conglomerate is largely made up of clast-supported cobbles and boulders of granitic gneiss that are arranged in lensoid bodies separated by magnetite-bearing metawacke. Although these rocks are tightly folded and cut by a pervasive Paleozoic schistosity, their shapes and mutual relations suggest that they are paleochannels in which the basal part of the channel faces toward the Middle Proterozoic rocks beneath and east of the bridge. Tauvers (1982a) who was the first to recognize the nature of this deposit suggested that the clast-supported channels formed during periods of very rapid flow whereas the matrix formed during period of reduce flow. One such environment where deposits such as this can develop is on an alluvial fan in arid to semi-arid climates (Tauvers, 1982b). A submarine fan, however, might be equally plausible.

As shown on the geological map and profile section the layering in the Pinnacle Formation has been deformed into a series of reclined folds that become tighter as the fault contact is approached. These folds and the associated axial surface schistosity are coeval with gently-plunging folds along the unfaulted, western boundary of the massif. Their counterclockwise asymmetry indicates that they developed on the western, overturned limb of the Eastern Lincoln massif. Furthermore, we suggest that these folds have been rotated from their original low plunge to a steeper plunge as the thrust fault was developed on the overturned limb of the massif. The transport lineation on thrust faults throughout the massif plunges to the east or slightly south of east (DelloRusso and Stanley, 1986, pl. 4).

Deformation and metamorphism at "Crash Bridge" is Taconian as suggested from isotopic age analysis at this and nearby localities. A biotite K/Ar age of 410 m.y. was obtained from a biotite schist

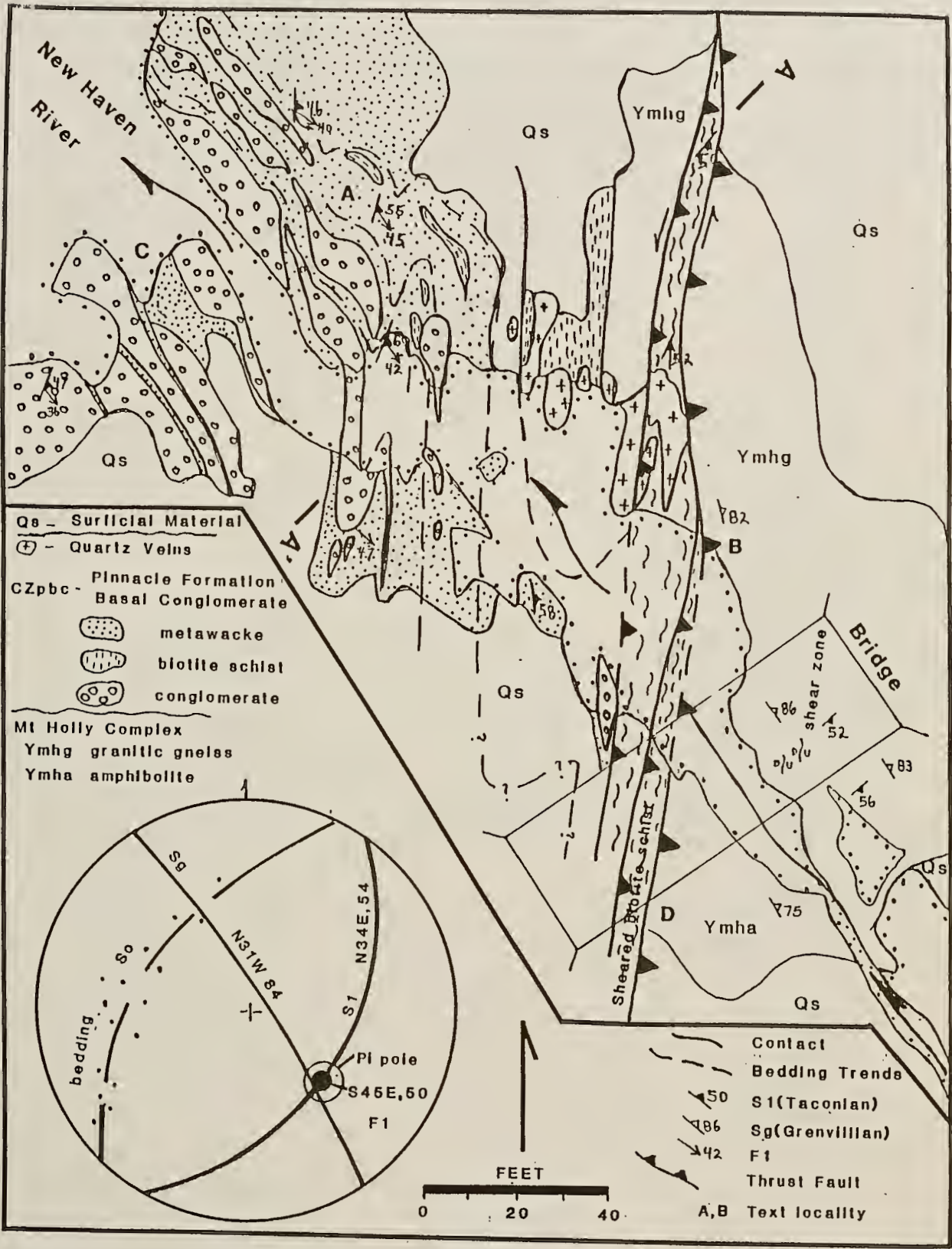
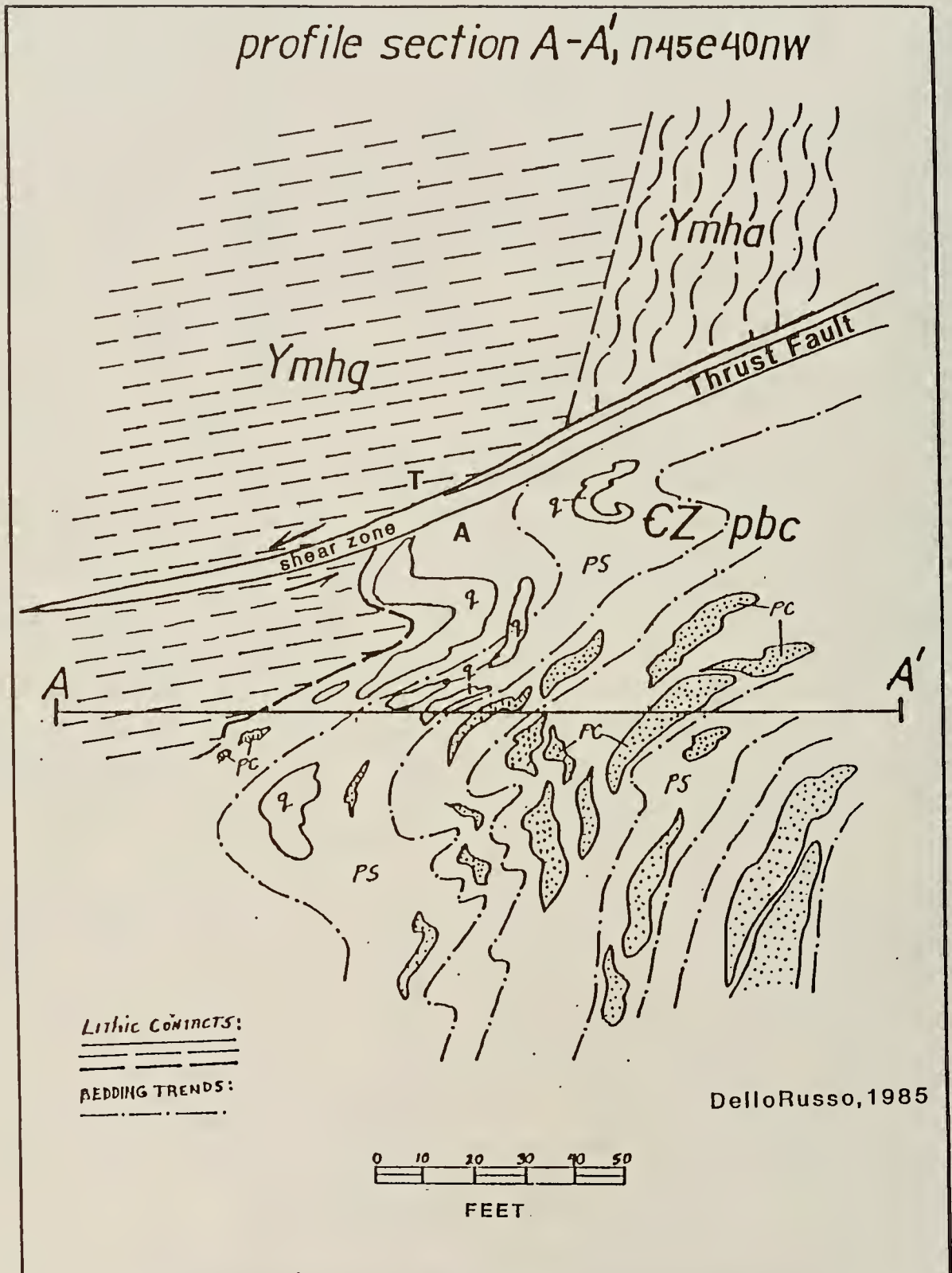


Figure 10 **BASAL LATE PROTEROZOIC UNCONFORMITY AT LINCOLN, VERMONT**

Tauvers, 1982 Stanley and DelloRusso 1985



FAULTED UNCONFORMITY AT LINCOLN, VERMONT

Figure 11

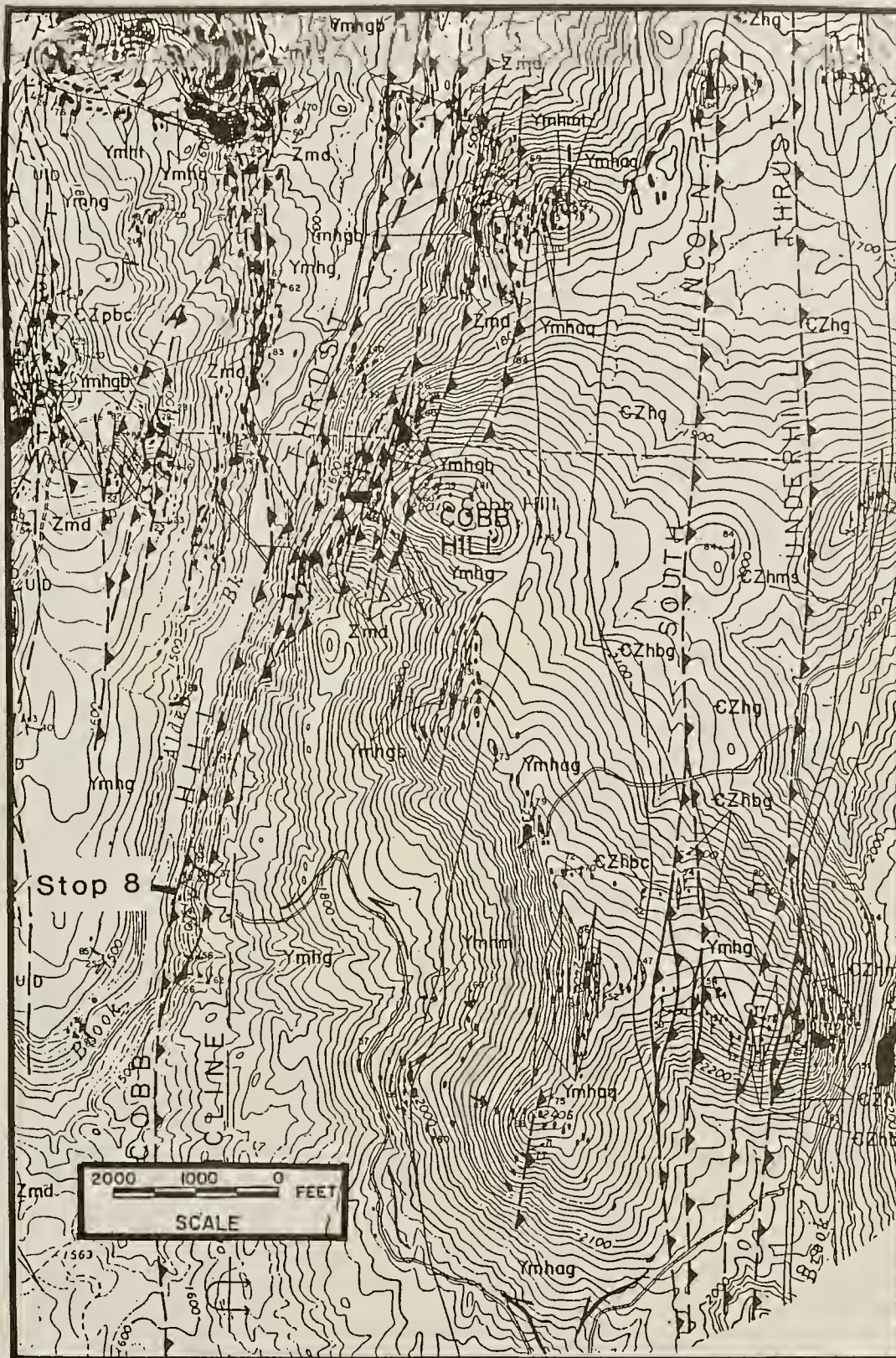


Figure 12

Geological map taken from DelloRusso and Stanley (1986, pl. 2) showing the location of Stop 8. Lithic symbols are listed in the captions for figures 3 and 7. Other symbols can be found on plate 2 of DelloRusso and Stanley (1986). Long arrows parallel to the roads show the route to Stop 8. This is private property and you must ask for permission to visit these outcrops.

at this locality (Cady, 1969). Directly to the west in the Pinnacle Formation other isotopic age determinations reported by Cady (1969) were recalculated by Sutter and others (1985) and fall within the range 370 to 397 m.y. with about a 20 m.y. error. They interrupt these data to represent cooling ages from Taconian metamorphism (about 465 m.y).

Stop 8 - Cobb Hill Thrust Zone (fig. 12; DelloRusso and Stanley, 1986, pl. 2 and 4). (DelloRusso and Stanley) - Drive south from "Crash Bridge" through South Lincoln (about 4 or 5 houses) along a dirt road for about 7 miles. Turn northwest on a Forest Service road. Park at the house at the very end of this road. The locality is located on the east side of Alder Brook and is difficult to find.

The Cobb Hill thrust zone is the major thrust zone in the Eastern Lincoln massif. A branch of this thrust passes through the "Crash Bridge" locality. Alder Brook, which is a major topographic lineament in the Lincoln massif is largely controlled by this fault zone.

NO HAMMERS PLEASE - YOU MAY COLLECT LOOSE MATERIAL ONLY.

The series of outcrops illustrates the development of a mylonitic fault zone in granitic gneiss. A description on the various phases and the supplemental discussion are published in DelloRusso (1986) and DelloRusso and Stanley (1986, pl. 4 specifically). Study the outcrop beginning with the southernmost outcrop. The evolution of the fault zone fabric is divided into 4 phases.

Phase 1 - Fragmentation of the granitic gneiss - In this outcrop we find pebble- to cobble-size clasts of coarse-grained gneiss in a finer grained, foliated matrix of recrystallized quartz, feldspar, and sericite that is compositional similar to the gneiss. The sericite forms from the alteration of feldspar with the addition of water and aluminum oxide. The only foliation present in the gneiss clasts is the coarse mineral layering of Grenvillian age. Although there is no evidence of cataclasis in the matrix, it is likely that fracturing may have been important in the very early stages of deformation.

Phase 2 - Pervasive mylonitic schistosity - Moving a bit to the north we find that a well developed mylonitic layering or schistosity is present in the matrix. The number and size of the gneiss clasts is greatly reduced and the grain size in the matrix is finer. The foliation is well developed and quartz grains and clusters are elongated into a pronounced lineation. We suggest that recrystallization and recovery has continued to reduce the grain size and eliminate the original gneiss clasts.

Phase 3 - Folded mylonitic schistosity - Moving still farther north we find that the mylonitic schistosity is now folded into tight, reclined folds. These folds are asymmetrical with a clockwise or north-over-south asymmetry. Like the folds at "Crash

Bridge" these folds plunge slightly south (S65E, 35) of the mineral lineation (S71E, 30). We suggest that the mylonitic fabric at this stage has become unstable because the grain size has been reduced to such a size that it is easier to deform by folding than by further grain-size reduction. In essence, the mylonite has strain hardened.

Phase 4 - Shear bands and fragmented mylonite - Moving still farther north we find that shear bands are present on the attenuated short limbs of the reclined folds. The shear bands are very planar, thin, micaceous layers that clearly truncate the older mylonitic foliation at a low angle. Continued deformation and shear movement along the bands offsets and rotates this older foliation and results in a new "breccia" consisting of mylonitic fragments. The development of sericite along the shear bands indicates that retrogression of feldspar is an important softening mechanism during this phase of deformation. The absence of recognizable deformation of the shear bands indicates that they represent the youngest phase in the development of the fault zone.

Continue to the north for 50 ft. Here you will find an outcrop with many of the features that we have already described. Here, however, several large quartz veins intrude the zone parallel to the layering. Near the quartz veins are sericite-rich schists or phyllonites. We have seen the occurrence of sericite at a number of stages in the evolution of the fault zone. Here they are formed on a larger scale and represent the final product of alteration along the fault zone where silica-rich solutions invade the zone. It is likely that much of the strain along the Cobb Hill thrust zone is concentrated in these sericite-rich zones.

This fault is a postmetamorphic thrust zone in that it overprints the older Grenvillian metamorphism. We estimate that it formed under "ductile" conditions with temperatures in the order of 400°C to 450°C and pressures of 4 or 5 kbars (10-15 km), although the indicators are certainly not precise.

Stop 9 South Lincoln Bridge (fig.13) (Stanley) - Return north to the "village" of South Lincoln. Try not to block traffic across the bridge. This outcrop, which consists of mylonitic gneiss and basal conglomerate, was originally mapped by Cady and others (1962) as the contact of the eastern cover sequence and the Middle Proterozoic rocks. The presence of chlorite schist, biotite schist, and garnet schist identical in composition and aspect to the overlying Hoosac Formation to the west and structurally below the outcrop of gneiss and conglomerate indicates, however, that a major thrust fault occurs along the lower contact of the gneiss. This relation was first recognized by Tauvers (1982a, 1982b) who considered it to be part of the Underhill thrust zone. DelloRusso and Stanley (1986) have recognized similar relations along the eastern border of the Lincoln massif. This particular fault they named the South Lincoln thrust. Petrofabric analysis has shown that the gneiss is mylonitic with distinct layers of quartz and feldspar (Strehle and Stanley, 1986). A very prominent mineral

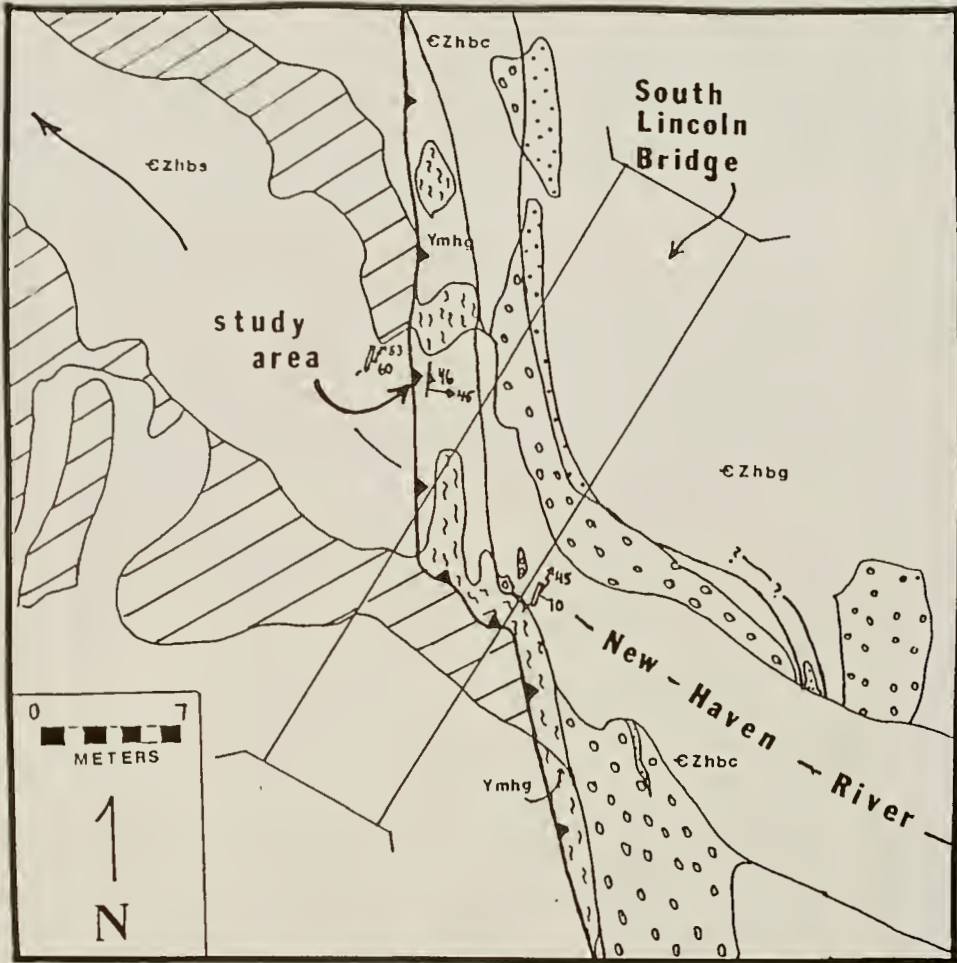
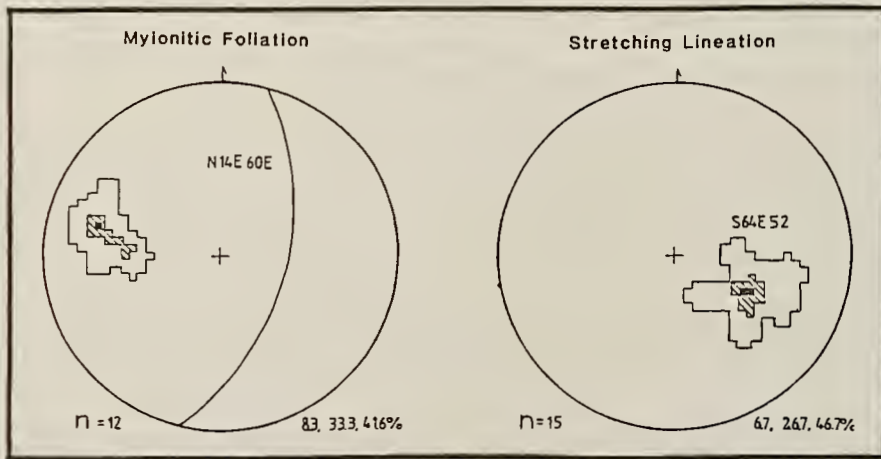


Figure 13

At the bridge at South Lincoln, the mylonitic gneiss of the Mt. Holly Complex (Ymhg) is in thrust contact with the Hoosac Formation (CZhbs). A pervasive, platy mylonitic foliation (Sm) and penetrative mineral lineation dominates the gneiss (represented by symbols in gneiss sliver, Ymhg, and shown by the contoured data in Figure 33). The fault is deformed by younger folds (F2) represented by the symbols to the east and west of the sliver. In Jerusalem, the fault-related deformation along the Underhill thrust fault (Utf) is distributed over a much wider zone and numerous tectonic slivers are juxtaposed in the fault zone (fig. 45, Plate 4). Geology mapped by Tavers (1982) and modified by Stanley.



Lower hemisphere equal area projections of n poles to the mylonitic foliation and n stretching lineations. Tick marks north. Contours are given as the percentage of points per 1° area. The stretching lineation is defined by elongation of quartz and feldspar grains.

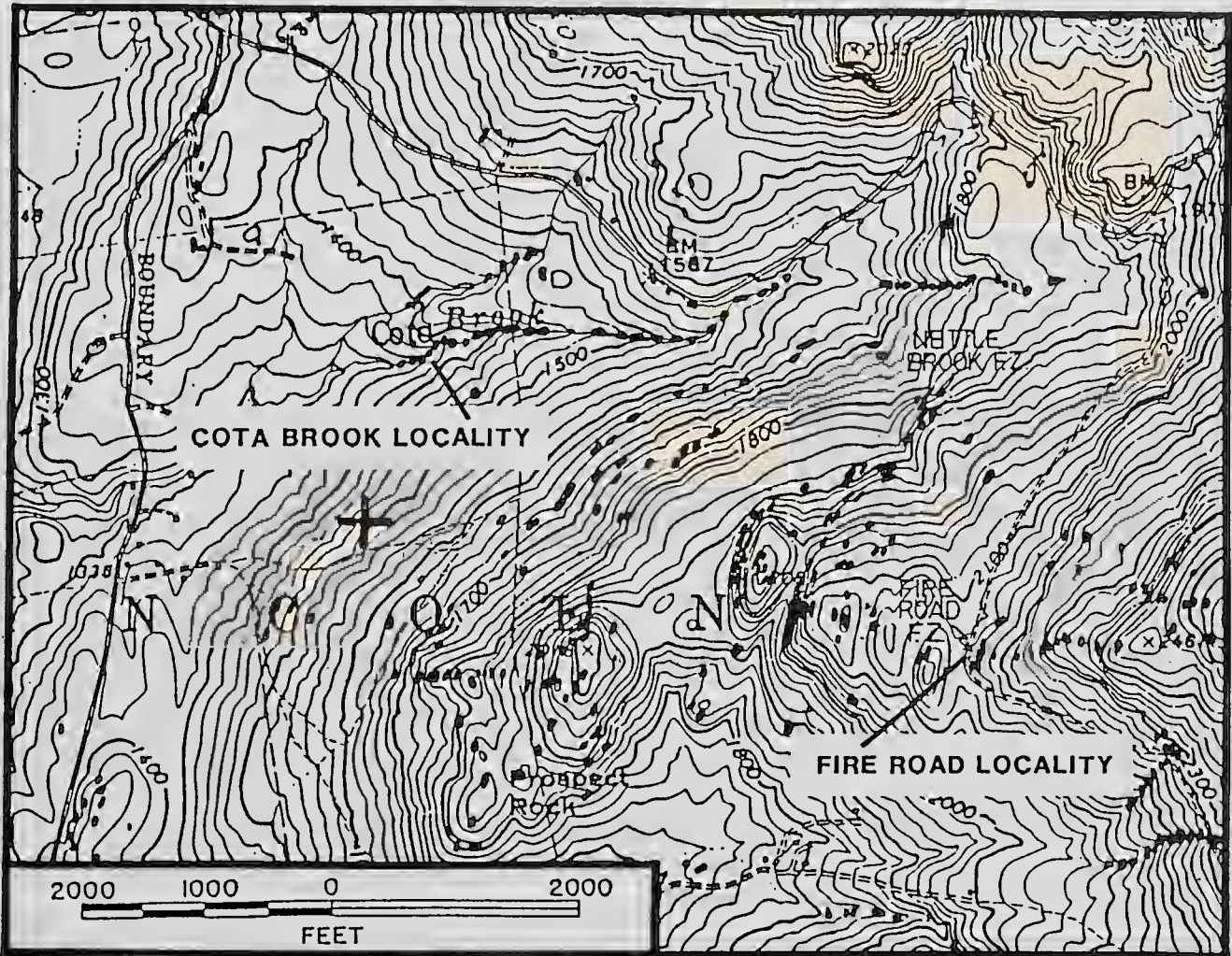


Figure 14

Location map for the Cota Brook Sequence and the Fire Road Fault Zone. Topographic base is from the Lincoln 7.5 minute topographic quadrangle. Elevations are in feet above sea level. Cross mark located below the label for Cota Brook locality is at longitude 72.57.30 west and latitude 44.05.00 north. The road in the northern (upper) part of the map is the Lincoln Gap Road. Lincoln Gap is off the map to the east. A part of Gerry Road is seen along the western boundary of the map. Black spots represent observed outcrop.

elongation plunges to the southeast (S64E, 52) in the plane of the mylonitic foliation. The orientation of recrystallized grains and the quartz C axis fabrics indicate east-over-west movement. The dominant deformation mechanism is dislocation creep resulting in the recrystallization of both quartz and feldspar. Based on these fabrics and the presence of garnet in some of the surrounding rocks Strehle and Stanley (1986) estimate that the fault zone developed under ductile conditions comparable, but perhaps somewhat deeper than the Cobb Hill thrust zone. Again, Taconian deformation is inferred based on available isotopic age information and the similarity in orientation of fabrics across the massif. We will see that this same fabric orientation continues to the east into the pre-Silurian cover.

Stop 10 - Cota Brook Sequence (figs 14 and 15) (O'Loughlin and Lapp) - The Cota Brook sequence is a stratigraphically coherent section of metawackes and mafic schists of the Hoosac (Pinnacle) Formation. Outcrop is exposed in the stream valley and banks of Cota Brook at elevations of 1350 to 1390 feet above sea level. The Cota Brook sequence is bounded both upstream and downstream by zones of fissile schists with abundant graphite pods and layers. These zones are interpreted to be "pre-peak" or "syn-peak" metamorphic faults (fig. 15). Rock units surrounding these fault zones are lithologies in the Underhill Formation. Faults do not appear to be present within the sequence. A tight to isoclinal fold within the sequence is defined by mafic schists and changes in the dip of the dominant schistosity. The geochemical data and amphibole petrology will be discussed by Coish and Laird (refer to their respective papers in this guidebook). Lithic descriptions and estimated modes are given in Table 2.

Stop 11 - Fire Road Fault Zone (figs 14 and 16) (Lapp and Stanley)

This group of outcrops shows the typically complex contact relationships of the area and is easily accessible from a forest service road 2 km southwest of Lincoln Gap (see location map). This area is distinctly marked by a 20 m x 1-2 m milky quartz vein (see outcrop sketch). Black carbonaceous schist of the Hazens Notch Formation (EZhnc) occurs along the lower (western) part of this outcrop, while chloritoid-white mica Mt. Abraham Schist (E2a) occurs above. Outcrops of the latter unit are, here, anomalously rusty weathering and contain flattened garnets with chlorite pressure shadows and kyanite pseudomorphs composed of white-mica.

Contacts between the two schists are extremely sharp and highly "interfingered". One exposure shows folds in the rusty schist truncated by the carbonaceous schist along a pre- to syn-metamorphic fault. This contact was later folded and intruded by the large quartz vein. Reactivated faulting further displaced the schists to the west, over the quartz vein along a corrugated surface. The east-west trend of the hinge line of the corrugations is parallel to quartz lineations on the carbonaceous schist (EZhnc), which is also found as lineated slivers within the quartz vein.

Table 2

Cota Brook Sequence

METAWACKES

In general, the metawackes are light tan to brownish in color and are less resistant to stream erosion than the mafic schists. Four lithic types have been distinguished.

- Im laminated metawacke
well laminated or foliated, quartz rich, little plagioclase, white mica rich, with "quartzite" beds and layers of coarse biotite porphyroblasts.
- m mica poor metawacke
quartz and feldspar rich, with coarse biotite porphyroblasts, less mica than other lithic types.
- q massive "quartzite"
primarily quartz and feldspar, little or no mica.
- gm grey green laminated metawacke
well laminated or foliated, grey green in color, with coarse biotite porphyroblasts.

Estimated Modes of Metawackes

Minerals	Location		216	219	219Q	
	Sample	a	b			
Quartz		15	30	30	45	Modes are mineral percentages. tr = trace percentage observed dash (-) = mineral not observed
White Mica		60	55	10	-	
Chlorite		7	5	tr	1	
Plagioclase		4	2	45	45	
Garnet		1	1	3	tr	
Biotite		2	-	7	2	
Graphite		-	5	-	-	
Opaques		1	2	1	tr	
Carbonate		-	-	-	tr	
Epidote		-	-	-	tr	

MAFIC SCHISTS

In general, the mafic schists are green to dark green in color and are more resistant to stream erosion than the metawackes. Geochemical work done on some of these mafic schists (Gavigan 1986) has identified them as Type A metabasalts (Coish and others 1985) which were formed during the initial stages of continental rifting prior to the formation of Iapetus (proto-Atlantic). The exposures of mafic schist are each relatively distinct in appearance; important characteristics are noted below by location (Loc.) number.

- Loc. 216Z fine grained.
- Loc. 217 coarse to very coarse grained, amphibole rich layer at upper contact.
- Loc. 218A well laminated or foliated. Similar to mafic schist at Loc. 221.
- Loc. 218B coarse grained, massive, grain size fines toward lower contact.
- Loc. 220 coarse grained, with a 2-3 inch thick "quartzite" bed, very amphibole rich.
- Loc. 221 very coarse grained, well laminated or foliated, very amphibole rich, with small pods of carbonate(?) material. Similar to mafic schist at Loc. 218A.
- Loc. 222 "lumpy" texture, with small pods of carbonate(?) material.

Estimated Modes of Mafic Schists

Mineral	Location		217	217	218A	218A	218A	218A	220
	Sample	a	b	a	b	c	d		
Quartz		40	30	10	15	27	27	20	
Plagioclase		3	-	10	35	1	28	5	
Epidote		5	10	2	1	5	3	-	
Garnet		-	-	tr	tr	-	tr	-	
Amphibole		-	-	60	2	60	25	70	
Chlorite		40	35	10	15	3	10	tr	
Biotite		-	-	tr	-	-	-	-	
Carbonate		10	25	5	30	3	2	-	
Opaques		2	tr	3	tr	1	1	3	

Modes are mineral percentages. tr = trace percentage observed
dash (-) = mineral not observed

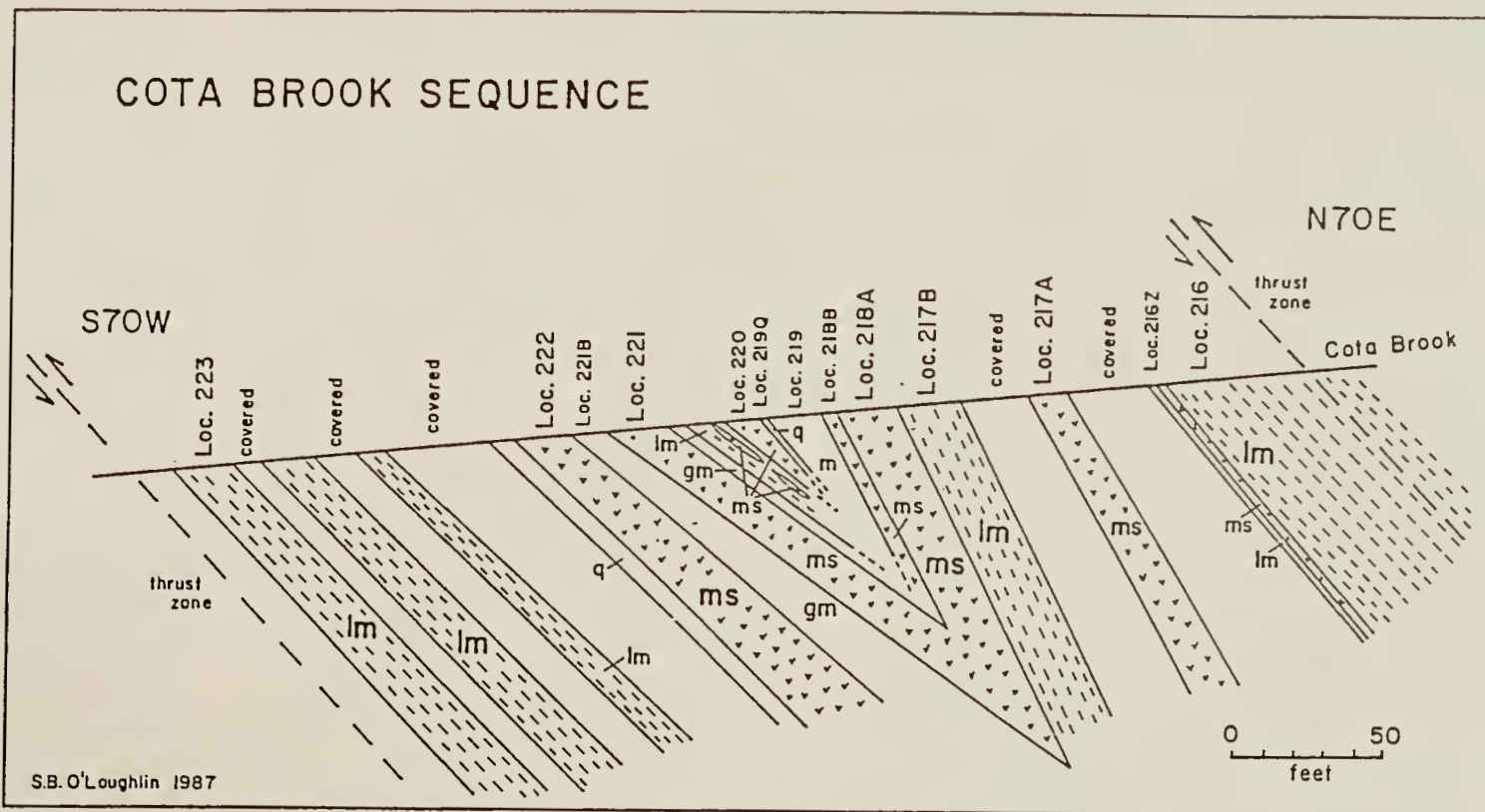


Figure 15

Cross section of the Cota Brook sequence looking north. The sequence consists of stratigraphically coherent metawackes and mafic schists of the Hoosac (Pinnacle) Formation which are bounded by thrust zones (pre-peak metamorphism, locations approximate) and surrounded by lithologies of the Underhill Formation (not shown). Contacts, where exposed, are sharp, generally parallel the foliation, and may be isoclinally folded. The synform (Fn+1 age) shown in the sequence is based upon changes in the orientation of the dominant schistosity (Sn). The geometry of the bounding fault zones is unclear. Three possibilities for this geometry are:

1) Both the eastern and western thrust zones are structurally upright and merge above and below the Cota Brook sequence to form a sliver of the Hoosac in the Underhill. 2) The western thrust zone is overturned and continuous with the eastern thrust zone. The thrust

zone is isoclinally folded (Fn age) and closes above the Cota Brook sequence at the hinge of an antiform. An erosional window through the Underhill exposes the Hoosac. 3) The eastern thrust zone is overturned and continuous with the western thrust zone. The thrust zone is isoclinally folded (Fn age) and closes below the Cota Brook sequence at the hinge of a synform. A klippe of the Hoosac rests on the Underhill. In all three variations, the thrust movement sense is east over west for both the eastern and western thrust zones. Elevation of location (Loc.) 216 is 1390 feet above sea level; Loc. 223 is at 1350 feet. No vertical exaggeration. Lithic designators: lm - laminated metawacke, m - mica poor metawacke, q - massive "quartzite", gm - grey green metawacke, ms - mafic schist. For more information, see text.

THE FIRE ROAD FAULT ZONE

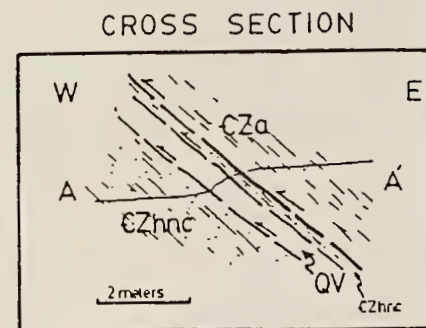
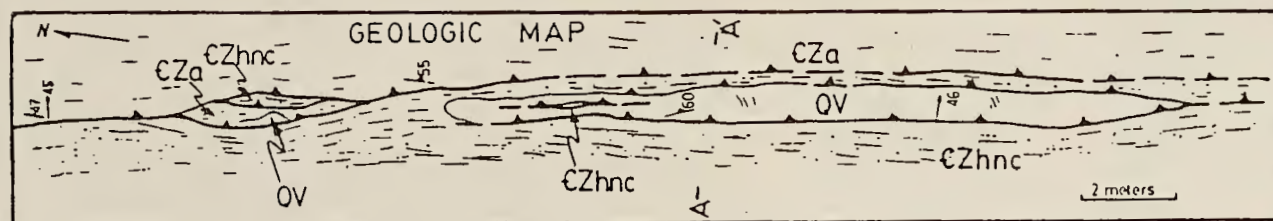
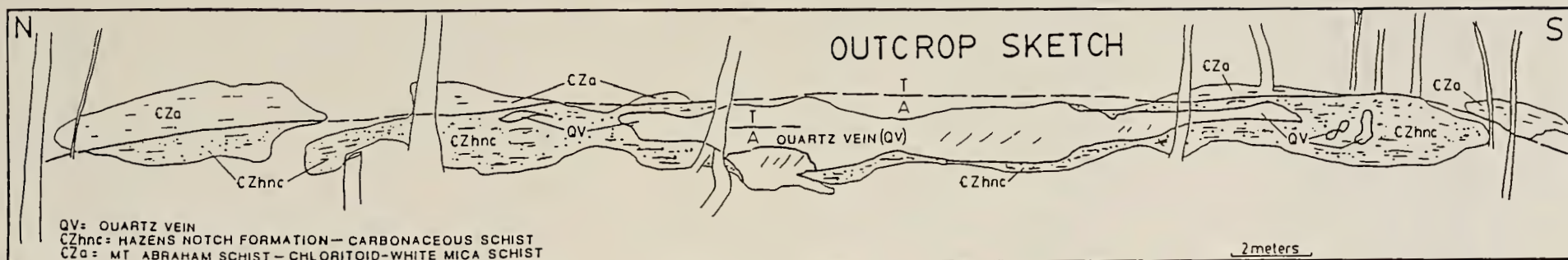


Figure 16

Outcrop sketch, geologic map and cross section of the Fire Road Fault Zone. Longitudinal profile view in outcrop sketch is looking east. Long, heavy line represents an interpreted syn- to post-metamorphic fault contact between black carbonaceous schist of Hazens Notch Formation (CZhnc) and chloritoid-white mica Mt. Abraham Schist (CZa) (at this location, rusty weathering). T = toward; A = away. The large, milky quartz vein postdates the dominant folding but is cut by discontinuous post-metamorphic faults which bear slivers of the carbonaceous schist. The moderately east-dipping, dominant schistosity and east-west trending mineral lineations are symbolized on the geologic map. Cross section shows interpretation of east over west sense of motion for this reactivated fault zone.

These observations suggest a history of repeated deformation and metamorphism at this fault zone. Early faulting of the carbonaceous and rusty schists was followed by folding and truncation along faults during kyanite-grade metamorphism. The dominant folds were intruded by the quartz vein, which was displaced over the schists and cut by late (post-metamorphic) faults. A similar fault zone between the carbonaceous schist and the Underhill Formation occurs in an area called the Nettle Brook Fault Zone (shown on the location map).

THE PRE-SILURIAN HINTERLAND ALONG THE VALLEYS OF THE WHITE AND MAD
RIVERS, CENTRAL VERMONT

By

Rolfe S. Stanley, Thomas Armstrong, Jerome Kraus, Gregory Walsh,
Jeffery Prewitt, Christine Kimball, and Athene Cua
University of Vermont
Burlington, Vermont
05401

ITINERARY

Meeting Place: The General Store in Warren Village just east of the junction of Route 100 and the Lincoln Gap Road. Warren can be reached from Northfield by traveling south on Route 12a to Roxbury. Turn west on the Roxbury Gap Road and follow the signs on the west side of the Northfield Mountain to Warren. For those that happen to arrive late, the field trip will be along Route 100 for most of the day. We will begin to the south near Hancock and gradually work our way northward to Waitsfield.

Departure time 8:30 AM.

Please read the introductory part of this paper entitled "A Transect through the Pre-Silurian rocks of Central Vermont".

GEOLOGY OF THE GRANVILLE-HANCOCK AREA, CENTRAL VERMONT

by

Thomas R. Armstrong

The bedrock geology of the Granville-Hancock area was first mapped by Osberg (1952) at the scale of 1:62,500. The contacts between the major units were interpreted as depositional due to the intercalation of lithologies along contact zones, and the lack of sharp boundaries between many formations. Osberg interpreted the eastwardly dipping stratigraphy as a coherent depositinal sequence younging to the east, consisting of the Monastery, Granville, Pinney Hollow, Ottauquechee, and Stowe Formations. Several of these formations contain minor lithologies which are similar or identical across strike. For example, carbonaceous schist which is shown on the Vermont state map as a member of the Pinney Hollow Formation, was mapped by Osberg as the Ottauquechee Formation. The present detailed mapping of this region has clearly demonstrated the similarity of this unit to the black schist of the Ottauquechee Formation and is thus in agreement with Osberg's interpretation.

Based on the distribution of the Ottauquechee and Stowe Formations, and the greenstones within the Pinney Hollow Formation, Osberg interpreted the structure as consisting of a series of west-verging, overturned folds. These folds were believed to be responsible for deforming the stratigraphy into the map pattern dominated by large scale folds.

New mapping at the scale of 1:12,000 has demonstrated the tectonic nature of contacts between the Mt. Abraham, Hazens Notch, Pinney Hollow, Ottauquechee, and Stowe Formations. Many of the gradational contacts mapped by Osberg are now interpreted as early-metamorphic thrust faults (Table 1). These thrusts always appear to juxtapose discreet lithologies of different formations. Fault zone fabrics are absent due to annealing during taconian peak metamorphism. The tectonic nature of these surfaces, however, can be demonstrated by the lithic truncation of various units along these contacts. The greenstones (The Hancock and Brackett Members of Osberg) are shown truncated along early-metamorphic contacts in numerous localities. Without detailed mapping of this geometry, it would be impossible to distinguish tectonic contacts from depositional boundaries.

The silvery-green quartz-sericite-chlorite-albite schist and albite-epidote-chlorite-calcite greenstone of the Pinney Hollow Formation are demonstrably overlain by the carbonaceous albitic schists of the Hazens Notch Formation (more correctly called "Granville") which in turn, are overlain by the graphitic black phyllite, black quartzite, and quartzose schist of the Ottauquechee Formation. Both of these contacts are early-metamorphic thrusts, shown by the map scale truncation of upper and lower plate lithologies. In some instances, the Ottauquechee thrust slice lies directly above the structurally lowest Pinney Hollow Formation, with the Hazens Notch thrust slice absent. This relationship suggests that after emplacement the carbonaceous schist of the Hazens Notch (Granville) and Pinney Hollow slices were deformed prior to the emplacement of the Ottauquechee slice. This geometry occurs in both the eastern and western parts of this area, and thus rules out a simple stratigraphic climb of the Ottauquechee slice along a ramp through the Hazens Notch and Pinney Hollow slices. Serpentinite bodies, found within the carbonaceous Hazens Notch and Ottauquechee Formations, were emplaced either during this phase of thrusting or an earlier phase associated with the development of an accretionary wedge (Stanley and Ratcliffe, 1985). Many of these bodies were also incorporated in subsequent taconian syn-metamorphic thrusts.

Following the development of the early-metamorphic thrusts and synchronous with Taconian metamorphism was the formation of the first phase of syn-metamorphic folds (F_{n-1}) and associated axial plane schistosity (S_{n-1}). This phase is preserved exclusively as relict fold hinges and schistosity, transposed and sheared along a second syn-metamorphic schistosity (S_n). Infrequently, F_{n-1} are exposed in structures refolded by the second phase of syn-metamorphic folds (F_n) and associated axial plane schistosity (S_n) (see Stop 17). The transposition and refolding of F_{n-1} and S_{n-1} is frequently seen at the microscopic scale.

The second phase syn-metamorphic schistosity (S_n) is the dominant fabric within this region, with an average trend of N05E, 65SE. This schistosity is axial planar to the second phase of syn-metamorphic folds (F_n). The hinges of F_n within the schists are oriented parallel, and sub parallel, to the dominant mineral lineation (L_n) which rakes steeply down the plane of S_n (about 85 from both the south and north strike lines). Some hinges are shallower, and at a considerable angle to the mineral lineation. The arcuate nature of these F_n hinges suggests that F_n are sheath folds, although only a few demonstrable closures can be seen in the field. Hinge orientations within the more

competent greenstones and quartzites are consistently at high angles to the mineral lineation. F_n folds developed within these units were more resistant to rotation during this high shear strain event, and thus never formed into sheath configuration, but rather into inclined, isoclinal folds (Armstrong, 1987; Kraus, 1987). This phase of deformation is primarily responsible for the map pattern geometry which shows the early-metamorphic thrusts being deformed into isolated, infolded klippe. This interpretation is very similar to the infold model of Osberg (1952) except that the contacts are now interpreted as faults.

The first phase of syn-metamorphic thrust faults (SF_n) developed coaxial to S_n and sheared out both F_{n-1} and F_n structures, producing a pervasive mylonitic fabric. Kinematic indicators, such as asymmetric porphyroclasts, pressure shadows, rotated porphyroblasts, and C-S fabric consistently display an east-over-west sense of motion. These faults displace the sheath fold klippe, juxtaposing both similar and different lithologies. The magnitude of displacement along SF_n was much less than that along the early-metamorphic faults which always juxtapose different lithologies. SF_n faults are further demonstrated by map scale lithic truncations, many of which include the greenstones within the Pinney Hollow and Stowe Formations. Many of these faults also contain tectonic slivers of rock-types foreign to either of the adjacent plates.

A third phase of syn-metamorphic deformation occurred coaxial to the second phase, and includes east-over-west asymmetric folds, a local axial plane schistosity, and several coplanar thrust faults. A progression in the development of this phase of schistosity (S_{n+1}) can be recognized where the associated asymmetric folds (F_{n+1}) are pervasive. In these locations S_{n+1} is the dominant fabric, shearing out F_n folds and folding S_n into F_{n+1} . In a few locations, S_{n+1} strain produces a second phase of syn-metamorphic faults (SF_{n+1}). Kinematic indicators, identical to those found along SF_n , consistently show east-over-west motion along SF_{n+1} .

This phase of deformation is considered a continuum of the second phase, to which it is coplanar. The development of F_{n+1} , (and possibly F_n) are most likely related to strain hardening during the development of S_n and S_{n+1} . SF_n and SF_{n+1} may represent periods of strain softening, the result of mineralogical alteration by a fluid phase during contemporaneous retrograde metamorphism.

Late stage, late-metamorphic upright, open folds (F_{n+2}) deform the syn-metamorphic fabrics and may be related to either late Taconian or Acadian deformation. Late and post-metamorphic faults, producing pervasive sheared, brittle fabric, probably developed during, and subsequent to F_{n+2} . Sense of motion along these faults has not been determined. It appears, however, that they occur as reactivated faults along older syn-metamorphic thrusts.

A current working hypothesis suggests that the Pinney Hollow, Stowe, and Mt. Abraham Formations, along with the white albitic schist member of the Hazens Notch Formation, developed as a continuous clastic sequence within a rift basin during the opening of Iapetus (fig. 6). The Ottawaquechee Formation, separating the westerly Pinney Hollow Formation from the easterly Stowe Formation, is interpreted as a large pre- or early-metamorphic klippe structurally overlying the continuous former units. The Stowe Formation is

therefore interpreted as an eastward continuation of the Pinney Hollow Formation. This is based upon the similarity between the quartz-sericite-chlorite-albite schists and albite-epidote-chlorite-calcite meta-igneous units in both formations. This is also in agreement with the rare earth and trace element geochemical data of Coish and others (1985, 1986) which places the Stowe Tholeiitic greenstones in an island arc/ MORB environment, and the Pinney Hollow alkaline basalts within an intraplate setting. The present interpretation would depict the Stowe greenstones forming as mafic flows and tuffs within highly attenuated continental crust ($\beta \sim 5.0$), and the Pinney Hollow meta-igneous units within thicker, more westerly, continental crust ($\beta \sim 2.5$) where assimilation would occur more readily.

The present interpretation depicts the Ottawuechee Formation as forming within an accretionary wedge environment. The numerous quartzose, albitic, and sandy lithologies could be incorporated following subduction of various clastic sequences, and their subsequent accretion into the wedge. Slivers of ocean crust (altered to serpentinite) would also be incorporated into the wedge. The black phyllite of the Ottawuechee Formation, comprising the majority of the wedge, would be the actual correlative sequence for the Taconic lithologies. The black quartzites may have been tectonically thickened during their accretion, represent part of the clastic sequence incorporated into the wedge, or a deposit shed from an entirely different source, to the north or possibly the east.

The carbonaceous schists of the Hazens Notch Formation would represent a sequence derived from the accretionary prism. Much of the sandy albitic schist within the Ottawuechee Formation is identical to the albitic matrix within the Hazens Notch carbonaceous units. These units, therefore, may be an amalgamation of the black phyllite and the incorporated sandy schists, derived during subaerial exposure of the wedge. This interpretation is in close agreement with that described by Stanley and Ratcliffe (1985). The Hazens Notch white albitic schist would be more properly depicted as an intermediate facies between the albitic schists and metawacke of the Hoosac Formation and the more easterly lithologies of the Pinney Hollow, Mt. Abraham, and Stowe Formations.

STOP 12 - DEPOSITIONAL CONTACT BETWEEN THE SCHIST AND GREENSTONE MEMBERS OF THE PINNEY HOLLOW FORMATION (Armstrong)- This outcrop affords one of the finest examples of the contact relationship between the quartz-chlorite-muscovite-albite schist and albite-epidote-chlorite-calcite greenstone members of the Pinney Hollow Formation. Due to the lack of a pervasive, fault-related schistosity, and the lack of lithic truncations, this contact is interpreted as depositional. Note the zone of alteration of the schist along the contact. This effect is most likely a result of fluid alteration during peak metamorphism, rather than a contact metamorphic effect during deposition of the meta-igneous greenstone. Since many of the greenstones within this area display planar compositional layers and laminae of albite-quartz and epidote-chlorite, their precursor may be ash, or water-lain tuffs. The contact between the schist and greenstone would therefore represent a conformable, contemporaneous sequence, possibly deposited in a rift basin. Greenstone is present at both ends of the outcrop, representing the limbs of a large syn-metamorphic fold (Fn) associated with the axial planar dominant schistosity (Sn) seen at this exposure within the schist.

Geochemical analysis of similar greenstones within the Pinney Hollow Formation by Coish, et. al (1985) yielded both alkaline and tholeiitic compositions, produced within intra-plate and MORD environments respectively.

A current working hypothesis places the Pinney Hollow Formation within a rift basin, distal from its western cratonic source region, upon very thin continental crust (Armstrong, 1987). Rift related igneous rocks (greenstones) of both tholeiitic and alkaline affinity could be produced in such an environment, where assimilation with continental crust would be minimum.

STOP 13 - Thatcher Brook Thrust (figs. 17, 18) (Armstrong and Kimball) - A 100 ft. thick sequence of carbonaceous albitic schist of the Hazens Notch Formation is tectonically juxtaposed between albite-chlorite-quartz-sericite schist of the Pinney Hollow Formation (fig. 17).

The eastern contact is a syn-metamorphic fault, displaying a well developed mylonitic fabric, and the truncation of schist sequences in the Pinney Hollow Formation. Asymmetric porphyroclasts, oriented within the dominant syn-metamorphic schistosity (S_n), display east-over-west motion. Microfabric analysis of C-S fabric, rotated porphyroblasts, asymmetric pressure shadows and porphyroclasts, and rotated schistosity, within syn-metamorphic fault zones, consistently display the same sense of motion, and are believed to be localized zones of intense strain development during S_n deformation. The fault zone fabric, at this locality, is oriented N07E, 65SE; similar to the regional S_n . The fault zone is characterized by the progressive intensification of S_n within the Pinney Hollow schist toward the contact, and by a tan weathering mylonitic quartzose schist within the fault zone. This fault (the Thatcher Brook Thrust) can be traced both north and south along strike where similar fault fabrics and tectonic slivers of foreign rock-types can be found.

The Thatcher Brook Thrust is deformed by a secondary phase of syn-metamorphic folds (F_{n+1}), the largest order of which consistently display counter-clockwise rotation. The average orientation of F_{n+1} is \sim N24E, 30° (fig.18). Axial plane orientations for this phase are similar to the undeformed orientations of S_n within this locality. This secondary fold phase is therefore believed to be a product of one syn-metamorphic deformational event, which produced the dominant schistosity and the primary phase of folds (F_n). Several syn-metamorphic faults within this region truncate F_{n+1} , having a fault zone fabric oriented similar to the F_{n+1} axial plane schistosity (S_{n+1}). The pervasiveness of S_{n+1} increases toward the fault zone where only limbs of earlier F_{n+1} can be recognized within a very planar fabric. This progression of fabric intensity demonstrates the relationship of S_{n+1} and many syn-metamorphic faults within this region. In many areas S_n and S_{n+1} form a composite schistosity, with S_{n+1} actually being the continuation of S_n deformation.

The western contact is not marked by a strong fault fabric and does not exhibit any mylonitization. Some small F_n folds within thin quartzites of the Hazens Notch Formation are sheared out along S_n ; this is a common feature within these highly strained rocks, but is not indicative of a major syn-metamorphic fault zone. The contact, interpreted as pre-metamorphic, has been highly altered during subsequent peak Taconian metamorphism. Kinematic indicators are absent. Fault motion is generally interpreted as being east to

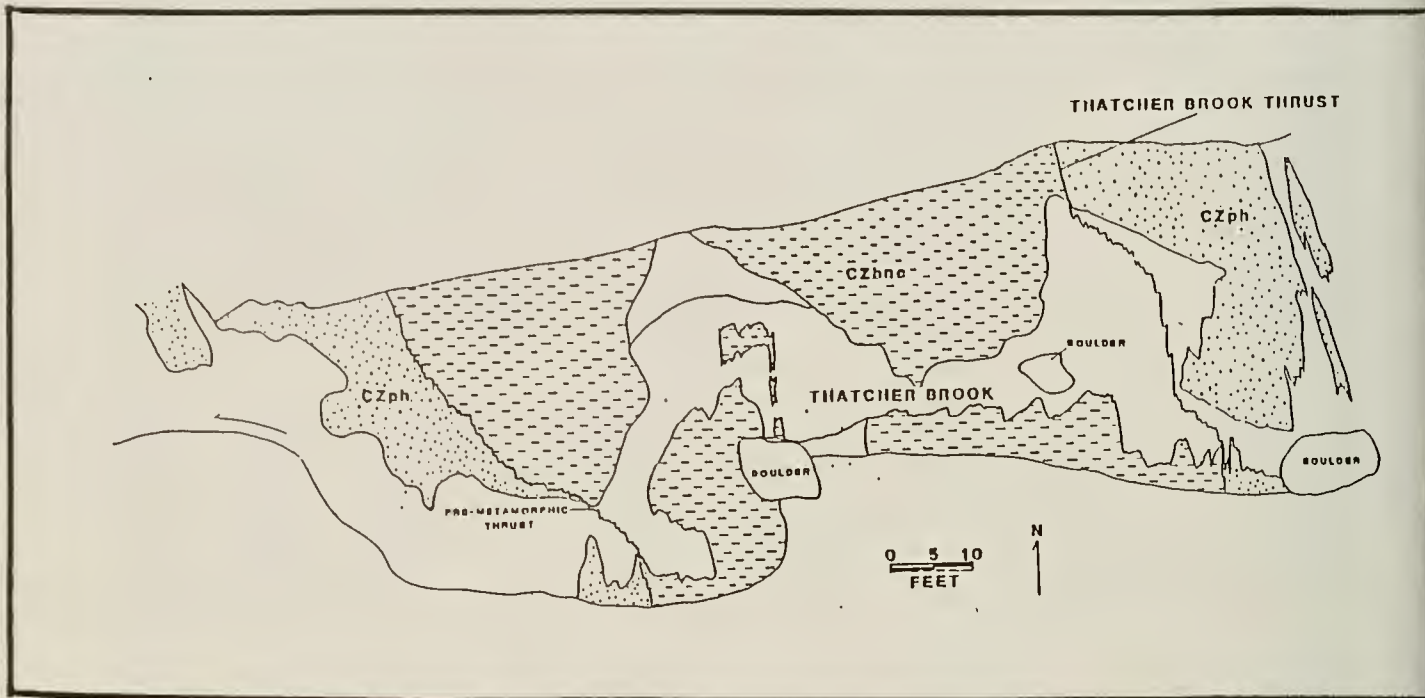


Figure 17

Detailed geologic map of the Thatcher Brook Thrust, a syn-metamorphic fault (eastern contact), and an earlier pre-metamorphic fault (western contact). CZph - quartz-chlorite-sericite-albite schist of the Pinney Hollow Formation; CZhnc - graphitic quartz-sericite-albite-chlorite schist of the Hazens Match Formation. The pre-metamorphic contact does not have a fault-related fabric due to annealing during subsequent Taconian metamorphism. This contact is interpreted as a thrust fault from relationships found along this same contact sequence, where the truncation of several greenstones of the Pinney Hollow Formation occur. The syn-metamorphic thrust displays a well developed mylonitic fault zone fabric. Asymmetric porphyroclasts along this contact show east-over-west motion. A small sequence of Pinney Hollow schist is truncated along this contact, evidenced by the emergence of a small greenstone body (within the Pinney Hollow) and the fault contact towards the southern end of the outcrop. Both faults are folded by the third phase of syn-metamorphic folds (Fn+1), pervasive throughout this exposure.

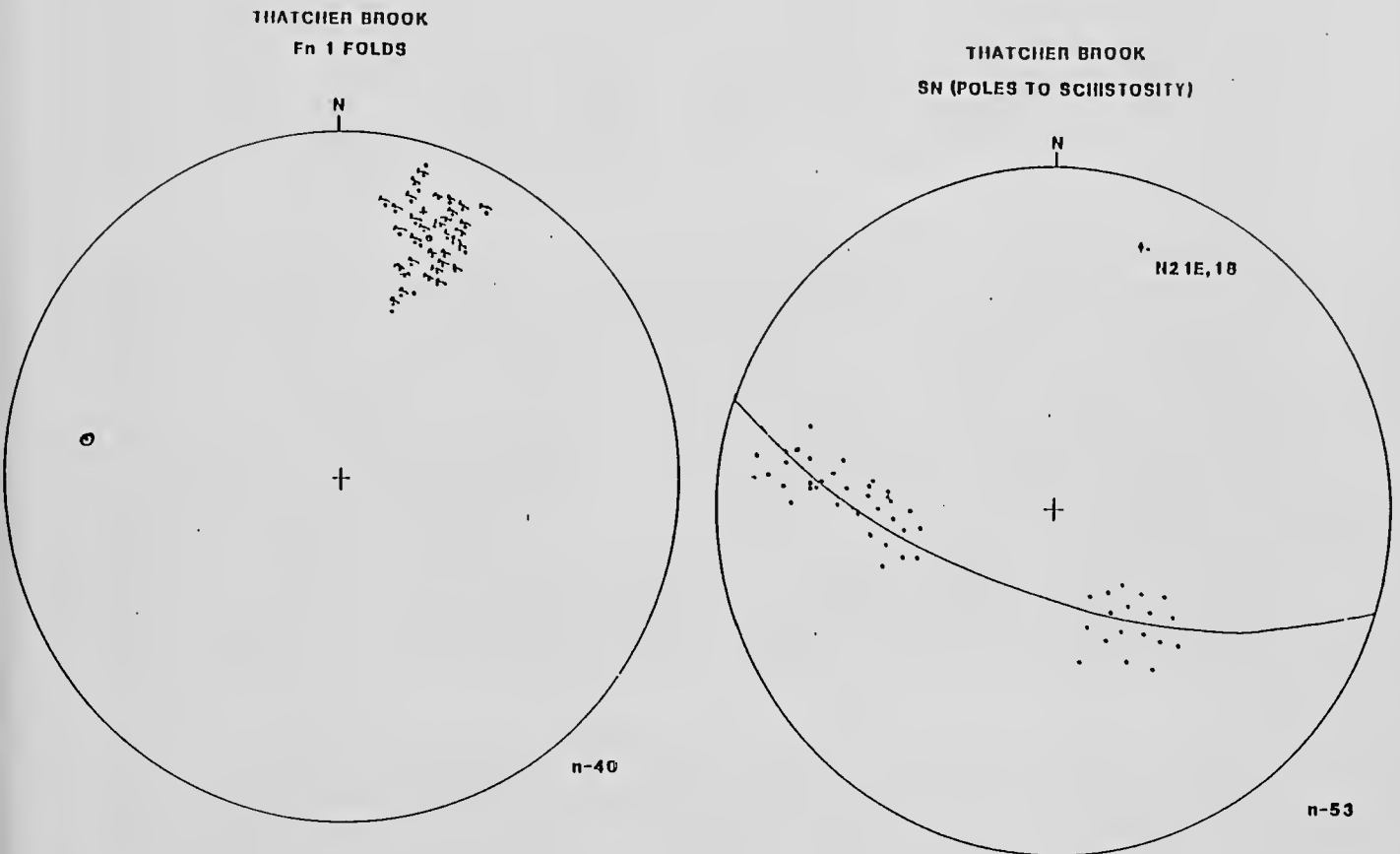


Figure 18

Equal area net projection of poles to the dominant schistosity (S_n) and F_{n+1} folds at Thatcher Brook. S_n is folded by F_{n+1} at this locality, forming a girdle around a best-fit great circle the pole to which is oriented N21E, 18. This orientation is shown as the cross within the F_{n+1} fold hinge net. The open circle denotes the average orientation of F_{n+1} at this locality. Arrows indicate the sense of asymmetry. The circle with the dot indicates the average orientation of undeformed F_{n+1} axial planes (S_{n+1}) oriented N06E, 67 SE.

west (Stanley and Ratcliffe, 1985). Distinctive quartzite beds within the carbonaceous schist are parallel to S_n . The distance between these quartzites and the pre-metamorphic contact is variable along strike, demonstrating the early tectonic nature of this contact. The tectonic character of this contact is also recognized to the north and south where lithic truncations, primarily of greenstones within the Pinney Hollow Formation, are present without fault zone fabric. The associated fault zone fabric of pre-metamorphic faults was obliterated during peak Taconian metamorphism, coeval with the development of the dominant schistosity (S_n). The carbonaceous schist of the Hazens Notch Formation is interpreted as structurally overlying the Pinney Hollow Formation above this pre-metamorphic fault. This interpretation stems from observations at other localities within the region. The western contact is also deformed by the later F_{n+1} .

Small shear bands cutting across S_n and F_{n+1} are the youngest deformational features found within this exposure. In some cases large albite porphyroblasts and pyrite cubes overgrow the shear band foliation, demonstrating a syn-metamorphic development. The foliation is parallel to F_{n+1} axial planes and appears to represent the initial development of a second schistosity (S_{n+1}). This S_{n+1} schistosity is exclusively found as a local fabric in association with a second phase of syn-metamorphic faults (S_{n+1}). Rotation of S_n by S_{n+1} consistently shows east-over-west motion. Kinematic indicators along S_{n+1} faults (identical to those found along S_n) also give a consistent east-over-west sense of motion. This third phase of syn-metamorphic deformation appears to be a continuum of the second phase with F_{n+1} folds occurring in strain hardened zones within regions of continued compression and secondary fault development (Armstrong, 1987; Kraus, 1987; Kimball, 1987; Ratcliffe, 1987, pers. comm.).

The evolution of this exposure can be summarized;

1. The development of pre-retrograde metamorphic thrust slices which always juxtapose discreet lithologies, usually from different tectonic environments. The carbonaceous schist of the Hazens Notch Formation thrust east to west over the Pinney Hollow Formation. The western contact was formed during this event.

2. The onset of Taconian retrograde metamorphism with associated deformational events producing an early schistosity (S_{n-1}) and associated folds (F_{n-1}) present as relict fold hinges and transposed foliation within the dominant schistosity (S_n) and associated sheath folds.

3. Development of the first phase of syn-metamorphic faults along zones of intense shearing within S_n . Movement was consistently east-over-west, incurring the juxtaposition of both discreet and similar lithologies. This phase produced the eastern contact which truncated the earlier pre-metamorphic structure

4. Continued deformation producing secondary folds (F_{n+1}), and associated S_{n+1} in many localities where deformation was intense. This event is responsible for the folding of both contacts at this locality.

5. Development of the second phase of syn-metamorphic faults along zones of intense S_{n+1} development. Movement was consistently east-over-west in a

manner similar to the first phase of faulting. Although not present at this locality, this event is observed throughout the region.

STOP 14 - SYN-METAMORPHIC THRUST BETWEEN THE PINNEY HOLLOW SCHIST

AND GREENSTONE MEMBERS (fig. 19) (Armstrong) - Two greenstone bodies are separated by a thin band of quartz-chlorite-muscovite-albite schist (fig. 19). The schist displays a well developed fault zone schistosity and has been altered to a tan, quartz-feldspar mylonite in many localities. This schistose mylonite is diagnostic of many syn-metamorphic fault zones throughout this region. Kinematic indicators, such as asymmetric porphyroclasts, C-S fabric, and rotated porphyroblasts, consistently show east over west displacement along these syn-metamorphic thrusts.

The map pattern of the eastern greenstone body displays east over west asymmetric, syn-metamorphic folds (F_n) which have an associated axial plane schistosity. F_n folds within the schists of the study area are sheath-like, with hinges that plunge at steep angles down the plane of the S_n schistosity. The folds within the greenstones vary from reclined to inclined, isoclinal folds. Variation in fold geometry is directly related to the competency of the lithologies involved. The competent greenstones are more resistant to the simple shear mechanism responsible for sheath development in the pelitic assemblages within this region. F_n fold axes within the greenstones have undergone less physical rotation and are not oriented parallel to the elongation mineral lineation as are the hinges of the pelitic sheath folds. The east over west asymmetry of the minor folds indicates that the eastern greenstone body is on the western limb of a synform. The closure of this F_n fold can be traced toward the northwest where it is truncated along the syn-metamorphic fault approximately 275 feet to the north of Allbee Brook bridge (fig. 19; pt. A).

The western greenstone body is seen in contact with the fault zone schist along the road, directly south of the bridge (fig. 19; pt. B). The greenstone appears to be continuous, and has been mapped in a large isoclinal fold directly west of this locality. The greenstone does not appear to be truncated, but only mildly deformed by the adjacent thrust fault.

The two greenstones are separated by a thin band of schist that displays a pervasive fault zone fabric. A quartzose, tan weathering mylonitic schist is found within this zone at many of the schist outcrops (pt. C, for example). Microfabric analyses of C-S fabric, and asymmetric porphyroclasts and pressure shadows, along syn-metamorphic faults, consistently show east to west displacement. The displacement along the syn-metamorphic faults must be much less than that along the pre-metamorphic thrusts since they juxtapose similar, as well as discreet lithologies.

The fault zone schist is continuous over the entire length of this suite of outcrops, and can be traced to the north where the western greenstone is abruptly truncated (~2500 ft north), and on into Granville Gulf.

A tan weathering schist is also found immediately southeast of the bridge along the folded eastern edge of the eastern greenstone (fig. 19; pt. D). This schist, although not a fault zone, does display a pervasive fabric which cross-cuts the F_n fold geometry. This fabric is most likely the result of large localized strain, which would likely occur along this contrast in

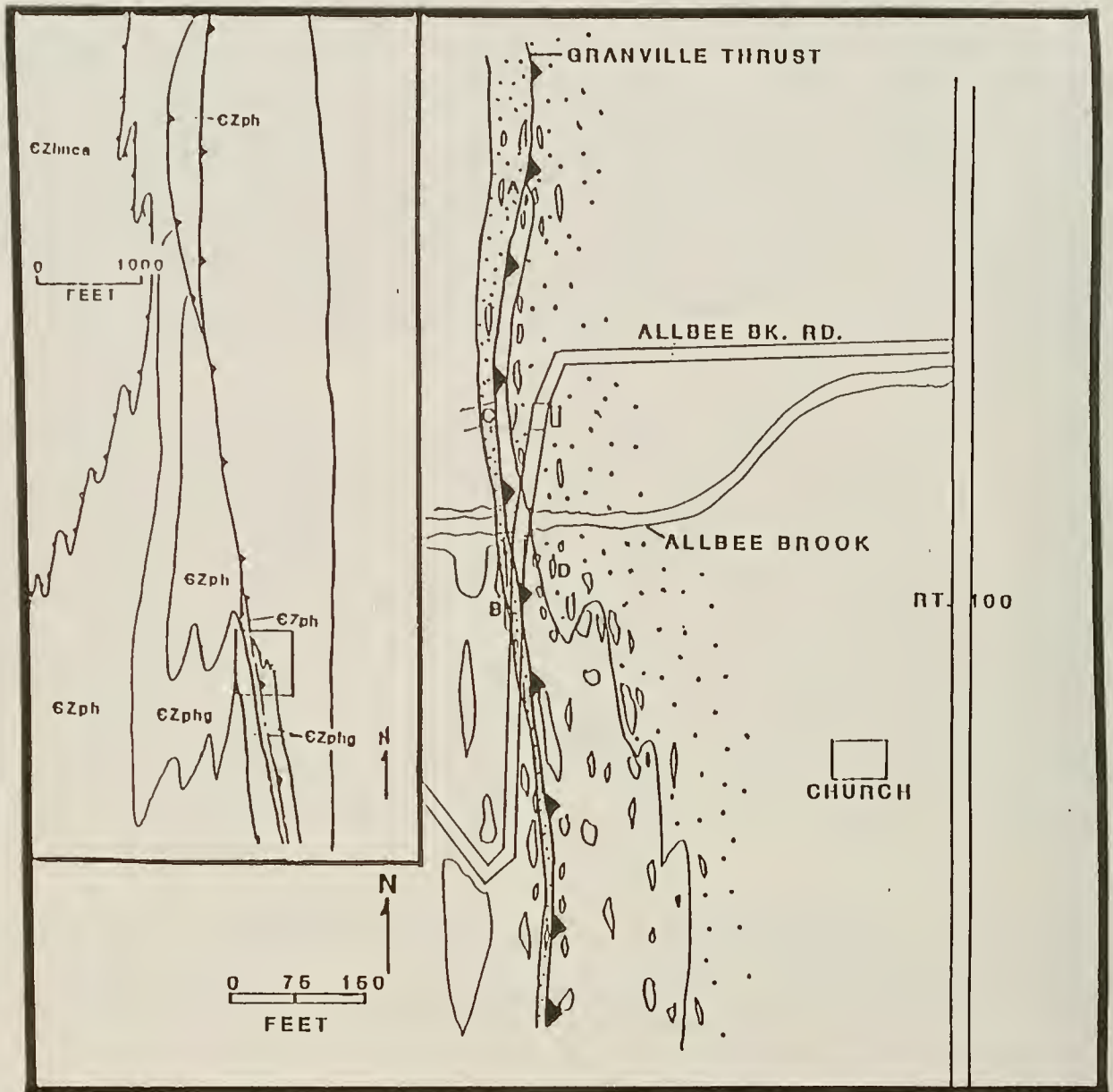


Figure 19

Detailed and regional geologic maps of the Granville Thrust, located in Lower Granville, central Vermont. The Pinney Hollow sliver-green quartz-chlorite-sericite-albite schist is stippled in the detailed map. The Pinney Hollow albite-epidote-chlorite-calcite greenstones are unpatterned. The eastern greenstone body is separated from the larger western greenstone by the fault zone schist contained within the Granville thrust. The map pattern east-over-west folds within the eastern greenstone plunge to the north, and are parasitic folds along the eastern limb of a large antiform. The folded greenstone is truncated along the fault over its entire length and is terminated to the north at pt. A. The western greenstone body is also truncated, to the north, along this fault as seen in the regional inset.

competency between the resistant greenstone and the pliable schist.

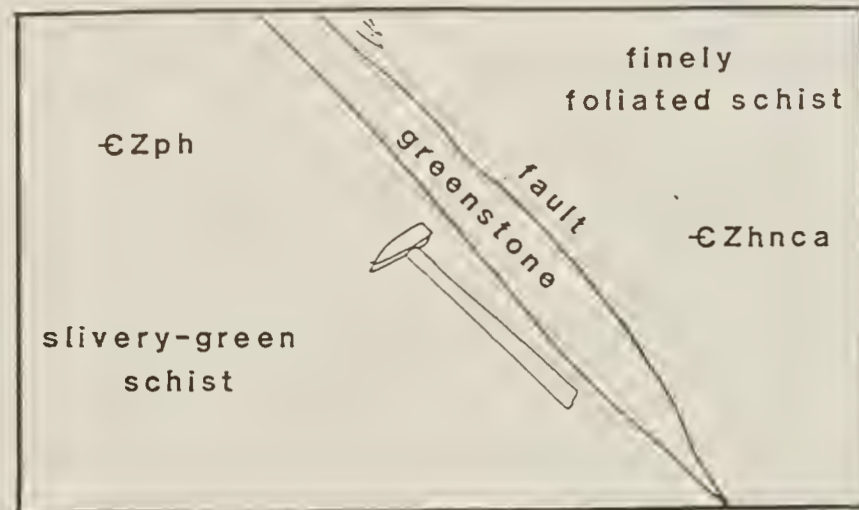
This large scale syn-metamorphic thrust (the Granville Thrust Zone) is the primary feature that controls the White River Valley and Granville Gulf. Ongoing work along the eastern side of this valley, to the south near Hancock, has demonstrated the truncation of a large greenstone body (~ 1 km long x 300 m wide) along an ancillary thrust fault rooting from this zone. Numerous lithic truncations, tectonic slivers, and fault zone fabrics may be found along the entire length of this zone.

Stop 15 - Granville Gulf - (fig. 20 and 21; map will be available at conference) - (Stanley) - BE VERY CAREFUL AND STAY OFF THE ROAD BECAUSE THIS IS A DANGEROUS CURVE ON ROUTE 100. This stop is very important because it shows the complex geology along the boundary between the Pinney Hollow and the Ottauquechee Formations. Here the Pinney Hollow consists of silvery green chlorite - muscovite - albite - quartz schist interlayered depositionally with two belts of mafic schist. The eastern contact (fig. 20) is a synmetamorphic fault in which the carbonaceous albitic schist of the Hazens Notch (CZha) truncates a thin lense of chlorite-rich schist with brown-weathered material on the schistosity (mafic schist?). The schistosity in the carbonaceous schist is so well developed and closely spaced that it weathers very readily. In thin section the foliation wraps around albite porphyroblasts. This "papery-schist" fabric is characteristic of late metamorphic or post metamorphic faults and has been observed at several localities along the western boundary of the Ottauquechee Formation. Note that the foliation in the carbonaceous schist next to the silvery green schist of the Pinney Hollow is more widely spaced and the rock is more coherent. This contact is a synmetamorphic fault because the albite porphyroblasts grew across the foliation. Thus this zone is one that has a compound history (table 1) - an earlier synmetamorphic fault that was reactivated as a late or postmetamorphic fault.

Continuing eastward along the outcrop you will find a tan - weathering schist with brown spots. This rock is found along several belts in this zone. In thin section it consists of fine-grained, well oriented muscovite and quartz. In sections cut parallel to the lineation the foliation is planar. Small grains of muscovite and quartz, which are oriented at a small angle to the dominant foliation, form a C-S fabric and indicate east-over-west movement. In sections cut perpendicular to the lineation the foliation anastomoses. I consider these rocks to be mylonitic schists which developed from the Pinney Hollow schist along synmetamorphic faults. Continuing eastward you will find another mafic schist, then another belt of carbonaceous schist which passes into black carbonaceous schist typical of the Ottauquechee Formation. The fabric throughout this belt is well represented by figure 21.



The Ottawaquechee thrust fault zone at Granville Gulf, central Vermont. A close-up photograph of the fault contact showing a thin greenstone lense (above the hammer) of the Pinney Hollow Formation (lower left) in contact with the well foliated, rusty-weathering schist (CZhnca) of the Hazens Notch Formation. Greenstones are common in the Pinney Hollow Formation but are absent in the caroonaceous, albitic schist (CZhnca) of the Hazens Notch Formation.



As you can see on the 1:12,000 scale geologic map there are 3 mappable greenstone bores in the Granville Notch area. Laird (1977) and Laird and Albee (1981) found barroisitic amphibolites with actinolitic rims in these greenstones and report $40\text{Ar}/39\text{Ar}$ ages ranging from 471 m.y. to 448 m.y.. The analyzed grains occur near quartz veins and in the matrix of the greenstone beds. The quartz veins throughout the Pinney Hollow are deformed and flattened parallel to the dominant foliation S_n , but cut across S_{n-1} where it can be observed. Although the barroisitic amphiboles probably formed during the development of S_{n-1} , since they are rimmed by actinolite, many of the grains are oriented in the dominant foliation S_n and appear to be oriented parallel to the lineation as viewed in oriented thin sections. These relations indicate, therefore, that the synmetamorphic faults formed during the Taconian Orogeny of Middle Ordovician age. This conclusion would suggest that all the structures that are displayed on the accompanying maps and photographs were formed during this time.

Figure 20

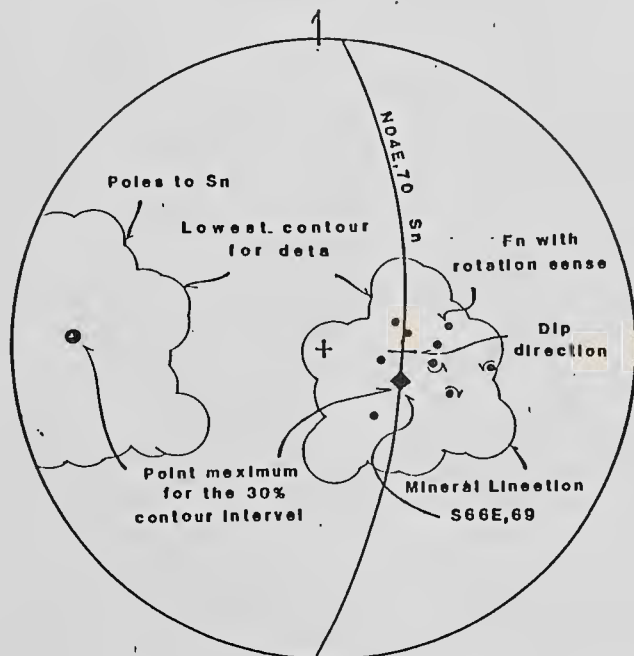


Figure 21

STRUCTURAL FABRIC FROM THE GRANVILLE GULF AREA

Lower hemisphere equal area net showing the orientation of 179 poles to the dominant schistosity (Sn), the orientation of 90 elongate grains, grain clusters, intersection lineations and quartz rods, and the orientation of 9 hinges of Fn folds. The center of the point maximum for the 30% contour interval corresponds to a surface (Sn) striking N04E and dipping to the east at 70 degrees. The corresponding point maximum for the mineral lineation plunges 69 degrees along a trend of S66E. This orientation represents the transport direction and it closely parallels the dip direction for the dominant Sn surface. Some thin sections cut parallel to this direction show asymmetrical fabrics that indicate an east-over-west sense of movement. See the photographs of thin sections from the mylonitic schist from the Ottauquechee fault zone at Granville Gulf. Note that all the observed Fn folds are oriented close to this direction and therefore are reclined. None of these folds have a sheath-fold shape. Compare this fabric with the fabric from Waitsfield where the Fn folds have a more varied orientation. The Fn folds in these two diagrams suggest that Fn hinges are rotated toward the transport direction as the Ottauquechee thrust fault is approached.

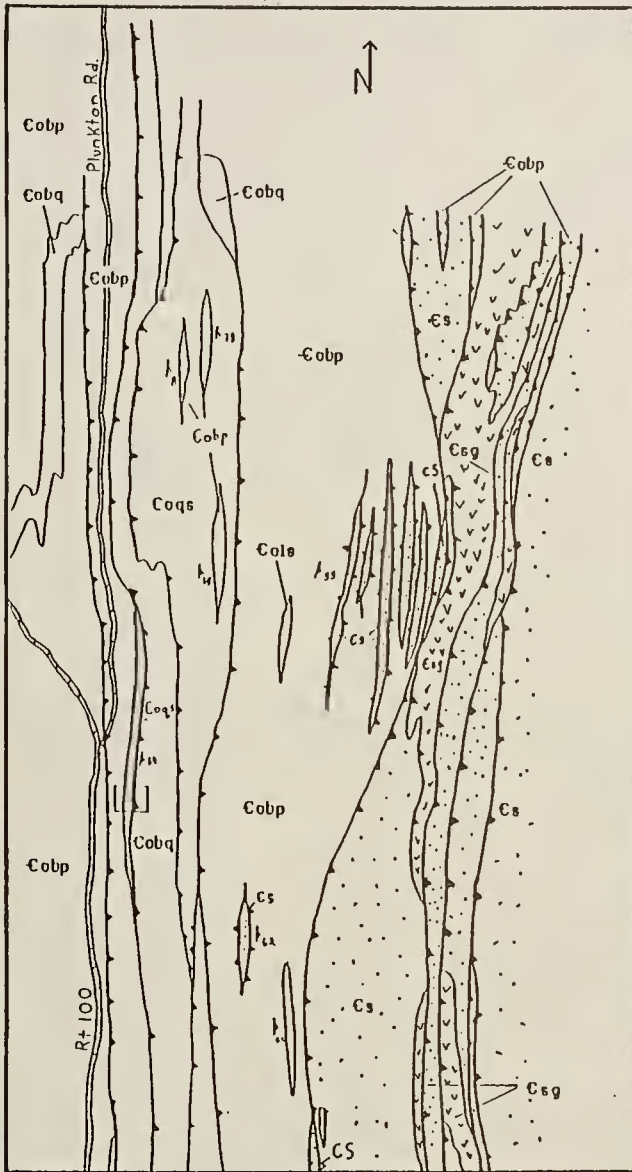
We will discuss the map and several cross sections that offer alternative interpretations of the geology.

The belts of mafic schist are important here because they have been geochemically analyzed by Coish and Masinter (1986). Their analyses of major elements and Y/Nb variations indicate that the mafic schists were originally basalts of transitional to tholeiitic character. The immobile trace element abundances suggest that they probably represent continental tholeiites that erupted through thin continental or transitional crust (Masinter, 1986). Laird has also studied the amphibole petrology and found barroisitic hornblende surrounded by actinolite (sample locality V14J of Laird and Albee, 1981; Laird and others, 1984; Laird this guidebook). The amphibole textures and compositions indicate that these rocks have undergone two metamorphic events, an earlier medium-high pressure metamorphism represented by the barroisitic cores and a lower greenschist facies metamorphism represented by the actinolite rims. Several samples of barroisite, one within the main part of the mafic schist and another near a quartz vein, gave $^{39}\text{Ar}/^{40}\text{Ar}$ ages of 471 and 448. In oriented thin sections the actinolite is oriented parallel to the S_n schistosity. The barroisitic hornblende appears to be flattened parallel to S_n and elongated parallel to the mineral lineation. Because the barroisitic core has been partially altered to actinolite, however, this relations may be deceptive. There is a distinct possibility that the barroisite grew during the formation of the older S_{n-1} schistosity. U-stage analysis has not been done on these rocks. We will discuss these implications as it pertains to the age of deformation and the fault classification shown on Table 1.

Stop 16 - Structural fabric in the Ottawaquechee Formation (figs. 22 and 23) (Kraus) - NO HAMMERS, PLEASE - This stop is located just east of Route 100 south of the junction of Plunkton Road and Route 100 between Warren and Granville Gulf. The outcrop is on the south side of a stream that forms a small ravine and is approximately 30 m east of Route 100 (fig. 22).

The outcrop is located in the quartzose schist member of the Ottawaquechee Formation. Other units in this formation are the black phyllite and the black quartzite members. The black phyllite is graphitic and is composed of quartz, muscovite, chlorite and pyrite. The black quartzite is graphitic, commonly contains veins of white quartz, and is found in layers 2 cm to 30 m thick. The white quartzose schist is composed of sand-size quartz, muscovite, chlorite and albite. It places it contains thin interlayers of graphitic phyllite. A few blue quartz grains can be found at this locality. All three members can be seen in the outcrops along the stream.

The main feature at this stop is a spectacular fold-structure exposed in the quartzose schist. Differential weathering has left a remarkable display of tightly-folded layers, three foliations, intersection lineations, slickengrooves, mineral lineations, and associated minor faults (fig. 23). PLEASE BE VERY CAREFUL AND DO



Es	Stowe Fm. chlorite-sericite-muscovite schist
Esq	Stowe Fm. greenstone
εobp	Ottawaquechee Fm. black phyllite
εobq	Ottawaquechee Fm. black quartzite
εoqs	Ottawaquechee Fm. quartzose schist
εots	Ottawaquechee Fm. talc schist

	Sn foliation (strike and dip)
	Pre- and Syn-metamorphic faults
	Sedimentary contact

- Csg
- Es

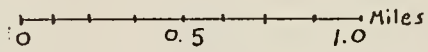


Figure 22

Location and local geology map. Stop location is indicated by the small square just off Rt. 100.

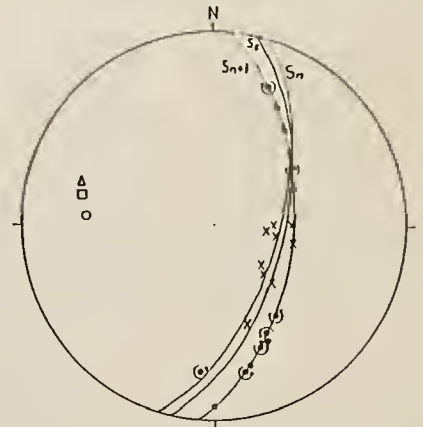


Figure 23

- = point maxima to S_f poles, n = 11
- = point maxima to S_n poles, n = 11
- = point maxima to S_{n+1} poles, n = 32
- = transport lineations
- = fold hinges
- = fold hinges with rotation sense

Lower hemisphere equal area projection of structural elements. No separation arc is implied due to uncertain fold to limb relationships.

NOT DISTURB THE FABRIC OR THE CURSE OF THE TROLL WILL BE UPON YOU!

The dominant schistosity, S_n , was formed in the rock during or near the peak of metamorphism. The S_n surfaces are spaced 1 to 5mm apart. An older S_{n-1} schistosity can be recognized in places between the S_n surfaces. A progression in the intensity of a third schistosity, S_{n+1} , can be seen beginning at the west end of the outcrop and increasing to the west where it becomes the dominant schistosity in the outcrop.

A number of discreet fault surfaces are present as thin layers of quartz that cut across S_n and S_{n+1} . Slickengrooves and clusters of elongate minerals form a prominent lineation oriented nearly parallel to the dip of the faults. This lineation represents the transport direction on the faults. The other lineations in the outcrop are formed by the intersection of schistosity and the folded layers. This lineation is related to folding and has no direct relation to the transport lineations which are confined to the fault surfaces.

Numerous minor folds can be seen in the outcrop. Many of the folds are represented as isolated hinges that have many different orientations in the plane of S_{n+1} (fig. 23). These folds represent parasitic folds on larger asymmetrical folds. Many of the folds are offset across faults and indicate eastward movement.

This outcrop is interpreted to occur on the faulted west limb of an overturned anticlinal structure of F_{n+1} age. Rotation of the parasitic folds and the formation of the transport lineations occurred during shearing and subsequent faulting. This outcrop is an exposure of a synmetamorphic fault that developed after the peak of metamorphism.

Stop 17 - Refolded fold of F_{n-1} and F_n age in the Pinney Hollow Formation (fig. 24) (Stanley, Armstrong) - This outcrop is located in the front yard of a private home in Alpine Village which is located just west of Plunkton Road. PLEASE, NO HAMMERS. This outcrop is quite rare in that it is one of the few places where refolded folds of F_{n-1} and F_n age are preserved in the mafic schist and silvery green schist of the Pinney Hollow. We believe that the geometry preserved in this outcrop is representative of the larger structure in the Pinney Hollow Formation.

Figure 24 is a geologic map of the outcrop drawn from a photomosaic. The relations are therefore portrayed with a fair degree of accuracy. Carefully trace the contact between the mafic schist (greenstone) and the silvery schist and compare your findings with figure 24. The axial surface (S_n) of the F_n fold trends to the north. It folds S_{n-1} which is well preserved in the hinges of F_n folds in the greenstone but is largely transposed in the adjacent schist. After you have recognized the structure see if the folds continue to the east and west across the outcrop. You will find that they are more difficult to follow in these directions. We suggest that the refolded fold in the center of the

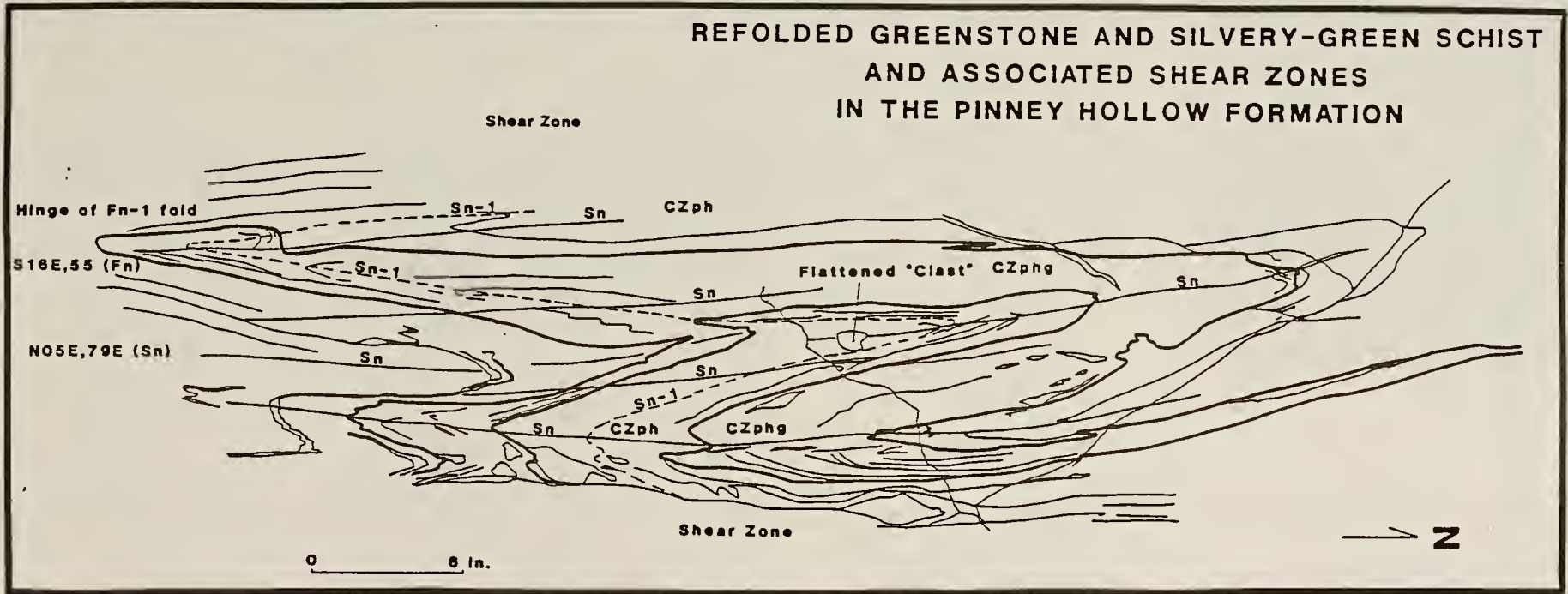


Figure 24

outcrop is truncated by shear zones to the west and east. The eastern one is marked by a string of quartz lenses that are oriented parallel to the Sn schistosity. Follow this zone northward (toward the house) and you will find that the zone is oriented parallel to the the older (Sn-1) layering. The western shear zone is less distinct but appears to have the same geometry. We believe that these refolded folds and associated shear zones are representative of part of the internal fabric in the Pinney Hollow Formation. The shear zones would represent the linear symetamorphic faults whereas the refolded fold, which here is nearly reclined in position, would represent the sheared-out "sheath" folds.

Also note the small dark lenses 4-6 cm in width. These lenses are composed of titanohematite with exsolution "ellipsoids" of ilmenite and separate grains of magnetite (Armstrong, 1986, class paper). The magnetite grains contain (111) lamellae of ilmeno-hematite. Pods such as these are derived from hematitic shales during metamorphism. The composition of the titaniferous hematite host and associated lamellae is hematite 80 percent and ilmenite 20 percent and is based on the analysis of 7 grains using the microprobe at Middlebury College. The significance of these analyses will be discussed by Armstrong.

Stop 18 - Fabric within the Pinney Hollow Schist (fig. 25a and 25b) (Stanley and Coleman) This outcrop is located at the junction of Route 100 and the Lincoln Gap Road. The contact between the Hazens Notch Formation is located just west of this junction. An outcrop of the carbonaceous albitic schist (CZhnca) of the Hazens Notch is located just west of the junction. The outcrop of the Pinney Hollow schist is on the east side of the road. NO HAMMERS PLEASE. JUST OBSERVE AND APPRECIATE WHAT NATURE HAS PRESERVED!

The fabric in the schist is very complex. Here the dominant schistosity (Sn), as we see it in strike section, is well developed and anastomoses. Prominent mineral lineations and reclined fold hinges plunge down the dip. This fabric is similar to the fabric found in many places in the Pinney Hollow (for example Stops 13, 14, and 15). Coleman and Journeay (1987) suggest that these fabrics record a history of rotational strain involving layer-parallel shear, flattening perpendicular to the Sn schistosity and extreme elongation parallel to the mineral lineation. In many places this fabric is offset by younger shear zones (C surfaces) that record either left-lateral or right-lateral displacement. These shear zones are not as obvious in dip sections in outcrop. In thin section, however, Coleman and Journeay (1986) observed microscopic shear bands and asymmetric albite porphyroblasts in both the strike and dip sections. These shear bands wrap around older albite porphyroblasts which preserve inclusion trails of clockwise and counterclockwise sense. In the dip section the down-to-the-east shear bands were more prevalent than the up-to-the-west bands. These data suggest several interpretations. The first is that they simply record late flattening across the belt. As a result we should find

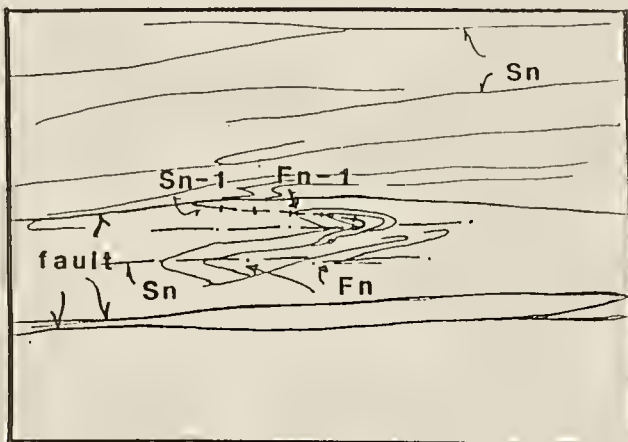


Figure 25a

Origin of synmetamorphic faults Fault zone fabrics in the silvery schist of the Pinney Hollow Formation showing the sheared off limb of an F_n fold which has in turn folded an older F_{n-1} fold (center of the diagram). These relations suggest that synmetamorphic faults develop through a process of intense shearing along the limbs of tight to isoclinal folds. Thus the fold developed before the fault unlike foreland folds that develop as a result of irregularities (ramps etc.) along a propagating fault surface. The difference in mechanism may be due to the absence of strong contrasts in ductility in the hinterland compared to the foreland where beds of distinctly different mechanical properties form major surfaces of weakness. The anastomosing foliation, commonly measured as the dominant foliation S_n , is really a composite foliation made up of S_n and the transposed remains of S_{n-1} and compositional layering. The schist in many synmetamorphic fault zones is commonly more strongly foliated than it is away from the faults. This indicates that movement has occurred not only during but slightly after metamorphism. Although original bedding is rarely preserved in these rocks, it can be found along contacts with beds of greenstone and quartzite.



Figure 25b

Foliated and fragmented pegmatite in a synmetamorphic fault in the silvery green schist of the Pinney Hollow Formation The foliation is the pervasive S_n foliation. Isolated quartz stringers and veins are strung out parallel to the dominant foliation. The relations are best seen directly above and to the right of the pen. The schist is composed of quartz, muscovite, chlorite and albite. The fabric in the pegmatite indicate that F_n deformation occurred at the ductile-brittle transition where quartz deformed ductily and feldspar deformed by brittle to semi-brittle fracture. Suggested temperatures and pressures are in the range of 350°C to 400°C and 3 or 4 Kbars (10 to 13 km.) based on experimental work on feldspar (Tullis and Yund, 1977) and the estimated stability of chlorite.

approximately equal numbers of shear bands with opposite senses of displacement. The second interpretation is based on Coleman's and Journeay's study where they found that 85% of the shear bands and asymmetrical albite porphyroblasts record down-to-the-east movement. In this interpretation these structures record late displacement in which greenschist facies hanging-wall rocks of the Pinney Hollow and Hazens Notch Formations moved down to the east over relatively higher-grade footwall rocks of the Mt. Abraham Schist. This movement may have been a result of the uplift and arching of the Lincon massif and its cover immediately to the east. Movement over a deep ramp may have been responsible for this uplift.

In one place in the outcrop an older Fn-1 fold is preserved between the Sn schistosity (fig. 25a). In another place an older pegmatite is disrupted by Sn and the younger shear bands (fig. 25b).

Perhaps the most important lesson that can be learned from this outcrop is that the fabric is clearly the result of very complex and intense strain. Originally, we thought that the fabric was largely the result of simple shear associated with faulting. Pure shear (flattening) certainly has played a role in the development of the shear bands. The earlier fabric may well have been the result of a large component of pure shear since asymmetrical structures with a consistent sense of displacement are absent. The rotated trails in the albite certainly suggest this, although more work should be done on these structures. Our present position is that one should be cautious in concluding that outcrops like this one are simply the result of synmetamorphic faulting. Other such criteria as the geometry of map units should be used in conjunction with the fabric.

Stop 19 - Fault zones in the Ottawaquechee Formation north of Waitsfield (fig. 26) (Walsh) - These series of outcrops are found on either side of Route 100.

Station A (fig. 26) - A folded contact between the Ottawaquechee black phyllite (Cobp) and the Ottawaquechee greenstone (Cobg) is well exposed at this locality. The black phyllite can be seen in the cores of several of the folds along the contact with the greenstone. The dominant schistosity (Sn) is axial planar to the folds, thus indicating that they are Fn in age. The geometry of the Fn folds is best seen in the greenstone where the folds are isoclinal and have hinges that plunge steeply to the south and the north. The geometry of these folds is consistent throughout the area and is most likely related to larger map-scale folds.

The contact between the black phyllite and the greenstone is interpreted as depositional. From the relationship between the contact, the Fn folds, and the dominant schistosity it is evident that the contact predates the later structures. No evidence, however, is seen for a pre- or early metamorphic fault along this contact.

Bedrock Geologic Map of the Northhorn Fayston Area in the Vicinity of Route 100

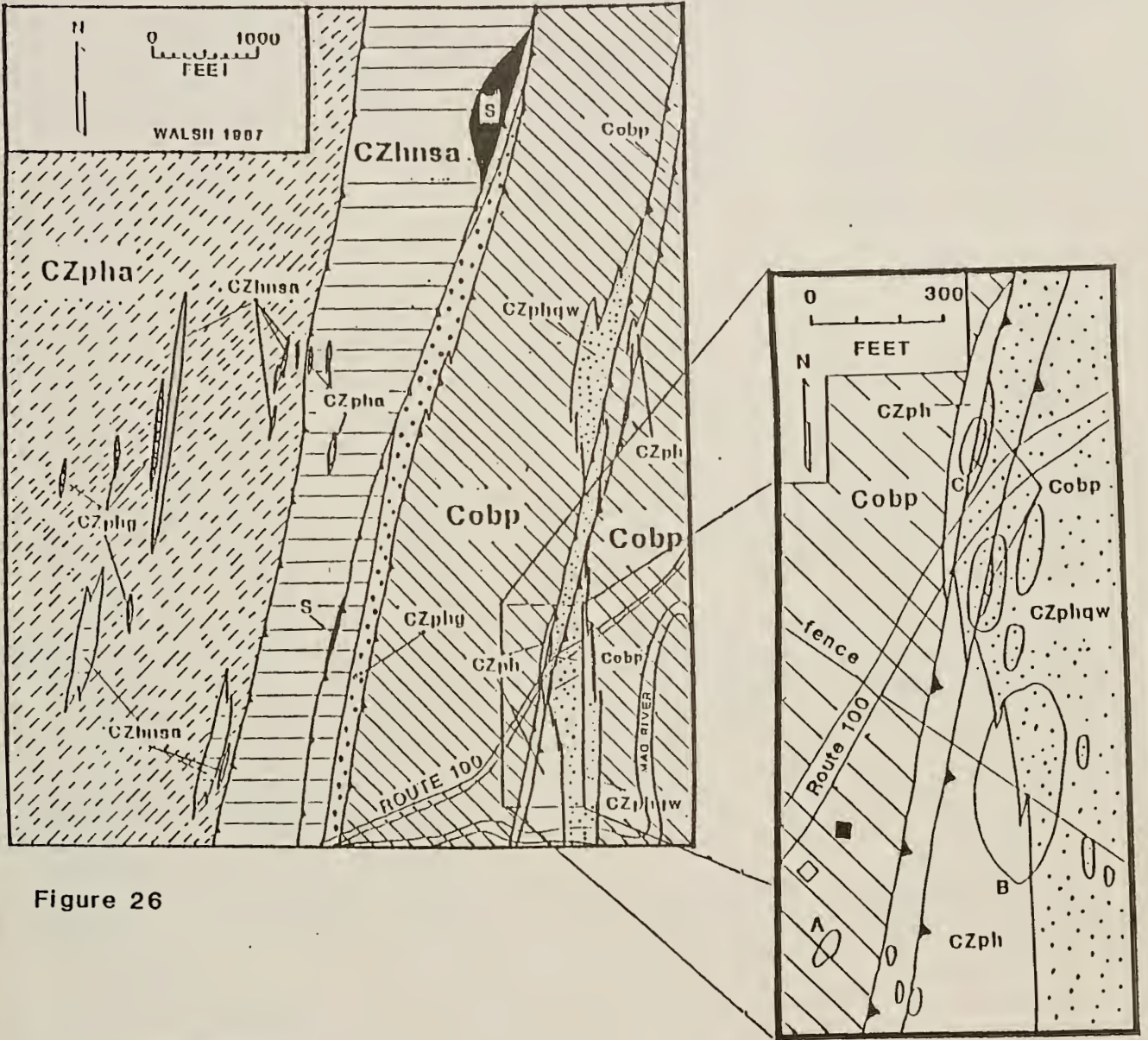


Figure 26

EXPLANATION

Lithologies:

- CZpha: albite-quartz-chlorite-muscovite schist
- CZph: quartz-albite-chlorite-muscovite-magnetite schist
- CZphqw: quartz-rich wacke/schist
- CZphg: albite-epidote-chlorite-calcite-actinolite schist/greenstone
- CZphsa: interlayered quartz-chlorite-muscovite-albite schist and albitic carbonaceous schist
- Cobp: graphitic pyritiferous quartz-muscovite-chlorite schist
- S: undifferentiated serpentinite and talc carbonate

Symbols:

- Thrust fault--teeth on upper plate
- Outcrop
- House
- Barn
- Station location

Stations B and C - The following series of outcrops will demonstrate the gradational nature of the contact between the quartz - albite - muscovite - chlorite - magnetite schist (CZph) and the quartz-rich wacke (CZphqw) of the Pinney Hollow Formation, as well as lithic truncations of the schist and the wacke against the fault-controlled contact with the Ottauquechee black phyllite.

Station B - The schist of the Pinney Hollow is exposed at this station. The schist is characterized by a silvery-green appearance on the foliation surface, well defined chlorite streaks, quartz rods, and magnetite octahedra.

The quartz - rich wacke grades from a light-colored sandy schist near the contact with the silvery-green schist to a well-foliated quartz-rich wacke toward the east in the pasture. Proceeding from Station B north towards Station C, outcrops of the quartz-rich wacke are encountered near the top of the hill. Continue northwest down the hill, out of the woods, and across the pasture towards the outcrops of the quartz-rich wacke on the east side of Route 100. Note how the contact between the schist and the wacke trends northwest, crossing the dominant schistosity. The contact on the map is inferred from the geometry of the outcrop-scale Fn folds similar to those seen at Station A.

Station C - The outcrop at Station C is located in a stand of trees on the west side of Route 100. The contact between the silvery-green schist of the Pinney Hollow and the Ottauquechee black phyllite is best exposed at the north end of the outcrop. The contact is sharp and parallels the dominant schistosity. Based on the mapped lithic truncation of the Pinney Hollow schist and the wacke against this contact and the presence of similar geology along their belt to the north (fig. 26) the contact is interpreted as a synmetamorphic fault. The foliations are seen near the contact at the north end of the outcrop. The acute angle between the two foliations indicates an east-over-west sense of shear. If the two foliations are related to the contact it would support the synmetamorphic fault interpretation.

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METAMORPHISM OF PRE-SILURIAN ROCKS, CENTRAL VERMONT

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(This note is meant to accompany field trip B-8 by Stanley and others, this volume. Please refer to that paper for geologic units, structural features, and field trip stops.)

Our understanding of the metamorphism in central Vermont is based on petrologic and isotopic studies of intercalated pelitic, mafic, and calcareous rocks. West of the Green Mountain anticlinorium (GMA) metamorphic grade increases from chlorite to kyanite zone with the highest metamorphic grade east of the Lincoln Massif (lm, fig. 1). Biotite to kyanite zone rocks also crop out east of the GMA, with the highest metamorphic grade along the Northfield, Worcester, and Elmore mountains (nm, wm, em, fig. 1).

Weakly metamorphosed limestones west of the Lincoln Massif record calcite-dolomite, oxygen, and carbon isotope temperatures between 210 and 295 C (Sheppard & Schwarcz, 1970), while biotite grade dolomite and marble give temperatures between 255 and 400 C. North of the Lincoln Massif biotite grade pelitic schist gives an oxygen isotope temperature of 435 C (Garlick and Epstein, 1967; all stable isotope temperatures have been recalculated using the fractionation curves compiled by Friedman & O'Neil, 1977). Garnet grade mafic schist gives calcite-dolomite and amphibole-plagioclase temperatures of 489 to 450 C and an amphibole-plagioclase pressure of 7 to >8 Kbar. At Mount Grant (m, fig. 1) oxygen isotopes indicate kyanite grade metamorphism at 420 to 465 C (Garlick and Epstein, 1967).

Pelitic assemblages reported by Albee (1965, 1968) indicate that metamorphism reached a higher temperature east of the GMA than farther west (Worcester Mountains vs. Green Mountains, figure 2). Amphibole and plagioclase compositions in mafic schist do not support this hypothesis, perhaps because of retrograde metamorphism. At Elmore Mountain, for example, the earlier pelitic assemblage kyanite + garnet + biotite can be observed even though all of these minerals are pseudomorphed (Albee, 1972, Stop I-2). Hornblende is pseudomorphed by actinolite (Laird and Albee, 1981) and has similar composition to hornblende in garnet grade mafic schist near Lincoln. If the highest grade hornblende is not preserved, there is no way to tell from electron microprobe analyses how high grade the mafic rock was before retrogradation. Plagioclase in mafic rocks is albite both east and west of the GMA.

One cannot distinguish the pressure of metamorphism east and west of the GMA from the pelitic samples alone. The composition of amphibole in amphibole + chlorite + epidote + plagioclase + quartz

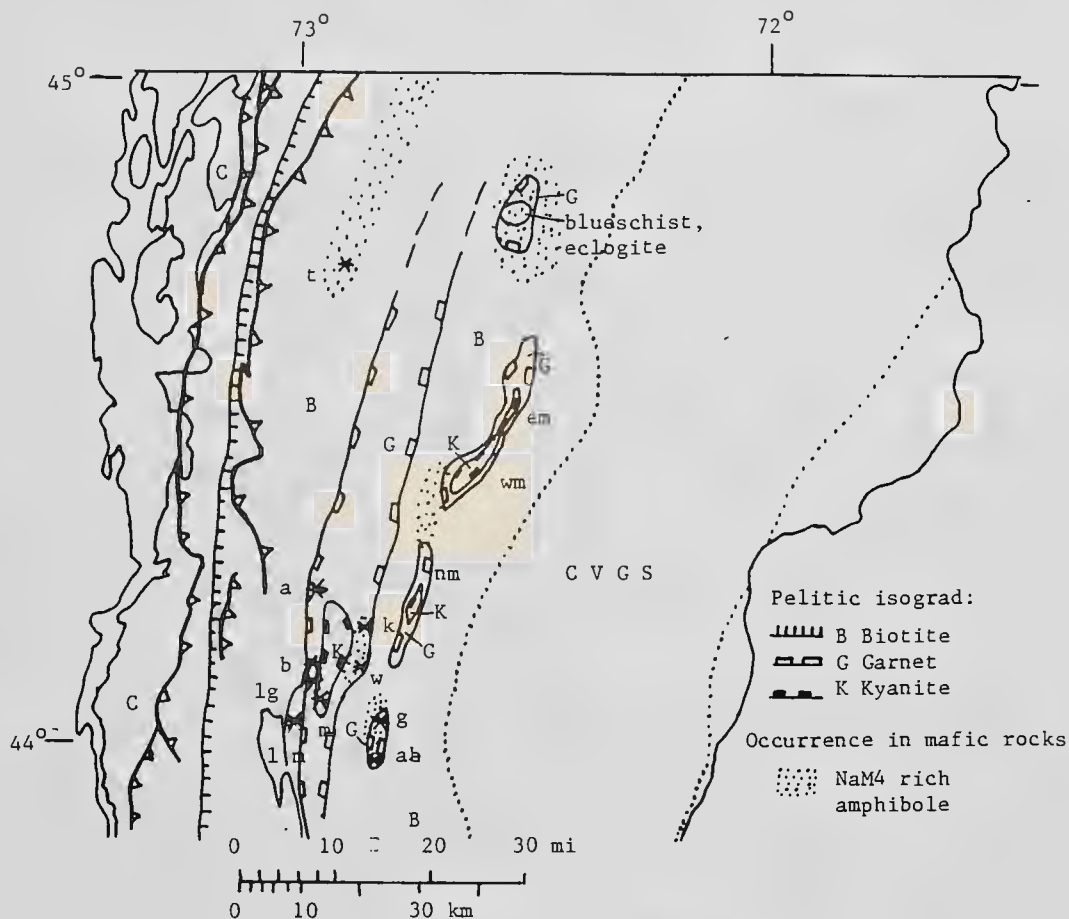


Figure 1: Metamorphism in north-central Vermont after Doll et al. (1961) and Laird et al. (1984). Stars refer to localities discussed in text: t(Tibbit Hill volcanics), a (Appalachian Gap), b (Battell greenstone), lg (Lincoln Gap road and south), lm (Lincoln Massif), m (Mount Grant), k (Kew Hill), w (Warren), g (Granville Gulf), ab (Allbee Brook). Other abbreviations: em (Elmore Mountain), wm (Worcester Mountains), nm (Northfield Mountains), CVGS (Connecticut Valley Gaspé Synclinorium). The blueschist and eclogite locality is discussed by Bothner and Laird (this volume).

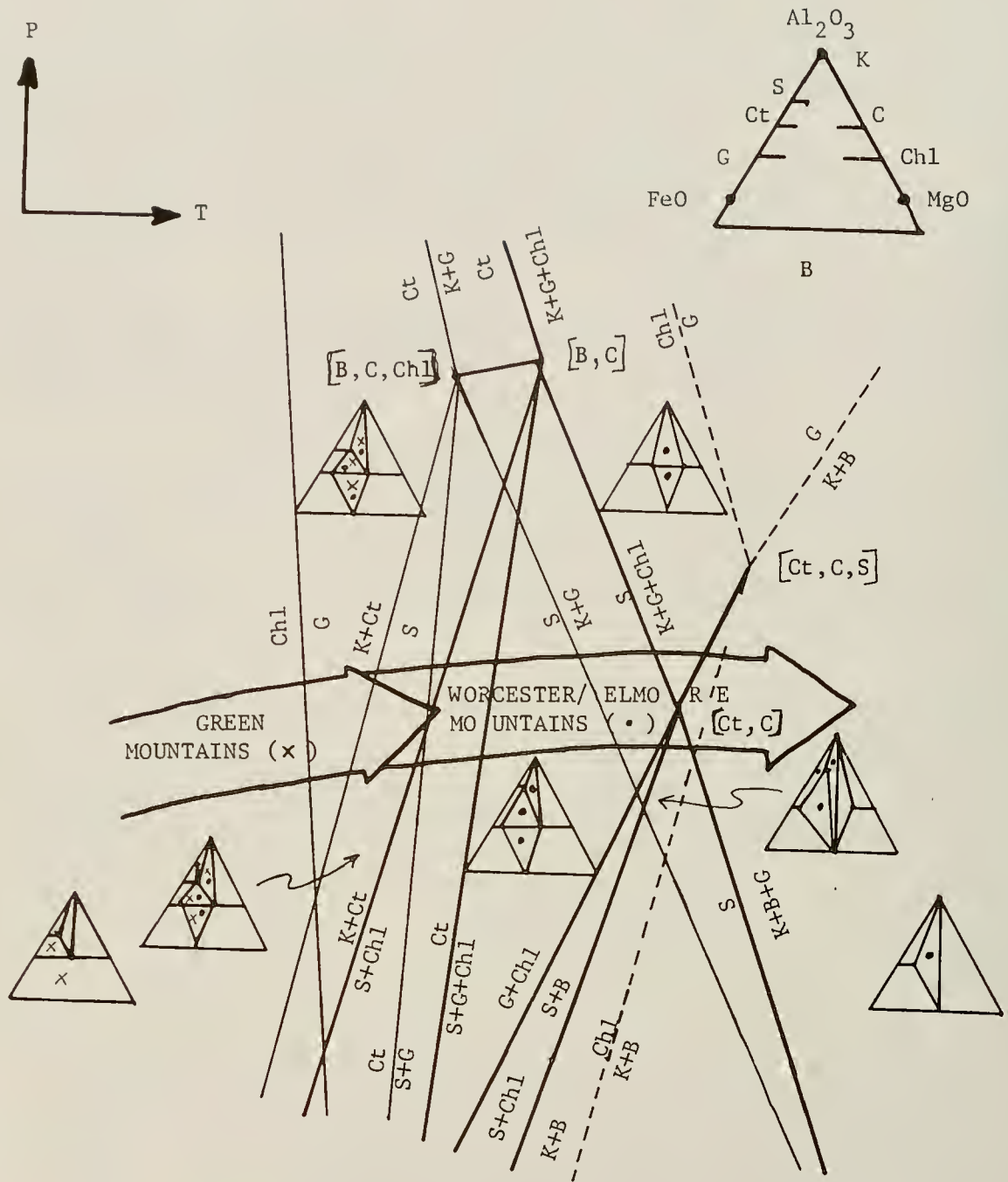


Figure 2: Mineral assemblages in pelitic schist from pre-Silurian rocks, north-central Vermont. Pressure-temperature grid is from Harte and Hudson (1979). Slopes of metamorphic reactions are generalized. The maximum grade of metamorphism in the Green Mountains west of the Green Mountain anticlinorium and in the Worcester Mountains (wm, fig. 1) is represented by the end of the appropriate arrow. Assemblages are shown on a projection from quartz and muscovite (Thompson projection). Abbreviations: B (biotite), C (cordierite), Chl (chlorite), Ct (chloritoid), G (garnet), K (kyanite), S (staurolite).

schist, however, shows very nicely that in central Vermont metamorphism was higher pressure east than west of the GMA (fig. 3). In contrast, high NaM4 in amphibole indicates that medium-high pressure facies series metamorphism is identified both east and west of the GMA in northern Vermont (fig. 1; compare sample LA426 with samples of medium-high-pressure facies series east of the GMA, fig. 3).

Mafic rocks east of the GMA have amphibole with core compositions indicating higher metamorphic grade than rim compositions. Good examples include our stop at Granville Gulf where barroisite is overgrown by actinolite (g, figs. 1 & 3) and at Elmore Mountain as described above. Because the actinolite rims grew (amphibole mode increased) the rims cannot have formed by simple cooling and must record metamorphism (Laird, 1986).

Decreasing temperature with time is also indicated in a mafic sample from the Lincoln Massif (south of Cobb Hill) by outward decreasing NaM4 and $AlVI + Fe3 + 2Ti + Cr$ in amphibole (lm, fig. 3), although Ti rich cores are suggestive of original igneous compositions. This sample records a maximum temperature of metamorphism less than the overlying Hoosac and Underhill fms. The retrogradation was reported earlier by Tauvers (1982) and may be correlative with the retrograde metamorphism documented by O'Loughlin and Stanley (1986).

In contrast, continuous zoning indicates progressive metamorphism in the Underhill Fm. west of the GMA along Lincoln Gap and Appalachian Gap roads (lg and a, figs. 1 & 3). Zoning is harder to interpret in the Hoosac Fm. (lg, EL 475) and in the Underhill Fm. south of Mt. Abraham (b) and at Castle Hill (lg, EL 66, fig. 3), in part because the zoning is irregular and may be related to miscibility.

$40Ar/39Ar$ age data on amphibole obtained by Marvin Lanphere (USGS, Menlo Park) show that Taconian age metamorphism at stations a and g (fig. 1) is 471 Ma (Laird et al., 1984). Laird et al. (1984) report 376 to 386 muscovite and biotite $40Ar/39Ar$ total fusion ages from the kyanite grade pelitic schist east of the Lincoln Massif studied by Albee (1965). Lanphere et al. (1983) also reported Devonian biotite (387 Ma) and amphibole (382 Ma) $40Ar/39Ar$ ages from the Lincoln Gap road (VJL340). Do these data mean that Acadian metamorphism is recorded east of the Lincoln Massif while Taconian metamorphism is recorded farther north and east?

Structural data presented by Stanley and his colleagues on this trip indicate a similar age of deformation and metamorphism east and west of the GMA. Sutter et al. (1985) interpret Lanphere's Devonian ages as cooling ages from a Taconian metamorphic event. Further isotopic data are needed along with our combined structural, geochemical, and petrologic studies to address this problem.

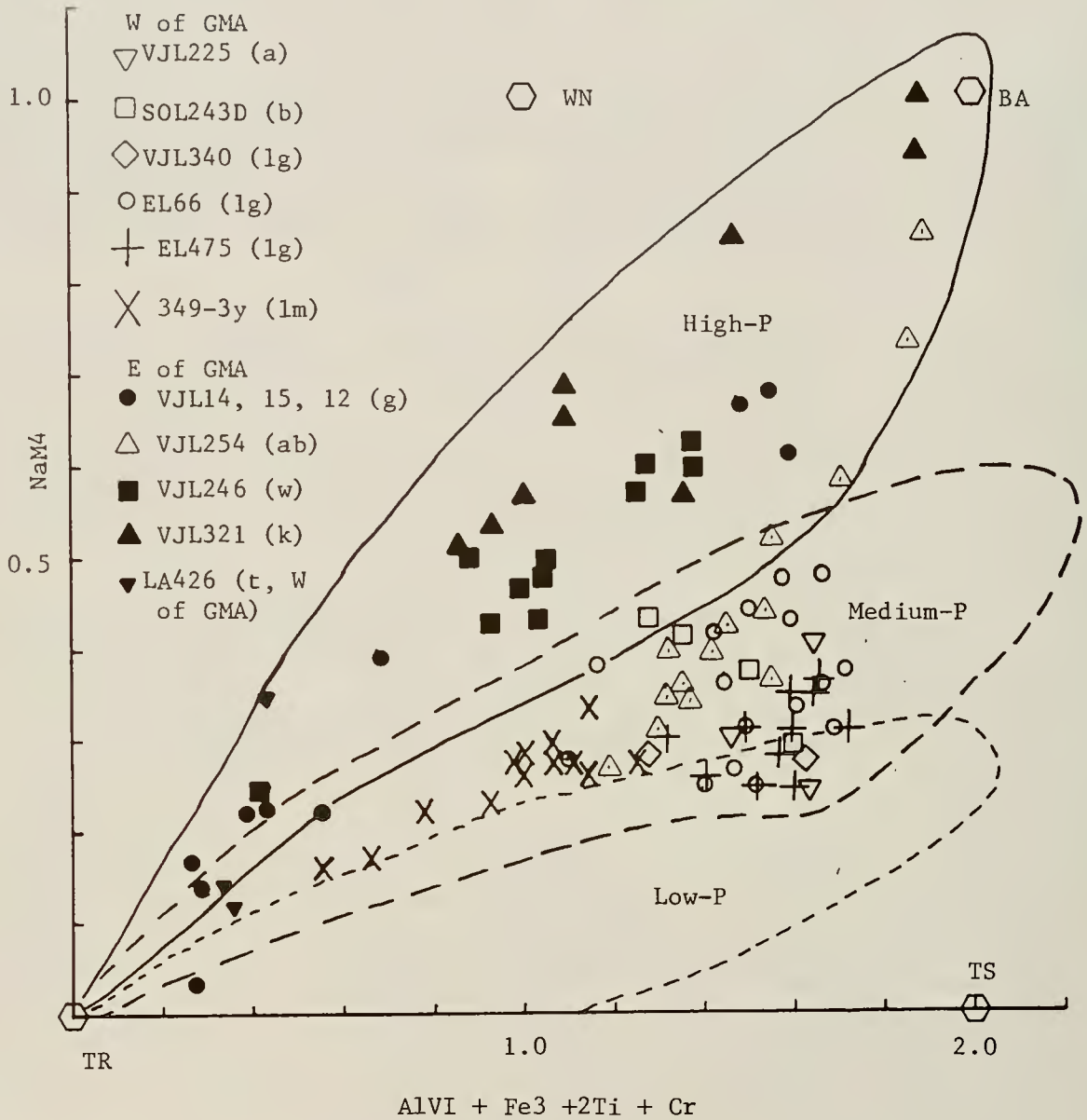


Figure 3: Formula proportion NaM₄ versus (AlVI + Fe₃ + 2Ti + Cr) for amphibole in amphibole + chlorite + epidote + plagioclase + quartz schist, central Vermont. Sample localities of VJL numbers from Laird et al. (1984). Localities in () from fig. 1. Envelopes for high-, medium-, and low-pressure facies series metamorphism are from Laird et al. (1984). Increasing temperature of metamorphism results in increasing advancement along both the X and Y axes. TR (tremolite), TS (tschermakite), WN (winchite), BA (barroisite). Analyses are electron microprobe data and normalized to total cations - (Na + K) = 13

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REGIONAL GEOCHEMICAL VARIATIONS IN GREENSTONES FROM THE CENTRAL VERMONT APPALACHIANS

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INTRODUCTION

Greenstones from the Mount Holly Complex, the Pinnacle, Underhill, Hazens Notch, Pinney Hollow, Ottauquechee and Stowe formations have been analyzed for their major, trace and rare earth element abundances. This is part of a long term project by students at Middlebury College begun seven years ago. I present here a summary of regional geochemical trends and discuss briefly their tectonic implications. Some of the ideas presented here are from Coish et al. (1985; 1986).

GREENSTONE GEOCHEMISTRY

Mount Holly Complex, Lincoln Massif

Three mafic dikes cut the Lincoln massif, just north of Ripton, Vermont (Delorusso and Stanley, 1986; Filosof, 1986; Stanley et al., this volume, **Stop 3**). The dikes are intrusive into felsic gneiss; one dike contains inclusions of the gneiss, and also has fine-grained apophyses chilled against the gneiss. The age of the dikes is deduced from the stratigraphy. The dikes cut the gneiss but do not intrude the overlying Pinnacle Formation, so they are pre-Pinnacle in age (Delorusso and Stanley, 1986). The dikes are near vertical and their strike ranges from N60E to about East. The dikes are 5 to 8 meters thick and all have a well-developed foliation. All three dikes contain plagioclase feldspar, biotite, actinolite, chlorite and magnetite. Compared to other Vermont greenstones, the dikes have a higher abundance of biotite. One dike has phenocrysts of plagioclase whereas another is biotite-phyric.

Geochemically, the dikes are basalts. They have high Ti, Zr, Y, P and rare earth element (REE) concentrations (Figs. 1, 2 & 3), and can be further classified as transitional basalts. Compared to greenstones in overlying formations, they have many chemical similarities; however, they do have higher K, and lower Ti/P and Ti/Y ratios. These differences may indicate that the dikes have undergone some contamination with continental crust. The dikes near Ripton are similar to basaltic rocks of Precambrian age in the Catoclin Formation in Virginia, and the Lighthouse Cove Formation in Newfoundland. The chemistry of the Ripton dikes, like the basalts from Virginia and Newfoundland, is consistent with their formation during rifting of the proto-North American continent during the late Precambrian.

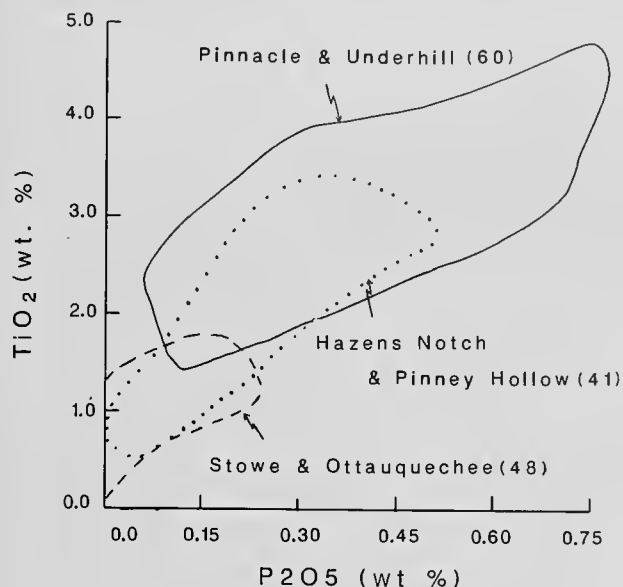


Figure 1. Variation of TiO_2 vs P_2O_5 in Vermont greenstones. # analyses in brackets.

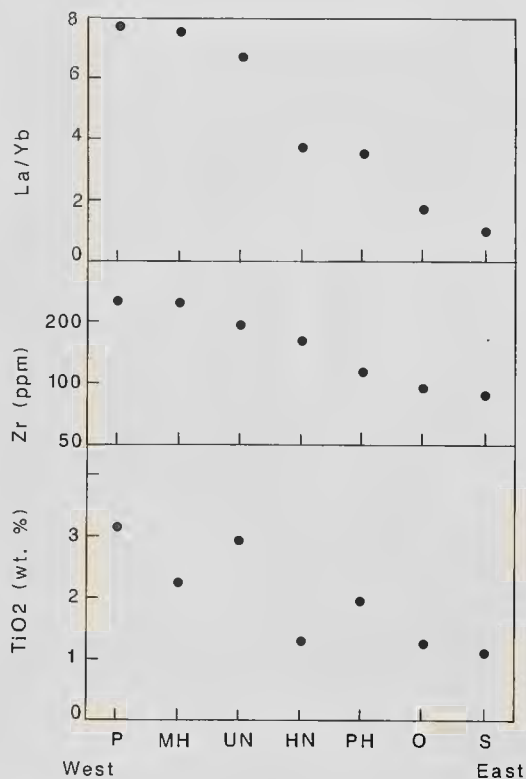


Figure 2. Average concentration of Ti, Zr & La/Yb in all units.

Pinnacle and Underhill Formations

Greenstones from the Pinnacle and Underhill Formations are grouped because they are geochemically very similar, and thus, probably have the same origin. Greenstones from the Pinnacle Formation have been sampled from the Tibbit Hill metavolcanics in northern Vermont (Fleming, 1980). Greenstones from the Underhill Formation are from two localities near Huntington, Vermont (Larsen, 1984; Seibert, 1984), and from six localities in the vicinity of South Lincoln (Gavigan, 1986). The Tibbit Hill outcrops are part of a more extensive unit in southern Quebec; pillow lavas are common, and relict volcanic textures are seen in thin section. These samples are interbedded with metagraywacke, including conglomeratic units, arkose and phyllite. The Huntington greenstones are coarse-grained, plagioclase-amphibole rocks that may be meta-gabbros. Greenstones near Lincoln are mostly plagioclase, chlorite, actinolite, epidote rocks. One outcrop is a 3-meter wide dike cutting pelitic schist. All greenstones from the Pinnacle and Underhill formations are in the greenschist metamorphic facies, and all contain abundant magnetite.

The chemistry of greenstones from the Pinnacle and Underhill formations indicates the samples were transitional basalts with very high concentrations of Ti, Zr, Y, and the rare earth elements (Figs. 1, 2 & 3). Furthermore, the rare earth element patterns are fractionated such that the light rare earth elements (LREE) are enriched relative to the heavy rare earth elements (HREE) (Fig. 3). These geochemical features are typical of basalts produced in early stages of continental rifting. Although there is variation in chemical

B-8
C-6

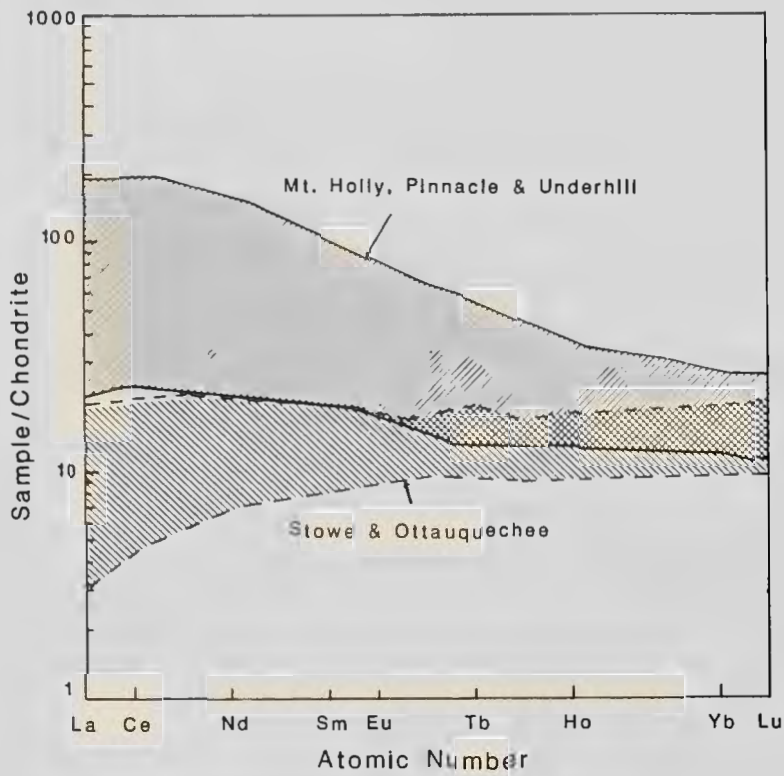


Figure 3. Rare earth element concentrations in Vermont greenstones. Stippled areas represent the range of values for formations indicated.

composition among all the samples, there is no clear difference between the Pinnacle & Underhill greenstones. This indicates that the Pinnacle and Underhill greenstones were basalts generated from the same type of mantle source rock. There could have been, and probably were, many different magmatic episodes over a wide range of time from this same mantle source.

Hazens Notch and Pinney Hollow Formations

Greenstones from the Hazens Notch Formation were sampled north and east of Lincoln Gap (Masinter, 1986). Greenstone samples from the Pinney Hollow Formation were taken from four outcrops along route 100, between Granville and Waitsfield (Masinter, 1986), and from a single outcrop along route 125, near Hancock (Poyner, 1980). Greenstones from these two formations were grouped because of their overall geochemical similarity, and their differences from greenstones to the west and east. Some of the greenstones in these formations are indistinguishable in the field from greenstones in the Pinnacle and Underhill formations in that they are dark green, chlorite-actinolite rocks. Other greenstones contain much epidote and thus are much lighter green, similar in fact to greenstones found in the Stowe Formation. Thus, the Hazens Notch and Pinney Hollow formations contain two types of greenstones. At present, we see no systematic distribution of these two types.

Geochemically, the greenstones are high Ti, Zr, Y and REE basalts; concentrations of these elements overlap with the low end of the envelope for the Pinnacle and Underhill greenstones and extend to lower values (Fig. 1 & 3). The greenstones with lower values overlap greenstones from the Stowe Formation. Samples from one particular body exhibit nearly the entire chemical range shown by the whole group. This suggests either 1) magmas erupted in a single place were generated at different depths in the mantle or from different parts of a single magma chamber, or 2) the different basalts were generated in disparate regions and intimately mixed later by faulting. Whatever the reason, it is clear that the Hazens Notch and Pinney Hollow formations contain basaltic greenstones with a wide range of chemical compositions.

Ottawaquechee and Stowe Formations

Greenstones from the Ottawaquechee Formation were sampled near route 100 between Waitsfield and Waterbury (Doolittle, 1987). Greenstone samples of the Stowe Formation were taken from three large bodies: near Braintree, Vermont (Bailey, 1981); near Moretown (Anderson, 1983); and near Waterbury Center (Perry, 1983). All greenstones are fine-grained, light-green rocks distinct from the darker greenstones to the west. The greenstones are in the greenschist facies; a high abundance of epidote results in the light green color. Many of the greenstones exhibit a distinct mineralogical layering with bands of albite & quartz alternating with bands of epidote, chlorite & actinolite.

Greenstones from the Ottawaquechee Formation are indistinguishable chemically from greenstones from the Stowe Formation. Samples from both formations are basaltic in composition and have lower concentrations of Ti, Zr, Y and REE than most greenstones from more westerly formations (Fig. 1, 2 & 3). Also, the rare earth element patterns show a slight to moderate depletion in the LREE relative to the HREE (Fig. 3), contrasting with the LREE-enriched patterns of the western formations. In almost all geochemical features, the Stowe and Ottawaquechee greenstones are similar to modern mid-ocean ridge basalts.

The depleted LREE pattern indicates that the greenstones must have been derived from a different mantle source than the western greenstones; in particular, the mantle source must have been depleted mantle. On all counts, the Stowe and Ottauquechee greenstones are different from **all** greenstones in the Pinnacle and Underhill formations and **many** greenstones in the Pinney Hollow and Hazens Notch formations.

TECTONIC IMPLICATIONS

Geochemical trends in modern basaltic rocks have been used to place volcanic rocks in a particular tectonic environment, i.e. mid-ocean ridge, subduction zone, or within-plate environment. Applying the same technique to ancient rocks, it can be shown that greenstones from the Mount Holly Complex, Pinnacle and Underhill formations formed in a within-plate environment. Furthermore, these greenstones appear to have been formed within a **continental** plate. Most modern within-continental plate volcanism occurs in regions where a plate is being stretched and pulled apart to form a rift valley. The tectonic environment of formation of the Ottauquechee and Stowe greenstones appears to be an ocean ridge. Greenstones from the Hazens Notch and Pinney Hollow formations show chemical characteristics of both within-plate **and** ocean ridge environments.

Before an attempt is made to describe tectonic events in the late Precambrian, it is instructive to look at the geographical distribution of some important elements in the greenstones (Fig. 2). In figure 2, the various formations are arranged in their current spatial distribution from west to east, and the average concentrations of TiO_2 , Zr and La/Yb (a measure of LREE enrichment) are plotted. There is an obvious decreasing trend from west to east. This trend can be attributed to a change in the environment of formation of the volcanic rocks either with time if the rocks young toward the east or with distance from a rift axis if the greenstones are all approximately the same age. Whatever the interpretation, it seems that the present arrangement of the formations roughly reflects the arrangement at the time of formation; amazingly, things are not completely scrambled. It is certainly possible and perhaps likely that considerable lateral shortening has occurred, but the geochemistry of the greenstones tells us that the formations have **not** been well shuffled.

The geochemistry of greenstones in central Vermont can be explained by a fairly simple model of splitting of a continent and formation of an ocean basin. During the late Precambrian (~ 650? m.y.a), the Adirondack continent began to stretch and pull apart presumably due to a thermal plume in the mantle below. Volcanism accompanied this stretching resulting in the formation of basalts in the Mount Holly, Pinnacle and Underhill formations. The continental crust at this time was distended but still fairly thick; this can be called an early stage of rifting. Slightly later (~600? m.y.a.), volcanism continues to produce basalts further east as the zone of magmatism migrates. The continental crust is now more stretched and an ocean basin is starting to develop. This is a transitional stage in the development of ocean crust; the basalts produced here will have transitional features between true within-plate basalts and ocean ridge basalts. The basalts erupted at this time become part of the Hazens Notch and Pinney Hollow formations. Still later (~550? m.y.a.), the continental crust has been completely stretched and true sub-oceanic mantle established; ocean ridge type basalts are erupted farther east and will later become part of the Ottauquechee and Stowe formations. The ocean basin at this time could have been very small. Eventual closure of the ocean basin during the mid-Ordovician results in metamorphism of all the basalts to greenstones.

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CLEAVAGE VS. FOLDING VS. THRUSTING:
RELATIVE TIMING OF STRUCTURAL EVENTS
IN THE CENTRAL CHAMPLAIN VALLEY

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INTRODUCTION

The gross structure of a thrust belt (e.g. western Vermont) is controlled by the geometries of the thrust surfaces and the magnitudes of the displacements (Boyer and Elliott, 1982; Suppe, 1983). Within every thrust belt, however, there are numerous other structures that form during the same tectonic event but are not obviously related to the thrusting. It is the temporal and spatial relations between these secondary structures and the thrusting that this field trip will investigate.

Regional Setting

The structure of the central Champlain Valley has long been thought to consist of a narrow line of thrusts separating undeformed strata to the west from multiply folded strata within a recumbent synclorium to the east (Cady, 1945). Recent mapping (Washington, 1981a, b, 1982, 1987b; Washington and Chisick, 1987; Harding and Hartz, 1987), however, has found that the area is a complex thrust belt (figure 1). In fact, the areal distribution of the strata that led to the synclorium theory (Dana, 1877; Cady, 1945) is the result of successively older strata being brought to the surface by the thrusts east of the "axis of the synclorium."

The Western Vermont thrust belt consists of four major thrust systems: the carbonate thrust sheets (herein called the Middlebury thrust system) at the base, the Champlain thrust sheet above, the New Haven-Green Mountain (NH-GM) thrust system overriding the trailing edge of both of these, and the Taconic allochthons to the south sitting atop the Middlebury and NH-GM systems. As Keith (1932) recognized, each of these thrust systems has a distinct stratigraphy. Recent work is redefining the correlations among the sequences (Washington and Chisick, 1987). Figure 2 summarizes the stratigraphy of the thrust belt and compares it with the autochthonous strata along the edges of the Adirondacks. This field trip will concentrate on the

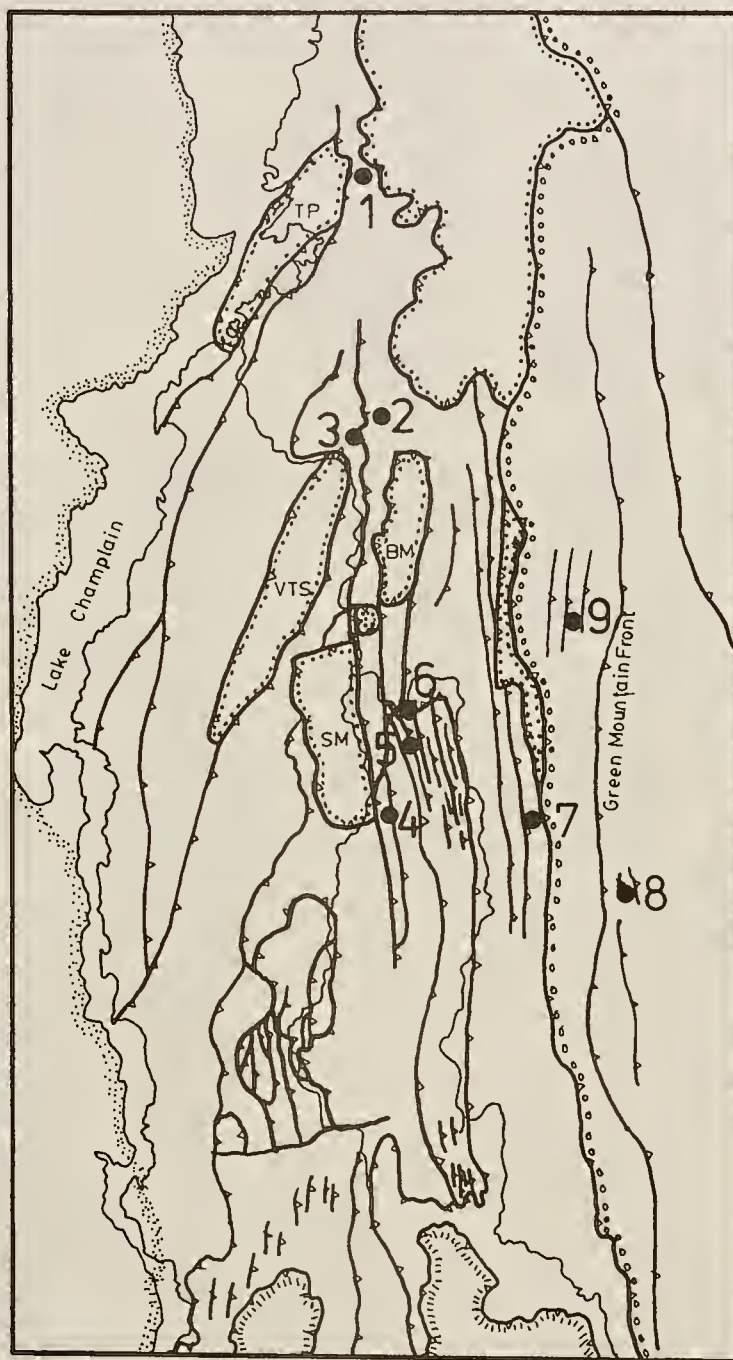


Figure 1 - Structure map of the central Champlain Valley. Autochthon extends west from fine stipple, New Haven - Green Mountain thrust system extends east from line of open circles, north end of Taconic allochthons are denoted by hatchures, and edges of Champlain thrust sheet (including isolated pieces) are marked by heavy dots; intervening area is the Middlebury thrust system. TP - Thompson's Point; VTS - Vergennes thrust sheet; BM - Buck Mountain; SM - Snake Mountain; field trip stops indicated by numbers.

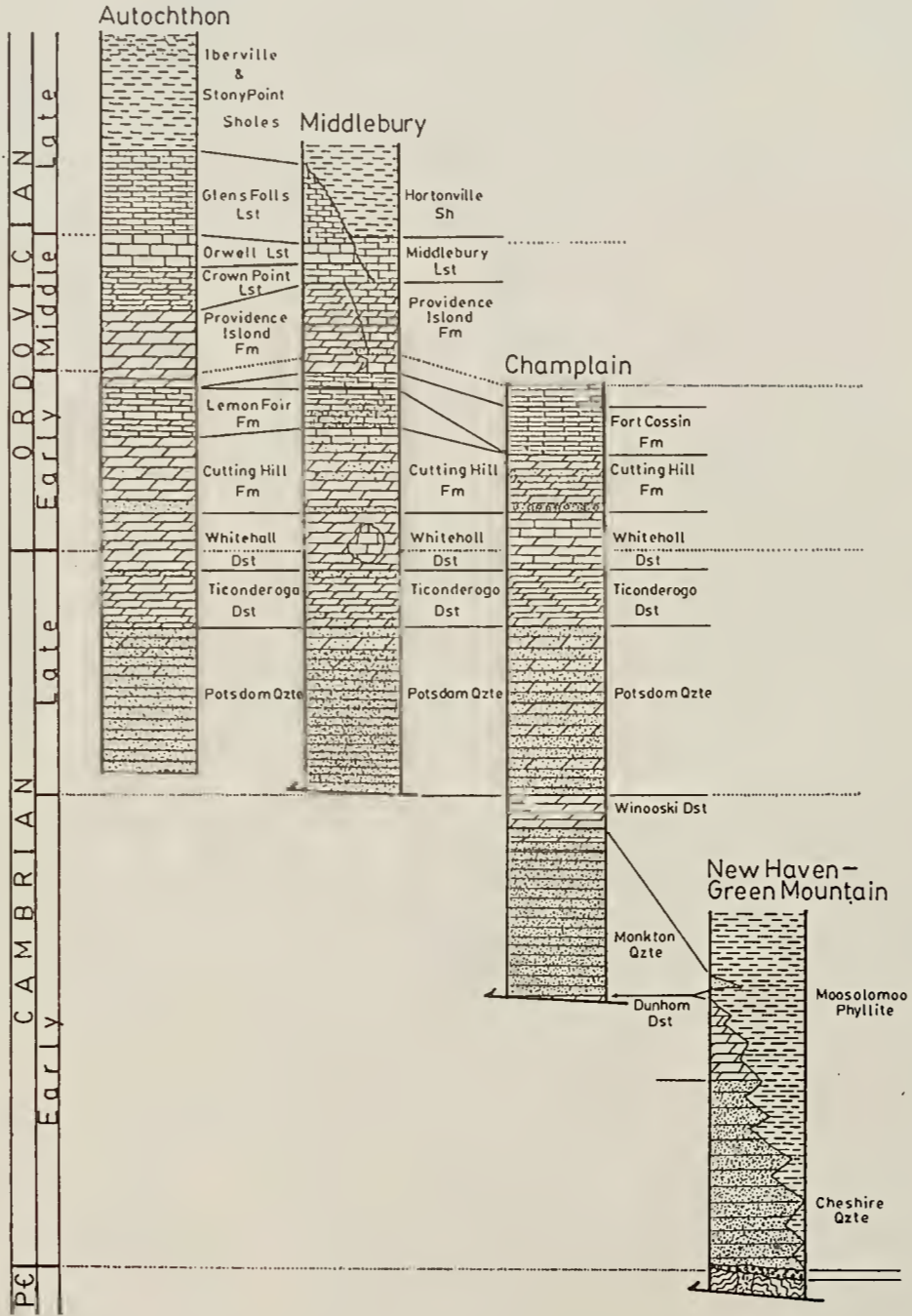


Figure 2 - Comparison and correlation of the stratigraphic packages encountered in the autochthon and the Middlebury, Champlain, and New Haven - Green Mountain thrust systems in the central Champlain Valley.

Middlebury thrust system, but STOPS 8 and 9 are within the New Haven-Green Mountain thrust system.

One of the key elements in the structure of this thrust belt is penetration thrusting. A penetration thrust breaches an overlying thrust surface and places structurally lower material onto a portion of the structurally higher sheet (Washington, 1985). The present positions of the western pieces of the Champlain thrust sheet (i.e. Snake Mountain, Buck Mountain, the Vergennes thrust sheet, and Thompson's Point) has led most prior workers (Logan, 1860; Dana, 1877; Cady, 1945; Coney and others, 1972) to infer that the Champlain thrust lies beneath the "Middlebury synclinorium." Actually, these portions of the Champlain thrust sheet are tectonically isolated by penetrating thrusts of the Middlebury thrust system. This will be seen at STOP 4 where the Weybridge thrust sheet overlies the trailing edge of the Snake Mountain portion of the structurally higher Champlain thrust sheet.

There is a high metamorphic gradient within the eastern parts of the Middlebury thrust system. This is produced by the juxtaposition of greenschist metamorphics of the NH-GM thrust system against carbonates of the Middlebury thrust system. The north end of the Vermont marble belt lies about the latitude of New Haven because north of there the trailing edge of the Champlain thrust sheet intervenes between the metamorphics and the carbonates. STOP 7 is the Middlebury marble quarry where metamorphic effects are superimposed on thrust-belt structures, obscuring the detailed structural relations.

Thrust Systems and Related Structures

Thrust surfaces are generally stepped, with alternating flats (segments parallel to layering) and ramps (segments oblique to layering) (Rich, 1934; Dahlstrom, 1970; Boyer and Elliott, 1982). Ramps can be perpendicular, oblique, or parallel to transport (see Wilson and Stearns, 1958, Dahlstrom, 1970, and Elliott and Johnson, 1980). Displacement over the steps produces "fault-bend" folds (Suppe, 1983) which are generally large, flat-topped, and asymmetric. At STOP 1 we will see some nice examples of such folds in small thrust systems.

There are two basic types of thrust systems - duplexes and imbricate fans (Boyer and Elliott, 1982). Duplexes are thrust systems in which the various thrusts branch off of a common basal (floor) thrust and join a common overlying (roof) thrust. Imbricate fans, on the other hand, are thrust systems in which the various thrusts branch off of a common basal thrust and climb up through the overlying

strata until they either reach the surface or go blind (end without connecting to the surface or another thrust). The Champlain thrust system is an emergent imbricate fan; the structurally lower Middlebury thrust system, however, is a duplex for which the Champlain thrust (and Taconic thrust to the south) was the roof. The tectonic dismembering of the Champlain thrust sheet occurred when imbricate thrusts of the Middlebury system breached the Champlain thrust rather than joining it.

Generally, fault-bend folds are the last major structural element to develop in a thrust sheet unless there are multiple tectonic events or out-of-sequence deformation (Coward, 1984). Because thrusting progresses downward and forelandward, deformation in lower (later) thrust sheets can affect higher (earlier) thrust sheets (Elliott and Johnson, 1980; Mitra and Yonkee, 1985). Where the lower thrust sheets are exposed, however, the source of this deformation is usually clear. The only other type of deformation that does not pre-date fault-bend folding is "ramp-bend" folding which will be discussed below.

Adjacent to thrust surfaces there are commonly zones of higher deformation. In brittle rocks these zones are marked by cataclasis, whereas in ductile rocks they are generally marked by drag folding and cleavage. The deformation within these zones diminishes in intensity with distance from the fault surface. Examples of this deformation will be seen at STOPS 4 and 5. Some thrusts do not develop noticeable zones of deformation, however, all of the displacement apparently being confined to the fault surface (e.g. STOP 2). Where conditions are just right, secondary thrust systems can form immediately below major thrusts. If the major thrust causes a great increase in overburden, cleavage can form within the secondary thrust systems after fault-bend folding (e.g. STOP 1).

Secondary Structures

There are three structures which are generally associated with thrusting: folds, cleavage, and joints. All of these record shortening of the material; the direction of shortening is roughly perpendicular to the fold axes and cleavage planes and perpendicular and parallel to the two orthogonal joint sets. Where these structures are penetrative (i.e. developed throughout a thrust sheet) they form prior to thrust displacement (Washington, 1987a); only localized structures form during displacement. Regional joints are most evident in competent strata, whereas folds and/or cleavage are dominant in ductile strata. In the Champlain Valley, most of the strata fall within the ductile range, so cleavage and folds are common.

As Henderson and others (1986) recently pointed out, the relative timing of cleavage and fold development is a function of the material properties. Since folding is strongly controlled by layer thickness and layer-boundary strength (Kuenen and de Sitter, 1938; Ramsay, 1967) while cleavage formation is primarily controlled by ductility and stress magnitude (Hobbs and others, 1976), folds will develop first if the material is thinly bedded and stiff and cleavage will develop first if the material is massive and ductile. Cleavage and folds can be coeval if conditions are intermediate between these two extremes.

Where cleavage forms first, subsequent folding will reorient the cleavage, creating a cleavage fan (Fisher and Coward, 1982). Where folds form first, cleavage orientation is consistent throughout (Henderson and others, 1986). The relative orientations of the fold axial surfaces and the cleavage planes indicates the relative orientations of the stress field during the formation of the two structural elements.

Although folds can only form where there has been some shortening parallel to layering, cleavage can form even where shortening is nearly perpendicular to layering. The orientation of cleavage is a good indicator of the environment in which it formed. Cleavages that are nearly perpendicular to layering (e.g. STOPS 2 and 8) are generally formed directly ahead of a step in an active thrust. Cleavages that are nearly parallel to layering (e.g. s_1 at STOP 5), on the other hand, are generally formed beneath a thick thrust sheet advancing across a flat; the increased load causes the shortening and the taper toward the foreland causes the deviation from horizontal.

The morphology of the cleavage is primarily a function of material properties (Hobbs and others, 1976; Cosgrove, 1976; Engelder and Geiser, 1979): a fine-grained rock will form a slaty cleavage (especially if it is monomineralic), an impure sandstone or limestone will form a spaced cleavage, and a strongly anisotropic rock will tend to form a crenulation cleavage (unless the stress is applied in the wrong orientation). Thus, in the Champlain Valley the shales and pure micritic limestones exhibit slaty cleavage (STOPS 1, 4, 5, 6, and 9); the impure limestones and sandstones contain spaced cleavage (STOPS 2, 3, 4, and 8), and second generation cleavages in the limestones and slates are crenulation cleavages (STOPS 5 and 6).

Ramp-Bend Fold Trains

Although folds and cleavage generally form prior to thrust displacement, small areas contain trains of late

folds, often accompanied by cleavage. These anomalous areas always occur at the junction between two differently oriented fault-bend folds. The folds are generally oriented somewhat oblique to regional strike and plunge rather noticeably toward the more hindeward end. Due to their peculiar location and timing, these fold trains are attributed to the room problems above ramp intersections during the fault-bend folding process; thus the term "ramp-bend fold trains" is applied. STOP 6 is an excellent example of a ramp-bend fold train.

Fault-bend folds form above hangingwall ramps as they advance up stepped thrust surfaces from their original seats (Suppe, 1983). The orientation of the hangingwall ramp has little bearing on the shape of the fault-bend fold since the shape is dictated by the necessity to maintain surface length and volume in the thrust sheet while it remains in continuous contact with the underlying fault surface. When two differently oriented ramps meet, the folding process becomes more complex; the theoretical fault-bend fold shape is not possible within the zone of overlap because it would necessitate drastically decreasing the surface area. Since the fault-bend folds must be accommodated, however, an effective reduction in surface area must occur. This is accomplished by shortening the layers with short-wavelength folds (fig. 3), often with accompanying cleavage.

The amount of shortening varies within the overlap zone, the greatest occurring in the central portion decreasing to zero towards the edges, especially the edge toward the more frontal ramp. Where the ramp-bend is created by the junction of a frontal and an oblique (or lateral) ramp, the shortening commonly extends along the oblique ramp farther than might be expected, making quantification of the strain produced by various ramp-bend geometries of limited

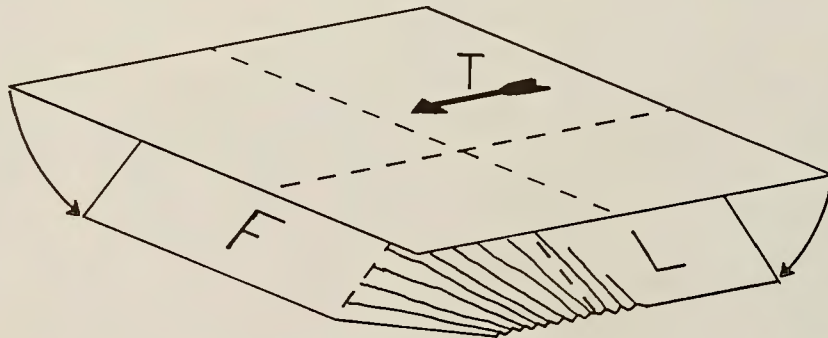


Figure 3 - Ramp-bend folds accommodate surface area decrease in zone of overlap between frontal (F) and lateral (L) fault-bend folds. T indicates transport direction.

use. This hindeward extension of the fold trains is probably due to drag along the tapered edges of the sheets.

Although there are many different possible fold patterns that could accommodate the diminution of the surface area within the overlap, the folds always have a trend between that of the two fault-bend folds. This consistency of orientation is not surprising, however; the stress field in an active thrust sheet would force the folds to be oriented tangential to the edge of the thrust sheet. The slight rotation of the folds toward the orientation of the more hindeward ramp can be attributed to the aforementioned drag on the tapered edge.

Discussion

From the preceding review of thrust belt structures, it is obvious that the relative timing of structural elements in a thrust belt is not constant. Rather, the relative timing is determined by material properties and the tectonic environment. To further complicate the picture, each thrust sheet should be considered separately since the tectonic history of each is slightly different.

Fortunately, there are many clues to the origins of the various structural elements. The orientation of a cleavage and its relation to accompanying folds gives some indication of the orientation and history of the stress field and the character of the material. The areal distribution of these elements provide further clues to the gross structure of the thrust belt.

In the Champlain Valley there are several sets of cleavage and folds (although prior workers [e.g. Crosby, 1963; Voight, 1965, 1972; Coney and others, 1972] have tended to lump them together). Generally the folds accompany cleavages formed prior to the cleavage, indicating that layer thickness and boundary strength were sufficiently low. By using the distribution of these elements, thrusts can be identified even when they are not exposed and detailed structural relations can be deciphered. Thus, the secondary structures are providing valuable insights into the structural development of western Vermont.

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Figure 4 - Route of field trip in western Vermont.

ITINERARY

Assembly point is the gas station (currently BP) at the intersection of Routes 7 and F-5 in Charlotte.
Time: 9:00 A.M.

Mileage

- 0.0 Go south on Route 7.
- 0.4 STOP 1: Charlotte Roadcuts. These roadcuts provide an excellent view of thrust structures developed in the shales directly beneath the Champlain thrust (Champlain thrust lies near top of hill). Thrust surfaces, fault-bend folds, and slaty cleavage can be seen in this roadcut. Note that the cleavage post-dates thrusting and fault-bend folding. The orientation of this cleavage indicates it formed in response to the imposition of a large overburden, probably the Champlain thrust sheet after it overrode these shales. The thrusts developed during movement on the Champlain thrust. This is the south end of Stanley's (1969) stop 5.
- Continue south on Route 7.
- 2.1 Hill to left is Mount Philo. Cliffs near top are Monkton quartzite of the Champlain thrust sheet. Thrust lies about at base of cliffs.
- 9.0 Turn left onto Satterly Road.
- 9.2 STOP 2: Hammond Quarry. North face of this quarry contains a small thrust in Glens Falls limestone. Note the strong spaced cleavage and detachment surfaces in the footwall. The cleavage and detachments are basically coeval. Both formed in response to horizontal tectonic compression during active movement on the overlying thrust (Washington, 1987).
- Turn around and proceed back to Route 7.
- 9.4 Turn left and continue south on route 7.
- 9.9 Turn right onto Route 22-A. Stop as soon as you are clear of the corner.
- 10.0 STOP 3: Vergennes Roadcut. The low roadcut along the south side of Route 22-A has well developed folds and cleavage. The cleavage is clearly unaffected by the folds, indicating that it formed later. The cleavage is parallel to the axial planes of the

folds, however, so the structures are probably penecontemporaneous.

Continue south on Route 22-A.

- 11.4 The Vergennes falls are the last falls on Otter Creek. The falls consist of Whitehall dolostone of the Vergennes thrust sheet (a tectonically isolated piece of the Champlain thrust sheet) with the thrust plane reaching the surface about the base of the falls. Otter Creek is navigable from here to Lake Champlain, so this is where Commodore Macdonough built his fleet that defeated the British fleet at the Battle of Plattsburgh in 1814. At the end of Otter Creek is the site of Fort Cassin (built to protect the fleet during construction), the type locality of the Cassinian stage (Early Ordovician).

For the next few miles Route 22-A follows the ridge formed by the Vergennes thrust sheet. The low outcrops along the road are Whitehall and Ticonderoga dolostone. The broad valley to the west is underlain by Late Ordovician shales.

- 16.8 Ahead and to the left is Snake Mountain, the first American locality to be recognized as having non-sequential stratigraphic stacking. The occurrence of the Early Cambrian Monkton quartzite atop Late Ordovician limestones and shales caused considerable controversy (see Emmons, 1842, and Adams, 1848) which was resolved only when Logan (1860) defined the Logan's line thrusts with this as the type locality. The Champlain thrust lies about at the base of the upper cliffs, which are Monkton quartzite.
- 24.8 Turn left onto Route 125.
- 27.8 The south end of Snake Mountain. This is the south end of the Champlain thrust as mapped by Cady (1945) and recorded on the state geologic map. All strata for the next 10 km south are of Ordovician age.
- 28.4 Lemon Fair River.
- 29.5 The cliffs that face you are the hangingwall of the Weybridge thrust, a penetration thrust which places Ordovician strata over the lower Cambrian strata of the Snake Mountain portion of the Champlain thrust sheet. Unfortunately, the thrust surface is not exposed.

30.4 STOP 4: The Ledges. These ledges provide one of the best exposures of the upper Beekmantown Group in the Group in the Champlain Valley. The limestones at the base are Lemon Fair formation (this is the type locality) and uppermost Cutting Hill formation. The shaly strata are Fort Cassin formation and the overlying limestones and dolostones are Providence Island formation. The Providence Island strata are highly deformed, whereas the underlying strata are basically undeformed; the floor thrust of the Sudbury duplex lies just above the base of the Providence Island formation. Note the decrease in cleavage intensity with distance below the Sudbury floor thrust (a true detachment). There is also some drag folding immediately below the detachment surface.

Continue uphill on Route 125.

31.0 Turn left onto Samson Road.

32.2 Bear right onto dirt road. The ledges to the north and south of this intersection contain folded and faulted Lemon Fair strata. The folds and faults are part of a ramp-bend fold train associated with a lateral ramp along the Weybridge thrust.

32.7 Lateral cutoff of the top of the Lemon Fair formation lies to left.

32.9 Lunch. Bittersweet Falls. The falls consist of Middlebury limestone, the contact with the overlying Hortonville slate occurring just above the top of the falls. Extending east from the top of the falls is a winding canyon providing a continuous exposure through the slate that forms the "core of the Middlebury synclinorium." The eastern boundary of the slate is a thrust fault (one of the imbricates of the Sudbury duplex) and several other small thrusts can be seen within the slate. Two generations of cleavage can be identified locally in the slate: 1) a nearly bedding-parallel cleavage (axial planar to folds near the top of the section) and 2) a crenulation cleavage. The former formed in response to the superposition of the Champlain thrust sheet (which formed drag folds in the upper strata). The latter formed in response to shear strains near thrusts formed during duplexing. The limestones at the falls only show the former of these cleavages.

After lunch continue north on road.

33.5 Turn left.

- 33.7 Turn right onto Route 23 (paved road). The monument at corner is to Silas Wright, a Governor of New York who came from this town (Weybridge).
- 34.6 STOP 5: Dubois Quarries. There are two small limestone quarries to be visited here, one on the northwest side of the barn and the other about 150 meters west on the south end of a hillock. The first of these is an excellent exposure of the relations between bedding and two cleavages. As with most of the strata in the duplex, these limestones are basically unfolded. The earlier cleavage cuts across bedding at a very low angle, dipping slightly to the east. The second cleavage is a crenulation cleavage which deforms the first cleavage planes. The floor of this quarry is essentially the surface of one of the larger displacement imbricate thrusts in the duplex. The exposed strata are the basal portion of the Middlebury limestone. Crenulation cleavage forms in lower part of the thrust sheet (here through the entire quarry) in response to the shear along the fault plane.

The other quarry contains upper Middlebury limestone with a small shear zone along the top (top of quarry face is approximately top of shear zone). This shear zone is also marked by crenulation cleavage. Whether this shear zone represents a thrust is not clear.

These quarries were opened to provide stone for the bridge in the center of Middlebury. Stone from the first quarry can be seen in the bridge abutments above street level. The quarries were abandoned prematurely after a boiler on a quarry machine exploded under one of the workers.

Turn around and proceed back northwest on Route 23.

- 35.5 Turn right onto Hamilton Road (dirt).
- 36.0 STOP 6: James Pasture. The pasture to the north of this road is one of the classic structure localities in the Middlebury area (see Crosby, 1963, and Coney and others, 1972). It contains an easily mapped fold train with an axial-plane cleavage. There is also a later crenulation cleavage. The south and west sides of this knoll coincide with the Sudbury thrust (the eastern part of the roof thrust of the duplex. The folds seen here are part of a ramp-bend fold train formed by the intersection of the frontal and lateral ramp of the Sudbury thrust along the northern edge of the duplex (the Sudbury nappe marks the southern end of the duplex). Note the southeasterly plunge of the

fold axes. These folds die out a short distance to the north. The boundary between the Middlebury limestone and Providence Island formation is marked by a thin shale layer, the marker bed used by Crosby (1963) to define the fold train.

Continue east on Hamilton Road.

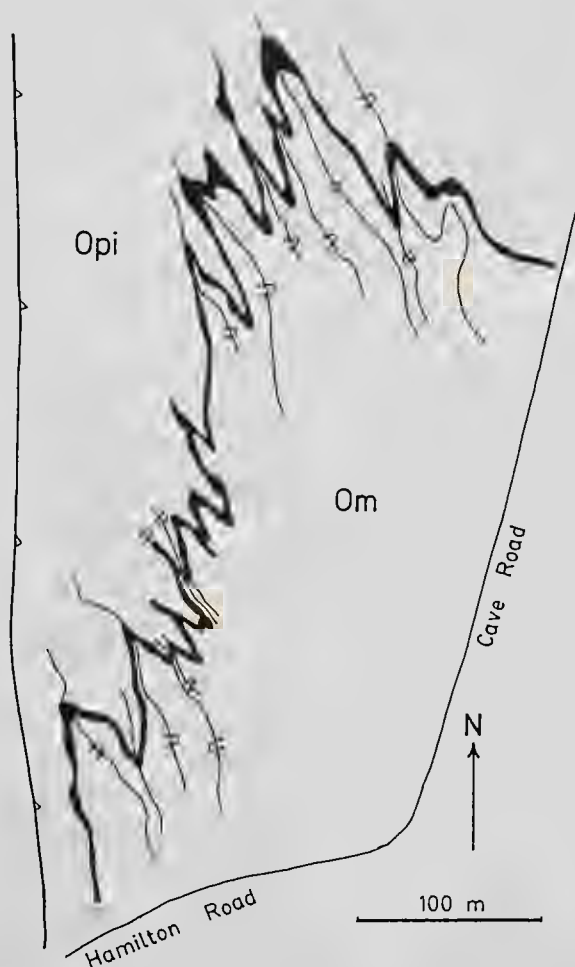


Figure 5 - Structure of James Pasture (after Crosby, 1963). The marker bed is a thin (.5 m) shale layer separating the Middlebury limestone (Om) from the dolostones and limestone of the Providence Island formation (Opi). Note the orientation of the folds relative to regional strike which is nearly due north.

- 36.9 Turn right.
- 37.5 UVM Morgan Horse Farm.
- 38.7 Bear left. Pulp Mill Covered Bridge - oldest covered bridge in Vermont and one of very few two lane covered bridges still in existence. If your vehicle is too large to go through:
- 0.0 turn right and continue southwest to Route 23
 - 0.4 turn left onto Route 23 and go to end
 - 1.2 turn left go to intersection
 - 1.3 turn left onto Main St. and proceed to Route 7 (note bridge rock from Stop 5)
 - 1.5 turn left onto Route 7
 - 1.6 turn right onto Seminary Street (by church).
= 39.8 below.
- 38.8 Continue straight as you leave bridge. You have just crossed the Sudbury thrust.
- 39.5 Turn left across train tracks.
- 39.5 Turn left.
- 39.6 Turn Right.
- 39.7 Turn right onto Route 7.
- 39.8 Turn left onto Seminary Street.
- 40.1 Continue straight (right fork).
- 41.1 Bear right onto Foote Street.
- 42.1 Turn left onto quarry access road.
- 42.3 STOP 7: Middlebury Quarry. As you enter the quarry, be very careful; this is a deep quarry and is still in operation with trucks constantly entering and leaving on the access road. A geologist of the Vermont Marble Co. will meet us at this stop.
- The marble of this quarry is cut by several thrusts which have been obscured by later recrystallization during metamorphism. The original layering can be recognized by interspersed pelitic layers which also show later cleavages. The stratigraphic position of the marble is not established, but I assign it to the Lemon Fair formation based on structural position relative to nearby fossil localities.
- Leave quarry.

- 42.5 Turn left as you leave access road.
- 43.4 Turn left onto Route 7.
- 44.7 Turn left onto Route 125.
- 46.3 Narrow bridge with dangerous curve.
- 46.5 STOP 8: East Middlebury Roadcut. Park to left of road in small parking area. The roadcut on south side of road shows a rotated early spaced cleavage. The rotation occurred as fault-bend folds formed during movement on a series of thrusts along the base of the mountain front (see Harding, this volume). The cleavage probably formed in an equivalent structural environment to that seen at Stop 2.

Turn around and proceed back down hill.

- 47.8 Turn right onto Route 116.

For the next few miles we follow the Green Mountain front. The base of the front is mostly hidden by glacial gravels, but wherever it is visible there is a series of thrusts separating the quartzites of the front from the carbonate-pelitic sequence of the adjacent valley. The hills to the west of the road are caused by the massive dolostones at the base of this sequence (probably equivalent to the dolostone lying directly atop the quartzite) being brought to the surface by thrusts.

- 56.6 Turn left onto dirt road.
- 57.6 Turn left.
- 58.0 STOP 9. New Haven Mills. Southwest facing ledges just into behind house provide a nice view of a fold and thrust system within the New Haven metamorphic complex. Note that most of the folds are related to thrusting. An early cleavage, seen best in the purer dolostones, pre-dates these folds.

Beware of the poison ivy growing along edge of woods.

End of trip.

To return to route 116, turn around, proceed back down road to end (across bridge), turn right, and go 0.8 mi.

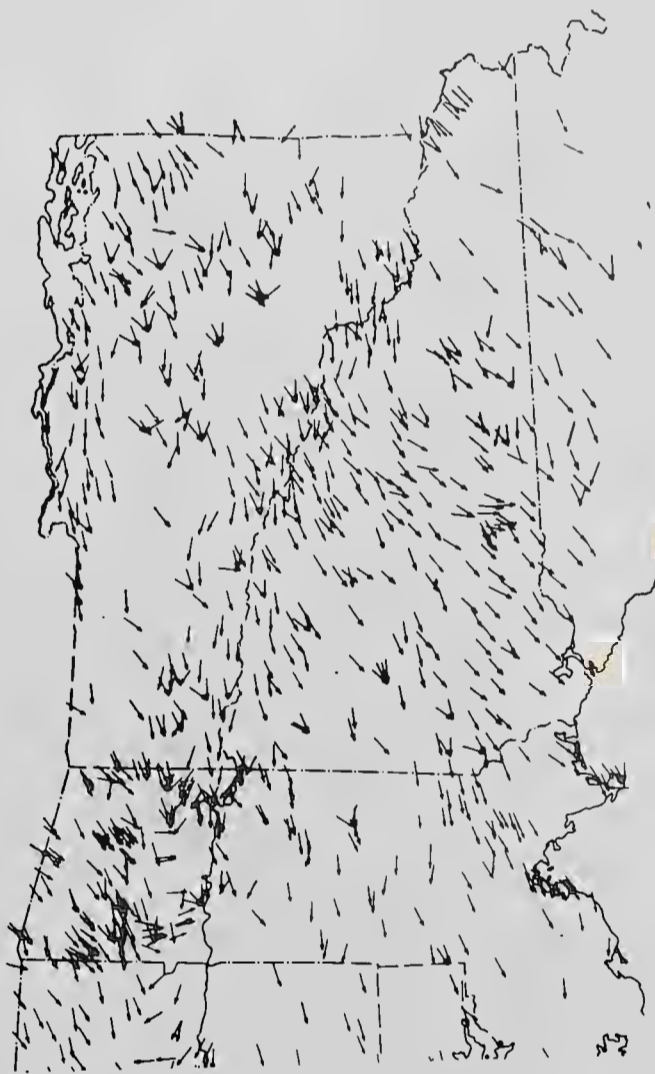


Figure 1. Compilation of striations in Vermont and New Hampshire by James W. Goldthwait (portion of Fig. 5-3 in Flint, 1957, p.60).

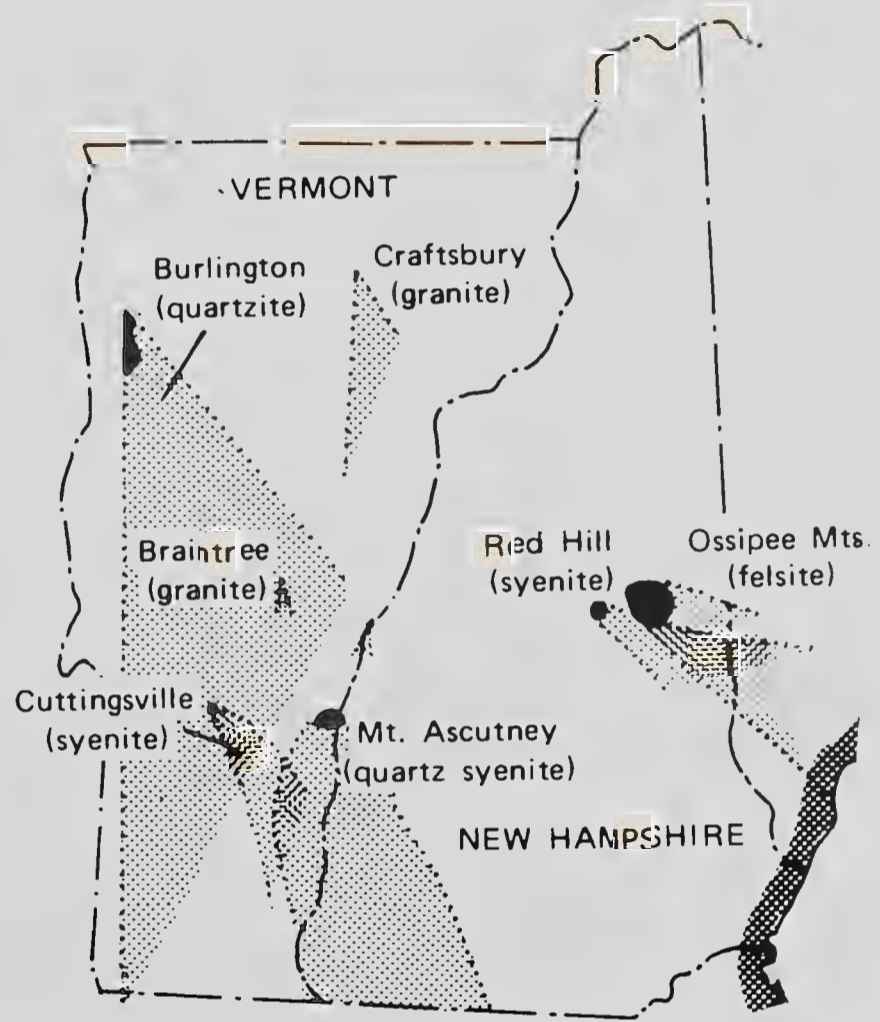


Figure 2. Indicator fans in Vermont and New Hampshire (portion of Fig. 7-19 in Flint, 1971, p.178).

SOUTHWEST-TRENDING STRIATIONS IN THE GREEN MOUNTAINS, CENTRAL VERMONT

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"That the highest points of land should have been scored by abrasions passing over them seemed to the older geologists better explained by floating than by glacial ice; for no one had then made clear how ice could move up hill to altitudes of thousands of feet. The ice of living glaciers moves down slopes: how then could the ancient ice have passed over the tops of the mountains unless the land itself had been so low that icebergs could have floated over them? The geologists had the credit of believing many strange stories, but even they hesitated to accept the doctrine that land ice could have been pushed over New England from the St. Lawrence valley."

Charles H. Hitchcock, 1904, Report of
the Vermont State Geologist, p. 67

INTRODUCTION

In 1861 Edward Hitchcock reported two directions of glacial striations in the Green Mountains, in the vicinity of Middlebury Gap. One set, oriented to the southeast, roughly coincided with striation directions in other parts of Vermont, showing that currents of ice (or icebergs, in Hitchcock's view) flowed southeast *out* of the Champlain Valley and across the mountains. The second set, however, directed to the southwest, was locally restricted to the valley of the Middlebury River and showed that ice flowed *into* the Champlain Valley. Hitchcock interpreted the southwest striations as evidence for local glaciers in the Green Mountains, though James Goldthwait argued in 1916 that the striations might have been produced by an ice sheet.

This field trip returns to Middlebury Gap, 116 years later, to reconsider the evidence of multiple striation directions in the Green Mountains. Recent mapping (work still in progress) shows that patterns of glacial striations in the Green Mountains are complex, recording multiple directions of ice movement.

REGIONAL PATTERNS OF GLACIATION

During the Late Wisconsinan glaciation, the Champlain Valley in western Vermont was a major conduit for ice flowing southward from ice domes in the Hudson Bay region. Ice radiated to the southwest into New York and to the southeast into New England from a locus in, or to the north of, the Champlain Valley (Mayewski et al., 1981; Hughes et al., 1985; Lowell and Kite, 1986). In Vermont, the southeast flow direction is indicated by patterns of



Figure 3. Indicator fans mapped by students at Norwich University. Source areas in solid black. Short lines down glacier from source areas represent 10% isopleths of source rocks in till samples



Figure 4. Distribution of some erratics of reddish-brown quartzite (solid dots) derived from the Monkton Formation (solid black areas).

glacial striations and indicator fans (Figs. 1-2). Indicator fans mapped by students at Northfield University show trends ranging from due south to S40E (Fig. 3). Many erratics of the reddish-brown Monkton quartzite have been carried southeast across the Green Mountains into southwest New Hampshire (Fig. 4).

The general pattern of deglaciation in Vermont was one of regional stagnation in the central highlands, and of oscillating or episodic ice margin retreat in the Hudson-Champlain Valley. The primary evidence for stagnation in central Vermont is the virtual absence of moraines and other ice-marginal features in highland areas. Ice-marginal features in the Hudson and southern Champlain Valleys, on the other hand, indicate active ice margin and stagnation zone retreat for the Champlain Valley lobe (Connally, 1970; De Simone and La Fleur, 1986). The Champlain Valley deglaciated rapidly, and was ice-free by about 13,000 years B.P. (Connally and Sirkin, 1973).

HISTORY OF ICE-FLOW STUDIES IN VERMONT

Striation patterns have been an important topic of study in Vermont since the early state surveys of the 19th century. Charles Adams (1846) and Edward Hitchcock and others (1861) noted the prevailing southeastward trend of striations across the state, though the cause of the pattern was controversial. Icebergs and "glacio-aqueous" agencies, acting in concert with epeirogenic movements of the land (submergence and elevation), were favored theories of 19th century geologists (Upham, 1894). Continental glaciation was widely accepted in the northeastern United States by the late 19th century, and Charles Hitchcock (1897) synthesized much of the striation data known at that time in his early description of the "Eastern Lobe" of the ice sheet. The "Eastern Lobe", now the Hudson-Champlain lobe, was considered as an eastern counterpart to the glacial lobes of the mid-western United States.

Multiple striation directions in Vermont have generated considerable controversy. As noted above, Edward Hitchcock (1861), and his son Charles Hitchcock (1904), considered multiple striation directions near Middlebury Gap as evidence for local glaciation in the Green Mountains. However, James Goldthwait (1916, p.68) argued that multiple striations, "though so strange as to excite much speculation in the '60's", could be produced along the lobate margin of an ice sheet. Stewart and MacClintock (1969) considered multiple striations as evidence for multiple ice advances across Vermont, both from the northeast and from the northwest, but this concept has not been substantiated by later workers (Larsen, 1972; Wagner et al., 1972).

RESULTS OF RECENT WORK

Recent mapping has revealed a rather large region of southwest-trending striations in the Green Mountains extending for at least 60 kilometers along the crest of the range from Appalachian Gap in the north ($44^{\circ}12'30''$) at least as far south as Sherburne Pass ($43^{\circ}40'$) (Figs. 5 and 6). One of the best and most accessible places to observe the southwest striations is at Middlebury Gap where they were first reported by Hitchcock and others in 1861 (Fig. 7). Early attempts to locate striations to the north and to the south of Middlebury Gap were thwarted by a thick cover of till on the Green Mountain summits (Upham, 1889; Hitchcock, 1904). Much of the present mapping was possible because of recent bedrock exposures on the Long Trail.

The crest of the Green Mountains in central Vermont lies mainly above 2,000 feet, and

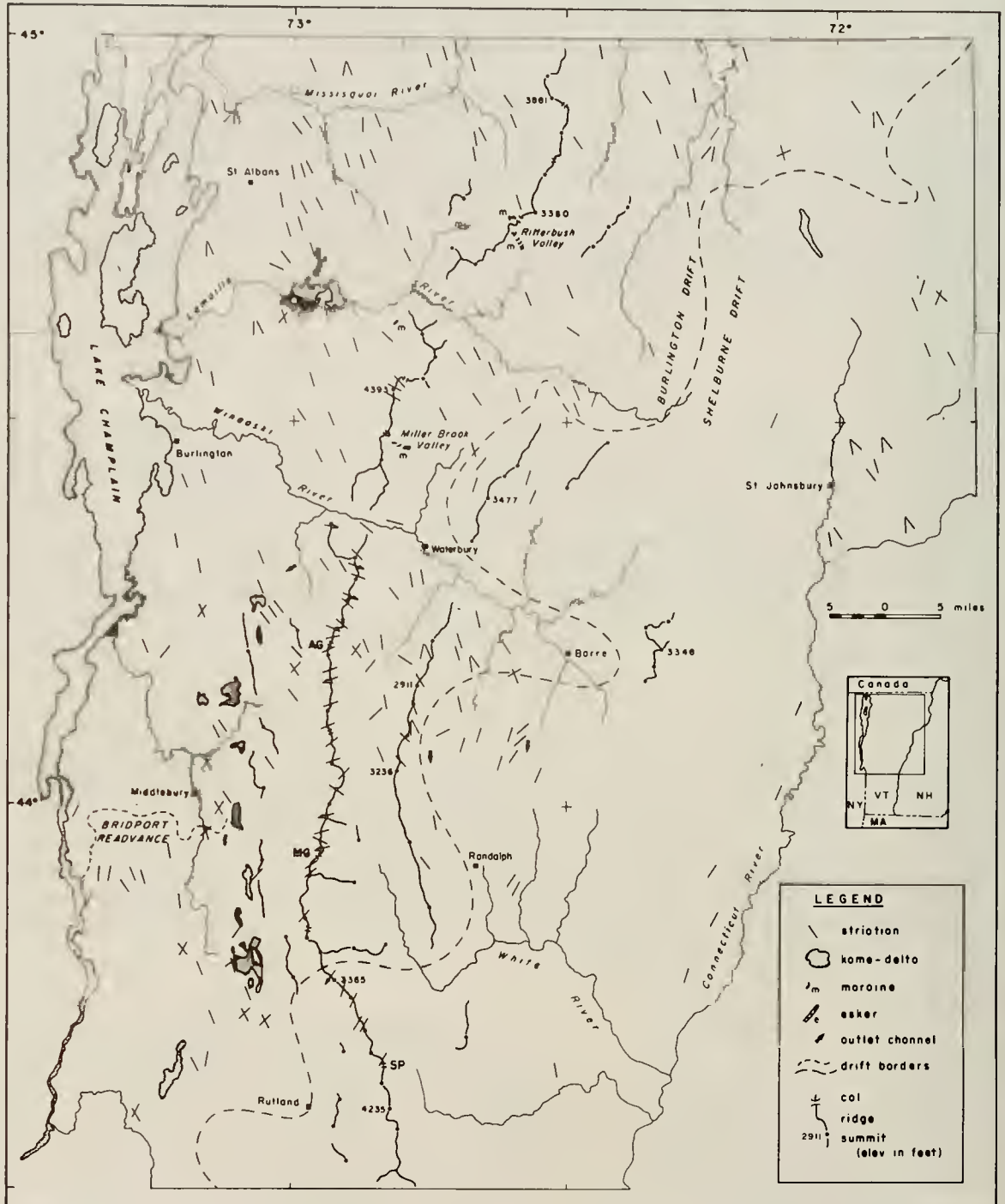


Figure 5. Glacial features in northern Vermont. Southwest striations occur along the crest of the Green Mountains from Appalachian Gap (AG) south to Shelburne Pass (SP) (MG = Middlebury Gap) (sources: Stewart and MacClintock, 1970; Connally, 1970; Wagner, 1970).

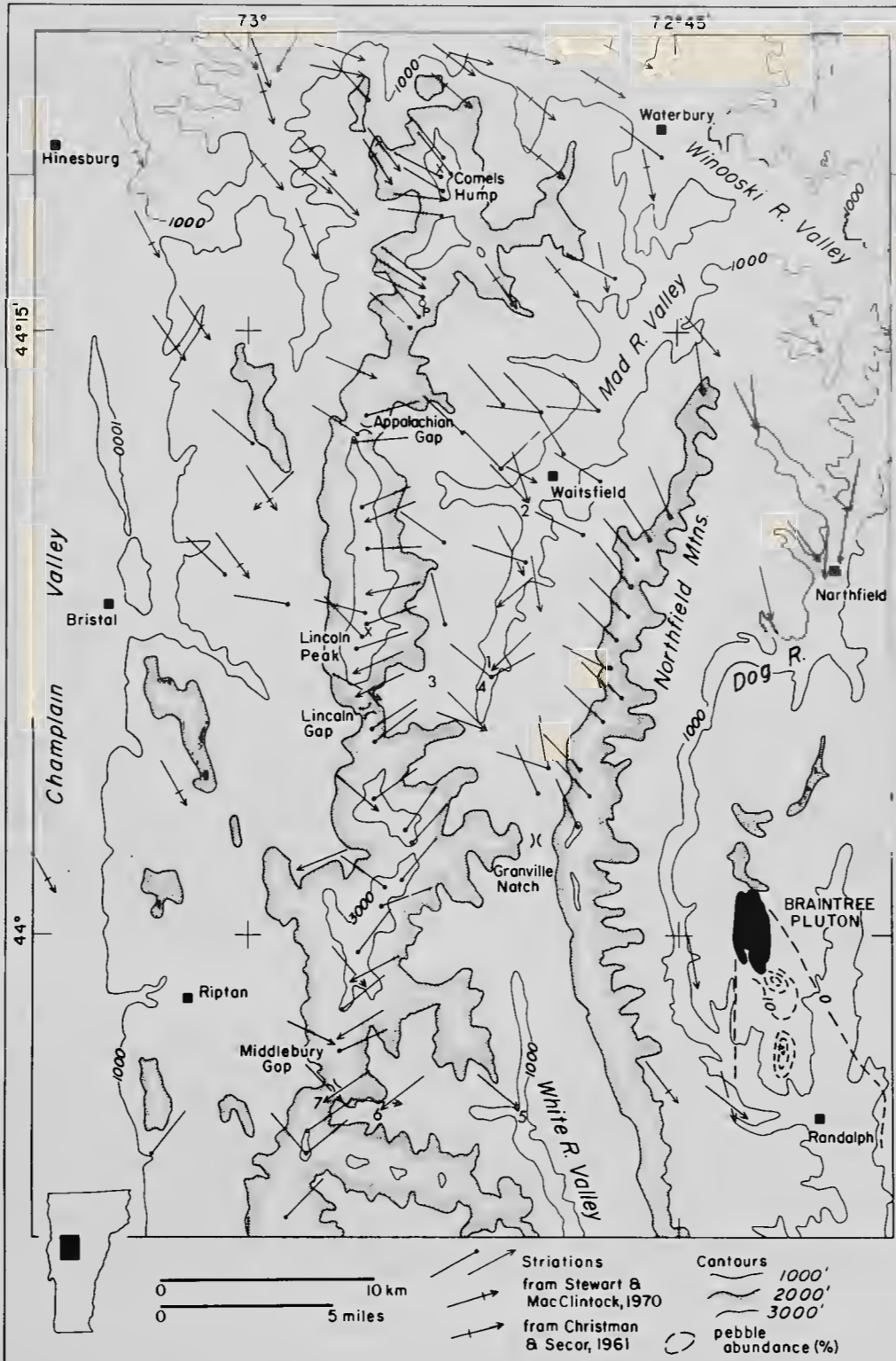


Figure 6. Topography and patterns of glacial striations in the Green Mountains, central Vermont. Striation location at dot or arrow; arrows show direction of crag and tail features. Numbers refer to STOPS in field trip guide. "P" is location of 2 ft. diameter pothole near summit of Burnt Rock Mtn. (Doll, 1936).

is flanked on the west by the Champlain Valley lowlands, and on the east by the Mad River Valley and the Northfield Mountains. North-south trending valleys and ridges are the products of differential erosion of metamorphic rocks of the Green Mountain Anticlinorium. The range is breached by the west-flowing Winooski and Lamoille Rivers (superimposed drainages), and roughly north-south trending tributaries to these rivers form a crude trellis drainage pattern (Figs. 5 and 6).

The predominant striation direction in northern Vermont is to the southeast (Figs. 2, 5) except along the very crest of the Green Mountains between Appalachian Gap and Sherburne Pass where striations are to the southwest (Figs. 5, 6). Southeast striations on the highest summits (Mt. Mansfield, Camels Hump, the Northfield Mountains, Killington Peak) are generally oriented S40E to S50E. At lower elevations, striation directions show some conformity to valley bottoms (as in the Winooski River Valley, Fig. 6) but striations may also be oblique to valley trends (as in the Mad River Valley, Fig. 6). Peaks cause some deflection of ice flow, as at Camel's Hump where flow on the peak is to the southeast, shifting to east-west in the col about two kilometers to the south.

Southwest striations occur primarily along the very crest of the Green Mountains and locally in the Mad River and Champlain Valleys. The boundary between southeast and southwest-trending striations near Appalachian Gap is abrupt, changing from about S60E north of the gap to about S70W south of the gap, in a distance of less than three kilometers. Following south along the range, the southwest striation direction changes gradually to about S50W near Middlebury Gap and about S30W near Sherburne Pass (Fig. 5). Cross-cutting relationships of striations in the vicinity of Middlebury Gap, and at one locality near Lincoln Gap, show that southwest striations on the crest of the Green Mountains are younger than (cut across) southeast striations. This is opposite to what Stewart and MacClintock (1969) and Connally (1970) report in the Champlain Valley, where southwest striations are apparently older than southeast striations. There is no evidence for southwest movement of clasts in the Braintree indicator fan, located 21 kilometers east-northeast of Middlebury Gap (Figs. 6 and 8).

DISCUSSION

The sequence of glacial, or deglacial, events that produced the striation patterns in the Green Mountains is seemingly complex. The regional pattern of striations in the Green Mountains, and the dispersion direction of erratics in central Vermont, indicate ice flow to the southeast from a source area in the St. Lawrence and Champlain Valleys. However, at least along the very crest of the Green Mountains, from Appalachian Gap south to Sherburne Pass, southwest striations show that the last flow of ice was into the Champlain Valley. At least locally, the ice surface also must have been sloping into the Champlain Valley. Either there was a local buildup of ice in the central Green Mountains, or a drawdown of ice in the Champlain Valley. There is no compelling evidence for a local ice cap in this region.

Studies of striation patterns in many areas of northern New England and adjacent Canada show that local ice flow reversals were common during retreat of the last ice sheet (eg. Lamarche, 1974; Chauvin et al., 1985; Lowell and Kite, 1986). The mechanism of flow reversal is mainly related to rapid thinning of ice in ice-streams and ice lobes, especially in the St. Lawrence Valley. Thinning and drawdown of ice in the Champlain Valley could account for local reversals of flow to the southwest along the Green Mountain crest, thereby explaining the localized distribution of the southwest striations immediately adjacent to the Champlain Valley. At least for a short time there was probably an ice divide in the region of

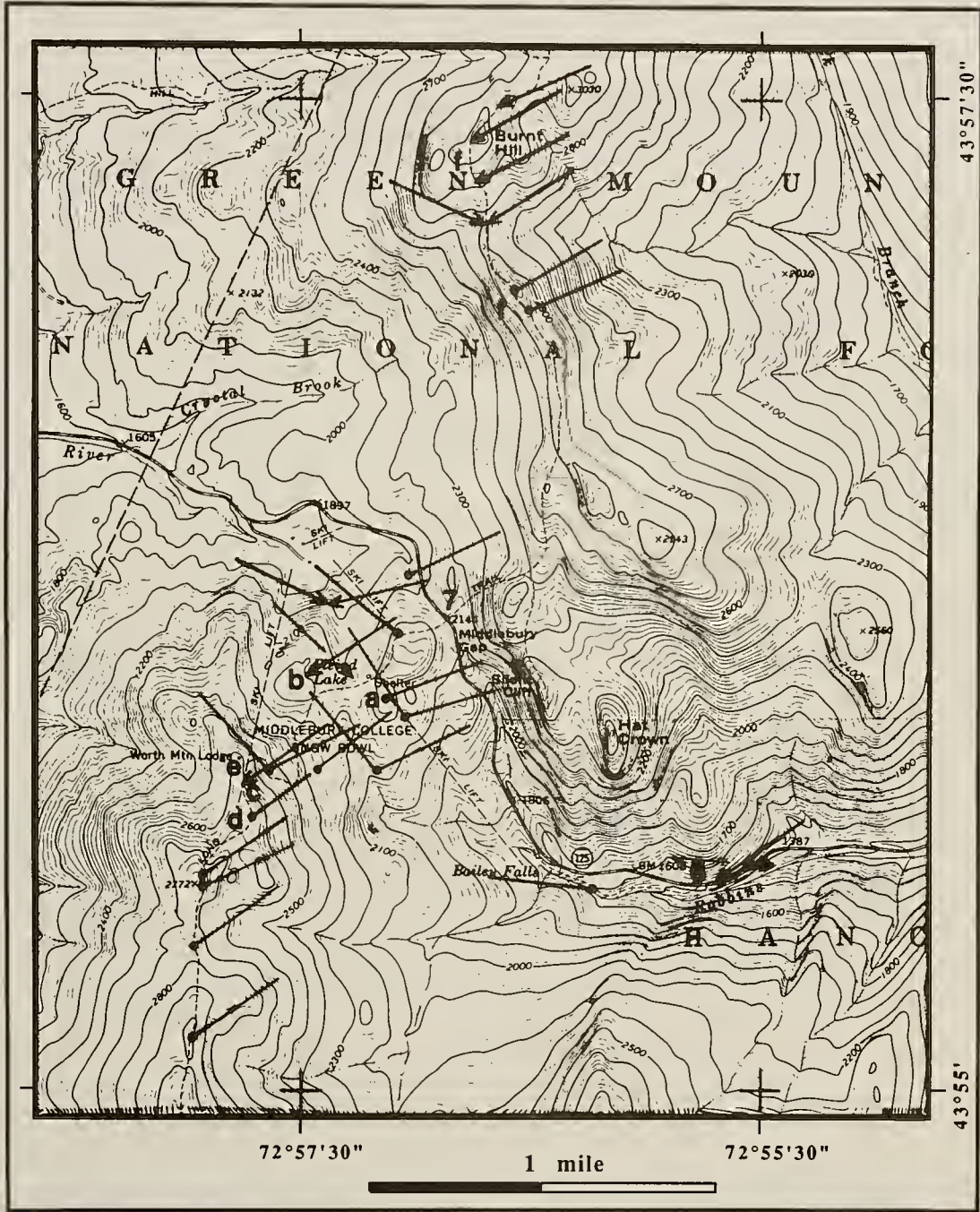


Figure 7. Striation patterns in the Middlebury Gap area. Striation location at dot or arrow; arrows show direction of crag and tail features. Numbers refer to STOPS in field trip guide. The common occurrence of southeast striations at Middlebury Gap is probably because the gap is in the lee of significant bedrock knobs to the northeast: Burnt Hill, Silent Cliff and the Hat Crown.

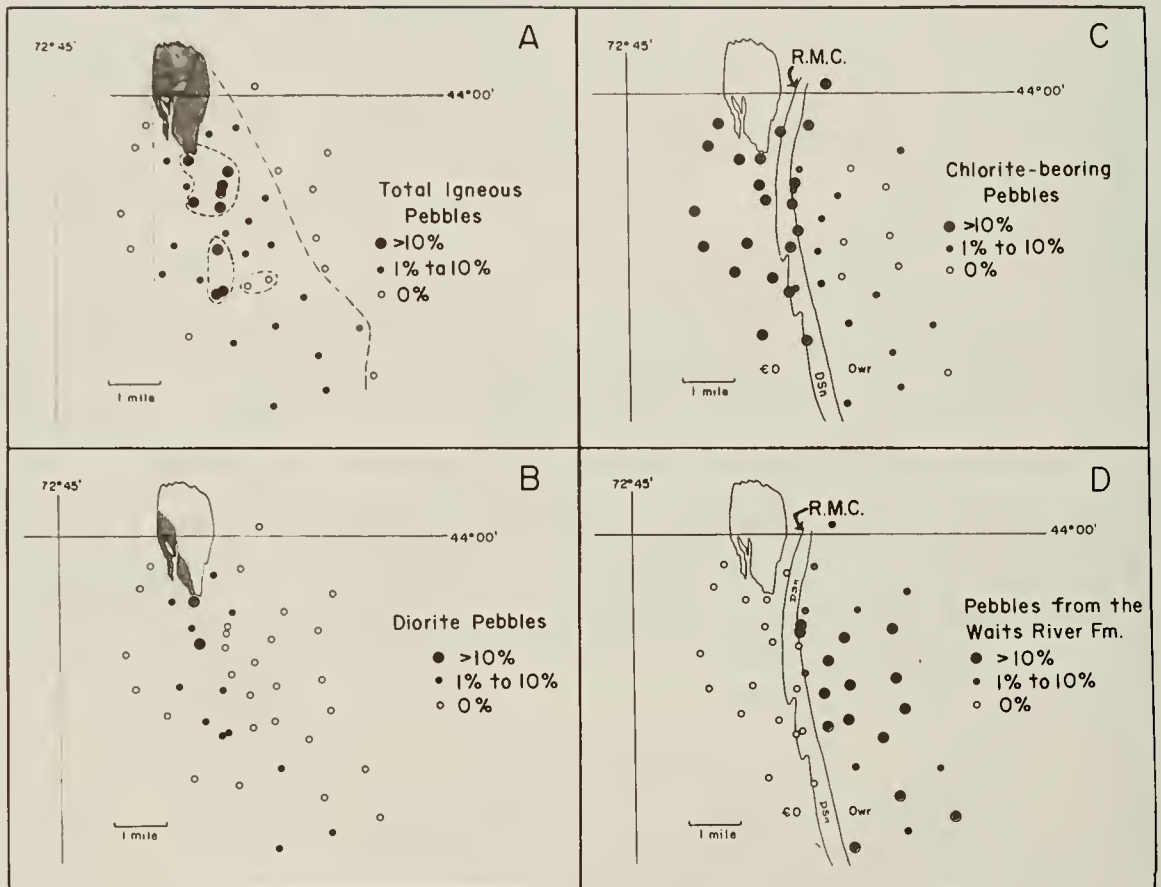


Figure 8. Distribution of indicator clasts in northwest part of Randolph 15' quadrangle, (A) Braintree indicator fan based on percent of total igneous pebbles (granite, diorite, and gabbro); stipple pattern, Braintree Pluton, (B) indicator fan based on percent of diorite pebbles; stipple pattern, area of diorite outcrop, (C) distribution of chlorite-bearing pebbles derived from the Cambro-Ordovician, -EO; DSn, Northfield Formation; Dwr, Waits River Formation; R.M.C., Richardson Memorial Contact, (D) distribution of pebbles of brown-weathering calcareous quartzite derived from the Waits River Formation.

the Mad River Valley, with ice flowing both to the southeast and to the southwest. The divide may have sloped to the south, resulting in the more southerly component of the southwest flow near Sherburne Pass, but more work must be done in this area. The hypothesis outlined above is anticipated by the theoretical ice surface reconstructions of Hughes et al. (1985, Fig. 3) in their model for about 16,000 years B.P. showing flowlines diverging to the southeast and to the southwest in central Vermont.

The hypothesis of ice drawdown in the Champlain Valley, combined with localized restructuring of the ice sheet, raises a number of important, and as yet unanswered, questions. Where was the southern margin of the ice sheet when ice flowed southwest into the Champlain Valley? Did ice in the Champlain Valley remain active after the mountains became ice-free? Did the margin of southwest flow retreat up the west flank of the Green Mountains, contributing sediment to kame deltas in the Champlain Valley (see Connally, 1982), or did the southwest flow stagnate in a "top to bottom" mode, essentially collapsing in place? Does the "boundary" between southeast and southwest flow directions near Appalachian Gap represent the southern margin of late-stage ice flowing southeast across the Green Mountains in northern Vermont? These questions provide the focus for on-going studies in the Green Mountains.

ACKNOWLEDGMENTS

We thank Stewart F. Clark for unpublished data on striation directions in the Northfield Mountains.

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ITINERARY

Assembly point is at the U-Haul garage on the east side of Vermont Route 100 at the junction with the Lincoln Gap Road, located about 19 miles south of both Exits 9 and 10 on Interstate 89 and about 0.4 miles south-southwest of the Warren General Store. Assembly time is 9:00 A.M. Topographic maps: Warren, Hancock, and Bread Loaf 7.5 minute quadrangles.

Mileage

0.0 **STOP 1:** The outcrop at the junction of the Lincoln Gap Road and Route 100 shows striations oriented S43W on stoss surfaces, and meltwater scour on lee surfaces, probably representing pressure-melting and regelation (freezing) processes at the base of the ice. These SW striations may or may not be related to SW striations along the crest of the Green Mountains to the west. Bedrock is Cambrian Pinney Hollow formation (Doll, 1961).

STOP 2 (Optional): 5.3 miles north of STOP 1 on Route 100, at bridge across the Mad River, just south of the village of Irasburg. Strong striations to the SE, with large crag and tails.

STOP 3 (Optional): Follow Lincoln Gap road west from STOP 1 for 2.6 miles. Turn right just beyond mailbox and brook crossing, follow narrow dirt road to pit.

The Hartshorn pit is located in a small delta on the projected shoreline (N20W-0.9m/km) of glacial Lake Granville that formed to the north of a 1,410-foot threshold at Granville Notch. Deltaic topsets consist of 2.0 meters of poorly sorted pebble gravel with cobbles interbedded with layers of fine sand, medium sand, and pebbly coarse sand with dune crossbeds dipping south. Deltaic foreset and collapsed beds are greater than 2.0 meters thick under a trimmed surface and dip to the south. The foreset beds consist of layers of poorly sorted fine to medium grained sand and pebble gravel.

Proceed south on Route 100, from Stop 1.

0.1 Take a right (west) on a small dirt road immediately before (north of) a bridge over the Mad River. Keep to the right and enter into gravel pit.

0.3 **STOP 4:** The Don Moore pit is made up of two active faces. The higher northwest face is about 160 meters (525 ft) long and up to 27 meters (89 ft) high. The lower southwest face is about 81 meters (265 ft) long. A measured section located near the center of the northwest face has four main units in a fining-upward sequence that is capped by colluvium. Unit 1 at the base is poorly sorted pebble gravel with cobbles and is interpreted to have been formed in a subglacial tunnel that drained south into glacial Lake Granville.

Unit 2 is 8.7 meters thick and is composed mainly of interbedded fine and very fine sand. Ripple crossbedding dips both to the northeast and southwest. In the middle of Unit 2 are 1.6 meters of granule to pebble gravel interbedded with fine sand. Unit 2 is interpreted to be proximal lake-bottom sediments deposited near the ice margin.

Unit 3 is 4.5 meters thick and is composed of fine sand in massive grain flows interbedded with very fine sand. A-type ripple cross-lamination near the middle of several grain flows indicates transport to the northeast. Unit 3 is interpreted to have been deposited by turbidity currents possibly kicked into action by meteoric events in tributary valleys to the south or west.

Unit 4 consists of 3.1 meters of laminated gray silt and orangish-brown to brown, fine to very fine sand in beds about 1 centimeter thick. Minor crossbeds in fine sand dip to the southwest. The inferred environment of deposition was the quiet bottom of glacial Lake Granville when the ice margin was located well to the north. The depth of Lake Granville was about 115 meters when the uppermost lacustrine beds were deposited.

The pebble gravel with cobbles that caps the western half of the southwest face underlies a stream terrace that extends 90 meters to the west. This terrace was formed after Lake Granville drained and may have been graded to Lake Winooski. Sediments under the stream-terrace gravels have crossbeds that dip toward S10W to S25W, up the Mad River valley. In the past, 2.0 to 3.0 meters of south-dipping trough crossbeds 20 to 50 centimeters thick were exposed in the southwest face.

Proceed back to Route 100.

- 0.5 Turn right (south) on Route 100, cross the Mad River.
- 6.5 Height of land at Granville Notch. Spillway of Lake Granville at an approximate elevation of 1,410 feet.
- 7.2 Potholes(?) in the left (east) wall of the notch.
- 10.6 Town of Granville. Continue south on Route 100. The valley floor south of Granville is the outwash(?) plain for glacial meltwaters flowing through Granville Notch.
- 14.9 Pass outcrop on right (west) side of road, immediately north of the junction of Routes 100 and 125.
- 15.0 Turn right (west) onto Route 125 and park on right (market on south side of road).

STOP 5: Outcrop on the west side of Route 100, immediately north of junction with Route 125. Striations are S50E.

- 18.0 Pass road to Texas Falls picnic area (nice potholes). Continue west on Route 125.
- 19.8 Park in pullout on left (south) side of road, immediately uphill (west) of glacially polished outcrop on right (north) side of road.

STOP 6: Ascend up and to the left (west) of the roadcut about 40 feet to a small slab showing striations (with stoss and lee) trending east-southeast (S85E) and southwest (S53W). The SW striations cut across the SE striations.

A section of the old Middlebury Gap Road (?) can be followed from the pullout uphill to the west for about a half mile. In 1987 a slump along the river exposed laminated silts and fine sands containing small dropstones.

Continue west on Route 125.

21.2 Park at the height of land, Middlebury Gap (elevation 2144 feet).

STOP 7: Hike south on the Long Trail, pass top of ski lift, pass shelter (0.3 miles), come to ski trail just beyond shelter.

STOP 7A: Good SW striations at top of this ski trail about 150 meters east of Long Trail. Continue on Long Trail about 0.1 miles to junction, take side trail west for 0.1 miles to Lake Pleiad.

STOP 7B: South shore of Lake Pleiad, SE striations and grooves (stoss and lee) crossed by weathered SW striations. Return to Long Trail and continue south. Cross ski trail and continue upwards. Cross another ski trail.

STOP 7C: Excellent SW striations on ledges, at ski trail crossing. Continue up to top of spur. Cross two ski trails.

STOP 7D: Strong SW striations at second ski trail crossing; SW striations predominate at higher elevations to south along the range. Leave Long Trail and follow ski lift-line down and west to lookout near Worth Lodge.

STOP 7E: View N of Lake Pleiad and Middlebury Gap. Return to Middlebury Gap.

STOP 7F: Walk north on Long Trail to Burnt Hill, strong striations to the SW.

STRUCTURE AND METAMORPHISM AT TILLOTSON PEAK, NORTH-CENTRAL VERMONT

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We wish to dedicate this work to A. H. Chidester
whose regional mapping of the Tillotson Peak area
provided the basis for this detailed study.

PROLOGUE

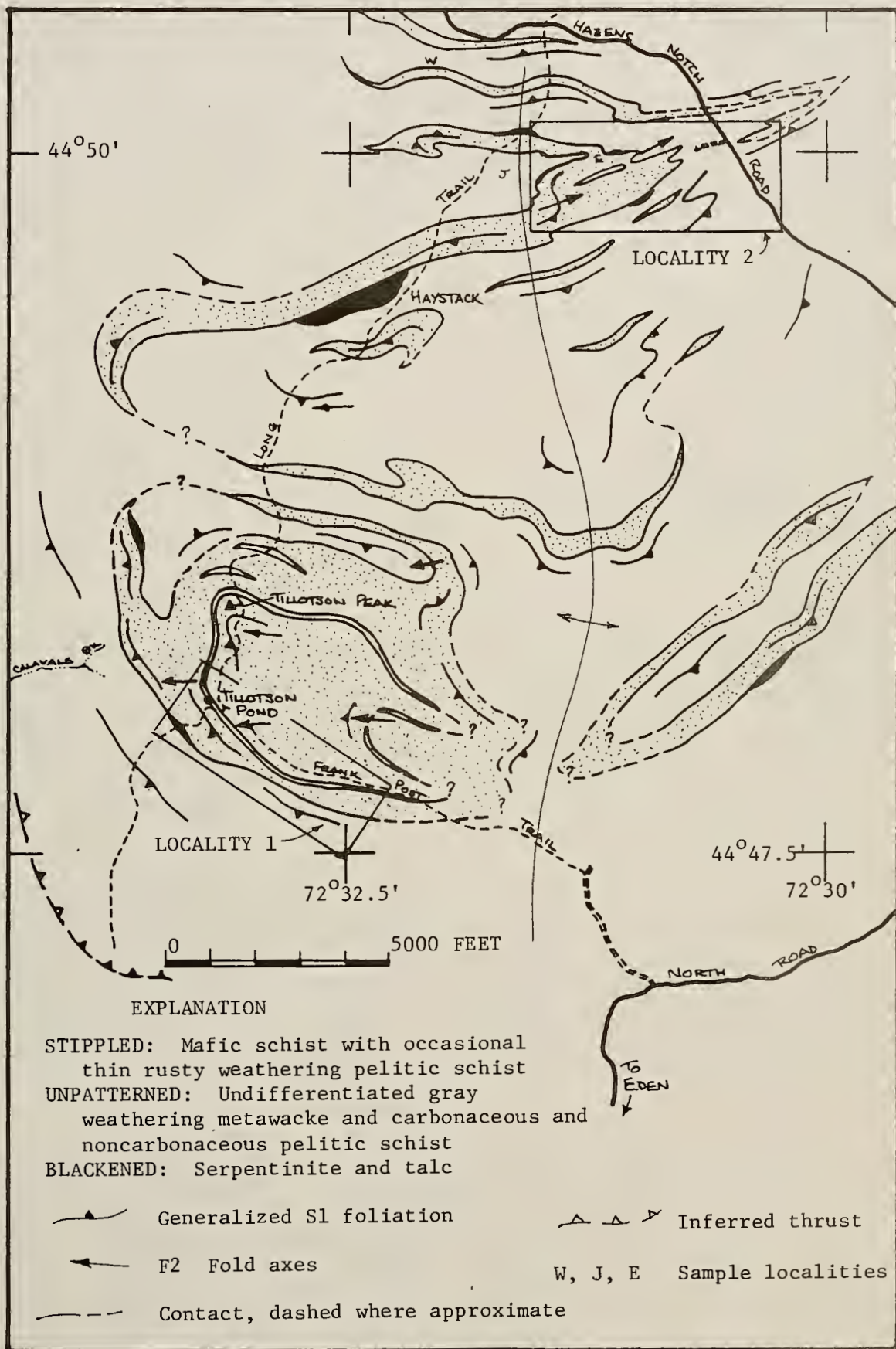
The unique occurrence of glaucophane + epidote blueschist and of eclogite in the U. S. Appalachians hides among trees, nettles, and ferns in the vicinity of Tillotson Peak, Vermont (fig. 1). One cannot see these rocks in roadcuts, or even in outcrops near roads. Consequently, this field trip will be different from most NEIGC trips. We will make traverses along two brooks and nearby trails and brush to see the range of rock types and their structural setting. If the foliage is anything like that when these rocks were stumbled upon in October, 1973, we should have a beautiful trip. If you come during the Spring or Summer, bring plenty of bug dope!

GEOLOGIC SETTING

Rocks in the vicinity of Tillotson Peak, Vermont, were mapped by Cady et al. (1963) in their study of the ultramafic rocks in northern Vermont. These ultramafic rocks are within a N-S trending zone on the east limb of the Green Mountain anticlinorium (GMA) and other Grenville-age basement massifs farther south in the northern Appalachians. They are generally considered to be ophiolitic fragments emplaced during Taconian closure of the Iapetus Ocean (Doolan et al., 1982; Stanley et al., 1984). The largest exposure of ultramafic rocks is at Belvidere Mountain (Chidester et al., 1978; Gale, 1986), the southern margin of our geologic mapping.

Laird and Albee (1981a) discovered high-pressure facies series mafic schist within the Belvidere Mountain Amphibolite Member of the Hazens Notch Formation (nomenclature from Doll et al., 1961) at Tillotson Peak. Farther south and east of the GMA Laird and Albee (1981b) reported medium-high-pressure facies series mafic schist within the Hazens Notch, Ottauquechee, Stowe, and Pinney Hollow formations (fig. 1; Laird, this volume). West of the GMA Laird and Albee (1981b) and Laird (this volume) reported medium-high-pressure facies series metamorphism in the Pinnacle

FIGURE 1. GEOLOGIC MAP OF THE TILLOTSON PEAK AND HAYSTACK MOUNTAIN AREA



Formation and medium-pressure facies series metamorphism in the Hazens Notch and Underhill formations.

Laird et al. (1984) reported a total fusion $^{40}/^{39}\text{Ar}$ glaucophane age of 468 ± 6.4 Ma at Tillotson Peak and a total fusion $^{40}/^{39}\text{Ar}$ barrosite age of 490 ± 8.0 Ma at Belvidere Mountain. The glaucophane age is consistent with other $^{40}/^{39}\text{Ar}$ total fusion and step-wise heating ages in northern Vermont and farther south in the northern Appalachians that indicate Taconian metamorphism at 465 ± 10 Ma (see Sutter et al., 1985 for a recent summary of isotopic data). Laird et al. (1984) suggested that the older age at Belvidere Mountain may mean that these rocks were metamorphosed before being emplaced tectonically, consistent with the subsequent mapping of Gale (1986) where the dated mafic unit is interpreted as being fault-bounded.

As discussed by Cady et al. (1963) the structure of the Tillotson Peak area is distinct in that E-W folds are preserved. These are not unique in the Taconian regions of Vermont nor in Grenville basement rocks, but they are unusual in that the dominant folds in Vermont are generally N-S. The reasons for these distinct orientations are speculative and include fold interference by yet undefined E-W graben structures in the underlying, presumably Grenville basement (Cady et al., 1963), oblique collision towards a relict failed arm (the Ottawa graben) or aulacogen (Doolan et al., 1982), and rotational fold-thrusting caused by large included masses (like the Belvidere ultramafic body, or seamounts; Bothner and Laird, 1986). A recent observation by Doolan (oral communication, 1987) that E-W folds seem always related to large ultramafic bodies in both the U. S. and Canadian Appalachians warrants further consideration.

To understand the structural setting of these high-pressure facies series rocks, we have mapped the Tillotson Peak area at a scale of 1:10,000 and have expanded earlier structural, petrologic, and isotopic studies to provide a basis for interpreting the tectonic history and evaluating the emplacement hypotheses (Cady et al., 1963; Doolan et al., 1982; Stanley et al., 1984; Stanley and Ratcliffe, 1985) of this vestige of the Iapetus Ocean.

LITHOLOGIES

The Tillotson Peak area is composed primarily of mafic schist referred to as the Belvidere Mountain Amphibolite Member of the Hazens Notch Formation (Doll et al., 1961). The mafic rock that crops out along Traverse 1 (Lockwood Brook and north, fig. 1) is typically dark blue gray, fine- to medium-grained, massive to schistose amphibolite with epidote + garnet porphy-

roblasts + chlorite + titanite + magnetite + gold-colored sulfide. Phengitic muscovite, plagioclase, and quartz are not uncommon. Calcite and/or dolomite and paragonite occur locally. Three amphiboles are seen in thin section: glaucophane, barroisite, and actinolite (see section on metamorphism). Glaucophane is widespread in the Tillotson area; omphacite + garnet occur locally (Fig.1). Neither glaucophane nor omphacite is easily recognized in hand specimen, or in thin section for that matter, because they are generally very pale lavender and pale green, respectively (due to low Fe^{3+} content).

Farther north, Cady et al. (1963) mapped another mafic unit that they distinguished as the amphibolite member of the Hazens Notch Formation. This rock will be seen on Traverse 2 (Eclogite Brook, fig. 1). We do not see any lithologic differences between these rocks and those previously mapped as the Belvidere Mountain Amphibolite. Our mapping shows that they are the same unit as foreseen by Cady et al. (1976, p. B-13), repeated by folding (and thus further ornamenting the "Trilobite"). Amphibolites mapped farther north than the Hazens Notch road (VT Route 58) contain neither glaucophane nor omphacite; garnet is replaced by chlorite.

Within the mafic rocks are interlayers of silver gray, medium-grained pelitic schist containing white mica + quartz + chlorite +/- garnet +/- plagioclase, and often glaucophane altered to a fine-grained symplectite. Both phengitic muscovite and paragonite occur. Chloritoid is rare. Some of these layers are thick enough to be mapped at a scale of 1:10,000 (fig.1). We will see infolded mafic and pelitic schists on both traverses.

The Hazens Notch Formation as defined by Cady et al. (1963) is typically brown-weathered, gray, medium-grained graphitic and non-graphitic schist with plagioclase porphyroblasts, white mica, quartz, chlorite, and gold-colored sulfide. Magnetite and polycrystalline quartz lenticles are common north and east of Tillotson Peak. In several places (eg., along the Long Trail between Tillotson Camp and Belvidere Mountain), the carbonaceous Hazens Notch Formation is distinctively black and orange weathered pelitic schist with white mica + quartz + plagioclase porphyroblasts + titanite + graphite. It has a strong C/S shear fabric and no obvious layering.

Cady et al. (1963) mapped "albite gneiss" at the contact of the Belvidere Mountain Amphibolite and the carbonaceous schist typical of much of the Hazens Notch Formation. This unit is massive to schistose and medium-grained with light gray plagioclase (probably all albite) + quartz + epidote +/- garnet +/- magnetite layers and green chlorite-rich layers. Piemontite occurs at Lockwood Brook and at several other nearly inaccessible areas (but on strike with the Lockwood occurrence). Chloritoid occurs on Traverse 2. We prefer

not to use the term "gneiss" for this rock, in spite of its obvious compositional layering and frequent development of plagioclase porphyroblasts (in both felsic and pelitic layers), but rather a feldspathic metawacke. In most outcrops, the nature of the layers and the degree of deformation (excellent fold preservation) suggest that the layers represent original bedding.

Sparse layers of pelitic schist with knots of magnetite and quartz occur in Lockwood Brook at the beginning of Traverse 1 and also near Traverse 2. We wonder if these represent futile attempts at the formation of iron formation. Similarly, garnet-rich layers commonly seen in some mafic rocks suggest coticles (Mn-rich coatings on original basalt or mafic tuff deposits).

Ultramafic rock crops out as isolated lenses along the contact of the Belvidere Mountain Amphibolite and the feldspathic metawacke in several localities. The rocks are generally strongly sheared and are composed primarily of serpentine + talc + carbonate with knots of actinolite and chlorite. Little of the cross-fiber serpentine of the types observed in the Belvidere Mountain ultramafic body is present.

STRUCTURE

The Hazens Notch Slice of Stanley and Ratcliffe (1985) consists of mafic and pelitic schists and feldspathic metawacke of the Hazens Notch Formation, in and near its type locality at Hazens Notch (Cady et al, 1963). These rocks are multiply deformed. We recognize two clear folding episodes within the Hazens Notch package and infer a third based on map pattern. Faulting within the Tillotson - Haystack region is locally important, but through-going faults have not been clearly established.

The dominant foliation, S_1 , is for most rocks a transposition surface that is interpreted to parallel or closely parallel original lithologic layering and is axial planar to small isoclinal folds (F_1) variably preserved in mafic schists and feldspathic metawacke. Presumed original layering (S_0) is locally preserved in the coarser feldspathic metawacke as 2 to 5 cm thick layers, some possibly graded, and as strong compositional layers of alternating garnet (the "coticles") and of amphibole in mafic rocks. S_1 is refolded about approximately east-west axes into nearly reclined mesoscopic, and inferred macroscopic folds, that produce the overall map pattern (F_2). That pattern closely resembles Type 2 interference folds of Ramsay (1967). Outcrop-scale mesoscopic nearly reclined folds are best represented by mafic schist along the southern contact of the mafic rocks comprising Tillotson Peak in and near Lockwood Brook, west of Haystack Peak on the ridge toward Burnt

Mountain, and in feldspathic metawacke along the Long Trail approaching Haystack Peak from the south (fig. 1).

F₃ is a broad, open macroscopic antiform, referred to by Cady et al. (1963) as the Gilmore anticline. No mesoscopic structures have been recognized that can be directly tied to this regional fold. Its confirmation is supported by comparison of the trends and plunges of F₂-fold axes from the Lockwood Brook area to the south and the Eclogite Brook area to the north. Those data are presented as equal area plots in figures 4 and 5 and clearly show that the different orientations of the same structures from the Lockwood Brook area and the Eclogite Brook area are related by a simple north-trending regional fold.

S₂ is a crenulation cleavage best developed in mafic schist and weakly developed in feldspathic metawacke. Pelitic schists occasionally contain this surface. It is approximately axial planar to F₂ mesoscopic reclined folds. The intersection lineation (L_C) formed by this cleavage trends approximately parallel to reclined fold axes, but does show a spread that is consistent with later folding about a north-south axis as suggested above. Preferred orientation of amphibole is often well-developed in mafic schists (L_M). Mullion structure is locally well developed in some of the feldspathic metawacke near contacts with mafic and ultramafic rock.

Contacts between mafic schist and feldspathic metawacke in the Tillotson area are locally marked by discontinuous ultramafic bodies (Long Trail south of Tillotson Pond, Haystack Peak, and Eclogite Brook, for example). Foliation is typically parallel. Contacts between mafic rock and intercalated pelitic schist are gradational within the Tillotson Peak - Haystack area and suggest original (but likely modified by internal shear during folding) layering. Strong C/S shear fabrics are observed most commonly in areas underlain by carbonaceous, orange-weathering pelitic schists of more typical Hazens Notch Formation south of the mafic mass that comprises Tillotson and the area including Hazens Notch. These rocks generally separate the areas dominantly underlain by mafic schist and feldspathic metawacke and appear to "confine" the distribution of glaucophane and omphacite bearing assemblages. We suggest that this unit marks the position of one or more faults, the full extent of which awaits further analysis.

The overall distribution of intercalated mafic and pelitic schist, feldspathic metawacke, and carbonaceous pelitic schist of the Tillotson - Haystack area confirms the general east-west structural form originally defined by the mapping of Cady et al. (1963). Several smaller ultramafic bodies have been

recognized along the contact between mafic schist and "albite gneiss." A significant difference, however, is the recognition that Tillotson is underlain by a complexly refolded system rather than the simpler doubly plunging syncline of Cady et al (1963). In addition, tectonic contacts separate carbonaceous pelitic schist and feldspathic metawacke and separate feldspathic metawacke and mafic rock that dominate this region. The latter is marked by ultramafic slivers in several places. Mafic and pelitic rocks within the main mafic mass are intercalated, and the gradational contacts commonly separating them within single outcrops argue effectively that they were originally deposited together. Equally important is the observation that the feldspathic metawacke and intercalated mafic and pelitic schist record the same metamorphic history.

METAMORPHISM

Glaucophane is abundant in mafic, pelitic, and felsic rocks from the Long Trail between Lockwood and Eclogite brooks and east to North Road (fig. 1). Mineral assemblages could have formed in equilibrium with different assemblages distinguishing differences in bulk rock composition (fig. 2A). Equilibrium is not attained on a thin section scale, however, and a record of the pressure-temperature path of many samples is nicely preserved.

At Lockwood Brook (Traverse 1) and several other localities barroisite is present only as inclusions in garnet and as cores of amphibole grains with actinolite rims (Laird and Albee, 1981a, Plate 1B; Laird and Bothner, 1986, fig. B-6). Based on theoretical, empirical, and experimental considerations (Laird, 1986), this relationship implies a decrease in metamorphic temperature with time. The pressure must still have been reasonably high (>7 Kbar using fig. 11, Maruyama et al., 1986) as glaucophane appears to be stable with actinolite.

Later retrograde metamorphism caused alteration of glaucophane and omphacite to fine-grained symplectite (Laird and Albee, 1981a, Plate 1C). Garnet was altered to chlorite. This alteration is pervasive east of the Gilmore antiform and near Hazens Notch Road. Thin sections of samples from Traverse 2 shown this retrogradation.

Omphacite inclusions in garnet from Traverse 2 (E, fig. 2A) record increasing pressure with increasing temperature. From garnet cores to rims temperature-pressure estimates based on the garnet-clinopyroxene geothermometer and on isopleths of the reaction albite = jadeite + quartz (fig. 3) are 380° C, 9.5 Kbar to 550° C, 12.5 Kbar. Garnet rims and omphacite grains adjacent to garnet give temperature and pressure from 540° C, 12.5

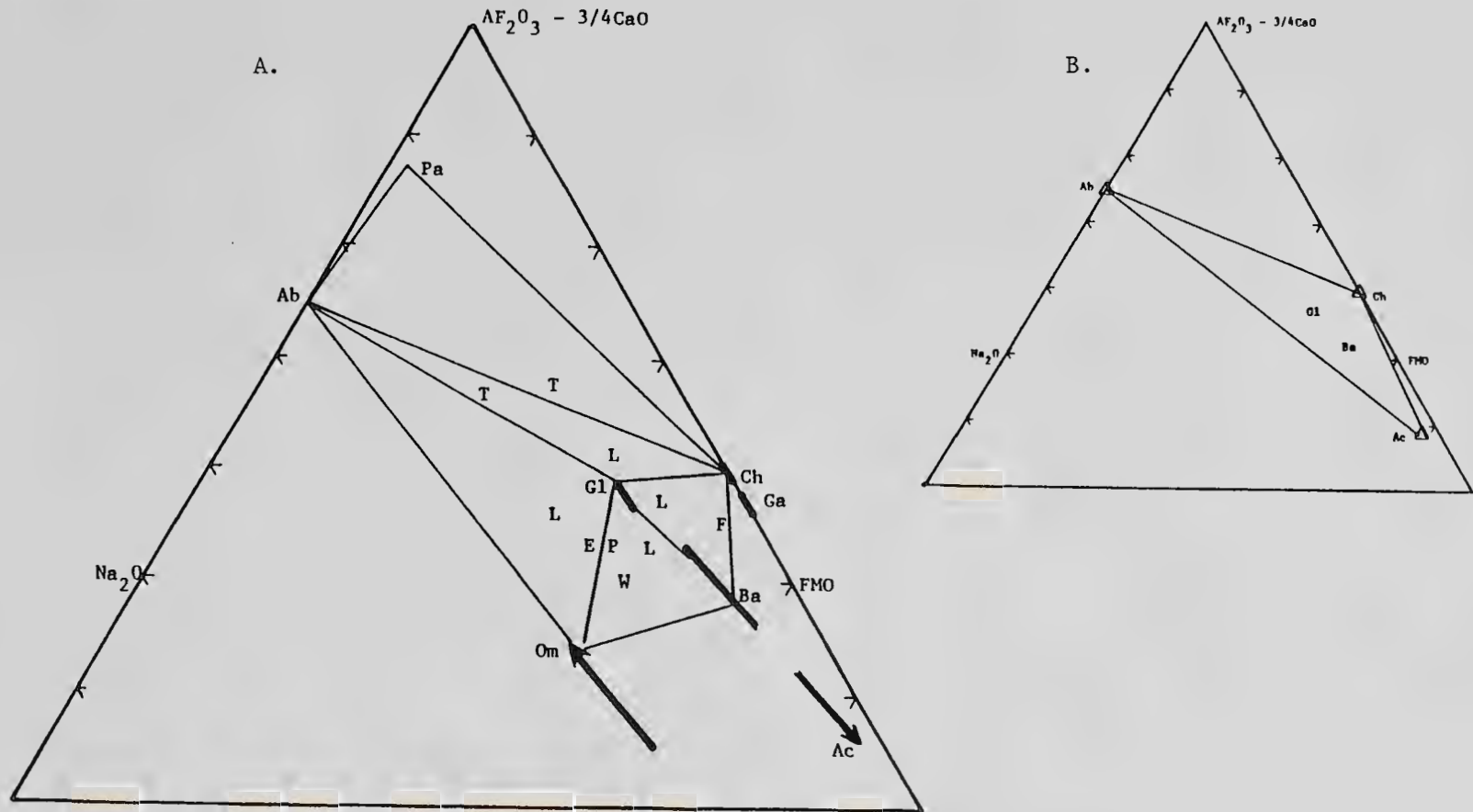


FIGURE 2. Electron microprobe mineral compositions in mafic schist projected from epidote onto $(Al_2O_3 + Fe_2O_3) - Na_2O - (FeO + MgO + MnO)$ in mol% units. Bold lines show variation in mineral composition of chlorite (Ch), glaucophane (Gl), omphacite (Om), barroisite (Ba) to actinolite (Ac), and garnet (Ga). Arrows indicate core-to-rim zoning in Om and Ba to Ac. (A) shows mineral assemblages from: E (Eclogite Brook), F (Frank Post Trail), L (Lockwood Brook), P (Tillotson Pond), T (Tillotson Peak), and W (west of Long Trail, fig. 1). Tie lines to garnet are omitted for clarity. (B) shows mineral compositions from Calavale Brook. Gl and Ba are glaucophane and barroisite compositions for reference. Other abbreviations: Ab (albite), Pa (Paragonite).

Kbar to 590° C, 13.5 Kbar. Omphacite + garnet assemblages from Traverse 1 give 520° C, 12 Kbar to 540° C, 12.5 Kbar. A lower temperature is given by garnet and omphacite at W (fig. 1): 440° C, 10.5 Kbar to 500° C, 11.5 Kbar (fig. 3). These data imply that the highest grade of metamorphism was perhaps 50° C and 1 Kbar greater at Lockwood Brook than at W and at Eclogite Brook than at Lockwood Brook.

Pelitic schist intercalated with feldspathic metawacke west of Eclogite Brook within the same (or similar) coarse-grained epidote layer seen at Eclogite Brook is composed of muscovite + quartz + garnet + paragonite + glaucophane (altered to fine-grained symplectite) + chlorite + epidote. Chloritoid occurs as inclusions in garnet only. The minimum temperature of metamorphism given by the reaction chloritoid + albite = paragonite + garnet + H₂O is about 440° C at 10 Kbar (Ghent et al., 1987). Kyanite is not observed, indicating that the maximum temperature of metamorphism is less than that for the reaction chloritoid + quartz = kyanite + garnet + H₂O (567° C at 10 Kbar; Ghent et al., 1987). Chloritoid pseudomorphed by white mica and garnet pseudomorphed by chlorite occur in metawacke seen on Traverse 2 (stop 8).

The pressure of metamorphism is constrained by the reaction paragonite + chlorite = chloritoid + glaucophane. Brown and Forbes (1986, fig. 14) show this reaction at about 13 Kbar, 450° C and 15 Kbar, 550° C. If chloritoid and glaucophane coexisted, the pressure of metamorphism was above this reaction and the metamorphism was similar to high-pressure metamorphism seen in the Sesia Zone, Italy, and the island of Sifnos, Greece (see the compilation of mineral assemblages by Brown and Forbes, 1986, fig. 14 and references therein). If chloritoid and glaucophane were not stable, the pressure was below this reaction and the metamorphism was similar to that seen in the omphacite-zone metasedimentary rocks from New Caledonia described by Ghent et al. (1987 and references therein). Glaucophane + chloritoid do coexist in pelitic schist from the Long Trail north of Tillotson Pond.

Within the errors caused by interpreting equilibrium mineral compositions in retrograded rock, by using mineral equilibria calculated for minerals of different composition from the Tillotson area (pelitic schist and metawacke) and by having to use experimental data obtained at much higher temperature and pressure (garnet-clinopyroxene geothermometer, mafic rocks), the mafic, pelitic, and felsic rocks were metamorphosed at the same pressure and temperature.

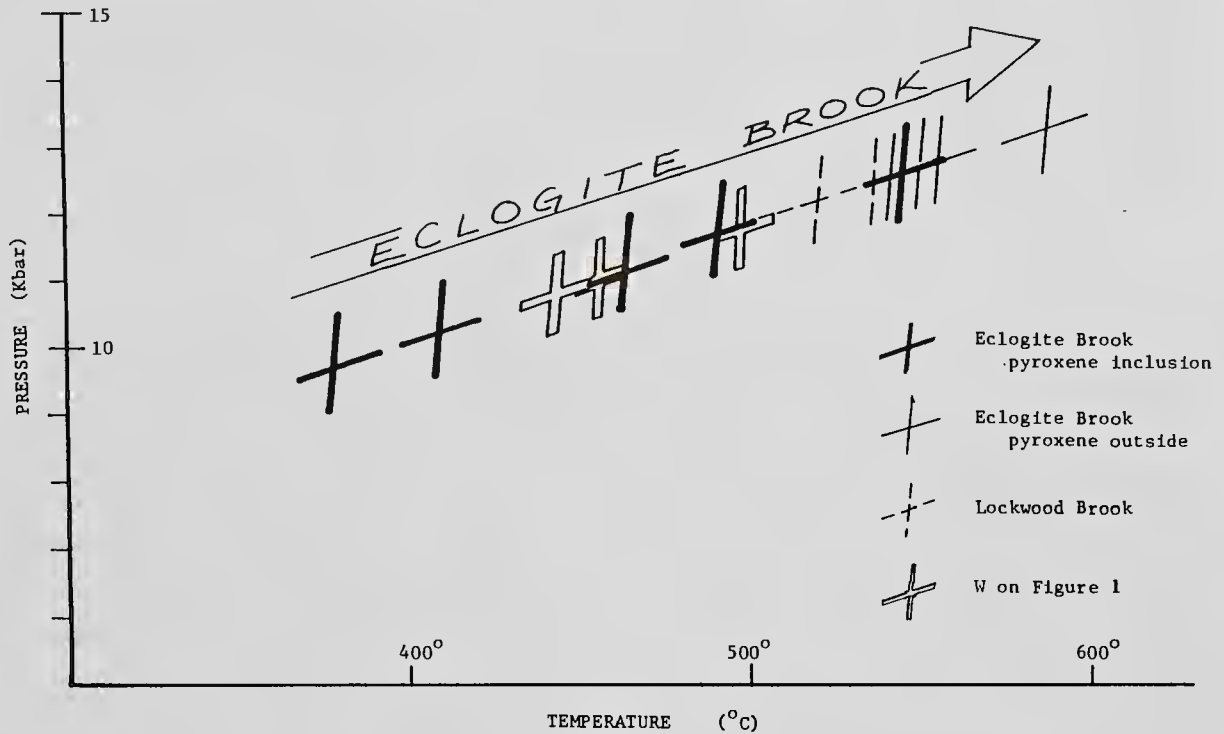


FIGURE 3. Intersection of garnet - clinopyroxene geothermometer of Ellis and Green (1979) and isopleths on the albite = jadeite + quartz geobarometer of Holland (1979, 1980). Arrow for Eclogite Brook is based on coexisting compositions obtained by electron microprobe analysis of omphacite included in garnet and of omphacite outside of, but adjacent to, garnet.

West of Hazens Notch and Tillotson in structurally higher rocks glaucophane is rare and amphibole is generally actinolite. Omphacite has not been observed. The presence of actinolite rather than barroisite or glaucophane in amphibole + chlorite + epidote + plagioclase + quartz schist indicates lower grade metamorphism from rocks to the east. Figure 2B shows that the mafic assemblage from Calavale Brook (fig. 1) could form by the breakdown of glaucophane or barroisite.

SPECULATIONS AND ARM-WAVING

Taconian high-pressure facies series mineral assemblages in mafic and pelitic schists and metawackes of the Hazens Notch Formation are preserved in a very restricted area between Tillotson Peak and Hazens Notch. Early isoclinal folding (F1) appears to correlate with the development of glaucophane + omphacite + barroisite (often in cores of garnet) in rocks of appropriate composition and record increasing P suggestive of crystallization in the down-going slab. A geothermal gradient estimate of 12 -14° C/km (fig. 3) is consistent with Newton's (1986) estimates for Paleozoic, Mesozoic, and Cenozoic subduction environments. Nearly reclined folding (F2), largely responsible for an unusual Type 2 fold interference map pattern, postdates F1 (by how much is unresolved, but probably not long) and correlates with the development of glaucophane + actinolite in the same rocks.

Rapid ascent at least through the ~500° C isotherm (to prevent complete retrogradation of glaucophane and omphacite) probably along west directed thrusts are inferred to occur within the Hazens Notch Slice in this area. The apparent replacement of high-pressure facies assemblages by medium-high (and lower) pressure facies series assemblages to the north and west of Hazens Notch and south of Tillotson Peak (including Belvidere Mountain) may reflect in part those emplacement surfaces (strongly sheared, carbonaceous pelitic schists). Alteration may also be related to the formation of the Gilmore antiform (F3) and to structurally higher positions to the west.

The origin of the east-west folds within the Hazens Notch Slice remains a delightful puzzle. Consideration of the various published hypotheses reviewed earlier in this manuscript leads us to favor the involvement of an irregular western Iapetan shelf edge underlain by Grenville basement during the final stages of Taconian closure.

ACKNOWLEDGMENTS

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ITINERARY

Assembly points: (a) Entrance to Elmore State Park, on Vermont Route 12, Elmore, Vermont at 8:30 am and/or (b) Eden General Store at the intersection of Vermont Route 100 and North Road at 9 am. Topographic maps: Jay Peak 15' or Lowell 7.5' quadrangles.

Mileage

- 0.0 Eden General Store, intersection of Route 100 and North Road
 Take North Road.

- 1.7 View of Belvidere Mountain, Eden and Belvidere asbestos
 dumps to the west. Chidester et al. (1978) present detailed
 maps and discussion of this ultramafic body. Gale (1986)
 presents alternative arguments for the emplacement of this
 body and associated mafic and pelitic rocks.

- 3.7 Entrance to the Belvidere Mountain asbestos mine (Vermont
 Asbestos Group, VAG). Dumb and open pit are obvious.
 Belvidere Mountain amphibolite is exposed beneath the fire
 tower (490+-8 Ma, Laird et al., 1984).

- 5.2 Dirt road (driveway) to the left (west) to the base of the
 Frank Post Trail.

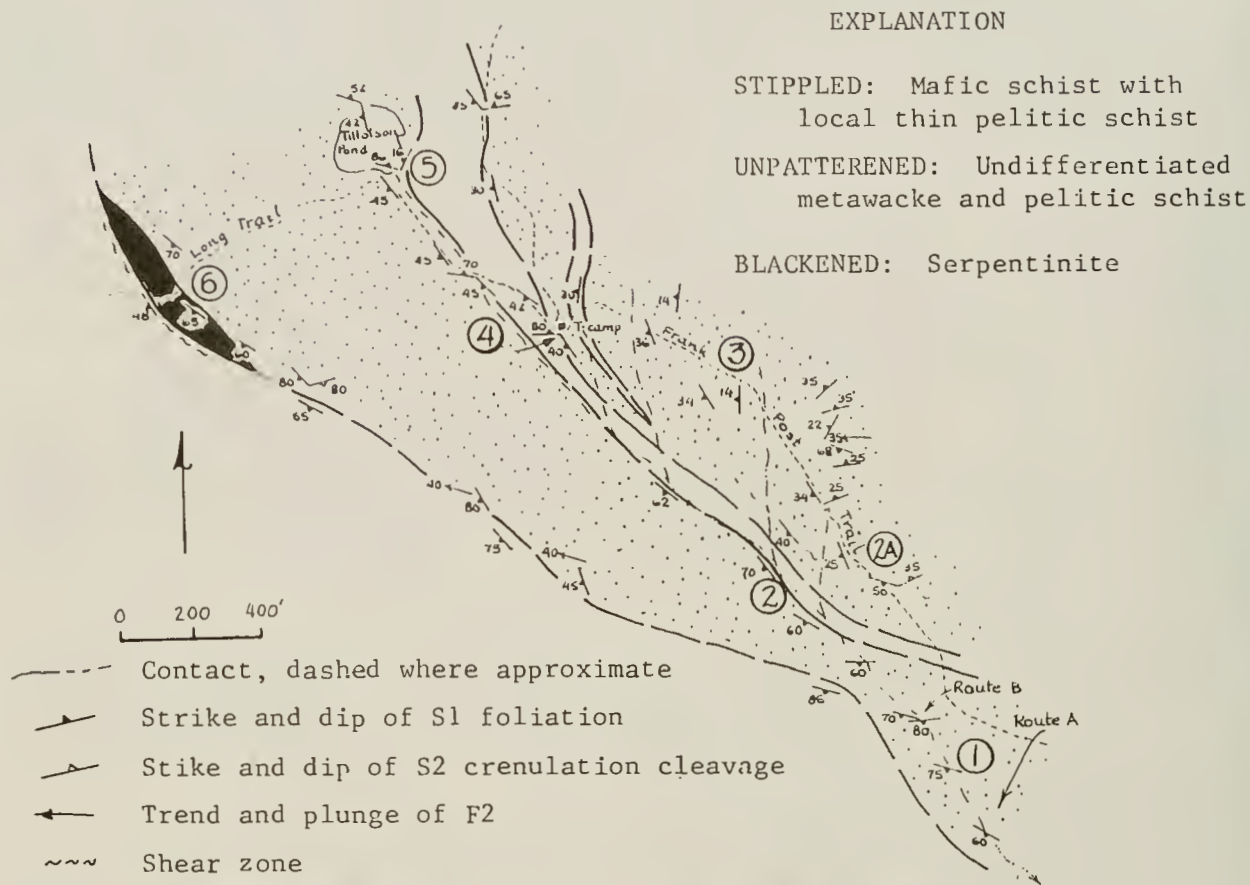
- 5.7 Park beyond Jack Stark's house on the old logging road
 heading north. The trail head west initially parallels
 Lockwood Brook. Follow the hike to Tillotson Camp and
 bring your lunch.

LOCALITY 1 - LOCKWOOD BROOK TRAVERSE

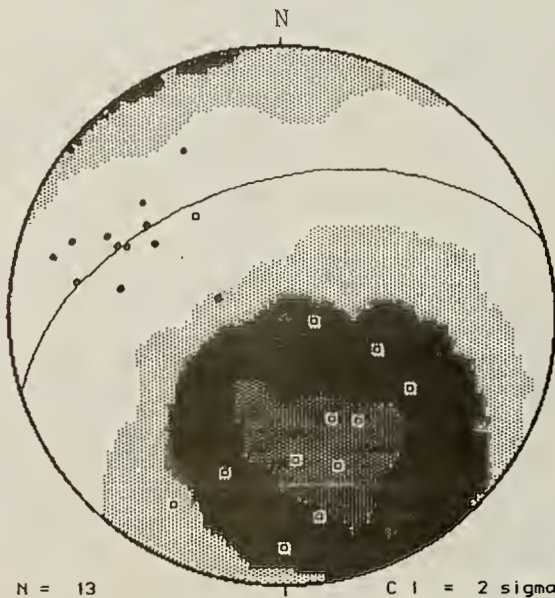
Beginning at an elevation of ~1370 ft, follow the Frank Post Trail (blue blaze) initially NW along an old logging road. The hike to the first outcrop will take about an hour. Above about 2000' outcrop is very good. The pace and compass map in Figure 4 will help locate several numbered crops discussed below in Lockwood Brook, in the Frank Post Trail, at Tillotson Camp, and along the Long Trail.

Approximate distance from base and elevation (in feet)

FIGURE 4. PACE AND COMPASS MAP OF THE LOCKWOOD BROOK AREA



S1 foliation from the Lockwood Brook area
Great circle based on F2 concentration; black square is fold axis, 32°, N76W



S2 crenulation cleavage from the Lockwood Brook area, scatter a function of position on the west limb of the Gilmore antiform. Great circle is average attitude, black dots are F2 fold axes

500 (~1410') Fork in trail. Keep right, away from the brook; follow the blue blaze.

2950 (~1650') Trail intersects and crosses a small brook.

6235 (~2000') Large tree across trail.

Continue uphill on Frank Post Trail and take either Route A or Route B to Lockwood Brook. The routes are marked for this trip.

6780 (~2100') Route A. The trail reaches a fairly open, flat area at the base of a steep section of trail. There is a small drainage coming in from the north. Two downed trees may still be apparent. A rock in the trail is marked by two arrows pointed SW. Proceed into the woods ~S20W, 330' to Lockwood Brook at the base of STOP 1

7030 (~2150') Route B. A rock with a blue blaze in the trail is marked with an arrow pointing SW and labeled 84-10. Proceed into the woods about 100' ~S20W to Lockwood Brook near the top of STOP 1

STOP 1. (~2030') Continuous outcrop in stream for about 300'. Rock is usually *very slippery*.

The rock is primarily gray-brown weathered, bluish gray-green, well layered (< 1 cm to ~10 cm), massive to foliated, fine- to medium-grained amphibolite composed of amphibole + carbonate + chlorite + epidote + garnet + plagioclase + titanite + phengite + magnetite + apatite +/- omphacite +/- pyrite + chalcopyrite. Representative mineral analyses are given in Laird and Albee (1981a, ABM100) and Laird and Bothner (1986, Table B-3A). Glaucofanite occurs in most samples and is commonly pseudomorphed by a fine-grained symplectite composed of chlorite + plagioclase + white mica + calcic amphibole. Omphacite occurs locally (eg., 25' up from the base of the exposure and at the top of the falls and is also pseudomorphed by a fine-grained symplectite.

Also present at this outcrop are minor pelitic layers. A 35 cm thick layer near the bottom of this outcrop contains garnet porphyroblasts up to a cm across. Above this layer is a ~40 cm thick layer of phengite + quartz + garnet + chlorite + titanite + magnetite + apatite +/- glaucofanite schist with black, pod-like masses flattened in the plane of foliation. These pods (clasts?) are composed primarily of magnetite + quartz + plagioclase and may represent iron-rich concretions. Epidote here and along strike is locally Mn³⁺-rich (piemontite).

Multiple periods of mineral growth and of deformation are nicely shown in this outcrop. Zoned amphibole with barroisite cores overgrown by rims zoned outward from actinolite to actinolitic hornblende record a decrease in temperature followed by an increase in temperature of metamorphism. If omphacite formed at the same time as barroisite, the first period of metamorphism is estimated to have occurred at about 500°C and 11 kbar based on garnet - pyroxene geothermometry and isopleths on the reaction albite = jadeite + quartz (Fig. 3). The compositions of barroisite and glaucophane and of actinolite and glaucophane may define miscibility gaps. One sample of pelitic schist contains inclusions of glaucophane within garnet. Fine-grained pseudomorphs may thus be after glaucophane. Alternatively, these pseudomorphs may have replaced barroisite, perhaps explaining why discontinuously zoned garnets in this sample have Ca-richer rims than core.

Over this nearly continuous outcrop, layering attitudes remain fairly constant. Early F₁ isoclinal folds occur locally on surfaces normal to layering. Warps in the mineral lineated main foliation are related to mesoscopic F₂ re-clined folds. At the top of STOP 1 layering in mafic schist is locally discordant, a feature noted at several other places in the region, that may be interpreted as a premetamorphic fault or an originally cross-cutting relationship.

Continue upstream: at ~2230', about 200' from the top of the falls at STOP 1 and ~100' SW of the brook, is an outcrop of layered, schistose feldspathic metawacke composed of albite + quartz + chlorite + epidote + white mica + magnetite + biotite with probable pseudomorphs of glaucophane. Darker colored layers are more chlorite rich, commonly with better developed S₂ crenulation cleavage than the more felsic layers. The trace of this contact, not exposed here, is marked farther upslope and on the Long Trail by an ultramafic sliver (STOP 6) that suggests a fault.

STOP 2 (~2400', about 800' from the top of the falls at STOP 1). Well exposed contact between mafic and pelitic rocks. Series of outcrops in brook of infolded steel blue-gray, massive, fine-grained amphibolite and orange-brown weathered, dark gray pelitic schist. Mafic layers include the assemblages: glaucophane + carbonate + chlorite + epidote + garnet + quartz and actinolite + glaucophane + epidote + garnet + quartz + titanite +/- chlorite, carbonate, and gold-colored sulfides. Pelitic layers contain white mica + garnet + quartz + glaucophane + titanite +/- chlorite, carbonate, tourmaline, and epidote.

The contact between mafic and pelitic schist from here upstream is everywhere parallel and assumed conformable. This pelitic schist is commonly intercalated with the mafic schists and is more or less traceable around

Peak (Fig. 1). Both rock types are folded about F2 (~EW, see Fig.) whose nearly reclined character is apparent at this stop.

Return to Frank Post Trail via a side drainage (~170', ~N20E) in which brown-weathered, sheared amphibolite is exposed. Follow the trail to about 2480'.

STOP 3 (~2480') Several outcrops of brown-weathered, gray-green, medium- to coarse-grained, massive to schistose amphibolite composed of amphibole, garnet, epidote, titanite, quartz, chlorite, and, locally, glaucophane. Amphibole is distinctly coarser grained than seen at STOPS 1 and 2. It is generally zoned with magnesio-hornblende cores and actinolite rims (F, fig. 2A). Compositions are similar to zoned amphibole seen in the tributary above STOP 2, but amphibole and chlorite are more Mg-rich and glaucophane, white mica, and magnetite are less abundant.

Several mesoscopic F2 folds are exposed in the ledges 100-200' NE of the trail that support the presence of a macroscopic WNW-plunging, approximately reclined fold (see Fig. 4). Crops here occur on the "upper" limb. On to Tillotson Camp. We will try to point out two key outcrops within about 300' of the camp where the Post Trail crosses small ledge-bounded tributaries. At the first "crossing", well-preserved (though weathered) F1 isoclinal folds are refolded about broad open warps on the upper limb of an F2 fold. Please use a camera, instead of a hammer. At the second, intercalated mafic, pelitic, and an unusual green, coarse-grained, crenulated chlorite schist are exposed.

STOP 4 Tillotson Camp (2560'; MAGNIFICENT VIEW AND SOME FINE SITTING SPOTS FOR LUNCH HERE OR AT THE POND). Outcrops at Tillotson Camp are gray weathered, medium-grained glaucophane + calcic/sodic-calcic amphibole + epidote + garnet + white mica + titanite + quartz + apatite +/- carbonate +/- plagioclase schist. The rock is well foliated and contains well developed S2 crenulation cleavage. Early F1 isoclinal folds are exposed on the east facing joint surface below the camp. Because this outcrop is right on the camp site, PLEASE DO NOT SAMPLE IT. The same mineral assemblages occur at previous stops and in other outcrops in the immediate vicinity.

Join the Long Trail "behind" the camp (near the "facilities", if needed) and walk southwesterly roughly along the upper reaches of Lockwood Brook to Tillotson Pond. Minor intercalations of pelitic schist occur within the dominant mafic schist.

STOP 5 Tillotson Pond. Cross over (with some care) an occasionally precarious beaver dam to small but important pavement outcrops on the east shore of the pond. Light gray weathering, strongly lineated schistose amphibolite containing glaucophane + epidote + garnet + chlorite + titanite + pyrite is intercalated with silvery pelitic schist and thin pale pink layers with fine-grained garnet. Pink garnet layers are considered good examples of cotecules and may represent metamorphosed Mn-chert. S₁ dips shallowly to the west; west plunging minor F₂ folds "decorate" the pavement outcrop. No hammers please.

An outcrop hiding in the brush contains epidote + glaucophane + omphacite + quartz + dolomite + calcite + titanite + magnetite + apatite + chalcopyrite (P, fig. 2A).

Time and interest permitting, we will continue another 1000' or so south along the Long Trail towards Belvidere Mountain across a number of small crops of intercalated mafic and pelitic schist until a small crop of serpentinitized ultramafic is reached (thanks in large measure to the loss of a large tree).

STOP 6 The ultramafic marks the contact between the Tillotson "mafic mass" and coarse, frequently well-layered felsic metawacke. The ultramafic is typically light brown-gray weathering, bluish-black calcareous serpentinite and is strongly sheared. To the north gray-brown weathered, blue-gray medium-grained porphyroblastic garnet amphibolite with glaucophane +- omphacite crops out. The garnets, some 1 cm across, contain glaucophane inclusions. "Structurally above" the ultramafic (to the south) medium- to coarse-grained, layered feldspathic metawacke (quartz + plagioclase + white mica + chlorite + epidote + piemontite + biotite + magnetite + apatite) is in contact with the serpentinite. The north-facing surface of this crop can be followed 50 - 60 feet along strike to the west where actinolite knots up to 25 cm across and irregular crinkled actinolite + chlorite-rich folded layers occur. We interpret this surface as a fault.

Return to Tillotson Camp and follow the Frank Post Trail to the vehicles.

- 6.2 Intersection of Stark driveway and North Road. Turn left (north)
- 9.7 Intersection North Road and Vermont Route 58, Hazens Notch Road. Turn left (west) toward Hazens Notch. Several dirt roads will join from the northside.....stay on Route 58

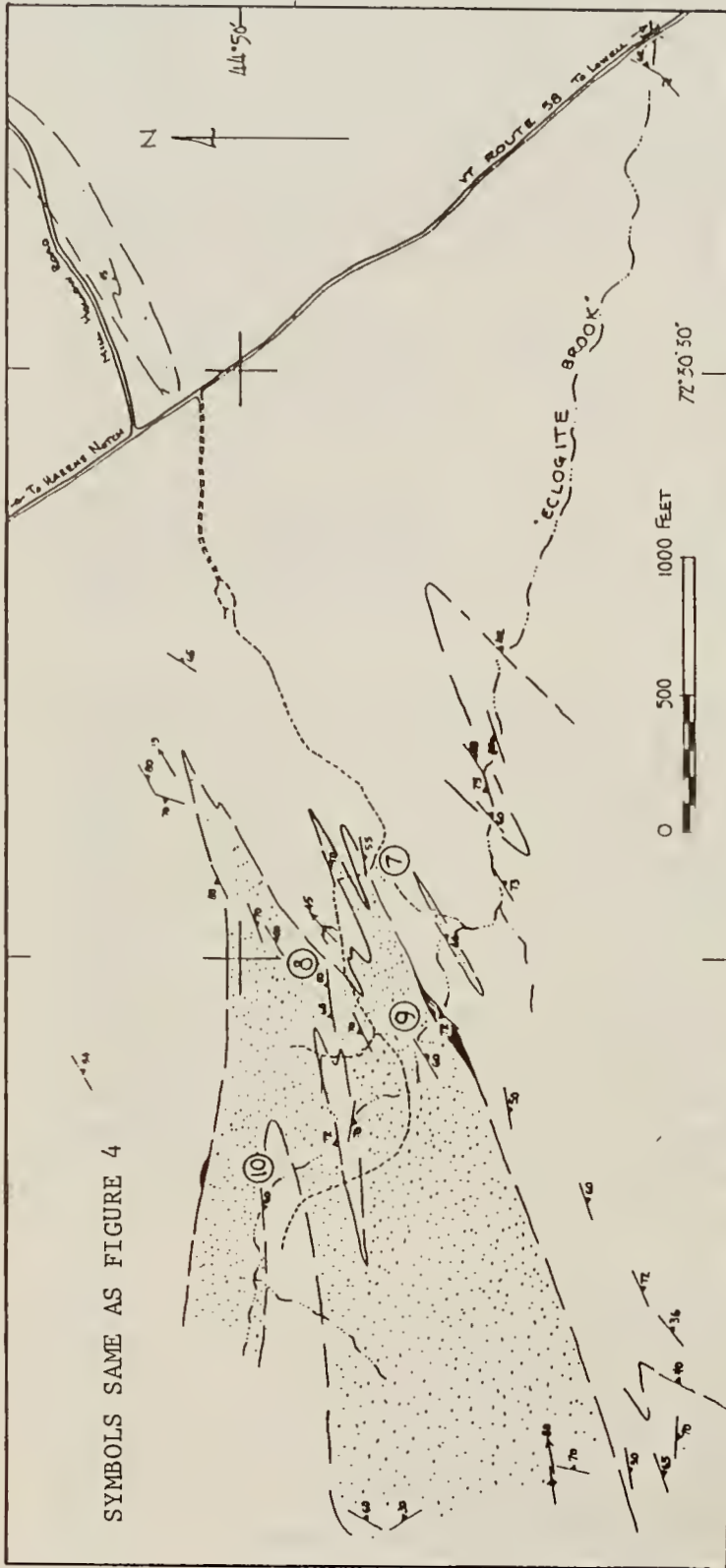
- 13.4 Turn left on logging road, just (0.05 miles) east of the Lowell/
Westfield town line and a road to the north to McAllister
Pond.
- 13.55 Park somewhere in the "staging area" at the end of the road

LOCALITY 2 -- ECLOGITE BROOK TRAVERSE

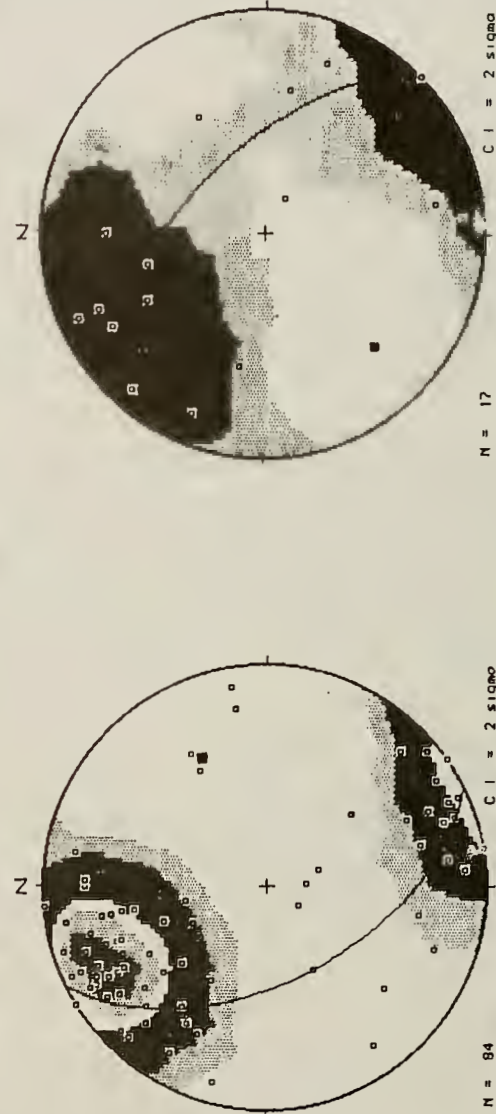
Figure 5 is a pace and compass map of the Eclogite Brook locality (informal usage, and unlikely to be adopted by local residents). We will follow the main haulage road to examine recently exhumed pavement outcrops of intercalated mafic and pelitic schists (some with large, abundant epidote porphyroblasts) and feldspathic metawacke, then follow the brook to see sheared ultramafic rock at the contact with mafic and pelitic schists, and finally layered (in a small stream outcrop) eclogite intercalated with porphyroblastic garnet pelitic schist.

STOP 7 (~1000' up the logging road, elevation ~1710') Base of nearly 200' continuous pavement consists of intercalated brown-gray weathered, medium- to coarse-grained quartz + plagioclase + garnet + gold-colored sulfide schist, with abundant quartz veins, and dark green chlorite-rich layers that are tightly folded. Garnet is altered to chlorite. Shallow dipping, isolated boudin-like masses (canoes) of thinly layered mafic schist occur within this rock. Epidote knots 5 -15 cm long parallel S_1 . The mafic layers are generally brown weathered, blue-gray, fine-grained, massive to weakly layered with abundant magnetite euhedra. A coarse-grained layer composed primarily of epidote and amphibole poses the question of a layered mafic intrusion as possible protolith, but there remains no obvious discontinuity. The amphibole is barroisite with altered cores of probable glaucophane. Coarsely crystalline epidote-rich rock (J, fig. 1) is also exposed just east of the Haystack and is considered correlative with the rock here.

STOP 8 (~1370', elevation ~1775'; about 100' along a due west stretch of the road) Blasted "roadcut" and pavement crop of brown-gray weathered, light gray, medium-grained white mica + porphyroblastic plagioclase + quartz + chloritoid (in thin section) schist with garnet porphyroblasts partially altered to chlorite. Early layering (S_0 ?) is neutrally folded and cut by clear AP cleavage, which we interpret here as S_1 because of its parallelism with the contacts with nearby mafic rocks. Plunges to the east and northeast are common in this area. Open flexural F_2 folds involving both mafic and felsic rocks occur a short



SYMBOLS SAME AS FIGURE 4



N = 84
C I = 2 sigma
S1 Foliation from the Eclogite Brook area
Great circle based on F2 concentration
Black square is fold axis, 35°, N65E

N = 17
C I = 2 sigma
S2 crenulation cleavage from the Eclogite Brook area
Cleavage fan interpreted as F3 (the Gilmore antiform),
here plunging moderately southwest (black square)

FIGURE 5. PACE AND COMPASS MAP OF THE ECLOGITE BROOK AREA

distance to the northeast in the woods and are correlated with the reclined folds at locality 1.

The contact with very fine-grained, mafic schist with small quartz pods and very thin (<5mm) light green epidote layers (laminations?) occurs in the logging road about 20' to the west. The foliation in both units is parallel. This unit is traceable along strike into Eclogite Brook.

At the crest of the road, where a less obvious skidder trail enters from the north, is an outcrop of thinly layered, massive mafic schist with a series of climbing small-scale F₂ folds that have a mullion-like appearance. Crenulation cleavage intersections define lineation that is subhorizontal and parallel to the northeast-trending fold axes. Thin epidote-rich layers enhance the layering character of this rock. Garnets are sparse in some areas but may be concentrated in thin layers suggestive of cotecules.

Continue along the main road to the stream. Head downstream about 300'

STOP 9 (elevation ~1810') Intercalated mafic and pelitic schists in the main drainage are broadly warped about northeast-trending and shallowly plunging F₂ axes (see Fig. 5). These rocks contain the same assemblages as those exposed at STOPS 7 and 8.

Along a small tributary to the south is exposed strongly sheared talcose serpentinite that occurs approximately along the contact with the intercalated pelitic and mafic schists and more massive feldspathic metawacke exposed downstream. The serpentinite marks a similar fault boundary to that exposed on the Long Trail south of Tillotson Pond.

Return to the main haulage road (upstream) and either continue upstream along the stream or follow that logging road approximately parallel to the stream for about 800'. A red flag marks a turn north to the stream and the eclogite outcrop. Crops in the stream of dominantly mafic schist with some intercalated garnet and nongarnet bearing pelitic schist are infolded. Schistosity is folded about gentle east-plunging F₂ axes. A late right-lateral fault offsets mafic and pelitic layers by about a meter.

STOP 10 (elevation ~1920') Small outcrops of type C eclogite and porphyroblastic garnet pelitic schist. Brown weathered, silvery-gray, medium-grained pelitic schist with garnet porphyroblasts 1 to 1.5 cm across crops out on the south side of the stream. In thin section, garnets are rotated and contain inclusions of probable glaucophane. Foliation surfaces are coarsely crenulated with subhorizontal EW lineation. The attitude is parallel to foliation in the eclogite.

The dark gray-green to brown weathered, dark blue-gray, well-layered, medium-grained amphibole + garnet + white mica + omphacite schist provides important information on the pressure-temperature path of this rock (see metamorphism section). Omphacite occurs in thin green layers adjacent to garnet rich layers. Garnets are choked with inclusions of epidote, titanite, and omphacite.

The contact between the pelitic schist and eclogite is covered by about a meter of stream debris. The foliation in the two is, however, parallel and we consider this to be a conformable contact. To the south the pelitic schist is intercalated with mafic schist. To the north sparse outcrops of mafic rock occur, with one small "skidder crop" of ultramafic between the last occurrence of mafic and the first occurrence of pelitic schist, a relationship not unlike that at STOP 9, or at STOP 6 earlier in the day.

The easiest route back to the vehicles is about a 100' jaunt to the south where we intersect the main haulage road, and head downhill. A few crops of folded mafic schist occur in the "trail."

Leave the parking area, turn right (east) onto Hazens Notch Road (Route 58) to the center of Lowell and Route 100 or left (west), through the Notch, and to Montgomery Center. From there, you're on your own. Have a safe trip home.

CARBONATE-CLASTIC SYNTHEMS
OF THE MIDDLE BEEKMANTOWN GROUP
IN THE CENTRAL AND SOUTHERN CHAMPLAIN VALLEY

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INTRODUCTION

The Champlain Valley is the type-locality for the Beekmantown Group (Clarke and Schuchert, 1899), originally called the Calciferous Sandrock (Eaton, 1824; Emmons, 1839, 1842, 1855; Brainerd and Seely, 1890a, b). These units are the Lower Ordovician carbonate-clastic sequence of the Appalachian miogeosyncline (Ulrich and Schuchert, 1902; Ulrich, 1911, 1913; Schuchert, 1943). Despite considerable geologic work for more than 150 years, the internal stratigraphy of the Beekmantown group is still in dispute. This field trip will visit some localities where the stratigraphic and temporal relations of the middle Beekmantown can be established (fig. 1).

Although mapped by the early New York and Vermont geologists and their respective surveys, it was not until Augustus Wing, a local schoolteacher and amateur geologist, undertook his studies of the central Champlain Valley that the importance and internal stratigraphy of the Beekmantown strata was recognized (Wing, 1858-75; Dana, 1877a, b; Seely, 1901). Unfortunately, the full results of Wing's investigations were never published; only a summary by Dana (1877a, b) and two sketch maps in Cady (1945) have appeared (the originals of the notebooks disappeared in the early 1940s [Cady, written communication to Washington, 1977]). Brainerd and Seely's (1890a, b) work, the basis for subse-



Figure 1 - Route of field trip with locations of stops.

quent Beekmantown stratigraphic systems, was mostly based on Wing's mapping and stratigraphy, although they modified it somewhat. They used Wing's stratigraphic section (after remeasuring and redescribing it) from eastern Shoreham, Vermont, (the Bascom ledges) as their type-section.

Despite the general ignorance of Wing's work, most disagreements over the internal stratigraphy of the Beekmantown Group in the Champlain Valley have revolved around Brainerd and Seely's modifications of Wing's stratigraphic system (Stockwell, 1986). Except for Flower (1964, 1968) who reestablished the paleontologic foundation for the Beekmantown stratigraphy (unknowingly confirming Wing's work), most Beekmantown studies languished or were reconstituted from Brainerd and Seely (1890a, b) (Seely, 1906, 1910; Foyles, 1924, 1927, 1928a, b; Perkins, 1908). In 1945 Cady presented a general regional synthesis which Welby (1961, 1964), Coney and others (1972), and numerous Senior Theses at Middlebury College have built upon.

Recently, detailed analysis of the Beekmantown strata in the southern Champlain Valley has led back toward Wing's original stratigraphic system (Fisher, 1977, 1984; Fisher and Mazzullo, 1976; Fisher and Wharthen, 1976; Mazzullo, 1975, 1978; Mazzullo and Friedman, 1975, 1977; Chisick and Friedman, 1982a, b; Chisick and Bosworth, 1984). This work has now been extended into the central Champlain Valley (Washington and Chisick, 1987) with minor revisions that bring the stratigraphy even more nearly into alignment with Wing's original work. This field trip will synthesize much of this recent work, especially within the complex middle Beekmantown (i.e. the Bascom Subgroup of Washington and Chisick, 1987).

Stratigraphy

Brainerd and Seely's (1890a, b) stratigraphy included all of the strata between the Potsdam sandstone and the "Trenton" (actually the base of the Chazy) in the Calciferous. Five separate lithologic divisions (the formations of Rodgers, 1937, 1955, and Cady, 1945) were defined and labelled A to E from the base upward. Divisions C and D were further divided into four units each (labelled 1 to 4 from the base upward). No stratigraphic hiatuses were recognized although distinct faunas were known (Wing 1858-1875; Whitfield, 1886, 1887, 1889, 1890a).

Clarke (1903) dropped division A with its obvious Cambrian faunal assemblage, a practice adopted by Cushing (1905), Ruedemann (1906), Ulrich (1911, 1913) and most subsequent workers (Wing considered A to be "Upper Potsdam"). Rodgers (1937), following Ulrich (1911, 1913), suggested

that the Cambrian-Ordovician boundary lies within division B; this has now been confirmed by Taylor and Halley's (1974) careful analysis of trilobite assemblages. Detailed work by Chisick (Chisick and Friedman, 1982a, Chisick and Bosworth, 1984) has confirmed Wing's (1858-1875, notebook 5, p. 8-10) conclusion that the top of the Canadian lies within division E. Thus, the type locality of the Beekmantown group shows it extending from Late Cambrian to early Middle Ordovician. Over the last two decades there has been a move to redivide C and D (Washington and Chisick's, 1987, Bascom Subgroup) into two formations comprised of C-1 - D-1 and D-2 - D-4 (Flower, 1964, 1968; Fisher and Mazzullo, 1976; Fisher, 1984). Detailed fossil analysis (Chisick and Friedman, 1982a; Chisick and Bosworth, 1984; Repetski, 1982) shows that the major disconformities and faunal breaks lie within D-1 and D-3; furthermore, the intervening strata are not lithologically divisible on a regional basis. Thus, Washington and Chisick (1987) lumped these enigmatic strata together in a synthem, the Lemon Fair formation.

The stratigraphy of the units seen on this field trip is presented in figure 2. Figure 3 is a partial presentation of the conodont assemblages obtained from these strata in the central and southern Champlain Valley (from Repetski, 1982, Chisick and Bosworth, 1984, and Chisick, unpublished data).

Regional Setting

During most of Cambrian time, the Pre-Cambrian basement (not just the Adirondack region) maintained sufficient relief that it kept the shelf restricted. By the end of the Cambrian, however, the marginal portions of the Pre-Cambrian basement had been penneplained, inundated by shallow seas and capped by an onlapping shelf sequence. From Late Cambrian into Middle Ordovician time, the Champlain Valley was the site of relatively low-energy shelf deposition dominated by carbonate with lesser amounts of generally coarser-grained, well-sorted, bimodal, quartzofeldspathic sands. Generally, the shelf strata are laminated fine-grained limestones and dolostones, calcareous quartz sands, probably pelloidal limestones, and occasional algal boundstone and oolitic limestone. The sandier sediments spread across the shallow shelf with dominantly dolomitic carbonates concentrated along the high-energy shelf margin edge. As one ascends the Beekmantown stratigraphic sequence, the amount and grain-size of clastics generally decreases. Strikingly sharp stratigraphic contrasts between clastics and carbonate facies suggest juxtaposition of dissimilar sedimentologic regimes.

M. ORDOVICIAN

SNAKE HILL FORMATION

HORTONVILLE FORMATION/
BENSON FORMATION

GLENS FALLS FORMATION Shoreham Member
 Larrabee Member

ISLE LA MOTTE FORMATION Ameterdam Member
 Orwell Member
 Lowville Member
 Pamela Member

MIDDLEBURY FORMATION "sandy"
 "homogeneous"
 "banded"

~~~~~ disconformity ~~~~~  
CROWN POINT FORMATION

~~~~~ Knox Unconformity ~~~~~

PROVIDENCE ISLAND FORMATION Bridport Member
 { E - Rhynchonella beds } Beldens Member
 Weybridge Member
 Burcharde Member

PORT CASSIN FORMATION Wing's Conglomerate Member
 { D₃₋₄ - Conglomerate & Emereon Schoolhouse Member
 upper Trilobite Beds } Sciota School Member
 Ward Member

~~~~~ disconformity ~~~~~

LEMON PAIR FORMATION/  
PORT ANN FORMATION  
    { D<sub>1-3</sub> - lower Trilobite Beds }

~~~~~ disconformity ~~~~~

CUTTING HILL FORMATION/
GREAT MEADOWS FORMATION
 { C_{1-D₁} - Ophileta Beds }

Smith Basin Member
Port Edward Member
East Shoreham Member
Kingsbury Member
Winchell Creek Member

~~~~~ disconformity ~~~~~

WHITEHALL FORMATION  
    { B - Subcrystalline Limestone }

Skens Member  
Rathbunville School Member  
Steve's Farm Member  
Warner Hill Member  
Hoyt Member  
Pinch Marsh Member  
Mosherville Member

TICONDEROGA FORMATION  
    { A - Upper Potsdam }

POTSDAM FORMATION

U. CAMBRIAN

B E E K M A N T O W N   G R O U P

L. ORDOVICIAN

Bascom Subgroup

Saratoga Springs Subgroup

Figure 2 - Stratigraphy along the field trip route.

| NORTH AMERICAN CARBONATE PLATFORM |                    |                            |                                  |
|-----------------------------------|--------------------|----------------------------|----------------------------------|
| STAGE                             | Trilobite Zonation |                            | Conodont Zonation                |
|                                   | M. ORDOVICIAN      | Vttrka                     | Zone K                           |
| Zone J                            |                    |                            | -----                            |
| Zone i                            |                    |                            | MO Founa 1                       |
| Zone H                            |                    |                            | LO Founa E                       |
| Canadian                          |                    | Zone G2                    | -----                            |
|                                   |                    | Zone G1                    | -----                            |
|                                   |                    | Zone F                     | Founa D                          |
|                                   |                    | Zone E                     | -----                            |
|                                   |                    | Zone D                     | -----                            |
|                                   |                    | Zone C                     | Founa C                          |
| L. ORDOVICIAN                     | Strophodonta       | Zone B                     | -----                            |
|                                   |                    | Zone A                     | Founa B                          |
|                                   | Strophodonta       | S. BULBOSA Subzone         | -----                            |
|                                   |                    | S. BREVISPICATA Subzone    | CLAVOHAMULUS HINTZEI Subzone     |
|                                   | Strophodonta       | M. TYPICALIS Subzone       | HIRSUTODONTUS SIMPLEX Subzone    |
|                                   |                    | M. DEPRESSA Subzone        | CLAVOHAMULUS ELONGATUS Subzone   |
|                                   | Strophodonta       | CORBINIA APOPSIS Subzone   | FRYXELLODONTUS INORNATUS Subzone |
|                                   |                    | SAUKIELLA SEROTINA Subzone | HIRSUTODONTUS HIRSUTUS Subzone   |
|                                   | Strophodonta       | SAUKIELLA JUNIA Subzone    | -----                            |
|                                   |                    | SAUKIELLA PYRENE Subzone   | PROCONODONTUS Zone               |
| UPPER CAMBRIAN                    | TELEMELEAUAN       | SAUKIA Zone                | PROCONODONTUS Zone               |
|                                   |                    | ELLIPSOCEPHALOIDES Zone    | PROCONODONTUS Zone               |
|                                   | FRANCORIAN         | IDAHOIA Zone               | PROCONODONTUS Zone               |
|                                   |                    | TAENICEPHALUS Zone         | PROCONODONTUS Zone               |
|                                   | DREBACHIAN         | ELVINIA Zone               | PROCONODONTUS Zone               |
|                                   |                    | DUNDERBERGIA Zone          | PROCONODONTUS Zone               |
|                                   |                    | APHELASPIS Zone            | PROCONODONTUS Zone               |
|                                   |                    | CREPICEPHALUS Zone         | PROCONODONTUS Zone               |
|                                   |                    | CEDARIA Zone               | PROCONODONTUS Zone               |
|                                   |                    | NO ZONATION ESTABLISHED    | PROCONODONTUS Zone               |

Figure 3 - Conodont Biostratigraphy. Conodont zonation chart for the North American carbonate platform (above) and conodont stratigraphy chart for the middle Beekmantown carbonates in the central and southern Champlain Valley (next page). Samples for conodont processing were collected on an as-needed basis throughout the study area. Primary concerns were: 1) establishing the temporal and faunal relations of the Providence Island formation, 2) clarifying the middle Beekmantown stratigraphy. Most of the conodont elements encountered in these strata in the central and southern Champlain Valley are Midcontinent Province with a secondary North Atlantic Province subset. The majority of mixed Midcontinent/North Atlantic Province samples were obtained from the Rysedorph Hill terrain. Identification of conodonts was by S. Chisick (unpublished data), with assistance by E. Landing (unpublished data) and J. Repetski (1982 and unpublished data).

## PROVIDENCE ISLAND FORMATION

## Whiterockian (Fauna I)

|                                              |                              |
|----------------------------------------------|------------------------------|
| Acontiodus coniformis                        | ?Prioniodus slagana          |
| ?oplacognathus sp.                           | Protopanderodus asymmetricus |
| Oothodus communis = Oepikodus communis       | Protopanderodus gradatus     |
| Histioidella altifrons = Histiodella sinuosa | Protoprioniodus aranda       |
| Miorozarkodina parva                         | ?Pteraoontiodus              |
| Multiostodus compressus                      | Reutterodus andinus          |
| Oepikodus quadratus                          | Scolopodus paraornuformis    |
| Oistodus multiostrugatus                     | ?Tripodus laevis             |

## Whiterockian/Canadian (Fauna I/E)

|                         |                            |
|-------------------------|----------------------------|
| Juanognathus jaanussoni | Oistodus oorrugatus        |
| Juanognathus variabilis | Oistodus longicauda        |
| Jusudontus ganada       | Rhipidognathus sp.         |
| ?Oepikodus evae         | Scolopodus acontiodiformis |

## Canadian (Fauna E)

|                                              |                        |
|----------------------------------------------|------------------------|
| Acodus delicatus                             | Oistodus pseudoramis   |
| Cristodus loxoides                           | Prioniodus sp.         |
| Drapanodus planus                            | Protopanderodus reotus |
| Juanognathus variabilis                      | Reutterodus andinus    |
| Miorozarkodina marathonsis                   | Walliseroda oomptus    |
| Oepikodus communis                           | Walliserodus sthington |
| Oistodus bransonii = Paltodus jaffersonensis |                        |

## PROVIDENCE ISLAND/PORT CASSIN FORMATIONS

## Canadian (Fauna E/uD)

|                            |                            |
|----------------------------|----------------------------|
| Aoodus deltatus (deltatus) | ?Protopanderodus leonardii |
| Disphorodus delicatus      | Scandodus pipa             |
| Drapanodus arotatus        | Scolopodus abruptus        |
| Drapanodus protus          | ?Scolopodus toomeyi        |
| Oelandodus oostatus        | Scolopodus variabilis      |
| Oelandodus elongatus       | ?Triangulodus brevicauda   |
| ?Oistodus oomptus          | ?Tropodus oomptus          |

## PORT CASSIN FORMATION

## Canadian (Fauna uD)

|                            |                          |
|----------------------------|--------------------------|
| Drapanodus concavus        | Scolopodus cornutiformis |
| Drapanodus conulatus       | Scolopoda emarginatus    |
| Drapanodus gracilis        | Scolopoda parabructus    |
| Drapanodus parallelus      | Scolopodus triplicatus   |
| Glyptoconus quadruplicatus | Ulrichodina abnormalis   |
| ?Macrodus dians            | Ulrichodina cristata     |
| Oistodus parallelus        | ?Ulrichodina viaconensis |
| Scandodus flexuosus        |                          |

## PORT CASSIN/LEMON PAIR FORMATIONS

## Canadian (Fauna uD/1D)

|                          |                             |
|--------------------------|-----------------------------|
| Drapanoistodus eubrectus | ?Protopanderodus longicauda |
| Oistodus inaequalis      | ?Scandodus furnishi         |
| ?Oistodus gracilis       | Scolopodus gracilis         |
| Oneotodus simplex        | Ulrichodina sp.             |

## LEMON PAIR FORMATION

## Canadian (Fauna 1D)

|                            |                              |
|----------------------------|------------------------------|
| Drapanodus concavus        | Scolopodus filiosus          |
| Drapanoistodus basiovalis  | ?Scolopodus quadruplicatus   |
| ?Drapanoistodus inaequalis | Scolopodus rex               |
| Histiodella donnae         | Scolopodus rex paltodiformis |
| ?Oistodus ischeguilensis   | Ulrichodina deflexa          |

## LEMON PAIR/CUTTINO HILL FORMATIONS

## Canadian (Fauna 1D/uC)

|                                              |                           |
|----------------------------------------------|---------------------------|
| Drapanodus pseudoconcavus                    | Protopanderodus elongatus |
| Drapanoistodus forceps                       | ?Rossodus highgateensis   |
| Onsotodus variabilis                         | ?Scandodus flexuosus      |
| Paraordylodia gracilis = ?Onsotodus gracilis | ?Scolopodus bolites       |

## CUTTINO HILL FORMATION

## Canadian (Fauna C)

|                      |                             |
|----------------------|-----------------------------|
| Aoanthodus lineatus  | ?Cordylodus lindstromi      |
| Acanthodus uncinatus | Iapetognathus prearengensis |
| Acodus onsotensis    | Loxodus bransonii           |
| Aoodus triangularis  | ?Paltodus bassleri          |
| Clavohazulia densa   | Rossodus manitouensis       |



The slope to base-of-slope complex was dominated by fine-grained terrigenous sediments with occasional interbeds of lime mud that was carried off of the shelf in suspension. Coarse-grained carbonate or clastic sediment was either trapped on the shelf or carried past the slope into deeper water by turbidity flows passing down submarine canyons transecting the slope and cutting into the shelf. Periodically, carbonates would build out onto the upper slope shales, producing the ragged slope-shelf boundary now represented in the Rysedorph Hill terrain.

### Sedimentation

The Beekmantown shelf (Mazzullo and Friedman, 1975, 1977; Mazzullo, 1978) is punctuated by many pure quartzofeldspathic sands and very arenaceous carbonates formed in littoral zones leading out (eastward) into generally peritidal flats of lime muds with localized siliciclastic accumulation. Algal mat remains (stromatolites) were ubiquitous. Reef-like accumulations (bioherms) of stromatolite heads and thrombolites frequently occurred. Oolitic shoals are encountered near the shelf edges and along the edges of deeper channels.

The peritidal shelf-edge shows abundant features of shallow-water deposition, e.g. dessication cracks, rip-up clasts, flat-pebble conglomerates, cryptalgal-laminites, probable tidal channels, and flaser bedding. Peritidal cycles reflect depth-controlled oscillations in sedimentation rate coupled to long-term gradual subsidence and short-term oscillation in sea-level. The deposits are shallowing-upward cycles (SUCs) (Mazzullo and Friedman, 1975, 1977; Mazzullo, 1978; Chisick and Friedman, 1982a), asymmetric units of very fine calcarenite overlain by a thicker section of coarse calcilutite. Upwards through a SUC, there is a decrease in fossil taxa, grain-size, intraclasts, and pelloids, and an increase in dolomite and fenestrae or a change from current laminates to disrupted planar algal laminates. Subaerial overprints (marked by crusts with wavy shale partings) are common near the top of middle Beekmantown SUCs.

Although dominated by peritidal carbonates, the Beekmantown does contain subtidal sequences formed during repeated drowning of the outer portions of the shelf. Regionally, cyclic peritidal sequences encase subtidal limestones and bioherms which pass landward into cyclic peritidal facies. Basal transgressive sands formed during submergence, followed by ribbon carbonates and intraclastic wave-agitated layers or mixed flats in shallow intertidal settings. Reworking is evident in abundant fining-upward layers, channel-lags, and wave-formed structures. Episodic

surges of sediment laden storm currents overwhelmed developing algal-mats with very rapidly deposited silt and sand, temporarily terminating mat growth.

The Beekmantown sediments found in the Champlain Valley attest to mild epierogenic basin deposition. Gradual subsidence during sedimentation maintained the depositional surface, producing subtle switchback regressive-transgressive packets which are time-site specific but not time-basinal correlative. Vertical sequencing of conformable lithologic units reflects lateral juxtaposition of corresponding environments, thereby implying large-scale trends involving regionally significant changes in marine conditions, thus demonstrating Walther's "Law of Facies". The asymmetry of the total vertical section must represent a more complex change in environmental/sedimentological parameters over time.

In the Champlain Valley correlations of immediately adjacent lithologies are difficult, at best, because of facies changes, repetitive lithic sequences, sedimentologic discontinuities, localized and regional diagenetic events (esp. dolomitization and dedolomitization), and structural complexities. In addition, the strata generally have sparse and biostratigraphically nondiagnostic macrofossil assemblages. Thus, the Beekmantown is difficult to unravel in the Champlain Valley.

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#### ITINERARY

The assembly point is the Williams Old Brick Store in the village of Charlotte (just west of Route 7 on Route F-5).  
Time: 8:30 a.m.

#### Mileage

- 0.0 Go south from blinker.
- 0.7 On the right is Barber Hill, an alkaline igneous intrusive complex (syenite) that cuts the Paleozoic sedimentary sequence of the Champlain Valley (Welby, 1961; Laurent and Pierson, 1973), here the Iberville (Clark, 1934) member of the Stoony Point formation (Ulrich, 1911) (Middle Ordovician). To the left is a panorama of the Red Sandrock range, a line of hills capped by Monkton quartzite (Keith, 1923) (Lower Cambrian) of the Champlain thrust sheet. The thrust trace lies at the base of the cliffs near the top of each hill.
- 1.8 Hill on right contains Burchards member (Cady, 1945) of Providence Island formation (Ulrich and Cooper, 1938) (earliest Middle Ordovician) intruded by a swarm of igneous dikes (Perkins, 1908).
- 2.0 Turn right onto Thompson's Point Road.
- 2.1 Railroad crossing.
- 2.3 Thorp Brook thrust fault. The roadcut is through Bridport member (Cady, 1945) of the Providence Island formation in the footwall.



- 2.4 Emerson Schoolhouse thrust (Brainerd and Seely, 1890a) separates the Providence Island strata from the Fort Cassin strata of Stop 1.
- 2.5 STOP 1: Emerson Schoolhouse. The roadcut just west of intersection contains excellent exposures of the Emerson Schoolhouse member (Welby, 1961) of the Fort Cassin formation (Whitfield, 1890a) (the type section is the ledges just south of road). The old Emerson schoolhouse, on northwest corner, has recently been converted into a house. The Emerson Schoolhouse member is a light bluish-grey weathering-enhanced dolomargillaceous limestone and silty quartzofeldspathic calcitic dolostone. Note the abundant desiccation and solution-collapse features, indicating episodic emergence. The nodular (?stylonodular) nature of the outcrop suggests syndepositional compaction and stretching that produced sedimentary boudins. The slight overlapping of nodules in the same horizon and folding-contortion of some nodules may indicate minor down-slope movement. Note the basal detachment surfaces, typically planar and parallel with the underlying bedding. Locally the surfaces are undulating and at times truncate underlying units. Occasional black cherty dolostone layers are interspersed within the Emerson Schoolhouse. We feel Welby's (1961) distinction of a Thorp Point member is artificial (Washington and Chisick, 1987) since the type-strata differ from the Emerson Schoolhouse only slightly in amount and character of the constituent lithologies. The top of the Fort Cassin, here at the Emerson Schoolhouse intersection, has been faulted out.

Turn around.

- 2.9 Railroad crossing.
- 3.0 Turn right (actually a double right) onto road that parallels railroad.
- 5.5 Continue straight on Greenbush Road.
- 7.5 Turn right onto Route 7. Roadcut is Larrabee member (Kay, 1937) of Glens Falls limestone and Orwell (Cady, 1945) member of Isle la Motte formation (Emmons, 1842) (Middle Ordovician).
- 9.2 Ferrisburg Four Corners. Road to right leads to Fort Cassin headland on Lake Champlain, the type locality of the Fort Cassin formation (Whitfield, 1890a; Cushing, 1905; Ruedemann, 1906) and the fauna of the Cassinian stage (Whitfield, 1886, 1890a,b,c; Foyles,

1923). It is now private property and the landowners are hesitant to allow geologists onto the rocks.

- 9.8 Outcrops on left are Glens Falls limestone.
- 10.6 Intersection with Route 22-A. Continue straight on Route 7. Outcrops on right are Crown Point limestone (Cushing, 1905) (Middle Ordovician).
- 11.4 Railroad crossing. Beldens (Cady, 1945) member of Providence Island formation in fields to left.
- 13.1 Monkton quartzite along road. This is the north end of the Buck Mountain massif, a tectonically isolated piece of the Champlain thrust sheet (Washington and Chisick, 1987; Washington, 1987c).
- 15.1 Sciota School (Fisher, 1977) and Emerson Schoolhouse members of Fort Cassin formation in field to left.
- 15.3 White Pigment's New Haven Junction plant. The rock is brought here from quarries in Middlebury and Shelburne. The Middlebury quarry is in Lemon Fair and Fort Cassin strata. The Shelburne quarry is in Shelburne marble (Lower Cambrian [Keith, 1923]). Originally, Lemon Fair strata from nearby quarries supplied the plant.
- 15.6 Turn left on Route 17.
- 16.3 STOP 2: New Haven Roadcut. Lemon Fair formation. The outcrops on the north side of the road are exposures of the upper Lemon Fair as it approaches the formational contact with the overlying Fort Cassin (which forms a few outcrops along the crest of this ridge). The Fort Cassin in this area contains Sciota School and Emerson Schoolhouse members, but the Ward siltstone (Fisher, 1977), which lies at the base farther south, is absent.

Lithologically, the upper Lemon Fair approaches the Ward member of the Fort Cassin in its increased dolomitic and sandy nature. However, the conodonts Histiodella donnae, Oneotodus simplex, Oneotodus variabilis, and Ulrichodina deflexa peg the Beekmantown  $D_2$ - $D_3$  nature of this exposure. This creates a stratigraphic problem since the Fort Cassin has been defined as  $D_3$ - $D_4$  (Cushing, 1905; Ruedemann, 1906) and is based on a major change in the fauna (Flower, 1964, 1968; Fisher and Mazzullo, 1976; Chisick and Bosworth, 1984) between Cutting Hill (Washington and Chisick, 1987, modified from Cady, 1945) and Fort Cassin. A stratigraphic solution does exist, the synthem (see Chang, 1975; Salvador, 1985).

With synthem, neither the lithologic character of the rocks that compose the unit nor their fossil content nor the time span represented figure into the definition and recognition of the new unit. This avoids excessively inflexible terminology in stratigraphic classification and recognizes the special cases that always exist in the field with unconformity bounded geometric controlled rock bodies.

The Lemon Fair is a classic synthem, bounded above and below by major regional disconformities. Although primarily composed of  $D_1$ - $D_2$  strata, the Lemon Fair does extend nearly to the top of  $D_3$ . Despite an internal hiatus that extends for nearly the entire Jeffersonian stage, there is no apparent lithologic boundary within the section. Additionally,  $D_1$  is split between the Smith Basin member (Flower, 1964) of the Cutting Hill formation and the lower Lemon Fair depending on the local position of the regional disconformity. Under ISSC guidelines, the Lemon Fair meets the special requirements of the synthem and helps to resolve a serious impediment to stratigraphic and structural studies in the Champlain Valley.

Sedimentologically, the upper Lemon Fair consists of laminated, cross-laminated, and cross-stratified dolomitic quartzofeldspathic siltstone with thin wisps of fine sandstone on argillaceous siltstone. Ripple-trains/ripple-marks are characteristic features of these beds. This exposure has been structurally deformed. Small chevron folds with associated bedding-plane shear zones and incipient cleavage abound. The dark bands and blebs within and between laminations/bedding planes show dessication and tension cracks filled with anthraxolite (i.e. spent petroleum), indicating that these beds were once excellent source beds for petroleum.

Continue.

- 17.0 Turn right (first right) by monument onto Town Hill Road. This is New Haven Village.
- 18.9 Turn left onto Route 7 and continue south.
- 20.0 Turn right onto Campground Road.
- 20.4 Railroad Crossing. The outcrops for the next couple miles are Providence Island formation and Middlebury limestone.
- 21.2 Turn left onto Pearson Road.

- 58.8 To left is view of Great Ledge and Rattlesnake Mountain. The Taconic Frontal thrust lies about at the base of the cliffs.
- 59.1 Beldens member of the Providence Island formation on right.
- 59.7 To the right lies strata ascribed to Forbes Hill conglomerate by Zen (1961), Poughkeepsie melange by Fisher (1977; 1985), and Rysedorph Hill conglomerate by Cushing and Ruedemann (1914) (these are just different names for the same unit).
- 60.7 To the right on Forbes Hill is an igneous dike-swarm cut by cross-faults in both carbonates and shales.
- 60.8 Zen's (1959) stop 2.
- 60.9 Bear left onto Main Road (east).
- 61.6 Bridge over Hubbarton River.
- 61.7 Outcrop of Bridport member of the Providence Island formation on left.
- 62.1 Turn right onto Hackadam Road (dirt).
- 62.3 Bear right. Outcrops on right are Emerson Schoolhouse member of the Fort Cassin formation and Burchards member of the Providence Island formation.
- 62.7 To right is upturned Bridport member of Providence Island formation thrust over Hortonville slate.
- 62.8 Turn right onto River Road.
- 62.9 To right by line of birch trees, note flat lying Providence Island overlying Hortonville. Hortonville has small limestone boulders and cobbles in matrix and along bedding planes.
- 63.3 STOP 6: Forbes Hill Thrust System. The rock face in the old gravel pit across the Hubbarton River (this is the south end of Forbes Hill) shows highly imbricated Burchards member of the Providence Island formation thrust over Snake Hill shale. The imbricate thrust system exhibits all the characteristics of a duplex (see Fermor and Price, 1976; Elliott and Johnson, 1980; Boyer and Elliott, 1982; Cooper and others, 1983; Washington, 1987a), although the roof thrust sheet has been removed by erosion. As was recognized by Cadell (1838) a century ago, imbrication of this sort necessitates stiff (i.e. lithified)

materials. Furthermore, the mechanics of duplex formation (Washington, 1987a) require cohesive overlying strata (for experimental confirmation, see Chamberlin and Miller, 1918). Thus, we interpret the deformation herein to have occurred in lithified materials (Chisick and Washington, in prep.), not soft sediments as Zen (1961, 1967) and Rodgers (1932) have claimed. Further confirmation of our interpretation is provided by the occurrence of fractures, cleavage, and cataclasis in both limestone and shale.

Continue.

- 63.4 Hortonville slate on right and left.
- 63.5 Sand pit with varved sand couplets of Fort Ann Stage of Lake Vermont (Chapman, 1942).
- 63.8 Turn left onto Main Road and cross bridge.
- 64.5 Bear left.
- 64.6 Hitchcock Cemetery on left.
- 64.7 To left is a thrust fault cutting Providence Island strata.
- 64.8 Thrust fault cutting Providence Island near leading edge of Forbes Hill duplex.
- 65.0 Continue straight.
- 65.1 Snake Hill shale.
- 65.5 West Haven thrust.
- 66.0 West Haven Village. Continue straight for Stop 7.
- 66.2 Orwell member of the Isle la Motte limestone (Ulrich and Schuchert, 1902).
- 66.4 Snake Hill shale on right.
- 66.6 An unnamed thrust follows the break in slope.
- 66.8 Turn right onto Burr Road.
- 66.9 Coggman (aka Codman) Creek.
- 67.1 Continue straight on Burr Road.
- 67.3 Lower Lemon Fair strata on left.

locally cherty. In the outcrops at this locality cross-beds and cross-stratification features abound and classic tidal channel features are common. Intermittently, the sand content increases to form dolomitic arenite layers. The chert, black to blue-black, occurs within the more mottled sections. The East Shoreham is a most conspicuous and superb mapping unit, especially when accompanied by the underlying Winchell Creek siltstone member (not exposed here). Ripple marks, desiccation cracks, and (most notably) Scolithus burrows are characteristic features of the Winchell Creek.

Continue.

- 40.9 Continue straight.
- 41.0 Richville dam on right, deformed Ticonderoga on left. River valley follows a thrust within the Shoreham duplex.
- 41.3 Turn left onto Shoreham-Whiting Road.
- The type section for the Beekmantown (Brainerd and Seely, 1890a) was measured in the ledges extending north from this road. Although the outcrop is excellent, it would require the rest of the day to show a good composite section in this area.
- 42.0 East Shoreham Cemetery. This is the south flank of Cutting Hill, the type section for Cady's (1945) Cutting dolostone (see also Fisher and Mazzullo, 1976) and the Cutting Hill formation (Washington and Chisick, 1987). The ledges to the west of the cemetery are the type locality for the East Shoreham member.
- 42.2 Turn right onto East Shoreham Road (also called Shoreham Depot Road).
- 42.9 Old Addison Railroad bed. To right is a covered railroad bridge, recently restored.
- 43.0 Bridge over Richville reservoir (Lemon Fair River). This marks the very southern end of the Pinnacle thrust sheet.
- 43.8 Turn right onto Royce Hill Road.
- 45.5 Magnificent view of the Adirondack front. The tectonic cause of this front, which is apparently related to recent uplift of the Adirondacks is not known.

- 45.5 Ledges of Burchards member of the Providence Island.  
 46.4 Turn left onto North Orwell Road.  
 47.1 Old quarries in Orwell member of the Isle la Motte formation and Glens Falls limestone along the side of Deignault Hill to left. These quarries show nice examples of various types of faulting.

47.3 Orwell Cemetery.

48.2 Turn right onto Route 73.

48.3 LUNCH. Park in drive in front of school.

Continue.

48.5 Turn left onto Route 22-A.

Just to the west of this intersection is the Orwell duplex described by Chisick and others (1984).

49.1 Outcrops of Orwell member of the Isle la Motte formation.

49.4 Outcrop on left is Crown Point limestone.

50.0 East Creek Marsh marks the East Creek thrust, the roof thrust of the Orwell duplex.

50.4 Outcrop on left is Glens Falls limestone.

54.6 Cliffs to left are the southwest edge of the Sunset Slice of the Taconics.

54.9 Outcrop on right is Emerson Schoolhouse member of the Fort Cassin thrust atop Hortonville slate (Keith, 1932).

North end of the Rysedorph Hill terrain.

55.8 Turn right onto Benson Landing Road.

56.0 Outcrops of Hortonville slate.

56.6 Benson Corners. Turn left onto Stage Road.

56.9 Snake Hill. Outcrop on right is Benson slate (=Hortonville), not Lower Cambrian "Bull formation" as Zen (1961) claimed.

dant vertical burrows, dessication cracks, intra-clasts, fenestrae, and other sedimentological features indicative of deposition in peritidal settings. Salinity varied and circulation was poor. Siliciclastic pulses of fine argillaceous material and minor silt-sized quartz, feldspar, and mica flakes were transported east onto tidal flats and out distal tributary channels. Temporary shut-off of the siliciclastic pulses allowed a more marine and carbonate dominant deposition to occur. The biota diversified and bioturbation (mottling) became more pronounced. Deeper conditions resulted in more limestones and massive bedding. As conditions shallowed, emergent fabrics and dolostones, both primary and dolomitized earlier sediments, formed. Bedding at this stage was controlled by precursor stratification and intensity of dolomitization dynamics.

Continue.

- 29.3 Cliffs to left are dolostones of the East Shoreham member of the Cutting Hill formation. The trace of the Weybridge thrust follows the base of the cliffs. To the right across the Lemon Fair Valley is the Snake Mountain massif containing Lower Cambrian strata of the Monkton thrust sheet. The Weybridge thrust breached the Champlain thrust sheet and superimposed the Ordovician carbonates onto the trailing edge of the Snake Mountain massif.
- 30.7 Turn left onto West Street (if you cross the bridge on Route 125, you have gone too far).
- As you drive south along West Street, note the hills on the other side of the Lemon Fair Valley (to your right). These hills constitute the type locality for Cady's (1945; Cady and Zen, 1960) Bridport dolostone, now included as a member in the Providence Island formation (Washington and Chisick, 1987; also see Ulrich and Cooper, 1938; Schuchert, 1943).
- 34.2 Turn right onto Route 74.
- 34.7 Roadcut through Lemon Fair formation.
- 35.0 Cliffs of sandy Lemon Fair strata on right; Washington and Chisick's (1987) stop 2. The Pinnacle thrust trace lies at the base of the cliff. For the next 6.5 miles the route lies within the Shoreham duplex (Washington, 1985).
- 35.1 Lemon Fair River.



- 35.8 Turn left onto Quiet Valley Road.
- 36.1 The hills to the left are Ticonderoga (Welby, 1959, 1961, after Rodgers, 1955) (Upper Cambrian) and Whitehall (Rodgers, 1937) (Upper Cambrian -- Lower Ordovician) formations; to the right is Lemon Fair formation. A thrust separates them.
- 37.7 Bascom's Ledge lies across the valley to the east. This is the type locality for the Beekmantown group (Dana, 1877a, b; Brainerd and Seely, 1890a; Clarke and Schuchert, 1899).
- 38.1 Continue straight onto Shacksboro Road at corner.
- 38.2 Bridge over Lemon Fair River. The "falls" to the right supported a small village in the 19th century. This village bore the names of Shacksbury and Newell's Mills. The strata exposed in the falls is Lemon Fair formation.
- Turn right.
- 38.6 Crown Point Road marker on the left. This road was built in 1760 (the end of the French and Indian Wars) to connect Fort Crown Point on Lake Champlain with Fort Number Four on the Connecticut River. It crossed this spot because it crossed the Lemon Fair River at the "falls" we just saw, the only feasible ford across the Lemon Fair (the name comes from the French "Limon faire" = "to make mud").
- 39.3 Cliffs to left are Lemon Fair formation. This is the west side of the Pinnacle.
- 40.4 STOP 5: The Euber Ledges. The outcrops to be visited extend down the hill from behind the house just southwest of where we stop. The "canyon" of the Lemon Fair River exposes several ledges of the upper Cutting Hill formation, mainly the Smith Basin and East Shoreham members. The uppermost member, the Smith Basin (Flower, 1964; redefined by Fisher and Mazzullo, 1976) is a massively-bedded medium dark grey limestone with wispy laminae of silt-size feldspar and quartz. Some irregular nodules of black chert are locally present in the rocks, usually replacing macro-fossils (Flower, 1964). In both Vermont and New York the Smith Basin is characterized by large solution cavities. Many of the clasts within these paleokarsts are derived from the overlying Lemon Fair formation. Immediately below the Smith Basin is the East Shoreham dolostone. This medium to thin bedded dolostone is very quartzose and

21.9 The hill to left is Crosby's (1963) station 7 and Coney and others (1972) Stop 1.

22.1 Huntington Falls Bridge.

22.2 STOP 3: Huntington Falls. Weybridge Problem. Since Cady (1945; supplemented by Cady and Zen, 1960) the Weybridge siltstone has been considered the basal member of Providence Island formation (nee Beldens, nee Bridport, nee Chipman). This section has been thought to consist of a "flat-lying" sequence of ribbon-thin limestones separated by equally thick dolofeldspathic silts. Careful observations at Huntington Falls, however, shows that the type section is structurally complex. These observations were confirmed during the recent construction of a new power station at the falls, when fresh, continuous cuts were made through the ledges south of the falls proper.

The strata above the falls are definitive "flat-lying" Beldens member of the Providence Island formation, but the falls themselves coincide with a major thrust. Imbricate thrusts form the steps in falls. The small thrust sheets contain mainly Middlebury limestone (Washington, 1987b), often overlain by Hortonville. As has been noted by prior workers in the Champlain Valley, the striped-bedding commonly observed within the limestones is not stratigraphically definitive. The exposures during construction of the new power plant showed that most of the small thrust sheets are capped with Hortonville slate. Although the stratigraphic position of some of the siltstones is not certain, most lie within the upper Middlebury and can be correlated with Washington's (1982) sandy facies of the Middlebury. Thus, the type-locality for the Weybridge siltstone is mostly not (if at all) within the Providence Island. This creates a terminological problem, as yet unresolved, since there truly is a siltstone member that lies near the base of the Providence Island formation.

Continue.

23.5 Outcrops in the woods are Beldens member of the Providence Island formation and those in the fields are primarily Middlebury limestone (Middle Ordovician) of the Sudbury duplex (Washington, 1981a, b, 1987b). This is the northern end of the duplex.

23.9 Turn right onto Hamilton Road.

- 24.9 James Pasture to right (Crosby, 1963, station 10; Coney and others, 1972, stop 2; Washington, 1987c, stop 6).
- 25.4 Turn right onto Route 23 and then immediately bear left around monument onto James Road (south).
- 27.9 Turn right onto Route 125.
- 28.1 Bear left.
- 28.4 Bear right.
- 28.8 STOP 4: The Ledges. The an excellent exposure of the Bascom subgroup (Washington and Chisick, 1987) except for the lower part of the Cutting Hill formation (Winchell Creek [Fisher and Mazzullo, 1976] and lower East Shoreham [Washington and Chisick, 1987] members), overlain by the lower Providence Island formation (primarily Burchards member). The Providence Island strata (except for a thin veneer of Burchards adjacent to the Lemon Fair contact) are highly deformed, being caught up in the Sudbury duplex. The floor thrust of the duplex lies within the lowermost Providence Island and forms the lower boundary for the major deformation.

The Fort Cassin is well-developed here with Wing's Conglomerate (Seely, 1906), Emerson Schoolhouse, and Sciota School members represented stratigraphic succession as one descends through the section. The conodonts Ulrichodina abnormalis, ?Scolopodus toomeyi, Glyptoconus quadraplicatus, and Drepandodus concavus have been identified from the middle of the section. All are good indicators of Cassinian fauna, hence the Fort Cassin formation.

The Ward member is absent from the base of the Fort Cassin here. Just as at Stop 2, the underlying Lemon Fair transcends the Jeffersonian hiatus and Fort Ann (Flower, 1964, 1968) fauna mixes with Fort Cassin fauna. The Lemon Fair is thin at here, so its contact with the uppermost Cutting Hill (Smith Basin member) can be seen near the downhill end of the roadcuts. Since the expected lithologic and faunal breaks can not be seen, this roadcut presents an excellent case for the application of the stratigraphic term synthem. Here the Jeffersonian stage is at least mostly absent, but the sedimentologic logic sequence does not reflect the break.

The Bascom subgroup deposits of mixed carbonate-siliclastic sediments began on a broad tidal flat composed of supratidal and intertidal flats which were dissected by tidal channels. These fine-grained argillaceous units contain a restricted fauna, abun-

- 67.9 Gate on right. Access path for Shaw Mountain.  
 68.3 Turn around in barnyard and proceed back along road.  
 68.6 Park on right far enough over to allow milk truck by.

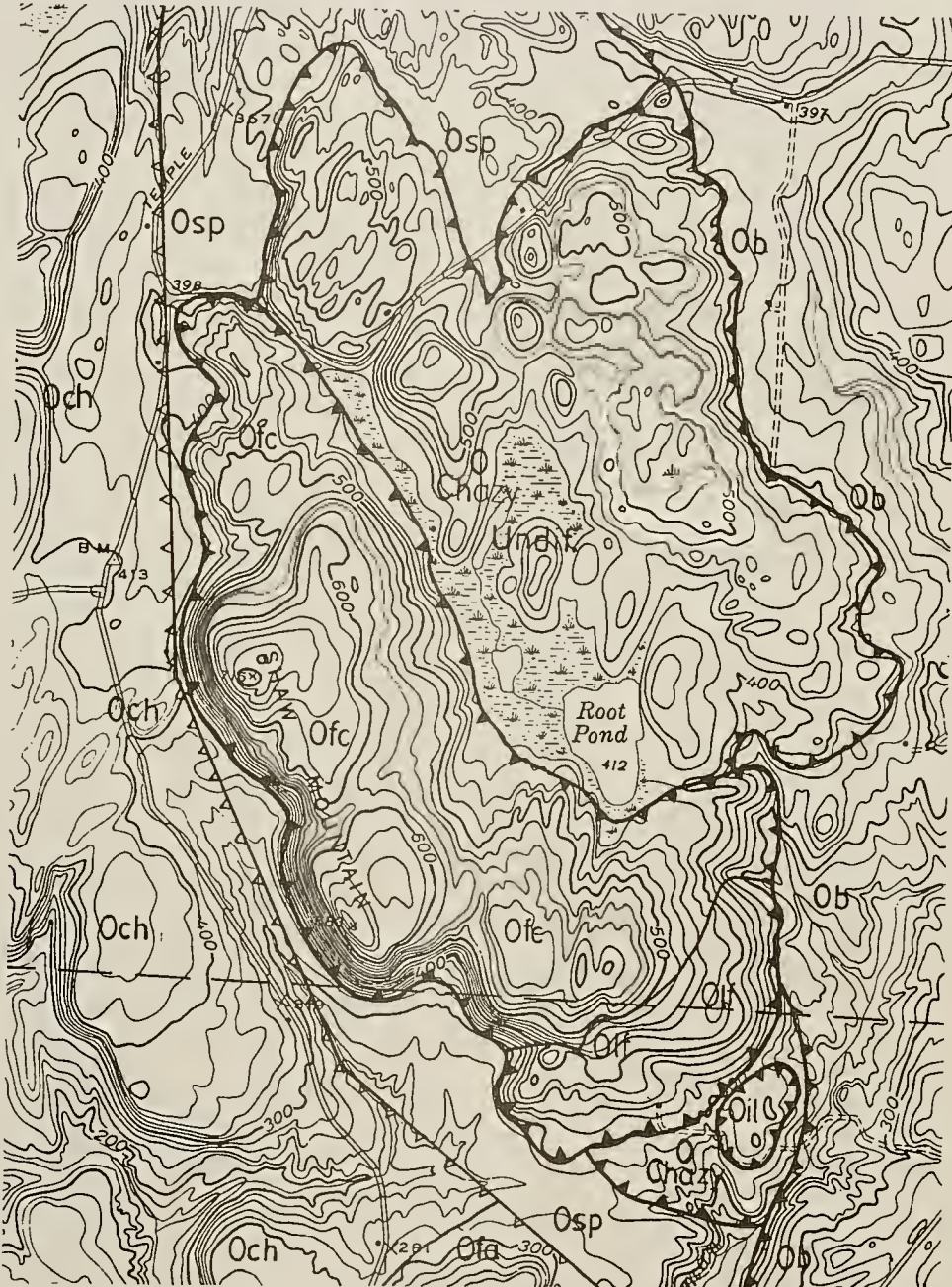


Figure 4 - Geology of the Shaw Mountain area. Paleozoic thrusts denoted by closed teeth, active high-angle reverse faults by open teeth. Och - Cutting Hill, Ofa - Fort Ann, Olf - Lemon Fair, Oil - Isle la Motte, Ob - Benson, Osp - Stony Point, O Chazy - Chazy undifferentiated.

68.7 STOP 7 (optional): Shaw Mountain Klippe (fig. 4). The Shaw Mountain klippe consists of Fort Cassin and upper Lemon Fair strata and abuts Cutting Hill and lower Lemon Fair to the west and southe. On the east, the klippe is underlain by Hortonville (called the Benson by Wheeler, 1941) slate, and on the north it is underlain by Trenton and Chazy limestones and an unnamed Chazy shale (proposed Benson Landing shale of Chisick and Friedman, 1982b) which also abuts the Cutting Hill to the west. We interpret the Cutting Hill and lower Lemon Fair block to be a modern (active) east-directed thrust sheet associated with the present uplift of the Adirondacks. The strata within the Shaw Mountain klippe is overturned, dipping about  $40^{\circ}$  east and younging westward. The underlying strata dips eastward between  $5^{\circ}$  and  $12^{\circ}$  and youngs upward. Thus, Shaw Mountain is structurally and stratigraphically discordant with the underlying rocks. The most reasonable explanation for its origin is as the recumbent limb of a fault-bend fold above a hangingwall ramp in a thrust sheet overlying the Rysedorph Hill terrain.

Shaw Mountain is probably the reason that no detailed maps have ever been published of this area. The maps published by Dale, Ruedemann, Walcott, Wheeler, Rodgers, and Zen are so generalized as to be meaningless. Recent detailed mapping by Chisick (see Chisick and Friedman, 1982b) has elucidated the stratigraphic relations, but only recently have we been able to resolve the structure into a workable model. It should be noted that Keith also mapped this area (Washington, 1987d) and correctly identified the stratigraphic relations, but he never published his findings. The Centennial Geologic Map of Vermont (Doll and others, 1961) gives a relatively detailed view of this area, but much of the stratigraphy is misidentified, as fossils found by Chisick and Keith show.

Continue.

69.7 Turn left onto Main Road.

70.4 Turn right onto Book Road.

70.6 East-west normal fault of Rodgers (1937).

70.8 Lemon Fair formation.

70.9 Note thrust at base of roadcut on left. Lemon Fair is thrust over Snake Hill shale. This thrust was not recognized by Rodgers (1937) or Rodgers and Fisher (1969).

- 71.0 Sugar house on right.
- 71.7 Bridge over Poultney River; Vermont - New York boundary. Note terminology changes.
- 72.4 Sciota Cemetery on right. Outcrops of Fort Ann (= Lemon Fair) strata.
- 72.7 Continue straight.
- 73.3 Original site of Sciota School on left. Base of type section for Sciota School limestone member of the Fort Cassin formation.
- 73.5 Sciota School member on left, Ward member on right.
- 73.6 Type locality for Ward siltstone member of the Fort Cassin formation (Fisher, 1977, 1984) named for Ward Road to right. Continue straight.
- 73.7 Fort Ann on right, Ward member of Fort Cassin on left. Here the Jeffersonian hiatus is the formation boundary so the Lemon Fair name is dropped in favor of Fort Ann (Flower, 1964).
- 74.3 Turn left onto Washington County Route 11.
- 74.6 STOP 8: Westcott Corner Thrust Model. This is Bosworth and Kidd's (1985) stop 2. The carbonate blocks within the Rysedorph Hill terrain generally consist of upper Beekmantown strata (Cutting Hill and above), with a few of Crown Point and Isle la Motte strata. All of these blocks contain very limited sections of the carbonate succession and are overlain by Snake Hill slate. Historically, these blocks have been considered olistostromes of a major melange (Cushing and Ruedemann, 1914; Wheeler, 1942; Zen, 1961, 1967, 1972a; Rodgers and Fisher, 1969; Fisher and Wharton, 1976; Fisher, 1977, 1984; Bosworth and Kidd, 1985), but their areal distribution is not random and they are all underlain by thrust surfaces, so we do not accept the old interpretation (see Leonov, 1983). Rather, we feel that their distribution into successive uni-stratigraphic rows of blocks, with the rows lying in proper stratigraphic order, indicates that these blocks are remnants of carbonate stringers built out onto upper rise shales (the Snake Hill) and were deverticulated and deformed during thrusting of this upper rise sequence onto the shelf.

As Bosworth and Kidd (1985, Stop 2) point out, this outcrop shows a block of Burchards member of the

Providence Island formation that has been internally deformed and thrust onto Snake Hill shale. This deformation probably preceded major displacement on the Mettawee River fault (the frontal thrust for the Rysedorph Hill terrain) (see Washington, 1937a, 1937c, for discussions of relative timing of thrusting and deformation). The penetrative structures within the shale and limestone and the cataclastic textures adjacent to the faults indicate that this deformation was definitely post-lithification. Thus, not only is the melange origin of this belt considered unlikely, but the major deformation did not occur in "soft sediments" as so many prior workers have claimed.

Continue.

- 74.8 Turn right onto Wescott Road.
- 75.0 Snake Hill shale on left in burrow pit.
- 75.3 On left, Burchards member of Providence Island thrust over Snake Hill shale. Major thrust in valley to right.
- 75.4 Larrabee member of Glens Falls limestone on left.
- 75.7 Isle la Motte limestone thrust over Snake Hill shale.
- 75.8 To left, Snake Hill shale capped by a klippe of Taconic slates.
- 75.9 Turn right onto Carlton Road.
- 76.1 Beldens and Burchards members of Providence Island formation.
- 76.3 To the south (left) along Mud Brook can be seen a ramp anticline with Burchards member of Providence Island formation overlain by younger limestones. Mud Brook flows along the thrust fault which formed the Mud Brook gulf.
- 76.5 Cross Bosworth and Kidd's (1985) Taconic Frontal thrust. Although this is indeed a major thrust, it lies within the Rysedorph Hill terrain and is not marked by either a lithologic or metamorphic boundary so we do not feel this should be called the Taconic Frontal thrust.
- 76.6 Bear left. Snake Hill shale on left after corner. This is Bosworth and Kidd's (1985) stop 3.

- 76.8 Carlton School on left, Carlton thrust on right.
- 77.0 Emerson Schoolhouse member of Fort Cassin formation.
- 77.1 Snake Hill shale.  
At this point we pass westward out of Rysedorph Hill terrain onto parautochthonous carbonates.
- 77.4 Turn right onto Fairhaven Turnpike.  
Just west of the corner lies the Carlton thrust, Fisher's (1984) high-angle normal fault boundary separating carbonate autochthon from the allochthonous melange.
- 77.6 Fort Ann formation on right, Sawmill Pond normal fault on left.
- 77.9 Turn left onto Buckley Road.
- 78.2 Emerson Schoolhouse member of Fort Cassin formation on left.
- 79.6 Park on left by abandoned truck scales.

Stop 9: Tri-County Stone Quarry. Exposed within the abandoned quarry are the Winchell Creek (somewhat more limey than usual) and Kingsbury members of the Great Meadows formation (approximate equivalent of the Cutting Hill formation in Vermont). These units are stereotypical of sabkha-imprinted tidal flats (Mazzullo and Friedman, 1975). Along the north-eastern rim of the quarry lies the Sciota School member of the Fort Cassin underlain by a thin layer of Lemon Fair. The limey nature of the Winchell Creek and near absence of the Lemon Fair indicate that this is along stratigraphic strike from Thompsons Point (Stop 1).

During the Lower Ordovician, several withdrawals of marine conditions temporarily exposed broad expanses of continental shelf (Braun and Friedman, 1969). These short-lived exposures are reflected in abrupt changes in types of sedimentation, wavy erosional surfaces, and karstic features. However, one wonders how abrupt is abrupt!

Along the western wall of this quarry, a marked change in sedimentation can be seen. Here the upper Winchell Creek boundary is marked by an iron-stained disconformity separating it from the overlying Kingsbury limestone. The change is abrupt and nicely preserved, but within the realized norm for geology. Along the northeastern rim, wavy erosional surfaces identify major disconformities along the top of the Kingsbury limestone (C<sub>2</sub>) of Great Meadows formation



and the base of the Sciota School member (D<sub>4</sub>) of Fort Cassin formation. Note the rolling nature of the Fort Cassin bedding. This is not thrust fault controlled, rather it is depositional and geometrically expected.

Within the upper Kingsbury and lower Lemon Fair are two sets of enigmatic vertical features. Fisher (1984, from an oral communication by S. Schammel, 1983) interprets these as "Neptunian fissures" caused by undersea earthquakes which "cracked" the limestone and filled these chemically enlarged (i.e. by instantaneous pressure solution) voids with calcarenite. We do not agree.

We interpret these vertical features to be solution-enhanced dessication cracks formed on abruptly emergent intertidal flats. The cracks and subsequent chemical enhancement must have occurred when the material was in a nearly to completely lithified state since no collapse structures have been found adjacent to these features. The "U-ed" bases of the fissures were definitely created by karstic processes, and the concave layering of the included sediments indicate they were infilled by particulate deposition from above. In addition, the cyclic nature of the fill stratigraphy indicates the material arrived by normal sedimentologic processes.

The fissures are regularly arranged, forming polygons. They reach to relatively constant depths and tend to be bulbous toward the bottom, indicating that depth was controlled by a base water table. They may have been quite deep originally, but erosion of the uplifted material would have decreased the apparent fissure depth. The fissure sets may appear to be quite large, but even larger sets have been observed in West Texas, Arizona, Nevada, and Abu Dhabi in areas where excessive lowering of the water table has occurred in relatively rapid, truly catastrophic events. Thus, we interpret these as recording two catastrophic lowerings of sea-level at some time between lower Cutting Hill and Fort Cassin time. The depth of the fissures indicates the approximate level of the lowered sea-level. From observations elsewhere, we prefer to correlate these with the karst events at the Cutting Hill/Lemon Fair formational boundary and the Jeffersonian hiatus. Finally, the fill bears strong lithologic affinity with the sandier Lemon Fair/Fort Cassin strata, but no fossils have yet been obtained from the fissure fill.

Continue.

79.7 Turn right onto Route 4.

Roadcut on right contains limey Winchell Creek siltstone member.

- 79.8 Valley marks boundary between Great Meadow formation (east) and Whitehall dolostone (west).
- 80.0 Quarry in Whitehall formation; both Skene and Steve's Farm members present.
- 80.1 Skene Mountain on right. Type section of Whitehall formation (Rodgers, 1937).
- 80.3 Ticonderoga-Whitehall formational contact on right.
- 80.4 Potsdam-Ticonderoga formational contact on right.
- 80.5 Turn left onto South Williams Street (by Armory).
- 80.7 Cross railroad tracks.
- 81.0 Bridge over Mud Brook.
- 81.4 Continue Straight.

On left, Greenmount Road; Greenmount Cemetery on hill of Whitehall (Skene member). On right, Adirondack front.

Note the topography - numerous right angle bends in streams, square hills. This is an area of modern block faulting associated with the active uplift of the Adirondacks.

- 82.0 Old cupolas from destroyed barns on right. Bear right onto Upper Turnpike Road at corner.
- 82.3 Mettawee River - following Tub Mountain fault here.
- 82.5 Upper Turnpike Road follows spur of Tub Mountain fault. On right is Fort Edward dolostone and Winchell Creek siltstone members of the Great Meadows formation. On left is Beldens member of Providence Island formation. Hill in distance is Tub Mountain, mostly Whitehall with a cap of Great Meadows.
- 83.0 Bear left on Upper Turnpike Road (paved).
- 83.2 Contact between Bridport member of Providence Island formation and Orwell member of Isle la Motte limestone. The Isle la Motte lies in core of a small syncline.
- 83.3 Bridport member of Providence Island formation.

- 33.4 Thrust fault placing Potsdam sandstone onto Providence Island formation.
- 33.8 Bear left. Dick Hyatt Road to right lies along Potsdam-Ticonderoga formational contact.
- 34.3 Mettawee River thrust, the frontal boundary of the Rysedorph Hill terrain.
- 34.7 Farmhouse on left served as C. D. Walcott's field station in 1886. On right is Whitehall underlain by Snake Hill shale.
- 35.6 Rodgers' (1952) suggested stop for private vehicles (road was unfit for buses at time).
- 35.9 Mettawee River to left follows frontal thrust of the Rysedorph Hill terrain. In this vicinity, Rodgers (1952), Sellick and Bosworth (1933), and Bosworth and Kidd (1935) have called this thrust the Taconic Frontal thrust. Sciota School and Emerson Schoolhouse members of Fort Cassin on right.
- 36.3 Intersection of Comstock Cemetery normal fault with Mettawee River thrust fault. Hill on right contains Providence Island strata; south and to left is Fort Cassin strata.
- 36.5 Turn right onto Rathbunville Road.

As we proceed along this road, we pass strata equivalent to those at Stop 4, i.e. Providence Island, Fort Cassin, and Fort Ann.

- 37.4 Turn around and park.

STOP 10: Rathbunville Road. While driving down Rathbunville Road, you have passed down section from the Providence Island and are now stopped alongside the Skene member of the Whitehall formation. Two short walks will be taken:

Walk A: Reef in the Sciota School member of the Fort Cassin formation. On the crest of the hill north of the parking area. This is an in situ domal stromatolite-thrombolite reef with various cephalopods (Cassinoceras, Tarphyceras, and Eurystonites), trilobites, and ostracods (Isochilina) can be seen in a "life-assemblage" along with an excellent paleo-karst. Please look but do not hammer!

Walk B: Tidal Flat Sedimentology. A walk west along the ledges next to the abandoned road leading through

the Steve's Farm, Rathbunville School, and Skene members of the Whitehall formation will show several sedimentological features common to all tidal-flat environments. Tempestites with flaser-bedding (very rarely preserved) can be seen. Other features seen are horsetail stylolites, laminated crusts, herring-bone cross-beds, sinusoidal ripples, and general supratidal-intertidal environments.

Proceed back to Upper Turnpike Road.

88.3 Turn right onto Upper Turnpike Road.

88.7 Bear left.

88.8 Turn left and park in lot.

STOP 11: Mettawee Falls. This stop illustrates both tidal flat sedimentation/diagenesis and thrust structures of the Rysedorph Hill front. Along the west side of the river lies a thrust sheet containing Burchard's and "Weybridge" members of the Providence Island formation. The riverbank is bedding-plane surfaces of these strata with some of the best examples of tidal-flat features found anywhere. Among the features to be seen are:

- a. mega-ripple trains;
- b. several other types of ripple marks;
- c. Liesegang-banding desiccation cracks;
- d. animal trails;
- e. teepee structures;
- f. rip-up clast conglomerates;
- g. vanished evaporite nodules;
- h. sabkha chertification;
- i. paleokarsts.

Structural features present here include:

- a. very small ramp anticlines above very small thrusts (some may be doubly blind);
- b. tension-gash jointing;
- c. incipient cleavage;
- d. tectonokarst;
- e. the Rysedorph Hill front separating multiply deformed deep-sea shales from the upper rise on the east bank from only slightly deformed, platform carbonates on the west bank.

Some of the structural and sedimentological features (e.g. small ramp anticlines and teepee structures) look very similar. Generally, however, the deformation in the carbonates is so minor that the sedimentary features are undisturbed and spectacularly exposed.

Turn right (west) onto Upper Turnpike Road when leaving parking lot.

83.9 Turn left onto Thomas Road.

Thomas Road follows the Providence Island-Fort Cassin formational contact; Fort Cassin on right, Providence Island on left.

90.0 In woods on right is another reef. This one is in the Fort Cassin formation.

90.5 Turn right onto Route 22.

For the next three miles we follow Emmons (1842) transect through the Beekmantown strata. Rodgers (1952) and Rodgers and Fisher (1969) followed this same route (from west to east) and this served as the type locality for the definition of the Canadian of Flower (1964).

91.2 Fort Ann behind Dot and At's Tavern.

STOP 12: Comstock Traverse. This is the "Calciiferous" section described by Emmons (1842). Historically, the Beekmantown units have played an important part in the development of North American geologic thought, especially in the development of American stratigraphic nomenclature. This historic traverse still presents new data that leads to new insights and interpretations.

STOP 12a: Winchell Creek siltstone member of the Great Meadows formation. This outcrop, although somewhat disturbed by thrust deformation, shows the typical Winchell Creek lithology as defined by Fisher and Mazzullo (1975). The weather-enhanced cross-bedding, herring-bone cross-stratification, and soft-sediment deformation (slumps and folds) attest to the unstable depositional and diagenetic conditions of this section of Winchell Creek. Note the flame structures and bubble-escape (blow-out) features. Disc-structures can also be seen. Does this deformation indicate that these strata are from a portion of the Winchell Creek that had built out to the edge of the carbonate platform, or did it result shock-waves from syn-depositional seismic activity? We favor the former interpretation.

Continue.

91.9 Cross Comstock Cemetery normal fault.

- 92.0 Start of relatively complete Beekmantown section on right side of road. First is Burchards member of Providence Island formation followed by "Weybridge" member, which is underlain by Wing's Conglomerate member of Fort Cassin.
- 92.4 Boundary between Fort Cassin (east) and Fort Ann (west) on right. On left are dolostones of the Ticonderoga and Whitehall formations.
- 92.7 Park in space on left of road.

STOP 12b: Comstock Traverse (continued).

1) Skene member of the Whitehall formation. An excellent example of a solution-collapse breccia. Note its deceptive appearance as a massive dolostone.

2) Skene member of the Whitehall formation. A thrombolite mound with a highly dolomitized channel to the east. From across the road, one can see the ghosts of rounded gravels that made up the channel-lag. Note the power of the dolomitizing fluids as they moved through these rocks.

3) Winchell Creek member of the Great Meadows formation. Here are carbonate sand dunes dolomitized to look like massive dolostone. Only frosted quartz grains escaped the dolomitization and outline the original sedimentary layering.

Continue.

- 92.8 Whitehall behind prison guard homes on right.
- 92.9 On right, Comstock school. On left, Comstock Prison Quarry Road (quarry still active). Boundary between Whitehall and Ticonderoga formations.
- 93.2 Potsdam outcrops on right. Great Meadows Prison on left.
- "Little boys who pick up rocks either go to prison or become geologists." (Ambrose Bearse)
- 93.5 Warden's House on left sits on Precambrian gneiss.
- 93.6 Bridge over the Champlain Canal.
- 94.0 Intersection of Route 22 and Route 4. We are sitting in the Precambrian of the Adirondacks.

End of Field Trip.

STRATIGRAPHY AND STRUCTURE OF THE SUTTON AREA, SOUTHERN QUÉBEC:  
CONSTRUCTION AND DESTRUCTION OF THE WESTERN MARGIN  
OF THE LATE PRECAMBRIAN IAPETUS.

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Introduction:

The purpose of this excursion is to introduce the reader to the Late Precambrian/Early Cambrian stratigraphy of the Oak Hill Group in southern Québec, and to evaluate the effects of the Taconic orogeny west of the Green/Sutton Mountains anticlinorial axis (GSMA).

The Sutton area has been subject to numerous studies between 1930 and 1960 (Clark, 1934 and 1936, Eakins, 1963, Osberg, 1965, and Rickard, unpubl.). These works, particularly those of Clark (1936), established the stratigraphy of the Oak Hill Group west of the Enosburg Falls/Pinnacle Mountain anticline (EFPMA, fig.1). Fossils found within the intermediate formations of the Oak Hill Group indicate a lower Cambrian age (Clark, 1936). These authors also mentioned the structural complexities that arise east of the EFPMA. Sudden change in the structural and thermal history occurs as one crosses the "Mansville Phase" (Clark, 1934). Rocks to the east, known as the Sutton Schists, have tentatively been correlated to the Oak Hill Group (Table I), although no real evidence is yet proposed.

The goals of current studies, north and south of the international border, are to re-evaluate the tectonic evolution along the southern arm of the Québec reentrant, to provide better constraints on possible stratigraphic link across the Sutton/Richford syncline (SRS) and to document the structural/thermal evolution west of the GSMA.

Stratigraphy:

The Oak Hill Group, as defined by Charbonneau (1980), includes 8 formations (fig.2). Because this excursion will primarily emphasize on the lower Oak Hill stratigraphy (rift-related volcanic/sedimentary facies), only five (5) formations will be describe below. The reader is refered to Clark (1936) and Charbonneau (1980) for descriptions of the upper Oak Hill formations.

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<sup>1</sup>: with permission of le ministère de l'Énergie et des Ressources, Québec.

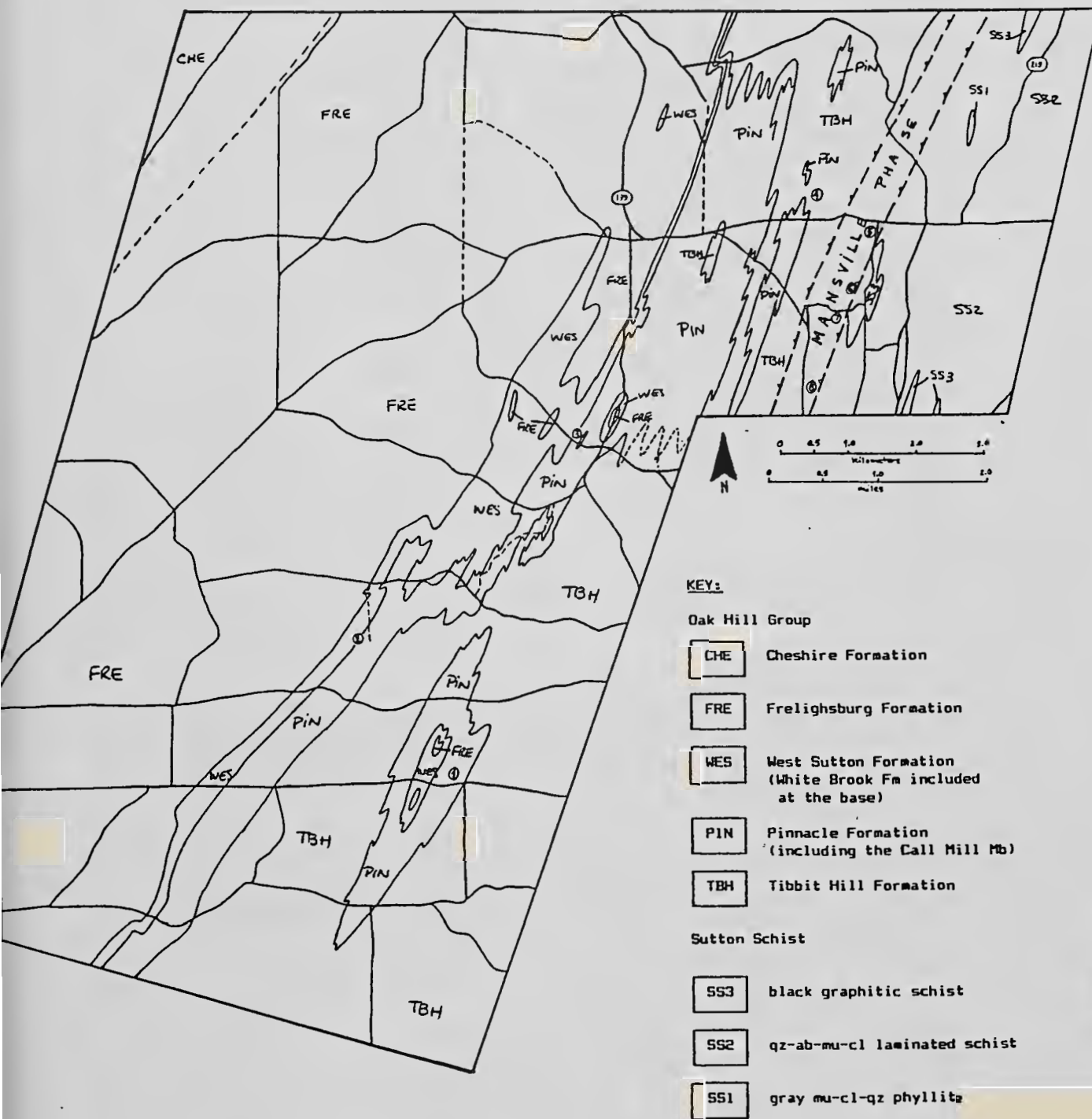


Figure 1: Geologic map of the Sutton area, southern Québec (modified after Colpron, in progress, Dowling, in progress, and Eakins, 1963).



| Clark (1934)*      |                                                                                                                         | Clark (1934)                                                            | Cady (1960) |                                      | Rickard (1965)     | Osberg (1965)                               | Clark & Eakins (1968)                           |                                    |            |
|--------------------|-------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------|-------------|--------------------------------------|--------------------|---------------------------------------------|-------------------------------------------------|------------------------------------|------------|
| PMA                |                                                                                                                         | Monts Sutton                                                            | SRS         | SMA                                  | Monts Sutton       | Monts Sutton                                | SRS                                             | SMA                                |            |
| Groupe de Oak Hill | Vail<br>Sweetsburg<br>Scottsmore<br>Oak Hill<br>Dunham<br>Gilman<br>West Sutton<br>White Brook<br>Pinnacle<br>Call Mill | Schistes de Sutton<br>incluant la Phase Mansville du Groupe de Oak Hill | Sweetsburg  | Mansville inférieur (~Ottawaquechee) | Schistes de Sutton | Sweetsburg<br>incluant Scottsmore et Dunham | Schistes de Sutton                              | Ottawaquechee                      |            |
|                    | Tibbit Hill                                                                                                             |                                                                         | Tibbit Hill | "Bonsecours"                         |                    | Schistes de Sutton                          | Gilman<br>Bonsecours<br>Pinnacle<br>Tibbit Hill | Phase Mansville (un bordure ouest) | Bonsecours |
|                    |                                                                                                                         |                                                                         |             |                                      |                    |                                             |                                                 |                                    |            |
|                    |                                                                                                                         |                                                                         |             |                                      |                    |                                             |                                                 |                                    |            |

\* la nomenclature de Clark (1934) a été retenue ici, étant celle adoptée par les auteurs mentionnés.

**Table 1:** Across strike correlation chart for the study area; PMA: Enosburg Falls/Pinnacle Mountain anticlinorium, SRS: Sutton/Richford syncline, SMA: Green/Sutton Mountains anticlinorium.

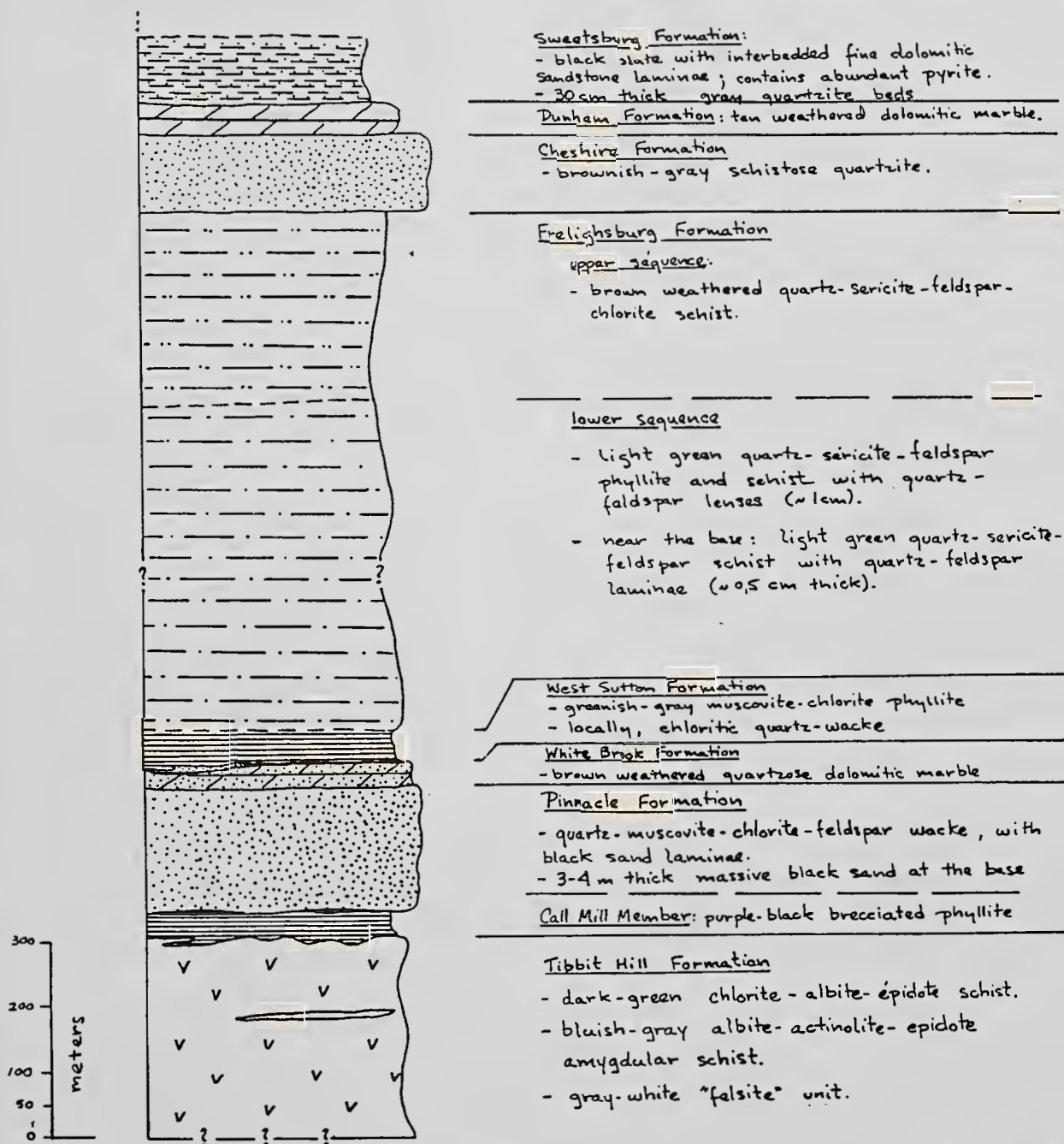


Figure 2: Stratigraphic column for the Oak Hill Group in southern Québec.

Rocks of the Oak Hill Group occupy a 15-25 kilometer wide belt that trends N20E. They lie on the western limb of the GSMA from Danville, Québec, to at least Lincoln, Vermont. The pre-Gilman (Clark, 1936) section of the Oak Hill stratigraphy is correlative with the eastern facies of the Camels Hump Group in Vermont (Table II).

In southern Québec, the Tibbit Hill Formation is the known base of the sequence. It is essentially a green chlorite - epidote - albite schist and a blue-gray amygdular schist. Chemically, the mafic rocks of the Tibbit Hill formations are metabasalts of alkalic affinities that belong to a "within-plate" tectonic setting (Pintson *et al.*, 1985, and Coish *et al.*, 1985). The presence of an albite porphyry felsic rock indicate the bimodal nature of this volcanic suite.

The Tibbit Hill is unconformably overlain by the Pinnacle Formation. Its lowermost unit is the Call Mill Member (Clark, 1936). It consists of a gray to purple-black phyllite that frequently contains phyllitic, chloritic and slaty clasts. It is often characterized by a smooth glacially polished outcrop surface. The Call Mill is laterally continuous and has a maximum thickness of 30 meters.

The coarse clastics of the Pinnacle Formation consists of two main units. The lower Pinnacle is a coarse-grained quartz - muscovite-chlorite - magnetite wacke. The abundant matrix support is the result of the degradation of feldspars and lithic fragments (of probable volcanic origin) that composed an original lithic arenite. Locally, a basal massive black sandstone is present. It is composed of 80-90 % of well-sorted and well-rounded magneto-ilmenite grains. This facies is usually 4-7 meters thick and constitute lenses of about 25-30 meters long, which transgress over coarse graywacke. The black sandstone is interpreted to be a beach sequence.

Bedforms in the lower Pinnacle Formation are generally restricted to parallel laminations and thin beds of magneto-ilmenite less than 1 cm thick, although they may be as thick as 10 cm. Slumps and loadcasts are locally present in relatively thick black sandstone beds. Slate clasts, of the same composition as the Call Mill, are found within the lower 5 meters of the Pinnacle wacke and black sandstone.

The transition between the lower and upper units is locally marked by another black sandstone horizon. Where this horizon is not found, this passage is marked by the introduction of a finer grained quartz wacke and a muscovite-rich matrix. Unlike the lower Pinnacle, there is little mixing between the magnetite and quartz grains. The above results in better defined black sandstone beds and "cleaner" quartz wacke.

Bedforms are more abundant and varied in the upper Pinnacle. Load casts and slumps develop in magnetite beds greater than 2 cm. The crest of symmetrical ripples are rarely higher than 2 cm. Larger mega-ripples are truncated by tabular cross beds which indicate a unidirectional flow. Parallel beds and laminations are numerous. Laminations can be as thin as one grain, attesting to the remarkable sorting in this unit. The above bedforms do not correspond to any stratigraphic horizon. However, channels and dolomitic lenses are restricted to the upper 7 meters of

| V E R M O N T |                     |                     |                                |               | Q U E B E C  |                     |                                                     |                                  |                       |
|---------------|---------------------|---------------------|--------------------------------|---------------|--------------|---------------------|-----------------------------------------------------|----------------------------------|-----------------------|
|               | Tauvers (1982)      | Doil (1961)         | Dennis (1964)                  | Booth (1950)  | Clark (1934) | Charbonneau (1980)* | Cooke (1962)                                        | Marquis (1985)                   |                       |
|               | Lincoln             | Centre-Sud          | Enosburg Falls                 | Nord-Ouest    | Sutton       | Sutton (W)          | Shefford                                            | Richmond                         |                       |
| 6m? — 0.1?    |                     | Clarendon Springs   | Sweetsburg                     | Skeel Corners | Vall         | Sweetsburg          | Sweetsburg                                          | Sweetsburg                       |                       |
|               |                     | Danby Winoski       | Bridgeman Hill                 | Rugg Brook    | Sweetsburg   |                     | Scottsmore                                          |                                  | Upper Mansville Phase |
|               |                     | Monkton             | Rugg Brook<br>Parker<br>Dunham | Parker        | Oak Hill     |                     | Oak Hill                                            | Scottsmore<br>Oak Hill<br>Dunham | Dunham                |
| E1.           | Dunham              | Dunham              |                                |               | Dunham       | Dunham              |                                                     |                                  |                       |
|               | Cheshire            | Cheshire            | Cheshire                       |               |              | Cheshire            |                                                     |                                  |                       |
| HADRYNIEN     | Fairfield Pond      | Underhill           | Fairfield Pond                 | Gilman        | Gilman       | Frelghsburg         | Gilman                                              | Gilman                           |                       |
|               |                     |                     | Forestdale                     | West Sutton   | West Sutton  |                     | West Sutton                                         | Lower Mansville Phase            |                       |
|               | Pinnacle            |                     | White Brook                    | White Brook   | White Brook  | White Brook         | West Sutton<br>White Brook<br>Pinnacle<br>Call Hill |                                  | Pinnacle              |
|               |                     |                     | Pinnacle                       | Pinnacle      | Pinnacle     | Pinnacle            | Pinnacle                                            |                                  |                       |
|               |                     |                     | Tibbit Hill                    | Tibbit Hill   | Call Hill    | Call Hill           | Call Hill                                           | Call Hill                        |                       |
|               |                     |                     | Tibbit Hill                    | Tibbit Hill   | Tibbit Hill  | Tibbit Hill         | Tibbit Hill                                         |                                  |                       |
| PE            | Mount Holly Complex | Mount Holly Complex | ?                              | ?             | ?            | ?                   | ?                                                   | ?                                |                       |

\* nomenclature adoptée pour ce rapport

Table 2: Along strike correlation chart for the Oak Hill Group between Vermont and Québec.

the Pinnacle Formation. The Pinnacle Formation is 150-190 meters thick.

The Pinnacle Formation is overlain by the White Brook Formation. This unit is a brown-weathering sandy dolomitic marble. Detrital magnetite is never present. The White Brook is one of the best stratigraphic markers because of its high resistance to erosion. Locally, a basal dolomitic sandstone is present in the White Brook formation. In some place, a black hematiferous slate is abundant. The White Brook has a maximum thickness of 30 meters, but is highly variable and becomes discontinuous near the international border.

Another hematiferous slate horizon (30 cm thick) is locally present at the base of the West Sutton Formation. However, the West Sutton is typically a silver- to greenish-gray phyllite. A chloritic wacke is also associated with the West Sutton Formation. This formation has a maximum thickness of 30-40 meters.

The West Sutton pass into the Frelighsburg Formation. In the course of this excursion, only the lower Frelighsburg will be observed. It is a pale green phyllite that weathers orange. The lower Frelighsburg is characterized by the presence of millimetric to centimetric quartzofeldspathic lenses. Fine euhedral crystals of magnetite or pyrite are abundant. Larger pyrite cubes are replaced by limonite, conferring a "spotty" aspect to this rock. The quartzofeldspathic lenses are the result of two successive transposition of thin silty beds.

The presence of pillow structures in the Tibbit Hill Formation indicates that at least part of the volcanic pile was extruded in a subaqueous environment. Whether the entire Tibbit Hill was subaqueous is problematical. The chemical affinity of these rocks, their bimodal nature, and the geometric distribution and thickness indicates that the Sutton area was the focal point of rifting in the Québec re-entrant during late-Hadrynian time (Kumarapeli *et al.*, 1981, Williams, 1978, and Rankin, 1976). In such an environment high heat flow will prohibit the initial subsidence of surrounding terrane. Evidence for slow subsidence is seen in the overlying clastic sequence.

The lateral continuity of the Call Mill Member indicates that volcanic activity ceased prior to Call Mill time. The upper contact is interpreted to be erosional because of its sharp nature and the presence of slate clasts in the lowermost Pinnacle Formation. The Pinnacle time represents the beginning of clastic sedimentation. The abundance of chlorite matrix in the lower Pinnacle section suggests the subaerial exposure and erosion of volcanic terranes (Tibbit Hill Formation). The peneplanation of the volcanic terranes occurred during middle Pinnacle time, as indicated by the scarcity of matrix in the upper unit.

Several lines of evidence suggest a static, shallow water, high energy environment through Pinnacle time. The presence of heavy minerals and the deposition of these as beach placers is found throughout the Pinnacle Formation, indicating the proximity to a shoreline. A shallow water environment is further suggested by the appearance of dolomitic lenses in the uppermost Pinnacle. The absence of "shale" horizons and the bedforms in the upper Pinnacle implies constant reworking and

winnowing of sediments.

The calcareous cement of the uppermost Pinnacle and the occurrence of dolomitic lenses indicates that there is no time gap between the reworking of the last Pinnacle sediment and the deposition of White Brook lithologies. The presence of hematiferous slate within and on top of the White Brook suggest a deeper environment. Passage into the West Sutton "shale" agree with this interpretation. The chloritic wacke associated with the West Sutton Formation probably results from storm generated reworking of Pinnacle sediments deposited in a more distal environment, attesting for the paucity of heavy minerals.

We believe that a "rapid" episode of subsidence began during White Brook time. Subsequent units record a deeper or more distal environment. In this context, the Frelighsburg Formation is interpreted as a distal turbidite, where thin silty horizons are interbedded in a shaly matrix. Transition from initial to thermal subsidence occurred during middle Frelighsburg time as indicated by the coarsening upward sequence of this formation.

The same stratigraphy is preserved within the "Mansville Phase", east of the EFPMA. However, the units are considerably thinner there (table III). Only the Pinnacle Formation presents a different aspect. It is a dark gray dirty sandstone with millimetric clean quartzite laminae. On the east side, the "Mansville Phase" is bordered by a rusty weathering black graphitic phyllite. This phyllite often contains millimetric to centimetric quartzo-feldspathic beds. Pyrite molds are common.

East of the "Mansville Phase", the rock assemblage is known as the Sutton Schist. This "group" includes some black graphitic schist (similar to those of the "Mansville Phase"), a quartzo-feldspathic gneiss (?), a quartz - albite - tourmaline meta-arkose, an albite porphyroblasts greenstone, a silver-gray muscovite - quartz schist, some laminated quartzite and a quartz - feldspar - muscovite - chlorite laminated schist (similar to the Frelighsburg). No stratigraphy of the Sutton Schist have been established yet. However, current work indicate that some of these units are continuous and may eventually lead to the definition of a stratigraphy. The aim of such work is to compare a possible stratigraphy of the Sutton Schist with the well established series of the Oak Hill Group.

Historically, correlations with the Oak Hill Group (Table I) have been supported by the low chemical maturity of metaclastics and the presence of metavolcanic and some marble horizons (Clark, 1934). The chemistry of greenstones indicates that they were extruded through a thinner continental crust than the Tibbit Hill (Coish *et al.*, 1985). Therefore, the Sutton Schist may record a later stage of rifting and should not be envisaged as time correlative with the Oak Hill, at least for the pre-drift section. However, some distal equivalent of the drift facies of the Oak Hill Group might be present in the Sutton Schist (e.g. black graphitic schist  $\approx$  Sweetsburg Formation; qz-fd-mu-cl laminated schist  $\approx$  Frelighsburg Formation ?). Constraints on environment may be provided by the presence of tourmalinite laminae in several lithologies.

TABLE 3: Comparison of thicknesses between the "normal" sequence of the Oak Hill Group (e.g. west of the EPMA) and the "Mansville Phase". Thickness in meters.

| Normal  | Lithology       | Mansville Phase |
|---------|-----------------|-----------------|
| 606 ±   | Frelighsburg Fm | ?               |
| 10-40   | West Sutton Fm  | ?               |
| 6-30    | White Brook Fm  | < 17            |
| 150-190 | Pinnacle Fm     | < 10            |
| 0.3-30  | Call Mill Mb    | 0-6             |
| ?       | Tibbit Hill Fm  | ?               |

Chown (1987) reports tourmalinite from environment subjected to partial evaporation. A similar origin for the tourmalinites of the Sutton Schist is compatible with a rift environment.

#### Structural geology:

The Sutton area has undergone three phases of deformation. The first one is represented by the  $S_1$  schistosity.  $S_1$  is best developed in phyllitic rocks. In the "Mansville Phase", the early schistosity is ubiquitous. Although  $F_1$  folds are rarely observed, the map pattern suggest the presence of an early folding event. The transposition of  $S_1$  along the  $S_2$  cleavage is probably responsible for the obliteration of the early folds at mesoscopic scale.

The dominant cleavage,  $S_2$ , is a crenulation cleavage axial planar to tight to isoclinal folds.  $F_2$  folds dominate the map pattern. Muscovite recrystallization along the  $S_2$  plane increases eastward.  $S_2$  defines a cleavage fan centered on the EFPMA (fig.3). This anticline is a second phase structure. In the "Mansville Phase" important slip along  $S_2$  results in the shearing of the limbs of  $F_2$  folds.

A late "fracture cleavage" ( $S_3$ ) is sparsely developed west of the EFPMA. To the east,  $F_3$  open folds and undulations deform anterior structures. The third phase is responsible for the formation of the GSMA. The last two phases of deformation are dated to be Taconic in age (Rickard, 1965).

The change in plunge of  $F_2$  axis, from northeast to southwest, is interpreted here as the result of the interference of the first two phases of deformation. This issue in basin and dome (type 1) and hook (type 2, Thiessen, 1986) interference patterns.

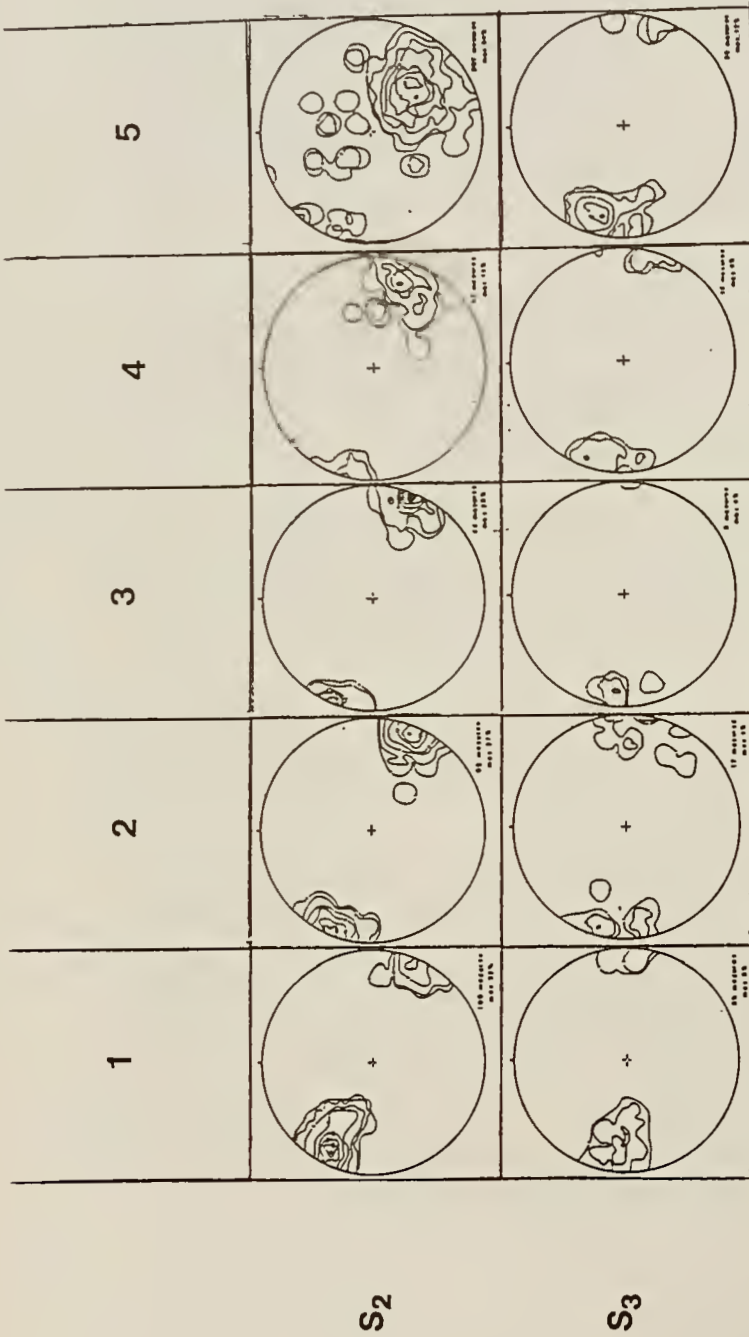
The main brittle structures in the area are present in the "Mansville Phase". In fact, this "Mansville Phase" is a fault-bounded zone within which intense shearing and stretching did occur. Fault zones are illustrated by stratigraphic truncations, fault slivers and shear fabrics (C/S). Small magnetite octahedron are often present on both side of the fault contact. The dominant faults are interpreted as second phase structure, because of the intense shearing observed along  $S_2$  and the fact that sides of shear bands are parallel with the dominant cleavage. Fault zones are not as easy to recognize within the Sutton Schist, due to the more recrystallized nature of these rocks. The best indicator of fault zone is the presence of serpentine slivers.

#### Discussion:

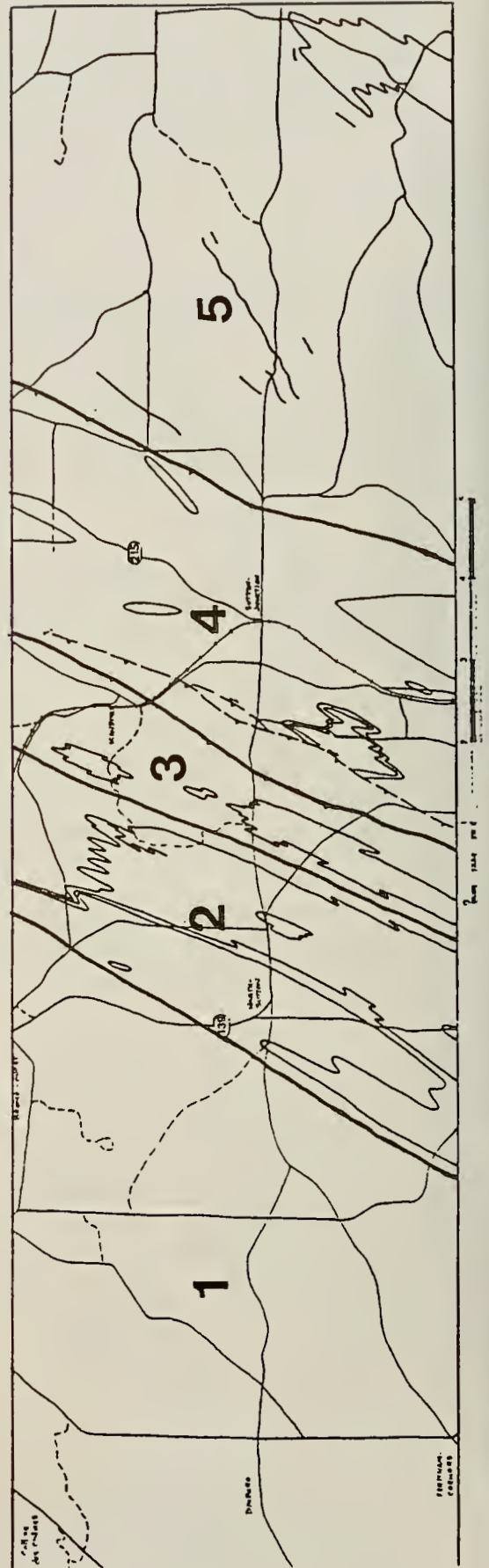
The first two phases of deformation are related with the accretionary stage of the Taconic orogen. Geophysical data suggest that the Oak Hill Group may still be rooted (St-Julien **et al.**, 1983), being



Figure 3: Equal-area stereonet of poles of  $S_2$  and  $S_3$  for the different structural domains. Contours are of 1, 2, 5, 9, 15, 20 and 30 percent per 1 percent area.



# DOMAINES STRUCTURAUX



transported on a basement slice in a later stage of accretion.

Although few indications of the first phase ( $D_1$ ) are observed in the field, we believe that  $F_1$  folds and  $S_1$  schistosity are related with the emplacement of nappes. Relative timing of the  $D_1$  episode with respect to the external domain is constrained by the fact that the  $D_1$  stage Stanbridge nappe is thrust over by the Oak Hill Group along a second phase fault (Charbonneau, 1980).

$S_2$  fanning developed contemporaneously with folds and thrusts of second generation ( $D_2$ ). This is indicated by parallelism of this cleavage with fault structures. Backthrusting along the "Mansville Phase" is interpreted to be coeval with westward thrusting at the toe of the Oak Hill "slice". Conjugate fault system appears to develop preferentially in the surficial part of the orogen. Away from the master décollement (e.g. the sole of the Oak Hill slice), the backward component becomes more important. The Sutton area is considered to lie at intermediate crustal level, where both types of structure are developed. In this context, the "Mansville Phase" is a 500 meter-wide shear zone where deformation took place by intense shearing and stretching.

Preliminary work on metamorphism supports this evolution. Rocks of the Oak Hill Group were first subject to higher metamorphic grade (upper greenschist ?) during  $D_1$  thrusting, being buried beneath the allochthons.  $D_2$  deformation brought up the Oak Hill Group in a "pop-up" fashion along the conjugate fault system, as attested by the chlorite grade, lower greenschist metamorphism. More intense recrystallization of the Sutton Schist results from deeper environment during  $D_1$  as well as  $D_2$  underthrusting.

The last phase ( $D_3$ ) records the final collision of the island-arc with the continental margin. Broad arching, in the Sutton area, exposed biotite grade rocks in the core of the GSMA.

#### Conclusion:

The lower Oak Hill Group records the early stage of rifting during the late Precambrian opening of Iapetus. Sedimentology of the Pinnacle Formation indicates a static shallow water and high energy environment, attesting to a low rate of initial subsidence. This contrasts with the sedimentology of the Pinnacle sequence in central Vermont (Dowling *et al.*, 1987). Such contrast is explained by the relative position of the basin to the focal point of rifting, where a thermal bulge is expected to form. The Sutton area is then interpreted to be in proximity to the paleo-position of the triple junction that generated the Québec reentrant.

These rocks were later involved in the three phases of deformation of the Taconic orogeny. The dominant structural features of the area result from the second phase. Deformation was concentrated within specific zone of high strain like the "Mansville Phase".

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#### Itinerary:

Assembly point is the restaurant Chez Camil, Sutton, Québec, on route 139, 10.8 km north of the Richford custom station. Parking space is available in the commuter lot, few meters to the northeast of the restaurant (fig.4). Starting time is 9:00 AM. Topographic map: Cowansville 30' quadrangle (NTS 31H/2; 1:50 000).

#### kilometers

- 0.0 From the parking lot in Sutton, take left on route 139 (south).
- 4.4 Take right on Alderbrook road.
- 5.1 Turn left on Perkins road.
- 9.3 STOP 1: 50 m after the junction of Perkins and Three Parish roads (to the left) park cars on the side of the road near the driveway that lead to the Ross farm. Take driveway to the farm house and ask permission to get in the woods in back of the farm.
- The hill back to the farm display the complete stratigraphy of the lower Oak Hill Group exposed on the overturned limb of a basin. Climb the hill from the eastern side; refer to text for details on the stratigraphy and sedimentology of the lower Oak Hill Group.
- At the end of the traverse, return to cars and continue westward on Perkins (at this point, Grande Ligne) road.
- 13.6 Take left on Russel road.
- 14.8 At the intersection of Russel and Jordan roads, turn right and then left on Dymond road (50 m).
- 17.0 Turn right on McCullough road.
- 19.6 Take right on Strobl road. Drive up to the end of this road.
- 20.2 STOP 2: Park cars off the road near the house of Kara and Gail Chaplin-Szathmary. Ask permission to get at the outcrop located in the backyard. **HAMMERS ARE PROHIBITED ON THIS OUTCROP.**

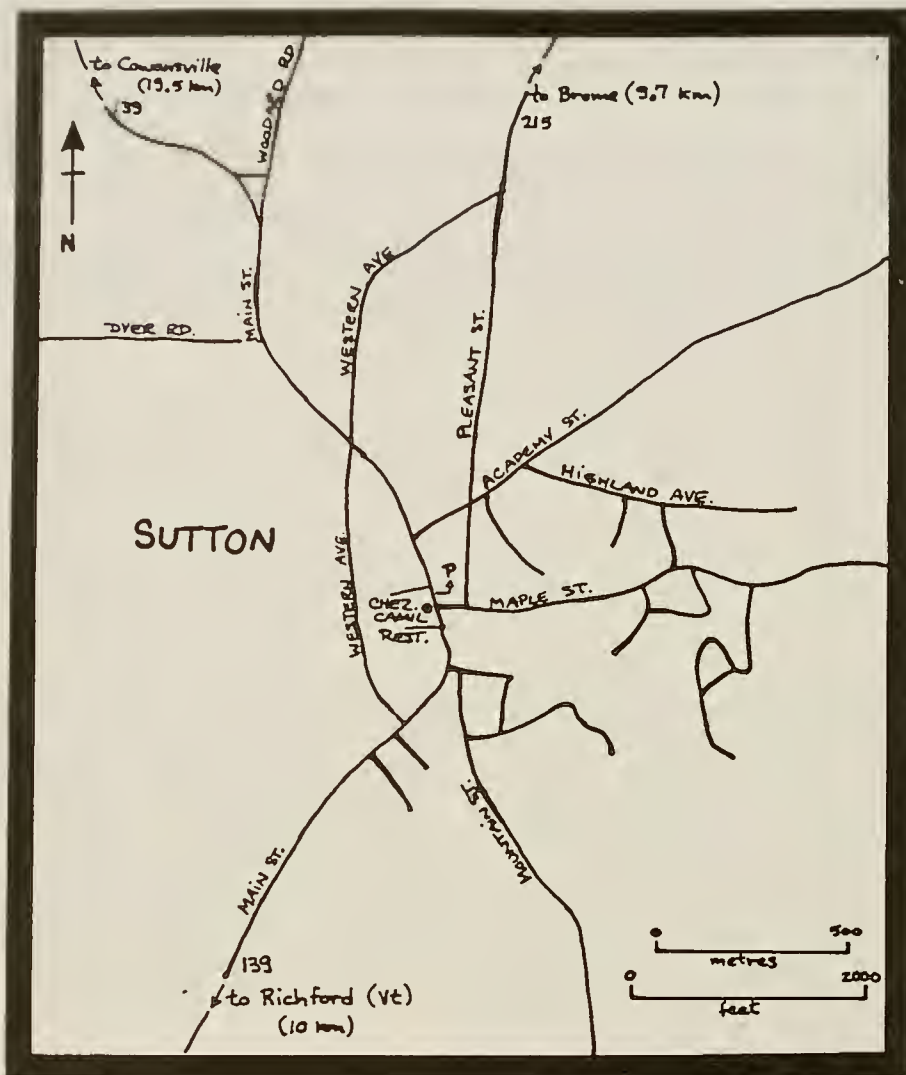


Figure 4: Detail map of the town of Sutton, Québec. The commuter parking lot is indicated by a "P".

This locality exhibits the numerous sedimentary features encountered in the upper Pinnacle Formation. Contact with the White Brook Formation is exposed at the northwest end of the yard. Note the abundance of black sandstone, the presence of dolomitic sandstone lenses in the Pinnacle and the transitional clean sandstone near the contact.

We will eat lunch at this locality. After lunch drive back Strobl road and take right on McCullough.

25.1 At the T intersection, turn left on Alderbrook road.

27.8 Take left on Macey road at the road crossing at West Sutton.

28.1 STOP 3: Park cars in front of the outcrop located on Mr. Hamel's property. This is only a short stop to illustrate the style of folding associated with  $D_2$  structures of the western limb of the EFPMA. Folds are defined by black sandstone beds of the upper Pinnacle Formation. At this locality, a small basin is cored by the White Brook and West Sutton Formations.

Going southeatward from the Pinnacle exposure, get in the White Brook Formation. Note the presence of thin "seams" of hematiferous slate and the numerous quartz veins. Then, going northeastward, get in a small open pit. The floor of the dugged area is composed of chloritoid-bearing West Sotton phyllite.

From this locality continue westward on Macey road.

31.0 Take right on North Sutton road.

33.9 Cross the intersection at North Sutton and get on route 139 eastward.

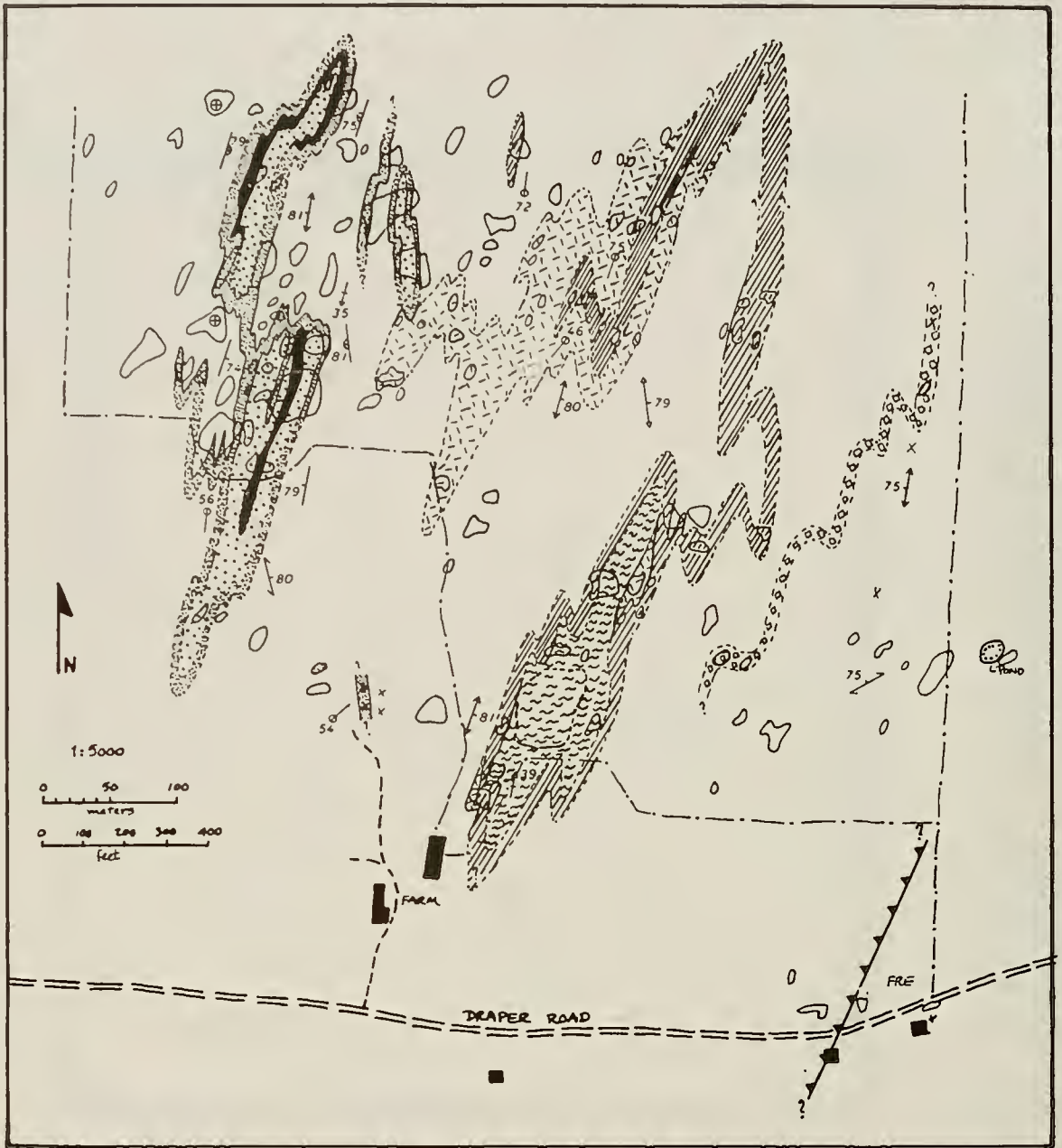
35.0 Take left on Draper road at the "Y" intersection.

36.4 STOP 4: After the end of paved road, take left on the second driveway (Hathaway Farm). Park cars on the left side of the barn. Ask permission to get in the pasture.

This locality display the various facies of the Tibbit Hill Formation. See figure 5 for the location of the different facies and particular features. The western end of the pasture exhibits a small double basin cored by the lower Pinnacle Formation.

From the Hathaway Farm, take left on Draper road. The next four stops will illustrate the structural features of the "Mansville Phase".

37.4 Turn left on Woodard road.



KEY:

- |                                                           |                   |                                             |                   |                                                   |
|-----------------------------------------------------------|-------------------|---------------------------------------------|-------------------|---------------------------------------------------|
| FRE                                                       | FRELIGHTSBURG Fm. | gray amygdule schist<br>(TIBBIT HILL Fm.)   | $\circ$ x         | outcrop.                                          |
| $Q_2$ -CL-Mt wacke<br>(lower PINNACLE Fm.)                |                   | blue amygdule schist<br>(TIBBIT HILL Fm.)   | $\oplus$          | pillows location.                                 |
| massive black sandstone<br>(lower PINNACLE Fm.)           |                   | laminated CL-Ab schist<br>(TIBBIT HILL Fm.) | $\frac{1}{2}$     | bedding plane, with top.                          |
| black phyllite<br>(CAL MILL Mb.)                          |                   | felsic rock<br>(TIBBIT HILL Fm.)            | $\rightarrow$     | early schistosity ( $S_1$ )                       |
| massive CL- <del>Ep</del> -Ab schist<br>(TIBBIT HILL Fm.) |                   |                                             | $\leftrightarrow$ | slip cleavage ( $S_2$ )                           |
|                                                           |                   |                                             | $\rightarrow$     | late cleavage ( $S_3$ )                           |
|                                                           |                   |                                             | $\rightarrow$     | $S_1/S_2$ intersection lineation<br>( $L_{1/2}$ ) |
|                                                           |                   |                                             | $\rightarrow$     | stretching lineation                              |

Figure 5: Detail geologic map of the pasture at the Hathaway Farm (STOP 4).

- 37.6 STOP 5: Park cars on the right side of the road at the end of the first field. Walk up to the second field, then, along the northern edge, reach the corner of the field. Follow the red flags up to the trail. Refer to figure 6 for outcrop location.

This locality present a small fault-bounded refolded syncline that exhibits the stratigraphy of the Oak Hill Group within the "Mansville Phase". Note the thickness of the different units.

After this stop, continue southward on Woodard road.

- 38.6 Take right on Hivernon road, then left on Godue street (38.7). Climb up the hill and turn left on Harvey street (39.1). Park cars on side of the street.

STOP 6: Walk back to the intersection of Godue and Harvey. A black graphitic schist containing centimetric quartzofeldspathic laminations is exposed in the right-side ditch.  $S_1$ ,  $S_2$  and  $S_3$  are observed in this outcrop. The same rock is exposed on both side of the house facing Godue street. On the left side, at the end of the driveway, the same structures are present. From this outcrop westward, go to the end of the yard. A few greenstone outcrops are visible. Regional mapping indicates that they are the exposed part of a fault sliver of the Tibbit Hill Formation. Here the Tibbit Hill is bounded by the Frelighsburg Formation and the black graphitic schist (a possible correlative of the Sweetsburg Formation).

From this locality, continue westward on Harvey street.

- 39.7 STOP 7: Take left on Hivernon road, then park cars on the side near a trail going south from Hivernon, 30 m before Asa Frary street. Walk down the trail. Outcrop is in the woods, on the right side, 20 m before the bee hives.

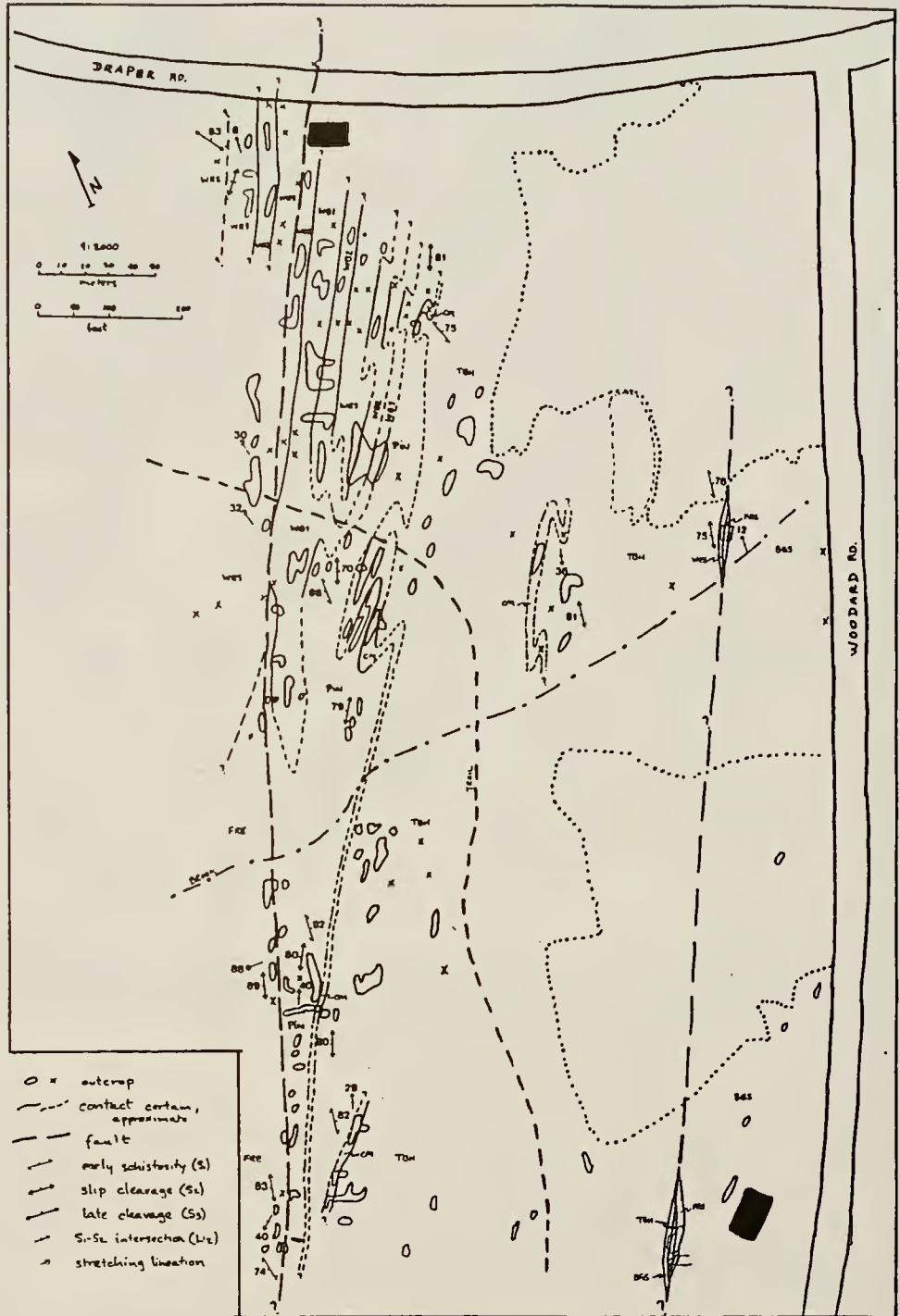
This locality exhibits a fault contact between the Frelighsburg (east) and the White Brook (west) Formations. Note that magnetite octahedrons are present only in the immediate proximity of the contact and developed on both sides of the fault. Note also the presence of shear bands in the Frelighsburg near the contact.

Go back to cars and drive back Hivernon road to Harvey street, and take left toward route 139. Go southward on route 139.

- 41.2 STOP 8: Turn left in the parking lot at the Rocher Bleu golf course. Park cars.

The lawn in back of the parking lot exhibits several small outcrops of White Brook dolomite. At the eastern edge, it is possible to follow the same contact between the White Brook and the Frelighsburg as observed at locality 7. Note again the presence of magnetite octahedrons in both lithologies.





- = outcrop
- contact certain, approximate
- - - fault
- early schistosity (S)
- ↘ slip cleavage (Sc)
- ↙ late cleavage (Ss)
- ⊥ S-Ss intersection (Wz)
- stretching lineation

KEY

|     |                                    |     |                                       |
|-----|------------------------------------|-----|---------------------------------------|
| BGS | black graphitic schist             | WB1 | dolomitic sandstone (White Brook Fm.) |
| FRE | FREELIENBURG Fm.                   | PIN | Pinnacle Fm.                          |
| WES | WEST SUTTON Fm.                    | CA  | Call Mill Mb.                         |
| WB2 | dolomitic marble (White Brook Fm.) | TBH | TIBBET Fm.                            |

Figure 6: Detail geologic map for part of the "Mansville Phase" (STOP 5). Approximate scale: 1:2500.

Going down the hill, along the southern side of the fairway, the first outcrop present is a highly sheared lithology that may belong to either the West Sutton or Frelighsburg. Note the nose of small  $F_2$  folds preserved between two  $S_2$ -slip planes.

From the golf course, take route 139 south to Sutton.

44.0 Return to the commuter parking lot in Sutton.

The itinerary for Trip C-6 starts on page 314. The introductory article for this trip, the itinerary for Trip B-8, and the articles by Laird and Coish are all assembled as a package.

# TECTONIC SETTING OF THE NORTHERN PART OF THE GREEN MOUNTAIN MASSIF, VERMONT<sup>1</sup>

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## INTRODUCTION

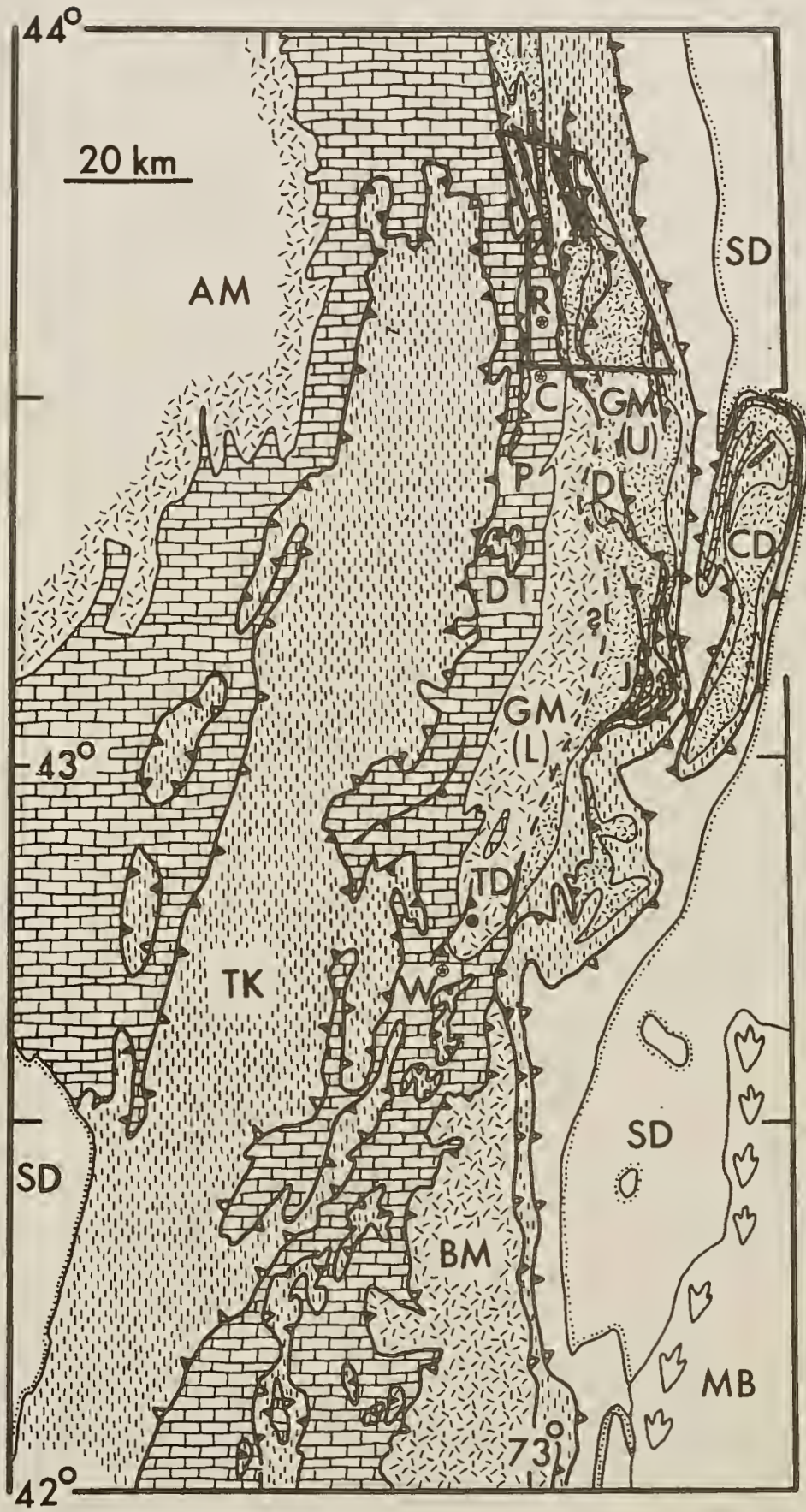
The Green Mountain massif in southern Vermont is cored by Middle Proterozoic basement rocks of the Mount Holly Complex (Doll et al., 1961; Fig. 1). The massif is bordered on the west by a sequence of Late Proterozoic to Middle Ordovician conglomerates, quartzites and marbles containing minor amounts of phyllite. This western sequence of cover rocks was deposited in shallow water on the continental shelf of ancient North America, and is blanketed by Middle Ordovician synorogenic flysch, which heralded the arrival of the Taconic thrust sheets (Cady, 1945; Rodgers, 1968). The basement rocks of the massif are bordered to the east by a very different cover sequence of Late Proterozoic to Lower Cambrian conglomerates, graywackes, and pelitic and mafic schists containing minor amounts of quartzite and marble. Farther east of the eastern cover sequence are mafic and pelitic schists and bimodal metavolcanic rocks of unknown but presumed Cambrian to Middle Ordovician age (Doll et al., 1961; Zen et al., 1983). These rocks may be remnants of an accretionary wedge and island arc complex (Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985).

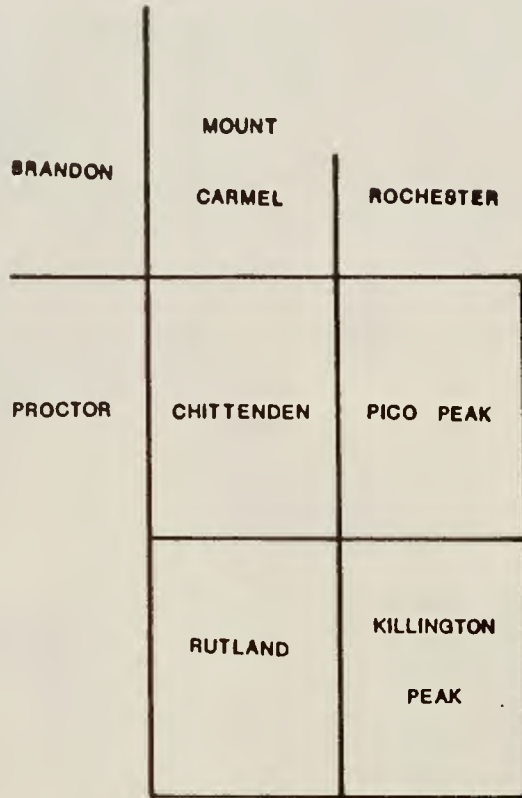
The basement rocks of the Mount Holly Complex were deformed during the Middle Proterozoic Grenville orogeny. These rocks, together with the cover sequences, were deformed and metamorphosed during the Ordovician Taconian and Devonian Acadian orogenies (Zen, 1967; Rosenfeld, 1968; Hepburn, 1975; Laird and Albee, 1981; Sutter et al., 1985). Within the map area of Figure 2, the Paleozoic metamorphisms reached no higher than biotite grade, except for some rocks in the northwest part of the area which reached garnet grade. The Grenville metamorphism occurred at higher temperatures, however, with the result that pegmatites are common in basement rocks of the Mount Holly Complex in the

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<sup>1</sup>Publication authorized by the Director, U.S. Geological Survey

Figure 1. Generalized tectonic map of western New England and eastern New York. Brick pattern- Late Proterozoic to Middle Ordovician western shelf sequence and Middle Ordovician synorogenic flysch; vertical dashed pattern- Late Proterozoic to Middle Ordovician slope-rise sequence; coarse hatchured pattern- Middle Proterozoic basement unconformably overlain by western cover sequence rocks; fine hatchured pattern- Middle Proterozoic basement unconformably overlain by eastern cover sequence rocks; unpatterned unit- presumed Cambrian to Ordovician remnants of an accretionary wedge and island arc complex; SD- Silurian and Devonian formations; MB- Mesozoic basin. Major tectonic features: AM- Adirondack massif; BM- Berkshire massif; CD- Chester dome; GM(U)- structurally higher tectonic unit in the Green Mountain massif; GM(L)- structurally lower tectonic unit in the Green Mountain massif; TK- Taconic klippen. Minor tectonic features and locations: C- Clarendon, VT; D- Devils Den exposure of cover rocks; DT- Dorset thrust sheet; J- Jamaica, VT; P- Pine Hill thrust; R- Rutland, VT; TD- The Dome; W- Williamstown, MA. Polygon shows area of Fig. 2. Heavy lines- thrust faults, teeth point to upper plate. Based on Doll et al. (1961), Zen et al. (1983), Thompson et al. (1982), Karabinos and Thompson (1984), Stanley and Ratcliffe (1985), and Thompson and McLelland (in press).





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

EASTERN COVER SEQUENCE

- ↵zph: Pinney Hollow Formation
- ↵zh: Hoosac Formation
- ↵zt: Tyson Formation
- ↵zp: Pinnacle Formation  
(uncertain affinity)

----- unconformity -----

- Mount Holly Complex  
(Relative ages uncertain)
- Ymf: Felsic gneiss
- Yma: Microcline augen gneiss
- Ymm: Marbles and calcsilicates
- Ymq: Quartzites and feldspathic quartzites

Figure 2. Geologic map of the north end of the Green Mountain massif based of field mapping 1981-1985. Structure in the western cover sequence south of Rutland, VT (R) based on Brace (1953) and Doll et al. (1961). N- Nickwacket Mountain. Heavy lines- thrust faults, teeth point to upper plate.





Green Mountain massif. This contrast in grade of metamorphism provides the most useful method for distinguishing between basement and cover rocks, although it can be difficult to identify basement rocks where the effects of Paleozoic deformation and retrograde metamorphism are especially intense.

Doll et al. (1961) interpreted the structure of the Green Mountain massif as an anticlinorium with a facies transition between the western and eastern cover sequences occurring over the eroded crest of the massif. Such a facies transition is not observed, however, at the north (Figs. 1 and 2) and south (Zen et al., 1983) ends of the massif where the two cover sequences are in fault contact. Stanley and Ratcliffe (1985, Plate 1) suggested that the basal units of the western cover sequence are related to the basal Tyson and Hoosac Formations on the east side of the massif by a facies transition, but proposed that units structurally above the Hoosac Formation were transported westward by thrust faults.

Based on detailed mapping near Jamaica (Karabinos, 1984) and Rutland, Vermont (Karabinos and Thompson, 1984; Karabinos, 1986), together with reconnaissance work elsewhere, I suggest that the Green Mountain massif is a more complicated structure composed of two different tectonic units (Fig. 1). The northeastern unit is composed of Middle Proterozoic basement and unconformably overlying eastern cover sequence rocks. It is structurally higher and more highly-transported than the southwestern unit. The southwestern unit is composed of Middle Proterozoic basement and unconformably overlying western shelf sequence rocks and is relatively less transported. As discussed later, a minimum relative displacement of 16 km between the two tectonic units appears to be necessary to explain the current structural geometry.

At the latitude of Rutland the eastern boundary of the massif is an unconformity, whereas the western boundary is a thrust which carried basement rocks of Middle Proterozoic age and rocks of the eastern cover sequence westward over the western shelf sequence (Figs. 1 and 2). The massif is made up of several thrust sheets composed of Middle Proterozoic basement rocks and eastern cover rocks and the stratigraphy of the cover rocks can be correlated between thrust sheets. The main purpose of this field trip is to examine the basement-cover relationships which provide evidence for these proposals.

Other important questions which may help fuel debate during the trip are: 1) How can we distinguish between the basal parts of the western and eastern cover sequences?, 2) what deformational features are attributable to the

Ordovician Taconian orogeny vs. the Devonian Acadian orogeny?, and 3) what is the relationship between thrusting in the Green Mountain massif and the Taconic Range to the west?

### Stratigraphy

The map area in Figure 2 is mostly contained in the Rutland, Vermont, 15' quadrangle mapped by Brace (1953) and the Rochester, Vermont, 15' quadrangle mapped by Osberg (1952, unpublished manuscript map). These works were used by Doll et al. (1961) in compiling the state map of Vermont. These authors along with MacFadyen (1956), Skehan (1961), Hewitt (1961), Chang et al. (1965), and Thompson (1967) describe the stratigraphy in and adjacent to the Green Mountain massif in detail. What follows is a brief summary of the lithologic units in the Rutland area. The principal differences between the present and past interpretations of the stratigraphy are that: 1) I correlate the basal cover rocks along the west boundary of the massif with the Tyson (Doll et al., 1961) and Hoosac Formations of the eastern cover sequence, whereas Brace (1953) mapped these rocks as the Mendon Formation. (Indeed, this belt includes Whittle's (1894) type locality of the Mendon Series, in his terminology); 2) I have mapped marbles west of Nickwacket Mountain (Fig. 2) as part of the western shelf sequence instead of the Forestdale Marble Member of the Mendon Formation (Brace, 1953). The possibility of this latter interpretation was first suggested to me by P.H. Osberg and J.B. Thompson, Jr..

#### 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.
- Ymq, Quartzite and feldspathic quartzite: Clean, massive, vitreous, blue quartzite. 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 Geisses: Heterogeneous plagioclase, quartz, +K-feldspar, biotite, chlorite, epidote, muscovite gneiss; may contain altered garnet.

#### Eastern Cover Sequence Rocks

##### Tyson (Doll et al., 1961) Formation

- €Ztc, Basal Conglomerate: Sand-sized detrital grains to boulders of blue quartz or quartzite, feldspar, and less

- commonly lithic fragments of gneiss in a matrix of quartz, albite, muscovite, biotite, chlorite schist.
- €Zts, Schist: Quartz, albite, muscovite, chlorite, biotite schist; commonly contains weathered pits and nodules of carbonate minerals. Light gray carbonate-rich quartz, muscovite schist.
- €Ztq, Quartzite: Light gray, fine-grained quartzite containing minor amounts of muscovite and feldspar.
- €Ztd, Dolomite: Fine-grained, buff-weathering, massive dolomite; many beds contain detrital quartz grains. Magnetite-rich near upper contact with Hoosac Formation.

#### Hoosac Formation

- €Zh, Albite Schist: Albite porphyroblast, quartz, biotite, muscovite, chlorite schist near base. Dark gray graphitic quartz, albite, muscovite, biotite, chlorite phyllite; commonly contains beds of dark quartzite, rarely contains thin beds of dolomite.

#### Pinney Hollow Formation

- €Zph, Green phyllite: 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

- €Zp: 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

#### Cheshire Quartzite

- €c: White, vitreous, massive quartzite. Interbeds of dark gray quartz, muscovite, biotite, chlorite phyllite and feldspathic, micaceous quartzite common near base.

#### Dunham Dolomite (Doll et al., 1961)

- €d: Buff-weathering dolomite containing siliceous partings and detrital quartz grains.

#### Monkton Quartzite and Winooski Dolomite

- €mw: 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

- €da: Vitreous quartzite interbedded with gray calcitic dolomite. Dolomitic quartzite and quartzose dolomite. Cross-bedding common.

#### Clarendon Springs Formation

- €cs: Gray calcitic dolomite.

#### Shelburne Marble

Os: White calcite marble. Intermediate gray dolomite unit.  
Ira Formation  
 Oi: Dark gray quartz, muscovite, biotite, chlorite  
 phyllite; contains beds of blue-gray calcite marble.

#### Basement-cover relationships

The contact between the basement rocks of the Middle Proterozoic Mount Holly Complex and the Late Proterozoic to Lower Cambrian Tyson Formation is a well documented unconformity along the east margin of the Green Mountain massif. Dale (1916) interpreted this contact as an unconformity and so has every other worker who has mapped it in detail, including Thompson (1950), Rosenfeld (1954), Brace (1953), Skehan (1961), Chang et al. (1965), Karabinos (1984), and Karabinos and Thompson (1984). We will see this contact at Stop 2.

The extent of stratigraphic continuity upward from the unconformity, within the cover sequence, is an important issue to resolve. Doll et al. (1961) showed the cover sequence east of the Green Mountain massif as a homoclinal sequence in which rocks become progressively younger to the east, with a major unconformity between Ordovician and Silurian formations. Zen et al. (1983), however, mapped the equivalent rocks along strike to the south in Massachusetts with numerous thrust faults dissecting the sequence. Ratcliffe and Hatch (1979) and Stanley and Ratcliffe (1985) proposed that the cover sequence east of the Green Mountain massif and around the Chester dome also contains faults at about the same position as the Hoosac Summit and Whitcomb summit thrusts of northern Massachusetts. Recent mapping (e.g. Karabinos, 1984; Thompson et al., 1982; Thompson and McLelland, in press) shows that thrust faulting was important, but more mapping is needed to correlate faults in southeastern Vermont with thrusts mapped elsewhere.

Based on detailed mapping near Jamaica and Rutland, Vermont (Karabinos, 1984, 1986; Karabinos and Thompson, 1984) and the work of others (Osberg, 1952; Brace, 1953; Skehan, 1961; Chang et al., 1965; and Thompson, 1972) I interpret the Tyson, Hoosac, and Pinney Hollow Formations as being stratigraphically continuous. In the Jamaica, Vermont area and at the north end of the Green Mountain massif the contact between the Hoosac and Pinney Hollow Formations is gradational and is not marked by evidence for strain gradients indicative of thrusting (Karabinos, 1984; Karabinos and Thompson, 1984). There are also many rock types common to the Tyson, Hoosac, and Pinney Hollow Formations suggesting that they were not deposited in dramatically different environments. Furthermore, the contacts do not show the persistent stratigraphic truncations used by Knapp and Stanley (1978), Stanley (1978, 1982), and Ratcliffe (1979) as evidence for thrusting in the

Rowe Schist in Massachusetts. Primary sedimentary structures are rarely preserved in the Tyson, Hoosac, and Pinney Hollow Formations, but their lithologies suggest that they were deposited in a deeper water environment than the quartzites and marbles of the shelf sequence. Zen (1967) and Thompson (1972) correlated these formations with rocks in the Taconic sequence to the west, some of which were deposited in a slope-rise environment (Friedman, 1979; Rowley et al., 1979). If this interpretation is correct, discontinuous quartzites and marbles in the Tyson and Hoosac Formations may have originated as sedimentary lenses (Skehan, 1961; Chang et al., 1965) and may have been derived from shelf 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. I have correlated cover rocks occurring in thrust sheets in the northern part of the massif with the Tyson and Hoosac Formations (Fig. 2) based on the similarity and sequence of lithologies.

On the west side of the massif south of Clarendon, Vermont (Fig. 1) the contact between the Mount Holly Complex and the Late Proterozoic to Lower Cambrian Dalton Formation is an unconformity (MacFadyen, 1956; Thompson, 1959; Skehan, 1961; Hewitt, 1961; Doll et al., 1961; Zen et al., 1983). The Dalton Formation is conglomeratic near the unconformity and phyllitic near its upper contact with the Cheshire Quartzite. It is not uncommon, however, for the conglomerate to grade directly into massive quartzite beds of the Cheshire Quartzite and for the intervening phyllite to be absent. The Dalton Formation is particularly thin and lacking in phyllite in the Wallingford, Vermont, 15' quadrangle (J.B. Thompson, Jr. and E. Downie, personal communications, 1985) just south of the map area shown in Figure 2.

Detailed mapping in the Killington Peak, Rutland, Pico Peak, Chittenden, Mount Carmel, and Brandon, Vermont, 7 1/2' quadrangles during the summers of 1981-1985 indicates that the western boundary of the Green Mountain massif north of Clarendon, Vermont is a major thrust fault (Figs. 1 and 2; Karabinos and Thompson, 1984; Karabinos, 1986). Rocks of the western shelf sequence, varying in age from the Early Cambrian Cheshire Quartzite to the Early Ordovician Shelburne Marble, structurally underlie basement rocks of the Mount Holly Complex or cover rocks which I interpret as basal units of the eastern cover sequence (Fig. 2). The cover rocks in the hanging-wall of the fault typically contain the sequence, upward from the basement contact: pebble to boulder conglomerate in a graywacke matrix; graywacke, commonly containing weathered carbonate grains and nodules and beds of quartzite and phyllite 1 m to 100 m thick; light gray quartzite; dolomitic marble; and albite-

bearing phyllite or schist locally interlayered with chloritoid-paragonite phyllite. These lithologies and their sequence are very similar to the common units in the Tyson and lower part of the Hoosac Formations on the east side of the massif at this latitude.

As shown in Figure 2, the Green Mountain massif at the latitude of Rutland is made up of several thrust sheets composed of Middle Proterozoic basement and Late Proterozoic to Lower Cambrian cover rocks belonging to the eastern cover sequence. Despite folding of the thrust sheets by later deformation, it is still possible to recognize stratigraphic truncations along fault boundaries.

The basement-cover relationships suggest that the Green Mountain massif is made up of two different tectonic units. Middle Proterozoic basement rocks of a structurally higher, more highly-transported unit are unconformably overlain by the eastern cover sequence. The upper part of this cover sequence (Hoosac, and Pinney Hollow Formations) was probably deposited in a slope-rise environment, as discussed above. The lower part of the sequence probably represents rift clastic deposits. Basement rocks of a structurally lower, relatively less transported unit in the southwestern portion of the massif are unconformably overlain by the western cover sequence dominated by quartzites and marbles deposited in a shelf sequence (Cady, 1945; Rodgers, 1968). The boundary between these two tectonic units is poorly constrained in the central and southern interior of the massif where detailed mapping is incomplete and exposure rather spotty, and it need not be an east-dipping thrust along its entire length. E. Downie and J.B. Thompson, Jr. (personal communications, 1986) suggested that the boundary could instead be a normal fault west of the Devils Den exposure of cover rock (Fig. 1). Another possibility is that some portions of the boundary are west-dipping back thrusts with eastward displacement of western cover sequence rocks and the underlying basement.

#### Conditions and age of thrust faulting

Where the location of thrust faults is well bracketed and samples have been collected, thin section analysis shows that in fault zones quartz deformed in a ductile fashion with extensive recrystallization, whereas feldspar deformed brittlely (Karabinos, 1986). Paleozoic metamorphism did not exceed biotite grade conditions in the area of Figure 2, except for a small region in the northwest part of the map area. Thrusting, therefore, probably occurred at biotite grade conditions.

Without independent information on the facies relationships between the western and eastern cover sequences, it is difficult to estimate the displacement on

thrusts in the northern part of the Green Mountain massif required to produce the observed juxtaposition of the two cover sequences. The basal lithologies of the two cover sequences are similar (Dalton Formation on the west side and Tyson Formation on the east side) but the overlying Cheshire Quartzite and Dunham Dolomite of the western sequence are, in general, quite distinct from the correlative Hoosac and Pinney Hollow Formations of the eastern cover sequence. Thompson (1972), however, correlated the Plymouth member of the Hoosac Formation with the upper part of the Cheshire Quartzite and the lower part of the Dunham Dolomite because of striking similarities in lithologies and textures. It appears, therefore, that similar depositional processes may have operated in both sequences during part of their development. It is also worth noting that if Stanley and Ratcliffe (1985) are correct in their proposal that the contact between the Hoosac and Pinney Hollow Formations east of the Green Mountain massif and around the Chester dome is a thrust, then the contrast between the cover sequences on the west and east side of the massif need not be too great. As described above, however, I interpret the Tyson, Hoosac, and Pinney Hollow Formations as being stratigraphically continuous. If this interpretation is correct, the significant lithological differences between the pelitic and mafic schists in the Pinney Hollow Formation and correlative units in the western shelf sequence (see Thompson, 1972) indicate that the eastern and western cover sequences were deposited in quite different environments when the Pinney Hollow Formation formed.

An important problem is the possibility of north to south facies variations in the cover rock sequences in addition to east to west variations. Dowling et al. (1987) described north to south facies variations in the Late Proterozoic Oak Hill Group of southern Quebec and correlative units in the Camels Hump Group of northern Vermont. Such north to south variations may be important north of the Green Mountain massif (cf. Tauvers, 1982), however, they do not appear to be dramatic at the latitude of the Green Mountain massif. Along the western boundary of the massif, south of Clarendon to the thrust fault in Pownal, Vermont exposed on the mountain called The Dome (Fig. 1), near the Vermont-Massachusetts border, the Dalton Formation and the Cheshire Quartzite maintain a fairly uniform thickness (Doll et al., 1961) and the most important variation is the presence or absence of the phyllite unit in the upper part of the Dalton Formation discussed above. South of thrust on The Dome, the Dalton Formation is significantly thicker than it is to the north (Zen et al., 1983), but the thrust has truncated the details of the transition in the Dalton Formation. Along the eastern boundary of the massif from its northern end south to Jamaica, Vermont, the Tyson, Hoosac, and Pinney Hollow Formations maintain an approximately uniform thickness (Doll

et al., 1961) and the most important variation is the presence or absence of the Plymouth Member of the Hoosac Formation. From Jamaica south to approximately the Vermont-Massachusetts border, two distinctive sequences of the basal cover rocks are separated by thrusts (shown by Doll et al. (1961) without intervening faults as the Cavendish Formation vs. the Tyson and Hoosac Formations). Skehan (1972) suggested that these two sequences could be coeval, represent east to west facies variations of each other, and be separated by thrusts. Karabinos (1984) presented structural evidence in support of these suggestions based on mapping near Jamaica, Vermont and recent mapping by N.M. Ratcliffe (shown in Zen et al. (1983) and unpublished) in southern Vermont also indicates that two distinctive cover sequences are separated by thrusts. These east to west variations in the basal cover rock sequences are similar to those described by Ratcliffe (1979) and Ratcliffe and Hatch (1979) in Massachusetts and southern-most Vermont.

The age of faulting in the northern part of the Green Mountain massif is poorly constrained; it could be part of the Taconian or Acadian orogenies, or there may have been more than one episode of thrust faulting. Uncertainty in the correct age assignment of faults in the eastern Taconic klippen, the Berkshire and Green Mountain massifs, and in complexly deformed cover rocks east of the massifs imposes a major limitation on our tectonic and palinspastic reconstructions of western New England. Stanley and Ratcliffe (1985) proposed that during the Taconian orogeny, thrusting became generally younger to the east away from the transport direction of the thrust sheets. Thus, late Taconian (and possibly Acadian) thrusts may truncate early Taconian thrusts. Later thrusting would not only complicate the structural geometry of the Taconian thrust belt but also the palinspastic reconstruction of the relative depositional sites of cover rocks from different thrust sheets.

Until much-needed radiometric dating of fault zone material becomes available, I think it is reasonable to correlate thrusting in the massif with Taconian faulting elsewhere in western New England (Taconic Range: Zen, 1967; Ratcliffe, 1979; Bosworth and Rowley, 1984; Berkshire massif: Ratcliffe and Harwood, 1975; Norton, 1975; Ratcliffe and Hatch, 1979; northern Vermont: Stanley and Roy, 1982; Stanley et al., 1982), although Acadian faulting is also recognized in western New England (Ratcliffe, 1979).

In support of this tentative correlation, the stratigraphic separation on the east side of the Taconic Range near Rutland, Vermont is very similar to that along the west boundary of the massif as shown in Figure 2. On the east side of the Taconic Range, basal Cambrian or Late Proterozoic units of the Taconic sequence, Biddie Knob, Bull, and West Castleton Formations of Zen (1964) and Netop



and St. Catherine Formations of Thompson (1967), structurally overlie Ordovician formations of the western shelf sequence or the Middle Ordovician synorogenic flysch blanketing the shelf sequence. On the northwest side of the Green Mountain massif, Middle Proterozoic basement and Late Proterozoic to Cambrian eastern cover sequence rocks of the Tyson and Hoosac Formations overlie Cambrian to Ordovician rocks of the western shelf sequence (Fig. 2). Virtually all the rock types found in the Hoosac and Pinney Hollow Formations, with the exception of mafic schists, are present in the basal units of the Taconic sequence near Rutland. In particular, Thompson (1967, p. 87) noted the strong resemblance of the pairs Netop-St. Catherine in the Dorset thrust sheet and the Hoosac-Pinney Hollow Formations east of the Green Mountain massif described by Chang et al. (1965). Therefore, it is possible that the sole faults on the east side of the Taconic Range, at least that of the Dorset thrust sheet, and the northwest margin of the Green Mountain massif were once continuous and that the fault cut up section to the west in the transport direction. If the sole faults of the Dorset thrust sheet and the northwest margin of the Green Mountain massif were once continuous, a minimum thrust displacement of approximately 16 km is necessary to account for the current structural geometry (Fig 1). Naturally, a much larger thrust displacement is possible.

Although the basement-cover relationships indicate that the massif is made up of two tectonic units, it is unclear at present how much relative displacement between these units occurred. If the western Taconic klippen (group 1 and 2 of Stanley and Ratcliffe, 1985) were deposited east of the Chester dome and were thrust over rocks of the Green Mountain massif as suggested by Stanley and Ratcliffe (1985) the relative displacement between the southwestern and northeastern tectonic units of the Green Mountain massif may be as little as 16 km as discussed above. If, however, the western Taconic klippen represent a transitional cover sequence between the western shelf sequence and cover rocks presently east of the Green Mountain massif as suggested by Zen (1967) the relative displacement between the two tectonic units in the massif must be greater than approximately 60 km to account for the present structural geometry (Fig.1).

Thrust faults are also present in the western cover sequence in the Vermont valley such as the Pine Hill thrust (Fig. 1; Wolff, 1891; Dale, 1894; Brace, 1953; Zen, 1964; Thompson, 1967). Thrusts involving Middle Proterozoic basement and the western cover sequence may form a duplex structure (Boyer and Elliott, 1982) which appears to extend beneath the Green Mountain massif and the Taconic Range. Ando et al. (1984) presented seismic evidence compatible with this interpretation (see also Stanley and Ratcliffe, 1985). A duplex beneath the Green Mountain massif would

also help account for its anticlinorial structure.

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#### ITINERARY

Assembly point is parking lot on east side of Route 100 across from ski area in West Bridgewater, Vermont, 0.3 miles south of intersection of Routes 100 and 4.

#### Mileage

0.0 Drive south on Route 100.

This stretch of the valley followed by Route 100 is underlain by dolomite. The steep slope to the west is held up by resistant rocks of the Middle Proterozoic Mount Holly Complex and the Late Proterozoic to Lower Cambrian Tyson Formation. The steep slopes to the east are formed by the aluminous schists of the Late Proterozoic to Lower Cambrian Hoosac and Pinney Hollow Formations.

1.1 Quarry in dolomite of the Tyson Formation east of road. Outcrop of quartzite of the Tyson Formation next to road on east and in stream to west.

1.6 North end of Woodward Reservoir

2.8 STOP 1. Parking loop on west side of Route 100 just south of Woodward Reservoir.

On west side of outcrop, where the steep slope meets the parking area, is east-dipping, light gray, clean quartzite of the Tyson Formation. Heading east, the rock types are variable: quartz-rich muscovite, biotite, chlorite schist; thin layers of albite porphyroblast schist; quartz-pebble conglomerate in a quartz matrix, a schistose dolomite matrix, and dolomitic quartz matrix.

On the east side of Route 100 in an abandoned quarry is a good exposure of the dolomite member of the Tyson Formation. Many layers contain detrital quartz grains. The dolomite is typically massive and buff weathered, and sedimentary structures are

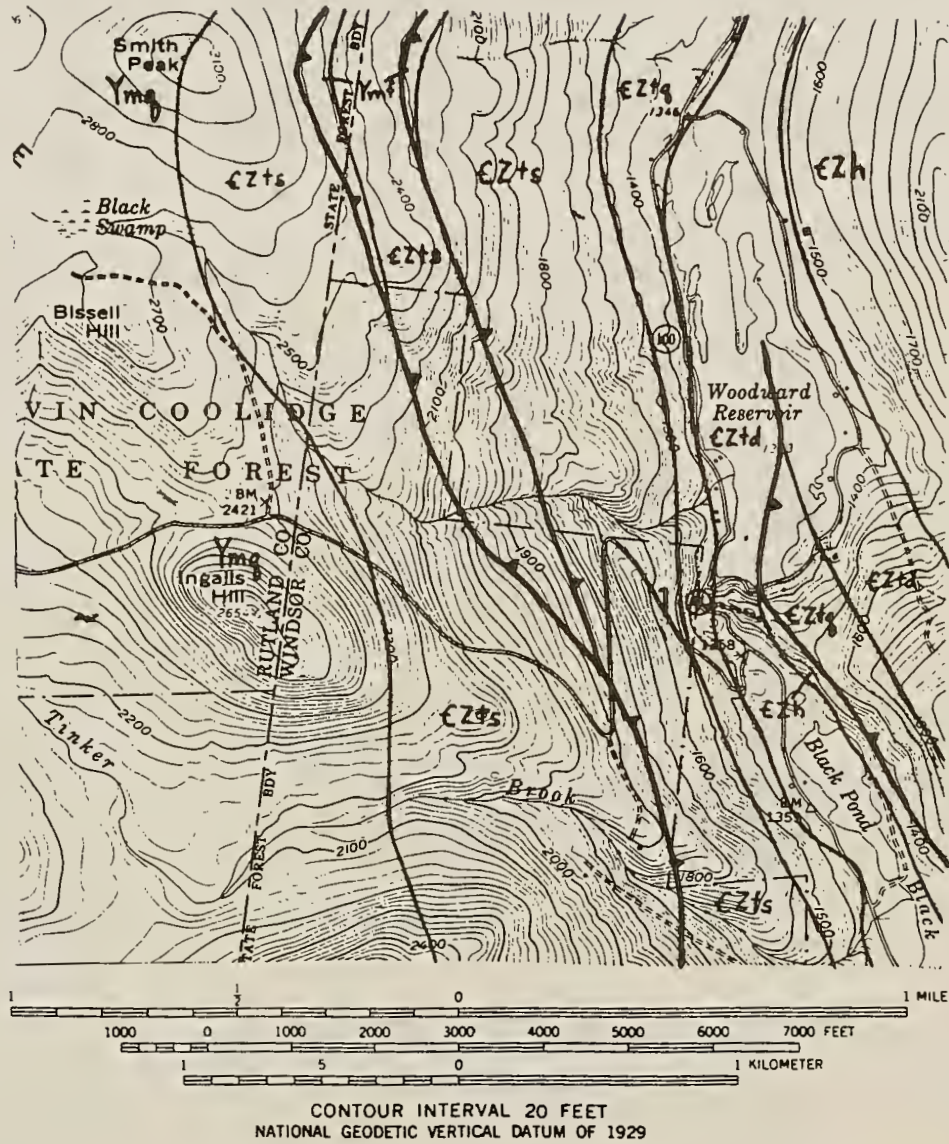


Figure 3. Photocopy of a portion of the Killington Peak 7 1/2' quadrangle showing the location of Stop 1. See stratigraphy section in text for description of geologic units.



rarely preserved. The depositional environment of this unit is difficult to reconstruct. It is widely but not continuously exposed: in the Chester and Athens domes, on the east side of the Green Mountain massif from the south near the Vermont-Massachusetts border to the north end of the massif near Pittsfield, Vermont and in some of the cover rock exposures within the massif as seen later at Stop 5.

(A good understanding of the depositional environment of this dolomite unit would be most helpful in tectonic reconstructions. Stanley and Ratcliffe (1985) used this unit as evidence that the continental shelf extended east of the Chester and Athens domes and then argued that the Taconic thrust sheets must have come from still farther east because their sediments were deposited in deeper water, presumably a slope-rise environment (Friedman, 1979; Rowley et al., 1979). The available evidence also permits the interpretation that the dolomite unit of the Tyson Formation formed as slump deposits derived from the shelf. It commonly contains beds with detrital quartz and dolomite grains, and at its lower boundary the unit grades into dolomite-bearing schist, graywacke, or quartzite of the Tyson Formation. It is worth noting that dolomite beds are locally present near the base of the Taconic sequence (e.g., Zen, 1967), therefore, the distribution of the dolomite unit of the Tyson Formation may not prove that the Taconic thrust sheets were derived from east of the Chester dome.)

Walk east just south of quarry to examine dolomite.

Continue into woods to outcrop of albite porphyroblast schist of the Hoosac Formation. Contact not well exposed here, but the lower part of the Hoosac contain minor magnetite. This is very common where the contact between the dolomite unit and the Hoosac Formation can be closely bracketed (see also Thompson, 1972; and stop 5, this trip).

Return to cars and drive north on Route 100.

5.9 Intersection of Routes 4 and 100, West Bridgewater.

Continue north on Route 100 and west on Route 4.

9.2 STOP 2. Pull over on east side of road and cross highway to outcrop on west side.

The description of this outcrop is from Thompson (1972). It is duplicated here because the contact between basement rocks of the Mount Holly Complex and cover rocks of the Tyson Formation is exposed.

The basement lithology is quartz, K-feldspar, plagioclase, muscovite gneiss. Upslope are good outcrops of quartzite and quartz-rich gneiss with abundant pegmatites.

The Tyson Formation is here composed of quartz, albite, muscovite, chlorite schist.

A folded gneissic fabric in the basement rocks is truncated at the unconformity.

Continue north on Route 100.

10.1 Turn right onto River Rd. Again, the valley which the road follows is underlain by the dolomite member of the Tyson Formation.

13.3 STOP 3. Pull off road on right and walk west to sharp bend in road.

At bend in road is augen gneiss of the Mount Holly Complex. Walk back, northeast, and examine lithologies along road. The augen gneiss becomes very sheared near the contact with blue quartzite of the Mount Holly Complex. The blue quartzite is coarsely crystalline and contains pegmatites. Farther northeast is a conglomerate of the Tyson Formation interbedded with albite schist and carbonate-rich schist.

The same sequence of lithologies can be seen walking northwest from bend in road or going straight up cliff. The augen gneiss here is in the core of a small antiform.

Return to cars and continue west.

13.9 Low outcrops of augen gneiss in pasture north of road.

View of Pico and Killington Peaks to south

14.1 Turn south (left) onto Route 100.

15.9 Turn left into parking area for Kent Pond.

16.0 Stop 4. Park near east end of lot. Head south along shore to where Kent Brook flows into pond (approx. 50 m). Walk upstream to exposures of

interbedded quartzite, marble, and calc-silicate rock of the Mount Holly Complex.

This package of interbedded quartzite, marble, and calc-silicate rock is very common on the broad terrace which runs north-south and is located east of Pico Peak and west of the Route 100 valley.

Return to cars and Route 100.

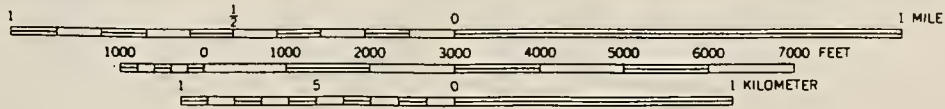
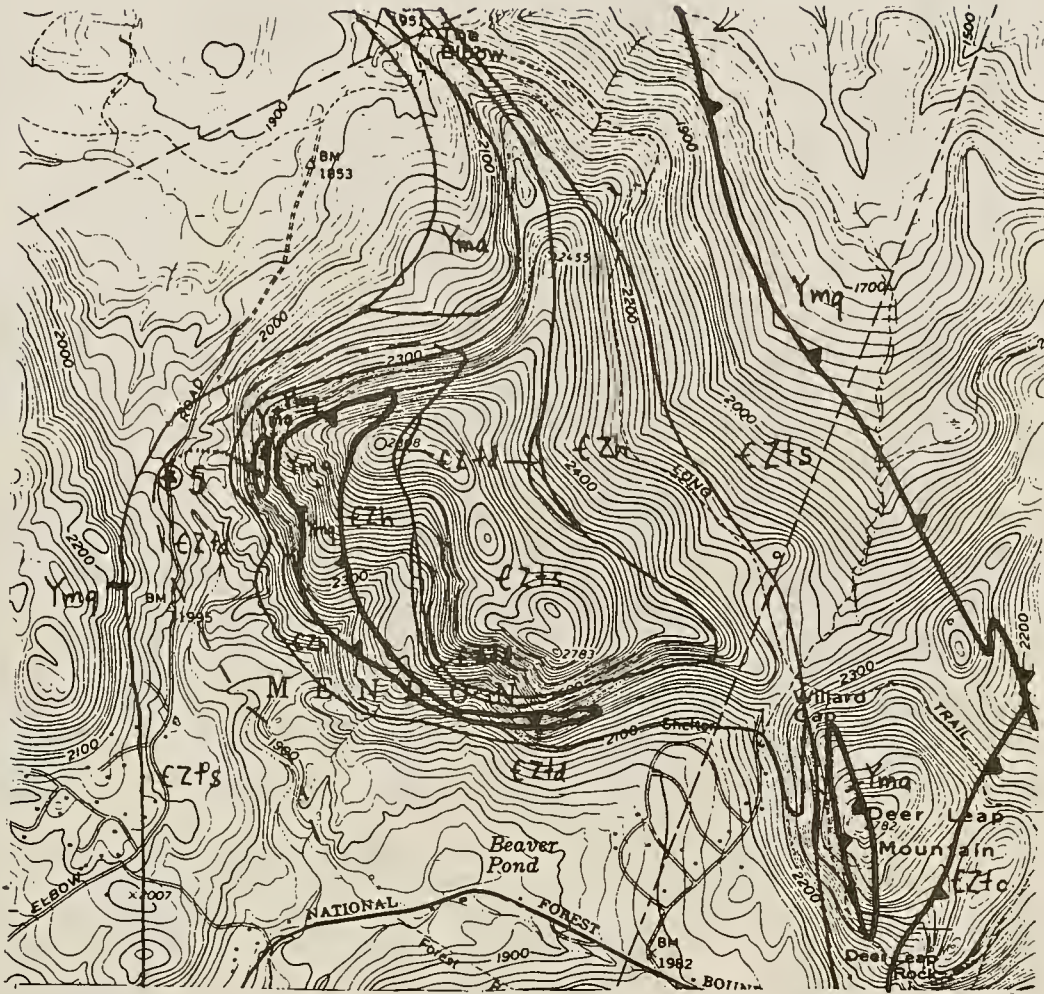
- 16.1 Turn south (left) onto Route 100.
- 16.3 Turn west (right) onto Route 4.
- 17.8 Sherburne Pass and Long Trail crossing. Cliffs on north side of road, called Deer Leap, are composed of cover rocks of the Tyson Formation. Some rather spectacular conglomerates are present near the top of the cliffs.
- 19.0 Beaver Pond on right side of road.
- 20.9 Turn right onto Old Turnpike Road, opposite Killington-Pico Motor Inn.
- 22.1 Pavement ends.
- 22.6 Road narrows, continue straight ahead.
- 23.2 STOP 5. Pull off to side of road.

Fine-grained dolomite of the Tyson Formation in road bed. Walk north 0.1 mile and turn east (right) into woods along obscure dirt track for approx. 45 m to abandoned quarry. Fine grained, buff-weathered dolomite.

Head due east 150 m to base of steep slope. Here is magnetite-rich albite schist of the Hoosac Formation. Upslope are layers of quartzite and quartz-rich schist. Further upslope are exposures of augen gneiss.

I interpret the augen gneiss here to be in thrust contact with the Hoosac Formation and to occupy the core of a synform (Fig. 4). The outcrop pattern on this unnamed hill is complicated by at least two sets of post-thrusting folds. The first set forms an east-closing, nearly recumbent synform. The second set of folds is more upright and open and is variably oriented.

Return to cars. Drive south to return to Route 4.



CONTOUR INTERVAL 20 FEET  
NATIONAL GEODETIC VERTICAL DATUM OF 1929

Figure 4. Photocopy of a portion of the Pico Peak 7 1/2' quadrangle showing the location of Stop 5. See Stratigraphy section in text for description of geologic units.

- 26.0 Turn west (right) onto Route 4.
- 27.1 Long outcrop on left of diverse lithologies in the Mount Holly Complex. Quartzites, rusty schists, calc-silicates rocks, and feldspathic gneisses.
- 28.3 STOP 6. Turn right into parking lot of Sugar and Spice Pancake House just past Meadowlake Drive. Walk 0.2 miles west along Route 4 to outcrop on left of road.

Blue, vitreous quartzite of the Mount Holly Complex containing pegmatites. Across the road is a pasture with no exposure (a rather unusual setting for the basal clastics of the Dalton Formation and Cheshire Quartzite). The pasture is probably underlain by carbonates of the western shelf sequence.

South of here on the west slope of East Mountain are good exposures of conglomerate, schist, and dolomite which I interpret as belonging to the Tyson and Hoosac Formations.

Return to cars and drive west on Route 4.

- 29.5 Turn south (left) onto Town Line Rd.
- 31.3 Turn west (right) onto Killington Rd. at stop sign.
- 32.3 Turn south (left) onto Stratton Rd. at stop sign.
- 32.9 Turn left onto Allen Rd. at traffic signal.
- 34.6 STOP 7. Pull off onto side of road or off of driveway on east side of road. Walk past Caboose and into tree farm. Head to the northeast corner of pasture and cross fence. Head due east and look for outcrop of light gray, vitreous quartzite with cross-beds showing tops to east. Cross-beds also present in dolomite beds nearby.

Quartzite beds are part of the Danby Formation. Farther east uphill are exposures of gray dolomite of the Clarendon Springs Formation and calcite marble of the Shelburne Marble.

Walk east through woods approx. 500 m to break in slope on west side of Bald Mountain. Outcrop of conglomeratic unit of the Tyson Formation.

Exposures of basement lithologies are present at the break in slope along strike to the north and south.

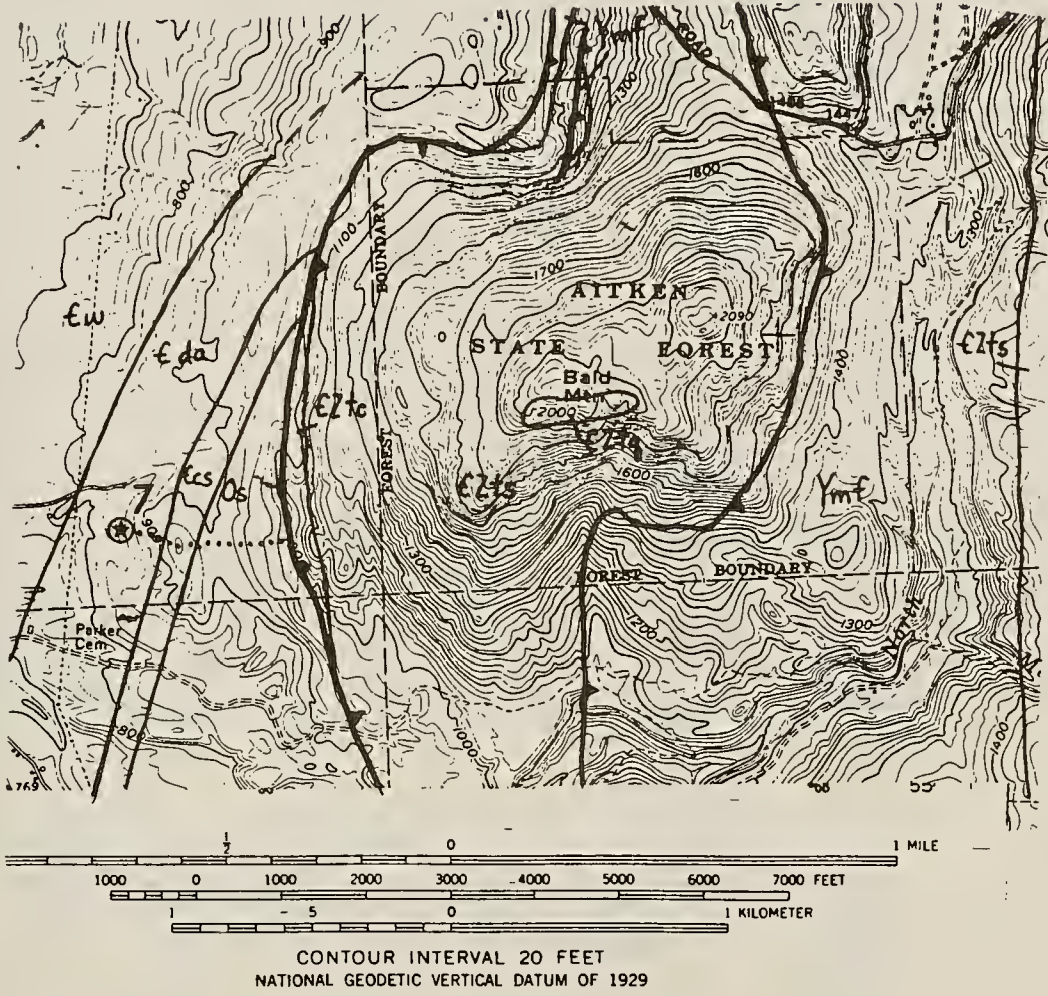


Figure 5. Photocopy of a portion of the Rutland 7 1/2' quadrangle showing the location of Stop 7. See Stratigraphy section in text for description of geologic units.

Brace (1953) mapped a thrust between the quartzite and marble outcrops in the pasture and the conglomerate exposures at the break in slope of Bald Mountain. He interpreted the thrust as a break in the western cover sequence, whereas I propose that it juxtaposes eastern and western cover sequence rocks. The evidence for this interpretation includes: 1) The similarity of lithologies and sequence of rock types between the Tyson and Hoosac Formations on the east side of the massif (and in thrust sheets within the massif) and the cover rocks exposed along the west boundary of the massif;. 2) nowhere along the west boundary of the massif in the area of Figure 2 is it possible to demonstrate stratigraphic continuity between the rocks I have mapped as Tyson and Hoosac Formations and the Cheshire Quartzite; 3) just south of the area shown in Fig. 2 in the Wallingford, Vermont, 15' quadrangle, where it is possible to demonstrate stratigraphic continuity from the basement-cover unconformity upward to the Cheshire Quartzite, the basal conglomerate grades very rapidly into the clean quartzites of the Cheshire Quartzite; only minor beds of phyllite are present, trivial in comparison with the thick sequence of graywacke, phyllite, and schist present along the west margin of the massif north of Clarendon, Vermont; and 4) in the northwestern part of Figure 2 a significant area underlain by dolomite and calcite marble shown by Doll et al. (1961) as the Forestdale Marble (a unit in the Mendon Formation of Whittle, 1894) contains lithologies identical to the Dunham Dolomite, Monkton Quartzite, and Winooski Dolomite, which I have mapped as part of the western cover sequence. This stratigraphic reinterpretation, if correct, indicates that the Tyson and Hoosac Formations are in thrust contact with the western shelf sequence in this area.

Return to cars. Return to Route 4 by reversing directions back to stop 6. To go to Route 7 return to road, drive south 0.4 mi. and bear right where road forks. Turn right at intersection and drive west 0.8 mi. to Route 7.

END OF TRIP

WINOOSKI RIVER TRANSECT:  
REFOLDED FOLDS AND THRUST FAULTS IN THE  
CORE OF THE GREEN MOUNTAIN ANTICLINORIUM

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INTRODUCTION

A transect across the Camels Hump Group parallel to the Winooski River is being mapped from Jonesville to Waterbury, Vermont, in the Camels Hump 15' quadrangle as part of the Vermont Bedrock Geology Mapping Project. This report is based on field work in the summer of 1986 and in June, 1987, and is presented as a progress report rather than final conclusions. The report incorporates detailed mapping by Eiben (1976) and Aubrey (1977), on Stimson Mountain and the northern ridges of Camels Hump, respectively. The Winooski River cuts a deep WNW valley transverse to the dominantly N-S trending structures, thus providing a deeper view of structures which, on the Green Mountain ridges, appear as parallel belts of rock due to low-plunging late folds.

GEOLOGIC SETTING

The late Proterozoic Camels Hump Group occupies the center of the Green Mountain anticlinorium, covering a twenty-mile-wide area at the latitude of the Winooski River (Doll *et al.*, 1961). The hinge of the anticlinorium lies approximately halfway between Jonesville and Waterbury, passing close to the summits of Camels Hump and Bolton Mountain (Christman and Secor, 1961). The hinge shifts en echelon eastwards from south to north, and along the crest there is a mile-wide zone of fairly flat-lying rocks. The apparent lack of symmetry in rock units across the anticlinorium was explained by Christman and Secor (1961) as facies interfingering between Underhill and Hazens Notch Formations (fig. 1). Eiben (1976) concluded that earlier phases of folding unrelated to the anticlinorium were important in controlling the map pattern. He mapped out greenstones and several types of schist around Stimson Mountain as different members of the Underhill Formation (fig. 2). Aubrey (1977) followed a similar interpretation, but suggested that some of the rocks might belong to other formations of the Camels Hump Group. Stanley and Ratcliffe (1985) concluded that early faults, as well as facies changes and folds, must be considered in the interpretation of map patterns within the group. The Camels Hump Group is continuous to the south with rocks



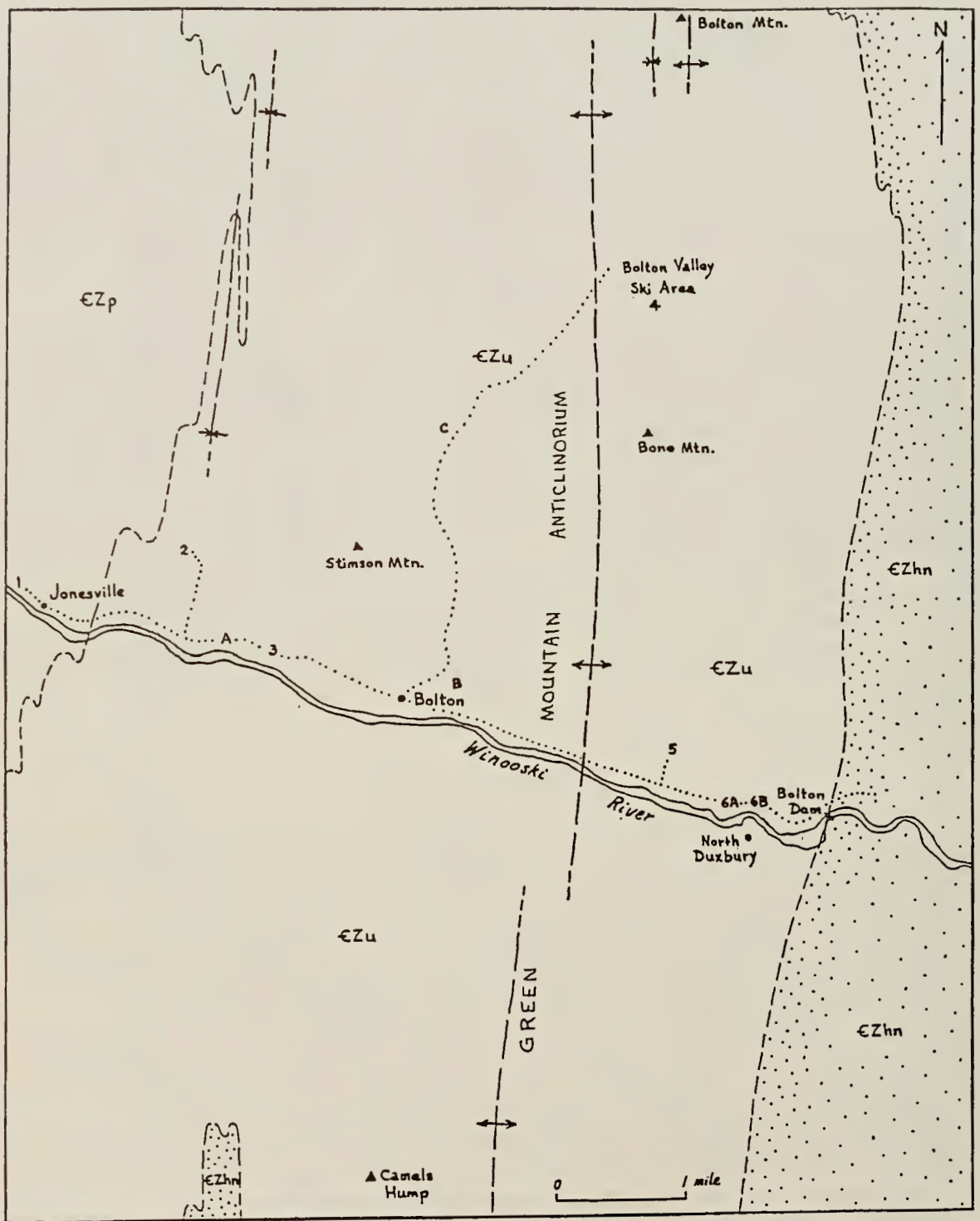


Fig.1 Geology of the Bolton area, after Christman and Secor (1961). EZp Pinnacle, EZu Underhill, EZhn Hazens Notch. Field trip route (dotted line) and stops are shown.

along the east side of the Green Mountain anticlinorium, which according to Stanley and Ratcliffe (1985) contain the root zone for the Taconic allochthons.

## STRATIGRAPHY

The Camels Hump Group (Doll et al., 1961) includes rocks deposited in three different, though adjacent, tectonic settings: clastic and volcanic rift facies deposited on North American continental basement (Pinnacle Formation and Tibbet Hill Volcanics), transitional continental slope and rise graywackes and shales (Underhill and Pinney Hollow Formations), and ocean floor eugeoclinal carbonaceous pelitic rocks and volcanics (Hazens Notch Formation). These rocks were juxtaposed and imbricated during the development of an accretionary wedge along the eastern margin of the continent (Stanley and Ratcliffe, 1985). Only the Hazens Notch contains local slices of metamorphosed ultramafics; its western contact thus corresponds to Cameron's Line.

### Pinnacle Formation

The Pinnacle Formation is characterized by pin-striped metagraywackes and greenish-gray phyllite and schist. Blue quartz grains and local quartz conglomerate beds have been reported (Christman and Secor, 1961; Thresher, 1972). Thresher's work showed that typical Pinnacle graywackes (his Huckleberry Hill Member) contain biotite and stilpnomelane, whereas others contain chloritoid but lack biotite (the latter he assigned to the Underhill Formation). The Pinnacle Formation crops out mainly west of the area we have mapped, but there are metagraywackes or granulites in the area which might belong to the Pinnacle.

### Underhill Formation, sensu stricto

We have mapped rocks as Underhill Formation using much stricter criteria than Christman and Secor (1961), Doll et al. (1961), Eiben (1976) and Aubrey (1977). The Underhill consists mainly of silver-green magnetite-bearing chlorite-muscovite-quartz(-albite) schist and gneiss. Biotite and garnet are rarely present. There are local lenses of quartz-feldspar granulite. Magnetite is present in nearly all outcrops of the Underhill, either as conspicuous porphyroblasts or as finely disseminated grains. Pyrite may be present, but is rarely abundant enough to impart rusty weathering to the rocks, in contrast to the Hazens Notch Formation. Graphite may be locally present, giving the rock a dark blue-black color. Magnetite and non-rusty weathering are thus the chief field criteria used in mapping the Underhill. Rocks of the Underhill are relatively resistant and form many of the cliffs north of the river in Bolton, as well as

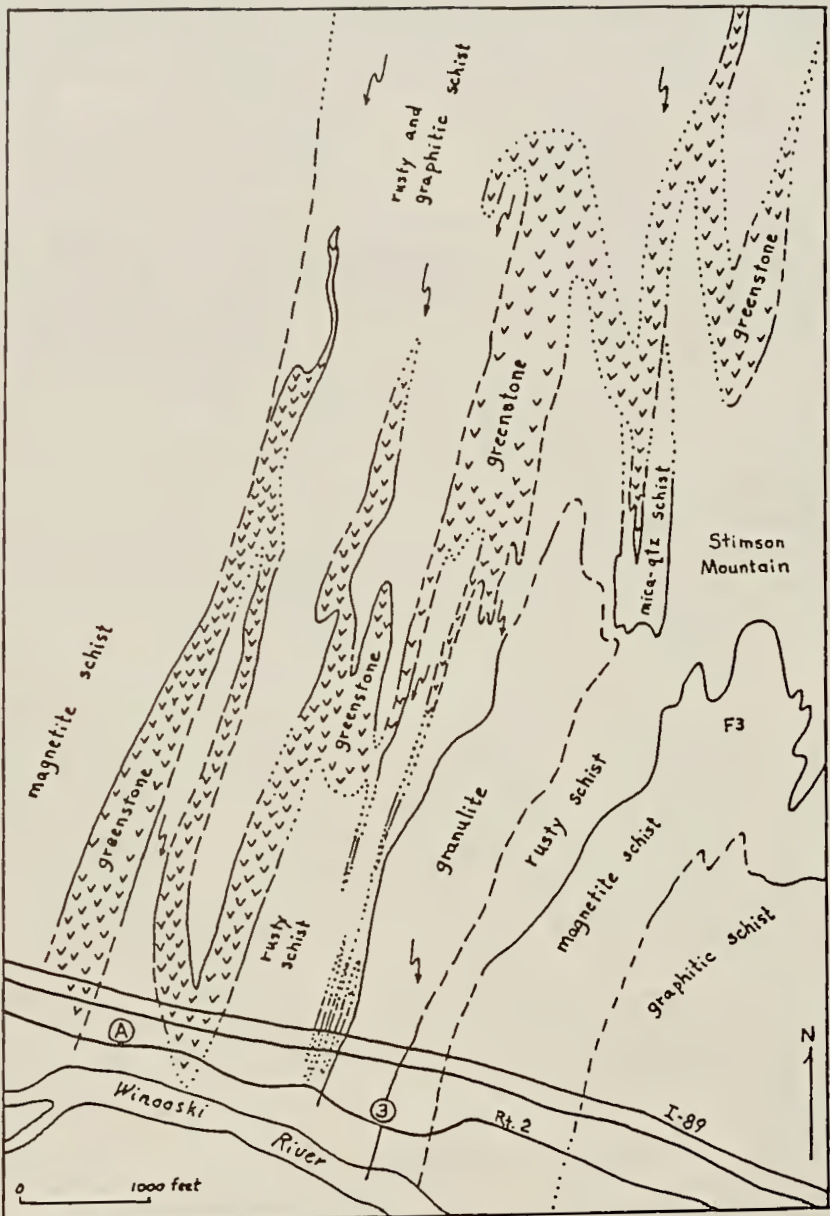


Fig.2 Geology of area west of Stimson Mountain, after Eiben (1976). Arrows indicate F2 fold rotation sense.



Fig.3 Sketch from outcrop of typical sheared F2 fold hinges.

the summits of Camels Hump, Bone Mountain and Mount Mansfield. We have not yet encountered greenstones within the Underhill in Bolton, but they are present elsewhere in the Underhill in the quadrangle (Christman and Secor, 1961).

#### Hazens Notch Formation

The Hazens Notch Formation is the most heterogeneous unit mapped. Christman and Secor (1961) mapped a contact between Underhill and Hazens Notch Formations east of, and parallel to, the anticlinorial hinge (fig.1). Rocks of the Hazens Notch Formation consist of rusty-weathering, sulfidic, chlorite-muscovite-biotite-quartz(-albite-garnet) schists interlayered with strongly graphitic, black schists with similar mineralogy. There are local horizons of tan-weathering carbonate. Several layers of greenstone and amphibolite occur within the rusty and graphitic rocks.

Rock types similar to those of the Hazens Notch are exposed in many places west of the anticlinorium, within what Christman and Secor (1961) mapped as Underhill Formation. They recognized this problem, and found that "each unit contained local layers similar to all the other units" (Christman, pers. comm., 1987). Eiben (1976) mapped a detailed stratigraphy on Stimson Mountain which we interpret as an imbrication of rocks from the Underhill, Hazens Notch, and possibly Pinnacle Formations. Those rocks belonging to the Hazens Notch are: rusty albite schist with local lenses of granulite and marble; rusty graphitic schist with 1-5 cm. black quartzites and local tan carbonates; deep red, rusty micaceous schist; light gray garnet schist with quartz laminae; and greenstones. Eiben (1976) reported a variety of metavolcanics and concluded that amphibolites were more commonly preserved in fold hinges whereas greenschists resulted from higher shear strain and retrograde metamorphism on fold limbs.

The metavolcanics in the Hazens Notch east of the anticlinorial hinge are also variable, including chlorite-albite-carbonate-epidote greenstone, chlorite-albite-quartz schist, and amphibole-albite-epidote-chlorite amphibolite. One occurrence of talc-carbonate rock was found within the Hazens Notch Formation in a roadcut along Route 2 near Bolton Dam.

We have attempted to map the Hazens Notch-like rocks separately from the Underhill, and the resulting pattern is shown in figure 4. Greenstone marks the main contact along the east edge of the Underhill east of Bone Mountain, but in many other places there is no greenstone between them. It is doubtful that the Underhill-Hazens Notch contact is anywhere a depositional contact. In some places the contact is clearly a fault, as discussed below. Elsewhere, rela-

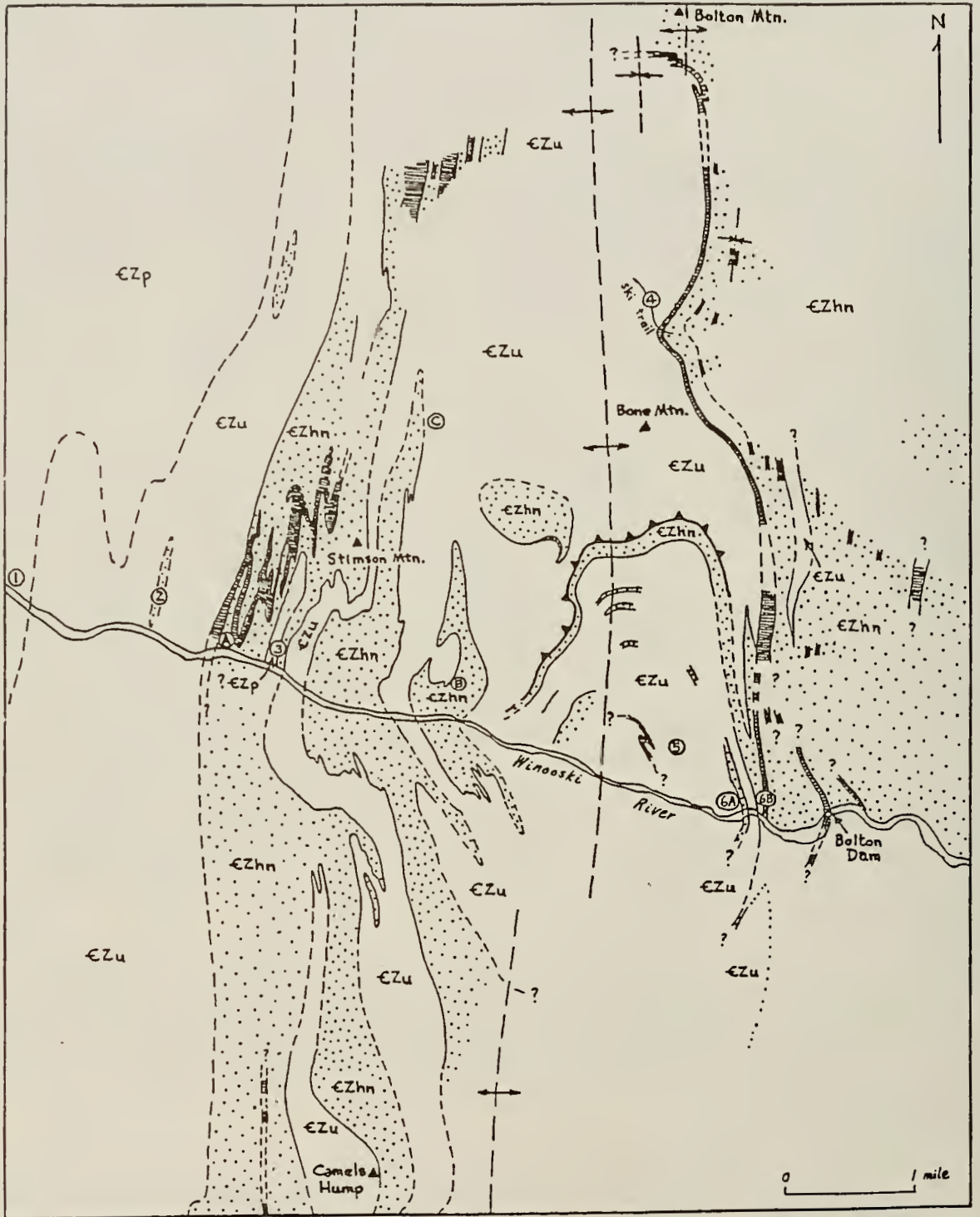


Fig.4 Geology of the Bolton area according to this study. Symbols as in figure 1. Greenstone is horizontally ruled.

tionships are uncertain, especially where both units become richer in albite near the contact (Eiben, 1976 and Aubrey, 1977), a phenomenon also noted in the Bakersfield-Waterville area to the north (Thompson, 1975). The albite porphyroblasts cut across the foliation and are interpreted as Acadian in age. Late metamorphism may thus obscure the original nature of the contact, making it appear transitional.

## STRUCTURE

### Phase Three Structures

The Acadian Green Mountain anticlinorium dominates the structure of the Bolton area. Along its hinge bedding and foliation are nearly horizontal. Most outcrops contain minor folds related to the anticlinorium. The folds plunge gently south or north, with axial planes dipping steeply to vertical. West of the hinge the folds have an asymmetrical shape and west-over-east rotation sense, and east of the hinge, the folds are east-over-west. Folds with both senses of rotation, or folds with symmetrical (neutral) shape, are found in the hinge area. In the more schistose rocks a crenulation cleavage or spaced cleavage parallels the axial planes. The Acadian structures have been called "phase three" because there is evidence for two older phases of deformation (Eiben, 1976).

### Phase Two Structures

The phase three folds (F3) discussed above deform both bedding and a pervasive foliation (S2) which is parallel to the axial planes of phase two folds (F2). F2 folds are generally isoclinal and very commonly show evidence of shearing along limbs (fig.3). F2 fold axes have a wide spread in orientation, but many are approximately parallel to the Acadian F3 fold axes. F2 folds are believed to be Taconian in age.

F2 folds are unfortunately much less common than F3 folds. Close attention to the rotation senses in map pattern and in outcrops near contacts aids in understanding the history of deformation. South of Bone Mountain layers of Underhill and Hazens Notch Formations are repeated, the Underhill forming cliffs and steep slopes and the Hazens Notch forming gentler slopes (fig.4). If the alternating layers were due to F2 isoclinal folds, the rotation sense of related minor folds should change back and forth. However, only west-over-east F2 minor folds have been found. Faults have apparently "sheared out" the east-over-west F2 folds. Eiben (1976) based his fold interpretation of the

greenstones west of Stimson Mountain (fig.2) in part on F2 rotation reversals. The same rock type is not consistently west of the greenstone, however, and the greenstone cannot be followed continuously north of Eiben's area (fig.4). An interpretation involving faults as well as folds is probably more correct.

### Faults

Evidence for faults is of three kinds: (1) contacts truncated by other contacts, (2) discontinuous slices, and (3) field evidence. Along the eastern contact of the main mass of Underhill Formation north of Bone Mountain there is consistently a greenstone, dipping east under the Hazens Notch Formation. To the south this contact appears to truncate layers of Hazens Notch within the Underhill. East of the contact is an exceedingly complex zone about a half-mile wide in which slices of greenstone, Underhill magnetite schist, and white quartzite occur within rocks typical of the Hazens Notch Formation. As of this writing detailed mapping of this zone is underway. We suspect it may represent a pre-metamorphic fault zone. One confusing aspect of the zone is that although the layers are discontinuous, the sequence of rock types across strike maintains some semblance of order for hundreds of feet along strike.

Direct field evidence of faults is best observed along the Underhill-Hazens Notch contacts south of Bone Mountain. F2 folds and pervasive foliation are locally truncated. Extremely sheared sulfidic schists with papery texture and brecciated quartz veins occur mainly within the Hazens Notch. F3 folds deform these features, indicating a late Taconian, pre-Acadian age for the faults.

### Phase One Structures

Strongly developed E-W mineral lineations and quartz rods and scarce isoclinal E-W fold hinges (F1) suggest a phase of deformation older than F2. F1 folds may have been rotated to their E-W orientations during F2 deformation, and this orientation may indicate the sense of movement during F2.

### METAMORPHISM

Platy minerals commonly are parallel to the pervasive foliation (Taconian metamorphism?), but locally biotite and muscovite grains (Acadian?) lie across the foliation. Christman and Secor (1961) showed the Bolton area mostly within a four- to five-mile wide garnet zone which coincides with the Green Mountain anticlinorium. However, the garnet isograd along the west side corresponds roughly with the

western limit of patches of Hazens Notch Formation, and thus may reflect the presence or absence of rocks of appropriate composition rather than P-T conditions. (The Underhill Formation rarely contains garnet or biotite.)

The present authors have done no thin-section work to date. Samples of amphibolite have been collected for study by Jo Laird to determine whether the amphiboles show evidence for a high-pressure Taconian metamorphism preceding the Acadian greenschist metamorphism.

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#### ITINERARY

Assembly point is Quinn's Store and Post Office, Jonesville, Vermont, on Route 2 between Richmond and Waterbury, at 8:30 A.M. Topographic Maps: Richmond, Bolton Mountain, and Waterbury 7.5' quadrangles. People approach-



ing Jonesville from Waterbury have excellent views of Bone Mountain cliffs as they leave Waterbury. Optional stops are intended to replace Stop 4 in case of inclement weather.

### Mileage

- 0.0      STOP 1: North side of Route 2 opposite Post Office. This is mainly a stop to inspect the Pinnacle Formation: chlorite schist and pin-striped metagraywacke granulite with blue quartz. Blue quartz is more abundant in massive granulite (Stop 8, Thresher, 1972) 0.1 mile east. At Stop 1 early folds are deformed by upright Acadian folds with associated spaced cleavage, and still younger (late Acadian?) kink bands.
- Consolidate transportation and purchase lunch materials if necessary. Proceed east on Route 2.
- 0.3      Long Trail crosses Winooski River and Route 2. Underhill Formation in roadcut on left.
- 1.4      Turn left (north) on Bolton Notch Road under Route I-89.
- 1.5      Small roadcut: Underhill Formation magnetite schist.
- 1.9      Where road becomes paved turn left onto road to gravel pit.
- 2.0      STOP 2: Park in gravel pit. Walk through woods along south edge of pit and follow former Long Trail 900 feet west along narrow glacial ridge (esker?) toward Duck Brook. First outcrops along trail are silver-green magnetite schist of Underhill Formation. F3 folds have a west-over-east sense of rotation, whereas F2 folds are east-over-west. Leave trail and go directly down to Duck Brook. Next large outcrops to west are Hazens Notch Formation: rusty and graphitic schist. These rocks are the westernmost exposures of Hazens Notch Formation we have found to date, and are believed to be a fault slice. Follow Hazens Notch outcrops west along south side of brook. Dark gray quartzite at 130 feet. Western contact with the Underhill (115 feet farther west) is marked by quartz veins which cut a tan-weathering "phyllonite". The actual contact is poorly exposed. Duck Brook descends over ledges of Underhill to the west. Return to vehicles and return to Route 2.

- 2.7 Turn left (east) on Route 2. View of Stimson Mountain to NE mapped by Eiben (1976).
- 3.1 Optional STOP A: Roadcuts both sides of road. Hazens Notch Formation: rusty-weathering, locally graphitic schist with black albite and pyrite.
- 3.3 Turnout: Rusty schist on north side of road. Projection from greenstone outcrops north of Route I-89 suggests that greenstone crosses Route 2 between this outcrop and Stop A to the west (see fig.2).
- 3.4 Turnout. View of north ridges of Camels Hump to SE mapped by Aubrey (1977).  
STOP 3: Roadcuts on north side of Route 2. West end of cut: quartz-feldspar-biotite granulite, massive with local (?) primary layering. This rock type is not typical of Hazens Notch Formation. Could it be a slice of Pinnacle Formation? It extends about a mile north from Stop 3. East end of cut: granulite interlayered with rusty Hazens Notch-like schist. The proportion of schist generally increases towards east end of cut. Quartz-calcite veins locally abundant. Joint surfaces coated with iron oxides. Foliation dips fairly steeply west. Proceed east on Route 2.
- 3.6 View north toward cut on Route I-89: Underhill magnetite schist. A layer of Underhill sandwiched within Hazens Notch forms most of the cliffs visible on the south slope of Stimson Mountain (fig.2).
- 4.7 Turn left (north) on road to Bolton Valley Ski Area (under Route I-89).
- 4.9 Optional STOP B: Devil's Washbowl, Joiner Brook. No hammers. Hazens Notch Formation: black albite schist with rusty horizons. Falls formed when glacier lowered base level in Winooski River valley.
- 7.0 Optional STOP C: Large, slanted outcrops north of road--smooth and slippery--use care! Underhill Formation: albitic layers define east-over-west dismembered F2 folds, some of which are deformed by F3.
- 7.5 Road crosses Joiner Brook. Underhill Formation. Foliation still dips west.
- 8.8 Small road cuts south of road. Underhill Forma-

tion. Foliation nearly horizontal.

- 9.0 Bolton Valley Ski Area. Bolton Mountain is highest peak to north. Park in public parking area below road.  
STOP 4: Hike up ski trail east of lift #3 (behind real estate office and swimming pool). Abundant outcrops of Underhill Formation from elevation 2240 to 2420', at top of lift #3, where we will go slightly down to the east to another ski trail which continues up to the south.
- Elevation 2440': View to north of Bolton Mountain. Conspicuous cliffs on Bolton (alternate Stop 4) are Underhill Formation with flat-lying foliation which is axial planar to large F2 folds. Farther east along the cliffs recumbent F2 folds deform greenstone found at the Underhill-Hazens Notch contact. The Long Trail follows the base of these cliffs and then goes to Bolton Mountain summit, which is rusty garnet schist of the Hazens Notch Formation.
- Elevation 2600': Ski trail reaches greenstone overlying Underhill Formation (contact is poorly exposed here). This contact trends northerly across the upper ski slopes to the Bolton Mountain location described above.
- Elevation 2630': Greenstone overlain by Hazens Notch Formation. Rusty schist is interlayered (or folded?) with greenstone at the contact.
- Elevation 2630 to 2780': Excellent exposures of typical Hazens Notch Formation (rusty biotite schist, locally very graphitic).  
 Return to vehicles and return to Route 2.
- 12.3 View of Camels Hump ahead to south: 3740 feet of relief.
- 13.3 Turn left (east) on Route 2. Bamforth Ridge to SE, across river: Underhill Formation.
- 13.5 Roadcuts to left (north): Underhill and Hazens Notch interlayered. Foliation dips west.
- 13.9 Cliffs to north behind trailer park are Underhill; the small valley between them and "Bolton Ledges" is Hazens Notch.
- 14.2 "Bolton Ledges": Tall cliff of Underhill.
- 14.6 Crossing anticlinorial hinge. Distant glimpse to north of cliff on Bone Mountain (elevation 2200');

flat lying late Taconian fault separates Underhill Formation in cliff from underlying Hazens Notch.

- 15.1 Turn left (north) on narrow dirt road.
- 15.3 STOP 5: Pinneo Brook. No hammers. Waterfall; potholes in Underhill Formation; foliation now dipping east. Albitic gneiss layer in east-over-west F3 fold. Return to Route 2.
- 15.4 Turn left (east) on Route 2, past roadcuts: mostly Underhill with minor rusty layers, dipping moderately east.
- 16.2 STOP 6: Turnout on north side of Route 2; roadcuts in woods if time permits. Watch traffic!

6A: Roadcut 0.1 mile west of turnout. Underhill Formation: silver-green magnetite schist and albitic gneiss. Atypical layer with large garnets toward west end.

At low east end of cut, "pseudo cross-bedding" formed by albitic layering (N4E, 43SE) and foliation (S1?) (N15W, 39NE). S2 foliation defined by chloritic partings is oriented N7E, 50SE. In remainder of cut, F3 folds are east-over-west and S3 spaced cleavage is nearly vertical.

30 feet from east end: compositional layering is cut by a fault surface that is hard to see in the fresh roadcut. Using extreme caution, follow along top of cut (easy access to top at west end) to find the fault on the natural outcrop, where it is deformed by an isoclinal fold (axial plane N7W, 51NE; hinge plunges east). Whether this is an F1 or F2 fold, the fault must be older (before peak Taconian metamorphism).

6B: Roadcut east of turnout. Be careful under tall roadcut--watch for falling rocks. Hazens Notch Formation. Continue on foot to greenstone exposed at county line sign. This greenstone can be followed to the north to Stop 4, but here it has Hazens Notch Formation on both sides. The details of this map pattern are being worked out as of this writing; a revised edition of figure 4 will be available in October. Return to vehicles and proceed east.

- 16.8 Long roadcut under power lines; second greenstone within Hazens Notch Formation. The same greenstone is also exposed below Bolton Dam and south of the Winooski River. Small body of talc-carbonate schist near east end of cut.