TECTONIC EVOLUTION OF THE TACONIAN HINTERLAND, GRANVILLE-HANCOCK AREA, CENTRAL VERMONT

A Thesis Presented

by

Thomas R. Armstrong

to

The Faculty of the Graduate College

of

The University of Vermont

In Partial Fulfillment of the Requirements
for the Degree of Master of Science
Specializing in Geology

May, 1993

TECTONIC EVOLUTION OF THE TACONIAN HINTERLAND, GRANVILLE-HANCOCK AREA, CENTRAL VERMONT

A Thesis Presented

by

Thomas R. Armstrong

to

The Faculty of the Graduate College

of

The University of Vermont

In Partial Fulfillment of the Requirements
for the Degree of Master of Science
Specializing in Geology

May, 1993

Accepted by the Faculty of the Graduate College, The University of Vermont, in partial fulfillment of the requirements for the degree of Master of Science, specializing in Geology.

Thesis Examination Committee:

Rolfe S. Stanley, Ph.D.	Advisor
Barry L. Doolan, Ph.D.	
Russell M. Agne, Ph.D.	Chairperson
Lynne A. Bond, Ph.D.	Dean, Graduate College

Date: February 12, 1992

ABSTRACT

Bedrock geologic mapping combined with structural, petrographic, lithologic, and stratigraphic analyses indicate that rocks within the Granville-Hancock area of central Vermont are part of the Late Proterozoic (?) - Lower Ordovician (?) Iapetan passive margin sequence deformed during the Ordovician Taconian and Devonian Acadian orogenic events. The sequence of deformation Taconian deformation events was (from oldest to youngest): **D1** Progressive westward imbriaction of discreet lithologic packages along at least two different thrust faults (**T1**); the Ottauquechee and Granville Formations were emplaced westward above the Pinney Hollow Formation and Mt. Abraham schist along the Ottauquechee Fault. The Pinney Hollow/Mt Abraham package was emplaced westward over the Hoosac, Monastery and Battell Formations along the Child's Mountain Fault. Both faults predate Taconian garnet/kyanite grade Barrovian metamorphism (**M1**). **D2** deformation is characterized by at least three distinct north striking, east dipping schistosities; **S2.1** fabric is syn M1,the dominant fabric in the western part of the field area, and a relict fabric in the central and eastern parts. **F2.1** folds are west verging and open in the west; relict F2.1 in the central and eastern areas are tight, reclined, and isoclinal. **S2.2** schistosity is preserved in the western area as a crenulation cleavage that increases in intensity progressively eastwards where it is a well developed mylonitic schsitosity, the dominant fabric in the central and eastern areas. S2.2 postdates M1 and is coeval with sub-garnet grade retrogression (**M2**) related to fluid infiltration within mylonitic fault zones (**T2**). **F2.2** folds in the west are open, upright, crenulate and kink folds; in the central and eastern parts of the field area F2.2 folds are tight, isoclinal sheath-like folds with hinges sub-parallel to an elongation mineral lineation. F2.2 folds are commonly truncated by an axial planar foliation within younger T2 fault zones, implying progressive development of syn-M2 fabrics within a similar (continuous) strain state. **S2.3** schsitosity is locally developed within the eastern part of the field area, and is typically a crenulation foliation, axial planar to weakly developed, west-verging , overtruned folds. **S2.3** is coplanar with undeformed S2.2, and probably a late-stage coninuum within the D2 strain state. Several late-stage deformations have been observed within the eastern part of the field area: **S3** northeast trending, steep west dipping pressure solution/crenulate cleavage is axial planar to mesoscopic upright chevron folds which appear to be parasitic to at least one map-scale , northeast trending and plunging antiformal structure along the crest of the Northfield Mountains. This structure is manifested by the changes in strike of S2 fabric in proximity to S3 deformation. Based upon regional structural considerations, S3 is believed to be Acadian. **T3** post-metamorphic faults appear to form within older D2 fabrics and are generally found within the eastern part of the study area and are characterized by brittle fabric, slickenlines and offset of brecciated S2 fabric. Kinematics within these zones indicate both reverse and normal motion; these zones may have formed during either later parts of the Acadian orogeny or during opening of the Atlantic ocean during the Mesozoic.

ACKNOWLEDGEMENT

Many people have contributed to my education and understanding of New England geology (and geology as a whole) over the last decade. This education started at the University of Massachusetts where Professors Peter Robinson, Don Wise, and the late Leo Hall provided me with the fundamental skills that are a requisite for bedrock geologic mapping and structural geology; the long hours spent with them in the field on various projects taught me how important it is to start any geologic project by "looking at the rocks". The field and mapping skills acquired from them certainly helped in my acclimatization to the subtle complexities of Vermont pre-Silurian geology. Time spent in the field with my friend David Elbert, in the Bronson Hill Terrane of southwestern New Hampshire, was valuable in that it was my first dose of bedrock mapping on a large scale. Dave served as both a teacher and colleague, and was very effective in demonstrating the importance of understanding the relationship between metamorphism and deformation in order to solve structural and tectonic problems.

The basic skills I learned at UMass were well honed during my time at Vermont. Field and laboratory work in several classes with Barry Doolan, Judy Hannah, and Jack Drake taught me a great deal about aspects of geology that significantly complimented my mapping skills. In particular, Barry's course on Appalachian geology provided a stimulating look at lithologic and stratigraphic similarities and differences between possible correlative rock units throughout New England. This information was combined with a treatise on passive margin evolution, which stressed that one must attempt to reconstruct a passive margin environment from lithologic sequences as an independent test of structural retrodeformations. I consider this class to be one of the very best, and most important, of my geologic career. In addition, much was learned from discussions with Jeff Prewitt, Sam Haydock, Vincent Dello Russo, Bill Dowling, Maurice Colpron, Tim Mock, Chris Kimball, Athene Cua, Jerome Kraus, Marian Warren, and especially, Greg Walsh, helped to clarify and organize alot of the facts, ideas and conclusions reached in this thesis.

Of course, there is always one person that stands out as having profoundly assisted in shaping a student's career. As with many others before me (and unfortunately, for me, many since !), Rolfe Stanley is that person who has played the largest role in directing my education and guiding me in the way I conduct scientific endeavours. Through a series of rigorous, but very illuminating and informative field-based courses, supplemented by necessary classwork on theoretical subject matter, Rolfe has been able to instill into several generations of geologists the importance of looking for, finding, and describing obvious to extremely subtle, simple to extremely complex geologic features. Time and time again, Rolfe was able to demonstrate the importance of these features and how necessary the task in properly identifying them in order to solve geologic problems. The most important lesson I have learned from him is the one I think of every time I'm mapping in the field; "what do the rocks tell you?". Although this may sound crazy to some (maybe to most), its importance lies in what it implies; make sure your conclusions are strictly founded on facts aquired from the rocks. This advice is unfortunately overlooked by many, and is a major reason for lapses in advancement of geologic mapping. The highest compliment that I can pay to Rolfe is that I will follow his advice, I will instill the same values into those whom I giude, and I will try to reach the high standards of professionalism that he has helped many geologists strive for, and which, through his own work, has helped define.

All too frequently, those closest to us make the biggest sacrifices, give us the greatest moral, financial and intellectual support but are left out when the time comes for giving credit where credit is due. My wife, Kathleen, has been the major driving force behind my finally finishing this thesis. Kathleen has played many roles, including draftsperson, artist, technical support staff, financial backer, editor, (a very critical one, I might add), and chairperson for field assistant recruiting (including our dogs Nikita and Natasha, and most recently, our son, Ryan Thomas). She has been the most understanding and patient friend anyone could ask for and has put many of her own professional plans "on hold" in order to provide support for mine. Although far from demonstrating the gratitude and love I feel for her, I dedicate this thesis to her.

Table of Contents

List of Tables

List of Figures and Plates

CHAPTER 1
INTRODUCTION

Location and physiography

The study area is located within the towns of Granville, Hancock, Ripton, and Rochester, in the Hancock and eastern part of the Bread Loaf 7.5 minute quadrangles, Vermont (figure 1.1). The field area is defined by a 45 square mile quadrilateral whose eastern border is the crest of the Northfield Mountains and the western border straddling the crest of the Green Mountain axis seven miles to the west of the eastern border. The northern border is the Hancock and Bread Loaf quadrangles northern boundary, approximately 1.3 miles north of the town of Granville. The southern border lies along Route 125, west of the center of Hancock, and follows the Howe Brook road east to the Northfield Mountains (figure 1.1).

The physiography of the study area varies dramatically, east to west; along the eastern boundary, the Northfield Mountains rise steeply above the valley floor to a maximum height of 2950 feet, with a mean vertical relief of about 2000 feet. The western slope of these mountains provides the best outcrop control in the entire area. Immediately to the west, a broad, elevated plateau extends for over 2 miles further west to the western branch of the White River Valley. Although small drumlin hills within the plateau do contain a few outcrops of graphitic schist and interbedded blue quartzite, the plateau itself is largely a network of swamps.

Dispersed mountain peaks, attaining heights up to 2900 feet, extend another 4 miles west to the crest of the Green Mountains. These mountains are pervasively incised by numerous tributaries of the White River, including the Hancock Branch (figure 1.1). Most of these stream valleys have steep slopes that cut down through the bedrock to the valley floor,providing excellent outcrop control. The eastern slopes of the Green Mountains are overlain by glacial till and varve clay up to an elevation of about 2700 feet. Below this elevation outcrop is scarce, and is usually found only in large stream valleys such as the Hancock Branch. Above 2700 feet, outcrop is more abundant, although sometimes difficult to discern from the abundant float, eroded from the ridge crest.

Access within the study area was extremely good. The north-south trending White River valley contains Route 100, from which many subsidiary roads originate, leading west toward the Green Mountain crest. Forestry route 55 heads west from the town of Granville, following the uppermost section of the White River, providing access to the Bread Loaf Wilderness area (figure 1.1). Farther south, Route 125 heads west from the town of Hancock, providing access to Texas Falls, the Hancock Branch watershed, and

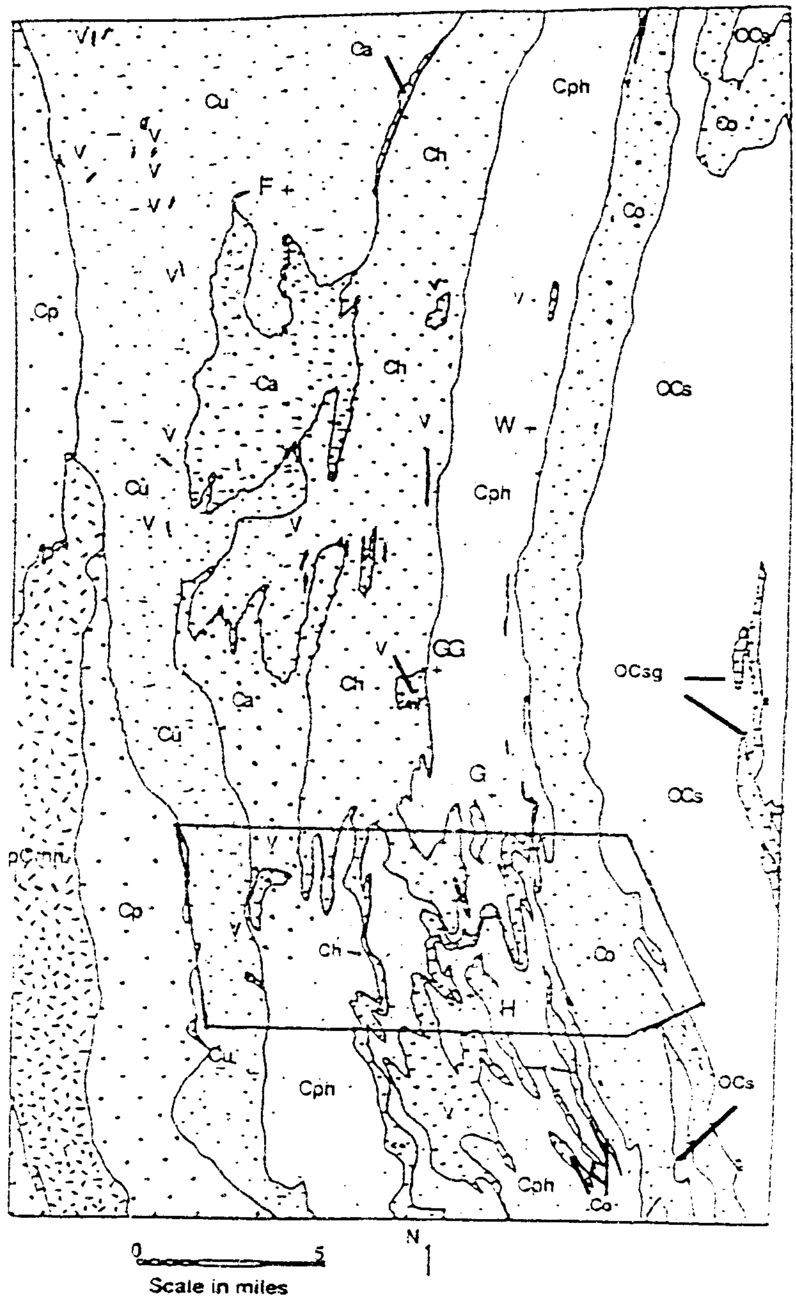


Figure 1.1 Regional geologic map of the Vermont eastern pre-Silurian sequence, compiled from Doll et al., 1961, and Cady et al., 1962. The Granville-Hancock study area is outlined. Contacts were interpreted as depositional with triple junction-like contact relationships interpreted as complex facies changes. Formation symbols are; pCmh- Precambrian (Middle Proterozoic) Mount Holly Complex, Cp- Pinnacle Fm., Cu-Underhill Fm., Ca- Mt. Abraham Mbr./Underhill Fm., Ch- Hazen's Notch Fm., Cph- Pinney Hollow Fm., Co- Ottauquechee Fm., OCs- Stowe Fm., OCsg- Stowe mafic volcanics. "V" symbols refer to mafic volcanics; black symbols are ultramafics. Reference points include: **G**- Granville, **H**- Hancock, **GG**- Granville Gulf, **W**- Warren, **F**-Fayston.

most of the southwestern part of the field area. East-west trending Clark Brook, Thatcher Brook, and Howe Brook roads connect with Route 100 and provide access to the Northfield Mountains along different latitudes within the area. Another north-south road along the immediate western slope of the Northfield Mountains also connects these roads (figure 1.1).

Regional geologic setting and history

The field area is located within the Vermont eastern pre-Silurian sequence, so named for its position along the eastern flank of the Middle Proterozoic age Green Mountain massif of southern Vermont and the Lincoln massif of central Vermont (Plate 3; Doll et al., 1961; Cady, 1969). The pre-Silurian cover sequence is traceable as far north as southern Quebec (Slivitzsky and St. Julien, 1987) and southward into the New York City area (Hall and Robinson, 1982; Stanley and Ratcliffe, 1985). The areal distribution of the pre-Silurian lithologies and associated tectonic fabric are believed to be related to North American continental collision with an easterly situated island arc during the Medial Ordovician Taconic Orogeny (Rowley and Kidd, 1981; Hall and Robinson, 1982; Hatch, 1982; Stanley and Ratcliffe, 1985).

First studied in detail at the end of the last century (Hitchcock et al., 1861, and numerous predecessors), the pre-Silurian cover-basement sequence received great attention during pioneering studies made before the advent of plate tectonic theory (Perry, 1928; Clark, 1934; Cady, 1945, 1956, 1969; Osberg, 1952; Brace, 1953; Chang et al., 1965). Perry (1928) was the first geologist to separate the eastern sequence into mappable lithologic units, including the Pinney Hollow and Ottauquechee Formations; names still used by present day workers. Clark's 1934 study included a stratigraphic description of the Oak Hill Group, the northern continuation of the Vermont sequence in Quebec. Cady's work in Vermont from 1937 until the early 1970's included mapping, lithologic identification, and stratigraphic correlation of a significant part of both the eastern and western Vermont pre-Silurian sequences. Brace (1953) and Chang et al. (1965), and Thompson (1972) conducted mapping, lithologic identification, and metamorphic petrologic studies within the Bridgewater and Woodstock areas, respectively. Although their structural and lithologic studies did little towards augmenting or modifying the earlier work of Perry (1928), their petrologic descriptions and discussions set the stage for more classic studies on mineral phase equilibria, chemographic projections, and petrogenesis of metamorphosed rocks (Thompson, 1957; Albee, 1965a; Thompson et al., 1977).

Osberg (1952) mapped within the Granville-Hancock area at 1:62,500 scale and included a structural, stratigraphic, and metamorphic analysis of the eastern sequence as part of a larger project involving a transect across the eastern sequence, the Lincoln massif, and the western sequence which includes clastic and carbonate sequences. Osberg continued the usage of Pinney Hollow, and Ottauquechee Formations of Perry (1928) for similar rock-types comprising part of an interpreted eastwardly younging homoclinal stratigraphic sequence (see chapter 2). Within his structural and stratigraphic models, Osberg considered the eastern sequence a deep water continuation of the western quartzite and carbonate sequence, both floored unconformably by basement rocks exposed within the Lincoln massif (figure 1.2). Although this cover sequence lacked stratigraphic symmetry, he considered the exposed massif to be the core of a large anticlinal structure, previously interpreted as regionally continuous and called the Green Mountain anticlinorium (Walcott, 1888).

Stratigraphic units within Osberg's study were compared to similar ones within the classic Taconic range of southwestern Vermont and eastern New York State. Because of lithologic similarities between the two sequences, Osberg proposed the eastern cover as a "root zone" for the supposedly allochthonous Taconic rocks. Prindle and Knopf (1932) had made a similar proposal for Taconic-like lithologies on top of marble and quartzite at Mt. Greylock in northwestern Massachusetts. This proposal was the first one within Vermont to consider the eastern sequence as an autochthonous correlative of the Taconic range units. Because he considered the exposed pre-Silurian units to be stratigraphically coherent, Osberg did not speculate as to the position of the root zone. Instead he portrayed the Taconic rocks rooting from a position farther to the east, below the Silurian-Devonian unconformity (Osberg, 1952; Osberg, pers. comm., 1988).

During the advent of plate tectonic models, several studies began to question many of the earlier interpretations concerning Taconic geology. Based on fossil ages, Dale (1917), Keith (1932), and Zen (1961) proposed that the Taconic slates were allochthonous with respect to underlying flysch-type slates and carbonate platform rocks. He later suggested that the structurally lower Taconic rocks (lower Taconic slices) were emplaced as unconsolidated sediments by way of sediment/fluid gravity sliding from an eastern uplifting source region and that the higher Taconic slices were emplaced above the lower slices along hard rock thrusts (Zen, 1967). Because of the abundance of soft sediment features, particularly within Zen's Giddings Brook slice of the lower Taconics, he favored a gravity slide model for the emplacement of the entire package from a previous "tectonic staging area" situated above the Green Mountain massif. Subsequent studies based on sedimentologic analysis of Taconic allochthonous and autochthonous rock units corroborated with Zen's proposal (Bird and Dewey, 1970).

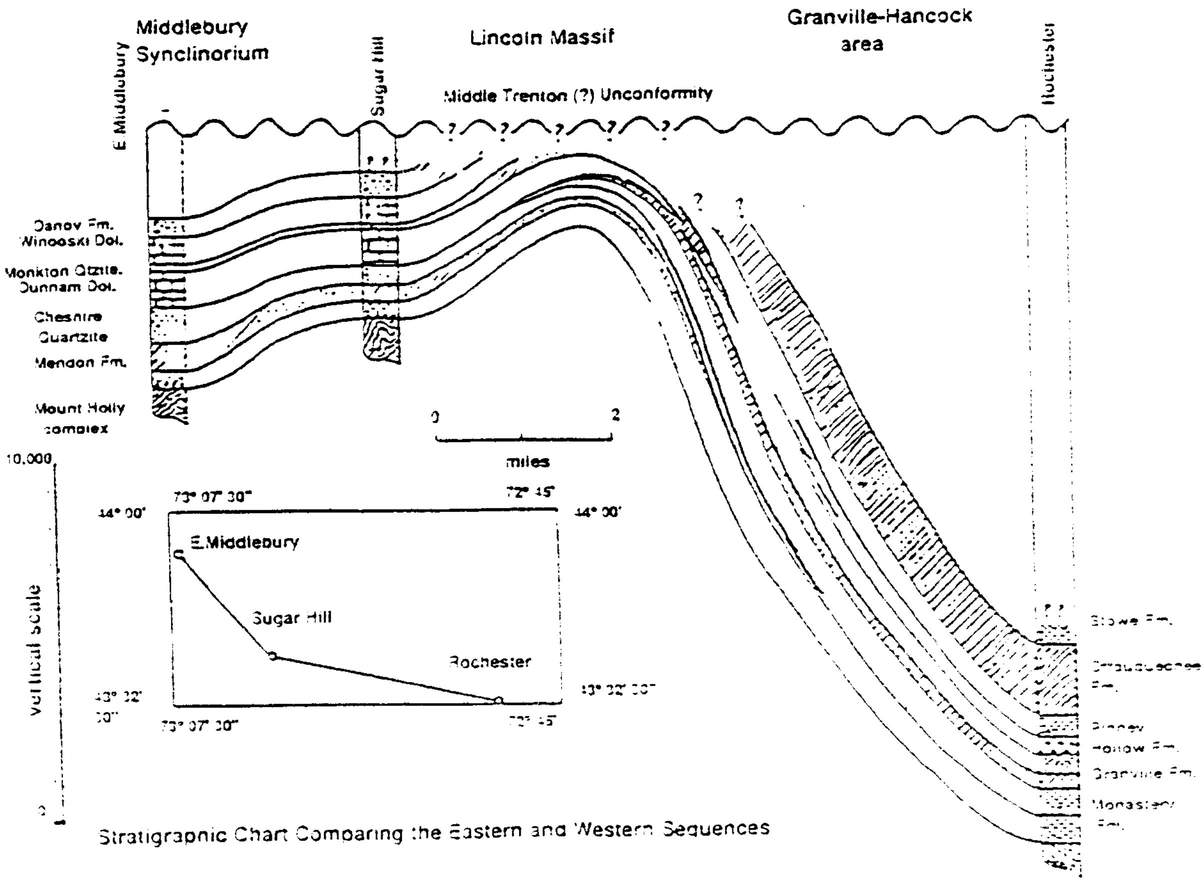


Figure 1.2 Osberg's 1952 stratigraphic correlation diagram of western and eastern Vermont pre-Silurian lithologies. Correlation of units is made across the Precambrian (Middle Proterozoic) anticlinal core of the Lincoln massif. Facies changes between quartzite and carbonate rocks of the Champlain Valley sequence and metasedimentary and volcanogenic sequences of the eastern Vermont sequence were believed to occur above the anticlinal core, although no such facies change was ever recognized on either the northern or southern closures of the massif (figure 6 from Osberg, 1952).

Later tectonic models favored hard rock thrusting and emplacement of the Taconic slices from a site on top of (Thompson et al., 1986; Karabinos, 1988) or east of the external basement massifs (Stanley and Ratcliffe, 1985). This implied that the eastern cover rocks were not a coherent homoclinal sequence, as portrayed by Osberg (1952), Doll and others (1961), Cady and others (1962), Chang and others (1965), and Thompson (1972). The development of an eastern cover "root zone" for the Taconic allochthon came directly from the work of several geologists working in Massachusetts and Connecticut during the 1970's (Ratcliffe, 1975; Harwood, 1972,- 1975; Zen et al., 1983; Stanley and Ratcliffe, 1985). These studies demonstrated the tectonic nature of contacts between many of the major lithologic units, giving rise to the term "lithotectonic sequence" (Stanley and Ratcliffe, 1985). Based on lithologic and structural arguments, Stanley and Ratcliffe (1985) suggested that the root zone for the Taconic allochthon lay along the Whitcomb Summit and Hoosac Summit Thrust Zones, as mapped within western Massachusetts (figure 1.3). This root zone site lay east of the more coarse grained clastic rocks of the Dalton and Hoosac Formations and west of the Rowe-Hawley ultramafic-bearing zone, interpreted as tectonized remnants of an accretionary wedge/forearc sequence with tectonic slivers of altered oceanic crust (Stanley et al., 1984; Stanley and Ratcliffe, 1985; figure 1.4). These thrust zones were extrapolated north into Vermont along the interpreted depositional boundaries of Doll and others (1961). Because of the speculative nature of this modified tectonic interpretation, Stanley and Ratcliffe (1985) discussed the need for future work in Vermont in order to substantiate or refute their tectonic model. The work of Roy (1982) in northern Vermont, and Tauvers (1981), and DiPietro (1982) in central Vermont, demonstrated the presence of numerous thrust faults within the eastern cover pre-Silurian sequence. Later, DelloRusso (1986), DelloRusso and Stanley (1986), O'Loughlin and Stanley (1986), and Lapp and Stanley (1987) demonstrated similar tectonic geometry within central Vermont Middle Proterozoic basement (the Lincoln massif), and eastern cover lithologies. These studies also discussed the presence of thrust faults both internal and bounding discreet lithologic packages. Stratigraphy immediately adjacent to the basement was believed to be coherent in many instances (DelloRusso, 1986), but became progressively tectonized eastward into the cover lithologies such that no stratigraphic cohesion was believed to exist and large lithotectonic units were interpreted as incorporation of numerous thrust slivers (Lapp and O'Loughlin, 1986).

Subsequent studies by Haydock (1988) and Prewitt (1989) within the Warren-Waitsfield, Walsh (1989) in the Fayston area, Kraus (1989) in the Northfield Mountains, and Armstrong and others (1988a, 1988b) in the Granville-Hancock area further

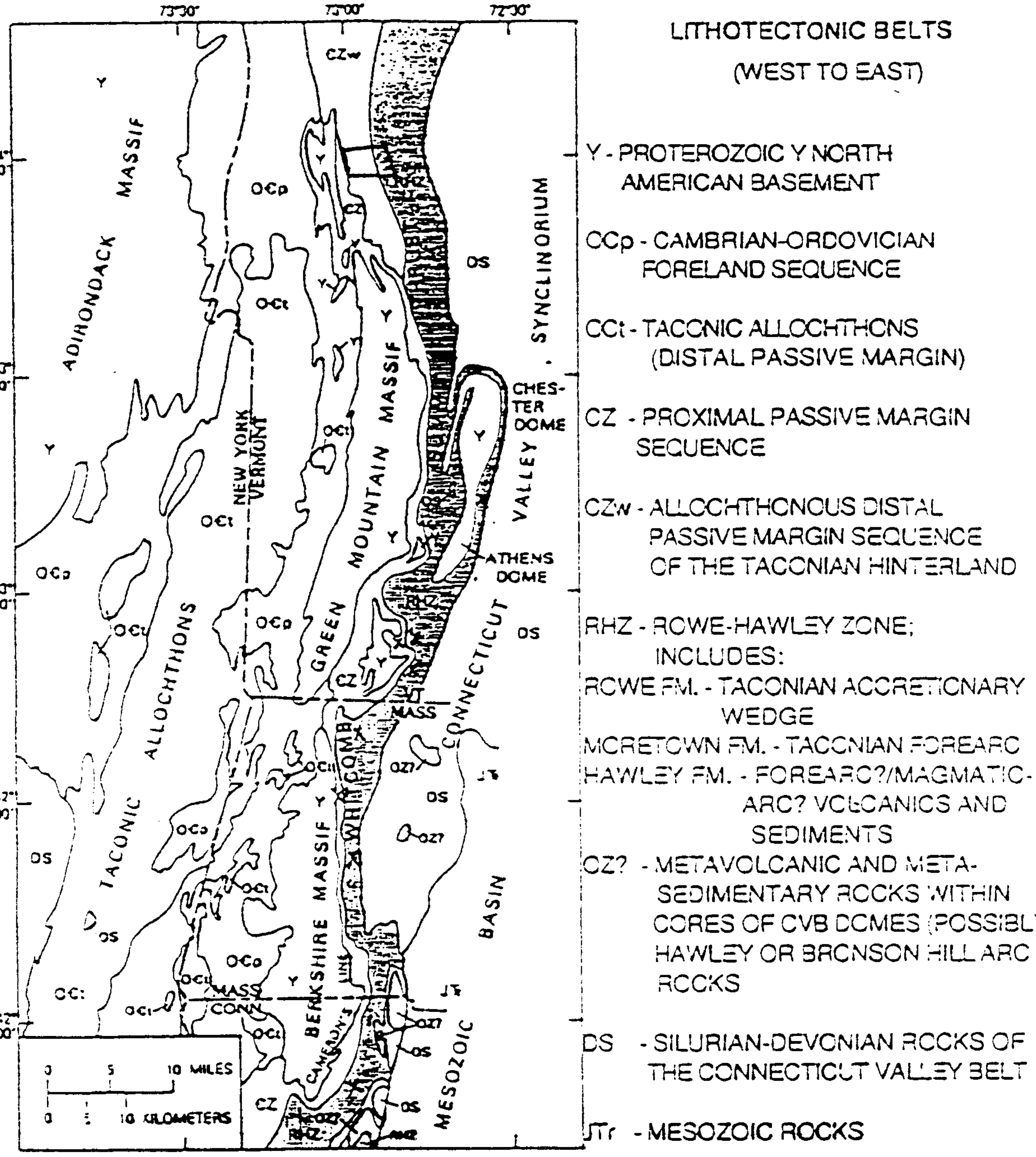


Figure 1.3 Lithotectonic diagram of western New England pre-Silurian rocks, showing the major Taconian thrust packages as interpreted by Stanley and Ratcliffe (1985). Taconic allochthons are believed to root underneath the Hoosac Summit Thrust (Cameron's Line of figure), flooring the allochthonous Hoosac sequence, and the Whitcomb Summit Thrust (WST) flooring the accretionary wedge rocks of the Rowe-Hawley Zone (RHZ; see text and Stanley and Ratcliffe, 1985 for discussion). CZw is the Pinney Hollow (PHS) and Hanzens Notch (HNS) Slices of Figure 1.4.

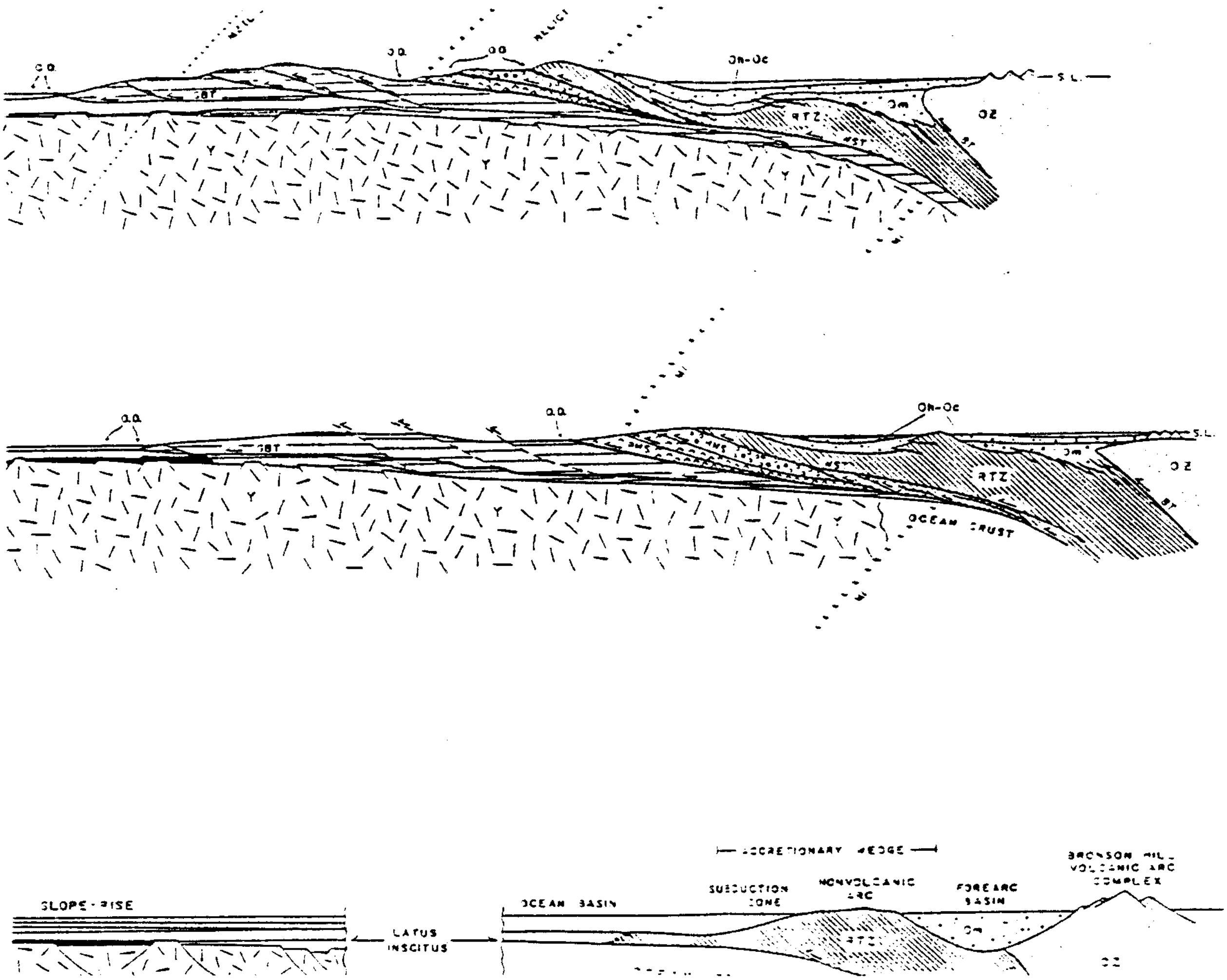


Figure 1.4 Interpretive geologic cross section across the palinspastically restored, Lower Ordovician Taconian convergent margin. The lower diagram depicts the early accretionary wedge (composed of Rowe Belt rocks) converging with the North American continental margin. The middle diagram shows the wedge following initial margin convergence with the development of the WST, underplating of the eastern Vermont sequence (HNS and PHS), and initial ejection of the Taconic allochthon lithotectonic assemblage (horizontally ruled). In Vermont, the Taconic root zone would thus be the fault flooring the eastern Vermont slices and not the WST. This fault would be the northern analog of the HST (figure 1.3). The upper diagram depicts further convergence; in Massachusetts, the eastern Vermont sequence is entirely truncated (buried) by the RTZ/WST, which becomes the root zone at that particular latitude (from Stanley and Ratcliffe, 1983). GBT-Giddings Brook Thrust, O.D.-olistostromal deposits, Oh-Oc- Hawle/Cobble Mtn. Slice, OZ-Bronson Hill arc complex.

substantiated the previous tectonic interpretations but also included several other generations of faults including ones predating the peak of metamorphism. These faults were believed to be responsible for most of the shortening across the belt, juxtaposing discreet lithotectonic assemblages. Tauvers (1982), DiPietro (1983), Armstrong (1989a, 1989b), and Walsh (1989) also described coherent stratigraphic sections within the western part of the cover sequence which they both compared to similar sequences within the Taconic allochthon. These eastern cover sequences were interpreted as more proximal (westerly) correlatives of Taconic lithologies and were further interpreted as immediate lower plate sections of the pre-peak metamorphic thrust (the Child's Mountain Thrust in the Granville-Hancock area) believed to be the Taconic root zone. Geochemical analysis of metamorphosed greenstones within the Middle Proterozoic basement, eastern cover sequence, and Taconic allochthon also suggests that the Child's Mountain Thrust is a likely candidate for the root zone (Coish et al., 1985, 1986; Stanley et al., 1988a, 1988c, 1989).

Purpose of this study

This study was initially developed as part of a more comprehensive bedrock geologic mapping project in central Vermont proposed and directed by Rolfe S. Stanley of the University of Vermont. The purpose of this and other central Vermont projects was to map the areal distribution of individual lithologic units and different generations of superposed deformational fabrics including fault zones, folds, and associated planar fabric. The conclusions brought forth from these projects would shed light as to the validity of both previous models regarding the belt as a coherent homoclinal sequence, and modern hypotheses which have described the belt as a highly tectonized, thrust-bound package of incoherent lithologies.

Significance of study

The conclusions derived from this study will contribute directly to the compilation and development of a new Vermont State Geologic map. Interpretations regarding the structural and tectonic evolution will hopefully be of assistance to future workers and useful to any subsequent models concerning the regional tectonic evolution of Vermont during the Taconic orogeny.

Working hypothesis

The rocks within the central Vermont pre-Silurian sequence are hypothesized to have been part of a Late Proterozoic-Middle Ordovician passive margin sequence along the eastern part of the North American continent. This passive margin sequence was initially destructed during accretion with a proposed magmatic arc terrane during the Middle

Ordovician, resulting in the development of an accretionary wedge above an eastwardly dipping subduction zone. Deformation and metamorphism of these rocks was a direct result of this accretion (B-Type subduction; Hodges et al., 1982) and subsequent continent collision (A-type) with the magmatic arc terrane (Bronson Hill arc complex of Stanley and Ratcliffe, 1985).

A broad review of the literature, including recent theses completed within the central Vermont eastern cover sequence, describes complex multiple episodes of deformation and metamorphism. Radiometric, fossil, and unconformity age relationships all have suggested that most of the deformation and metamorphism within this belt was the result of Taconian orogenic processes.

Based on these facts, bedrock geologic mapping was conducted assuming that the rocks of central Vermont were pre-Silurian in age and polydeformed and metamorphosed during the Taconic orogeny. Criteria were established in order to deduce the depositional and/or tectonic nature of contacts between major lithologic types (appendix 2). Finally, the tectonic evolution of the belt was then compared to modern day collisional environments.

Methods of study

Detailed mapping of lithologic and fabric distribution was conducted at a scale of 1:12,000 over the entire mapped area. Detailed maps of critical localities were done at larger scale, where necessary. Working field maps were photocopied from a 1:12,000 scale mylar base. This base was enlarged from standard 1:24,000 scale U.S.G.S. 7.5 minute quadrangle maps of the Hancock and Bread Loaf quadrangles.

Locations of outcrops were identified through standard pace and Brunton technique and the use of a barometrically calibrated altimeter. Structural measurements, including fabric and linear element orientations, were obtained through the use of a standard Brunton geologic compass.

Pertinent hand samples were collected from various localities in order to establish very general modal analyses of mineral types within the major mapped rock-types. These analyses were obtained through production of rock thin sections and petrographic study with the aide of a petrographic microscope. Microscopic observations of planar elements in and away from fault zones were also conducted in order to establish the deformation mechanism responsible for fault zone and schistosity fabric development.

Structural data were compiled and separated into three areal zones, roughly coincidental with transitions in fabric intensity, orientation, or metamorphic grade. The compiled data were then separated into chronologic order and plotted on Schmidt equal area nets. Where in order,data were contoured numerically for density in percent per area of net. Chronologically correlative data from the different areas were then compared. These data, along with subsequently developed structural, stratigraphic, and metamorphic interpretations, were then used to constrain across strike cross sections for the study area.

CHAPTER 2
STRATIGRAPHY

The pre-Silurian eastern cover stratigraphy will be described from west to east across the study area, starting with lithologies found on the east flank of the Middle Proterozoic Lincoln massif (figure 2.1). Due to a complex history of deformation, most of the pre-Silurian units are bound by several generations of thrust faults. Absence of any fossils, combined with a general lack of distinctive stratigraphic markers, precludes description of this sequence in a demonstrable chronologic order. The Middle Proterozoic gneiss of the Lincoln massif, with depositionally overlying cover, serves as a good reference point for describing the various eastern cover lithologies (see DelloRusso, 1986 and Warren, 1989 for a complete description). Although tectonic overlap of eastern cover units is significant, major north-south trending belts of discreet units have been recognized by earlier workers (Perry, 1928; Osberg, 1952; Cady et al., 1962). These belts allow objective west to east description of the pre-Silurian sequence without regard to relative ages. Where tectonic overlap of major belts is significant, the formations will be described in ascending structural order.

The stratigraphic descriptions will be divided into four sections for each formation:

1. An introduction into the origin of the formation name, type locality, a description of lithologies previously and presently included, rationale for stratigraphic grouping of lithologies, and any other pertinent information from previous studies. Emphasis will be placed on hand sample analysis, augmented by microscopic study. Modal analyses of particular samples will be included as accompanying figures. This section may include a stratigraphic summary of the pertinent lithologic characteristics.

2. A discussion of contact relationships with adjacent (overlying and underlying) units, specifically, those localities where the contacts are best illustrated and controlled.

3. This section will specify localities where particular lithologies of previously named formations are best exposed. It will also include type localities for newly proposed and/or redefined formations.

4. A brief discussion on possible depositional environments will be based upon both local and regional lithologic and compositional characteristics of the formations.

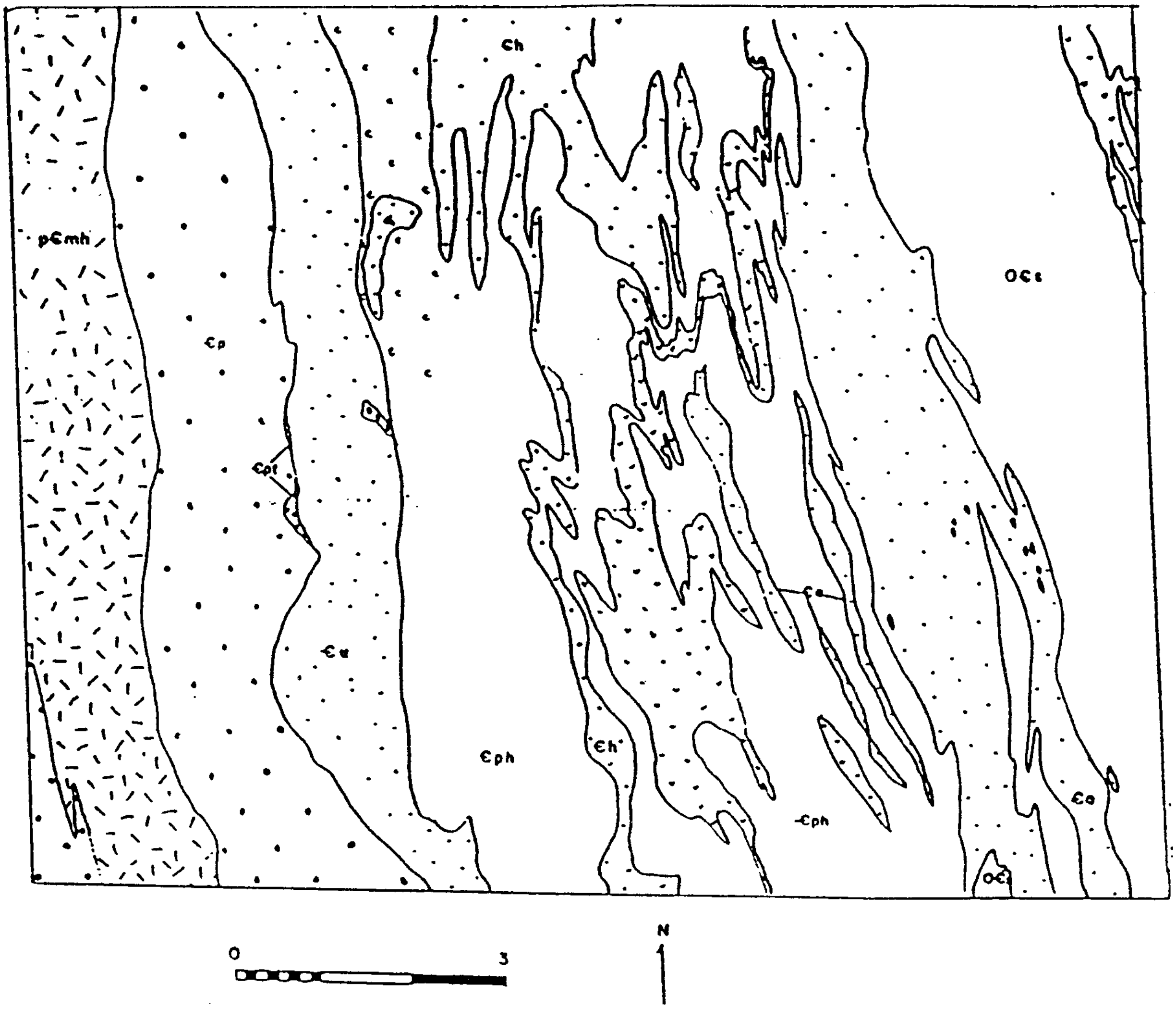


Figure 2.1 Geologic map of the Vermont eastern pre-Silurian sequence, as portrayed by Osberg (1952) and Cady et al.(1962). The sequence was interpreted as a coherent east dipping homocline, progressively younging from west to east, including: **pCmh:** Mt. Holly Complex (Grenville basement), **Cp:** Pinnacle Fm., **Cpt:** Tyson Mbr., **Cu:** Underhill Fm., **Cph:** Pinney Hollow Fm., **Ch:** Hazens Notch Fm., **Co:** Ottauquechee Fm., **Ocs:** Stowe Fm. "V" symbols designate intraformational greenstones; black spheres are ultramafic pods.

Hoosac Formation (CZh)

Previously included within the Middle Proterozoic basement of the Lincoln massif by Osberg (1952), this sequence of fine grained quartz-feldspar wacke, aluminous schist, quartzite, and polymict conglomerate was later included as part of the Upper Proterozoic/Lower Cambrian Hoosac Formation by Doll et al. (1961) on the Centennial State Map (figure 2.1). This name is acquired from the type locality section along the crest of Hoosac Mountain in northwestern Massachusetts (Prindle and Knopf, 1932; Ratcliffe, 1979). Only the easternmost part of this formation was studied along the crest of the Green Mountains (plate 1). The following description of lithologies does not include a significant part of this unit which extends to the east flank of the Lincoln massif in the Ripton area (Warren, 1989). This unit is described for the exclusive purpose of providing a comprehensive description of the entire pre-Silurian cover sequence (see plate 1).

Description:

The Hoosac lithologies observed during field mapping include the following:

1. Well foliated, tan weathering, fine grained quartz-feldspar-muscovite wacke, with rounded, presumably detrital blue quartz grains, indicative of a high grade (amphibolite to granulite facies) source area. This entire sequence, along with rare, thin (1 m thick), vitreous, white quartzite, is present as discreet horizons within albite schist, similar to the adjacent Monastery Formation, found primarily to the east of the Green Mountain's crest (plate 1).
2. A polymict conglomerate, containing well rounded and unsorted clasts of vein quartz and quartz-feldspar gneiss that is similar to basement rocks farther west, within the Lincoln massif. This particular conglomerate was found within the well foliated, tan weathering metawacke.

Contact relations:

The lower contact of the Hoosac Formation is very complex, and is either tectonic or depositional. In the Ripton area, Hoosac wacke depositionally overlies tectonic slivers of Grenville basement, comprising an anastomosing synmetamorphic shear zone geometry (Warren, 1989). This contact is usually marked by polymict conglomerate with variably sized clasts (Warren, 1989). In the South Lincoln area, Hoosac lithologies tectonically overlie Middle Proterozoic gneiss within the Lincoln massif along several large scale ductile thrust zones, including the South Lincoln Thrust Zone (DelloRusso, 1986; DelloRusso and Stanley, 1986).

The contact between the Hoosac and Monastery Formations is continuously marked by well developed fault zone fabrics, tectonic slivers of Hoosac conglomerate, and juxtaposed aluminous schist (Mt. Abraham Formation), creating a cartographic "triple junction" (plate 1, J6). This contact, the Underhill Thrust Zone, is traceable along the entire western border of the study area, and is believed to continue north into the Lincoln area, and south into cover along the east side of the Green Mountain massif, where it is called the Hoosac Summit Thrust Zone (plate 3; Stanley and Ratcliffe 1985; Ratcliffe et al., 1988).

Numerous meter thick layers of Hoosac-like wacke occur within the upper plate (eastern side) of the Underhill Thrust Zone. The frequency of these layers progressively decreases to zero over a 1km distance across the compositonal layering (parallel to a regional tectonic foliation). This progressive decrease may be indicative of a pre thrust, interlayered depositional Hoosac/Monastery contact zone.

Depositional environment:

Tauvers (1982), based on work in the Lincoln area, noted that the coarseness and detrital content of the Hoosac lithologies increased westward where the Hoosac grades into even coarser rocks of the Pinnacle Formation. He interpreted the Hoosac clastic sequence as the deep water part of an alluvial/fluvial deltaic fan complex originating from possibly several western sources. The abundance of detrital blue quartz, and relict calcium-rich plagioclase, as well as quartz-feldspar gneiss conglomerates, supports the notion of a high grade, Grenville basement source.

Monastery Formation (CZm)

The Monastery Formation was named for exposures of quartz-chlorite-garnet-biotite schist, quartzite, and graphitic schist, present along the northeast slopes of Monastery Mountain, south of the Robbin's Branch of the White River (Osberg, 1952, p. 42-45). The original description included all rocks lying between the Lincoln massif and rocks of the Granville Formation, to the east (figure 2.1). Included in this unit was a pebble to boulder conglomerate called the Tyson Member, for exposures identified within the Hoosac Formation in the village of Tyson, south-central Vermont (Perry, 1928). This conglomerate is found along the Underhill Thrust Zone, separating the Monastery from the Hoosac wacke. The relationship of the conglomerate with either formation is unclear due to enveloping fault zone fabric. Immediately above the conglomerate, Osberg (1952; p.42-45) described a sequence of fine grained wacke with minor albitic schist, interfingering eastward into a large volume of albite porphyroblastic, quartz-muscovite-chlorite schist, aluminous quartz-white mica-chloritoid-chlorite-albite schist, and white quartzite layers.

Grading upwards from the aluminous schist, Osberg (1952; p.46) also included a black graphitic schist with local laminations of carbonate (marble, dolomite) and locally thin to thick dolomite and marble; this was unit was named the Battell Member for exposures on the southeastern limb of Battell Mountain along the crest of the Green Mountains (plate 1). Doll et al., (1961), redefined the Monastery Formation as the Underhill Formation for along strike, correlative sequences to the north, mapped by Cady (1956, 1960).

Description:

The Monastery Formation is herein redefined to include albitic schist, interlayered aluminous schist, quartzite, newly discovered greenstone, and thin dolomitic metasandstone (dolarenite), interlayered with the albitic schist (figure 2.2).

The albitic schist includes light green to silvery gray, tan weathering, fine grained, quartz-albite-muscovite-chlorite-garnet-biotite-magnetite-ilmenite schist (figure 2.3). Quartz is the dominant constituent, comprising 20 - 40 modal percent. Albite occurs as equidimensional porphyroblasts, ranging in size from 1-6 mm in diameter. Other minerals vary in abundance and grain size (table 1). Some layers of the schist sequence contain garnetiferous horizons with porphyroblasts ranging in diameter from 2 mm to 3 cm. Locally, layers rich in magnetite porphyroblasts (1-3 cm in diameter) are found within the albitic schist and aluminous schist horizons. Chlorite-rich horizons are locally interlayered with thin (1 meter) horizons of biotite-rich granofels in the albitic schist member.

Aluminous schist is locally abundant within the albitic schist. This lithology is composed of quartz-white mica-chlorite-chloritoid with some zones abundant in garnet (table 1). Aluminous schist seems to be most abundant in the albitic schist near the contact with the overlying Battell Formation (figure 2.2). Albitic schist interlayered and in close proximity to the aluminous schist, appears to be more quartzose.

Locally interlayered with the aluminous schist and, in some localities, the albitic schist are highly recrystallized layers of vitreous white quartzite, without detrital grains (figure 2.2). The quartzite varies from massive to schistose, locally containing thin laminae of muscovite-rich schist. The matrix also contains a small amount of recrystallized albite, muscovite, and chlorite. Contacts with both schist lithologies are either sharp or gradational, with adjacent schist being very quartzose, and characterized as schistose quartzite.

Dolarenite Member (CZmd):

Locally interlayered with the albitic schist are thin (1 to 3m) discontinuous dolarenite (figures 2.2 & 2.4; table 1). This lithology consists of secondary dolomite, coarsely recrystallized with a granulite texture, and associated, highly oxidized and rounded, detrital

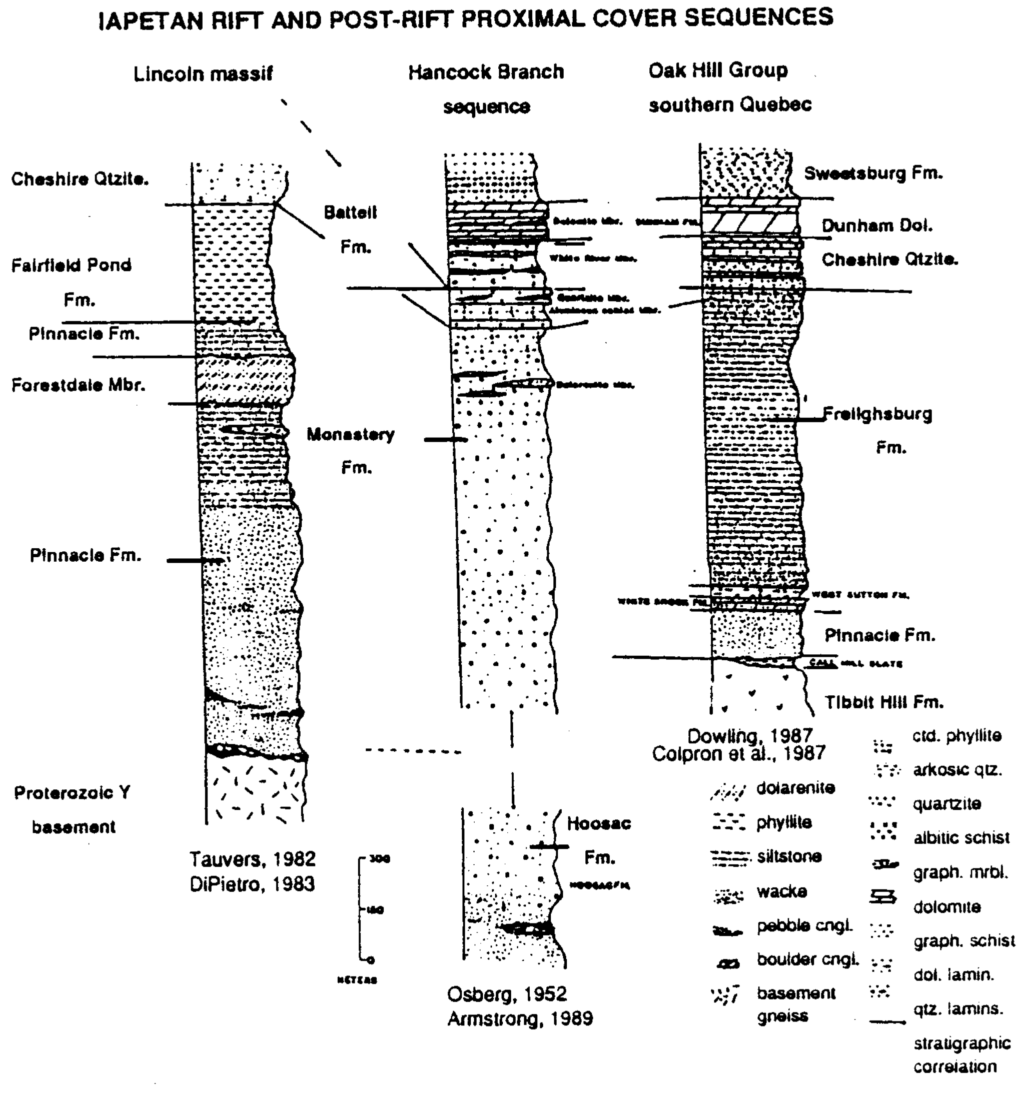


Figure 2.2 Comparative stratigraphic columns of proximal rift clastic and post-rift platform and slope-rise sequences. Arrows indicate lithologic datum planes, not necessarily time lines. The Lincoln massif sequence is from the west side of the massif, the Hancock Branch Sequence is from the study area, along the Hancock Branch of the White River, and the southern Quebec section is from the Oak Hill Group in the vicinity of Pinnacle Mountain and the town of Dunham. See text for a complete discussion of correlations and description of lithologies.

	HQ 511	HQ 515	HQ 516	HQ 519	HQ 521	HQ 522
Porphyroblasts						
garnet	-	-	-	1	3	-
albite	26	3	-	9	-	5
Groundmass						
quartz	15	69	62	24	35	30
albite	7	-	-	5	3	13
sericite	40	24	23	39	39	27
biotite	-	-	-	-	-	5
chlorite	8	3	4	20	18	17
chloritoid	-	-	6	tr	-	-
ankerite	tr	-	1	tr	tr	-
apatite	tr	tr	tr	tr	tr	-
alanite	tr	-	tr	-	-	-
pyrite	-	-	1	-	tr	-
tourmaline	1	tr	1	1	1	1
zircon	tr	-	-	tr	-	-
magnetite	2	1	1	1	1	1
ilmenite	1	tr	tr	tr	tr	1
Grain Size						
Porphyroblasts	1.0-3.5mm	1.0-2.0mm	-	0.5-2.5mm	1.0mm	1.0-3.0mm
Groundmass	0.2-0.5mm	0.2-0.5mm	0.1-2.0mm	0.1-0.4mm	0.1-0.5mm	0.1-0.5mm
Texture	schistose	schistose	phyllitic	schistose	schistose	schistose
Location	U 13	T 10	S 9	T 9	P 9	O 8

Table 1 Modal analyses of Monastery Formation samples.

Figure 2.3 Hand sample of tan weathering quartz-plagioclase-sericite-chlorite-tourmaline-ilmenite-magnetite schist from the Monastery Fm.

Figure 2.4 Discontinuous carbonate layers within nongraphitic, albitic schist of the Monastery Fm. These layers contain both fragmented, detrital quartz and magnetite grains, strongly dissimilar to the homogenous dolomite within the overlying Battell Fm. These dolarenites may represent small pulses of carbonate sands within a deltaic or lagoonal environment.

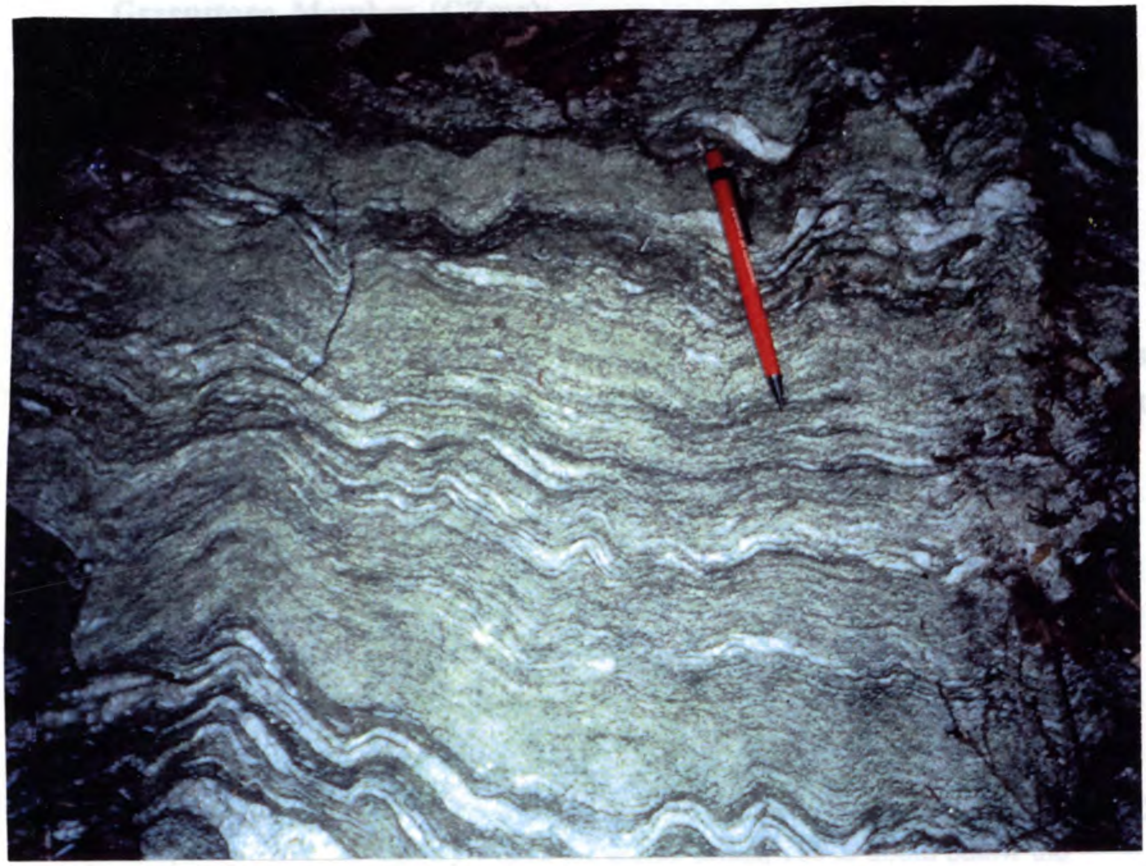


Figure 2.4 Discontinuous carbonate layers within nongraphitic, albitic schist of the Monastery Fm. These layers contain both fragmented, detrital quartz and magnetite grains, strongly dissimilar to the homogenous dolomite within the overlying Battell Fm. These dolarenites may represent small pulses of carbonate sands within a deltaic or lagoonal environment.

magnetite grains, angular to rounded fragments of feldspar, and few grains of blue quartz. These horizons, although found through-out the albite schist sequence, are generally separable as map scale units within the upper part of the very quartzose albitic schist near the interlayers of aluminous schist (figure 2.2).

Greenstone Member (CZmg):

Found within the easternmost exposures of the Monastery Formation albitic schist are several thin (1 to 3m wide), .5 to 1 km long dark green, coarse grained albite-chlorite-hornblende-carbonate-actinolite-magnetite metavolcanics or greenstones. Albite porphyroblasts occur pervasively in discreet layers and range in diameter from 0.5 to 2.5 cm. Hornblende and actinolite are present as relict cores surrounded by chlorite, which comprises the albite poor layers. Optical zoning of inclusion-filled albite is common and quite pronounced, with albite overgrowth rims being inclusion absent. These greenstones are found within albitic schist, devoid of aluminous schist, 100 to 200m below the contact with the overlying Battell Formation (plate 1).

Stratigraphic Summary:

The Monastery Formation includes a lithological succession composed primarily of the albite schist, becoming more quartzose near quartzite beds, which in turn, appear to be most abundant in the aluminous schist part of the sequence. The increase in quartz, quartzite, and aluminous schist progresses toward the contact with the overlying Battell Formation. Dolarenite is discontinuous and predominantly occurs within the albite schist near interlayers of aluminous schist and quartzite (figure 2.2). Greenstone is found in only one location, but occurs exclusively within the albitic schist, 100 to 200m below the aluminous schist and the contact with the Battell.

Contact Relations:

The contact with the overlying Battell Formation is characterized by the occurrence of the first distinctive graphite-rich schist layers, common to the Battell. Graphite occurs within quartz-chlorite-albite schist, interlayered with non graphitic schist, similar to the albitic schist of the Monastery. Graphite, however, is found in the majority of the Battell Formation, and although this interlayering appears to be a gradation between the two distinct formations, the contact is defined at the lowermost graphite schist horizon. The interlayered zone, commonly 10 to 100m thick, contains tourmaline-rich layers and graphitic marble, entirely absent in the rest of either formation. The graphitic / non graphitic layers occur as continuous 1 to 10cm thick horizons which, for the most part, parallel the dominant fabric. Graphite is progressively more pervasive away from the typical Monastery lithologies and into typical Battell homogenous graphitic schist. The along strike consistency of graphitic / non graphitic schist interlayering, the lack of any large-scale lithologic truncations, and the presence of graphitic marble and tourmaline-rich schist, not found in any of the other pre-Silurian lithologies, seem to indicate that this interlayered zone, and hence the Monastery / Battell contact, are most likely depositional and not the result of severe transposition of preexisting fault-related fabric.

Type Locality:

The Monastery Formation, as defined in this study, is best displayed in a 125m section within the Hancock Branch of the White River, starting immediately south of the fourth bridge, located at an elevation of 475 m, and continuing downstream. This section is accessible along a forestry road branching from the Texas Falls road approximately 1 mile north of Texas Falls (plate 1).

Depositional environment:

Although lacking original sedimentary features, several characteristics of this formation can be utilized for interpreting sedimentary protoliths and depositional environments:

1. The high modal percentage of quartz and feldspar within the schist units, coupled with the presence of similar lithologies in the Hoosac Formation, may indicate rapid deposition. The progressive decrease in grain size, detrital content, and feldspar modal abundance from the Hoosac, eastward into the Monastery, is probably indicative of deposition in relatively close proximity to the source region(s). This deposition site would be more distal to source, relative to Pinnacle and Hoosac lithologies, but more proximal relative to the finer grained, more aluminous rocks farther east.

2. Close association with Hoosac lithologies, interpreted as proximal deltaic and alluvial fan, rift clastic deposits (Tauvers, 1982), provides supportive evidence for a similar depositional environment. Greenstones within the Hoosac/Pinnacle and Pinney Hollow Formations are geochemically similar to modern day rift basalts, and provide further evidence for deposition of associated sediments within a rifting environment (Coish et al., 1985; Stanley et al., 1989).

3. The absence of abundant coarse wacke in the Monastery lithologies, where aluminous schist and albitic schist dominate, suggests finer grain sedimentation in a more distal part of the Pinnacle/Hoosac deltaic fan complex.

4. Dolarenite layers of the Monastery are similar to the western cover sequence Forestdale and White Brook dolostone of central Vermont and southern Québec, respectively. These units have been interpreted as rift clastic deposits formed within relatively shallow water, possibly within a subtidal to supratidal environment (Tauvers, 1982; Dowling, 1987).

Detrital quartz, feldspar, and oxides are present within all three dolostone units, and indicate concomitant clastic deposition.

5. Quartzite is found most commonly within the upper part of the Monastery as 5 cm to 1m beds. The lack of significant feldspar, carbonate, or oxides, in conjunction with the vitreous nature of the quartzite, indicates deposition of mature sands, possibly during transgression of the basin onto the craton, and/or peneplenation of the source area. The association of quartzite with fine grained aluminous schist, most likely metamorphosed fine grain silt and clays, provides further evidence for deposition far from the source region.

Battell Formation (Cb)

Previously referred to informally as the Battell Member of the Monastery Formation (Osberg, 1952, p. 42-43), graphitic schist, tourmaline-bearing schist, and carbonate structurally overlying the Monastery Formation are redefined as the Battell Formation. This formation presently includes the aforementioned lithologies previously mapped within both the Monastery and Granville Formations (Osberg, 1952).

The name Battell originated from exposures of graphitic schist with interlayered dolomite horizons on the southeastern flank of Battell Mountain (Osberg, 1952; p.46-48; figure 2.1). Detailed mapping has demonstrated that the Battell can be subdivided into three distinct members; interlayered graphitic/nongraphitic, tourmaline-bearing schist, presently defined as the White River Member (Cbw) which immediately overlies the Monastery Formation (figures 2.2 & 2.5), graphitic schist (Cb), and massive to bedded dolomite and marble (Cbd), interbedded with graphitic schist (Cb).

Description:

I. White River Member (Çbw):

The White River Member of the Battell Formation includes nongraphitic, rusty weathering, albite-quartz-muscovite-chlorite-tourmaline-pyrite schist and "Battell-like", graphitic, muscovite-chlorite-quartz-albite-magnetite schist (table 2). These two schists are interlayered on a scale ranging from 1 cm to 1 m. These layers, although transposed into the dominant foliation, can be traced along strike for significant distances (meters) without apparent truncation (figure 2.5). On a larger scale, the White River Member can be traced continuously around large fold structures, demonstrating its along strike continuity. This continuity, however, is obscured by at least one large thrust surface which cuts down into

Figure 2.5 Interlayered tourmaline-bearing, graphitic albitic schistose quartzite and dark graphitic marble of the White River Member of the Battell Formation. Interlayering is parallel to Sn foliation but is also relatively continuous and is interpreted as transposed bedding. Scale is 5 cm. (Picture from the Hancock Branch, White River, map coordinates T 11).

	HQ 467 Cbw	HQ 490 Cbw	HQ 509 Cb	HQ 513 Cb	HQ 528 Cb	HQ 575 Cbd	HQ 588a Cbd
Porphyroblasts							
albite	10	tr	4	6	6	-	-
Groundmass							
quartz	25	25	24	20	22	3	2
albite	14	20	15	10	12	1	-
sericite	28	38	29	36	33	4	2
biotite	tr	-	-	-	-	tr	-
chlorite	13	10	15	20	16	3	-
chloritoid	-	-	-	1	-	-	-
apatite	-tr	tr	-	tr	tr	-	-
calcite/dol.	2	1	7	2	4	88	95
pyrite	3	2	3	tr	2	tr	1
tourmaline	3	-	tr	2	2	-	-
graphite	tr	4	3	2	3	-	-
magnetite	-	-	-	-	-	1	tr
ilmenite	2	tr	-	1	-	-	-
Grain Size							
Porphyroblasts	1.0-2.0mm	< 1.0mm	0.5-2.5mm	0.5-1.0mm	0.5-2.0mm	-	
Groundmass	0.2-0.5mm	0.1-0.5mm	0.1-0.4mm	0.1-0.4mm	0.1-2.0mm	0.2-1.5mm	0.1-2.0mm
Texture	schistose	phyllitic	schistose	schistose	schistose	granulose	granulose
Location	D 17	B 17	V 14	T 11	U 14	K 8	H 7

Table 2 Modal analyses of Battell Formation samples (including the White River Member and dolomite unit).

this unit within several interpreted, pre-thrust antiforms (plate 1, I11).

Although compositionally similar to the Monastery Formation, the nongraphitic schist contains significantly less quartz and albite, and contains abundant tourmaline and sulfides, uncommon to most of the Monastery Formation. The graphitic schist horizons increase in abundance away from the nongraphitic albitic schist, aluminous schist , and quartzite in the upper section of the Monastery Formation (figure 2.2).

Discontinuous (1 cm to 1m) lenses of dark gray to black dolomitic marble with minor quartz, feldspar (albite), and muscovite are found in the more graphitic layers throughout the interlayered sequence (figure 2.2). These marbles do not appear to represent a consistent stratigraphic marker, but rather discontinuous horizons within the White River Member. Graphitic marbles are not present in any of the other formations within the study area, including the other graphitic schist units found comprising the Granville and Ottauquechee Formations, farther to the east. The marbles therefore appear to be a diagnostic lithology within the White River Member.

The White River Member ranges in thickness from 10 to 100m. It forms the base of the Battell Formation and is overlain by the dominant graphitic schist member (figure 2.2).

II. Graphitic schist member (Cb):

The graphitic schist member is composed of quartz-chlorite-muscovite schist with albite porphyroblasts, 1 to 4mm in diameter (table 2). Graphite, comprising 1-3 modal percent, is predominantly found as inclusions in the muscovite and albite, and as discreet seams along foliation planes.

The graphitic schist locally contains blocks or discontinuous lenses of dolomitic marble, with dolomite-calcite-epidote-quartz-muscovite (table 2; figure 2.6). These marbles are usually present 1 to 10m above the White River Member. Due to their discontinuous presence, they have not been included as discreet members.

III. Dolomite Member (Cbd):

At the type locality for the Battell Formation, carbonate is present as a 100m thick, bedded dolomite-marble sequence (plate 1, L7). This is the only mappable carbonate within the Battell Formation, and has been separated as a distinct member. Contacts with the surrounding Battell graphitic albitic schist are usually sharp with a few gradations between thin dolomite and schistose layers. Interbeds (mm to cm scale) of graphitic schist are common throughout this sequence (figure 2.2).

Although discontinuous, this dolomite, along with the smaller ones, seem to mark a consistent stratigraphic horizon approximately 1 to 10m above the White River Member and

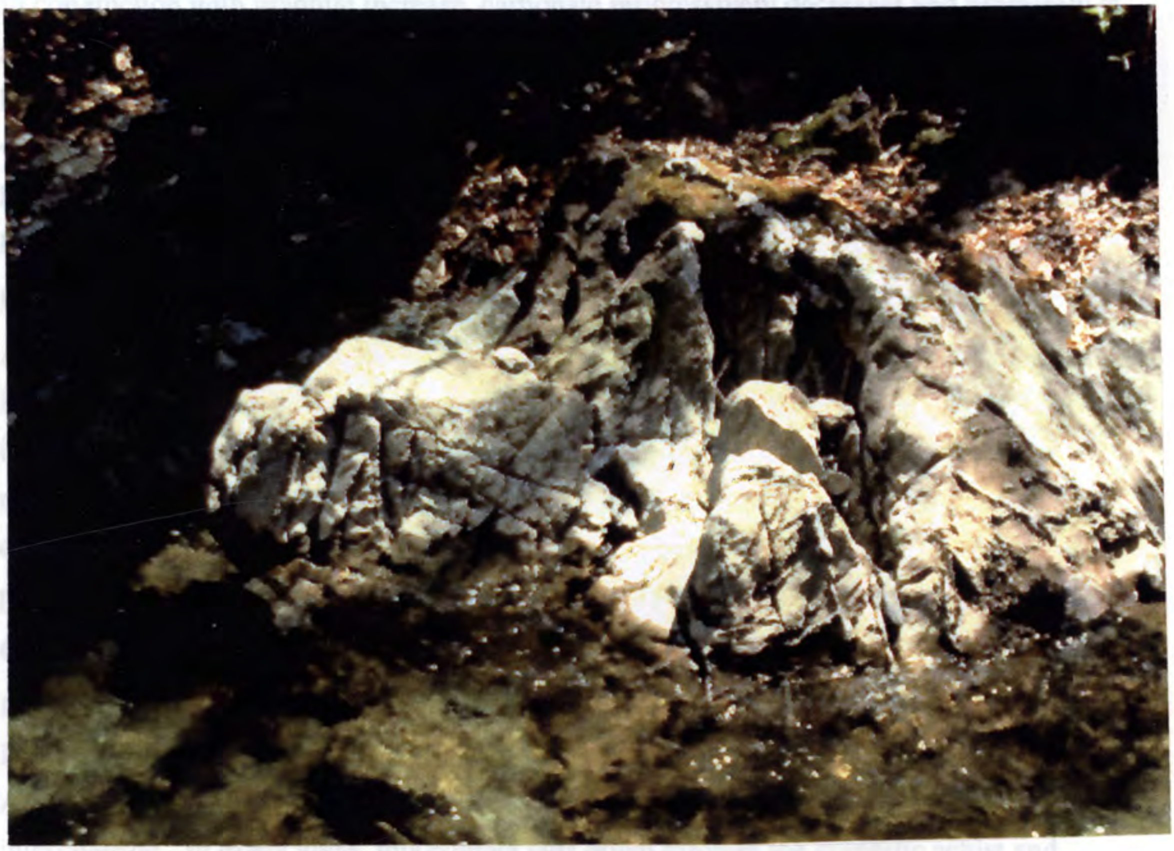


Figure 2.6 Blocks of dolomitic marble surrounded by non carbonate-bearing graphitic schist of the Battell Fm. Blocks appear to be slump deposits of previously coherent carbonate, possibly shed from the nearby platform (picture from Clark brook, 0.75 miles upstream from Forestry Route 55; plate 1, C21).

are interpreted as stratigraphically overlying the White River Member. Above the dolomite is a thick section of carbonaceous schist (Cb) with locally pervasive dolomitic laminations 1-2 cm thick (table 2; figure 2.7). This section appears to become more graphitic away from the dolomite, with pervasive seams of graphite occurring along foliation planes, and muscovite inclusions increasing from 1-2 percent, upwards to a maximum of 3-4 percent. In conjunction with graphite increase, carbonate progressively decreases and ultimately is absent. The carbonate absent section is also marked by an increase in quartz in particular horizons, usually 1 to 10cm thick. These quartz-rich horizons have a granulose texture and would be best described as graphitic, quartz granulites. Although not as quartzose, they have a similar appearance to the dark quartzites within the Granville and Ottauquechee Formations.

Lithologic Summary:

A presumed stratigraphic section for the Battell Formation consists of the lowermost interlayered nongraphitic and graphitic schist of the White River Member, grading upwards into a thin section of carbonaceous schist and interbedded and locally discontinuous massive dolomite and marble. The dolomite / marble is overlain by a dolomite-laminated carbonaceous schist section, grading upwards into non-carbonate-bearing, more graphitic, pyritiferous phyllite/schist with interlayers of graphitic, quartzose granulite (figure 2.2).

Contact Relations:

In contrast to the lower contact with the Monastery Formation, the upper contact of the Battell Formation, with aluminous schist, chlorite schist, and greenstone of the Mt. Abraham and Pinney Hollow Formations is quite sharp, without distinctive lithologic interlayering. In many areas, this contact cuts down through the graphitic schist and dolomite of the Battell Formation, and is characterized as a sharp break between either the White River Member or the albitic schist of the Monastery Formation (plate 1, K14). Due to the lithologic truncation of Battell lithologies, and the rare truncation of greenstone within overlying Mt. Abraham and Pinney Hollow Formations, this contact is demonstrably a fault surface. To the east this tectonic surface cuts down through the entire Battell Formation and juxtaposes Pinney Hollow on top of Monastery Formation albitic schist, greenstone, and dolarenite (plate 1, J27). The truncation of the Monastery greenstone and dolarenite, and the truncation of a thick Pinney Hollow greenstone along this surface, further demonstrates its fault character. This surface has been named the Child's Mountain Thrust for the Monastery / Pinney Hollow truncations mapped on the east flank of Child's Mountain (plate 1, O20).

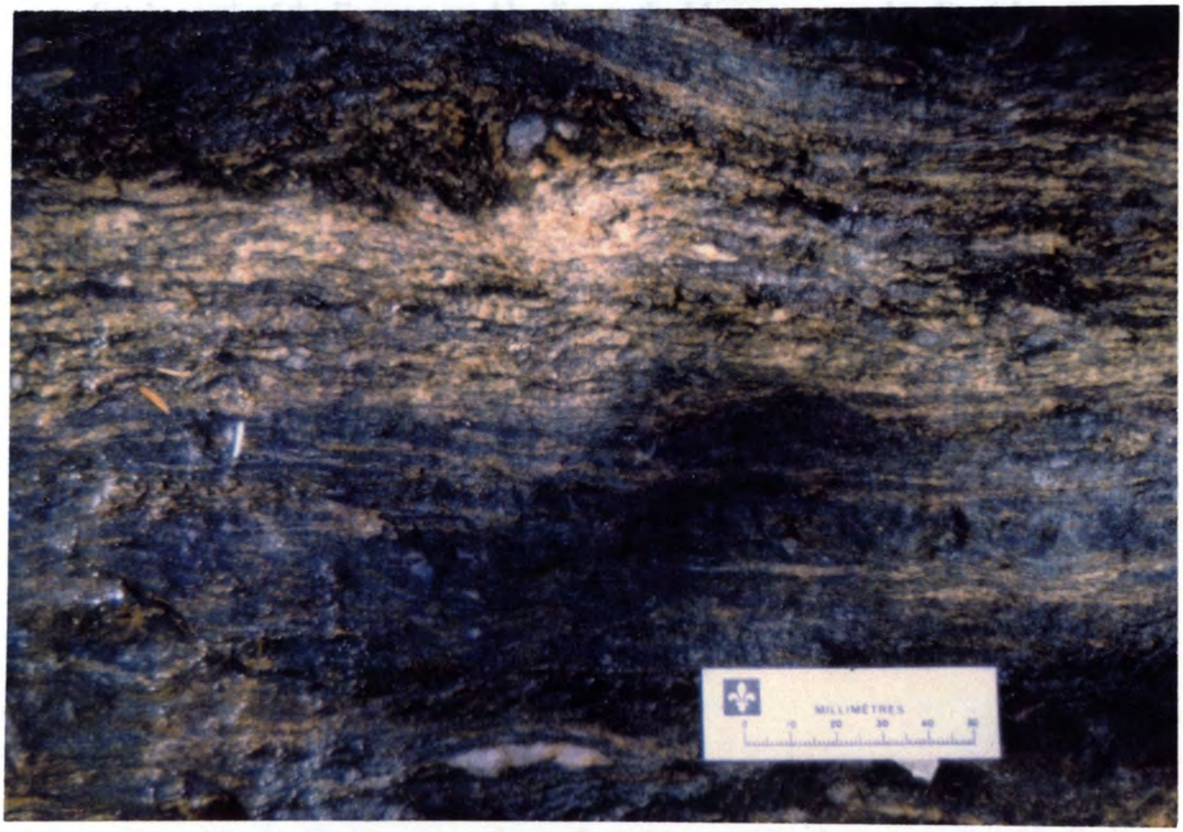


Figure 2.7 Dolomitic marble laminations within graphitic schist of the Battell Fm., at the confluence of the Hancock Branch and Texas brook (plate 1, W 14). Laminations are parallel to Sn foliation and are probably transposed turbidite laminae. Scale is five cm.

Type Locality:

Due to its inaccessible location, the type locality section for the Battell Formation should be appended to include a more accessible reference locality. A representative section is present within the Hancock Branch of the White River at the junction with Texas Brook, immediately south of the Forestry road leading to the Monastery type locality (plate 1, V14).

Because the present type locality is the only one containing thick dolomite, it is suggested that it remain the formal type section (plate 1, L7).

Depositional environments:

1. The consistent lithological succession of lowermost White River Member, into a thin sequence of graphitic schist, a massive section of marble, or blocks of carbonate within carbonate and noncarbonate-bearing graphitic schist suggests that the entire succession is a coherent sedimentologic package.

2. The interlayered schists of the White River Member appear to represent a sedimentological transition from nongraphitic Monastery albitic schist, aluminous schist, and quartzite, into homogenous graphitic schist and carbonate of the upper parts of the Battell Formation. Tourmaline, ubiquitously abundant in this unit, is generally envisioned as a consequence of borate deposition in a playa or sabkha environment (Reineck and Singh, 1975; Walker et al., 1978). The lack of abundant carbonate, and particularly gypsum mineralization (or metamorphic equivalents), seems to argue against this type of environment. The fine grained nature of these rocks, the presence of discontinuous carbonate, combined with the apparently conformable transition from an earlier rift clastic deltaic environment, does argue for a relatively passive, shallow water realm.

3. The dolomite unit, with interlayered marble, is extremely homogenous, unlike the heterogeneous dolarenite of the Monastery Formation. This homogeneity suggests quiescent depositional conditions, similar in type to those conditions for carbonate deposition within a siliciclastic platform (Milliman, 1974). Similar dolomites are found within the Cambrian carbonate platform of western Vermont, including the Lower Cambrian Dunham Dolomite and the Middle Cambrian Winooski Dolomite (Cady, 1945; Osberg, 1952). The massive to bedded dolomite sequence (Cbd), in conjunction with the discontinuous dolomite/marble blocks within the Battell graphitic schist east of the large carbonate member, best fit a slope rise / shelf edge environment, immediately outboard of the major part of the carbonate platform. The carbonate member would represent the distal part of the shelf edge, or possibly a slump/break off from the edge. The dolomite blocks,

also very homogenous and lacking interlayers of graphitic schist, could possibly be talus blocks or slumps broken off of the main carbonate body, and shed into the slope rise environment. These features are common within the post rift parts of the Afar region and Gulf of Aden (Illing et al., 1965). Definitive evidence, however, for correlation with the platform carbonates, such as fossil control, is lacking.

4. Graphitic schist above the dolomite member, contains pervasive 1 cm - scale continuous laminations of carbonate. Although highly transposed into the dominant foliation, these laminations may have been original sedimentary layers, quite similar to planar laminations in proximal turbidites or gravity flows, found within a slope rise environment (Bouma, 1962; Bouma and Brouwes, 1964; Walker, 1965; Lowe, 1982). This section could represent slope-rise deposition following basin deepening and drowning of the carbonate shelf edge. The lack of rip-up clasts, dismembered carbonate-bearing horizons, or cyclic metamorphosed siltstone and/or sandstones appears to rule out deposition within a tidal environment (Walker, 1978). Increase in graphite content upwards with a decrease in the amount of carbonate further suggests basin subsidence and deepening of the sediment/water interface below the calcium compensation depth.

Mount Abraham Formation (CZa)

This unit was first introduced as the Mount Abraham Schist Member of the Underhill Formation in the 1961 Centennial Geologic Map of Vermont (Doll et al., 1961; figure 2.1). Specific lithologies included:

> " Light gray sericite (muscovite-paragonite) quartz-chloritoid rock with silvery sheen; porphyroblasts of magnetite are common and porphyroblasts of chlorite, chloritoid, garnet, and kyanite occur locally"

The type locality for this member is on the summit of Mount Abraham, immediately north of Lincoln Gap, in the town of Warren, Vermont. It was defined by Cady (1960) as the aluminous schist unit occurring between albitic schists of the underlying Underhill Formation and the overlying graphitic schists of the Hazen's Notch Formation (see Doll et al., 1961). Due to the lack of similar lithologies immediately north of the type locality, the Mount Abraham was interpreted as representing a rapid and complex facies change with Underhill albite schist, and Monastery schist (later redefined as Underhill; Doll et al., 1961) to the south.

Osberg (1952; e.g., modal analyses) included this aluminous schist member as a lithology within the Monastery Formation. Since it was considered discontinuous along and across strike,and not large enough within the Granville - Hancock area to be displayed at 1:62,500-scale, it was not separated as a distinct formation. Doll and others (1961), Cady and others (1962) and Albee (1957) defined a similar unit to the north called the Foot Brook Member of the Underhill Formation:

> "Sericite (muscovite-paragonite) quartz-chloritoid schist with minor carbonaceous interbeds".

Both Cady and Albee interpreted this sequence as a western correlative of the Stowe Formation, since a very similar lithology was found farther to the east and interpreted as part of the youngest formation within the pre-Missisquoi eastern cover sequence (Cady et al., 1962; Albee, 1972). A similar paragonite-chloritoid-bearing schist makes up the majority of the Pinney Hollow Formation at its type locality (Perry, 1928; Brace, 1953). The Mount Abraham schist was given "Formation status" (delineated from the Underhill Formation) by Cady and others (1962) for a continuous exposure of quartz-white mica-chlorite-chloritoid-garnet phyllite and schist along the crest of Mount Abraham in the Lincoln Mountain 7.5 minute quadrangle.

Description:

Within the Granville-Hancock area, the Mt. Abraham Formation contains three different lithologies, the first two (pelites, described below) of which are inseparable at the 1:12,000 map scale, due to their discontinuous and intercalated distribution. The third lithology, a single greenstone body (CZag) has been mapped as a separate unit. These lithologies include (from most abundant to least):

Schist Member (CZa):

The most abundant unit is white to light gray, tan weathering, quartz-white mica-chlorite-chloritoid-ilmenite-garnet-epidote-magnetite-albite schist/phyllite (table 3). Ilmenite grains are quite abundant, can be seen in hand specimens, and are responsible for the rock's light steel gray color (Albee, 1965; Laird and Bothner, 1986; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987). Chloritoid grains occur as lathes and porphyroblasts usually 0.2-1mm in diameter, with the infrequent larger ones present sporadically. Garnet is fairly abundant throughout this unit. 1). Garnet is almost entirely absent east of this level, occasionally found as relict cores with pervasive chlorite overgrowth. The distribution of garnet within this unit is most likely the result of a progressive decrease in metamorphic grade from west to east, rather than a change in bulk composition (see chapter 4, metamorphism). Albite is almost always present in trace amounts, with very pervasive and complex inclusion trails of chloritoid, quartz, and rarely garnet. Unlike inclusions within garnet, many albites have inclusion trails which display an earlier, relict foliation.

	HQ 493	HQ 497	HQ 500	HQ 500a	HQ 535
Porphyroblasts					
albite	tr	-	1	tr	1
garnet	tr	tr	1	4	-
Groundmass					
quartz	57	35	33	23	48
albite	1	3	3	13	2
sericite	30	39	36	35	40
chlorite	6	9	20	14	6
chloritoid	2	2	4	5	1
calcite	-	5	-	-	-
graphite	-	3	tr	-	-
ankerite	1	tr	tr	2	1
apatite	tr	-	-	tr	tr
alanite	tr	-	tr	tr	-
pyrite	tr	1	-	1	-
tourmaline	-	tr	tr	tr	tr
zircon	-	-	-	tr	-
magnetite	1	tr	tr	1	tr
hematite	tr	-	-	-	tr
ilmenite	2	tr	2	2	1
Grain Size					
Porphyroblasts	0.5-1.0mm	0.5-1.5mm	0.5-1.0mm	1.0-3.0mm	0.4-1.0mm
Groundmass	0.2-0.5mm	0.2-0.5mm	0.2-0.5mm	0.2-1.0mm	0.2-0.5mm
Texture	schistose	phyllitic	schistose	schistose	schistose
Location	C 14	B 15	E 10	E 8	X 19

Table 3 Modal analyses of Mt. Abraham Formation samples.

The second most abundant lithology in the Mount Abraham Schist is a light gray to white, tan weathering quartz-chlorite-white mica-albite-magnetite-ilmenite schist, containing numerous disseminated quartz veins and large (1 to 4 cm) magnetite porphyroblasts. This lithology occurs as 1 to 5m thick layers within both the garnet-bearing and non-garnet-bearing parts of the previously mentioned lithology, although garnet is never found within this specific lithology. It is commonly found as 1 to 2 m wide zones, which are not reliably traceable along strike due to a lack of abundant outcrop. Because it is found both east and west of the proposed garnet isograd, the lack of garnet is most likely a reflection of bulk composition (chapter 4). The ubiquitous magnetite porphyroblasts, relatively abundant albite, and lack of chloritoid make this unit easily delineable from the more abundant chloritoid-garnet lithology. This unit is very similar to much of the quartz-chlorite-white mica-albite-magnetite phyllite and schist of the Pinney Hollow Formation farther east, which is in gradational contact with the easternmost Mt. Abraham Schist (plate 1, P20).

Greenstone Member (CZag):

A light to apple green chlorite-epidote-albite-hornblende-actinolite-quartz gneiss, or greenstone is found along Forestry Rt. 55, 3.4 km west of Granville (plate 1). It is surrounded by the magnetite-bearing lithology and the white mica-chloritoid phyllite, similar to the dominant Mt. Abraham lithology. This entire sequence is also intercalated with silvery, albitic, white mica-chlorite-quartz schist, identical to the major lithology within the Pinney Hollow Formation. Similar greenstones are found farther to the east, surrounded entirely by the Pinney Hollow silver schist. This greenstone is mapped within a large north plunging fold hinge, with limbs apparently truncating along the tectonic contact with the underlying Battell Formation (the Child's Mtn. Thrust; plate 1, C21). The greenstone appears to be in gradational contact with the surrounding schist units, due to the high amount of mafic constituents within the schist and the high white mica content of the greenstone within 1 m of the actual contact.

Contact Relations:

Although no units were mapped overlying the Mt. Abraham Formation in this area, graphitic albitic schist of the Granville Formation (Lincoln Gap Member, Cgl) is found tectonically overlying the Mt. Abraham Formation type section on Mount Abraham, north of Lincoln Gap (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987). This contact was interpreted as a pre-peak metamorphic thrust fault; the Mt. Abraham, with its tectonic lower contact (above Battell and Monastery Formations), represents a fault-bound lithotectonic package, locally referred to as the Mt. Abraham Slice (Stanley et al., 1988b). In the Fayston area, Walsh (1989) has suggested that graphitic schist of the Granville Formation may overly parts of the Mt.Abraham Formation along a pre-metamorphic thrust surface, due to the sharpness of the contact. Due to the sparseness of outcrop, further evidence for this fault in the Fayston area is not available (G.J. Walsh, pers. comm., 1990). A gradational layering of Granville and Mt. Abraham lithologies in some parts of this area have also been mapped, without any truncation of lithologies. This suggests that a depositional contact for at least parts of the Mt. Abraham / Granville sequence cannot be ruled out (Walsh, 1989).

The Mt. Abraham grades laterally eastward into the Pinney Hollow Formation chlorite-rich schist, containing abundant greenstone. The actual contact between the two formations is drawn along the contact between homogenous Mt. Abraham phyllite and the appearance of the first **continuous**, thick layer (> 1 m) of Pinney Hollow chlorite schist. The actual stratigraphic position of the Mt. Abraham Schist, with respect to the structurally higher Pinney Hollow Formation, is unclear since no diagnostic sedimentary topping criteria were either preserved or observed during mapping. The lack of significant greenstone within most of the Mt. Abraham, combined with its high alumina bulk composition, relative to the mafic-rich Pinney Hollow, suggests that it may form the stratigraphic upper section of the Pinney Hollow lithotectonic unit, hence, deposited subsequent to the main period of volçanism. Smaller bodies of Mt. Abraham-like chloritoid phyllite were found farther east within the main belt of Pinney Hollow schist (plate 1; G34, R35), and have been reported as thin horizons within the Stowe Formation (Albee, 1957). The lack of significant amounts of Pinney Hollow-like lithologies to the west may be the result of tectonic truncation, with the Child's Mountain Thrust ramping westward into the stratigraphically higher Mt. Abraham Schist.

Reference Section:

Due to the lack of any accessible sections in the Granville-Hancock area, the various lithologies of the Mt. Abraham Formation cannot be viewed together at any one particular locality. A representative section of the dominant white mica-chlorite-chloritoid phyllite can be seen within the White River at an elevation of 670 m (2200 ft), immediately south of a logging road which branches from Forestry Route 55 (near a large culvert) approximately 3.7 km (2.3 miles) to the east (plate 1). Since a major part of the Pinney Hollow type section, in Bridgewater, Vermont, contains paragonite-chloritoid-garnet schist indistinguishable from the central Vermont Mt. Abraham Schist, the Mount Abraham Schist lithologies should not receive any formal promotion to formation status until further work determines the along strike correlations of southern and central Vermont Pinney Hollow (and Mount Abraham) units.

Depositional environment:

Modal analyses and bulk compositions of various Mt. Abraham lithologies mapped within central Vermont, by various workers, consistently indicate high alumina content (Osberg, 1952; Cady et al., 1962; Albee, 1965b; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Walsh, 1989). High alumina usually indicates a clay-rich protolith for low and medium grade rock types (Folk, 1968; Frey, 1970; Wybrecht et al., 1985). The high alumina content, combined with a typical lack of feldspathic component or detrital material suggests deposition far from a source area (Reineck and Singh, 1975; Kinsman, 1975). The local abundance of ankerite carbonate (at least in the Granville-Hancock area) may be the result of metamorphism and Fe replacement of Mg from primary or secondary dolomite (Folk and Land, 1975). Carbonate layers indicate that deposition of sediments occurred in relatively shallow, warm, quiescent waters above the calcium compensation depth. The lack of pervasive greenstone, such as those within the Pinney Hollow, may reflect a period of relatively little volcanic activity, possibly marking the cessation of assumed, rift-related processes. If this hypothesis is correct, the gradational contact between the Mt. Abraham and the easterly adjacent Pinney Hollow Formation may represent a stratigraphic horizon. The Mt. Abraham lithologies would comprise the upper part of a depositional sequence with stratigraphically lower Pinney Hollow units.

Alternatively, the Mt. Abraham Schist may represent a rift-stage sediment deposited in far from the source region(s), or even a post-rift sequence deposited within a sediment-starved basin. Its lithologic similarity to the Taconic rift-clastic, aluminous shales with interlayered basalt, makes the first hypothesis the most attractive one.

The Pinney Hollow Formation (CZph)

This unit was originally identified at its type locality, Pinney Hollow, in the town of Plymouth by Perry (1928, p. 24):

> "During the preliminary examination of the rocks of Plymouth, the formation herein named and described as the Pinney Hollow Schist was first examined at several excellent outcrops a few yards east of West Bridgewater village. Subsequently it was found that the formation was much more typically exposed in Pinney Hollow from the Pinney Hollow School westward nearly to Plymouth Village, and it was consequently named from that locality".

The Pinney Hollow Formation was first described by Perry (1928, p. 25) as:

> "Megascopically, the typical Pinney Hollow Schist is pale green and thinly laminated, often with minute crenulations, in the general plane of schistosity, passing into lines of slip cleavage. Distinct layers of quartz and the micaceous minerals, chlorite and sericite, visible and magnetite grains are often of sufficient size to be recognized."

Brace (1953) and Skehan (1961) later recognized interlayered chlorite-epidote-albite-carbonate-amphibole metavolcanic greenstone and hornblende amphibolite (Chester Amphibolite Member) were included within this formation. In the Granville-Hancock area, Osberg (1952; p. 55-61) defined the Pinney Hollow as a chlorite-muscovite-quartz-albite schist with interlayered albite-epidote-calcite-chlorite greenstone (Hancock Member), and small, thin, discontinuous lenses of graphitic albitic schist interpreted as sedimentary lenses. Osberg also noted the presence of aluminous-chloritoid-white mica-chlorite phyllite as discontinuous lenses similar to that found within the Monastery Formation. Osberg also mentioned the similarity of this unit to the chlorite-muscovite-quartz-albite schist and associated greenstone (Brackett Member) of the Stowe Formation, although separated by the intervening graphitic schists of the Ottauquechee Formation (figure 2.1).

Lithologic Description:

Schist member (CZph):

Within the western part of the field area, the Pinney Hollow schist is fine grained, silvery green chlorite-muscovite-albite-quartz-magnetite-pyrite-garnet schist/phyllite. This unit is also very homogenous within the central part of the study area (Domain 2) where it is associated with the greenstone member (CZphg, below). Albite occurs as small (1 to 3mm diameter) porphyroblasts with numerous inclusion trails of fine grained oxides, quartz and chlorite. Quartz is found as interstitial grains and as abundant disseminated veins, disarticulated within the dominant foliation. Garnet is found within one particular locality as numerous, small euhedral to subhedral grains, overgrowing a pervasive fault zone foliation(Table 4a; HQ 256 B). Within this garnetiferous zone, quartz comprises the majority of the rock, a phenomena common to most synmetamorphic fault zones within the Pinney Hollow Formation. Garnets within this zone do not have the pink color, characteristic of almandine-rich garnets abundant throughout the western belt of Monastery and Mt. Abraham lithologies (see Osberg, 1952, and Albee, 1965b). These fault zone garnets may very well be of a different composition, and in equilibrium below garnet grade.

Farther east, some chlorite schist and phyllite is complexly layered on a centimeter to meter scale with light gray, tan, and dark gray weathering, albite-muscovite-chlorite-quartz-magnetite-schist with albite-rich layers creating a compositional layering parallel to the dominant foliation (table 4a). Albite porphyroblasts are more abundant and larger (3 to 6 mm) than those typically found in the majority of the Pinney Hollow chlorite schist\phyllite farther west. The two different schist/phyllite lithologies are also complexly interlayered with each other across a 2km zone within the central part of the field area, centered along the White River Valley (plate 1). Since both lithologies do occur in some amounts across the entire breadth of this formation, they have been grouped together as a single map unit (CZph). The albite-rich unit most common in the eastern part of the study area also contains discontinuous layers of white mica and chloritoid-bearing phyllite, similar to the Mt. Abraham Formation. Greenstone is also present within the albite schist as infrequent, thin layers. Pervasive albite-rich layers, intercalated with chlorite-rich layers, are somewhat similar to the albitic schist of the Fayston Formation (Walsh, 1989). It is The albite-rich schist is less aluminous than the chlorite-rich schist and phyllite, and would help in explaining the abundance of feldspar over white mica and therefore its coarser grain size (see chapter 4, metamorphism).

Associated with the albitic schist sequence are thin 1-2 meter white quartzites and 0.5-1 meter green quartzites, discontinuous along strike and presumably occurring at various stratigraphic horizons. These quartzites have been dynamically recrystallized and do not possess any observable relict detrital grains. In addition to quartz, they locally contain trace amounts zircon, magnetite, ilmenite, feldspar, and minor amounts of muscovite and chlorite (table 4a). Similar quartzite horizons are locally found within the Waitsfield-Warren area to the north (Stanley et al., 1986; Walsh and Stanley, 1987;

	HQ 71	HQ 96	HQ 124	HQ 133	HQ 187	HQ 256-6x
Porphyroblasts						
albite	3	-	2	3	6	3
garnet	-	-	-	-	-	2
Groundmass						
quartz	65	50	58	45	44	50
albite	5	4	4	4	8	5
sericite	15	39	30	37	15	17
chlorite	11	4	4	9	24	21
chloritoid	-	1	-	-	-	-
graphite	tr	tr	tr	-	tr	tr
ankerite	-	tr	-	-	tr	-
apatite	tr	-	-	-	-	-
alanite	tr	-	-	-	-	tr
pyrite	-	1	1	-	-	tr
tourmaline	tr	tr	tr	tr	-	-
magnetite	1	tr	1	2	2	1
hematite	tr	tr	tr	-	-	tr
ilmenite	tr	1	tr	tr	-	1
Grain Size						0.1-0.8mm gar
Porphyroblasts	0.5-1.5mm	-	0.5-2.0mm	0.5-2.5mm	0.5-3.5mm	0.2-1.5mm ab
Groundmass	0.1-0.3mm	0.2-0.4mm	0.2-0.6mm	0.2-0.5mm	0.2-1.0mm	0.1-0.3mm
Texture	schistose/ mylonite	schistose	schistose/ mylonite	phyllite	schistose/ gneissic	schistose/ mylonite
Location	E 34	G33 "CZa" on map	D 31	K 35	B 22	L 30 Allbee Brook

Table 4a Modal analyses of Pinney Hollow Formation pelitic samples.

	HQ 3	HQ 65	HQ 132	HQ 254A	HQ 254B	HQ 461	HQ 544
Porphyroblasts							
albite	2	-	-	-	-	-	-
hornblende	4	5	1	3	tr	-	4
Groundmass							
quartz	5	-	2	2	8	2	-
albite	15	14	21	24	8	23	15
sericite	tr	-	-	-	- -	-	-
biotite	-	tr	2	1	-	3	4
chlorite	50	55	34	39	58	34	45
epidote	10	23	36	26	22	30	23
calcite	10	2	3	3	2	6	5
rutile	-	-	-	·	tr	-	-
sphene	2	1	1	2	2	2	3
pyrite	-	-	-	-	tr	-	-
magnetite	2	tr	-	tr	2	-	1
hematite	-	-	-	-	-	tr	-
Grain Size							
Porphyroblasts	0.5-1.5mm	0.65-1.2mm	0.5-2.0mm	0.5-2.5mm	0.5-1.5mm	-	0.5-1.5m
Groundmass	0.1-1.0mm	0.2-0.4mm	0.2-0.6mm	0.2-0.5mm	0.1-0.5mm	0.3-1.5mm	0.1-0.3m
Texture	gneissic	schistose	schistose	gneissic pt. 15a	schistose	gneissic/ schistose	gneissic
Location	F 28 Route 55	D 30	J 34	L 30	L 30 Allbee Brook	T 41	K 35

Table 4b Modal analyses of Pinney Hollow Formation greenstones.

Haydock, 1988) and the Plymouth area (Karabinos, 1987; Perry, 1928). Prewitt (1989) has mapped similar white quartzites within the Pinney Hollow Formation of the Warren area, some ranging in thickness up to 2 to 3m.

Greenstones (CZphg):

Hornblende-actinolite-bearing, chlorite-epidote-calcite-albite-magnetite-quartz ± muscovite ± sphene ± rutile mafic schist / gneiss (greenstone), previously defined as a single stratigraphic unit (Hancock Member), occur primarily as 1-100 meter wide layers within the chlorite-muscovite-phyllite / schist unit (table 4b). These greenstones are either discontinuous where faulted or relatively continuous along strike within the field area (plate 1). Albite porphyroblast-bearing greenstone, less epidote-rich than the majority of greenstones, is found within the albitic schist at one particular locality (plate 1; Q35). This greenstone is similar in appearance and composition to those within the albitic schist of the Fayston Formation, mapped farther to the north in the Warren, Waitsfield, and Fayston areas (Haydock, 1988; Prewitt, 1989; Walsh, 1989). Both greenstone types usually show a mineral segregation of albite porphyroblast-rich horizons with associated laminations of chlorite- and epidote-rich layers parallel to the regionally dominant schistosity (Sn). These segregations may have been relict volcanoclastic bedding and/or epidosite lens-like segregations within the pre-deformation volcanic bodies.

Contact Relations:

Within the Child's Mountain area, the Pinney Hollow Formation is underlain by carbonate-bearing graphitic and nongraphitic schist of the Battell Formation White River Member. In a series of exposures near the Route 55 forestry road, a large greenstone within the Pinney Hollow chlorite schist abruptly terminates along the contact with the White River Member of the Battell Formation (plate 1, K25). In addition, a unique dark green, chlorite-rich albitic greenstone, within Monastery-like albitic schist, and carbonate-bearing schist layers (similar to the Monastery dolarenite unit) also terminate along this contact, documenting its faulted nature; this is demonstrated to be the eastern continuation of the Mt. Abraham/Pinney Hollow Thrust, cutting up section through the Monastery sequence into the overlying White River Member along a thrust fault.

The Pinney Hollow Formation is found intercalated with chlorite-rich phyllite of the Stowe Formation, within tectonic windows beneath the overlying Ottauquechee Formation (plates 1 & 2, S42-AA45). The Pinney Hollow/Stowe contact with structurally overlying Ottauquechee Formation redefines previous stratigraphic relationships and correlations which depicted the Ottauquechee as a stratigraphic intermediate between westerly situated (and presumably older) Pinney Hollow units and easternmost (and youngest) Stowe (Osberg, 1952; Doll et al., 1961; figure 2.1).

Reference Section:

Although all of the Pinney Hollow Formation units cannot be viewed at any one outcrop, a good example of the chlorite schist and greenstone lithologies, comprising the majority of this formation, can be seen along the west side Route 100, 3.7 km (2.3 miles) north of the intersection with Route 125 in the town of Hancock (plate 1; S31). The albitic schist unit is best displayed within a series of outcrops 125 m west of the first house on the west side of a dirt road, .75 km northeast from its branch point with the Howe Brook road (plate 1; Y36).

Depositional environment:

Modal analyses of the Pinney Hollow schist units show a significant amount of quartz and feldspar (albite), with a very high percentage of chlorite and white mica. A protolith for these units would presumably range from a fine grained siltstone (albite-rich schist) to a clay-rich siltstone or mud (chlorite phyllite/schist; Frey, 1970; Reineck and Singh, 1975). Albite porphyroblasts, common to many of the pre-Silurian units, are entirely of metamorphic origin and do not reflect inherent coarseness of the sedimentary protolith (Frey, 1970; Helmold and van de Kamp, 1984), although the <u>amount</u> of albite may indeed reflect original abundance of feldspar within the protolith. Detrital feldspar present within the protolith is commonly transformed during metamorphism into secondary feldspar of an equilibrium composition (Frey, 1970; Smith, 1983). Secondary composition is dependent primarily upon pressure, temperature, and fluid fugacity and composition (Ribbe, 1983; Smith, 1983; Yund and Tullis, 1983). Pressure and temperature conditions under greenschist facies metamorphic conditions favor albite compositions, due to the peristerite gap between An_5 and An_{16} (Smith, 1983;). This gap occurs only under moderately high pressure (> 5 to 6 Kb) or very high temperature (625 C; Amphibolite facies).

Albitization of calcic plagioclase occurs through the reaction:

$$\mathbf{Na + SiO_2 + CaAl_2Si_2O_8 = NaAlSi_3O_8 + Al + Ca + O_2}$$

Where Al is usually transferred internally within the feldspar structure, O_2 bonds with hydrogen to form water, and Ca is expelled from the feldspar, usually to be incorporated into an epidote group mineral, carbonate, or garnet (Frey, 1970). Na is brought into the feldspar as a free monovalent cation, which may have been produced through a breakdown of one or more sodic clays (ex: nontronite, montmorillonite, hectorite), either during diagenesis or epizonal to chlorite grade metamorphism (Helmold and van de Kamp, 1984). Through this process, original calcic plagioclase would simply be altered to albite through a calcium - sodium exchange with concomitant aluminum - silicon exchange, necessary for charge balance. Relict polysynthetic twins, common to triclinic calcic plagioclase, are actually preserved (optically) during and after exchange (Papike and Cameron, 1976). This explains the common observation of polysynthetic twin or lamellae within monoclinic albite porphyroblasts.

The significant amount of mafic material in both units (chlorite and epidote) further suggests intercalation of pelitic and volcanically derived sediments. The gradation from the pelitic units into greenstones substantiates the idea of greenstone deposition as volcanogenic sediments (water lain / ash flow tuffs). The lack of any type of primary, igneous texture argues against volcanic emplacement as lava flows or igneous intrusives. Several greenstones have been geochemically analyzed during previous studies and are similar in composition to modern day rift basalts, erupting through moderately distended continental crust (Coish et al., 1985; Coish, 1987).

Quartz is fairly abundant in both pelitic lithologies, especially in fault zone rocks. Quartzite is common in many parts of the formation, and it is quite possible that fault zone rock types are actually dynamically recrystallized mature or immature quartzite layers. These quartzose (and quartzite) layers range in thickness from 1 to 2 m and probably represent relict bedding. These layers may have formed as quartz-rich turbidite sequences, intercalated with the feldspathic siltstones and clays.

Mt. Abraham Schist-like aluminous phyllite is primarily in gradational contact with the chlorite phyllite of the Pinney Hollow Formation. This chlorite phyllite is usually associated with the larger greenstone bodies, and may represent volcanic contaminated fine grain sediments (clays). The phyllite has significantly less quartz and feldspar (albite) than the albitic schist. Although this might suggest that the coarser grain siltstone (albitic schist) fines into the chlorite phyllite (associated with the Mt. Abraham aluminous phyllite), several bodies of Mt. Abraham-like phyllite are found enveloped by albitic schist, in the eastern part of the belt (plate 1, H33, S35). The aluminous schist either represent local isolated deposits, totally distinct from the main belt of Mt. Abraham Schist, or these deposits are keel synclines which occur along the contact between the Mt. Abraham and Pinney Hollow Formations. Presently, it is impossible at this time to discern between the two interpretations considering the sparseness of aluminous phyllite exposures.

Granville Formation (Cg)

This formation was defined by Osberg (1952) for exposures of graphitic schist and quartzite found between the main bodies of then interpreted underlying Monastery and overlying Pinney Hollow Formations:

> "The name Granville is proposed for the graphitic quartz-muscovite schist lying above the Monastery Formation. It receives its name for exposures in the town of Granville. Although an unfolded section of the Granville has not been found in the Rochester quadrangle, its lithology is well displayed in outcrops in the White River 0.3 mile N25W of the village of Granville".

In previous studies, this rusty-weathering graphitic schist sequence was defined as part of the Talcose Schist Belt and the Albitic Schist Series, respectively (Hitchcock et al., 1861; Perry, 1928).

The Granville Formation was initially mapped by Osberg as rusty-weathering pyritiferous, patchy-graphitic, quartz-chlorite-muscovite-albite schist with interlayered, thin (1cm-1m), predominantly dark gray and black, and rare green and white quartzite layers. Associated with this graphitic unit were interlayered zones or lens-like bodies of nongraphitic, albite-quartz-muscovite-chlorite schist, similar to the interpreted underlying Monastery Formation and chlorite-muscovite-albite schist identical to the overlying Pinney Hollow Formation. Several thin, discontinuous buff to gray dolomitic marbles were mapped within this formation within the Granville area, and farther south within the town of Rochester.

Doll et al., 1961, renamed those rocks included in the Granville Formation as the carbonaceous schist of the Hazen's Notch Formation (as of Cady, 1960). The Hazen's Notch Formation, whose type locality is located in Hazen's Notch, north of Lowell, in northern Vermont, was defined as rusty weathering graphitic schist (similar to the Granville graphitic schist) **and** non-graphitic, albite-chlorite-muscovite-quartz schist, very similar to the present-day Pinney Hollow Formation.

This study redefines the graphitic-albitic rocks with interlayered dark and white quartzites, **exclusively**, as the Granville Formation. This definition of Granville is inconsistent with Osberg's in that only the graphitic schist sequence is included; non-graphitic rocks, also included within the Granville during subsequent mapping by Osberg, are redefined as vestiges of other lithotectonic units and thrust slices (Armstrong et al., 1988a; Stanley et al., 1988a). Dolomites and associated graphitic schist, mapped as part of the Granville Formation in an attempt to preserve stratigraphic continuity across large scale structures, are redefined as Battell Formation. This is due to their presence beneath the

Child's Mtn Thrust and the tectonic level of Mt. Abraham and Pinney Hollow lithologies. Graphitic rocks of the Granville Formation include all of those tectonically **above** the Pinney Hollow Formation. These rocks, including the type locality, do not contain any significant matrix carbonate or any discreet carbonate layers. The Granville is therefore discerned from the Battell through two distinct criteria:

1. Graphitic schist associated with carbonate layers, bedded carbonate, or carbonate blocks is mapped as Battell. Graphitic schist devoid of carbonate is mapped as Granville.
2. The presence or absence of the Battell White River Member interlayered graphitic and nongraphitic schist. This unit also contains distinctive black, graphitic marbles, not found in either the Battell graphitic / dolomitic unit or the Granville graphitic schist.

It is important to mention that all of the graphitic schist exposures which contain carbonate, and therefore meet criteria 1) as Battell Formation, also occur structurally above White River Member-type rocks, as well as Monastery lithologies. The graphitic rocks mapped as Granville also always obey both criteria. The author realizes that Granville may just be "carbonate-free" Battell, resting above Pinney Hollow instead of Monastery. The distinction between these two graphitic units, however, has led to the observation that the Battell/Monastery sequence is structurally lower than the Granville, and furthermore, lies beneath the Pinney Hollow/Mt. Abraham lithotectonic unit along the Child's Mountain Thrust. If the Granville and Battell Formations were once part of a continuous stratigraphic section, substantial structural telescoping must have taken place along the Child's Mountain Thrust to produce the present structural configuration.

Description:

The Granville Formation consists almost exclusively of a regionally pervasive patchy graphitic (1-2 modal percent) quartz-muscovite-chlorite-pyrite-apatite-ilmenite-magnetite schist with distinctive albite porphyroblasts ranging in diameter from 1-3mm (Table 5). Albite is also found interstitially between quartz and muscovite in the grain matrix. Albite porphyroblasts commonly overgrow an early foliation, preserved as graphite-rich inclusion trails. Quartzite, interlayered with the albitic schist, is discontinuous along strike and ranges in thickness from 1cm to 1m, with an average thickness being 6cm. There is a distinct variation in color, from dark gray or black through light gray, green, and white with the majority being within the light to dark gray range. In addition, many of the darker colored quartzites contain syntectonic white quartz veins, complexly folded, and diagnostic of quartzites within the similar Ottauquechee Formation, to the east (plate 1). Quartzite compositions tend to be homogenous; over 97 percent quartz with minor interstitial albite, muscovite, chlorite, epidote, magnetite, and zircon (table 5). Color is controlled by

	HQ 11	HQ 50	HQ 134	HQ 145	HQ 176
Porphyroblasts					
albite	5	2	-	3	1
garnet	-	-	-	-	-
Groundmass					
quartz	38	40	37	35	30
albite	3	12	8	13	2
sericite	42	27	35	31	56
chlorite	7	13	12	9	4
calcite	tr	-	-	1	1
graphite	2	4	5	5	5
apatite	-	-	-	-	tr
pyrite	3	1	3	1	1
tourmaline	-	tr	-	tr	tr
magnetite	tr	1	tr	2	tr
hematite	tr	tr	tr	tr	tr
ilmenite	-	-	tr	-	-
Grain Size					
Porphyroblasts	1.0-2.0mm	1.0-2.5mm	1.0-3.0mm	0.5-1.5mm	1.0-4.0mm
Groundmass	0.2-0.5mm	0.2-0.6mm	0.2-0.6mm	0.2-0.5mm	0.2-1.5mm
Texture	schistose/ phyllite	schistose	schistose	phyllite	phyllitic/ schistose
Location	E 29	D 29	K 34	G 34	C 26 Cgtm White River

Table 5 Modal analyses of Granville Formation samples.

variation in the percentage of graphite (gray vs light green / white) and epidote (green vs white).

Contacts between quartzite and the surrounding graphitic schist are quite sharp, consistently transposed within the regionally dominant schistosity of the central domain.

Contact Relations:

No overlying unit has been mapped above the Granville Formation. The upper parts of this sequence, however, are progressively more graphitic and quartzose, and have been separated to the north as the Lincoln Gap Member (Cgl of Stanley et al., 1989).

To the east, in the vicinity of the Northfield Mountains, Granville-like graphitic albitic schist is present above chlorite phyllite of the Stowe Formation, and is gradational with very graphitic phyllite of the Ottauquechee Formation (Kraus, 1989). Whether or not this albitic schist represents a tectonic vestige of the Granville sequence is impossible to determine, but it may shed some information on its relationship with other graphitic units across the belt.

Type Locality:

The section of graphitic albitic schist and thin dark quartzite layers within the White River, 0.3 km west of the town of Granville, will continue to be utilized as the type and reference localities although, as noted above, the Formation is redefined to include only the graphitic albitic schist and associated quartzite **without** carbonate or non graphitic schist.

Depositional environment:

Due to a lack of any type of sedimentary features or observable, relict stratigraphic markers, very little information concerning depositional environments can be accrued. The pervasive and homogeneous graphite and sulfides are the best source of information; both are diagnostic of anoxic, reducing conditions (Krumbein and Garrels, 1952). These conditions are generally regarded as indicative of restricted or closed, local basins, with deposition of anoxic mud in small discontinuous lenses. This type of deposition is common in many modern environments, particularly in rift systems such as the east African rift system (Bally, 1982; Burgess et al., 1987). Anoxic sediments are primarily restricted to localized areas because of the pervasive flow of oxygen-rich waters from high latitudinal polar ice caps (Wilde and Berry, 1987). High density, cold polar meltwater flows to lower latitudes in deeper parts of the oceanic water column, thus oxygenating otherwise oxygen-poor environments (Wilde and Berry, 1987). This type of oxygenation restricts anoxic deposition to those areas isolated from meltwater commingling. During geologic periods of glacial inactivity, oxygenation of deep water realms would not occur, consequently leading to an increase in anoxic sedimentation. Anoxic conditions during these periods may actually be quite far reaching, often occurring into the shallow water phototrophic zone (0.5 to 2.0 km depth; Wilde and Berry, 1987). During these periods, numerous benthic fauna and flora populations may be severely affected by the rising anoxic water interface. Rising anoxic waters ultimately lead to catastrophic death assemblages, resulting in large scale organic influx into the sediment source. This in turn results in large distributions of organic-rich shales, concomitant with periods of glacial (or polar ice cap) abatement (Wilde and Berry, 1987).

A major episode of glacial inactivity occurred during the lower Middle Cambrian, into the Lower Ordovician, which is synchronous with world wide occurrences of organic, graphitic slates and shales (Wilde and Berry, 1987). Rocks of the Granville, Battell, and Ottauquechee Formations are, at least in part, believed to be of similar age (see Clarke, 1934; Cady et al., 1962; Zen, 1967). The large scale distribution and homogeneity of these graphitic schists agrees well with the model of low oxygenation, even within relatively shallow waters. The lack of carbonate within the Granville Formation, and the greater amount of sulfides (relative to the Battell), would indicate deposition in relatively deeper waters, richer in sulfides, below the calcium compensation depth. The "upward" increase in graphite content, into the Lincoln Gap Member, may be the result of an increase in oxygen deprivation of waters, to higher bathymmetric levels. The upward progression, into the Lincoln Gap Member, of quartz abundance, and quartzite beds, may be indicative of a seaward regression, either by eustaic sea level lowering or through simple basin infilling (Kinsman, 1975). Regression would lead to greater landward erosion and an increase in coarse sediment influx, to once distal parts of a sedimentary basin. This might also explain the abundance of quartzite within the easterly adjacent Ottauquechee Formation.

The total lack of greenstone within the graphitic units in this area, and the lack of a demonstrable greenstone/graphitic schist depositional contact within the central Vermont belt, suggests deposition within a basin setting following (at least) the main phase of rifting. This would place the Granville Formation in a stratigraphic position somewhere above the Pinney Hollow and Mt. Abraham lithologies.

The Ottauquechee Formation (Co)

Within much of central Vermont the Ottauquechee Formation is characterized by a very graphitic black to dark gray phyllite interlayered with the characteristic thin to thick dark gray to black quartzites. These two lithological types were first defined by Perry (1928) for a similar sequence in the Ottauquechee River valley:

actually be quite far reaching, often occurring into the shallow water phototrophic zone (0.5 to 2.0 km depth; Wilde and Berry, 1987). During these periods, numerous benthic fauna and flora populations may be severely affected by the rising anoxic water interface. Rising anoxic waters ultimately lead to catastrophic death assemblages, resulting in large scale organic influx into the sediment source. This in turn results in large distributions of organic-rich shales, concomitant with periods of glacial (or polar ice cap) abatement (Wilde and Berry, 1987).

A major episode of glacial inactivity occurred during the lower Middle Cambrian, into the Lower Ordovician, which is synchronous with world wide occurrences of organic, graphitic slates and shales (Wilde and Berry, 1987). Rocks of the Granville, Battell, and Ottauquechee Formations are, at least in part, believed to be of similar age (see Clarke, 1934; Cady et al., 1962; Zen, 1967). The large scale distribution and homogeneity of these graphitic schists agrees well with the model of low oxygenation, even within relatively shallow waters. The lack of carbonate within the Granville Formation, and the greater amount of sulfides (relative to the Battell), would indicate deposition in relatively deeper waters, richer in sulfides, below the calcium compensation depth. The "upward" increase in graphite content, into the Lincoln Gap Member, may be the result of an increase in oxygen deprivation of waters, to higher bathymmetric levels. The upward progression, into the Lincoln Gap Member, of quartz abundance, and quartzite beds, may be indicative of a seaward regression, either by eustaic sea level lowering or through simple basin infilling (Kinsman, 1975). Regression would lead to greater landward erosion and an increase in coarse sediment influx, to once distal parts of a sedimentary basin. This might also explain the abundance of quartzite within the easterly adjacent Ottauquechee Formation.

The total lack of greenstone within the graphitic units in this area, and the lack of a demonstrable greenstone/graphitic schist depositional contact within the central Vermont belt, suggests deposition within a basin setting following (at least) the main phase of rifting. This would place the Granville Formation in a stratigraphic position somewhere above the Pinney Hollow and Mt. Abraham lithologies.

The Ottauquechee Formation (Co)

Within much of central Vermont the Ottauquechee Formation is characterized by a very graphitic black to dark gray phyllite interlayered with the characteristic thin to thick dark gray to black quartzites. These two lithological types were first defined by Perry (1928) for a similar sequence in the Ottauquechee River valley:

"The most typical exposures are found within the Ottauquechee River valley about halfway between West Bridgewater and Bridgewater Corners, and consequently it is proposed to name the formation after this stream."

Perry (1928; p. 27-28) described this unit as:

"Two rock types in intimate association are typical of this formation, a massive, slate-grey quartzite and a dark grey or black phyllite. These are in distinct contrast with the green schists above and below, i.e., to the east and west, and are readily distinguished from them. ... The quartzite occurs in sharply defined beds from a few inches to ten feet in thickness, interstratified with the phyllite which makes up the major part of the formation. ... The phyllite of the Ottauquechee Formation is a black, fine-grained, slightly micaceous schist with distinctly platy parting. The phyllite, as a rock type, is very uniform in appearance throughout the belt of outcrop. ... Light colored sandy or micaceous layers a few millimeters thick are occasionally present in the phyllite and in southern Plymouth beds of light colored chlorite-mica schists and micaceous quartzite appear in considerable numbers."

Although varying in the amount of quartzite, the Ottauquechee Formation in the Granville-Hancock area, as mapped and specifically described by Osberg, shows similar relationships between the black phyllite and dark quartzites.

Lithologic description:

In addition to the units described by Perry and Osberg, the Ottauquechee Formations, as mapped in the Granville - Hancock area, also includes; a white to gray weathering, massive, well foliated quartzose wacke with rare detrital grains of blue quartz and magnetite. Centimeter-scale interlayered quartzose-wacke (as described above) and black phyllite, defined by Osberg (pers. comm., 1988) as "bacon rock" and similar to the sandy layers described by Perry (1928). Heterogeneous, compositionally diverse, greenstone are locally present as discontinuous bodies within the black phyllite, usually associated with ultramafics. Ultramafic bodies, including serpentinite and associated talc schist are present along the fault contacts with the structurally underlying Stowe and Pinney Hollow Formations and within the black phyllite sequence.

I. Black phyllite (Cobp):

Within the Granville-Hancock area, the Ottauquechee Formation is predominantly composed of a dark gray to black, rusty weathering, very graphitic quartz-muscovite-chlorite $\pm$ pyrite $\pm$ ankerite $\pm$ ilmenite phyllite with graphite comprising 3-4 bulk modal percent as inclusions within muscovite and chlorite, and as discreet grains along foliation planes (table 6a). Pyrite occurs as millimeter to centimeter scale framboids,

Black phyllite	HQ 141	HQ 325	HQ 425	HQ 460	Black qtzite.	HQ 130	HQ 290	HQ 455
Porphyroblasts								
albite	-	-	tr	-		-	-	-
Groundmass								
quartz	32	36	25	33		95	86	83
albite	12	13	9	8		3	6	13
sericite	40	40	45	35		2	6	2
chlorite	15	5	10	15		tr	1	-
apatite	-	-	-	-		tr	-	-
pyrite	5	2	4	3		tr	tr	1
graphite	6	4	7	6		-	-	-
magnetite	-	-	-	-		tr	1	tr
ilmenite	tr	tr	tr	tr		tr	tr	1
Grain Size								
Porphyroblasts	-	-	0.5mm	-		-	-	-
Groundmass	0.1-1.0mm	0.1-0.4mm	0.05-0.4mm	0.1-0.4mm		0.1-2.0mm	0.1-2.0mm	0.1-2.0m.
Texture	phyllitic	phyllitic	phyllitic	phyllitic		granulose	granulose	granulos
Location	K 36	V 38	X 42	U 41		S 41	N 38	Y 45

Table 6a Modal analyses of Ottauquechee Formation phyllite and quartzite samples.

usually deformed within the closely spaced regional dominant schistosity of this eastern domain (Sne). These pyrites commonly show framboidal tails of fine grained quartz and sericite. Pyrite is homogeneously found throughout the black phyllite and locally present as fine grains within some of the quartzite horizons.

Westward from the main belt of the Ottauquechee Formation, the black phyllite member becomes progressively richer in albite as ground mass and as 1-3 millimeter diameter porphyroblasts. This progression is also characterized by an increase in the chlorite content with a corresponding decrease in the muscovite content.

Another sequence found in within the Ottauquechee Formation in close proximity to the contact between the black phyllite and the sandy quartzose wacke bodies is an interbedded black phyllite / tan wacke sequence defined by Osberg as the "bacon rock". The lack of continuity of this sequence and its variable thickness, excludes it from being defined as a discreet member.

Greenstone:

Within the eastern Ottauquechee belt, numerous albitic gray to green albite-chlorite-biotite-quartz-muscovite ± magnetite ± zircon ± ilmenite ± calcite mafic schists (greenstone) or "gray stones" are found as 1-3 meter thick discontinuous horizons in sharp contact with surrounding black phyllite. Many of these mafic units do not appear to be part of a larger continuous stratigraphic horizon or vestiges of any other larger mafic body / ultramafic unit, and are too small to be portrayed as a separate map unit. They have therefore been included within the black phyllite map unit.

II. Black quartzite (Cobq):

The quartzite member of the Ottauquechee Formation includes massively bedded, dark gray and black, fine grained, dynamically recrystallized, quartz with minor amounts of fine grained muscovite with graphite inclusions that give this unit its distinctive gray to black color. Trace amounts of calcite, magnetite and ilmenite are also present as secondary grains (table 6a). These quartzites, similar to those found in the Granville Formation, are diagnostically cross cut by syn-tectonic quartz veins, complexly folded and deformed.

Thicknesses of these quartzites are quite variable; thin discontinuous quartzose layers within the black phyllite are centimeter-scale in thickness, whereas zones of pervasive interbedded thick quartzite with minor interbeds of black phyllite have discrete quartzite layers obtaining a maximum thickness of 30 meters within the type reference section along Jeep Trail Mountain in the western Ottauquechee belt (see Plate 1, L35). Quartzites are commonly interlayered with black phyllite, occurring predominantly in pervasive zones which, although discontinuous, are traceable along strike (plate 1). This type of interlayering may represent actual interbedding of quartzite and graphitic schist protoliths.

III. Quartzose wacke (Coss):

Sandy, tan, gray to white weathering quartz-albite-muscovite-chlorite-ankerite ± magnetite ± ilmenite ± pyrite wacke occurs as discontinuous, often, lens-like bodies, within the black phyllite (table 6b; figure 2.8). Some of these bodies obtain thicknesses of 10-50 m and usually occur within the eastern Ottauquechee belt although some thin ones, interlayered and containing phyllite laminations, are present within the western belt south of Jeep Trail hill (Plate 1 M35). These discontinuous bodies appear to make up once coherent stratigraphic units that have been disarticulated during orogenesis. Their along strike discontinuity and local pervasiveness seem to indicate that they once comprised a much larger lens-like body and were not a regionally continuous internal stratigraphic horizon.

Thatcher Brook Member (Cotb):

The previously mentioned transition of black phyllite, westward into a graphitic albitic schist, richer in chlorite and poorer in muscovite, has been separated as a distinct member called the Thatcher Brook Member for a section of exposure along Thatcher Brook, 1 km up Thatcher Brook road, eastward from its intersection with Route 100 (plate 1, S35; table 6b; figure 2.9). This unit also includes numerous talc schist and serpentinite bodies, small discontinuous quartzites, discontinuous greenstones, and heterogeneous, sandy, albitic schist as discontinuous lens-like bodies or as actual blocks surrounded by a graphitic albitic schistose matrix.

The Thatcher Brook Member is consistently situated between the homogenous graphitic black phyllite of the Ottauquechee Formation to the east and the graphitic albitic schist of the Granville Formation to the west. The matrix of black phyllite, progressively increasing in albite porphyroblasts and groundmass content to the west, marks a transition between the two regionally mapped formations.

Contact Relations:

The contact between the Ottauquechee, Pinney Hollow, and Stowe Formations has been redefined during this study. Previously interpreted as stratigraphy underlying the Pinney Hollow Formation, the Ottauquechee is reinterpreted as a pre-peak metamorphic thrust slice tectonically overlying a continuous Pinney Hollow/Stowe depositional sequence (plate 1; Armstrong, 1989; Armstrong et al., 1988a, 1988b, 1989b; Kraus, 1989). A well defined lithologic transition between albitic chlorite-rich schist (Pinney Hollow) and chlorite-rich phyllite (Stowe) has been carefully mapped within several tectonic windows beneath the overlying Ottauquechee Thrust Slice (Armstrong et al., 1988a; plates 1&2, S42-AA46).

Sandy schist	HQ 139	HQ 381	Thatcher Bk. Mbr.	HQ 137	HQ 304	HQ 331	HQ 369
Porphyroblasts							
albite	-	-		3	2	2	-
Groundmass							
quartz	36	33		25	20	18	19
albite	43	44		18	30	20	25
sericite	13	5		37	25	40	35
chlorite	3	11		13	10	10	9
ankerite	2	1		-	2	tr	-
apatite	-	-		-	-	tr	-
calcite	-	-		-	-	tr	tr
pyrite	tr	1		1	3	2	3
tourmaline	tr	tr		tr	tr	-	-
graphite	1	2		3	6	6	6
magnetite	1	tr		-	1	tr	1
ilmenite	1	3		tr	1	2	2
Grain Size							
Porphyroblasts	-	-		1.0mm	1.0-1.5mm	0.5-2.0mm	-
Groundmass	0.1-2.5mm	0.1-1.0mm		0.1-1.0mm	0.1-0.4mm	0.05-0.4mm	0.1-0.4mm
Texture	schistose	schistose		schistose	schistose	schistose	phyllitic
Location	K 36	Z 40		M 35	S 36	V 38	U 30

Table 6b Modal analyses of Ottauquechee Formation sandy schist and Thatcher Brook Member samples.

Figure 2.8 Quartzose (sandy) schist unit (Coss) of the Ottauquechee Formation, showing interlayering of detrital blue quartz-bearing, white quartzose layers with brown weathering graphitic schist/phyllite horizons. Pictured is an Fn fold, folding Sn-1 foliation. Pencil is 5 cm long. Picture taken behind white house along Jeep Trail Hill road, map coordinates K 36.

Figure 2.9 Thatcher Brook Member of the Ottauquechee Formation, displaying finely laminated dark gray graphitic phyllite and light gray non graphitic, albitic schist. The albitic schist also occurs as small "blocks" which may either be olistoliths or boudinage. This unit typically contains pods of ultramafic and is interpreted as tectonic melange. Note the Sn+1 folds which deform Sn and have axial planes coplanar to undeformed Sn regional orientation. Picture is from the type locality in Thatcher Brook, map coordinates M 35. Pencil is 4 cm in field of view.

The tectonic character of this contact is well demonstrated by upper and lower plate truncations of dark quartzite and greenstone, respectively (plate 1, T43). In addition, the presence of serpentinite and talc schist "exotic" tectonic slivers **along** this contact, further documents this tectonic interpretation.

No units were found overlying the Ottauquechee Formation in the study area, although quartz-feldspar granulite, garnet-bearing phyllite, and felsic and mafic volcanics of the Moretown Formation, make up the next highest structural unit to the east (Hatch, 1987; Westerman, 1987).

Reference section:

The black phyllite (and associated dark quartzite, described below) is characteristically observed within a sequence of small cliffs, approximately immediately east of the Jeep Trail road, 1.5 km south of its intersection with the Kendall Brook road (plate 1, L35). The sandy quartzose schist is well represented in a series of exposures, 0.3 km south of the black phyllite locality, within several large fields (plate 1, M35). The Thatcher Brook Member can be seen at its type locality, along Thatcher Brook (plate 1, S35), at an elevation of 350 m (1150 ft).

Depositional environments:

The very graphitic black phyllite was deposited within an anoxic environment, similar to that described for the Granville and Battell Formations. The significantly greater amount of graphite (3 to 4 modal percent), sulfides, and phyllosilicates (chlorite and sericite), suggest deposition of Ottauquechee sediments under even more anoxic conditions farther from the source area. The lack of any carbonate within the black phyllite also suggests deposition below the calcium compensation depth.

The abundance of quartzite within this formation has always posed a great problem for workers. The similarity of the black phyllite to Granville and Battell graphitic schist makes east to west correlation of the three units (in that order) attractive. The general lack of quartzite in the two western graphitic units (Granville and Battell) poses a problem of Ottauquechee quartzite source. This problem may be explained in several ways:

1. The Ottauquechee quartzites formed as moderate to deep water cherts, derived from the sea bottom death assemblages of silica-shelled organisms. This model would also apply to the quartzites from the Granville and Battell Formations as well. This model is almost certainly invalid since oxygen isotope analyses have shown that the silica was derived as quartz grains from a continental source (Kraus, 1989; Stanley et al., 1989). In addition, microscopic analysis of individual quartz grains has shown that their blue color is a result

of an exsolved titanium-bearing phase (rutile), a metamorphic process that occurs within continentally derived quartz at upper amphibolite to granulite facies (Bohlen et al., 1983).
2. The Ottauquechee does not correlate with the other units and is either older or younger with a discreet sediment source. This is not favored since albite-bearing, Granville-like schist is found within the Ottauquechee Formation along much of its western limit, along the White River valley, in close proximity to the main belt of Granville Formation. These two formations also appear to occupy the same structural level immediately above the Pinney Hollow (and Stowe) Formation (plate 2).
3. The Ottauquechee quartzites were turbidites or gravity flows derived from channelized sources to the west, which bypassed the majority of the intervening material. Again, this is not favored due to the lack of any thick quartzite deposits within the Granville or Battell that may be relict channels. Turbidites usually suspend numerous grain-types including feldspar and mica, although a quartz-rich source may be devoid of any other phases. In addition, modern day studies of turbidites and especially gravity flows, show that some bedload suspension will be randomly or sporadically deposited during basinward transport from source to farthest bedload deposit (Lowe, 1982). This means that some large quartzite deposits should be found within at least part of either the Battell or Granville units.
3. Quartzite was deposited from longitudinally channelized sources, either from a continental promontory to the north or south. This model requires erosion from a promontory source area with marine transport either to the north or south into a deep part of the basin (Ottauquechee) outboard of relatively more proximal sediments. If this model is correct, southern (or northern) correlatives of the Granville and Battell graphitic units should show an increase in quartzite abundance. Unfortunately, the majority of both northern and southern correlatives are either structurally overlain by large thrust sheets of more distal lithologies, or have not been distinguished as separate units. More mapping needs to be done in order to characterize and separating Battell and Granville correlatives, both to the north and south of the type section in central Vermont.
4. The quartzite were **not** shed from a western source during passive margin development, but were instead accreted into the Ottauquechee from an **eastern** source during the early part of the Taconian orogeny (Kraus et al., 1988). Although little evidence substantiates this model, very little refutes it either. Several points that may agree with this model include, the general westward decrease in quartzite abundance and the presence of continental-like basement within the core of the Bronson Hill Terrane; the presumed Taconian island arc complex (Hall and Robinson, 1982; Robinson et al., 1989). This basement could have supplied the quartz during orogenic uplift and subsequent transport of

detrital material into a growing accretionary complex to the west (Stanley and Ratcliffe, 1985). The main argument against an eastern source is that the majority of rocks presently exposed within the Bronson Hill Terrane are more feldspathic than quartzose (Zen et al, 1983). In addition, eastern derived material requires that the island arc terrane be in close proximity during Ottauquechee deposition. Such deposition would almost certainly dictate that the Ottauquechee was receiving sediments within a forearc or marginal basin setting. Such environments usually contain an abundance of immature sediments composed of quartz, feldspar, and mica minerals along with a variety of clast types and sizes (Hamilton, 1988). Furthermore, Ottauquechee-like rocks to the north in southern Quebec (Sweetsburg Formation of Clark, 1934) are part of a coherent passive margin sequence believed to have been deposited during the Cambrian through Lowermost Ordovician (Slivitzsky and St. Julien, 1987). These facts seem to rule out an eastern, island-arc, or magmatic arc continental basement source.

Blue quartz-bearing quartzites, with abundant recrystallized plagioclase grains, occur within the Upper Cambrian-Lower Ordovician Hatch Hill Formation of the Taconic allochthon (lower slice, Potter, 1979). These thin, 0.5 to 1m thick quartzite beds are interlayered with carbonaceous schist, similar to Battell and Granville lithologies. The presence of abundant feldspar within the quartzite may indicate either deposition within a proximal environment (relative to the source area(s)), where suspended bedload contained material other than just mature quartz grains, or deposition of mature blue quartz grains and more heterogeneous, feldspar/mica-rich sediments from different source areas. In either case, the presence of blue quartz within this relatively proximal, post-rift passive margin sequence supports a western, Grenville source for the blue quartzites within the different carbonaceous/graphitic schists.

Detrital quartz and carbonate within the sandy quartzose schist unit indicate deposition within a relatively proximal and shallow water environment, respectively. Similar rock types are not found in any of the graphitic lithologies to the west or in any of the underlying lithologies. The interbedded nature of the contact of this unit with the black phyllite, and the discontinuous presence of the sandy schist argue in favor of deposition within isolated areas. In addition, the general lack of graphite within the majority of these lensoidal bodies favors deposition from a different source than the quartzites or black phyllite units. Carbonate could have been deposited during quartz sediment deposition in a shallow water environment as reworked deposits shed from more proximal (western, southern, or northern) source(s), possibly the carbonate platform. This model of formation coincides with a probable western source model for the Ottauquechee quartzite, although the two could certainly be mutually independent.

Stowe Formation (CZs)

This sequence of chlorite-quartz-sericite phyllite and associated chlorite-epidote-albite-carbonate mafic schist or greenstone was first named by Cady (1956) for a sequence of outcrops in the town of Stowe, Vermont. Because it overlay the Ottauquechee Formation, the Stowe was interpreted as the stratigraphically highest and youngest part of the pre-Silurian eastern cover sequence found within central Vermont (Osberg, 1952). This sequence was mapped to the south by Perry (1928) and, initially, by Richardson (1924) who named this sequence the Bethel Schist, described as:

> "... fine grained, greenish, schistose, highly metamorphosed sedimentary rocks which are more or less intimately associated with chlorite. These schists are, furthermore, characterized by numerous lenses, or eyes, and stringers of granular quartz" (from Perry, 1928; p. 29).

The Stowe Formation within the Granville-Hancock area as initially mapped by Osberg(1952) included all schist and greenstone found immediately above or to the east of the major mass of black phyllite of the Ottauquechee Formation (Plate 1). Because of this mapping constraint, many small bodies of Ottauquechee-like black phyllite and quartzite were included within the Stowe Formation.

Contact relationships between the Stowe Formation and the overlying, presumed Lower Ordovician Moretown Formation of the Missisquoi Group, were not described in detail for the central Vermont sequence. This contact, to the north in the area of Montpelier, was described as a conformable, depositional sequence and to the south, in the vicinity of Bridgewater, as an erosional unconformity (Perry, 1928). In the Jay area of northern Vermont, this contact was described as a synmetamorphic fault, with numerous tectonic slivers of Moretown-like lithologies intercalated with Stowe lithologies along a thrust zone (Roy, 1984; Stanley et al., 1984). Current 1:24,000 scale mapping by the author within the West Dover and Jacksonville 7.5' Quadrangles has shown that the Stowe is bound to the east (structurally upwards) by a pervasive mylonitic shear zone, the upper plate containing quartz-feldspar granulite, garnet phyllite, and feldspathic and mafic volcanic rocks mapped as Moretown Formation. Within the town of East Dover, this shear zone envelopes a 3km long by 1.5-2km wide dunite body, separating Stowe from Moretown lithologies.

Lithologic Description:

The Stowe Formation as defined by Cady (1956), interpreted by Osberg (1952), and as defined within this study area, consists of two predominate lithologies; chlorite-quartz-sericite phyllite and chlorite-epidote-quartz-albite-ilmenite-carbonate-amphibole greenstone.

I. Chlorite phyllite (CZs):

The predominant unit within the study area is a fine grained,dark green to light gray phyllite consisting of chlorite-quartz-muscovite ± albite ± pyrite ± magnetite ± ilmenite ± hematite ± ankerite ± epidote ± garnet, with characteristic iron-stained granular quartz (the "quartz eyes" of Richardson, 1924) with rare albite porphyroblasts, ranging in diameter from 0.5 to 1 mm, and as interstitial groundmass 0.1 to 0.4mm in diameter (table 7a). Although locally present throughout the Stowe Formation in other parts of Vermont, no quartzites were found within the Stowe chlorite phyllite in this study area. Interlayered white mica-chloritoid-biotite-quartz phyllite was found as discontinuous horizons within the chlorite phyllite and does not appear to make up a particular stratigraphic horizon, although it may be associated with other white mica-chloritoid bearing phyllites (Mount Abraham Formation) found farther west in the study area. The presence of biotite at this particular grade (sub- to biotite grade) requires a bulk composition below the garnet - chlorite join of an AFM. This is dissimilar to Mt. Abraham lithologies which are consistently above the garnet - chlorite join (chapter 4).

II. Greenstone (CZsg):

Associated with the predominant chlorite-rich phyllite are locally pervasive 1 to 20 m thick, locally continuous, light to dark green chlorite-epidote-quartz-albite-sphene-carbonate ± amphibole ± biotite ± magnetite greenstones (table 7b). Albite porphyroblasts are locally present and occur in diameters of 0.1 to 2mm. Although thin, these greenstones are locally continuous within the study area and are regionally thick (<1 kilometer in width) along the east flank of the Northfield Mountains (Kraus, 1989). They are compositionally very similar to Pinney Hollow greenstones, except that they lack (other than sphene) any significant Ti-bearing phase (usually rutile and/or ilmenite, both common to the Pinney Hollow greenstones).

Small lensoidal bodies of chlorite-rich phyllite, dark gray to black in color, consisting predominantly of chlorite with minor muscovite and quartz, are interlayered with quartzose granulite, in layers 1-3mm thick). This sequence of interlayered dark chlorite phyllite and quartzose wacke looks very similar to the intercalated black phyllite and wacke sequence described by Osberg as the "bacon rock". The lack of graphite (and abundance of chlorite) within this zone does distinguish this rock from the Ottauquechee bacon rock.

	HQ 411	HQ 418	HQ 422	HQ 462	HQ 464	HQ 466
Porphyroblasts						
albite	-	3	2	1	-	-
garnet	-	tr	-	-	-	-
Groundmass						
quartz	33	24	36	24	16	29
albite	17	7	5	10	8	14
sericite	10	23	29	19	33	17
chlorite	36	40	28	44	39	35
chloritoid	-	1	-	-	-	-
ankerite	tr	1	tr	tr	1	-
apatite	tr	tr	tr	tr	tr	tr
alanite	tr	-	-	-	tr	tr
pyrite	-	-	tr	-	1	-
tourmaline	-	tr	tr	1	1	1
zircon	tr	tr	-	tr	-	-
magnetite	1	1	tr	1	1	1
ilmenite	3	tr	tr	tr	tr	3
Grain Size						
Porphyroblasts	-	0.5-1.0mm	0.5-1.5mm	0.5-2.5mm	-	-
Groundmass	0.2-0.5mm	0.2-0.5mm	0.1-0.5mm	0.1-0.4mm	0.1-2.0mm	0.1-0.5mm
Texture	schistose	phyllitic	phyllitic	schistose	phyllitic	phyllitic
Location	X 43	X 44	Y 46	T 42	S 43	R 44

Table 7a Modal analyses of Stowe Formation pelitic samples.

	HQ 414	HQ 437	HQ 459
Porphyroblasts			
albite	-	1	-
Groundmass			
quartz	-	2	1
albite	25	21	24
epidote	33	30	23
biotite	-	1	-
chlorite	31	40	49
calcite	10	4	3
hornblende	-	tr	-
actinolite	tr	-	-
magnetite	1	1	tr
Grain Size			
Porphyroblasts	-	0.5-1.0mm	-
Groundmass	0.2-1.0mm	0.3-0.9mm	0.4-2.0mm
Texture	schistose	gneissic	schistose
Location	X 44	W 42	T 42

Table 7b Modal analyses of Stowe Formation greenstone samples.

Contact Relations:

The contact with the presently overlying Moretown Formation was not mapped during this study. Richardson (1924) and Perry (1928) described this contact as an unconformity. Recent workers, however, have reinterpreted this boundary as a tectonic contact, possibly developed during either the Taconian and/or Acadian orogenies (Hatch, 1982; Westerman, 1987). The Stowe Formation can be observed within several tectonic windows through the overlying Ottauquechee Formation. Within these windows, Stowe chlorite phyllite grades westward into more albitic and sericitic schist, with less chlorite and apparently little or no iron stained quartz layers. This albitic, sericite-rich schist is identical to, and traceable into a large belt of albite-sericite-chlorite-quartz schist and associated greenstone mapped as Pinney Hollow Formation (plate 1). This Pinney Hollow belt is also tectonically overlain by Ottauquechee. The similar structural position and, most significantly, the demonstrable lithologic gradation over a distance of less than 200 m, strongly argues in favor of the Pinney Hollow and Stowe Formations being a single lithotectonic unit, separable through a gradational facies change, tectonically overlain by the Ottauquechee Formation. This relationship redefines the stratigraphic (and structural) models depicting the Stowe as the stratigraphically highest, and youngest, pre-Missisquoi unit.

Ultramafics (s)

Numerous lens-like bodies of talc-steatite-calcite-magnesite phyllite, and serpentinite bodies, of ultramafic origin, appear either along syn-metamorphic fault zones as tectonized fault slivers, along the contact between the Ottauquechee and Pinney Hollow/Stowe Formations, or internal to the Ottauquechee Formation within the black phyllite sequence (Plate 1, table 8). These ultramafics occur as thin, 1-5 meter lenses with the rare occurrence of larger serpentinite bodies, some achieving diameters over 100 meters wide (Rochester Verde-Antique mine).

The origin of the ultramafics is highly questionable. Their presence within numerous syn-metamorphic shear zones, without any intervening Grenville-like lithologies, could possibly indicate an oceanic crustal origin with subsequent incorporation into accreted passive margin rocks (Ottauquechee Formation) during the Taconian Orogeny (Stanley and Ratcliffe, 1985). Another possible origin would be as ultramafic cumulates within a mafic igneous host that might have developed during Late Proterozoic\Lower Cambrian rifting of North America. Taconian orogenic processes could also account for their incorporation into the Ottauquechee. Although many of the ultramafics in central Vermont are not associated with mafic rocks, many of the serpentinites and dunites within southern Vermont do occur with a thick sequence of amphibolite (Skehan, 1961; Ratcliffe

	HQ 382A	HQ 416	HQ 434X	HQ 446X
Groundmass				
talc	95	33	93	11
chlorite	1	6	tr	3
carbonate	3	15	5	9
serpentine	tr	46	tr	75
magnetite	1	tr	2	1
chromite	tr	tr	tr	1
Grain Size				
Groundmass	0.5mm	0.5-1.0mm	0.5-1.0mm	0.3-1.0mm
Texture	phyllitic	schistose/ massive	schistose	massive
Location	AA 39	W 46	V 46	Z 43

Table 8 Modal analyses of ultramafic samples.

and Armstrong, mapping in progress). Whether or not the ultramafics are indigenous to true oceanic crust or highly rifted North American Laurentian basement is important towards understanding the magnitude of Iaptean rifting and the potential width of the passive margin basin.

LOCAL AND REGIONAL STRATIGRAPHIC CORRELATIONS

Intense and regionally pervasive deformation of the pre-Silurian belt, coupled with a total absence of fossil control, requires correlation of stratigraphic units solely on local lithologic analyses and observation of similar stratigraphic sequences in less deformed parts of the Taconide Zone, particularly the Taconic allochthon of New York and Vermont, and the Oak Hill Group of southern Québec. Observations of unique stratigraphic, lithologic, and paleontologic characteristics in the less deformed regions can be extrapolated to the more deformed central Vermont rocks in order to deduce pre-orogenic stratigraphic sequences, correlations, and gross ages of various lithofacies.

Several important lithologic, stratigraphic, and paleontologic characteristics are consistent in both the Taconic allochthon and the Oak Hill Group:

1. Homogeneous organic/graphitic slates, both carbonate- and non carbonate-bearing, overlie the majority of heterogeneous, coarse and fine grained clastic rocks.
2. Metavolcanic basalts and/or tuffaceous sediments are generally restricted to discreet stratigraphic horizons within the clastic rocks.
3. The graphitic/nongraphitic boundary, usually sharp but in some places intercalated over a small zone (1-50m), contains immature and mature white quartzite (continuous in Québec, lens-like in the Taconics) with fauna coeval with the Lower Cambrian Cheshire Formation, the basal part of the Iaptean carbonate platform (Walcott, 1888).
4. Dolomite within the Taconic section, immediately above the quartzite, occurs as small plugs and/or channel deposits, and contains fauna identical to those within the Québec Lower Cambrian Dunham Dolomite (figures 2.2 & 2.10; Potter, 1979). The Dunham progressively thins eastward into carbonate-bearing schist of the Sweetsburg Formation, very similar to the graphitic Lower Cambrian - Lower Ordovician West Castleton / Hatch Hill section of the Taconics.
5. Coarse clastics and metabasalt of the Taconic Rensselaer Formation is very similar to coarse clastics of the Québec Pinnacle Formation, immediately overlying

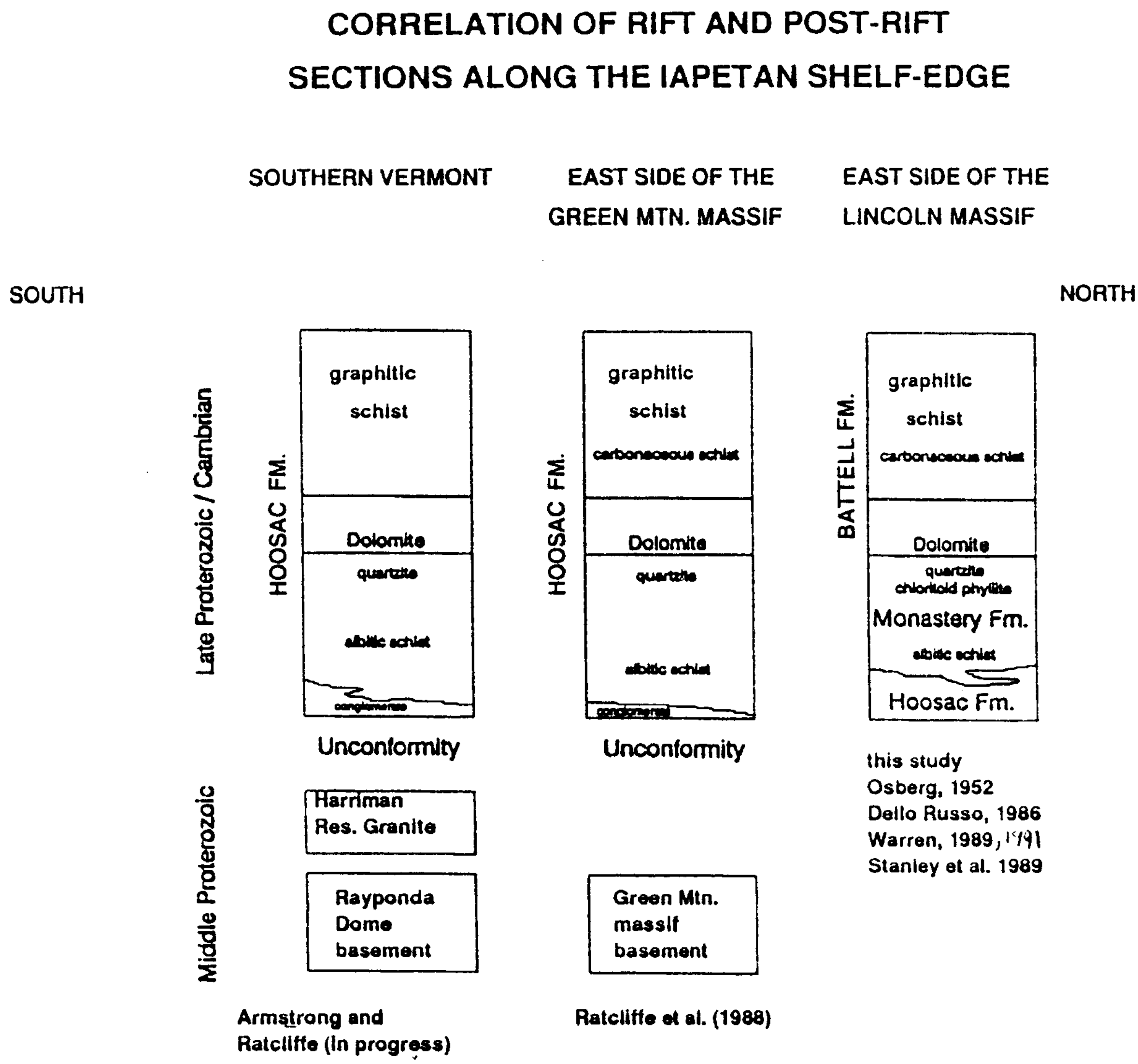


Figure 2.10 Stratigraphic correlation of distal rift/post-rift sections. Note the correlation of graphitic rocks (interpreted as post-rift), interlayered basal dolomite (the lower part of the carbonate platform, possibly Dunham correlative), and underlying basal quartzite and rift clastics. References are listed below each column.

metabasalt of the Tibbit Hill Formation.

6. Coarse clastics of both areas generally grade upward and eastward into aluminous slates and fine grained wacke without metabasalt. Rensselaer wacke does overlie fine grained aluminous slate but does indeed grade both upward and eastward into fine grain wacke and slates (Bomoseen wacke and Truthville slate, respectively).

Utilizing these characteristics, a working model for the correlation with the pre-Silurian of central Vermont can be established. Following stratigraphic correlation, the entire, presumably coherent, section can be placed within a passive margin model, following established theories on rifting, continental breakup, and subsequent post-rift subsidence. This will ultimately provide an interpretation into the original positions of all lithofacies within the Late Precambrian / Lower Ordovician Iaptean passive margin (plate 4).

I. Central Vermont stratigraphic correlations

1. Based upon the previously discussed idea that all graphitic rocks are parts of a once continuous stratigraphic sequence (above the coarse clastics), the Battell, Granville, and Ottauquechee Formations are shown as west to east facies. The presence of carbonate within the Battell and the lack thereof within the Granville and Ottauquechee, antithetic to increasing graphite content, indicates that the Battell was deposited in relatively shallow, quiescent water, whereas the Granville and Ottauquechee were deposited below the calcium compensation depth in relatively anoxic water. Furthermore, the presence of ultramafics within the Ottauquechee suggests its close association with either oceanic or severely attenuated continental crust, brought together during accretion of the extreme continental margin or during deposition upon transitional to oceanic crust (Armstrong et al., 1988a, 1988b; Kraus et al., 1988). The eastward increase in graphite and sulfides further indicates that the units are still within their proper west to east (proximal to distal, relative to the shelf edge) basin positions. Similar sections have been mapped in the Wilmington Dome region as Hoosac Formation (figure 2.10; Skehan, 1961). These dolomites were either interpreted in earlier studies as Dunham (and Battell) correlatives (Brace, 1953; Chang et al., 1965) or lower stratigraphic sections, part of the Tyson or Readsboro Formations (Skehan, 1961; Thompson, 1972). The Plymouth Member also consists of interbedded dolomite and marble, in some places over 150 m thick, overlying a thin wacke and schist section capped by vitreous white quartzite. This section is very similar to the Battell carbonate-bearing

graphitic schist, and was correlated with it by Doll and others (1961) on the Centennial State Map.

Stratigraphic positioning of the various carbonate-bearing lithologies was based entirely on the premise that all major contacts were depositional. More recent studies have demonstrated that many of these contacts are actually tectonic (Roy, 1984; Stanley and others, 1984; Stanley and Ratcliffe, 1985; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Haydock, 1988; Prewitt, 1989; Walsh, 1989). Many of these previously interpreted lower stratigraphic sequences are lithologically similar to those found in stratigraphically higher sections, and may be stratigraphic equivalents tectonically repeated across strike. Future work within these sections will be critical in determining which contacts are truly depositional and which are tectonic; some may eventually turn out to be undeterminable. These sections demonstrate the problems associated with attempts to interpret stratigraphic sections within highly deformed areas, particularly those with fault bounded contacts.

2. Fine grained, somewhat heterogeneous, metavolcanic- (greenstone) bearing albitic schists and chlorite phyllite of the Fayston (Walsh, 1989), Pinney Hollow, and Stowe Formations are correlated as a west to east (respectively) facies transition. The eastward fining of Pinney Hollow albitic schist into Stowe chlorite phyllite, and the similarity of albitic schist within the Pinney Hollow and Fayston Formations, is the primary lithologic evidence for this interpretation. This gradation is augmented by a progressive west to east (Fayston, Pinney Hollow, Stowe) decrease in quartz-feldspar granulose metawacke, some containing detrital feldspar and rutilated quartz. These units are predominantly interlayered with albite schist and most likely represent interbeded wacke and psammite sequences, the distal parts of thicker proximal sequences that may have been deposited to the south and west along the rifted margin. The progressive eastward decrease in grain size, although a primary result of metamorphism, is also controlled by changes in bulk composition; rocks within the eastern part of the belt tend to be less rich in quartz and particularly feldspar, and richer in alumina-bearing phyllosilicates (Tables 1-8). Modern day quartz-feldspar-rich analogues are most commonly found within actively rifting continental environments (Kinsman, 1975). Geochemical analyses of metavolcanic rocks within these units suggests a rift-related origin (Coish, 1987).

3. Aluminous phyllite of the Mt. Abraham Formation is found in gradational contact with the Pinney Hollow schist (plate 1). Similar aluminous lithologies are found in the upper part of the Monastery Formation, and to the north within the Foot Brook Member of the Underhill Formation (Albee, 1957; Doll et al., 1961). Mt. Abraham-like lithologies are also found in depositional contact with albitic schist, greenstone, and wacke of the Fayston

Formation (Walsh, 1989). Graphitic schist is also found within this area, **above** the aluminous schist, in gradational contact. This sequence of Fayston albitic schist grading into aluminous schist, which grades into graphitic schist, is very similar to the Monastery - Battell section of albitic, aluminous, and graphitic schist sequence, described in this chapter. The similarities of these two sequences leaves open the possibility that the two aluminous sequence are actually parts of a discreet stratigraphic horizon, possibly continuous across a large part of a rift basin. Although possible, this seems unlikely since aluminous phyllites of the Taconic range occur at various stratigraphic levels within the rift-clastic section (Ratcliffe, 1975; Potter, 1979; Stanley and Ratcliffe, 1985). The general lack of greenstone within the aluminous schist suggests that the aluminous rocks were deposited after the main phase of rifting and rift volcanism. The total absence of greenstones demonstrably in depositional contact with graphitic rocks, suggests that the graphitic units were deposited even later, possibly following termination of rifting.

4. Bounded by the Underhill Thrust zone, the Monastery Formation contains local thin layers of pebble conglomerate and fine grained wacke, similar to the Hoosac Formation. Albitic schist horizons, some with carbonate, have been mapped within the Hoosac Formation, west of the thrust zone (Tauvers, 1982; Warren, 1989). Although the two formations are bounded by a complex and pervasive synmetamorphic thrust zone (the Underhill Thrust Zone), this wacke/schist intercalation suggests a pre-tectonic association of Hoosac and Monastery lithologies, probably as a west to east fining grain-size transition (plate 4). Hoosac lithologies are traced westward around the northern end of the Lincoln massif into similar and coarser, grittier rocks mapped as Pinnacle Formation (Tauvers, 1982; DelloRusso, 1986). Although original stratigraphic thickness cannot be properly estimated for the Monastery Formation, the abundance of albitic schist and its large spatial distribution indicate that it is probably much thicker than its western counterparts (figure 2.2; plate 4).

II. Correlation of Taconic/Oak Hill and Central Vermont sequences

1. Interlayered graphitic and nongraphitic schist with graphitic marble lenses of the Central Vermont Monastery/Battell section (Hancock Branch sequence) is very similar to the southern Quebec Oak Hill, Lower Cambrian Frelighsburg/Sweetsburg contact, also containing graphitic marble (figure 2.2). The western part of the Oak Hill section is separated by the Lower Cambrian Cheshire Quartzite / Dunham Dolomite sequence, marking the shelf edge/slope-rise transition (Clark, 1934; Rodgers, 1968; Colpron et al., 1987).

Bedded, 10-100m thick dolomite and marble occurs within the Battell Formation at the central Vermont type locality (plate 1). Based on the similarity with the Lower Cambrian sequence of southern Québec, the White River Member, and thus, the Battell Dolomite, are interpreted as Lower Cambrian shelf edge units. The presence of vitreous white quartzite in both the Frelighsburg and the interpreted upper part of the Monastery sections, probably represents distal deposition of the Lower Cambrian Cheshire Quartzite (figure 2.2; Colpron et al., 1987).

Similar dolomite/graphitic schist sequences are found in southern Vermont. The Plymouth Member of the Hoosac Formation overlies a thin coarse clastic sequence (Tyson and Hoosac Formations) along the eastern flank of the Green Mountain massif (plate 3; Perry, 1928; Brace, 1953).

2. Based on the positioning of **1** along the shelf edge, proximal slope-rise turbidite sequences within the Taconic Lower Cambrian - Lower Ordovician West Castleton and Hatch Hill Formations, devoid of thick, bedded dolomite, are correlated as an eastern (distal) facies of the Battell Formation (figure 2.11). The similarity of the Battell with the Quebec Sweetsburg Formation, **outboard** of the Dunham Dolomite, supports this interpretation. Similar rock-types are absent to the south, probably due to their truncation along the continuation of the Child's Mtn. Thrust (plate 3). This section would lie west of the noncarbonate-bearing Granville and Ottauquechee formations which probably represent outer-rise and abyssal plain deposits, deposited below the calcium compensation depth (plate 4).

3. The interlayered contact between aluminous fine grained wacke of the Taconic Truthville Member of the Nassau Formation, Metawee Formation, and graphitic slates of West Castleton/Hatch Hill Formations (see Potter, 1979; Ratcliffe, 1979; Rowley et al., 1979) is similar to the sequence of albitic and aluminous schist, and quartzite (Monastery Formation) and overlying interlayered graphitic and nongraphitic schist of the White River Member of the Battell Formation (figures 2.2 & 2.12). The Taconic section also includes the underlying Rensselaer coarse wacke and interlayered basalt. The mapped eastward fining of Rensselaer clastics into chloritoid-bearing phyllite of the Chatham sequence (Ratcliffe, 1979, 1987) is markedly similar (albeit lower grade) to the Mt. Abraham Formation, aluminous phyllite of the Monastery Formation, and aluminous rocks above the Fayston Formation. This grain size change and compositional change (less quartz and feldspar, more micaceous sediments) may mark the eastward transition into the finer grained eastern Vermont clastics (Fayston, Pinney Hollow, and Stowe Formations). It is

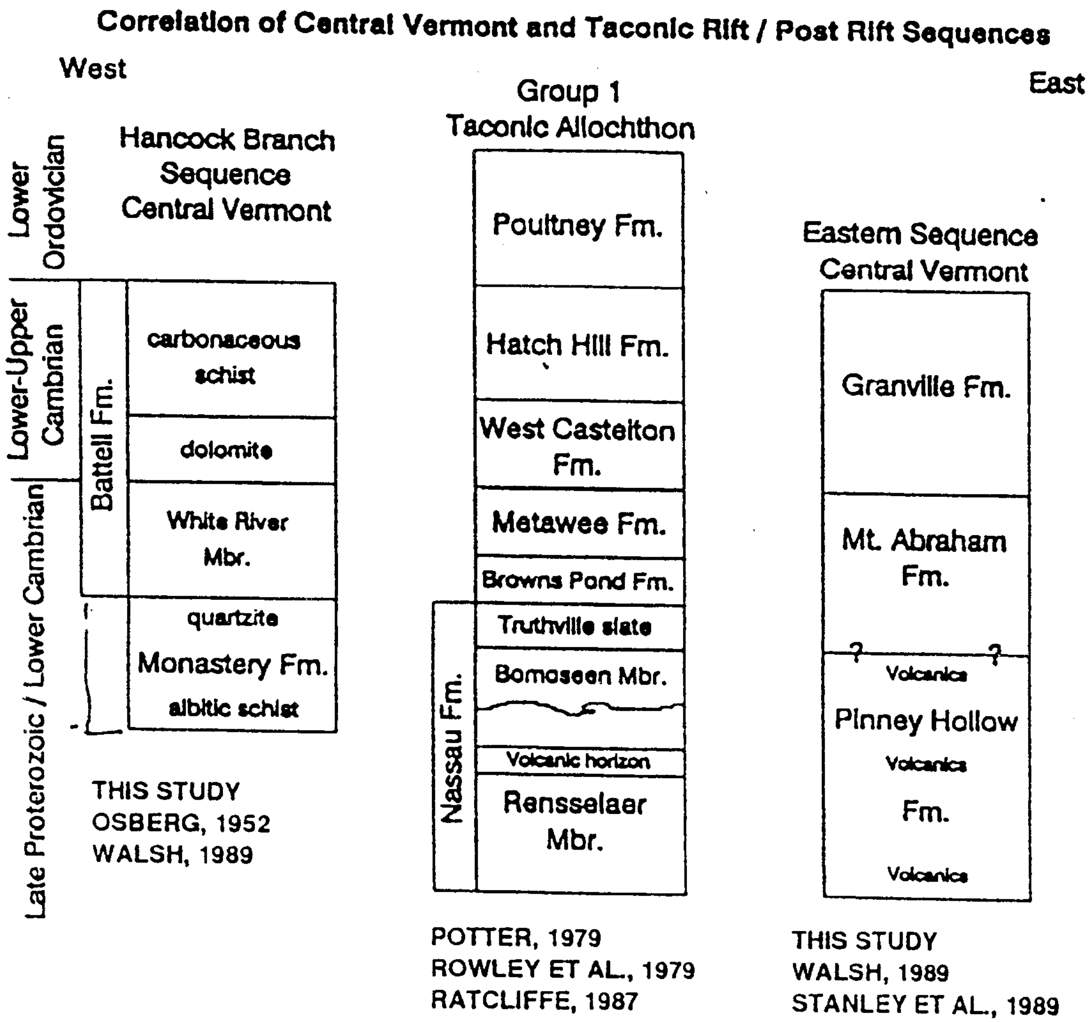


Figure 2.11 Stratigraphic correlation of central Vermont and Taconic allochthon sequences. Correlation is based upon similarity in graphitic/dolomitic sequences and underlying quartzite and aluminous phyllite/slates. Note that the Hancock Branch dolomite dictates positioning of the Taconic allochthon to its east, but west of the non carbonate-bearing Granville-Ottauquechee sections. This position generally agrees with the Stanley and Ratcliffe (1985) Taconic root zone position.

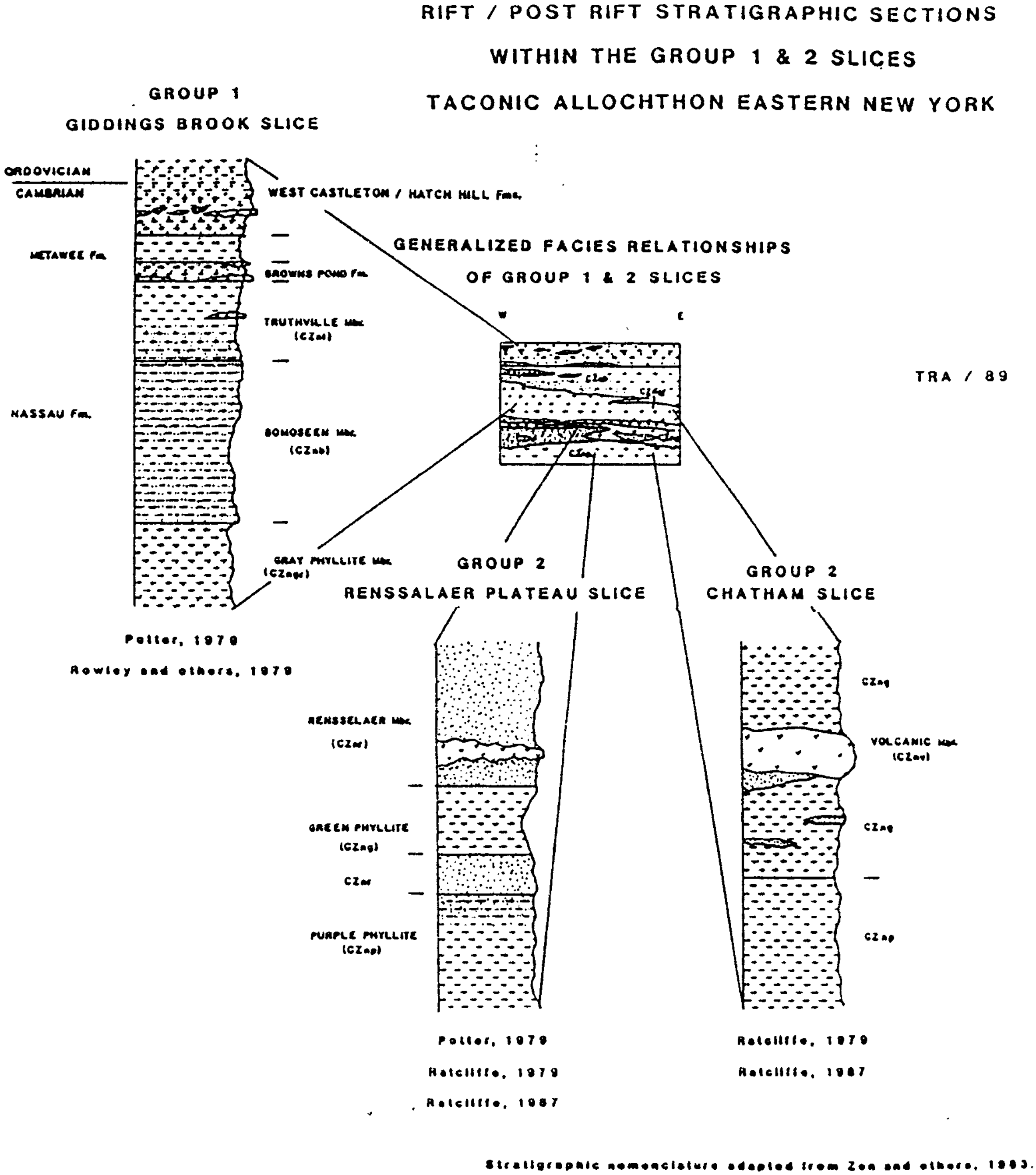


Figure 2.12 Stratigraphic columns for the Groups I and II Taconic allochthon. Post-rift dolomite occur as thin layers within the Brown's Pond and West Castleton Formations. References are listed below each column.

important to note from plate 1 that the Taconic clastics lie in their **projected** position; a northward along strike decrease in grain size or positioning of the coarse clastics **beneath** the Monastery sequence is required, in order to explain the placement of coarse rocks (Rensselaer) outboard of relatively fine grained ones (Monastery). This requirement demonstrates the complexities in attempts to correlate coarse clastic lithologies.

4. Coarse clastics of the Oak Hill Pinnacle Formation are presumed to correlate directly with the central Vermont clastic Pinnacle/Hoosac sequence, immediately overlying the Lincoln massif (figure 2.2). The rapid increase in thickness along strike to the south must be the result of increasing basin depth prior to, or during, clastic deposition (Dowling, 1987). Beach facies characterizing the northern Pinnacle, give way to fluvial and alluvial shallow water marine deposits southward, and support this interpretation (Tauvers, 1982; Dowling, 1987). The transition from Hoosac to Monastery may reflect an added deepening of the basin, and more distal, fine grained deposition along the outer margin of an alluvial fan/deltaic complex (Walker, 1978). A thin sequence of coarse arkosic conglomerates, wacke, and flaggy quartzites of the Dalton Formation, found along the western (and parts of the eastern) flank of the Green Mountain massif, represent the southern continuation of the Pinnacle/ Hoosac sequence (plate 3; fig. 3 of Stanley and Ratcliffe, 1985; Ratcliffe et al., 1988). Fine grained wacke, carbonate-bearing, nongraphitic albitic schist, and aluminous schist, mapped as Hoosac Formation along the east side of the Green Mountain massif, overlie a thin Dalton-like sequence (Tyson Formation), and directly correlate with the Monastery section (plate 3; Doll et al., 1961). Farther south, albitic schist with interlayered metavolcanics have been mapped as the allochthonous part of the Hoosac Formation (Stanley and Ratcliffe, 1985); for the most part, these rocks best represent the continuation of the eastern Vermont clastic sequence, specifically the Fayston and Pinney Hollow Formations.

III. Passive margin evolution

Following the previous discussion on correlations, summarized in plate 5, a model depicting the evolution of the Late Proterozoic/Lower Ordovician Iaptean passive margin is proposed. This model is consistent with present theories regarding the rifting and breakup of a continent with consequent development of an oceanic realm (Dewey, 1982; McKenzie, 1976: Royden et al., 1980).

Seismic reflection studies of modern day margins show essentially two types of deposits; a lower heterogeneous section without prominent bedding reflectors, lying within and above observed rift grabens, and a second, homogenous, well bedded sequence traceable over large areas. The lower sequence represents coarse and fine clastics with mafic (and minor felsic) volcanics, deposited during rift subsidence. The upper sequence temporally overlaps the upper sequence and represents deposition of fine grained sediments during late to post rift thermal subsidence (Dewey, 1982; McKenzie, 1976).

Theories depicting development of a passive margin are also two-fold; a first phase of rifting of continental crust leads to isostatic subsidence or uplift as a result of density and volume changes, and the influx of large amounts of heat from an uprising mantle. Subsidence is a consequence of low differential stretching of the mantle and crust, with small heat flow changes(LePichon and Sibuet, 1985). This initial subsidence results in rapid deposition of heterogeneous clastics and possibly, rift volcanics. In contrast, high differential stretching and high heat flow will lead to either little subsidence or even initial uplift. This phase begins to wane over time as the mantle upwelling begins to cool, and the uprising asthenosphere reequilibrates towards a steady state geotherm. This is the cause of the second phase of subsidence called thermal isostatic subsidence (McKenzie, 1976). In many rift margins, this second phase is synchronous with the breakup of the continent and initial establishment of ocean crust, resulting in a cessation of heat flow into the rifted continent. It also leads to the thermal stabilization of the hinge line separating stretched, subsided crust from unstretched, relatively unsubsided crust (McKenzie, 1976). This hinge line (established after cessation of rifting and during crustal stabilization) marks the position of the subsequent shelf edge, inboard of which develops the initial carbonate platform (Dewey, 1982). In some rift margins, the thermal subsidence mechanism does not correspond with breakup and occurs during rifting as a result of low strain rates (LePichon and Sibuet, 1985). This leads to low heat flux and consequent thermal decay of the asthenosphere, with thermal subsidence **in conjunction** with initial rift subsidence. It is therefore likely that the transition from initial rift-related to thermal-related subsidence actually represents a transition, marking decrease in the rift subsidence with a simultaneous increase in thermal subsidence. The rock record should reflect this slow transition as a gradational change in sedimentation pattern, progressively away from the unstretched margin edge ("tectonic hinge" of McKenzie, 1976) into the deeper basin where high crustal attenuation has resulted in initially large heat flow and subsequent conduction (Quinlan, 1988).

The pre-Silurian rocks of central Vermont, the Taconic allochthon, and southern Quebec,placed within their interpreted stratigraphic positions, agree well with modern day passive margin models:

1. The graphitic rocks, some demonstrably coeval with the carbonate platform sequence, represent post-rift thermal subsidence deposits. The presence of serpentinite within the

Ottauquechee Formation indicates its close association with ocean crust and is believed to represent deposition within the post-rift oceanic realm (Armstrong et al., 1988b). Ages within this sequence range from Lower Cambrian (Dunham and Dunham correlative rocks) to Lower Ordovician (Arenigian) for zone 6 graptolite-bearing shales of the Taconic Poultney Formation (Berry, 1960; Rowley et al., 1979).

2. Heterogeneous nongraphitic clastic rocks, some metavolcanic bearing and of rift origin (Coish, 1987), represent deposition during rifting of the ancient North American craton in the Upper Proterozoic to Lower Cambrian, prior to continental breakup and development of ocean crust. These deposits have an upper age constraint of Lower Cambrian from fauna within the Cheshire Quartzite (Walcott, 1888). A lower age constraint for the initiation of rift clastic sedimentation does not exist; the only valid radiometric age (554 +4/-2 Ma, discordant Pb/Pb age) was recently acquired from a felsic unit within the **upper part** of the Tibbit Hill Formation volcanics (Kumarapeli et al., 1989). Rather than indicating that rifting within southern Quebec was active until the Middle Cambrian, it probably indicates that the absolute Cambrian fossil ages are not well constrained.

3. The interlayered graphitic/nongraphitic zones, along with the aluminous slates and phyllites comprising the upper part of the clastic sequence (Fairfield Pond, Truthville/Upper Metawee, and Mt. Abraham Formations, from west to east across the basin), represent the transition from rift to post-rift subsidence mechanisms. This is supported by the change from heterogeneous to homogeneous sedimentation with a lack of rift volcanics and the initial appearance of white quartzite, similar to the basal quartzite of the platform (Cheshire Quartzite).

Comparison of the pre-Silurian section to modern day generalized passive margin sequences allows interpretation of the evolution of the Late Proterozoic to Lower Ordovician Iaptean passive margin (plate 5):

1. Late Proterozoic/Lower Cambrian rifting, with first phase extension of the upper crust by differential simple shear (following the model of Lister and others, 1986). This resulted in consequent initial subsidence with early coarse clastic deposition within isolated rift grabens. Continued subsidence led to the development of a larger continuous rift basin, with clastics overlapping graben margins, forming a continuous rift clastic sequence which become finer grained to the east, away from the western source area. This stage produced the clastic sequence of Vermont and the Taconic allochthon. Large differential stretching and attenuation of upper crust led to initiation of volcanism with deposits representing both pillow basalts (Rensselaer/Chatham) and volcanogenic water-lain deposits (Fayston, Pinney Hollow, Stowe Formations). Although shearing on the master décollment is down to the east (present day directions), graben propagation in most modern day regions initiates along the rift axis and progresses inboard into less deformed, thicker crust (Bonatti, 1987; Bosworth, 1986). According to this model, propagation and initial deposition during this phase would have occurred near the present position of the Stowe Formation, progressing westward with time.

2. Second stage rift subsidence during initiation of a second master décollment within the outboard middle and lower crust, would lead to continued initial subsidence, deposition of finer grained volcanic-devoid sediments, and development of a late rift (pre-shelf) sequence (Mt.Abraham, Bomoseen, Truthville, Monastery, Fairfield Pond, and upper parts of the Hoosac and Pinnacle Formations), thinning westward onto relatively unstretched crust. Conglomeratic horizons and coarse wacke of the Dalton, Pinnacle, and Hoosac Formations may represent earlier rift fluvial deposits prior to drowning of the migrating tectonic hinge line during subsequent onlap.

3. The interlayering of graphitic and nongraphitic lithologies heralds the transition from initial- to thermal-dominated subsidence mechanisms contemporaneous with initial continental breakup, development of the basal quartzite of the proto-carbonate platform, and the thermal equilibration of the asthenosphere with concomitant rapid thermal subsidence.

4. Deposition of homogenous anoxic mud, represented in the rock record by graphitic, sulfidic shales and schist, occurred during full-fledged thermal subsidence. The lack of thick organic-rich sections in both the Taconic and Oak Hill sections (relative to modern day margins), coupled with a Lower Cambrian to Lower Ordovician age bracket (Taconic section), indicates slow deposition and formation of a sediment starved post-rift sequence.

The presence of thick carbonate and carbonate laminated schist, within the Battell Formation, demonstrates its close association with the carbonate shelf edge. Outward, proximal to distal turbidites (or distal gravity flows; e.g. Lowe, 1982) of the West Castleton and Hatch Hill Formations represent slope-rise deposition. Eastern correlative, albeit nongraphitic, Granville and Ottauquechee Formations represent anoxic mud deposited below the calcium compensation depth with outer-rise to abyssal plain realms, respectively.

5. Thermal subsidence and thus passive margin development may have continued into the *Middle Ordovician when foredeep development and subsequent exhumation occurred* during Taconian orogenesis, producing the present unconformably overlying flysch sequence of the Indian River, Pawlet and Austin Glen Formations (Zen, 1967; Potter, 1979).

This interpretation points to the obvious need for detailed mapping and lithologic analysis in order to understand fully the stratigraphic sequences necessary for such a model. Additionally, the along and across strike heterogeneity of the rift clastics and the homogeneity of the post-rift sections show the need for linking particular clastic sections to their overlying post rift cover. Without this link, palinspastic positioning of along strike, projected clastic sequences, such as the Taconic section, can be quite erroneous and misleading, with profound consequences on attempts to constrain or check the validity of structural-based retrodeformation.

CHAPTER 3
STRUCTURAL GEOLOGY

Introduction

Osberg (1952) was the first worker within the Granville - Hancock area to conduct a structural analysis of various foliations and fold generations. He concluded that the rocks were deformed by essentially one distinct, regional folding event creating the dominant structural geometry. These folds had an axial planar schistosity in the eastern part of the area that folded compositional layering, interpreted as both a relict schistosity and primary bedding. This same foliation strongly paralled bedding in the western part of the eastern cover sequence. An earlier schistosity, as presently interpreted, paralled folded bedding in the crests of the folds. Osberg interpreted this foliation as mimetic cleavage, coeval with the regional foliation (Osberg, 1952; p.84). This type of cleavage was thought to occur during folding along planar anisotropies, such as bedding. (Billings, 1942; p. 218).

Regional folds trend quite uniformly to the north-northeast, but had plunges both shallow and steep to either the north or south. The actual fold data shown on Osberg's map, however, differ from the interpreted map pattern structures which were consistently shown plunging moderately to the south. A later set of post-metamorphic folds were interpreted as minor perturbations, superposed on the earlier regional deformation. These folds were generally observed within the eastern part of the area with trends ranging from slightly northwest to true northeast. The majority of the folds plunged less than 45°. These folds usually had an axial planar slip cleavage that in some areas was fairly penetrative, in other areas present only as a weakly developed crenulation cleavage (Osberg, 1952, p. 84-85). Only the regional fold generation had a related mineral lineation usually trending to the southeast and plunging 45° - 75° southeast. This was common to the extreme eastern part of the study area (Ottauquechee Formation) and particularly within the western part of the eastern cover sequence (Osberg, 1952; plate 2).

The major structural feature of Osberg's map was a the Green Mountain anticlinorium. It consisted of three parallel anticlines, giving rise to a major culmination situated within the core of the Middle Proterozoic basement (Osberg, 1952; p. 72-101). Although a structural feature, the culmination and three related anticlines were actually mapped on the basis of the lithologic distribution. The premise was that stratigraphy was coherent and symmetrical on both sides of the Lincoln massif basement rocks (Osberg, p. 66-69 and map) even though stratigraphic symmetry required correlation of rocks very dissimilar in composition (Osberg, 1952, p.69). This was explained as the result of abrupt, complex facies changes although the cross sectional distance between interpreted correlatives was no more than 16 km (~10 mi.). Additionally, all of these facies changes fortuitously occurred at the longitude of the Green Mountain anticlinorial crest, where the entire section was eroded such that direct observation of these facies changes was impossible.

Another problem of this stratigraphically-based structural interpretation based upon stratigraphic correlations was that minor structural data such as fold hinge orientations, bedding orientations, and foliation attitudes rarely coincided with map scale structures. Regional cross sections did not match the style of the various mesoscopic fold generations and also did not abide to the measured foliation and fold orientations. Complex lithologic distributions, including the juxtaposition of three or more major rock units, was regarded as a consequence of the complex facies changes rather than tectonic processes. The lack of knowledge concerning ductile fault contacts created rigid constraints on preserving stratigraphic continuity, both along and across strike (figure 2.1). These problems are well justified since Osberg's study was conducted prior to the advent of plate tectonic models, and relied primarily on the working theory that structural distribution of lithologies was subordinate to their depositional distribution.

Present structural analysis:

Structural analysis of the Granville - Hancock area indicates that the pre-Silurian eastern cover sequence has been deformed by several generations of folds, associated schistosity, and faults (Armstrong et al., 1988a). Superposition of deformation fabrics, interference patterns of fold generations, and dynamic recrystallization of various ages of mineral assemblages, add complexity to the structural history of this belt. Variations in fold and fault style, as well as the dominant mode of deformation, appear to be related to the variation in metamorphic grade and strain level across the belt. Correlation of various deformation fabrics is complicated by style variation and by the colinearity and coplanarity of the various generations of fabric.

The various synmetamorphic deformations will therefore be discussed in spatial domains, in which specific deformational features have relatively similar style, orientation, and developed at similar metamorphic grade. Three domains have been defined during this study and will be discussed in order; western domain, central domain, and eastern domain (plate 1). Following these descriptions an interpretation of feasible syn-metamorphic fold and fault fabric correlations across the three domains will be attempted based upon the available data. This will be followed by sections on the post or late metamorphic deformation, and the earliest observed deformation; pre/early metamorphic faults (see Appendix I for a description of deformation terminology). A concluding section will cover a brief interpretation of the overall structural evolution of the belt. This will combine all of the previous structural information as well as any pertinent metamorphic and/or stratigraphic data.

Western Domain

The western domain consists of a dominant foliation (Snw, Appendix I) deformed into a series of map pattern-scale upright to slightly west overturned, chevron-like antiforms and synforms. These folds (Fn+1) have smaller orders of coeval, asymmetrical folds along their limbs, usually with consistent senses of rotation on each limb and an axial planar synmetamorphic schistosity or crenulation cleavage (Snw+1). A later phase of deformation (Snw+2) may exist which gently warps Snw+1 near the crest of the Green Mountains (plate 1).

Snw:

The dominant schistosity of this domain is closely spaced (1mm) within all of the schistose rock-types and is defined by compositional layering and alignment of sericite, chlorite, ilmenite, and chloritoid lathes. It may also be recognized by the orientation of opaque, chloritoid, garnet, and quartz inclusions within albite porphyroblasts (figure 3.1). Within quartz-feldspar and rare mafic rock-types, Snw is defined by a penetrative, closely spaced (~2-3mm) compositional layering of quartz/feldspar (semi pelites), plagioclase (mafic), and micaceous minerals.

Minerals associated with this fabric vary according to bulk composition. Most pelitic rocks within this domain, however, are fairly aluminous and contain garnet, chlorite, white mica and possibly chloritoid. Garnet within these rock-types is generally younger and/or coeval with Snw. Relative age relationships are based upon the fact that garnets contain inclusions of quartz,chloritoid, plagioclase, ilmenite, rutile, and graphite that are generally planar with a few exceptions that show early Snw+1 crenulate folds. Garnets commonly show that the included Snw fabric is continuous with the external fabric, except in areas where a subsequent synmetamorphic fabric (Snw+1) is well developed and associated with mylonitic shear zones. In these areas Snw within garnets is truncated at the garnet edge and transposed into Snw+1 chlorite grade matrix (figure 3.2). These relative age relationships suggest that Snw development is therefore **roughly** synchronous with garnet grade metamorphism, regionally pervasive throughout this longitude within central

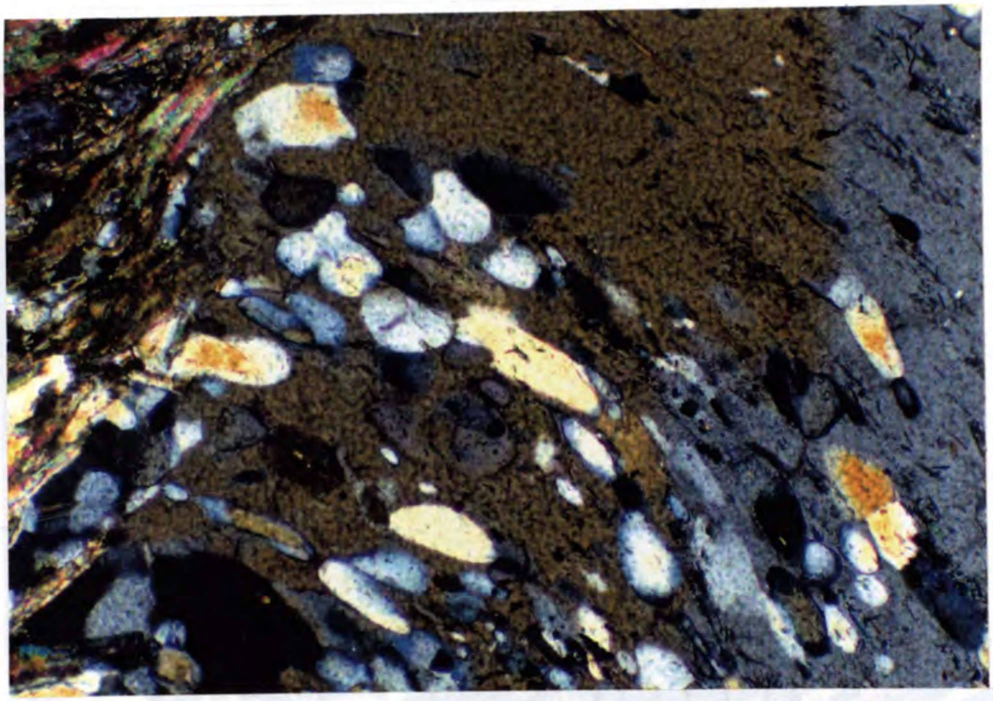


Figure 3.1 Photomicrograph of albitic schist, Pinney Hollow Formation, showing Snw foliation as "helicitic" inclusion trails defined by inclusions of quartz, ilmenite, and rutile within post-Snw albite porphyroblasts. This foliation actually merges with Snw+1 matrix foliation and represents preserved Fnw+1 crenulation fold hinges within the porphyroblasts. Matrix crenulations are transposed by continued Snw+1 deformation. Sample from Texas Brook area; plate 1, Y15. Field of view is 2.4 mm

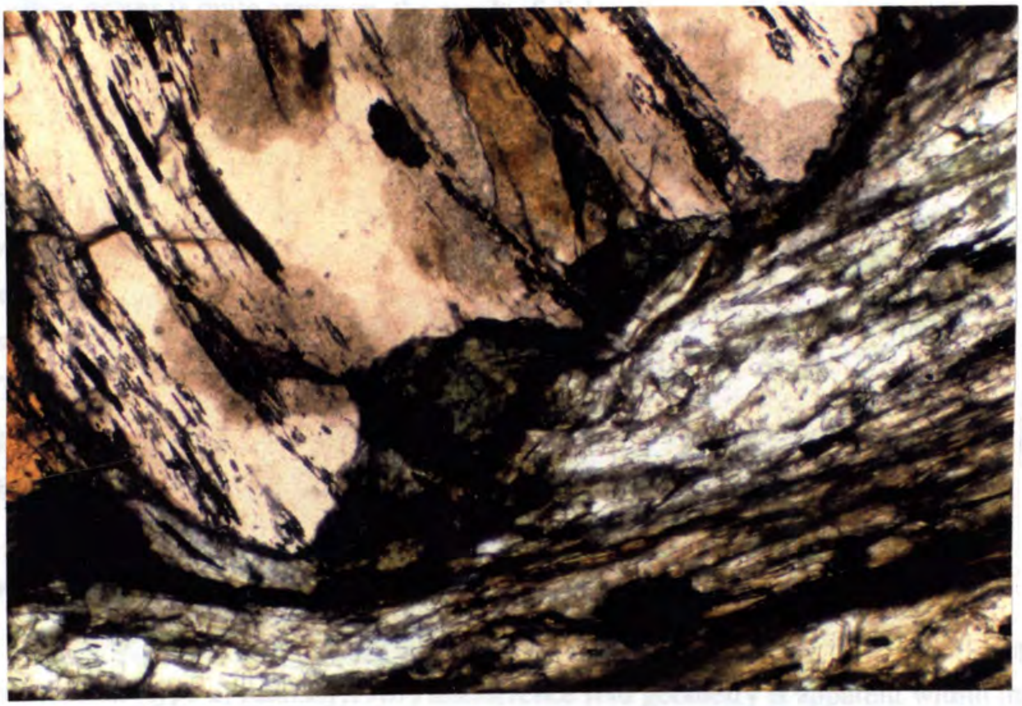


Figure 3.2 Photomicrograph of syn- to post-Snw (M1) garnet and matrix grains within chlorite-sericite-quartz-rich matrix. Matrix minerals are aligned within Snw+1 foliation and truncate along garnet edge. Notice Snw inclusion trails within garnet truncate along its resorbed edge where younger, Snw+1 age of chlorite is growing. These relationships indicate that garnet growth predates Snw+1 and postdates Snw. Sample from the Monastery Formation, HQ 521; plate 1, location P 9. Field of view is 2.4mm

Vermont (see Tauvers, 1982; DiPietro, 1983; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Haydock, 1988; Prewitt, 1989; Walsh, 1989).

Microstructural features within Snw schistosity are quite simple, lacking the pervasive asymmetry so common within well documented ductile shear zones (White, 1976). Snw-age quartz grains show polygon geometry with 120 triple junctions, a common texture within recovered quartz grains. Undulose extinction within both quartz and albite grains is quite common, the result of dislocation creep during relatively ductile, medium metamorphic grade conditions (Porier, 1980; Schmid, 1984a, 1984b). Subgrain development, grain size reduction, and boundary migration were not recognized. Their absence is most likely a consequence of pervasive post recrystallization recovery of deformed grains (Porier, 1980; Schmid, 1984a). Because of the pervasive folding by Snw+1/Fnw+1 crenulation cleavage, Snw is variably oriented but on the 1:12,000 map-scale it is generally north striking and shallowly dipping east in the extreme western part of the study area, and moderately east dipping in the eastern part of the western domain (plate 1). The general, broad lithologic distribution of the major formations within this region suggests that Snw was subhorizontal across the western domain prior to subsequent Fn+1 upright folding (plates 1 & 2).

Fnw:

Few mesoscopic Fnw folds, associated with Snw, were found during this study. Map pattern coaxial (type 2; Ramsey,1967) interference fold geometry is apparent within the large dolomite of the Battell Formation (Cbd; plate 1, K8). This pattern suggests a folded axial plane and fold hinge (Fn?), coaxially refolded by later Fn+1. Further map pattern type 2 geometry is not demonstrable within this domain, but it is suggested by the map pattern contact between Mt. Abraham and Monastery/Battell units (plate 1). The lack of Fn geometry may be the result of transposition of early folds into the Fn+1 structure, especially within the less competent pelitic rock-types.

Lnw:

Due to the highly (and tightly) folded nature of Snw, Lnw mineral were neither observed nor measured in this domain, although they have been reported to exist in other areas (DiPietro, 1983; DelloRusso, 1986).

Snw+1:

Superposed over the dominant Snw foliation is a younger non penetrative, wavy, widely spaced crenulation cleavage or schistosity, Snw+1. This foliation is axial planar to a phase of upright to slightly westward overturned chevron folds, Fnw+1. Snw+1 is defined by aligned chlorite and sericite lathes, rare chloritoid lathes, and several oriented elongate grains of quartz and albite. Snw+1 foliation appears to be coeval with **retrograde** metamorphism of the earlier Snw garnet grade mineral growth to biotite/chlorite grade assemblages since syn-Snw garnet is consistently altered to chlorite, which, along with a second phase of white mica, is commonly aligned within Snw+1 (chapter 4). No garnet porphyroblasts were observed overgrowing Snw+1 fabric, although this relationship is seen farther south in the Rochester area where garnet is related to Acadian prograde metamorphism (Jo Laird, pers. comm., 1990, 1991).

In the western part of the western domain, Snw+1 is generally a non penetrative, wavy to anastomosing, crenulation or spaced cleavage defined by sericite and chlorite lathes. Orientations of this foliation are fairly consistent, with a point maxima defined plane of N08E, 49SE (plate 6). This foliation is axial planar to the dominant set of folds, Fnw+1, which tightly deform the Snw compositional layering and foliation. Thin sections of pelitic units commonly contain Snw-age garnet with severe retrograde chlorite overgrowths in areas of well developed Snw+1 (figures 3.3 & 3.4).

Within the eastern part of the western domain (plate 1), Snw is pervasively deformed by Fn+1 folds and associated axial planar foliation (Snw+1). In many localities, Snw+1 is actually the dominant foliation, folding relict Snw. Snw+1 is always planar, closely spaced, and of a fairly consistent orientation (plate 6). Within the Snw+1 microlithons, Snw foliation can be readily recognized as a compositional layering (figure 3.5). Snw appears to be completely transposed into the plane of Snw+1 foliation along the limbs of Fnw+1 folds. Within crests of Fnw+1 folds, Snw is oblique to the axial planar Snw+1, and parallels the folded compositional layering or bedding. Thin sections of samples from this region show that the earlier Snw foliation either wraps around, overgrows, or truncates a garnet bearing assemblage in several different lithologies (figures 3.2, 3.3, and 3.4). This high grade assemblage is almost completely retrogressed by the younger Snw+1 foliation, defined by aligned chlorite and sericite, with recrystallized quartz and abundant albite porphyroblasts. Relict Snw-alligned grains of sericite, chlorite, and occasional chloritoid commonly show undulose extinction in areas where grains have been kinked. These kinks are quite common in the crest of small scale Fn+1 folds.

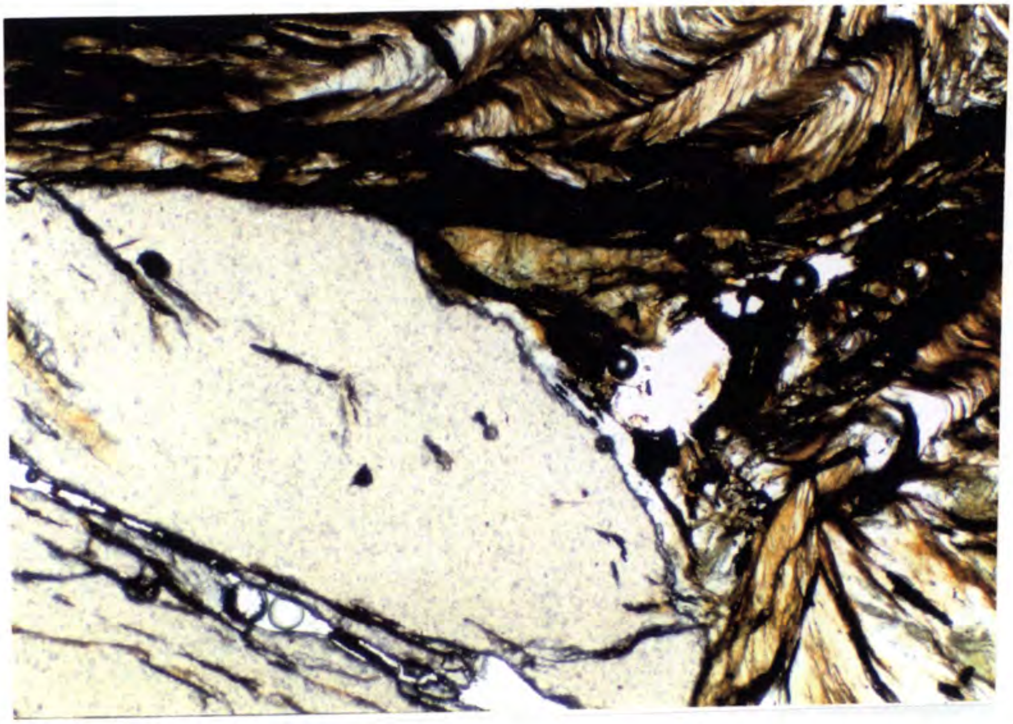
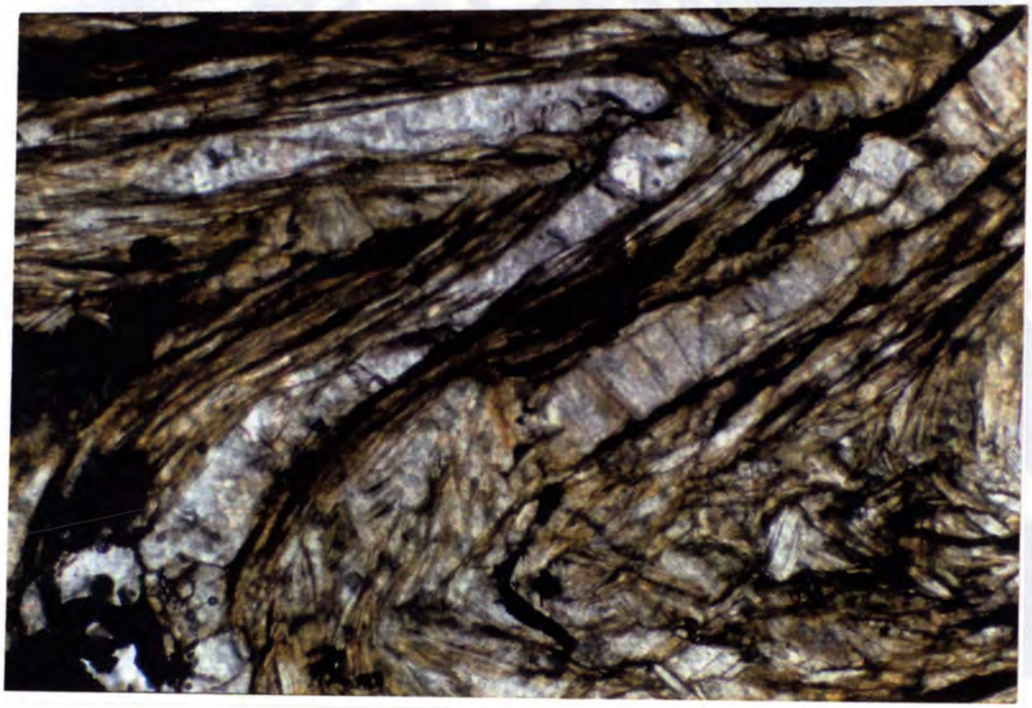


Figure 3.3a Photomicrograph of M1 (pre- to syn-Snw) garnet within pervasive, fault-zone Snw+1 schistosity composed of aligned chlorite, sericite, and ilmenite grains. Notice that the garnet does not contain any Snw+1 inclusions and that the matrix foliation forms a chlorite-rich tail on along the resorbed garnet edge. Fnw+1 crenulate folds are deflected around the periphery of the garnet; these relative age relationships demonstrate that garnet growth is pre-Snw+1 foliation and Fnw+1 fold development. Sample from Monastery Formation, Hancock Branch of White River; plate 1, T10.



3.3b (located 1.75mm below view in 3.3a) A small quartz veinlet, deformed into a tight Fnw+1 reclined fold with a steep southeast plunging hingeline, very similar in style to the Sn deformation farther east; this Snw+1 fault zone records a much larger magnitude of strain than the surrounding rocks which display a weak to moderate Snw+1 crenulation cleavage and upright Fnw+1 crenulate folds. view is 2.4mm in both pictures.

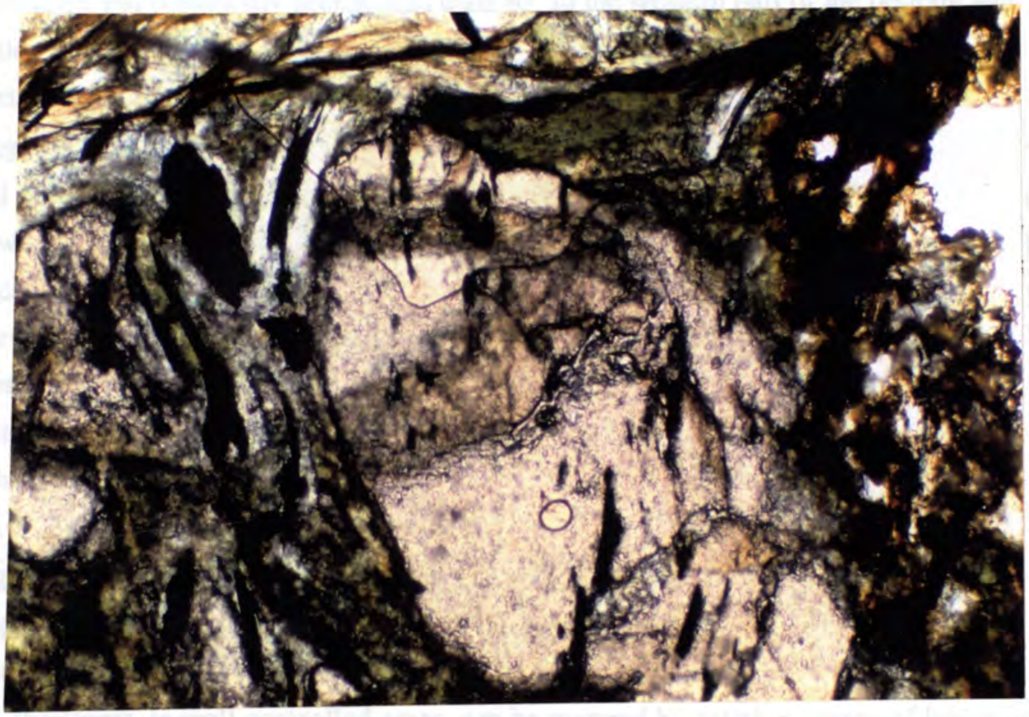


Figure 3.4 Photomicrograph of resorbed Snw-age garnet with Snw+1-age chlorite overgrowth. These M2 retrograde overgrowths are especially well developed in areas of pervasive Snw+1 fabric. Snw matrix fabric, usually defined by white mica and ilmenite lathes is completely transposed into Snw+1. Occasional Snw ilmenite can be seen bent into Snw+1 foliation. Sample from Mt. Abraham Schist; plate 1, J10, immediately west (high T side) of garnet isograd. View is 2.4mm.

Fnw+1:

Folds associated with the Snw+1 foliation have similarly oriented axial planes, coplanar with the pervasive Snw+1 (N08E, 46SE; plate 6). Fnw+1 fold hinges vary substantially in orientation, but commonly plunge less than 40° in the western part of the domain and greater than 40° in the eastern part (plate 6). The map pattern geometry and lithologic distribution within the western domain is primarily the result of a pervasive phase of upright to slightly westward overturned, isoclinal to chevron folds, Fnw+1 (figure 3.6). Fold geometry is recognized by folded compositional layering and/or dominant schistosity (Snw). An axial plane schistosity or slip cleavage (Snw+1 foliation) is defined by aligned sericite and chlorite. Although axial planes to these folds are quite consistent in orientation, Fnw+1 hinges are variably oriented, plunging anywhere from 5° to 85° (plate 6). Plunges are most commonly to the southeast, but northeasterly plunges are not rare. Within the westernmost part of this domain, hinges typically plunge from 10° to 35°; in the eastern part of this domain, hinges rarely plunge less than 35, achieve a maximum plunge of 85°, and an average plunge of ~60°. Within and in close proximity to Snw+1 shear zones, Fnw+1 hinges typically plunge steeper than surrounding folds away from the shear zone. Fnw+1 hinges within the shear zones are commonly sub parallel to the well defined stretching mineral lineation, oriented S70E and plunging at ~65° - 70° (plate 6). Large scale Fnw+1 fold structures, in well controlled areas, can be mapped by rotation senses of lower order parasitic folds. Some of these large scale structures are doubly plunging while others plunge exclusively to the southeast (plate 1; I10, I11, J16, J17).

Fnw+1 folds are almost entirely transposed by continued development of Snw+1 within the easternmost part of this domain. Where found, these folds deform the preexisting Snw foliation into tight isoclinal folds with southeasterly inclined to reclined hinges. These folds appear identical in style and orientation to folds mapped as Fn folds further to the east. The main difference between the two is that the Fn folds deform either transposed bedding or a rarely preserved relict schistosity (Sn-1) rather than a dominant schistosity (hence the relative fabric assignment Fn vs Fn+1; see Appendix I).

Nw+1 lineations:

Three types of lineations are present in this domain; a crenulation lineation, defined by a gentle crinkling of the Snw foliation surface, is infrequently present in all pelitic units. This lineation is the result of planar intersection of Snw with later Snw+1, and thus defines the orientation of Fnw+1 hinges; trend and plunge orientations of this lineation are similar

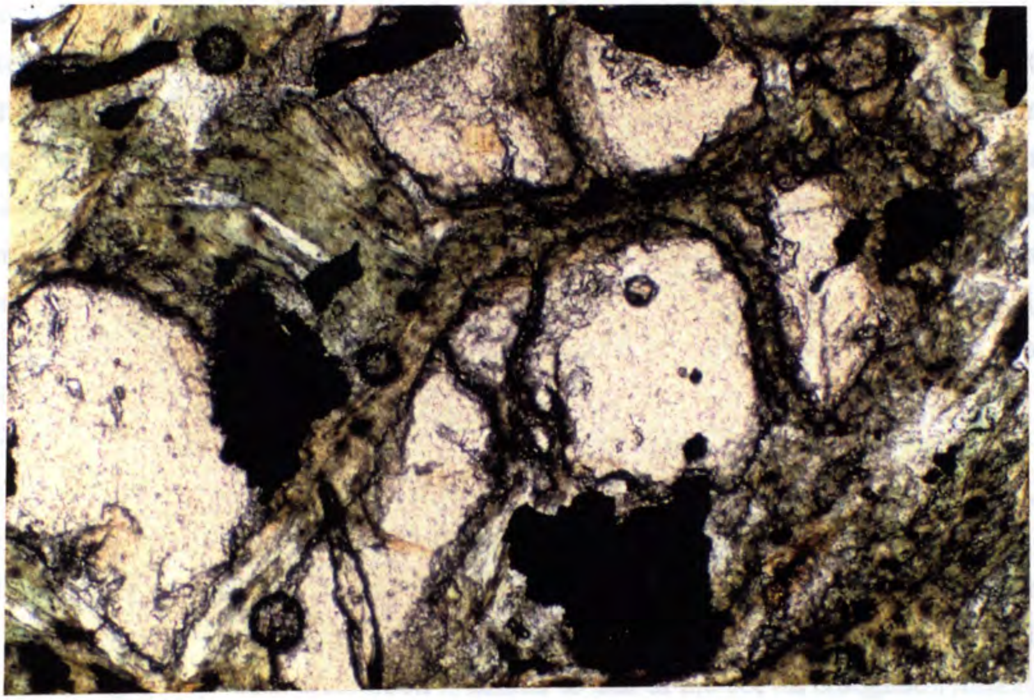


Figure 3.5 Photomicrograph showing relict foliation (Snw) within Snw+1 age chlorite that is replacing older Snw garnet. This section is from outcrop immediately along the garnet isograd. Note that M1 (high grade metamorphism) garnet has been almost completely replaced by chlorite pseudomorph during M2 retrograde metamorphism (chapter 4). Only a few small relict garnet fragments are preserved. Sample from the Mt. Abraham Schist; plate 1, F11. View is 1.2mm across.

to those of Fnw+1 fold hinges (plate 6). In areas where Snw+1 is well developed the crenulation is also more pervasive. The second lineation type is a strong mineral lineation, found along Snw+1 foliation planes. It is defined by aligned lathes of chlorite and sericite as well as rare biotite in some rock-types (figure 3.7). Recrystallized quartz is commonly found overgrowing chloritoid, garnet, and magnetite porphyroblasts within Snw+1 shear zones, and less commonly associated with the Snw+1 axial planar schistosity or crenulation schistosity. The asymmetrical tails are oriented with the long axis of the tails parallel to the mineral lineation. Within all 28 oriented thin sections from this domain, east-over-west tail asymmetry was observed, indicating that the mineral lineation defines the elongation direction in a east-over-west rotational strain state (Ramsey and Huber, 1983; Hobbs, et al., 1976; Nicholas and Poirier, 1976). Mineral lineations measured across this domain have a fairly consistent orientation, with a point maxima oriented S74E, 50° (plate 6). The third type of Nw+1 lineation is quartz rodding. This type of lineation may be the result of either new quartz growth or recrystallization of older quartz veins within an elongation direction (simple shear) or the least principle stress orientation (stretching direction for pure shear; e.g. Means, 1977; White et al., 1980; Bell, 1981; Law et al., 1984). Western domain quartz rod orientations show variation in orientation along a great circle roughly defining the Snw+1 point maxima plane (plate 6). The point maxima of measured quartz rods is defined by a tight clustering of the majority of rods, oriented S81E, 47° (plate 6). This orientation is similar to that of the coeval elongation mineral lineations (plate 6).

Snw+1 kinematic indicators:

Asymmetric tails on albite porphyroblasts, preferred orientations of quartz grains oblique to the Snw+1 foliation, and quartz pressure shadows on opaque porphyroblasts are common kinematic indicators related to Nw+1 deformation within several of the shear zones coplanar to Snw+1 (plate 1; T10, U11, X20). Rocks associated with the lower strain Snw+1 axial planar schistosity do not contain abundant asymmetrical tails or shear fabric, although the mineral lineation is still well developed and colinear to the shear zone mineral lineation. The infrequent asymmetrical tails seen on garnet, plagioclase, and opaque porphyroblasts does indicate that some component of rotational strain at least locally developed. Field observations indicate consistent east-over-west sense of rotation is also defined by displacement of mm-scale compositional layering regardless which Fnw+1 fold limb it is developed upon (Walsh, 1989). No consistent west over east kinematic indicators were found within Snw+1, even where sampled from west over east parasitic

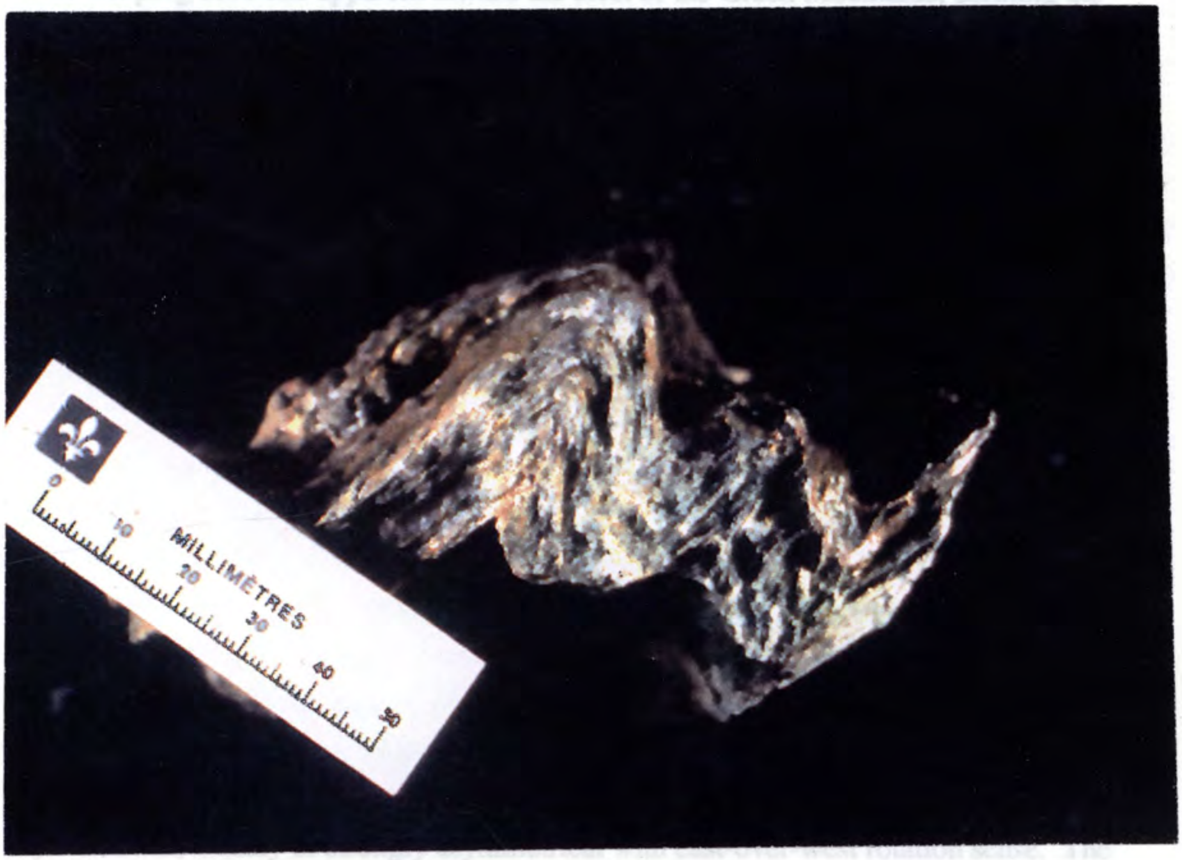


Figure 3.6 Hand sample of Monastery Formation schist showing dominant schistosity of the western domain (Snw) deformed into tight chevron folds (Fnw+1) with spaced axial plane schistosity (Snw+1). This style of fold occurs in many different orders of magnitude within the western domain, and is responsible for development of the Green Mountain anticlinorium at this latitude. Sample from Hancock Branch, White River; plate 1, V13.

Fnw+1 folds.

Post metamorphic deformation:

Deformation following Snw+1 development is locally present as a gentle warping of Sn. This warping becomes apparent toward the crest of the Green Mountains, defining the western border of the field area (plate 1). No fabric is associated with this warping, but it may be related to late deformation recognized to the north in the Fayston area by Walsh (1989), where it occurs as a pressure solution cleavage and associated open folds. This phase appears to be exclusively related to a zone around the axis of the Green Mountain anticlinorium (Cady et al., 1962; Walsh, 1989).

Central Domain

Snc:

The dominant foliation within this region is closely spaced (1mm), planar, and of fairly consistent orientation throughout the domain (point maxima orientation = N03E, 65SE; plate 6). Within the pelitic units, Snc is defined by quartz and phyllosilicate segregations. Quartz occurs as 0.2 to 2.0mm recrystallized grains while 1 to 2mm lathes of chlorite and sericite are pervasively aligned within the plane of foliation. Garnet (spessartine-rich) is only present at only one locality (Allbee Brook Fault Zone). Albite porphyroblasts (0.5 to 3mm) are either deformed within the foliation or statically overgrow it. Opaque porphyroblasts commonly have recrystallized quartz pressure shadows that can be symmetrical, weakly or strongly asymmetrical with east-over-west rotation sense. The associated down-dip Sn lineation (Ln) is defined by aligned lathes of chlorite and sericite, tourmaline, and quartz pressure shadows on magnetite grains.

A relict, earlier foliation (Snc-1) is commonly found within Snc microlithons and preserved within hinges of Fnc folds. Thin sections of schist from this region shows that the earlier foliation (Snc-1) contains chlorite clots, which may be pseudomorphs after garnet, similar to those seen in the western domain, but thoroughly resorbed within this region. Chlorite is similar in color, birefringence, and texture to Snc-age chlorite. No relict, pre-Snc garnet was found from this zone although some garnet inclusions have been observed within plagioclase porphyroblasts from the Northfield Mountains, immediately northeast of the study area (Kraus, 1989). These inclusions were most likely protected from resorbtion by the enveloping plagioclase during a syn- and\or post garnet-growth metamorphism.

Relict Snc-1 and rare Fnc-1, were observed within greenstones in many localities; compositional layering of epidote-rich layers, 1 to 5mm thick, and chlorite, carbonate, and albite-rich layers are locally folded into tight isoclinal folds with a secondary axial planar foliation (Stanley et al., 1987b; p. 329-331). The majority of Snc-1 appears to have been transposed into the plane of Snc. Thin sections show rare 0.1 to 0.3 mm anhedral cores of blue green hornblende overgrown by chlorite and small amounts of actinolite. Overgrowths are aligned within Snc, although they are relatively strain-free (Laird and Albee, 1981). This relationship suggests chlorite/actinolite growth during Snc development. Resorbed hornblende grew prior to Snc, probably during Snc-1 deformation.

Asymmetric grains or fabric are lacking within the mafic rock-types in this domain. Near the gradational contacts with the metapelites, greenstones display a more penetrative Snc foliation,evidenced by increase in abundance of aligned chlorite and sericite grains as well as pervasive quartz veins. Concomitant development of a pervasive chlorite, epidote mineral lineation, coaxial to the mineral lineation within the metapelites occurs within this gradational zone, usually marked by a high strain, closely spaced Snc schistosity.

Fnc folds:

These folds are temporally associated with an early phase of Snc foliation development. Snc is consistently observed in an axial planar orientation. Fnc fold hinges vary in orientation although axial planes are very coplanar. The point maximum for Fnc fold hinge data defines a trend and plunge coaxial with the domain quartz rod lineation (S49E, 50° for folds, S50E, 49° for lineation; plate 6). These fold hinges and associated axial surfaces define a reclined, isoclinal fold geometry. The colinearity of fold hinges and elongation lineation are also the major requirement for the definition of sheath fold geometry (Cobbold and Quinquis, 1980).

Fnc hinge orientations do display substantial variation within this domain. Variation, however, is independent of location within the domain, and appears to be primarily related to the lithologic type and the local magnitude of strain. Fnc folds within the metapelites are consistently elongate, and isoclinal in geometry. Fold hinges cluster around the elongation direction (plate 6) and describe reclined or sheath fold geometry (figure 3.8). Snc within the schistose rocks is very highly developed and closely spaced, displaying thoroughly recrystallized (mylonitic) texture. Fnc folds within the less micaceous (and more competent) greenstones have widely dispersed hinges, most of which are highly oblique to the elongation mineral lineation (figure 3.9).

Figure 3.7 Snw+1 mineral lineation (east-west trending) defined by aligned lathes of chlorite and sericite as well as aligned, acicular crystals of tourmaline. This type of lineation, associated with asymmetrical quartz tails on garnet and plagioclase porphyroblasts, records an incremental elongation direction within a bulk rotational strain environment (chapter 3). Sample from Battell Formation graphitic schist (Cb), Hancock Branch, White River (plate 1, V14).

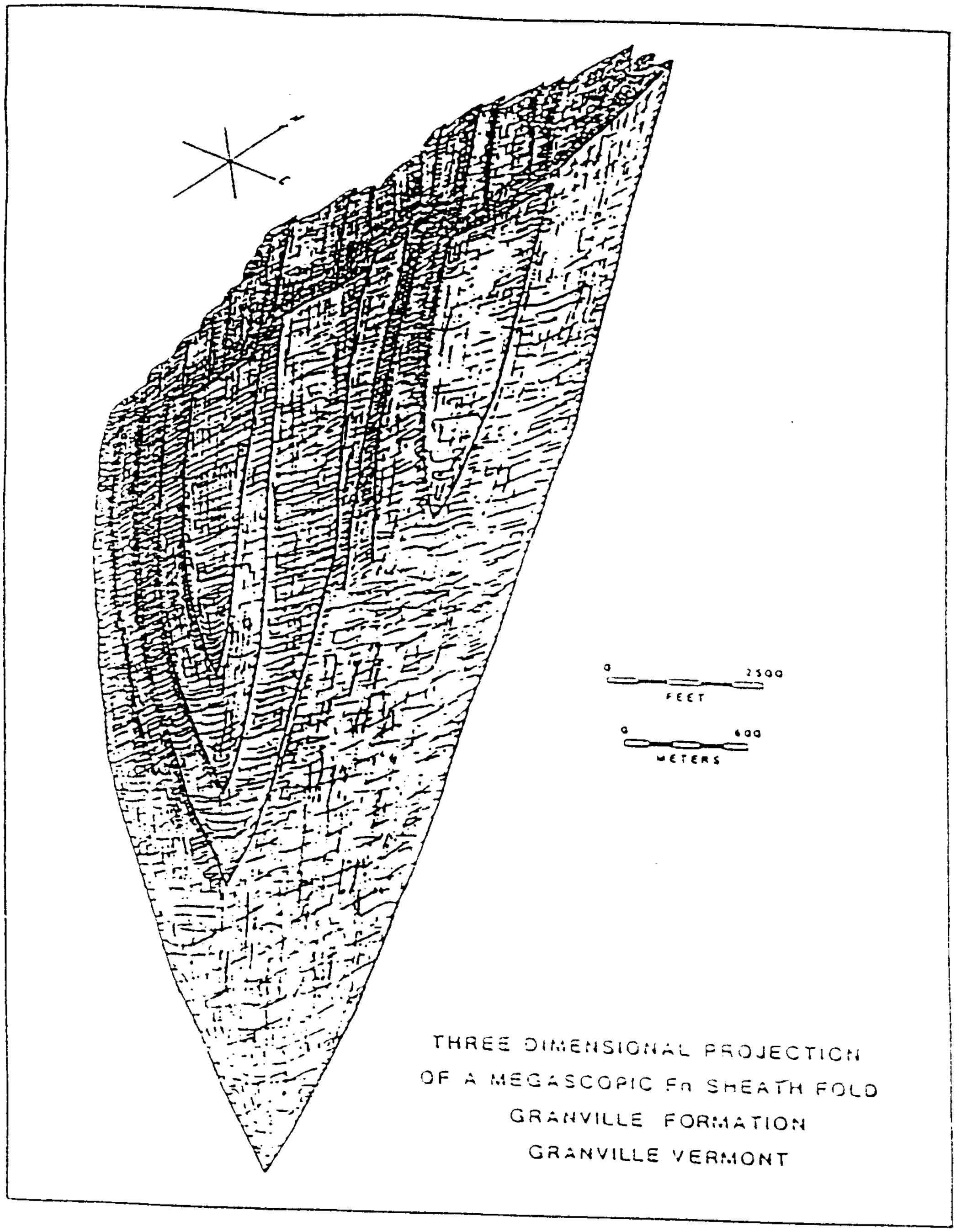


Figure 3.8a Diagram of a map-scale Fnc sheath fold, the surface of which is the contact between Pinney Hollow and Granville formations (cored by Granville). Parasitic folds have axes which plunge parallel to the Lnc mineral lineation (plate 6). Sheath folds may have developed during physical rotation of preexisting Fn-1 or early Fn folds during either a progressive rotational strain (Cobbold and Quinquis, 1983) or bulk flattening strain (Ramsay, 1970). This fold occurs within the proximity of the town of Granville (plate 1; M28-A30).

Nc lineations:

Mineral lineations and quartz rodding are the two most common lineations present within this domain. Intersection lineations are uncommon, due to the lack of a thoroughgoing secondary foliation, superposed upon Snc. The penetrative mineral lineation on Snc foliation within this zone is identical to that one described for the western domain. It is defined by strongly aligned chlorite and sericite grains, as well as aligned tails of quartz pressure shadows associated with magnetite porphyroblasts. Some tails, when viewed in the plane perpendicular to the lineation and foliation, display weak to strongly developed east over west asymmetry . Many of the recrystallized tails, however, do not have asymmetry and appear to be quite flattened (figure 3.10a). Asymmetrical tails are usually best developed within Snc-age mylonitic fault zones (figure 3.10b). The east-over-west rotation sense is consistent with rare quartz grain asymmetry relative to the Snc foliation, and to the asymmetry produced by the rotation of Snc foliation (S-surface of Lister and Snoke, 1984) into shear bands (C-surfaces) defined by dynamically recrystallized grains of quartz and new grains of chlorite and sericite (figure 3.11). The point maximum orientation of measured mineral lineations in this domain is S78E,66 (plate 6).

Quartz rod lineation:

Quartz rods are found within all of the pelitic units of the central domain. They are most common in zones of pervasive Snc foliation development, particularly in zones of mylonitic fabric development. Quartz rods, as described for the western domain, either form as recrystallized veins parallel to the elongation direction or as passively rotated Fnc fold hinges, recrystallized in an incremental elongation orientation. In contrast to the western domain, central domain quartz rods are very abundant with very large aspect ratios (20:1 for central domain, 5-10:1 in western domain). Quartz rod orientations are very similar within this domain (S50E, 49 ; plate 6). Very little fanning of data is apparent, even though the majority of all quartz rods are demonstrable Fnc fold hinges. Quartz rods are either absent or rare within the greenstone bodies of this domain. Where present, they are highly arcuate and at a large angle to the orientation of the elongation mineral lineation within the pelitic units. The lack of down dip oriented, abundant quartz rods is probably due to the general lack of high strain Snc fabric within the mafic rocks as well as a general lack of quartz, either as a matrix constituent or as syntectonic quartz veins. The arcuate, inclined nature of quartz veins that are present further demonstrates the lack of a strong

COMPARATIVE ANALYSIS OF

F2e & F2w

GRANVILLE - HANCOCK AREA

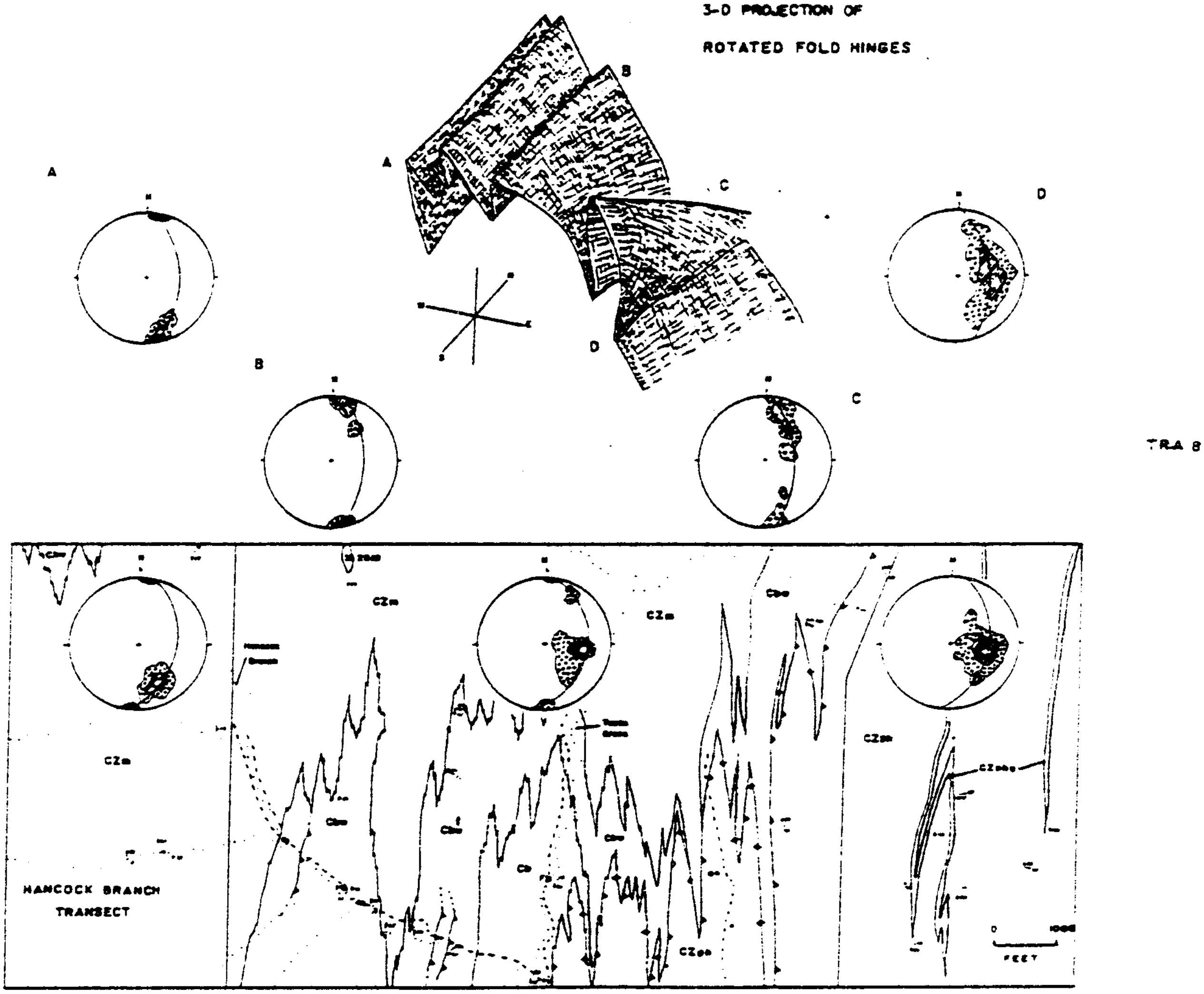


Figure 3.8b Diagram showing the progressive evolution of Fn+1 sheath folds in the western domain (see inset map), with progressive rotation of subhorizontal fold axes (western part of domain) into the n+1 elongation direction (eastern part of domain) as shown by the contoured plots of fold axes on the equal area nets. The upper four equal area nets show the predicted rotation of hinges for the model folded surface (upper part of diagram); the lower three nets show actual Fn+1 hingelines from the Texas Brook traverse across the western domain, wit progressive rotation that is in agreement with the model. Note that Fn+1 folds in the western domain are correlative with Fn folds in the central and eastern domains (see chapter 3 for further discussion).

Figure 3.9 Outcrop showing a westward overturned, inclined Fn fold within a greenstone of the Pinney Hollow Formation. Fn deforms Snc-1 compositional layering consisting of an epidote-amphibolite facies mineral assemblage. Folds within the more mechanically competent greenstones (relative to schists) are not reclined due to their effective resistance to rotation into the elongation direction during either a bulk rotational or flattening strain event. Outcrop along Forestry road 55, 25 yds. from the Town of Granville (plate 1; G29, thin section specimen HQ 265).

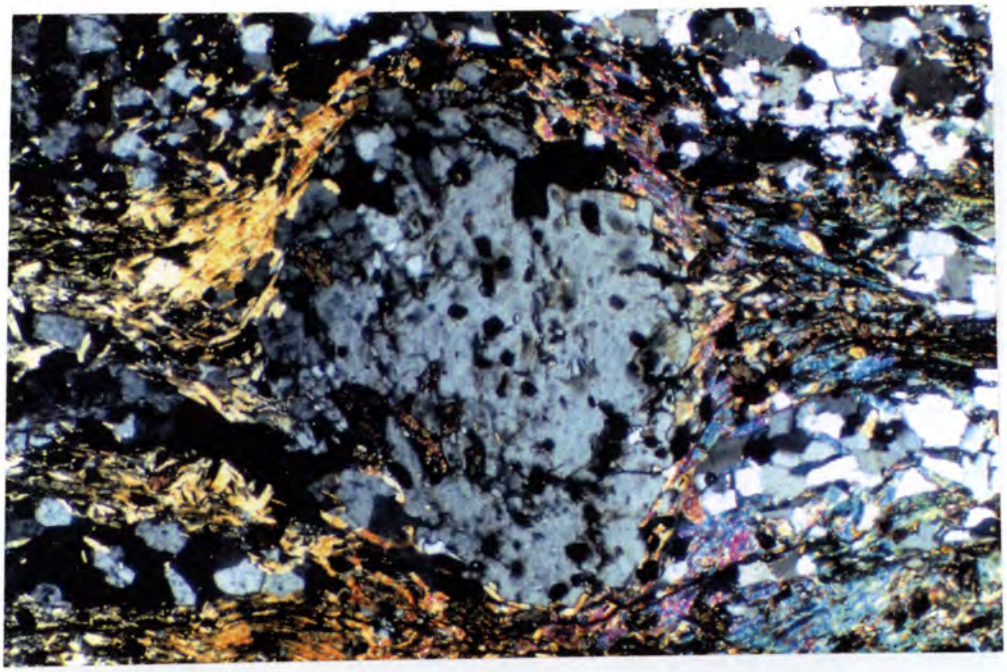


Figure 3.10a Photomicrograph of oriented sample (looking north) of mylonitic schist, Pinney Hollow Formation, within the Granville Thrust Zone, at Allbee Brook (figure 3.12). Dynamically recrystallized tails on plagioclase porphyroblasts are quite symmetrical, suggesting that Snc foliation development may have had a significant component of flattening; notice that other tails appear to be asymmetrical with east-over-west rotation sense (3.10b, next page), implying that both flattening and rotation were occurring synchronously or, more likely, as separate parts of a strain continuum. Allbbe Brook locality L 30, sample HQ 256-6x. View is 2.4mm

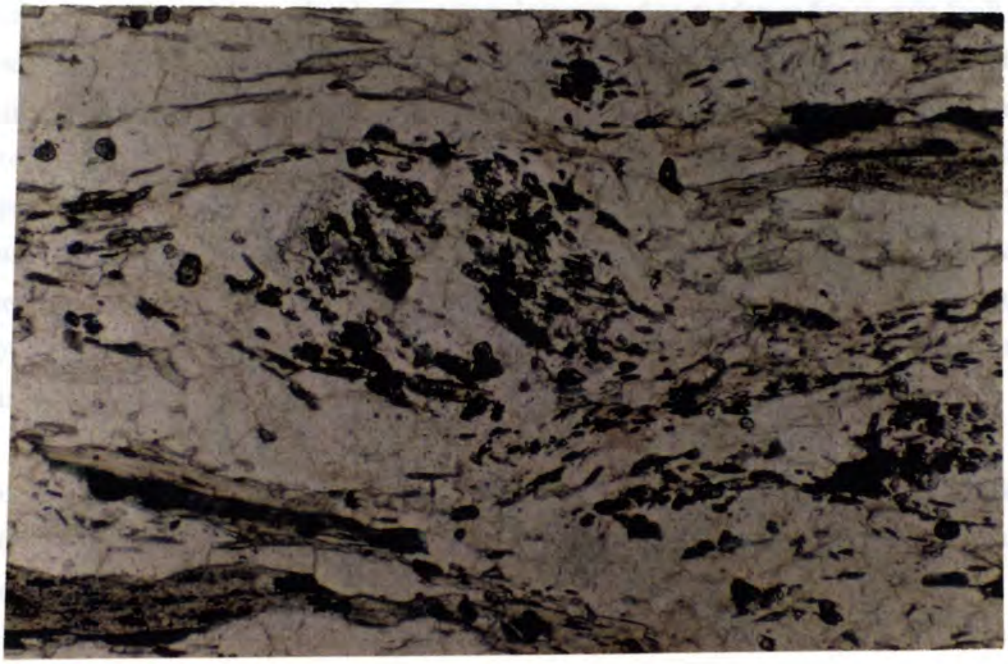


Figure 3.10b Photomicrograph of albite porphyroclast with Sn (?) inclusion trails that define a C-S relationship with east-over-west (sinistral) rotation sense. The "S" surface inclusion trails anastomose into the "C" surface inclusion trails that are incorporated within asymmetric tails on the porphyroclast. These tails are at a low angle to the Sn mylonitic foliation which is seen in the top of this view and in figure 3.10a (associated with flattening). These fabric relationships suggest that the "C-S" asymmetric fabric was preserved by the porphyroclast which was subsequently deformed by later flattening, and that an early phase of Sn simple shear was followed by a later phase of Sn pure shear. View is 2.4mm across; sample HQ 256-6 x, about 2.2 mm structurally above 3.10a. Sample oriented with west to the left and east to the right hand side of photograph.

strain within these competent rocks. These rods were not included within the central domain net data of plate 6.

Snc mylonitic fault zones:

Within the central domain, several discreet zones of intensely developed mylonitic Snc exist. Along these zones of high strain, lithologic units can be mapped terminating abruptly although in areas of lower strain these very same units are regionally continuous. The combination of intense mylonitic fabric and the abrupt termination (truncation) of otherwise continuous stratigraphic marker horizons are interpreted as evidence for ductile fault zones. These zones are best displayed within the Pinney Hollow Formation. Thin, but continuous greenstones within this formation grade into the adjacent schist. In areas of high strain mylonitic fabric, however, these greenstones also develop a pervasive fabric along their periphery, and abruptly terminate along the mylonitic fabric within the schist unit. This feature is best illustrated at the Allbee Brook locality where abundant outcrop provides control over the lithologic distribution and structural geometry (figure 3.12). Although the following discussion concerns only one locality the deformation mechanisms and kinematic indicators are common to all Sn fault zones throughout the study area.

At the Allbee Brook bridge exposure (location E), greenstone is found to the west of, and under the bridge. The western exposure is quite homogenous and over 0.6 km in width; the eastern greenstone exposure is highly foliated, compositionally segregated, and only 2m thick. These mafic exposures are separated by a schist belt, ~2m in width, displaying a pervasive mylonitic fabric (see section HQ256-6; modal analyses). The eastern greenstone is bound on its eastern side by a thick (1.5 km) belt of schist. The mylonitic schist belt varies in thickness along strike within the detailed map area (figure 3.12; A vs B). In some areas the mylonitized schist actually appears to be in gradational contact with the western greenstone, as evidenced by intermediate mafic/pelitic rock compositions along the schist/greenstone interface.

Fnc folds within the schist are rare and where present as folded quartz veins they are rootless, with limbs truncated along the mylonitic axial planar Snc foliation. The axial planar orientation of Snc in these zones suggests that it is associated with fold development; the fact that it also is responsible for shearing out of Fnc limbs indicates that Snc development continued through the cessation of Fn folding. This is characteristic of progressively developing high strain shear domains where fabric development and folds are part of cyclical strain hardening and softening (Wojtal and Mitra, 1986). Fnc hinge orientations in greenstones are oblique to the fault zone elongation direction, whereas those

Figure 3.11 Photomicrograph of C-S fabric relationship from same section as in figure 3.10. Snc mylonitic schistosity (S-surface) is defined by recrystallized quartz grains and sericite which comprise the "ribbon structures in this view. Shear band foliation (C-surface) is defined by second growth of chlorite (or realigned grains). These shear bands post-date the Sn flattening phase seen in 3.10a and probably represent late-stage simple shear within the Granville Thrust Zone. C-S relationship gives east-over-west rotation sense, consistent with asymmetrical tails (figure 3.10b) but inconsistent with plagioclase inclusion trail vorticity, if the plagioclase is interpreted to have rolled during growth (3.10b). Sample HQ 264 A; Allbee Brook locality; Plate 1, M30. View is 1.2mm, oriented looking south with top edge of photograph oriented parallel to the horizontal (ie; fabrics dip east).

within the schist are parallel to it. As stated before, this is evidence for differential rotation of hinges in a rotational strain environment where amount of rotation is dependent upon the mechanical strength of the deforming rock. Rotation senses of folds are consistently east over west throughout all units in the detailed map area. The relative age relationship between Fnc folds and Snc foliation indicates that folding developed prior to fault zone development; they are therefore not simply fault related folds.

The eastern greenstone eastern contact with the eastern schist can be seen in many areas (figure 3.12; eg. E,F,G). To the south, the contact is quite gradational, and therefore interpreted as a depositional contact. Within the southern part of the detailed map the width of the greenstone is well over 50m . As this contact is traced northward it appears to cut across strike rather abruptly as if folded by Fnc. This appearance is confirmed near the brook where an actual map pattern scale Fn fold can be readily seen folding the contact and causing its migration to the west (figure 3.12, H). The asymmetry of the fold is strongly east over west. The large scale fold has an axial plane orientation parallel to the surrounding high strain Snc with a hinge trend and plunge to the northeast at ~45°. The width of the greenstone at this point is less than 50 ft. and rapidly approaching the thickness as seen at the bridge (~3m). The foliation within the eastern schist becomes more pervasive north of this large fold with a strong mylonitic fabric containing east over west kinematic indicators as those of the mylonitic schist belt to the west of the bridge. The contact between the eastern schist and greenstone belts is quite sharp in this area and lacks the characteristic gradational zone. To the north of the bridge, the eastern greenstone terminates abruptly with the eastern and western schist belts becoming unified and of identical mylonitic texture. The western greenstone belt can be continuously traced another 1km to the north where it also terminates against the mylonitic zone within the schist belt (figure 3.12; A).

The combination of truncation along a discreet, undeformed plane and the mylonitic character of that surface are here considered demonstrable evidence for the tectonic truncation of continuous stratigraphy along a synmetamorphic fault zone. The consistent sense of fault zone fabric asymmetry, the rapid decrease in thickness of the eastern greenstone, and the eventual termination of that greenstone along the mylonitic surface all argue in favor of this interpretation. The progression in fabric intensity along the eastern greenstone/schist contact from south to north is the result of strain distribution across both sides of the tectonically thinned greenstone during progression of deformation. The lack of west over east Fnc fold asymmetry argues against the greenstone termination being a simple Fnc fold closure. This is further substantiated by the planarity of the western mylonitic schist belt, whose fabric transposes the limbs of Fnc folds within quartz veins.

Fault zone fabric:

Thin section analysis of oriented sections from the intermediate mylonitic schist belt show a pervasive, fine grained quartz-rich mylonite. Elongate quartz ribbons appear to have their long axes oriented at a very low angle to the pervasive mylonitic schistosity (Snc; defined by chlorite and sericite lathes) and defines an east over west asymmetry (figure 3.13). The near parallelism of the quartz ribbon long axis and the mylonitic foliation (parallel to the regional Snc), indicates that deformation included a large flattening strain or very high shear strain without dynamic recrystallization (Ramsay and Huber, 1983). The east-over-west imbrication of the ribbons, however, suggests that simple shear was operative sometime during ribbon development within this zone, possibly prior to the flattening deformation (figure 3.13). Some recrystallized tails of quartz on albite and pyrite porphyroclasts also show east over west asymmetry with the coalesced part of tails parallel to the Snc foliation (figure 3.10b). Fine grains of aligned chlorite and sericite define a shear band foliation which cross-cuts and bends mylonitic Snc into the shear plane (figure 3.13). The geometric relationship between mylonitic Snc foliation and the shear bands defines east-over-west rotation that is the strongest evidence for simple shear dominated deformation within the mylonitic fault zones of this area. These shear bands dip to the east at a shallower angle than Snc, producing an anastomosing, phacoidal geometry with Snc in thin sections and hand samples. The intersection of the two fabrics is observable in hand sample as a slight crenulation of the Snc foliation surface, sub horizontal in attitude. The widely spaced and weakly developed nature of the shear bands (with respect to the pervasive Snc) makes them impossible to record as a planar feature in the field.

Other kinematic indicators are rare due to the symmetrical nature of most matrix grains, either associated with discreet fault zones or in rocks containing the regional Sn fabric. Asymmetric recrystallized tails of quartz, chlorite, and sericite are commonly found on albite porphyroclasts and also consistently display an east over west asymmetry when viewed in a cut parallel to the fault zone elongation mineral lineation. Mylonitization within Snc shear zones appears to occur under sub-garnet grade conditions (possibly sub-biotite) due to the lack of garnet in rocks of similar composition to garnet-bearing lithologies in the western domain (ie: Mt.Abraham and Monastery Formations). Mn-rich garnet (stable at lower temperatures than almandine or pyrope; Tracy et al., 1990) within mylonite at Allbee Brook probably also formed at sub **garnet-grade** conditions.

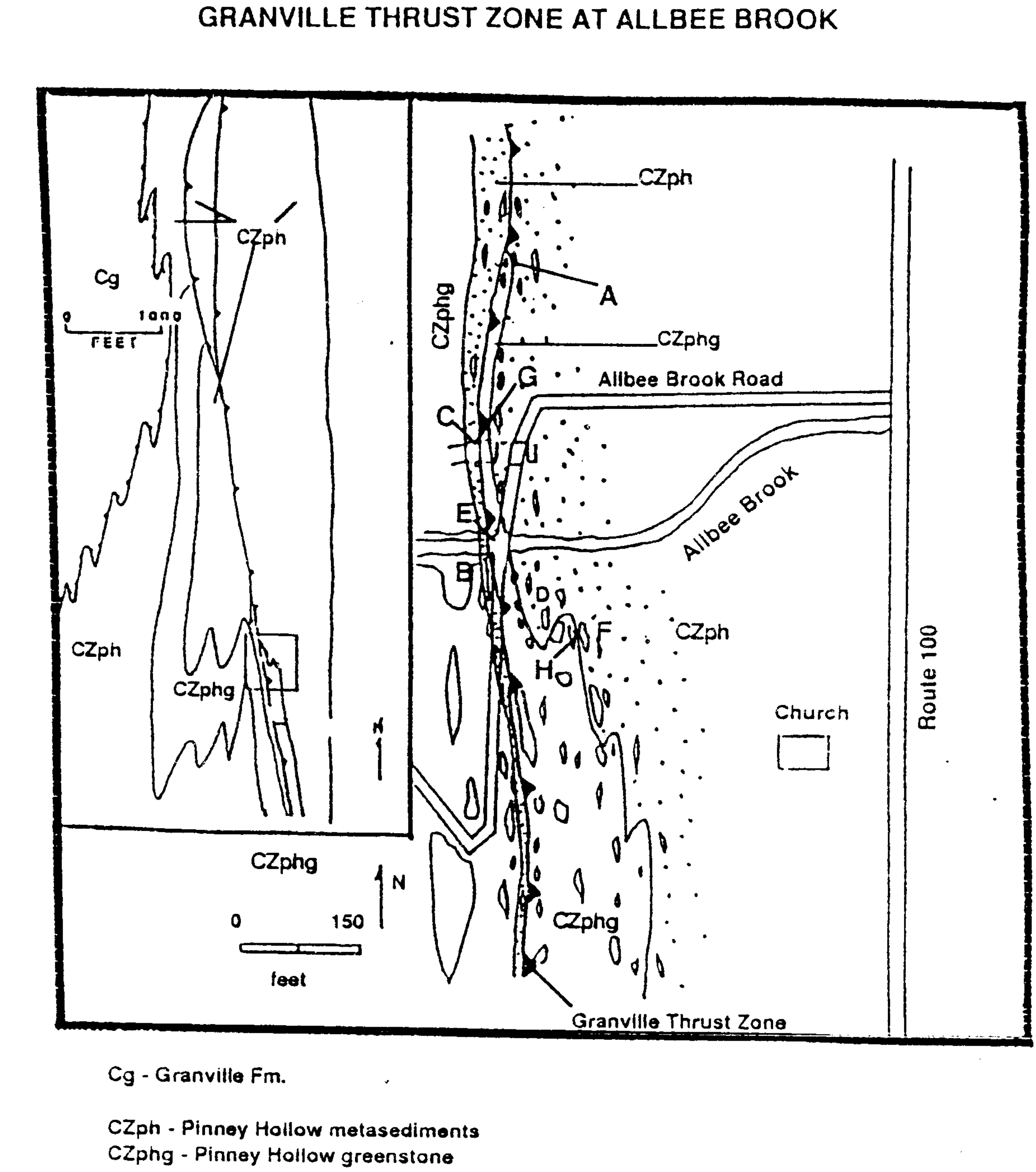


Figure 3.12 Detailed geologic map of the Allbee Brook locality and the Granville Thrust Zone (with regional inset map). This locality contains a well constrained map pattern truncation of a Pinney Hollow Formation greenstone (east of fault zone) along a synmetamorphic thrust fault (point A). Letters refer to locations discussed in text (chapter 3).

Fault zone lineation:

Mineral lineations from the Allbee Brook fault zone (also known as the Granville Thrust Zone), and all other N-age fault zones, are defined by chlorite, sericite, tourmaline, and rare biotite, as well as quartz pressure shadow alignment from overgrowths on earlier magnetite and pyrite porphyroblasts. Orientation of fault zone lineations are consistent with those measured outside of the fault zones.

Helicitic inclusions:

Plagioclase porphyroclasts within this zone contain inclusions of ilmenite, quartz, chlorite, and chloritoid defining a foliation which is continuous with the mylonitic Snc outside of the porphyroclast (figures 3.10a&b). These inclusion trails appear to be either helicitic or folded, sometimes defining an arc of ~90°. The inferred "axial planes" of the inclusion trails show a very consistent orientation, coplanar with axial planes of crenulate to isoclinal folds within the matrix; if the helicitic inclusions were the product of rotation, a large variation in these "axial plane" orientations would be expected since magnitudes of rotation would vary amongst differing shear planes (Bell, 1981). In addition, helicitic inclusion geometries occasionally show both east-over-west and west-over east asymmetry within the same thin section. These relationships suggest that the helicitic inclusions are the result of folding of the earlier Snc foliation prior to shear zone development and transposition of matrix Fnc crenulations. Growth of the porphyroblasts must have occurred following fold development and prior to matrix transposition during shear zone development (Bell, 1981). The common preservation of crenulation hinges within the porphyroclasts may not be fortuitous; Bell and Johnson (1989) suggest that porphyroblast growth will occur within low stress areas of particular folds which for quartz and feldspar are almost always the hinge areas. These porphyroclasts may actually preserve the early development of Fnc folds in the form of crenulations prior to their progressive rotation and attenuation into sheath-like folds and complete transposition of the crenulated schistosity (in this case, Snc-1).

Helicitic inclusion are common within many of the pelitic units of central Vermont, even those outside of the very high strain zones. This interpretation of inclusion geometry disagrees with previous interpretations in this belt that state helicitic inclusion development as a result of porphyroblast rotation (see Haydock, 1988; Kraus, 1989; Prewitt, 1989).

Fnc+1 folds:

These folds are found in discreet zones within this domain, usually in areas of reactivated Snc fault zones. Geometry of folds is quite consistent; upright and open, to slightly westward overturned, with predominantly shallow northeast plunging hinges. Occasionally, hinge orientations will be somewhat steeper with a few measurements southeast plunging parallel to the Nc-age mineral lineation (plate 6; Fnc+1 folds). No coeval foliation was observed with these folds, although the best fit great circle to the hinge data is relatively coplanar to undeformed Snc orientations (plate 6; N11E, 64SE). Fn+1 folds in the Granville Gulf area (immediately north of the study area), however, do have an axial planar foliation at a slight angle to the dominant Sn schistosity (Stanley et al., 1987c; Kraus, 1989).

Fnc+1 folds <u>deform</u> a sub-garnet grade (Snc) foliation into mesoscopic, asymmetric, westward overturned major folds (with parasitic folds showing both rotation senses) that contain microscopic crenulations of Snc mica-rich interlayers 100-300 microns in wavelength. Several analyzed thin sections contains chlorite pseudomorph textures, most likely after garnet which occur within fold hinges. The local and pervasive distribution of these folds within the central domain, is coincidental with coplanar Snc mylonitic fault zones. These folds are therefore probably the result of strain hardening following Snc fault zone development. Consistent east- over-west asymmetry of folds can be indicative of **pure** shear dominated deformation with a subordinate rotational strain component (Cosgrove, 1976). The coplanarity of Sn and Sn+1 deformation, the east over west vergence of of large scale Fnc+1 folds, and the progression in fold hinge orientation toward the the elongation lineation with increasing strain may indicate that these two deformations are continuous phases. These similar styles may also suggest that they form part of a strain continuum within a uniform stress state, possibly dominated by pure shear with a minor simple shear component, rather than punctuated phases of pure and simple shear (although the latter cannot be rúled out) (Cosgrove, 1976).

Sn+2 foliation:

A secondary foliation was observed in the central domain, superposed over the dominant Snc. It is not observed with Snc+1 and its relative age is therefore unclear. This foliation (Snc+2) is not homogenous over the entire domain. Rather, it is only locally developed, without any regard to preexisting structure. It consists of discontinuous 1 to 10mm anastomosing shear bands which bend the earlier Snc, creating either east over west and west over east asymmetry (figure 3.14). These Snc+2 shear bands dip steeply, either to

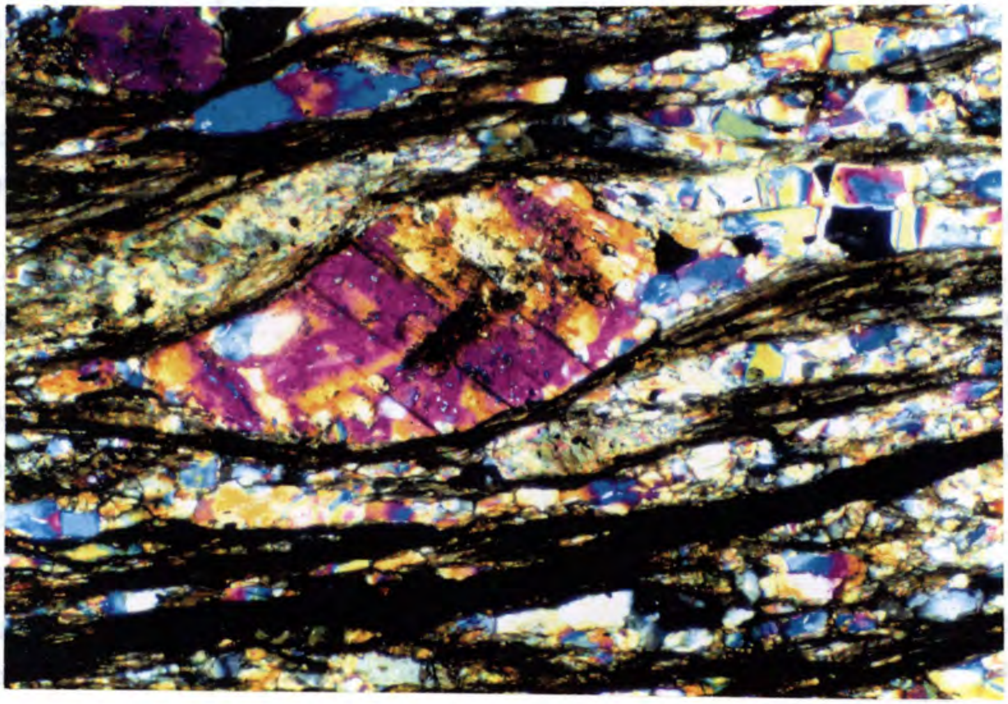


Figure 3.13 Photomicrograph of quartz ribbons whose imbrication vergence and long axis C-S relationship to the mylonitic foliation (defined by sericite and chlorite) suggest east-over-west (dextral) rotation. The lack of strong asymmetry of the ribbons, the symmetry of other matrix grains, and the presence of a younger shear band foliation that bends the mylonitic foliation (also producing sinistral rotation) indicates that a large component of flattening strain postdated the initial simple shear development of the ribbons, followed by a late-stage simple shear. Sample HQ 256 B from Allbee Brook locality. View is 1.2mm

the west or east, striking N10E. The presence of paired east and west dipping shear bands is common in most outcrops, creating a conjugate pair with the bisector of the obtuse dihedral angle normal to the Snc plane. Within ductile regimes, the dihedral angle associated with the maximum compressive stress commonly opens up and becomes the obtuse bisector (Ramsay, 1967; Ramsay and Huber, 1983). This would make Snc normal to the maximum compressive stress, and therefore indicate that the resolved shear stress, at least during the last increment of Snc development, was minimal to nonexistent; this would explain the large amount of symmetrical microfabric within these rocks, and support the hypothesis that flattening played a major role in the development of the regional deformational fabrics.

Another set of conjugate shear bands, and the set making up the majority of those measured within the study area, is locally found along the crest of Child's Mountain. This fabrics associated with cataclastic deformation and slickenline development. The conjugate pair strike N25E and dip steeply either west or east (plate 6). These shear bands cut across the Snc structural grain and may be related to a much later (possibly Acadian) age of deformation than Snc+2 found within the White River Valley and apparently associated with Snc.

Snc+3 foliation:

Locally found within the central and eastern parts of the central domain is a moderately spaced (2 to 6mm) crenulation and/or pressure solution cleavage. This cleavage is only locally present and is usually so weakly developed that it is not uniformly coplanar. It also does not cause any significant reorientation of earlier structures. Thin sections of pelitic rocks with Snc+3 contain either a carbonaceous selvage (in graphitic rock-types) or chlorite and sericite kinking, parallel to the cleavage plane. Although irregularly developed, the general orientations of the cleavage is fairly consistent, dipping predominantly steep to the west, although some steep east dipping cleavage was observed. The point maxima orientation of Snc+2 data defines a general orientation of N24E, 86NW (plate 6). This cleavage therefore does not have any spatial or stylistic association to the previously mentioned planar fabrics and is interpreted to have formed following a major shift in the regional stress state; it is similar to the orientation of the oldest recognizable cleavage within Silurian-Devonian rocks 16km to the east of the study area (Hatch, 1987; Westerman, 1987; Hatch, pers. comm., 1989).

Fnc+3 Folds:

Associated with the crenulation/pressure solution Snc+3 cleavage is a phase of weakly developed, small scale (1 to 10mm amplitude) open, upright folds, whose axial surface is coplanar to the Snc+3 cleavage (plate 6). Fold hinges are commonly inclined to the northeast, with a point maxima hinge oriented N18E, 40° (plate 6). A second cluster of data (1% contoured cluster) is based upon a few hinges which plunge steeply to the southwest. The lack of a well developed fold geometry with low amplitude folds, and the lack of a well defined girdle of data, suggests that little, if any, reorientation of the hinges by rotation, has occurred. Additionally, no subsequent deformation was observed that might be responsible for passive reorientation.

Both Fnc+3 and associated Snc+3 deformation become progressively better developed to the east within the eastern domain. This same fabric can be traced farther to the east within the Roxbury area, where it becomes the dominant deformational fabric in the pre-Silurian Stowe and Moretown Formations (Kraus, 1989).

Eastern Domain

Sne Foliation:

The dominant fabric within the eastern domain is a chlorite grade, closely spaced (0.5 to 1mm) penetrative mylonitic foliation that is associated with both isoclinal, reclined folds (Fne) and mylonitic fault zones. Within the pelitic units, Sne is defined by quartz segregations and fine folia of aligned chlorite and sericite. Sne orientations are fairly consistent, but variation does exist primarily as a result of reorientation by later deformation (point maxima orientation of N21E, 58SE; plate 6). Syn-metamorphic kinematic indicators are generally absent within this domain, in which rocks consistently display symmetrical grain shape and fabric geometries. Quartz ribbons in some of the quartzites do exist but are rarely imbricated. Shear band foliation, as described for the central domain, are also rare but do exist within several of the east-over-west mylonitic shear zones. Black phyllite of the Ottauquechee Formation is typically closely spaced but because of its phyllitic nature it does not tend to record much of the severe deformation. The lack of observable deformation is due to the high recovery rate of the phyllosilicate-rich phyllite. The severity of deformation is recognizable in the phyllite where it is associated with interlayered quartzite or silty laminations ("bacon rock") which ideally preserve the numerous phases of synmetamorphic folds and foliation.

Greenstones within this domain contain a variably developed Sne, of which the thicker and thus more mechanically competent ones do not have a highly mylonitic or pervasive Sne. The dominant foliation within these rocks is an earlier (Sne-1) compositional layering of chlorite, carbonate, epidote, and albite/quartz segregations. The thinner (meter scale) greenstones are less competent, and accommodate the pervasive Sne which thoroughly transposes the earlier, commonly folded Sne-1. This phenomenon is characteristic of the central domain as well.

Sne mylonitic fault zones:

Although kinematic indicators, and thus rotational strain, is present to some extent within all of the pelitic rocks, several discreet zones occur within the more competent schistose rocks which display pervasively sheared fabrics and quartz-rich mylonitic fabric (plate 1; solid thrust teeth). The development of the quartz rich zones may either be a result of inherent compositional differences or a product of strain induced chemical reaction involving the conversion of phyllosilicate and feldspar to quartz and accessory minerals (Knipe and Wintsch, 1985). Since many of the compositionally homogenous rocks record high strain features (ie; reclined folds, elongation mineral lineations, and coplanar shear bands and mylonitic foliation), without transitional quartz-rich zones, it seems doubtful that the actual fault zone rocks represent entirely chemically altered tectonite. It seems more likely that inherent compositional variations did exist (ie; quartzite beds) which provided accommodation of high strain rates and recrystallization necessary for large scale fault zone development. Fault zone fabrics and composition are therefore most likely the result of compositional variations rather than chemical alteration, although ample evidence exits that fluids played a significant role during deformation and metamorphism (see chapter 4). These fault zones all show equilibration of coeval minerals (quartz, chlorite, sericite, albite) at sub garnet grade conditions, evidenced by the inclusion of garnet within the growing porphyroblasts. As with synmetamorphic faults within the central domain, mylonitic fault zones truncate stratigraphic units either internal to major formations or those bounding formations. These stratigraphic units include numerous greenstones and several thin and thick dark quartzites within the Ottauquechee Formation (Cobq; plates 1 & 2).

Fne folds:

Folds which have Sne as an axial planar foliation, are consistently tight, isoclinal with large amplitudes, and are reclined, parallel the associated Sne elongation mineral lineation (plate 6; Fne point maxima = N88E, 64°). Most of the recognizable folds within this domain are

Greenstones within this domain contain a variably developed Sne, of which the thicker and thus more mechanically competent ones do not have a highly mylonitic or pervasive Sne. The dominant foliation within these rocks is an earlier (Sne-1) compositional layering of chlorite, carbonate, epidote, and albite/quartz segregations. The thinner (meter scale) greenstones are less competent, and accommodate the pervasive Sne which thoroughly transposes the earlier, commonly folded Sne-1. This phenomenon is characteristic of the central domain as well.

Sne mylonitic fault zones:

Although kinematic indicators, and thus rotational strain, is present to some extent within all of the pelitic rocks, several discreet zones occur within the more competent schistose rocks which display pervasively sheared fabrics and quartz-rich mylonitic fabric (plate 1; solid thrust teeth). The development of the quartz rich zones may either be a result of inherent compositional differences or a product of strain induced chemical reaction involving the conversion of phyllosilicate and feldspar to quartz and accessory minerals (Knipe and Wintsch, 1985). Since many of the compositionally homogenous rocks record high strain features (ie; reclined folds, elongation mineral lineations, and coplanar shear bands and mylonitic foliation), without transitional quartz-rich zones, it seems doubtful that the actual fault zone rocks represent entirely chemically altered tectonite. It seems more likely that inherent compositional variations did exist (ie; quartzite beds) which provided accommodation of high strain rates and recrystallization necessary for large scale fault zone development. Fault zone fabrics and composition are therefore most likely the result of compositional variations rather than chemical alteration, although ample evidence exits that fluids played a significant role during deformation and metamorphism (see chapter 4). These fault zones all show equilibration of coeval minerals (quartz, chlorite, sericite, albite) at sub garnet grade conditions, evidenced by the inclusion of garnet within the growing porphyroblasts. As with synmetamorphic faults within the central domain, mylonitic fault zones truncate stratigraphic units either internal to major formations or those bounding formations. These stratigraphic units include numerous greenstones and several thin and thick dark quartzites within the Ottauquechee Formation (Cobq; plates 1 & 2).

Fne folds:

Folds which have Sne as an axial planar foliation, are consistently tight, isoclinal with large amplitudes, and are reclined, parallel the associated Sne elongation mineral lineation (plate 6; Fne point maxima = N88E, 64°). Most of the recognizable folds within this domain are found within the very competent greenstones and quartzites, since phyllite recrystallization and transposition during continued Sne fabric development tends to obliterate Fn folds within this high strain domain. The fact that many of the hinges are parallel to the mineral lineation (reclined), even though within greenstone and quartzite, attests to the severity of N-age rotational strain within this domain (plate 6). This assumes that fold hinges have not passively rotated through recrystallization into incremental elongation directions during a pure shear dominated deformation. This type of pure shear passive recrystallization should produce quartz rods (fold hinges) which overgrow older quartz rods that grew in progressive elongation directions. This type of growth was not recognized within this area.

Mineral lineation:

Defined by aligned chlorite and sericite in schist and graphitic phyllite, as well as graphite and carbonaceous material, the N-age mineral lineation is identical to that one found in the central and western domains. It occurs parallel to the long axis of asymmetric fabric including many of the observable kinematic indicators. It is therefore also interpreted as an elongation mineral lineation associated with N-age strain. Orientations of the mineral lineation vary with change in orientation of the Sne foliation. The distribution of data across the Sne defined great circle demonstrates redistribution of the linear elements by later deformation (Fn+1 and Fn+3 folding). The point maximum of the data is oriented N39E, 44° (plate 6) quite divergent from that for undeformed planar and linear data in the central domain. The reorientation of the elongation direction from the southeast quadrant into the northeast quadrant parallels the reorientation of Sne from a north-northeast strike to a northwest one (plate 6). This redistribution is correct for northeast oriented folding of both Fn+1 and Fn+3 hinges.

Quartz rods:

As with those described from the western and central domains, quartz rod lineations define the orientation of either recrystallized quartz segregations or Fne fold hinges of quartz veins. Commonly they are parallel to the elongation direction in areas of high flattening and/or rotational strain. The lack of abundant asymmetrical grain and fabric geometry within this zone (excluding the discreet mylonitic fault zones) suggests that mineral lineations within the unmylonitized rocks are the result of elongation associated with a flattening strain; mylonitic fault zones with fabric asymmetry have mineral elongations resulting from a dominantly rotational strain. These two different strain-types produced colinear mineral lineations that have been indistinguishable in orientation and mineral-type and morphology. In either case, the colinearity of quartz rods and mineral lineations in the eastern domain is in agreement with the interpretation that preexisting fold hinges were transposed into the elongation direction (N38E, 38°; plate 6). The variation from reclined to inclined geometry, based upon the folding of either pelitic or mafic rock-type, with different mechanical competency, (described for the central domain) suggests that fold hinges were not initially developed in the reclined (parallel to mineral lineation) orientation. Fanning of the rods is also similar to that of the mineral lineation and Fne hinge data and further suggests redistribution by later noncoaxial folding.

Sne+1 foliation:

Within several areas, a second foliation is present in an axial plane orientation to a set of westward overturned, upright folds (Fne+1). Where present, this foliation (Sne+1) is wavy to quite planar, moderately spaced (1 to 3mm), and defined by aligned chlorite and sericite, and occasional carbonaceous material. Thin sections of Sne+1 show that the aligned grains are both realigned, kinked Sne mineral growth and later new grains. Although grain size kinematic indicators are absent, the foliation displaces N-age quartz vein segregations creating asymmetric tails with consistent east over west rotation sense (figure 3.15). The orientation of the Sne+1 point maxima defines a plane oriented N06E, 61SE (plate 6). This orientation is similar to that of undeformed Sn in the central domain and of Sne in areas of little or no later deformation. The coplanarity of Sne+1 with undeformed Sne suggests development of the two fabrics in a progressively evolving, but similarly oriented strain state. In order to fold Sne, Sne+1 must have actually developed slightly oblique to Sne so that a resolved component of compressive stress could act on the preexisting planar surface (Cosgrove, 1976; Bell, 1981; Ramsay and Huber, 1983). The lack of preferred grain shape, grain asymmetry, and C-S fabric associated with Sne+1 suggests that the strain was dominantly pure shear within the matrix; the asymmetric tails on quartz veins, oriented oblique to the Sne+1 fabric does indicate a rotational strain component was active during the bulk flattening and foliation development. This asymmetry also indicates east-over-west rotation.

Sne+1 fabric development increases in intensity toward discreet N+1 shear zones with mylonitic fabric and lithologic truncations. Although common in the eastern domain, these zones are discontinuous along strike and probably are not associated with significant fault displacements. For this reason, they have not been included on the geologic map as discreet synmetamorphic fault zones.

Fne+1 folds:

This phase of folding is associated with the Sne+1 foliation and is present within discreet zones of pervasive foliation development. Folds are usually westward overturned, upright to rarely isoclinal, with hinges commonly plunging to the northeast (plate 6; N24E, 35°). Hinge orientations fan significantly with some data plunging either steeply or moderately to the southeast, sometimes parallel to undeformed N age mineral lineations (plate 6).

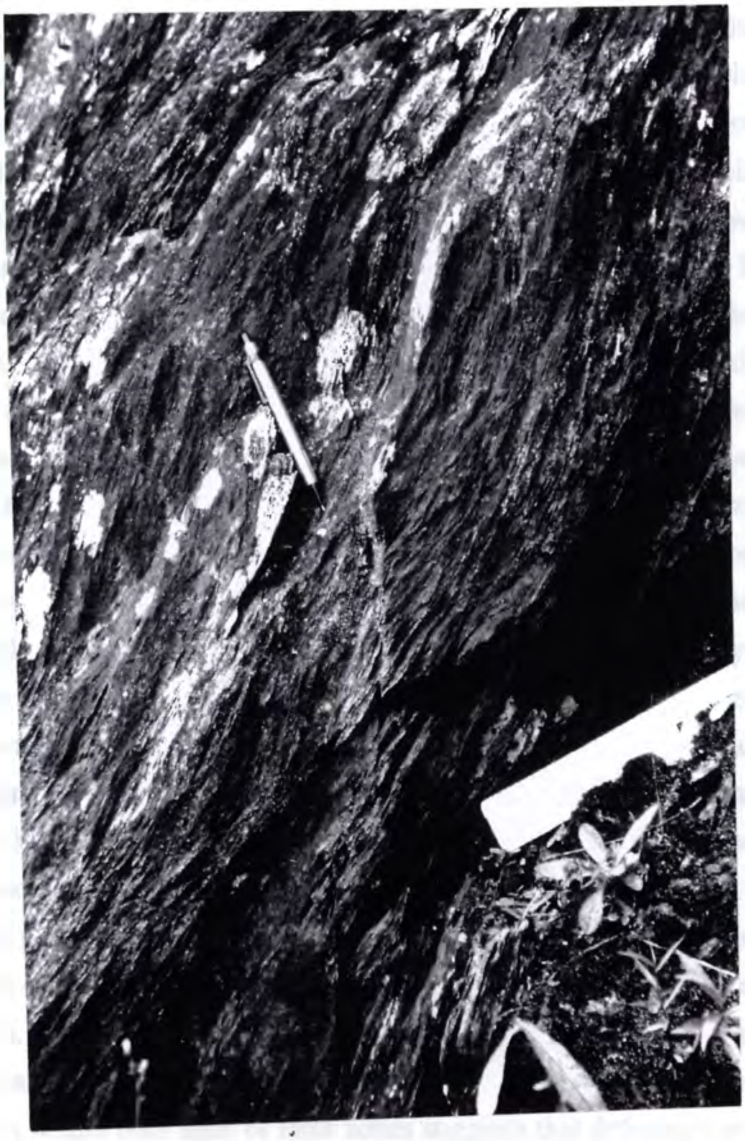


Figure 3.14 Snc+2 conjugate shear bands whose acute angle is bisected by Snc (between pens in photo). The obtuse bisector (sub horizontal position relative to the picture) may be the principle compressive stress axis, which is common for conjugate shears within ductile regimes (Ramsay and Huber, 1986; Cosgrove, 1976). Outcrop within the Pinney Hollow Formation, Route 100, immediately north of the town of Granville.

Because the undeformed Snc surface is strongly isoclinal with very large amplitudes, variation of hinge orientations is probably related to variation in the magnitude of rotational strain across any particular area. The cluster of hinges in the northeast quad of the equal area net most likely represents relatively unrotated linear elements. This orientation is in the proper attitude to cause the significant reorientation of earlier N-age planar and linear data.

Fanning of the Fne+1 fold hinges attests to some component of progressive rotational strain during development of at least Fne+1 lower greenschist facies deformation. The progression of rotational strain and resultant fabric is well displayed in Thatcher Brook where a pervasive Sne foliation and associated Fne reclined folds and lineation are folded by a coaxial phase of Fne+1 folds, all displaying east over west rotation sense (figures 2.9 & 3.16). Axial surfaces to the Fne+1 folds commonly show the initial development of a second foliation, Sne+1, coplanar to undeformed Sne. Folded Sne and associated mineral lineation and fold hinges show a prevalent redistribution into a northeast attitude (figure 3.16, plate 6). This entire zone of folding (~1 km across strike) occurs within a zone of earlier Sne fault zones (subsequently folded) and later Sne+1 shear zones, both with east over west asymmetrical fabric. The coplanarity of planar fabric, the progressive decrease in fold amplitude and rotation, and the distribution of Sn and Sn+1 mylonitic fault zones suggest that all of this deformation occurred within a single strain state as cyclical periods of strain hardening (folding) and strain softening (faulting). The cyclical nature of hardening and softening has been interpreted as a consequence of changing deformational and physiochemical conditions, in particular fluid activity, mineral chemical potential, pressure, and temperature (Schmid, 1984a, 1984b; Wojtal and Mitra, 1986). The presence of abundant pyrite and other sulfides within both Sn and Sn+1 fault zones (relative to surrounding rocks) may be related to fluid advection through these horizons during deformation. Similar sulfide-bearing assemblages are found in fault zones within the central domain and the eastern part of the western domain. The presence of similar mineral assemblages within both ages of fault zones suggests that deformation occurred under similar conditions during both phases of deformation, across the belt. In reality, the localization of Sne+1 deformation rules out any physical demonstration of it all being one discreet deformation; it may very well have occurred at different times in different places as a true phase of Sne work hardening.

Sne+3 cleavage:

Both N and N+1 synmetamorphic fabric are superposed by a late moderately spaced (1 to 5mm) crenulation or pressure solution cleavage, similar in style and orientation to Sn+3 of

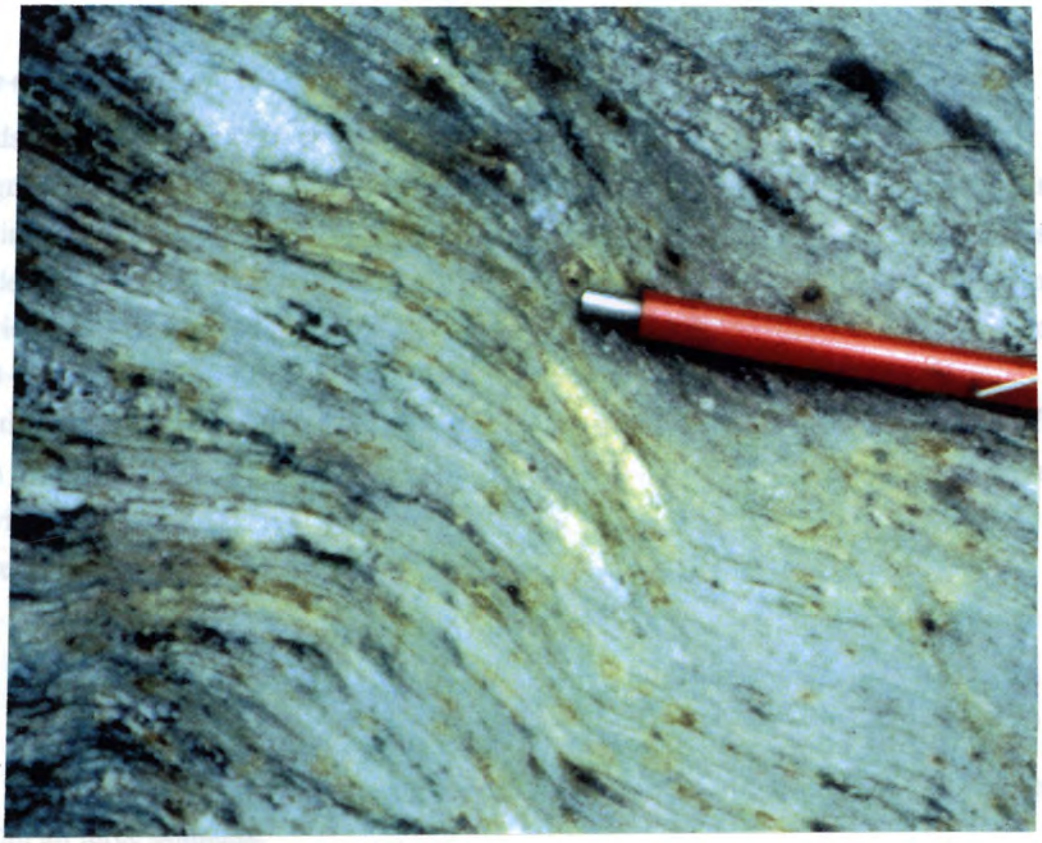


Figure 3.15 Outcrop of Pinney Hollow Formation albitic schist containing disseminated quartz vein with asymmetric tails indicating east-over-west rotation. Tail asymmetry is defined by obliquely oriented Sne schistosity (bounding the vein-schist interface) which is bent into the new foliation, Sne+1. This new fabric is locally developed in association with Sne+1 mylonitic fault zones, both coplanar to undeformed Sne fabric. This fabric is interpreted as a strain softened continuum of Sne following strain hardened, Fne+1 fold development. Photograph from Thatcher Brook, immediately west of Thatcher Brook Member (Ottauquechee Formation) type locality; plate 1, T35.

the central domain (figure 3.17; plate 6). Sne+3 is observed in thin section as either bent or kinked chlorite and sericite, or as selvages containing carbonaceous matter. The waviness of the cleavage surface and its punctuated distribution across the eastern domain, suggest its relatively weak development in the study area. No mineral lineations were found associated with this late cleavage.

Fne+3 folds:

Folds are associated with Sne+3, usually very open and upright, occurring at several magnitudes of order. The most common folds are micro- and mesoscopic, usually 1mm to 1m in amplitude (figure 3.17). Larger orders do exist, including a megascopic, ~1km scale moderately northeast plunging antiform which regionally deform the earlier Sne and Sne+1 fabric, causing its general redistribution to northwest trending strike lines (plate 6; plate 1, Z40-T45). Fold axial surfaces are defined by the Sne+3 cleavage. Fold hinges are relatively consistently oriented, with a point maxima of N17E 44 (plate 6). Fanning of the data is directly related to the weak development of this deformation phase, including the waviness of associated Sne+3. Fne+3 folds do appear to increase proportionally with increased Sne+3 development to the east (Kraus, 1989).

General fabric elements

This section includes three general structural features that are similar in character within all three domains.

Post metamorphic faults:

Many of the large scale N and N+1 synmetamorphic fault zones show brittle reactivation, evidenced by cataclastic fabric, gouge, tectonic breccia, slickensided surfaces on lozenge shaped strain zones, and very late chlorite clot overgrowth. These zones are usually restricted to a meter-scale width, even if earlier mylonitic zones were significantly wider. Motion sense on many of the faults is not determinable based on a general lack of kinematic indicators. Several late reactivated zones do show foliation asymmetry around low strain zones. The reactivated synmetamorphic fault bounding the western contact between the Pinney Hollow and Ottauquechee Formations (plate 1) shows foliation asymmetry around uncataclasized rock, with an east over west motion sense. This reactivation of an earlier N-age thrust fault is traceable northward for over 15km into the Lincoln Mountain 15 minute quadrangle (Kraus, 1989; Prewitt, 1989). This particular zone, unlike the majority of the

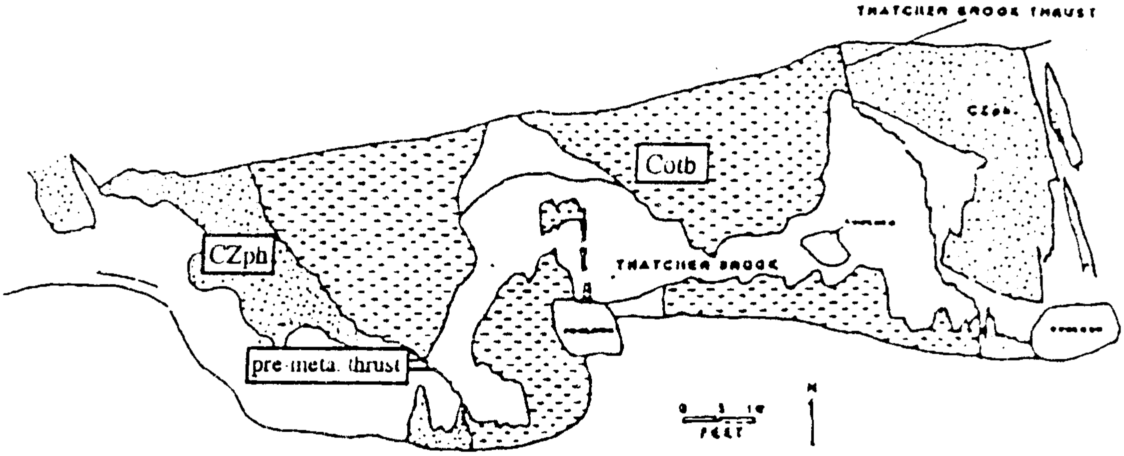


Figure 3.16 Detailed map and structural data for Sne+1 and Fne+1 elements at the Thatcher Brook type locality (plate 1; T35). Fne+1 progressively fold Sne into open upright, and tight isoclines with inclined fold hinges (net data and plate 6). The progressive decrease in the plunge of the Fne to Fne+1 fold hinges, despite the preservation of relatively steep, undeformed Sne, argues in favor of physical / passive rotation of hinges toward the elongation direction, rather than their growth in those orientations. It also argues in favor of a continuous Sn/Sn+1 strain state and cyclical syn-metamorphic (N/N+1) deformation. CZph-Pinney Hollow Formation; Cotb- Thatcher Brook Member, Ottauquechee Formation.

reactivated brittle fault zones, is regionally continuous and appears to have significant displacement (1km; plate 2; displacement based on the offset of upper- and lower-plate stratigraphy).

Most of the brittle fault zones are not continuous for more than a few outcrops, or wider than a few meters. The reactivation along the Ottauquechee Thrust Zone does not appear to be folded by the Fne+3 deformation; it is certainly possible that many of these zones are either Acadian or Mesozoic in age.

Sn-1 foliation:

Preserved as relict fold hinges within the dominant Snc and Sne schistosity is a closely spaced (0.2-1.0mm) quartz-feldspar-sericite-chlorite compositional layering within pelitic units and an epidote-chlorite ± actinolite ± barroisitic hornblende within the mafic units. This relict layering usually occurs as local vestiges within relatively low stress Fne fold hinges, between the mica-rich Sne folia (figure 3.18). It is usually completely transposed into Sn schistosity, therefore its original orientation relative to the later synmetamorphic deformations cannot be determined precisely. Sn-1 is also preserved within the Battell Formation as compositional layering of graphitic and carbonate-rich schist (Cb) and graphitic phyllite-marble interlayering (Cbd); this layering seems to represent transposed bedding. This fabric is the dominant fabric in the western domain (Snw) where it is temporally associated with the garnet grade metamorphism; relict lower amphibolite facies (garnet grade) barroisite is preserved with Snc-1 compositional layering of massive greenstone at Allbee Brook.

Pre Dn-1 fault zones (premetamorphic fault surfaces):

Although many of the contacts between the major lithotectonic belts in the eastern cover sequence are parallel to the dominant foliation of any particular domain (Sn), many others are folded by the related synmetamorphic folds (Fn-1, Fn, and Fn+1) and are thus earlier contact surfaces (Sn-2 or older). These earlier contacts are common between greenstone and schist within the central and eastern belts, where some earlier synmetamorphic folds (Fn-1) do fold an even earlier surface (Sn-2) which is always the contact between two different compositional units. These Sn-2 surfaces, interpreted as bedding (or compositional layering) in some cases and premetamorphic fault surfaces in others, appear to be the oldest planar feature preserved within the Vermont pre-Silurian sequence. The local age relationships between Sn-1 and Sn-2 relative to the garnet grade/amphibolite facies metamorphism is complex; barroisitic hornblende, garnet, and an

Figure 3.17 Fne+3 fold with axial planar Sne+3 spaced cleavage in interbedded quartzite and graphitic phyllite of the Ottauquechee Formation. Fne+3 folds are northeast trending and deflect Sne into a northwest trend (plate 6). Sne+3 cleavage is locally developed within core of fold only. Older Fne fold on western (left) limb of Fne+3 (arrow) is associated with axial planar Sne, which is the Fne+3 folded surface. The Fne fold deforms preexisting Sne-1 schistosity/foliation, defined by quartz-rich and mica-rich compositional layering. Scale bar is 6 inches long. Outcrop is located east of Jeep trail road on west flank of hill (plate 1, M35).

old phase of plagioclase porphyroblasts show local evidence of pre-, syn-, and rarely, post-Sn-1 growth (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Prewitt, 1988; Walsh, 1989). What seems clear is that Sn-2 surfaces existed before the peak of amphibolite facies metamorphism (M1, chapter 4).

Because dynamic recrystallization and mineral reequilibration were so pervasive in the schistose units during N-1, N, and N+1 deformation, most of the earlier N-2 fault contacts within the eastern and central domains have been fully annealed, and do not preserve either fabric or mineral growth. These early fault contacts are demonstrated solely on outcrop-scale and map pattern truncations of lithologies on both sides of the contact. Truncation of lithologies on only one side of the supposed premetamorphic contact could be interpreted as an unconformity rather than a fault. Because of this it is possible that many other premetamorphic faults do exist within this belt, but could not be distinguished.

The following contacts are considered to be premetamorphic (pre-Dn-1) faults:

1. In the extreme eastern part of the study area, Sn-2 contacts between lithologies of the Pinney Hollow/Stowe Formations and the Ottauquechee Formation show abrupt termination of internal stratigraphy within the upper (Ottauquechee) and lower (Pinney Hollow/Stowe) structural level. Quartzites within the Ottauquechee Formation can be continuously walked out in the field, where they are quite continuous and of general constant thickness, until their termination along the Pinney Hollow/Stowe contact surface. The same holds true for continuous, and usually 1 to 25m thick, greenstone within the Pinney Hollow/Stowe Formations. Where abrupt terminations occur, the dominant foliation (Sne) can be observed cross cutting the contact (figure 3.19). The combination of upper and lower stratigraphic truncation with a superposed oblique Sne foliation, requires the older (Sn-1 or older) surface to be a pre-Dn fault contact. This tectonic contact between the upper plate Ottauquechee thrust slice and lower plate Pinney Hollow/Stowe thrust slice, commonly contains 1 to 200m long lensoidal bodies of serpentine and/or talc schist (plate 1). These "exotic" tectonic slivers support the other evidence for early faulting since ultramafics originated as either oceanic crust or rift-related cumulates associated with mafic intrusives or extrusives (Coish et al., 1985, 1986; Tracy et al., 1984). The present intercalation of ultramafics into the surrounding metasediments requires their incorporation either by sedimentary (olistostromal) or tectonic (slivers) processes (Hamilton, 1988; Stanley et al., 1984; Doolan et al., 1982; Armstrong et al., 1988b; Armstrong and Colpron, 1989a). The fact that these slivers occur along fault surfaces argues in favor of a tectonic incorporation.

2. Thatcher Brook Member of the Ottauquechee Formation, graphitic schist, quartzite, sandy schist, and several ultramafics, overlie the Pinney Hollow Formation within both the

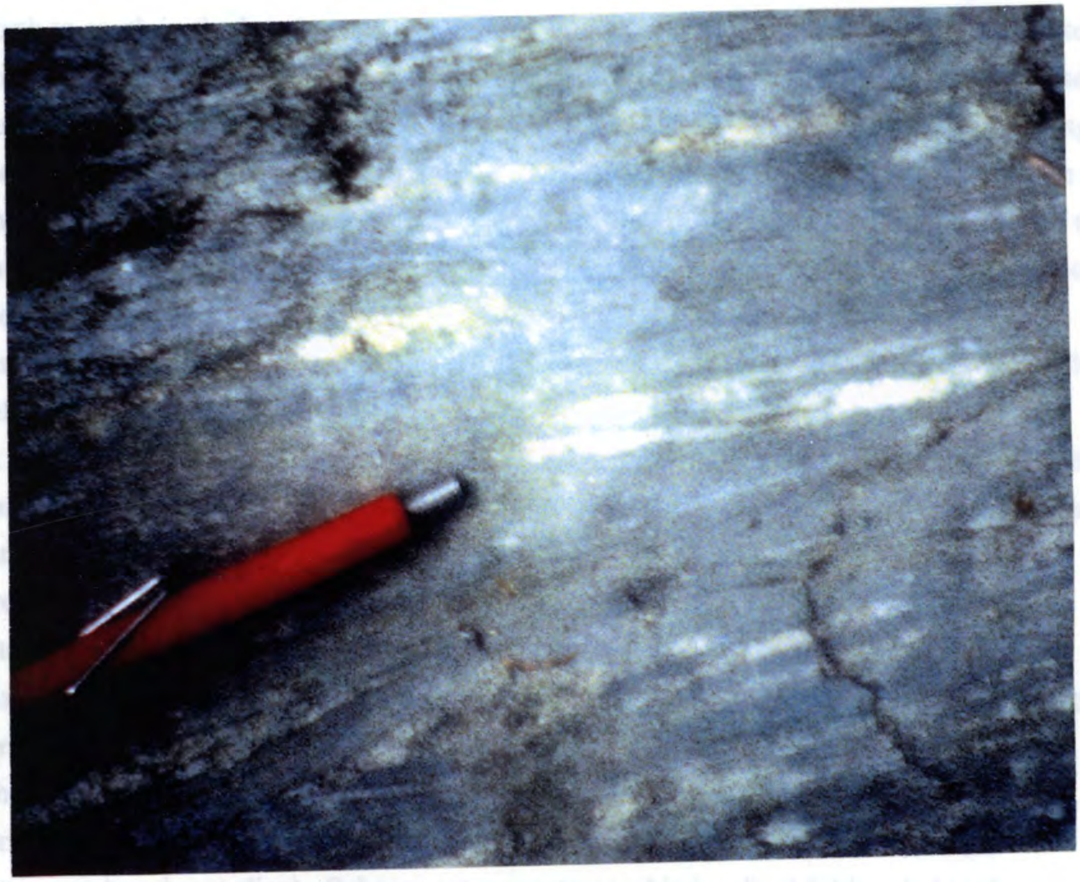


Figure 3.18 Relict Snc-1 compositional layering, defined by 0.5mm-2mm thick quartz-feldspar (light colored) and graphitic, sericite-chlorite (dark) layers. This layering is sometimes recognized as axial planar to Fn-1 fold hinges. Snc-1 layering is fully transposed into mica-rich Snc folia outside of the low stress energy quartz-feldspar-rich Snc microlithons. Pencil is 4cm long, oriented with point towards north. From Forestry Route 55, 200yds west of Granville town center; Granville Formation graphitic albitic schist.

central and eastern domains. This unit may be the western continuation (and lower structural level of) the Ottauquechee thrust slice. Greenstones of the Pinney Hollow Formation commonly truncate along the Thatcher Brook Sn-2 contact. Upper plate truncations within the Thatcher Brook lithologies cannot be distinguished at 1:12,000 scale mapping. At the Thatcher Brook type locality, however, graphitic schist underneath a continuous 4cm thick white quartzite progressively decreases in thickness along strike from north to south in the brook exposure (figure 3.16). Similarly, a 1m thick white quartzite within the Pinney Hollow truncates along this same contact. Sne foliation cross-cuts the quartzites, but locally transposes the schistose rocks underneath them, thereby obscuring the actual Sn-2 contact in some places. Although this exposure provides solid evidence of an early Sn-2 fault contact, it also shows how the contact geometry can be obscured by the subsequent and pervasive synmetamorphic deformation. Obviously, detailed mapping of well exposed outcrops, such as the Thatcher Brook exposure, is critical in recognizing premetamorphic fault zones. This is particularly true since Sn-2 fabric has also been either fully transposed or annealed by the later synmetamorphic deformation.

3. Within the central domain, Granville Formation graphitic albitic schist, with white and dark quartzite overlie the Pinney Hollow Formation, and appear to progressively grade into Thatcher Brook lithologies, farther east. This gradation is illustrated by the progressive eastward increase within the Granville, of discontinuous lenses of Ottauquechee-like dark gray quartzite, sandy schist, graphitic phyllite, and locally discontinuous, 1 cm to 1m thick, punky weathering carbonate rich greenstone. Both the Granville and Thatcher Brook (Ottauquechee) directly overlie the Pinney Hollow units. The lithologic similarities, gradation, and same structural position, suggest that the Granville and Ottauquechee Formations are parts of the same Sn-2 thrust slice (the Ottauquechee thrust slice). Subsequent synmetamorphic isoclinal folding led to the development of the antiform within the White River valley, cored by Pinney Hollow Formation, that separates the majority of the Granville from the eastern situated Ottauquechee Formation (plate 1 & 2). Since Granville Formation, homogenous graphitic schist does not contain any diagnostic stratigraphic markers, nor ultramafics along the Sn-2 surface, unequivocal demonstration of a Sn-2 fault surface with the structurally underlying Pinney Hollow Formation is not possible, even though Pinney Hollow greenstones do terminate along the Granville contact in several localities (plate 1). The truncation of greenstones may be due to an unconformity, or to faulting, but this faulting may be insignificant since the Granville is always present on top of the Pinney Hollow Formation and not other lithologies either to the west or east. If the Granville does depositionally overlie the Pinney Hollow Formation, then either;

A) The Granville is not the western continuation of the Ottauquechee Formation or;

B) The Granville is part of the Ottauquechee thrust slice, which terminates within the longitude of the Granville Formation, becoming a coherent depositional sequence with the underlying Pinney Hollow. Alternatively, the Granville is the western continuation of the Ottauquechee thrust slice which ramps up into, and over the Granville Formation in the vicinity of the White River valley. Both models have equal probability of being correct based upon the available information. Within the Granville Gulf, Warren, and Waitsfield areas, the Granville is in contact with the Pinney Hollow Fm. along the northern continuation of the Granville Thrust Zone (an Sn-age fault zone; Stanley et al., 1989).

4. The Mt. Abraham/ Pinney Hollow lithotectonic unit structurally overlies the Monastery Formation in several localities within the central domain, and the Battell Formation (White River Member) in others. Truncation of dolomitic schist of the Monastery and a thick greenstone within the Pinney Hollow, along this contact (the Child's Mountain Thrust) provides direct evidence for a tectonic surface, cross cut by the Snc foliation (plate 1; east side of Child's Mountain). The fact that the Pinney Hollow/ Mt.Abraham sheet overlies either Battell or Monastery in different localities, indicates major stratigraphic truncation of a coherent Monastery/Battell lithologic sequence (chapter 2).

Assuming early faulting does appear to be responsible for the gross distribution of many of the major lithotectonic units within the pre-Silurian section. The structural sequence of lowermost Monastery/Battell, intermediate Pinney Hollow/Mt. Abraham/Stowe, and uppermost Granville/Ottauquechee is the result of this early faulting (plate 4). Later phases of deformation simply deform this sequence with the reclined (sheath) folds and synmetamorphic faults primarily creating isolated tectonic inliers and windows defined by the pre-Dn contact surfaces. Later post-Dn/Dn+1 deformation Fn+3 folds and post metamorphic faults simply accentuate the preexisting superposed inlier and window tectonic geometry.

Pre-Dn-1 faults, controlling the early distribution of the major units, have been interpreted by Stanley and others (1984) and Stanley and Ratcliffe (1985) as being the result of accretionary wedge tectonics prior to later synmetamorphic, upright deformation, interpreted as heralding true continent-arc (continent) collision (Stanley and Ratcliffe, 1985; Armstrong and Colpron, 1989b). Regardless of the tectonic environment of formation, these faults may be of the same (or slightly younger) age as the emplacement of the Taconic allochthons. The superposition of a chlorite to sillimanite grade metamorphism amongst both allochthon and underlying autochthon rocks, indicates that faulting occurred <u>prior</u> to peak metamorphism (Sutter et al., 1985; Hames et al., 1990). Faulting within central Vermont also occurred prior to the thermal peak of metamorphism (chapter 4).

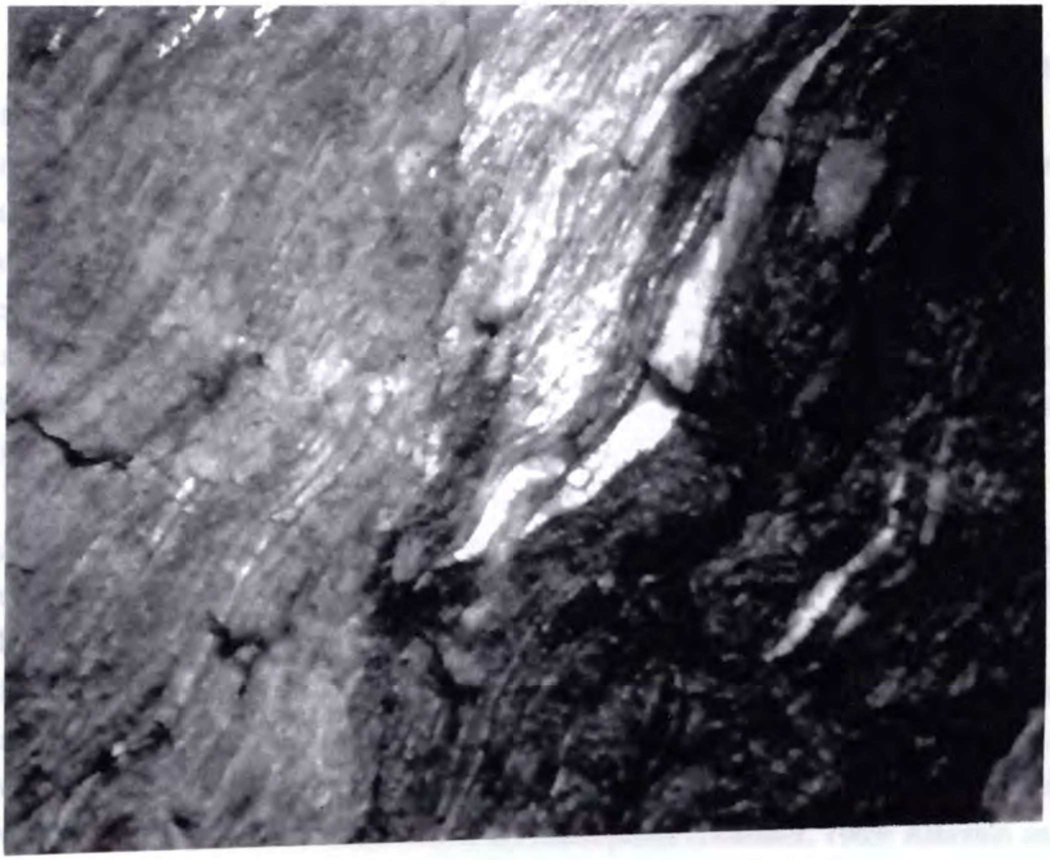


Figure 3.19 pre Dn-1, premetamorphic fault contact between black phyllite and quartzite of the Ottauquechee Formation (dark layers) and green, chlorite-phyllite, green albitic schist, and greenstone of the interlayered Stowe/Pinney Hollow Formations. The contact is extremely sharp and is cross cut at a shallow angle by the dominant schistosity, Sne. Fault fabric is nonexistent and is interpreted to have been fully transposed and annealed during the synmetamorphic deformations (Sn-1, Sn). Several meter thick black quartzite layers within the upper plate (Ottauquechee) truncates along this contact several yards north of where this photograph was taken (picture taken looking south). A 10m thick greenstone, continuous within the Stowe/Pinney Hollow Formations for more than 2km, truncates along this contact immediately south of this locality; upper(Ottauquechee) and lower plate (Stowe/Pinney Hollow) truncations demonstrate the tectonic nature of this surface. In some localities the contact is decorated by lensoidal slivers of talc schist and serpentine interpreted as tectonic slivers of either oceanic crust or rift-related ultramafic cumulates. (Plate 1)

CHAPTER 4
METAMORPHISM

Introduction

Laird and Albee (1981), and Laird, et. al (1984) have analyzed assemblages within pelitic and mafic units of the pre-Silurian eastern cover sequence. These studies have addressed the distribution of Ordovician and Devonian metamorphic isograds as well as pressure and temperature conditions during the development of Taconian (and possibly Acadian) metamorphic assemblages. In addition, Laird, et. al (1984), and Sutter and others (1985) have attempted to delineate the boundaries of Ordovician and Acadian metamorphism based on $^{40}Ar/^{39}Ar$ stepwise and total fusion spectra analyses, and K/Ar whole rock and mineral radiometric analyses of various assemblages throughout Vermont (figure 4.1).

The rocks of the Vermont pre-Silurian sequence record a complex, polymetamorphic history, believed to be primarily Taconian in age with the various phases of metamorphism have different ages relative to the synmetamorphic fabrics (Osberg, 1952; Cady et al., 1962; Laird and Albee, 1981; Laird et al., 1984; Sutter et al., 1985). Additionally, most of the mineral growth related to the various metamorphisms has been interpreted to as coeval with deformation. This phenomena of dynamothermal mineral growth was first hypothesized during the end of the last century (Hitchcock, 1888), but not proven until the 1960's with the advent of the microprobe (Hollister, 1965; Atherton and Edmunds, 1966). Recent petrofabric studies have depicted syn-deformational metamorphism as dynamic recrystallization of mineral assemblages under continuous reequilibration, rather than discreet, punctuated "metamorphic events" (Bell, 1981; Knipe and Wintsch, 1985). Continuous mineral growth and chemical reequilibration are manifested by compositional zoning of many minerals including garnet, sodic amphibole, plagioclase, tourmaline, pyroxene, chloritoid, biotite, and cordierite (Tracy, 1982; Tracy and others, 1990). Metamorphic mineral growth and syndeformational recrystallization are now understood to be a result of continuously evolving chemical reactions; the kinetics of which are controlled by several parameters including bulk composition, pressure, temperature, fluid fugacity, and oxygen fugacity (Tracy, 1982; Knipe and Wintsch, 1985).

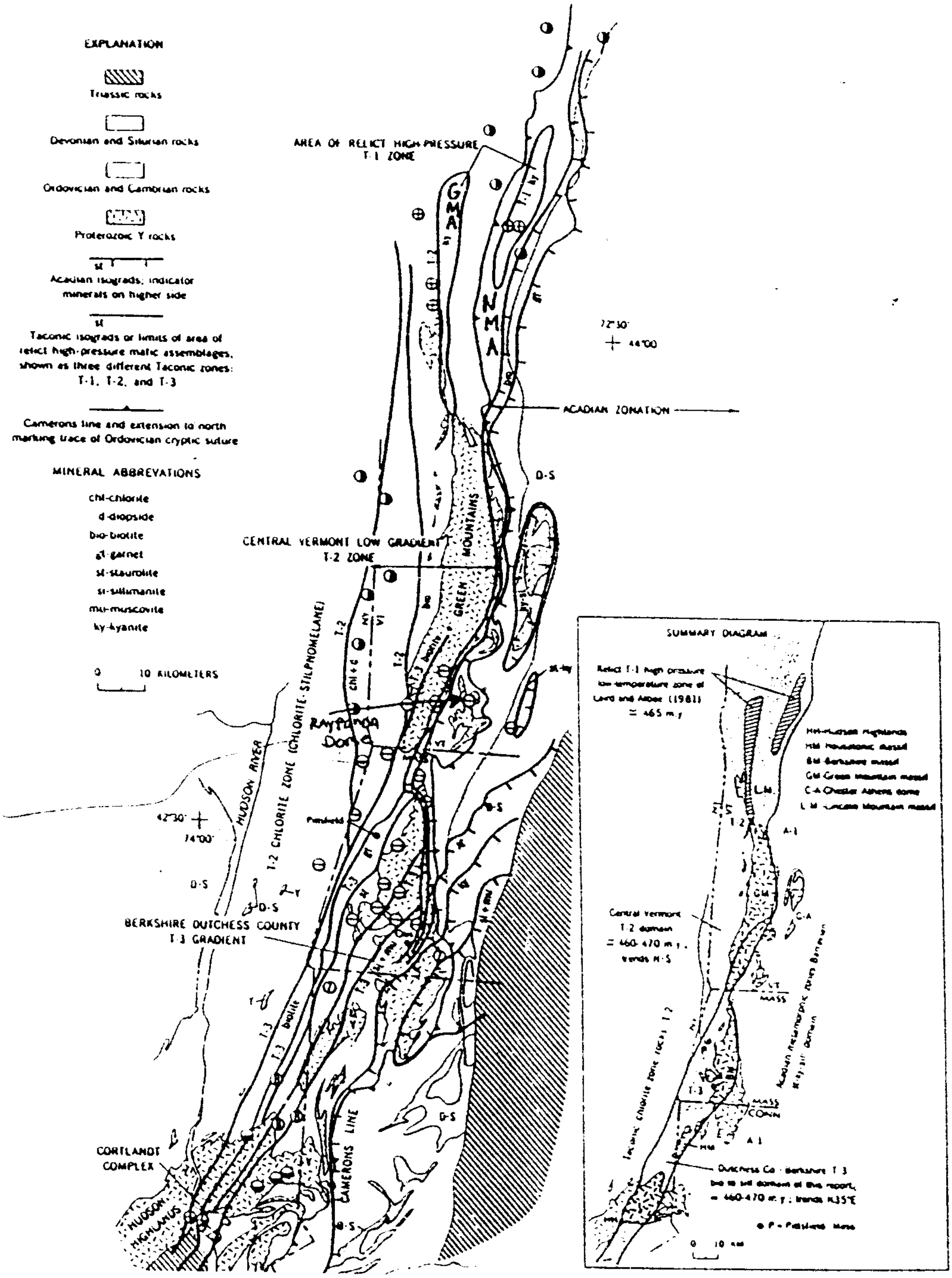


Figure 4.1 Taconian and Acadian metamorphic zones for western New England, including localities for K-Ar and Ar/Ar age analyses. Distribution of isograds is based upon the work of Laird and Albee (1981) and Laird et al (1984) for northern and central Vermont (figure from Sutter et al., 1985).

Causes for metamorphism in the hinterland and its time and space relations to deformation

The metasedimentary and metavolcanic rocks within the central Vermont sequence clearly contain several different mineral assemblages, each of which developed under different pressure, temperature, and, possibly, fluid conditions (Laird et al., 1984). High pressure-high temperature (Barrovian) metamorphism is generally regarded to be the result of thermal relaxation of perturbed geotherms by way of advection processes (England and Richardson, 1977; Thompson and England, 1984). The transgressive rate of metamorphism is therefore only as fast as the loading event itself. Metamorphism within the Vermont Taconian hinterland clearly post-dates the loading event (presumably related to the premetamorphic thrusting, pre Dn-1); this is because of the lag time between tectonic loading and the thermal advective processes (which usually is initiated as a conductive process; Thompson and England, 1984). Although this lag time may lead to disparate relative age relationships between metamorphism and specific deformation fabric development, the **magnitude** of metamorphism is usually uniform over widespread areas. Local variation in the intensive metamorphic (P-T) parameters can sometimes be caused by synmetamorphic deformation where extremely large-scale structures expose different crustal levels at a particular datum plane (e.g. the present day land surface). The landmark papers by England and Richardson (1977) and Thompson and England (1984) clearly point out that Barrovian metamorphism within an orogen does not occur as discreet, isolated pulses, but rather as a single thermobaric event that has four continuous phases that make up a "clock-wise" P-T path:

1. **Tectonic loading** - causing geologically instantaneous, isothermal compression (pressure increase).
2. **Thermal equilibration** - nearly isobaric heating over several million years (1 to 10's) toward peak thermal conditions. This is the prograde part of a P-T path.
3. **Tectonic denudation** - decompression during continued heating up to the peak thermal conditions. This indicates that peak pressure is attained before peak temperature; a common phenomena in metamorphic belts. This phase coincides with peak metamorphic conditions.
4. **Incipient cooling during continued denudation** - this is the retrograde part of the P-T path, or what some have referred to as a "remetamorphism" or retrograde metamorphism; it is crucial to note that this phase is directly related to the prograde metamorphism and is just an inevitable conclusion to a metamorphic cycle.

Because it is very difficult to place absolute age constraints on a single P-T path, most workers attempt to relate the relative age relationships between a metamorphic P-T path and various deformation events (Laird and Albee, 1981; Sutter et al., 1985). In conjunction with detailed structural analysis, a simple relative chronology of the various syndeformation mineral assemblages can be used to constrain interpretations of the structural and tectonic history of the pre-Silurian belt (chapter 3). Albee (1968a), Laird and Albee (1981), Laird and others (1984),Sutter and others (1985), Hames and others (1989), and Armstrong and Tracy (1991) have shown that particular mineral assemblages associated with specific metamorphic events show consistent relative age relationships to particular deformations over both local and regional extent. Specific, **radiometrically dated, synchronous** metamorphic mineral assemblages can therefore be **carefully** used in conjunction with deformation fabrics to distinguish between different phases of synmetamorphic deformations. Since ductile deformation fabrics are a **result** of dynamic recrystallization (a metamorphic process) these mineral assemblages (and related fabrics) can also be used in conjunction with numerous geothermobarometers in order to deduce pressure-temperature (P-T) conditions. This information can also be used to constrain the approximate P-T conditions under which the various synmetamorphic deformations developed which can then be synthesized into a model depicting the thermobaric and tectonic evolution of the studied area.

Diachroneity of metamorphism and deformation

In Vermont, deformation is generally believed to have been west directed during the Taconian orogeny (Stanley and Ratcliffe, 1985). Since metamorphism is the result of tectonic loading, it too must be diachronous, and therefore rocks within the western part of the hinterland **should** be of lower grade than rocks within the eastern part. The opposite relationships, however, are presently observed; rocks near the Green Mountain axis (anticlinorium) are Kyanite-chloritoid grade whereas rocks to the east, including the Northfield Mountain anitclinorium, are generally chlorite-grade (Laird et al., 1984; Sutter et al., 1985; Stanley et al., 1989). This is the case even though recent regional cross sections show that the Northfield anticlinorial structure is of much greater amplitude (figure 5.2), and the eastern part of the hinterland (Granville, Pinney Hollow,Stowe, and Ottauquechee Formations) were at deeper crustal levels than the western part prior to and during synmetamorphic deformation. Proposing a solution to this problem is one of the main objectives of this chapter.

Physical conditions during mineral growth

I. Mafic schist assemblages:

The use of mafic schists in deriving pressure and temperature conditions during metamorphism has only been recognized within recent years (Cooper, 1972; Myashiro, 1973; Graham, 1974; Harte and Graham, 1975; Laird and Albee, 1981). The usefulness of this rock-type in evaluating physical conditions of metamorphism stems from the fact that the mineral assemblage, commonly quartz-chlorite-plagioclase-epidote-amphibole ± K-mica ± Ti-phase, is present throughout greenschist, amphibolite, and blueschist facies series metamorphism (Miyashiro, 1961; Laird and Albee, 1981). It is the composition and modal variation of the minerals that changes with respect to changing physical conditions. The analysis of amphibole, chlorite, and plagioclase compositions within this common assemblage provides an estimate of pressure conditions during the development of the assemblage. Geothermometric analysis of coeval, and associated, pelitic assemblages provides constraint on temperature of metamorphism for the common mafic assemblage (Harte and Graham, 1975).

Amphibole: Amphiboles range in composition from actinolite in the biotite zone (as inferred from the coexisting pelitic assemblage) to ferro-magnesio-hornblende in the garnet zone to tschermakitic hornblende in the staurolite-kyanite zone (Leake, 1978). The general compositional variation is a product of three substitutions:

1. **Edenite** - $(Na+K)_A, Al^{iv}$ Si
2. **Tschermak** - $(Al^{vi}+Fe^{3+}+Ti+Cr), Al^{iv}$ (FeO+Mg+Mn)
3. **Glaucophane** - Na_{M4}, $(Al+Fe^{3+}+Ti+Cr)^{vi}$ Ca, $(Fe^{2+}+Mg+Mn)$

$Na_A, K, Na_{M4}, Al^{vi}, Ti, (Al^{vi}+Fe2O3+Ti+Cr)$,

and Al^{iv} content of actinolite increases continuously into the garnet zone. There is an increase in all of these constituents into the staurolite-kyanite zones, except for Na_A and K which decrease. The glaucophane and tsehermak substitutions therefore show a continuous increase with increasing metamorphic grade whereas the edenite substitution initially shows an increase into the garnet zone followed by a decrease into the staurolite-kyanite zones. With increasing grade, these substitutions produce a change from actinolite to hornblende as the amphibole within the common assemblage. The edenite substitution is favored under low pressure conditions, the tschermak under medium pressure, and glaucophane under high pressure conditions.

Plagioclase: Variation in plagioclase composition from albite to oligoclase is a result of phase separation at the peristerite gap (Ribbe, 1978; Smith, 1978). Under low pressure facies series metamorphism, the peristerite gap occurs within biotite grade assemblages with the sodium-rich component being consumed in the edenite amphibole substitution, leaving an oligoclase feldspar composition. Medium high pressure facies series metamorphism also contains the peristerite gap in garnet and higher grade assemblages. Lower grade assemblages have an albite plagioclase composition (An_{01} to An_{07}) with the calcic plagioclase component being consumed by the epidote, chlorite, and grossular reactions (Smith, 1978; R.J. Tracy, pers. comm., 1989). Under high pressure conditions, plagioclase is usually never more calcic than An 3%, the result of a continuous gap in plagioclase composition through all grades with the calcic component being fully consumed by production of epidote and/or calcite (Laird et al., 1984; Harte and Graham, 1975).

Composition diagrams:

The variation in mole percent Na, Al, K, Fe^{3+}, Ti, and Si, within different minerals can be plotted on composition diagrams. A graphical portrayal of the amphibole compositions found within Vermont mafic schists can be compared to those within well defined tectonic environments and formed under different physical conditions (fig 4.2). The pressure facies fields are drawn from modern and ancient orogenic belts where the pressure conditions and amphibole compositions are well constrained (Myashiro, 1973; see Laird and Albee, 1981a for discussion). Analyses of coexisting pelitic assemblages provide constraints on temperature of amphibole formation within the mafic rocks. The superposition of the metamorphic isograds on a pressure facies series composition diagram (illustrating P-T conditions) can be utilized in order to show the effects of both temperature and pressure on the composition of coexisting amphibole and plagioclase within the mafic schist (figure 4.2). Determination of plagioclase and amphibole composition by microprobe analysis allows the use of such a diagram for the quantifying of pressure and temperature conditions for that particular "phase" of mineral growth.

Mineral assemblages within mafic schist:

Analyses of mafic schist completed by Laird are from lithologies in central Vermont and show a general progressive decrease in pressure from east to west (figure 4.2; Laird and Albee, 1981; Laird et al., 1984). Individual amphibole grains sometimes show discontinuous zoning, indicating disequilibrium, between high to medium-high pressure facies series barroisitic hornblende cores and low to medium pressure facies series hornblende/actinolite rims. Although Laird and Albee (1981) argue that these discontinuous compositions were formed during a high pressure Taconian and low

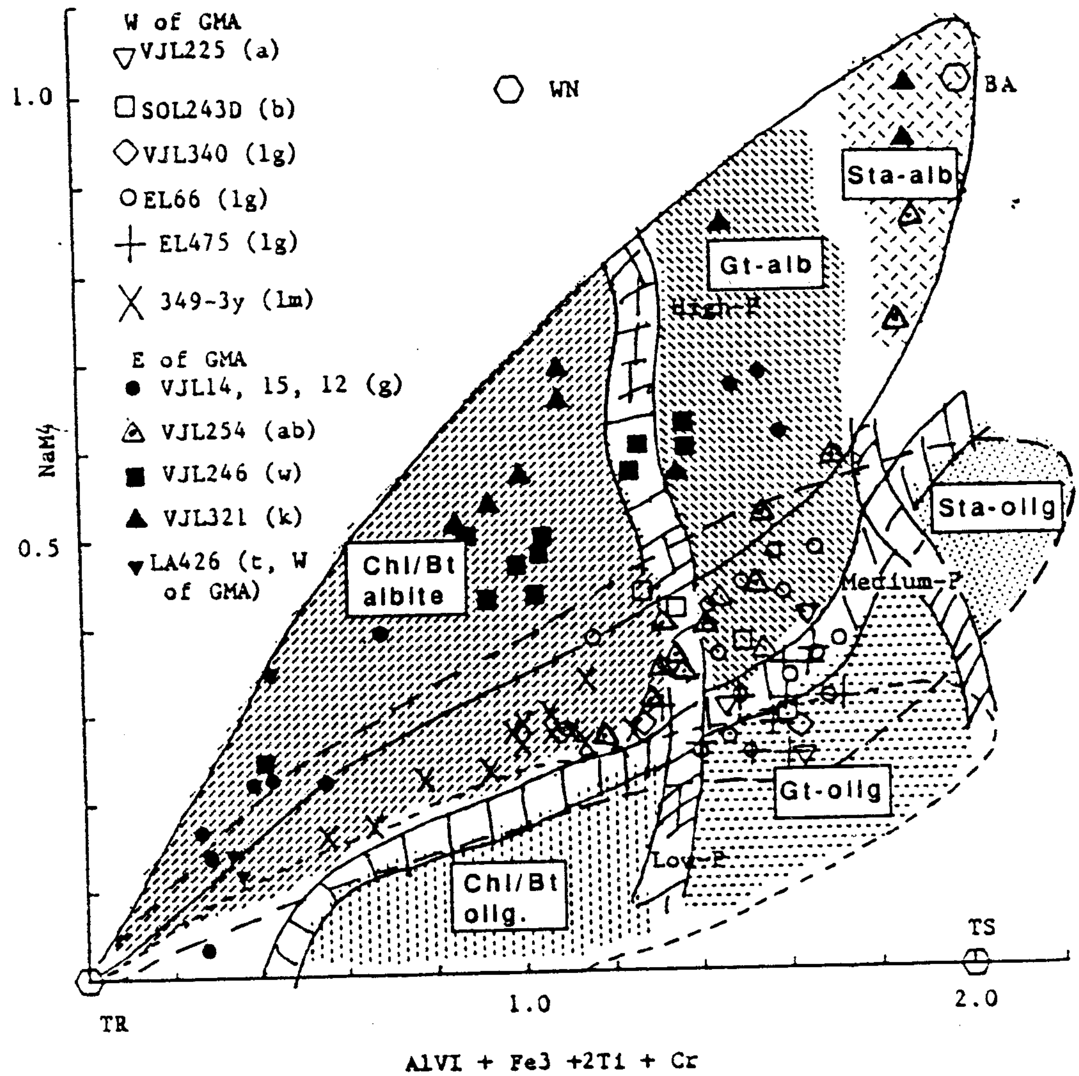


Figure 4.2 Pressure facies series diagram from Laird et al. (1984) showing variation in hornblende composition as a function of metamorphic pressure. The formula proportions on the ordinate and abscissa reflect different amounts of edenite, tschermak, and glaucophane substitutions at different metamorphic grades (see discussion in chapter 3). Samples in and near the study area include VJL254 (Allbee brook, Granville), VJL14 (Granville Gulf) and VJL246 (town of Warren). Isograds for pelitic assemblages, along with the plagioclase compositional isograd, are also displayed. Notice that the Allbee brook analyses consistently plot within the medium pressure, garnet-albite field whereas Warren and Granville Gulf samples show a fanning of points indicative of lower grade biotite-garnet grade transition at high pressure. This pattern may also be a consequence of differential cooling rates, with the lower T assemblages having slower cooling rates, and thus lower closure T. This diagram does not take into account the effects of chemical disequilibrium created by subsequent retrograde cooling, mineral zoning, twinning, or lamellae. Relict peak metamorphic pelite assemblages are absent or rare, thus the validity of variation in metamorphic grade cannot be locally tested.

pressure Acadian event, respectively (based on radiometric dates to be discussed in a later section), it is just as plausible that the actinolite-rich rims grew as a result of retrogression and reequilibration at lower P and T than the hornblende cores, during a cooling, decompression stage of a **single** Taconian metamorphism (Laird, pers. comm., 1990, 1991). The medium pressure facies series hornblende within rocks of the Hazen's Notch (Fayston) and Underhill Formations, west and northwest of the study area, are believed to have occurred within a garnet to kyanite grade environment.

Western Domain:

No greenstones or amphibolites were mapped within the major rock units in the western part of the Granville - Hancock area. In the Starksboro and Fayston areas, DiPietro (1982) and Walsh (1989) report significantly large bodies of amphibolite within the foliated metawacke of the Underhill (Monastery and Hoosac) Formation. Petrographic and microprobe analysis of amphibolite from the Starksboro area was conducted by Laird (1981). Amphibole composition and pelitic schist assemblages indicate medium pressure facies series and kyanite-chloritoid to garnet-oligoclase grade metamorphism (Laird and Albee, 1981). The hornblende, chlorite, biotite, epidote, carbonate, oligoclase mafic schist (greenstone) assemblage and the pelitic quartz, oligoclase, garnet, chlorite, white mica ± biotite ± Fe/Ti oxide assemblage were the oldest mineral phases recognized during studies in this region (DiPietro, 1982; Laird and Albee, 1981; Walsh, 1989). The observation of hornblende both overgrowing and being enveloped by the dominant foliation within the western region indicates that this medium pressure metamorphic event, M1, occurred during and after Sn-1 deformation (DiPietro, 1982; Walsh, 1989; Laird, pers. comm., 1989). Inclusions of epidote and plagioclase (of unknown composition) within many of the hornblende crystals indicate a still earlier phase of mineral growth, presumably during prograde conditions prior to the M1 peak conditions. Total fusion Ar^{39}/Ar^{40} analyses of M1w hornblende from the Starksboro amphibolite yielded a Taconian metamorphic age of 471Ma +5/-4 Ma (Laird and Albee, 1981).

Central Domain:

Petrographic analyses of mafic schist (greenstones) within the central domain show clear evidence of polymetamorphism; blue-green hornblende (barroisitic composition) shows various stages of resorption with strong overgrowths of chlorite without actinolite (figures 4.3 & 4.4). Weakly developed Snc foliation wraps around hornblende porphyroclasts and is defined by 1 to 2mm lathes of chlorite. Relict hornblende, plagioclase, epidote, rutile,

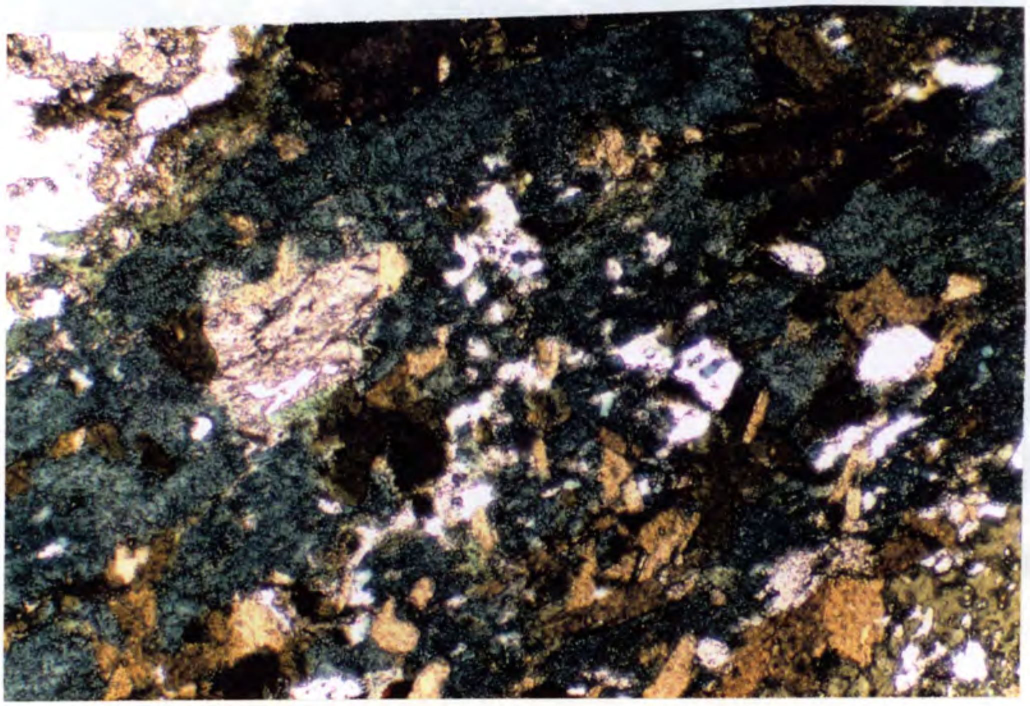


Figure 4.3 Photomicrograph of barroisitic (blue-green) hornblende from HQ 3, along Route 55, west of the town of Granville (plate 1; F 28). Hornblende from this locality is relatively pristine with little or no apparent retrogression; this may be due to the mechanical strength of the greenstone which does not accommodate Snc fabric nor coeval M2 retrograde metamorphism. Fabric within this rock is Snc-1 compositional layering. View is 2.4mm across.

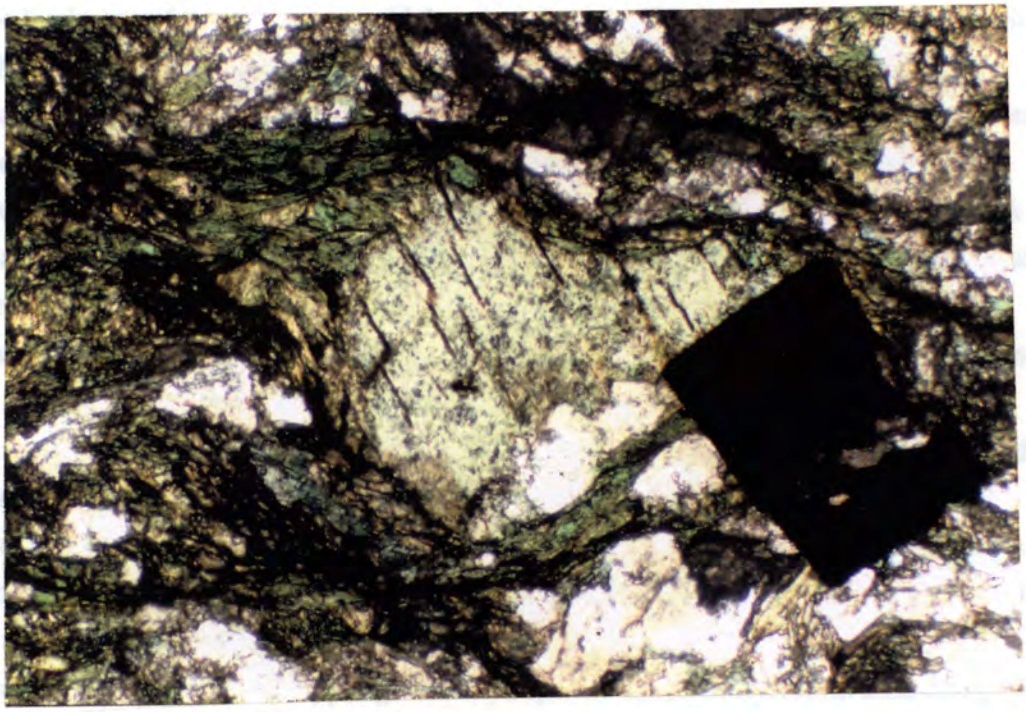


Figure 4.4 Photomicrograph of greenstone from Granville Thrust Zone at Allbee Brook (locality C, figure 3.12). Hornblende is highly resorbed, with chlorite overgrowth forming a pseudomorph texture. Well developed fabric is Snc mylonitic foliation. Hornblende is probably grew during M1 prograde metamorphism; chlorite overgrowth appears to have grown during M2 retrograde metamorphism, coeval with Snc. Chlorite pseudomorphs after hornblende are common in the central and eastern domains where Sn deformation is pervasive. Sample HQ 254B; view is 1.2mm across.

and ilmenite define the Snc-1 compositional layering. The magnitude of hornblende resorption appears to be related to the intensity of Snc foliation; greenstones that display a strong Snc fabric contain only chlorite pseudomorphs after hornblende (figure 4.4) whereas greenstones with weakly developed Snc generally contain relict hornblende that shows only moderate resorption and chlorite overgrowth (figure 4.3). The intensity of Snc, in turn, appears to be controlled by two major factors:

1. The thickness of the greenstone; thick greenstones tend to have weaker developed Snc than thinner ones. This is probably a function of the mechanical competency of the greenstones which tends to protect the internal parts from strain, related recrystallization, and fluid infiltration, necessary for chemical reequilibration and retrogression.
2. The proximity of the greenstone in regard to synmetamorphic fault zones; even thick greenstones (such as the ones at Allbee Brook) are susceptible to some Sne recrystallization in areas of extremely high stress and strain. Strain gradients within these greenstones, however, still appears to be large since recrystallization tends to decrease quickly across-strike and away from the specific fault zone(s) (compare figures 4.3 & 4.4).

Laird and Albee (1981) interpreted the blue-green hornblende at Allbee Brook (the Central Domain) to be the medium-high pressure equivalent of hornblende-bearing greenstones farther west (Western Domain; figure 4.2). $^{40}Ar/^{39}Ar$ and $^{40}K/^{39}Ar$ ages from blue-green (barroisitic) hornblende from Granville Gulf (immediately north and along strike of the Central Domain) yield an age of 471 Ma (Laird et al., 1984; Laird, 1987), supporting correlation of this upper greenschist/epidote-amphibolite facies assemblage with the M1 assemblage within the Underhill Formation. The segregation of hornblende, plagioclase, and opaques within the Sn-1 compositonal layering suggests that this central domain M1 metamorphism occurred either prior to and/or during Sn-1 deformation, similar to the relative age of M1 (with respect to the first synmetamorphic deformation) in the western Domain (DiPietro, 1982; Walsh, 1989). Subsequent retrogression including the development of chlorite overgrowths, secondary epidote, and possibly secondary ilmenite occurred prior to and/or during Snc fabric development. This relative age relationship (between secondary mineral growth and the second observed synmetamorphic deformation) is also present within the rocks to the west (Laird, pers. comm., 1989, 1990; DiPietro, 1982; Walsh, 1989).

Eastern Domain:

M1 hornblende within the eastern domain is rare and was found as small relict grains within only two different greenstones, both of which were over 400m thick (plate 1, figure

4.5; greenstones, CZphg, along Route 100). Chlorite overgrowths upon this grains are pervasive; chlorite is commonly found within most of the thick greenstones (> 100m) as spheroidal clots which probably developed as pseudomorphs after hornblende (figure 4.6). These greenstones contain a generally well developed, although somewhat wavy Sne synmetamorphic foliation in which rare vestiges of Sne-1 fabric may be observed (figure 4.6). Thinner greenstones (5 to 100m thick) tend to have a pervasive Sne fabric in which chlorite clots have been deformed and transposed along with nearly all Sne-1 compositional layering (figure 4.7). These observations are similar to those made within the central domain; retrogression of the primary hornblende-bearing assemblage is a function of the thickness of the greenstone and the magnitude of the subsequent Sn (and Sn+1) strain. The abundance of Sne and Sne+1 mylonitic fault zones within this domain, along with the abundance of mechanically weak graphitic phyllite, suggests that greenstones and albitic schist probably accommodated most of the strain within this domain that led to a more complete retrogression and reequilibration of the hornblende-bearing assemblage. The relative age relationships between supposed primary hornblende, overgrowths of secondary chlorite, and Sne deformation, indicate that primary (prograde) and secondary (retrograde) metamorphism within the eastern domain is correlative with primary (M1) and secondary (M2) metamorphism within the central and western domains.

II. Pelitic schist assemblages

M1 mineral assemblages:

The presence or absence of prograde pelitic minerals such as garnet,chloritoid, biotite, staurolite, kyanite, muscovite (paragonite), plagioclase, and Fe-Ti oxides, depends on the pressure and temperature of the system, and the rock bulk composition, graphically portrayable on a muscovite, quartz, water projected aluminum-iron-magnesium (AFM) ternary diagram (figure 4.8a). For example, garnet grade rocks which plot below the garnet-chlorite tie lines (the garnet-chlorite join) will contain chlorite and biotite, biotite and garnet, or garnet, biotite, and chlorite depending on the constituent mineral Fe/Mg ratios and on the amount of other components (such as Mn and Ca) within any or all of the particular (Graham, 1974; Thompson, 1957; Albee, 1965a). Garnet grade aluminous rocks, plotting above the join, will have; 1) garnet and chlorite, 2) garnet and chloritoid, 3) chlorite and chloritoid, or 4) garnet, chlorite, and chloritoid (figure 4.8a). The three phase assemblage kyanite, chloritoid, and chlorite would be present in rocks with higher alumina content and at higher grade; at garnet grade, pyrophyllite would supposedly substitute for

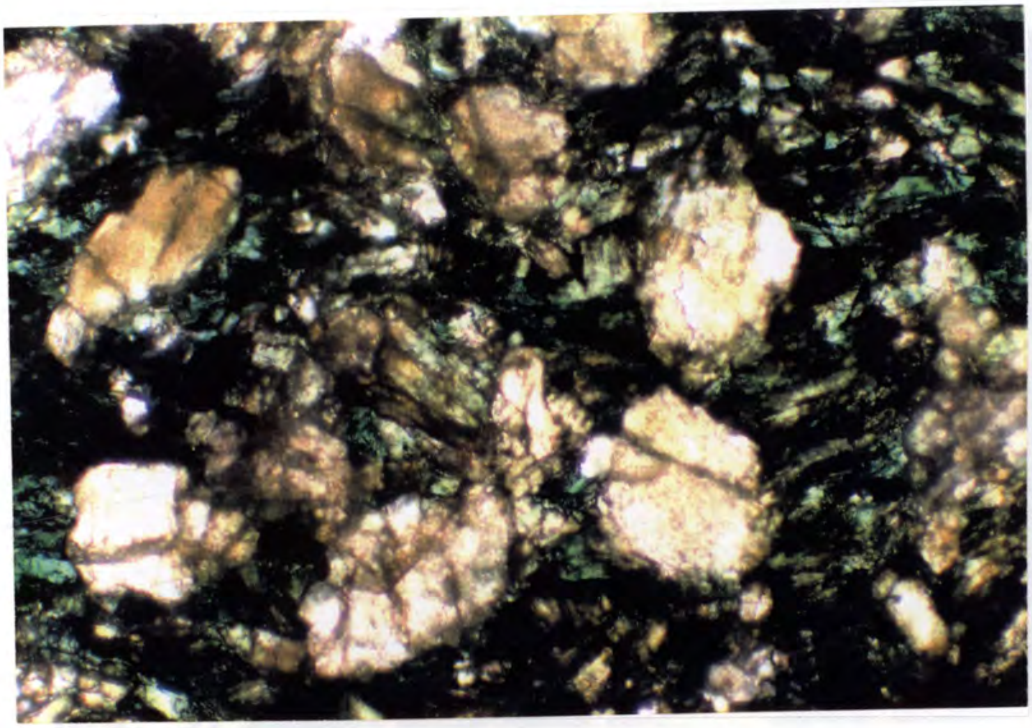


Figure 4.5 Photomicrograph of small (0.05-0.1mm) fragments of M1 hornblende which have been replaced by M2 chlorite during syndeformational (Sn) retrograde metamorphism. Two samples (including this) from greenstones over 400 feet thick were the only ones that had observable relict hornblende. Thinner greenstones have well developed Sn fabric and chlorite has entirely replaced hornblende (figure 4.6). Sample HQ 132; Clark brook (plate 1; J 34). View is 1.2mm across.

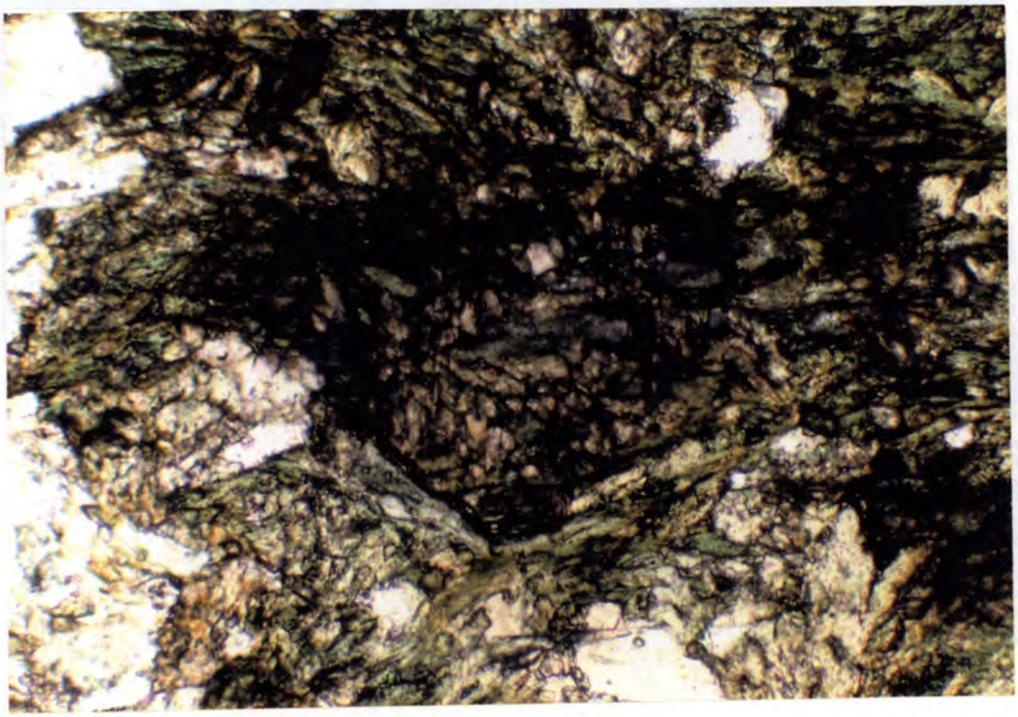


Figure 4.6 Photomicrograph of M2 chlorite pseudomoph presumably after M1 hornblende; no relict hornblende was observed. Sample contains a pervasive Sn fabric parallel to the top edge of the photograph. Relict Sn-1 is locally present as small, isolated fold hinges within the center of the pseudomorph whose axial surfaces dip moderately to the right. Sample from Pinney Hollow greenstone (HQ 461; plate 1 T 41). View is 1.2mm across.

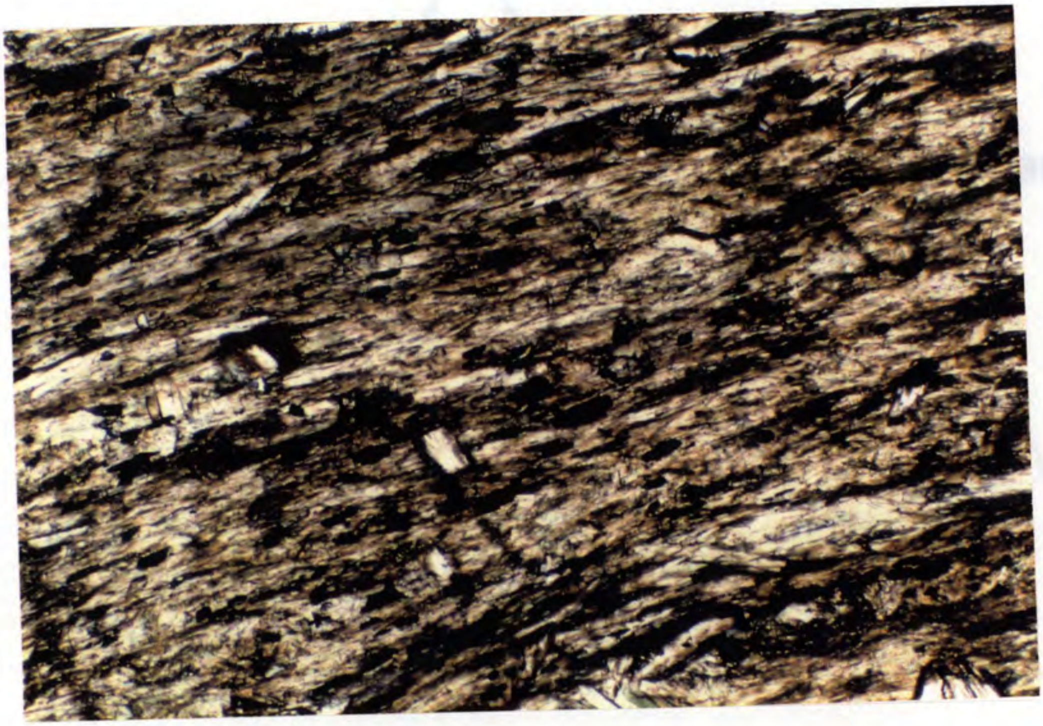
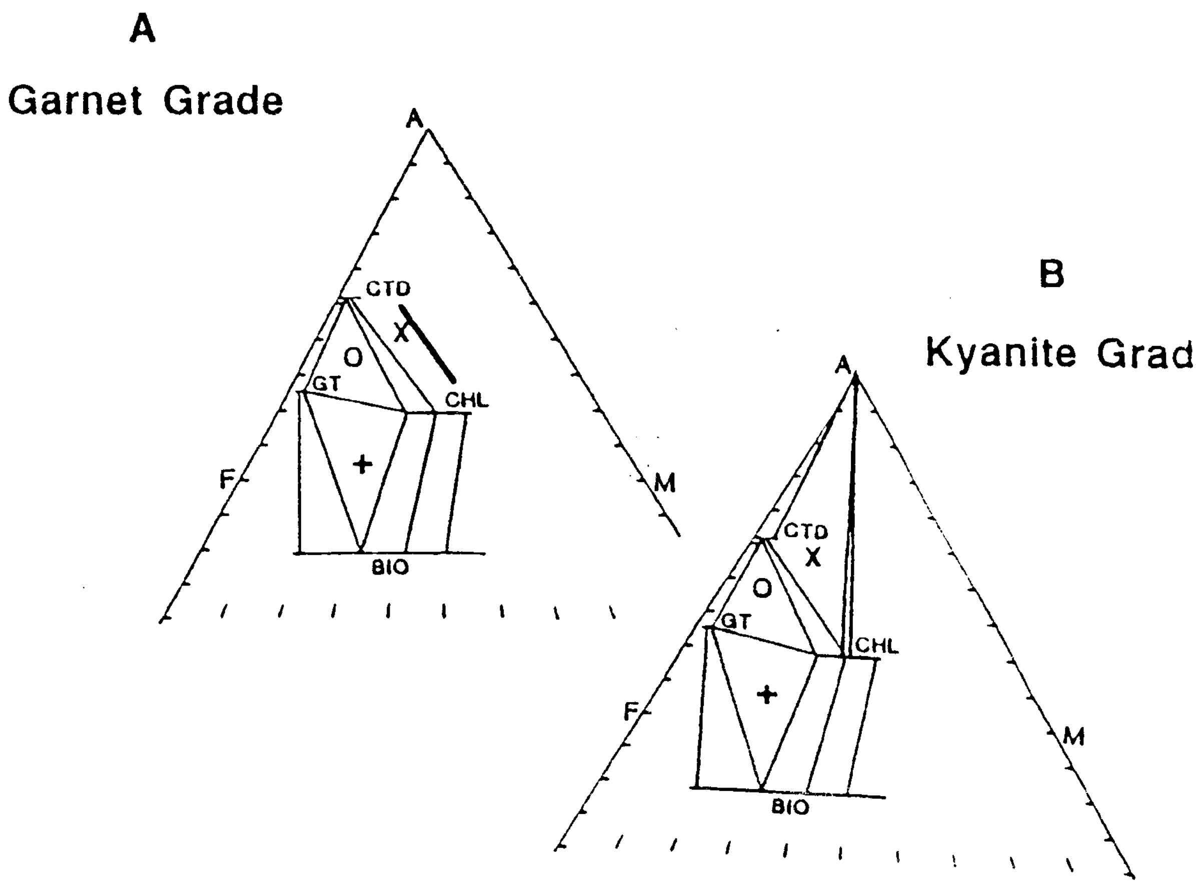


Figure 4.7 Photomicrograph of a Stowe greenstone (HQ 437; plate 1; W 42) with pervasive Sne fabric. M2 chlorite is abundant without any relict M1 hornblende. Chlorite pseudomorphs are not present possibly due to complete transposition into the Sne schistosity. Segregations of epidote and chlorite within Sne gives this greenstone a gneissic texture. View is 2.4mm across.



Symbols

X - Mt. Abraham Fm.; chloritoid-chlorite(A) and kyanite-chloritoid-chlorite(B) assemblages

O - Mt. Abraham and Monastery Fms.; garnet-chloritoid-(Fe)chlorite assemblages (also present in some of the Pinney Hollow Fm)

\+ - Monastery Fm. garnet-chlorite-biotite assemblages

Figure 4.8a AFM diagram projected through muscovite with quartz and H_2O as AKFM saturating phases for rocks at medium to medium-high pressures and at **A:** Garnet grade, and **B:** Kyanite grade. Notice that the minerals occurring in the rock assemblage are directly dependent upon both metamorphic grade and bulk composition. Most rocks within the eastern cover sequence are relatively high in alumina since they generally do not contain biotite; Aluminous bulk compositions include Mt. Abraham schist and aluminous parts of the Monastery Fm (X and O symbols), and most of the Pinney Hollow Formation. Biotite-bearing rocks are found within the less aluminous layers of Monastery (+ symbol) and the majority of the Fayston Fm., to the north (Walsh, 1989). Staurolite (and typical staurolite grade rocks) are absent in this region because of the increased pressure; kyanite is stable at lower temperatures than staurolite for this particular pressure range.

kyanite although reports of pyrophyllite within metamorphic rocks is rare (Thompson, 1957; Albee, 1965a). The presence (or absence) of garnet, chlorite, chloritoid, biotite, and kyanite is a function of variation in bulk composition and intensive parameters such as temperature, pressure, and chemical potential of the fluid phase (Albee, 1965a; Thompson et al., 1977). Determining whether a particular garnet-bearing assemblage (with a bulk composition too low in alumina to develop an aluminosilicate phase) grew at higher than garnet grade requires either the presence of a coeval aluminosilicate phase within a aluminous lithology near the garnet-bearing lithology, or detailed microprobe analyses to determine exact Fe/Mg ratios amongst garnet and biotite (Tracy et al., 1990), thermobarometric calculations in order to determine the P-T conditions of mineral formation, and extrapolation of this information onto a petrogenetic grid in order to determine if kyanite stability (or that of another phase) is possible).

Western Domain:

Estimates of rock bulk compositions within the western domain were determined by petrographic identification of coexisting AFM phases (figure 4.8a). Coexisting M1 chlorite and chloritoid, chlorite and garnet, and garnet, chlorite, and chloritoid were all found during petrographic analysis of different rocks from the western domain. These assemblages are constrained to the divariant and univariant fields of the quartz-muscovite-water projected AFM ternary diagram (figure 4.8a). All of these assemblages have relatively high alumina and iron compositions, reasonable for garnet grade conditions. No kyanite or staurolite was found within the study area even in highly aluminous muscovite-paragonite-bearing schist, indicating that metamorphic conditions were sub-kyanite grade. Kyanite-chlorite-chloritoid assemblages are present immediately north of the study area at Mt. Grant, in the Lincoln Mountain 7.5 minute quadrangle (Albee, 1965a; Lapp and Stanley, 1987). Kyanite does not coexist with garnet in this area; although this may suggest that the absence of kyanite in this study area is the result of bulk compositions that are too Fe-rich (within the garnet-chloritoid-chlorite field rather than the chlorite-chloritoid-kyanite field, figure 4.8a). The kyanite-bearing rock units that are found within the Mt. Grant area, however, are found along strike, and to the south within this study (Mt. Abraham Schist). It is therefore more likely that metamorphic conditions were slightly lower in this area rather than implementing a wholesale change of bulk composition of a contiguous rock unit. The lack of staurolite in these, and particularly, the Mt.Grant lithologies is probably the result of extra phases, particularly Mn which would displace the projection of staurolite (onto the AFM) toward the A-M join, to more Mg-rich compositions than the kyanite-chloritoid-chlorite field (figure 4.8a; Thompson et al., 1977; Armstrong and Tracy, 1991).

M2 Mineral assemblages:

M1 garnet-bearing assemblages in the western domain occasionally have overgrowths of chlorite, chloritoid, and quartz on M1 garnet (figure 3.4). In areas where Snw+1 foliation is present and defined by aligned M2 sericite and chlorite grains, M1 garnet is subhedral to anhedral (figures 3.2-3.5). Secondary growth of Nw+1 age chlorite and sericite (M2) does not include garnet development. M2 growth is therefore associated with M2 retrograde metamorphism at lower P/T conditions than M1. A probable retrograde reaction involving the breakdown of garnet to form chlorite, chloritoid, and quartz, requires the addition of water (Armstrong and Tracy, 1991):

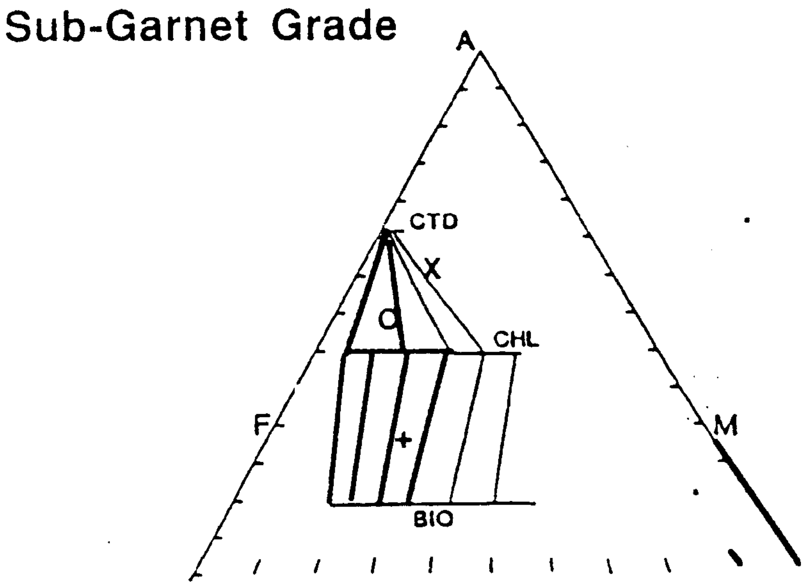
Garnet + Water = Chlorite + Chloritoid + Quartz

Variation in pressure may or may not be significant, due to a lack of precise geobarometers for M2 conditions. Temperature during M2 must be substantially lower than garnet grade conditions (500°C; based on muscovite - paragonite solid solution compositions and calcite-dolomite thermometry of Laird et al., 1984). Thin section analyses indicate that M2 retrograde metamorphism progressively increases from west to east across the western domain (e.g. figures 3.2 - 3.5). The alignment of chlorite and sericite within Snw+1 foliation, coupled with some sericite, chlorite, chloritoid, and quartz static overgrowth indicate that M2 occurred during and after Snw+1 deformation in this area.

Central and Eastern Domains:

M1 Mineral assemblages:

Equilibrium, garnet grade mineral assemblages are absent within both the central and eastern domains. Garnet is only locally found within the matrix of several analyzed samples, which have similar bulk compositions to the garnet-bearing rocks in the western domain (Osberg, 1952; figures 3.5 & 4.8b). All of these samples occur within central domain near the boundary with the western domain. Snc-1 and Sne-1 compositional layering in the pelites is generally preserved within the hinge regions of Fn folds (figure 3.18). Photomicrographs of this layering show that it consists almost exclusively of recrystallized quartz and plagioclase grains which are rimmed by fine grains of sericite and chlorite (figures 3.2 - 3.5). Relict garnet, indicative of M1 prograde metamorphism has either been completely replaced or was never present.



Symbols

X - Mt. Abraham Fm.; chloritoid-chlorite assemblage
O - Mt. Abraham and Monastery Fms.; chloritoid-(Fe)chlorite assemblage (also present in some of the Pinney Hollow Fm)
\+ - Monastery Fm. chlorite-biotite assemblage

Figure 4.8b AFM projection through muscovite (with quartz and H_20 as AKFM saturating phases) at sub garnet grade conditions. At high pressures (6-14 kbar) pyrophyllite ("A" apex) may be stable although it has not been recognized in this region. The aluminous assemblages (O and X symbols) also pertain to the majority of the Pinney Hollow and Stowe Formations within the eastern domain. In southern Vermont, the Stowe Formation also contains biotite, and thus its bulk composition may lie within the chlorite-biotite divariant field (+ symbol). The lack of biotite in the Stowe in this region is probably indicative of sub-biotite grade conditions during M2.

Small (1mm), euhedral garnets do occur within the Pinney Hollow Formation at Allbee Brook. These garnets have a lavender color and overgrow the dominant Snc and shear band foliation. Microprobe analyses of these garnets show that they are rich in spessartine (Mn) and grossular (Ca) components, whose stability field is significantly lower than most Fe-Mg garnets found in pelites (figure 4.9; Powell and Holland, 1990). The origin of these Mn/Ca-rich garnets may in part be due to the influx of Ca into the Allbee Brook mylonitic schist (the Granville Thrust Zone) from the breakdown of epidote and/or calcite within the adjacent greenstone. The Mn component may be indigenous to the rock; several outcrop exposures of Pinney Hollow schist, along Route 100 in Granville Gulf (immediately north of field area) contain thin (< 1m) quartzite horizons with garnet-rich layers. Similar sizes of garnet that cross cut the Sn and Sn+1 foliation are found within the Rochester area, and are believed to be Acadian in age (Laird et al., 1984; Laird, pers. comm., 1990). Relict garnet has also been reported as inclusions in plagioclase porphyroblasts from the Northfield Mountains (Kraus, 1989) and as relict, anhedral grains within metawacke of the Pinney Hollow Formation (Prewitt, 1989). These rocks all lie to the north, along strike of the eastern and central domains, respectively. Each presumably represents a relict vestige of the garnet grade M1 metamorphism; the lack of garnet outside of the plagioclase porphyroblasts (Northfield Mountains) and the anhedral shape of garnets within the metawacke indicate a state of disequilibrium between relict garnet and the matrix minerals, presumably of the retrograde M2 metamorphism.

Of particular exception are garnet porphyroblasts that overgrow the northeast trending crenulation cleavage in the Rochester area (R.S. Stanley, pers. comm., 1991). Small (0.5 to 1.0 mm) euhedral garnets have been reported in this area by Laird and others (1982), and are believed to represent the northwestern extent of the Acadian garnet-grade metamorphism (Jo Laird, pers. comm., 1991; Thomas Menard, pers. comm., 1991).

M2 Mineral assemblages:

Clots or streaks of chlorite (presumably after garnet), growing within the Snc and Sne schistosity, are quite common and suggests that thermal reequilibration of M1 assemblages during lower grade M2 was more complete from west to east. The lack of biotite, garnet, or even chloritoid from any of these rocks, regardless of bulk composition, indicates that the metamorphism which occurred either before and during Snc and Sne deformation, was sub biotite grade. The presence of biotite within the Sn fabric in the Warren, Waitsfield, and Fayston areas does suggest that metamorphic grade was slightly higher to the north

	X 16
SiO2	37.39
TiO2	0.32
Al2O3	20.14
Fe2O3	0.45
FeO	5.43
MnO	25.67
MgO	1.44
CaO	8.78
Total	99.62
	based on 24 Oxygens
Si	6.00
Al	3.91
Fe3+	.048
Ti	.039
Mg	.322
Fe2+	.740
Mn	3.534
Ca	1.478
Almandine	.120
Grossular	.22
Spessartine	.61
Pyrope	.05

Figure 4.9 Microprobe analyses of small, post Snc garnet from the Granville Thrust Zone at Allbee Brook (plate 1; L 30). Analyses were conducted on the Cameca Sx-50 probe at Virginia Tech (4-17-90). Sample current was 40na; accelerating potential was 15kv; beam resolution was 1.12 microns at 300,000x.

(Haydock, 1989; Prewitt, 1989; Walsh, 1989). Since the breakdown of garnet and the production of chlorite is a water consuming reaction:

$$30 \text{ Gar} + 3 \text{ Bio} + 84 \text{ H2O} = 14 \text{ Chl} + 3 \text{ Mus} + 33 \text{ Qtz}$$

Rocks lacking in biotite may include an Fe-Mg exchange reaction of phengitic muscovite in place of a biotite-muscovite net transfer reaction (Ghent and Stout, 1981). The production of the large amount of quartz agrees well with field observations of abundant small scale (1 to 10cm long by 0.5 to 3 cm wide) quartz veins which either cross cut or are transposed into the Snc and Sne schistosity. Such quartz veins are much more abundant in rocks of the central and eastern domains than those of the western domain.

Western Domain:

The M2 retrograde event seen in the central and eastern domains appears to be the same as the M2 farther west. M1 chloritoid and garnet within this domain are resorbed and overgrown by chlorite in various amounts with garnet found as subhedral to anhedral fragments and as relict cores with well developed M2 chlorite and quartz overgrowth; this growth sequence is similar to that for the other domains (figures 3.2 - 3.5). Quartz veins are also more abundant in areas with pervasive garnet retrogression. This retrogression is not well recognized in the western part of the domain; retrogression first appears and becomes pervasive within the central part of the domain where Snw+1 shear zones and locally pervasive Snw+1 crenulation cleavage occur (figures 3.2-3.5). Retrogression of garnet and its termination with respect to chlorite overgrowths occurs within the eastern limit of the western domain where Snw+1 crenulation foliation appears to either give way or transform into a dominant schistosity similar to Snc. The lack of a significant M1 assemblage in this area may attest to the eastward progression of M2 intensity and coeval synmetamorphic deformation.

Thermobarometry

Coexisting M1 muscovite and paragonite (sericite) compositions from the Mt. Abraham Schist (Monastery Formation of Osberg, 1952) were determined by Osberg (1952). When plotted on a calibrated muscovite-paragonite solvus curve, these analyses yield a mean temperature of 500°C. Although systematic problems do exist with this geothermometer, including closure temperature of Na-K exchange and phengite component activity within muscovite, this estimate corresponds well with temperatures derived from both coexisting M1 calcite - dolomite pairs in local mafic schist (Laird and Albee, 1981; 489°C) and amphibole-plagioclase geothermometry (460°C; Laird and Albee, 1981).

No thermobarometric data has yet been reported for the M2 retrograde metamorphism; this is primarily a consequence of a lack of mineral types amenable to net transfer and exchange thermometry and barometry (Essene, 1982). Based upon the fact that M2 is a garnet consuming, chlorite producing, water consuming event, temperatures around 400-475°C and pressures of 4.5 to 6 Kb would be likely (Spear and Cheney, 1989; Powell and Holland, 1990).

Distribution of metamorphic isograds

Sutter and others (1985) show their interpretation of the distribution of Taconian and Acadian isograds based on a synthesis of previous work within the northern Taconide zone in western New England and eastern New York (figure 4.1; Bence and McLelland, 1976; Cady, 1969; Dallmeyer, 1975; Dallmeyer and Sutter, 1976; Harper, 1968; Laird and Albee, 1981; Laird et al., 1984; Lanphere and Albee, 1974; Zartman and others, 1970). This interpretation shows the distribution of three distinct Taconian thermal phases; the high pressure, low to medium temperature T1 Barrovian assemblage, believed to be related to subduction zone metamorphism, T2 low gradient Barrovian metamorphism, a result of early thrust stacking of low grade (chlorite-garnet grade) rocks (with subsequent equilibration and thermal homogenization) of the pre-Silurian rocks within a westward advancing accretionary wedge, and T3 high gradient Barrovian metamorphism, a product of continued thrust stacking of basement rocks over westward situated sedimentary cover (Ratcliffe, 1975; Stanley and Ratcliffe, 1985; Hames et al., 1990). The Acadian isograds, related to a Barrovian low gradient, garnet-sillimanite / garnet-kyanite metamorphic event, extend west only as far as the Lower Ordovician Missisquoi Group lithologies in central Vermont. This interpretation is in close agreement with detailed field and petrologic studies completed and presently underway within the region (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Prewitt, 1989; Walsh, 1989). The T2 isograds are consistent with the M2 chlorite-biotite grade Barrovian assemblages found on the eastern limb of the Green Mountain anticlinorium. The earlier, Taconian Barrovian M1 garnet and kyanite-chloritoid assemblages within the Mt. Grant and Northfield Mountain areas (figure 4.1; T1) are probably relict M1 zones, predating the retrograde M2 (Sutter et al.'s. T2). T1 metamorphism would also be correlative with the relict medium and medium-high pressure amphibole cores, which probably formed under epidote-amphibolite facies conditions within a subduction zone environment or underneath a stack of pre-M1 thrust slices. Relict M1 cores occur within many of the greenstones within the central and eastern domains, overprinted by the later retrograde M2 (T2). Further detailed petrologic analyses may help to better document the areal extent of relict T1 phase. Because its areal extent is directly related to local magnitude of M2 reequilibration, a definitive delineation of T1 and T2 assemblages (and isograds) is a difficult task.

The variation in T2 metamorphic grade from central to southern Vermont may be related to the differential uplift of the various lithotectonic thrust slices within this region. There is no evidence for the T3 event within central Vermont since westward uplift and displacement of basement appears to be minimal in this region (Warren, 1989). Displacement of hot basement thrust slices over cool western cover rocks was more intense within the area of the Berkshire massif (figure 4.1). The deformed T2 and T3 isograd geometry corresponds to the deformed geometry of pre-M1 thrust slices and the inferred promontory of the North American continental margin prior to Taconian collision (Stanley and Ratcliffe, 1985; Armstrong and Colpron, 1989a, 1989b).

In an alternative petrologic and radiometric analysis, Laird and others (1984) discussed the distribution of Ordovician and Devonian metamorphism throughout northern and central Vermont. Ar/Ar total fusion ages from amphibole and muscovite in pre-Silurian lithologies produced ages of mineral growth (and argon diffusion closure) ranging from 471 to 385 Ma. Amphibole phase equilibria indicated medium high pressure facies series for Taconian metamorphism (M1) from the northern part of the Green Mountain axis (GMA), medium-high to high pressure (M1) along the eastern part of the GMA, and medium pressure (M1) conditions farther to the east within northern Vermont.

Medium to medium-high pressure M1 facies series rocks are at kyanite and garnet grade along the Northfield Mountain anticlinorium (NMA) in the north-central part of Vermont and along the GMA within the Mt. Grant area (figure 4.1). The rest of central Vermont is regionally metamorphosed to sub biotite grade by M1 (western sequence) and subsequent M2 in the eastern belt of medium high to high pressure M1. Medium pressure M2 and/or M3 (Acadian) facies series metamorphism is extensive along the northern part of the GMA, and appears to locally overprint the Ordovician M1 high pressure metamorphic distribution within central Vermont, specifically along the GMA. K/Ar, Ar/Ar total fusion ages of muscovite, biotite, and hornblende yield ages form 386 Ma to 355 Ma (Sutter et al., 1985). These ages were derived from samples within the kyanite-garnet isograds at Mt. Grant along the GMA (figure 4.1). These ages are somewhat similar to those derived from the lower grade Silurian-Devonian sequence in the Connecticut Valley-Gaspe synclinorium (CVGS; Harper, 1968). According to this interpretation, a significant part of the pre-Silurian sequence must have been regionally metamorphosed and thus deformed during the Acadian orogenic event (Laird et al., 1984). This is in disagreement with extensive field studies by various workers within central Vermont whom suggest that metamorphism producing the kyanite-chloritoid (?) and garnet grade M1, and subsequent M2 assemblages along the GMA, occurred during the Taconian orogeny (Stanley and Ratcliffe, 1985; Lapp and Stanley, 1987; O'Loughlin and Stanley, 1986; Stanley and others, 1987a, 1987b). It should be pointed that the majority of rocks within the Mt. Grant region and the western part of the Granville - Hancock area, do have a static overgrowth of muscovite, chlorite, and locally, biotite. This static growth post dates the syndeformational M2 metamorphism. The relative age of Laird's Mt. Grant biotite and muscovite to D1 and D2 deformation has not been properly documented. Amphibole from mafic rocks within this area is certainly M1 in age (Tauvers, 1982; DiPietro, 1982).

RADIOMETRIC ANALYSES

A synthesis of all radiometric analyses calculated within central Vermont was compiled and discussed by Sutter and others (1985; figure 4.10). This compilation emphasized the fact that $^{40}K/^{39}Ar$ and $^{40}Ar/^{39}Ar$ mineral analysis result in ages corresponding to the time of closure for that particular mineral, with respect to argon, rather than the time of crystal growth (Dodson, 1973; McDougall and Harrison, 1988). The approximate closure temperatures for biotite, muscovite, and hornblende within a slow cooling environment (~2-5°C/Ma) are 260°C, 320°C, and 480°C, respectively (Dodson, 1973, 1979). A muscovite, formed at garnet to kyanite grade (~500°C), would thus take 90 Ma in order to reach its closure temperature under cooling conditions estimated by Sutter (pers. comm., 1989) for Taconian metamorphism within this region (ie:~2°C/Ma). The measured age would therefore be quite younger than the actual age of metamorphism. In order to calculate the time of crystal growth, it would be necessary to analyze a specific mineral which formed under conditions near its argon closure temperature.

Muscovite and "hornblende" samples from the Mt. Grant area yield $^{40}K/^{39}Ar$ total fusion ages of 376 and 386 Ma, respectively (Laird et al., 1984). Since the assemblage is within the kyanite-chloritoid isograd, temperature of formation for the two phases was probably between 500°C to 550°C (Spear and Cheney, 1989; Powell and Holland, 1990). Assuming a metamorphic temperature of 525°C, a cooling rate of 2°C/Ma, and a closure temperature of 320°C, the time of muscovite formation would have been ~478 Ma, slightly high, but indicative of regional Taconian metamorphism within this region. The hornblende formation, assuming a closure temperature of ~480C, should have been ~408.5 Ma, which is not in close approximation to either Taconian or Acadian metamorphism as delineated by other radiometric dates within this region (Laird and others, 1984; Harper, 1968). This spurious measured age may be the result of impurities within the hornblende, or the analysis of an actinolite phase, commonly developed as a discontinuous overgrowth during conditions of lower pressure (Sutter, pers. comm., 1989).

Sutter's interpretation of the muscovite date as the age of cooling from Taconian metamorphism would appear correct if the analyzed muscovite and biotite are coeval with M2 metamorphism and related deformation. Field studies indicate that the synmetamorphic fabric associated with M2 is regionally extensive and traceable into the rocks immediately overlying the Middle Proterozoic Lincoln Massif (figure 2.1). This fabric is the only identifiable dominant foliation within this part of the pre-Silurian sequence. M2 related deformation, therefore, must be responsible for shortening of the massif during the Taconian collisional event. Other dates obtained through radiometric analysis in central Vermont were predominantly done on the medium-high pressure M1 cores of barroisitic and hornblende-rich amphiboles in various lithologies across the pre-Silurian belt. These dates range from 388 Ma, to the south along the SE spur of the Lincoln Massif, to 471 Ma, within the Pinney Hollow Formation at Granville Gulf (figure 4.10). These dates represent Taconian cooling ages and close approximations of metamorphic age for amphiboles formed near the argon closure temperature of 480°C.

The closest "true" Acadian date within central Vermont comes from the Umbrella Hill conglomerate of the Missisquoi group near Northfield, Vermont (figure 4.10). The calculated dates range from 349 Ma to 355 Ma, representing the time of closure following Acadian biotite grade metamorphic growth of the analyzed muscovite (Harper, 1968).

It should be mentioned that this "cooling age" hypothesis is indeed just the interpretation of Sutter and co-workers, and is actually contested to some extent by Laird (pers. comm., 1989). Laird contends that muscovite and, in particular, hornblende ages from Mt. Grant are indicative of Acadian <u>metamorphism</u> as a result of reequilibration of inherent argon, possibly upon cooling from a Taconian thermal maximum. Therefore, in Laird's interpretation, Taconian heating may have led to cooling, but equilibration occurred during Acadian time.

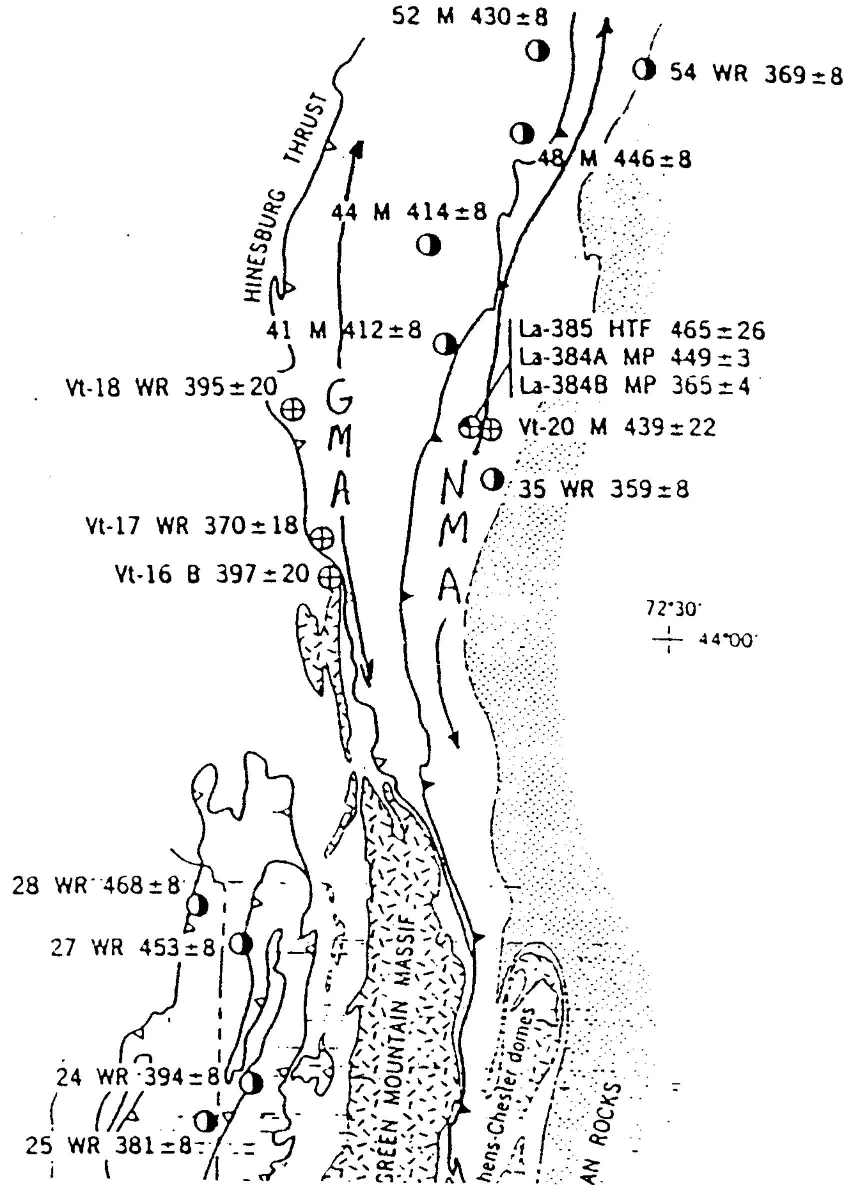


Figure 4.10 Compilation of $^{40}Ar/^{39}Ar$ and K-Ar ages for the Vermont pre-Silurian sequences. GMA and NMA refer to the Green Mountain and Northfield Mountain anticlinoria, respectively (from Sutter et al., 1985). WR- whole rock K-Ar ages; MP- $^{40}Ar/^{39}Ar$ muscovite plateau ages; M- K-Ar muscovite ages; B- K-Ar biotite ages; HTF- $^{40}Ar/^{39}Ar$ hornblende total fusion ages.

CHAPTER 5
STRUCTURAL AND TECTONIC SYNTHESIS

Introduction

The previous descriptions of the various deformations pose several important questions:

1. Is the synmetamorphic deformation primarily the result of simple shear or pure shear ?
2. How do the various deformations within the three domains correlate ?
3. What is the transport direction of pre-Dn-1 faults ?
4. Which deformations are Taconian and which are Acadian ?

This section will be followed by a tectonic synthesis which integrates the structural analysis with the metamorphic, lithologic, and stratigraphic information presented in the previous chapters.

1. Pure shear vs simple shear:

Petrographic analysis of the synmetamorphic deformation within all three domains shows that most of the grain shapes and porphyroblast geometries are symmetrical; this is especially true of rocks outside of the mylonitic fault zones (figures 3.3 & 3.5). Asymmetric quartz tails on garnet and plagioclase porphyroclasts do occur in these rocks (figure 3.4), but they are certainly few in number compared to those with symmetrical tails. Mylonitic rocks within fault zones display some tail asymmetry, but an equal number may still be relatively symmetrical (figure 3.13). Mylonitic fault zone asymmetry is occasionally displayed by the shear band foliation which bends the dominant foliation, creating a C-S relationship (Ramsay, 1967; Ramsay and Huber, 1983; Lister and Snoke, 1984). This shear band foliation is generally oriented at a low angle to the mylonitic schistosity, and is sometimes hard to distinguish. Fn (and Fn+1) folds within these fault zones commonly show higher amplitudes, tighter wavelengths, and steeper plunging hinges than those outside of these zones. In addition, many of the fold limbs within the fault zones are either highly attenuated or completely sheared and transposed by the fault zone fabric. The fact that attenuation and shearing occurs along the axial planar mylonitic Sn (or Sn+1) fabric, rather than along the later shear bands, suggest that Sn (Sn+1) age deformation continued to occur even after Fn (Fn+1) folding had ceased.
These relationships suggest that flattening (pure shear) and rotational strain (simple shear) may played major roles, at different times, in the development of the various synmetamorphic deformations. The fact that asymmetrical fabric, grain shape, and porphyroclasts do exist, both in and away from the mylonitic fault zones, does indicate that

a rotational strain component did accompany flattening; because it is also possible that the mylonitic (s-surface) schistosity was rotated into parallelism with the shear band cleavage during high intensity rotational strain, few unequivocal conclusions can be reached regarding the role of pure shear vs simple shear.

Biot (1961, 1965) presented a structural model which showed that both flattening (pure shear) and rotational strain (simple shear) can occur simultaneously within the same deformation regime. Cosgrove (1976) refined this model, incorporating better constrained constitutive equations for flattening and rotational strain, derived from deformation experiments using interlayered plasticene and vaseline (figure 5.1). This results of the deformation experiments led to the mathematical derivation of a unified constitutive equation for fold and cleavage development. The results of this derivation demonstrated that the relative magnitudes of flattening and rotational strain within an overall compressive environment are a function of:

1. The initial orientation of the stress field relative to the surface to be folded (anisotropy of figure 5.1).
2. The relative resistance of the deformed material (ie: the different layers of the anisotropy) to shearing and compression (L and M of figure 5.1).

These functions predict that preexisting foliation or bedding with a high degree of anisotropy (high M/L), oriented parallel to sigma one, will develop folds and deform by pure shear only. Conversely, a fabric with a low degree of anisotropy, oriented normal to sigma one, will develop a conjugate set of cleavages dominated by simple shear deformation.

In this study area, the predominance of symmetrical fabric within rocks outside of the fault zones, over asymmetrical fabric, the development of one axial planar cleavage and associated folds, suggests that the rocks deformed under a pure shear dominated strain state with sigma one oriented nearly parallel to the preexisting foliation and slightly diverging from the normal to the developed axial plane. The preexisting foliation within most of the rocks where it is still preserved shows a high degree of compositional layering and would be considered anisotropic on a mesoscopic scale ($M/L > 0.5$). Deformation would therefore approximate a geometry between cases a and d of figure 5.1 (closer to a than d).

The importance of this model and the interpretation of the synmetamorphic deformation is three-fold:

1. Sn deformation within the central, eastern, and eastern part of the western domain have fold hinges that plunge steeply down the foliation surface. This type of geometry could have developed either because Fn folded a steeply east dipping Sn-1 surface or Fn fold hinges progressively rotated into their present position. If Cosgrove's model can be

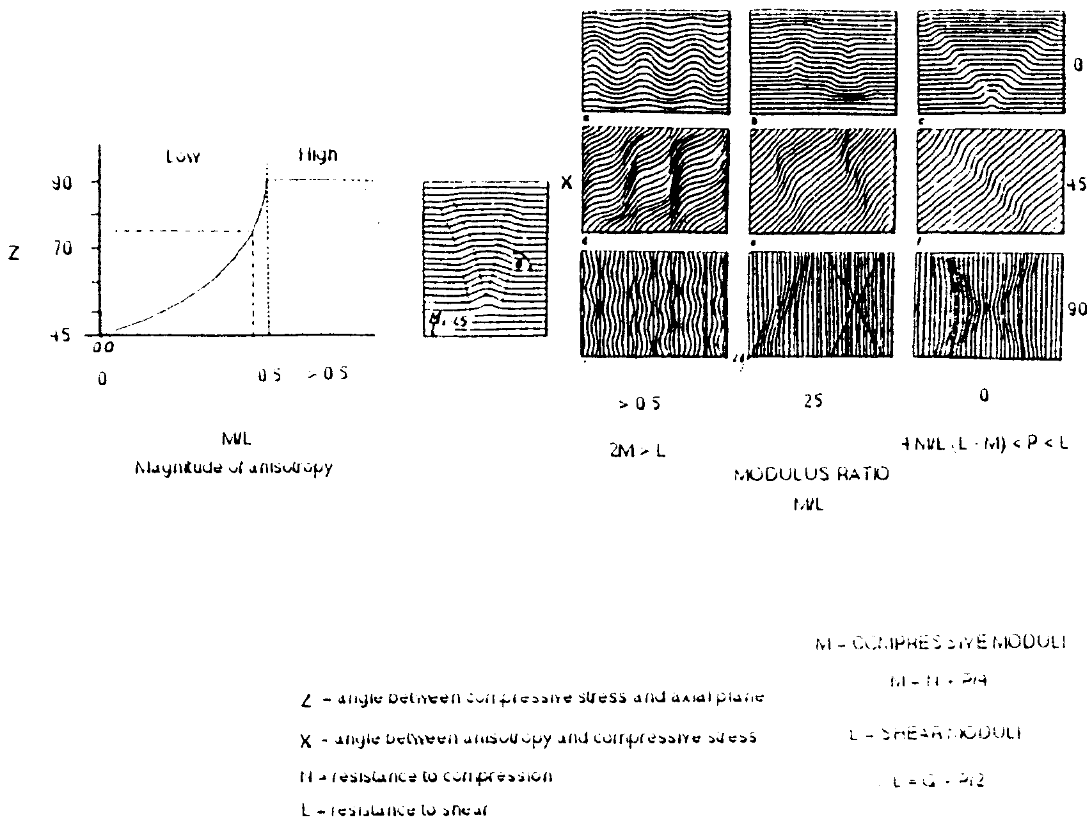


Figure 5.1 Graphical portrayal of Cosgrove's (1976) model for the development of crenulation cleavage and foliation under different strain regimes. **A)** Plot of Z (angle between compressive stress and axial plane of developing foliation) and M/L modulus ratio, an indicator of rock anisotropy. For low, but increasing M/L values, Z increases from 45 (simple shear) in an exponential manner. For M/L = .5, Z = 90 and resulting deformation is entirely simple shear. This assumes that anisotropy (bedding or preexisting foliation) contains the maximum compressive stress axis. **B)** Diagram showing how the initial orientation of preexisting anisotropy affects the development of subsequent foliation. For case where anisotropy is parallel to compressive stress (X angle) and M/L > 0.5, pure shear only occurs and axial plane develops normal to compressive stress. For same M/L, but X = 45, both pure and simple shear act upon folded surface. The other endmember case (M/L = 0, X = 90) shows the development of a conjugate cleavage set. This model shows that the magnitudes flattening and rotational strain components are controlled by the orientation of preexisting anisotropy (the surface to be folded) relative to the stress field, and the relative anisotropy of the material (its resistance to both shearing and compression; see text for discussion).

applied to this region, the type of fold geometry that characterizes Fn/Sn deformation indicates that Sn-1 was shallowly oriented prior to folding (assuming sigma one was not near vertical). The variation in Fn hinge orientations, with the more competent greenstones being inclined (relative to the Ln mineral lineation) and the less competent hinges within the pelites being reclined (parallel to the down dip mineral lineation) strongly suggest that Fn hinges may have originated in a shallow orientation and were subsequently rotated either through passive recrystallization (for pure shear) or physical rotation (simple shear) into the elongation direction (plate 6).

Fn sheath fold development would, following the above model, be a product of rotational strain and a subsequent flattening strain (Ramsay and Huber, 1983). Sn mylonitic fault zones, with a greater component of simple shear, may have developed following strain hardening of the rocks which would directly relate to a lowering of the M/L modulus ratio. This would initially cause the development of more asymmetric fabric within the same Sn foliation, but would eventually lead to **either** the development of a new foliation (the shear bands) at a low angle to Sn (case b, figure 5.1), or a recrystallization of the shear fabric during subsequent flattening .

Sn+1 and related Fn+1 folds have shallow hinges that plunge to the northeast (plate 6; figure 3.16). The asymmetry of these folds (the largest order is consistently east-over-west, in a down-plunge view) indicates that $M/L < 0.5$ and that sigma one was not normal to the anisotropy, Sn (the folded surface). These folds may have developed in response to two major factors:

1. The Sn foliation surfaces were rotated by Sne-related mylonitic fault zones such that a greater resolved component of shear was produced relative to flattening.
2. Work hardening of the mylonitic fault zones, probably as a result of significant grain size reduction without an equal rate of recovery, led to cessation of these mylonitic zones. A need for continued shortening was then accommodated by the initiation of a new set of folds, Fn+1 (and related axial planar schistosity, Sn+1) at a shallow angle to the Sn schistosity. Pi intersections from Sn+1 and undeformed Sn axial surfaces are colinear with Fn+1 hinge lines (plate 6). Because of the coplanarity of Sn and Sn+1 in the central and eastern domains, and their relative age relationships with metamorphism (chapter 4), these two generations are considered to parts of a strain continuum which developed with respect to a consistently oriented stress regime.

2. Correlation of structural fabrics:

Identification of deformation fabric by temporal relationships in each domain does not necessarily mean that the dominant fabric of one domain is the time equivalent of the dominant fabric within another. Since the dominant fabric nomenclature system is based upon the timing of fabric relative to the dominant one it is crucial to analyze the relative age relationships between fabrics and metamorphic assemblages within each domain before any attempt is made at a region-wide fabric correlation (Williams, 1984). Correlation of structural fabric must be based not only on local fabric relative age relations but on the spatial comparison of fabric age relative to regional mineral growth, comparison of local fold geometries, orientation of hinges, axial surfaces, and lineations. In addition, the necessity to observe and analyze the domain boundaries to check for any possible changes in fabric intensity across the study domain.

Previous studies of the central Vermont pre-Silurian have considered the dominant deformation across the entire belt to be either coeval or diachronous (younging to the west), and continuous (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Haydock, 1988; Prewitt,1989). Lapp and Stanley (1987) did make note of variation in Fn fold style and hinge orientation across the western and eastern parts of their study area. O'Loughlin and Stanley (1986) noted the decrease in metamorphic grade of mineral assemblages related to Sn, from west to east, characterized by an increase in chlorite overgrowths on garnet in the Mt. Abraham Schist. Compilation of the work previously completed in central Vermont shows that the eastern part of the pre-Silurian section (Ottauquechee, Pinney Hollow, Stowe, and Granville Formations) includes a chlorite grade dominant schistosity (Sn) axial planar to large amplitude isoclinal, reclined (sheath) folds and subsequent pervasive mylonitic fault zones. The western sequence, with a broadly expansive lithologic distribution (due to the relatively subhorizontal orientation of Snw-1 and pre-Snw-1 contacts) includes a garnet to kyanite-chloritoid grade dominant schistosity, Snw. This is locally axial planar to inclined overturned or isoclinal reclined folds with smaller half-wavelengths than those found farther east. In addition, the western domain Sn is itself folded by a chlorite grade secondary foliation (Snw+1) associated with upright open to slightly westward overturned chevron folds. Lapp and Stanley (1987) also noted that these folds have consistently shallow north or south plunging hinges in the western part of his area. Fn+1 in the eastern part became isoclinal, with hinges showing significant fanning with a cluster of reclined hinges parallel to a chlorite lineation, defined by overgrowth tails on garnet.

These relationships are similar to those found within the western domain of this study. Within the Granville - Hancock area, the east-west trending Hancock Branch and White River both provide almost continuous exposure through an otherwise glacial till covered area. Structural data on Snw, Snw+1, Fnw+1, and associated N+1 mineral lineation and quartz rods, comparison of fabric relationships with respect to mineral growth phases M1 and M2, and comparison of fabric geometry across the domain, provided some valuable insight into determining the relative age relationship of western Sn to that of the similar eastern and central domains. The critical relationships include;

1. Western domain Sn is associated with garnet grade M1 metamorphism. This is similar to that reported by all previous workers in along strike correlative rocks to the north. Sn within the other domains is consistently sub garnet grade. Chlorite and sericite aligned parallel to Snc and Sne and can be seen as overgrowths on earlier formed garnet porphyroblasts.

2. Western Sn is relatively shallow in orientation as evidenced by the broad distribution of major lithotectonic units and the shallow plunge of Fnw+1 fold axes which fold Snw surfaces (plate 6). This is in contrast to the steep dipping structures within the eastern and central domains (plates 1 and 6).

3. Sn+1 of the western domain, although only locally developed to the west, becomes a penetrative foliation within the eastern part of the western domain. This foliation is defined by chlorite, sericite, and quartz which overgrow M1 garnets (figures 3.3, 3.4, 3.5). These garnets show progressive resorption from west to east across the western domain. The transition between the secondary Snw+1 fabric and the Snw+1 schistosity that actually becomes the dominant fabric within the eastern part of this domain, occurs along a 1km distance, approximately west to east, centered upon the confluence of Texas Brook with the Hancock Branch (plate 1, V13). A similar transition can be seen along the east shoulder of Battell Mountain between the Green Mountain crest and Texas Gap (plate 1); because this traverse does not include continuous exposure, the Hancock Branch is considered a better traverse in order to observe this transition. Fn+1 folds within the western part of the domain are upright chevron folds, oblique to the N+1 chlorite/sericite elongation mineral lineation.

4. Fn+1 folds in the central and eastern domains are not reclined, do not have any associated mineral lineation or reclined quartz rods, or any pervasive axial planar Sn+1 schistosity. Fn+1 folds within the central domain occur midway through the domain and

progressively increase in frequency to the east. There are no Fnc+1 folds within the western part of the central domain.

The data indicate that Fn+1/Sn+1 of the western domain are not demonstrably coeval with Fn+1/Sn+1 of the other two domains. The fact that the central domain N+1 deformation is not present within the western part of the domain and increases in frequency to the east, argues against a fabric correlation with western domain N+1. Sn fabric of the western domain is demonstrably garnet grade (M1 metamorphism) and does not show pseudomorph texture or evidence of an earlier higher grade assemblage (chapter 4). Sn of the central domain and eastern domain is chlorite grade and in every sample containing relict higher grade minerals, shows clear evidence of polymetamorphic texture and reequilibration; chlorite overgrowths are abundant on progressively resorbed blue-green hornblende, and recrystallized quartz, plagioclase, epidote, rutile, and ilmenite occur within microlithons of aligned chlorite, sericite, and a second generation of quartz (figures 3.5 & 3.18). The earlier assemblage formed at garnet grade, and is associated with an earlier foliation (Sn-1). Well preserved greenstones contain medium-high to high pressure facies series barroisitic hornblende, with overgrowths of chlorite (chapter 4; see Laird and Albee, 1981). If the metamorphism is regionally expansive (as portrayed by all previous workers, including Bothner and Laird, 1986; Laird et al., 1981; Stanley and Ratcliffe, 1985; Sutter et al., 1985; Stanley et al., 1987a, 1988, 1989), then garnet grade Sn of the western domain should correlate with garnet grade Sn-1 of the central and eastern domains.

The change in style of upright Fnw+1 to reclined, isoclinal Fnw+1 (which appears to become the dominant fabric in the eastern part of the western domain) may record the transition of western Snw+1 to central Snc-age dominant deformation. The progressive increase in fold hinge plunge would thus be a result of rotation into steeper orientations through a combination of pure and simple shear. This strain increase would explain the increase in pervasiveness of Snw+1 from west to east as well as the increased development of quartz rods and associated elongation mineral lineation. This type of progressive strain phenomenon has been described for other orogenic belts (eg; Bell, 1981; Burg et al., 1981). The lack of regionally continuous Snw+1 reclined folds, quartz rods and mineral lineation, as well as the lack of pervasive Snc+1 within the central domain, strongly argues against correlation of central and eastern domain Sn+1 with western Snw+1.

5. Central and eastern domain Sn are believed to be relatively coeval and probably continuous since they both occur at chlorite grade, pseudomorph a higher grade fabric (Sn-1), have similar orientations, and magnitudes of rotation, have similar superposed deformations, and seem to be traceable across domain boundaries.

An alternative hypothesis for this deformation has been presented by O'Loughlin and Stanley (1986), Lapp and Stanley (1987), Haydock (1988), and Prewitt (1989). These investigators interpreted Snw+1 as a discreet and isolated deformation event possibly related to the development of the Green Mountain anticlinorium. This model was attractive to these workers because they had generally interpreted the Snw dominant fabric to be the western continuation of Sn within rocks farther east. In addition, this model was a simple and logical way to explain the lack of an apparent transition from Sn+1 of the west into eastern Sn+1. This interpretation disregards fabric-mineral chronologic relationships which clearly indicate that there was a single garnet-kyanite grade metamorphic event (M1) that was followed by a later retrograde (M2), chlorite-biotite metamorphism (Laird et al., 1984; Bothner and Laird, 1986). Even though it is not explicitly stated, all of the previous studies noted that garnet to kyanite grade metamorphism was either older than or synchronous with Sn deformation in the west and that this M1 metamorphism showed strong evidence of retrogression as evidenced by the development of chlorite overgrowths on garnet and hornblende (the chlorite "comet tails" on garnet, of Lapp and Stanley, 1987). These studies also observed that Sn deformation within the areas farther east (central and eastern domains of this study) was concomitant with chlorite to biotite grade minerals, that overgrew hornblende and in some cases was observed overgrowing preexisting garnet (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1987; Haydock, 1988). These relationships indicate that the Snw fabric developed in high grade rocks prior to retrogression, and that Sn fabric farther east developed in rocks actively undergoing retrogression. Even if one invokes the model that Snw+1 retrograde metamorphic fabric is distinct from Snc and Sne retrograde fabric, M1 garnet grade assemblages still predate and occurred during Snw development; Snc and Sne clearly post date it. If Snw+1 related metamorphism is younger than Snc/Sne related retrogression, **three** phases of metamorphism should be present in the western rocks; M1 garnet, prograde, the M2 retrogression seen in the central and eastern domains, and a subsequent retrograde event associated with the development of Snw+1. There is no evidence that two retrograde metamorphisms occurred within the western domain following M1 prograde metamorphism. For these reasons, the interpretation that Snw+1 is a western continuation of Snc and Sne deformation, with Snw probably being the western equivalent of Snc-1 and Sne-1, is favored.

3. Transport direction of pre-Dn-1 fault slices:

Direct kinematic evidence for transport directions of premetamorphic, pre-Dn-1 fault zones is absent because M1 prograde metamorphism obliterated the preexisting fabric by either dynamic or static recovery and annealing processes (Hobbs et al, 1976). Transport direction is inferred through sedimentologic and stratigraphic arguments, presented at the end of Chapter 2. The thrust slice sequence, from structurally lowest to highest, includes;

1. Monastery and Battell Formations, comprising the "autochthonous" sequence.
2. Pinney Hollow Formation and Mt. Abraham Schist structurally overlie (1) along the Child's Mountain Thrust (plate 5).
3. Granville and Ottauquechee Formations structurally overlie (2) along the Ottauquechee Thrust (plate 5).

The sedimentologic and stratigraphic arguments put forth in chapter 2 indicate that Pinney Hollow and Mt Abraham rocks were originally deposited as a rift clastic sequence outboard (east of) the rift-clastic Monastery and post-rift Battell sequence (plate 3). The Stowe Formation represents the eastern equivalent of the Pinney Hollow Formation. Granville and eastern equivalent Ottauquechee Formations represent carbonaceous sediments deposited within an anoxic, post rift environment (plate 3). The presence of ultramafics within the Ottauquechee suggests that it may have been deposited on or within close proximity of oceanic crust, within the distal-most parts of the Iaptean passive margin. Granville, which appears to be the proximal equivalent of the Ottauquechee, and the distal equivalent of the Battell, may have been deposited on Pinney Hollow, Mt. Abraham, and Stowe sediments (plates 3 & 5). These passive margin, pre-orogenic stratigraphic relationships dictate a westward transport of the evolving thrust stack, either as a diverticulated sequence (Stanley and Ratcliffe, 1985) or as a westward imbricating thrust package (Armstrong, 1989).

4. Taconian vs Acadian structures and metamorphism:

Radiometric age constraints presented in chapter 4 indicate that Snw, Snc-1, and Sne-1 related deformation, and all preexisting deformation within the pre-Silurian cover rocks, was Taconian (Laird et al., 1984; Sutter et al., 1985). The similarity in style of Sn and Sn-1 deformation suggests that the Snw+1, Snc, Snc+1, Sne, and Sne+1 deformation, and associated M2 retrograde metamorphism, was also Taconian. Since the first metamorphic event of any particular orogeny is usually (if not always) a prograde event, it is probable that the late stage garnet present at Allbee Brook and farther south within the Rochester area, represent the onset of Acadian metamorphism. Northeast trending Sne+3 deformation post dates garnet growth within the Rochester area (J. Laird, pers. comm., 1989, 1990). This deformation is the oldest Acadian fabric within Silurian-Devonian rocks east of the study area (Hatch, 1987; Westerman, 1987).

Structural evolution

The correlation of the moderately developed Snw+1 with the very high strain Sn fabric of the central and eastern domains agrees with the well accepted concepts of orogenesis. These concepts include time transgressive deformation and the progressive recycling of fabric associated with rotational strain in a constant strain state.

Based upon regional models of orogenic evolution and the facts posed in the previous section, an internally consistent model for the structural evolution of the studied rocks can be posed (figures 5.2 & 5.3);

1. Following the development of the Late Proterozoic/Lower Ordovician Iaptean passive margin (chapter 2), and initiating sometime during the Lower Ordovician, was the development of premetamorphic folds within the distal rift-clastic, post rift section of Pinney Hollow, Stowe, and Ottauquechee Formations (Fn-2, herein called F1; Appendix III). F2 folds are described by the gross distribution of greenstone within the eastern domain Pinney Hollow/Stowe belt. These folds were cut by;

2. A set of thrust faults (pre Dn-1; herein called T1 and/or S1) that probably developed during the transmission of shortening from an accretionary wedge into the Late Proterozoic/Lower Cambrian passive margin sequence (plate 5). These thrusts predated the M1 peak of metamorphism and placed post rift, oceanic, and accretionary rocks of the Ottauquechee Formation on top of the westerly situated rift clastic sequence of Stowe and Pinney Hollow Formations (B-type subduction of Hodges, et al., 1982). This could have also occurred within deep crustal levels of an evolving subduction zone/accretionary complex. This geometry suggests that any oceanic fragments may have been part of a preexisting oceanic lithosphere formed within a forearc environment or ocean crust of the passive margin (Hamilton, 1988). Such ultramafics may even have had an origin as a rift-related cumulate phase within distended continental crust (Tracy et al., 1984).

3. T1 Thrusting and expansion of an accretionary wedge (Armstrong et al., 1988a), continued with subduction of the continental margin (A-type). This continued thrusting resulted in development of the Child's Mountain thrust which placed Pinney Hollow and Mt. Abraham rocks over the top of more proximal lithified sediments (Battell Fm.) and underlying rift clastics (Monastery Fm.) whose geometry indicates that they were

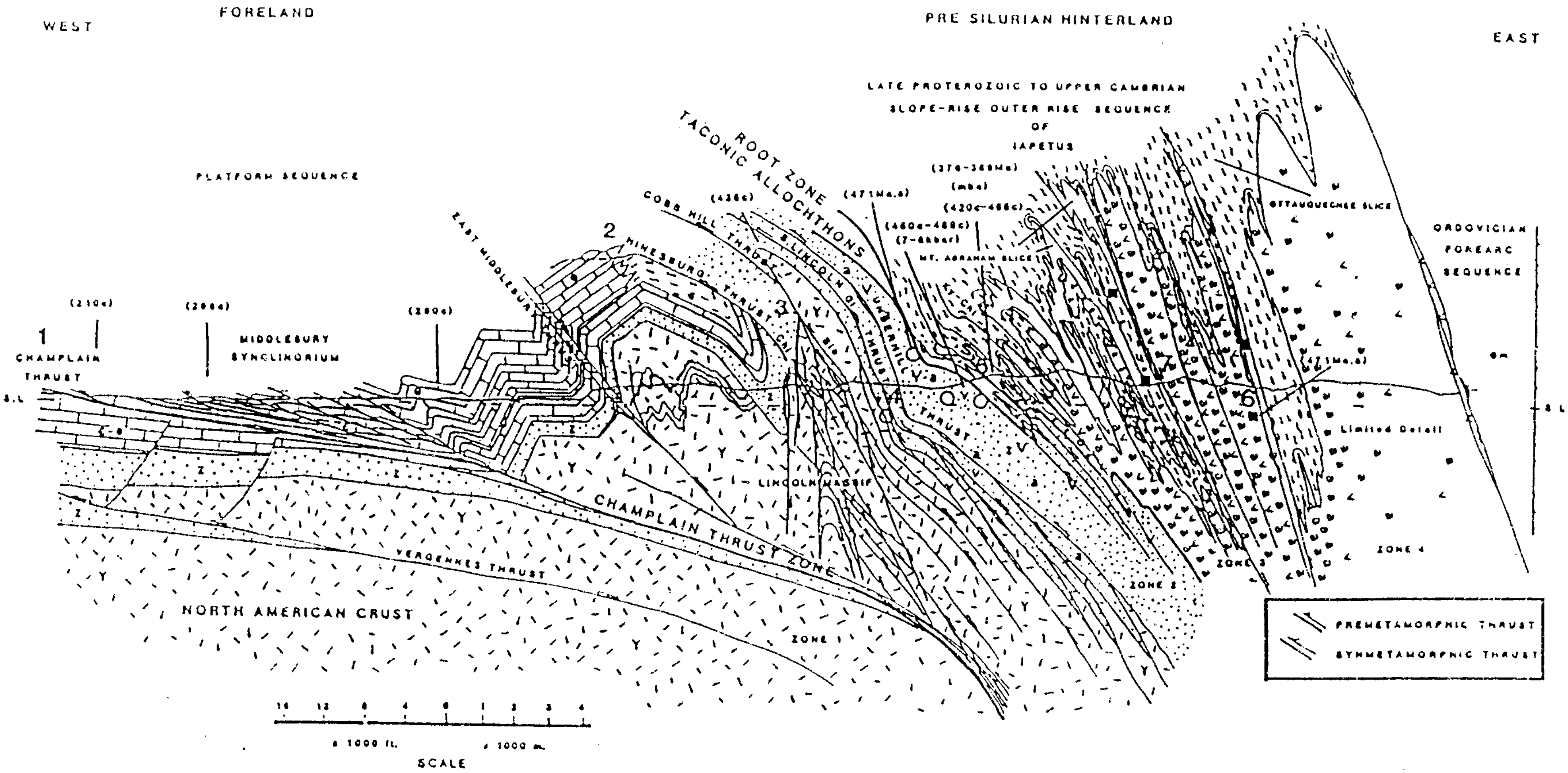


Figure 5.2 Regional cross section across the central Vermont foreland - hinterland sequence, displaying the major fault zones, the proposed Taconic Root Zone (of Stanley et al, 1989), and various metamorphic thermobarometric data and radiogenic ages.

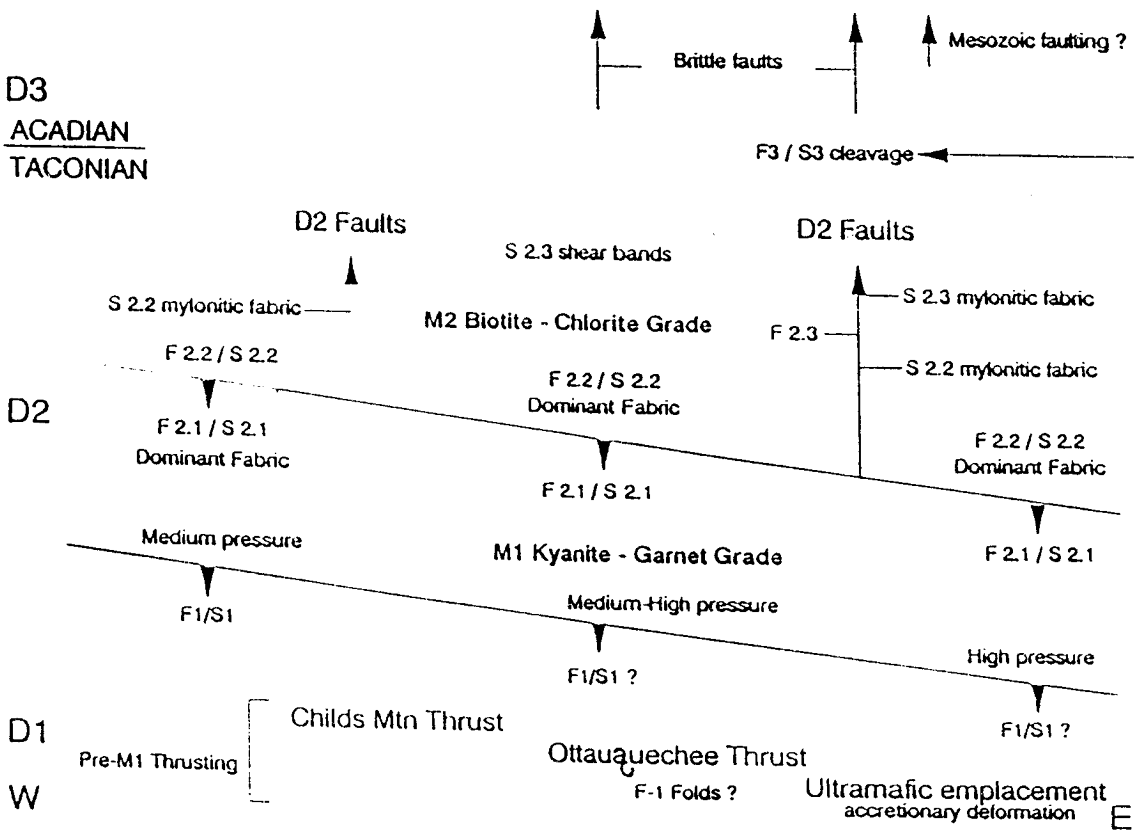


Figure 5.3 Time-deformation-metamorphism diagram showing the east to west temporal relationships between the various phases of deformation and metamorphism recognized in this region during this study. See chapter 5 for complete discussion and Appendix III for a description of the fabric nomenclature used in this figure and the text.

previously warped or folded; the present distribution of Pinney Hollow and Mt. Abraham rocks directly which locally lie directly on top of either Battell or Monastery units, necessitates that these underlying rocks previously underwent deformation such that subsequent pre-Dn (T1 and/or S1) faulting cut through anticlines and synclines cored by the Monastery Formation (Boyer and Elliot, 1982).

4. Following T1 faulting was the development of the first closely spaced schistosity, S2.1 (Snw of the western domain; Snc-1 and Sne-1 of the central and eastern domains), synchronous with at least garnet grade metamorphism (M1). M1 garnet grade metamorphic assemblages are present within both the upper and lower plates of the Child's Mtn. Thrust, further suggesting S2.1 foliation development following T1 thrusting. High grade metamorphism is most likely related to thermal equilibration of a perturbated geotherm following thrust-related tectonic loading (Thompson and England, 1984).

5. Superposed over or transposing the garnet grade S2.1 foliation was a second synmetamorphic foliation, S2.2, (Snw+1; Snc; Sne) which developed throughout the area at sub-garnet grade. This foliation was associated with a regionally pervasive strain event, whose shallow plunging hinge lines in the western domain, inclined to steep hinges in the central and eastern domains, and consistently oriented, steep east dipping axial planar schistosity within all three domains, are indicative of sub horizontal, layer parallel shortening. The orientation of F2.2 hinges (plate 6) suggests that S2.1 foliation was sub-horizontal prior to folding. The progression from shallow plunging F2.2 hinges in the west to steep F2.2 hinges to the east is probably the result of physical and/or passive rotation within and along the S2.2 foliation surface during continued flattening and rotational strain. This progression coincides with the idea that collision, and related deformation were time transgressive from east to west.

S2.2 fabric development evolved from as a weakly developed secondary spaced schistosity associated with upright folds with subhorizontal hinge lines (Sn+1, western domain). Folds may have developed either during early flattening dominated strain. Parasitic fold geometry associated with a predominance of symmetrical grain shapes, and recrystallized porphyroclast tails, indicating that fold development is a consequence of a flattening dominated strain. This style of deformation is presently observable within the western and central parts of the western domain. With continued time and strain, the fabric evolved into a pervasive schistosity, overprinting a higher grade S1 foliation to the point where both were of equal development (Texas Brook area, western domain; plates 1 & 2). F2.2 hinges, initially upright, were subsequently rotated with increasing strain toward the elongation direction, defined by a developing mineral lineation, L2.2. Preexisting quartz veins were folded by F2.2 during this place, with quartz rod hinges also being actively and/or passively rotated toward the elongation direction. The early axial plane foliation began to develop east-over-west slip along S2.2 schistosity surfaces. This sense of slip is present of both upright and overturned limbs on the F2.2 west verging folds. This asymmetry reflects a late D2 stage of rotational strain, resulting in the tightening of the previously developed F2.2 folds (initiated by flattening). Continued strain led to the shearing out of fold limbs and the development of discreet S2.2 (Snw+1, Snc, Sne) shear zones. This type of geometry is presently found within the eastern part of the western domain.

5. Continued strain within a constant strain state, led to further development of the (S2.2 or T2.2) shear zones as a result of strain softening. This softening phase was related to retrograde metamorphism, requiring the influx of substantial volumes of fluid in order to retrogress the garnet grade assemblage (chapter 4). Advection of substantial fluids may have been accommodated by the evolving Snw+1, Snc, and Sne shear zones, many of which show sulfide mineralization (pyrite, chalcopyrite). Regional retrograde metamorphism may have initially been controlled by fault zone distribution; later strain softening within the retrograde belt, and high magnitudes of strain were actually controlled by reaction enhance softening (Knipe and Wintsch, 1985). Strain softening-related shear zone deformation (T2.2) caused the shearing out of F2.2 reclined fold limbs, parallel to the plane of earlier S2.2 foliation. These T2.2 mylonitic fault zones contain either asymmetric fabric and/or a secondary shear band foliation, which along with the mylonitic S2.2 consistently show east-over-west asymmetry. Truncations of stratigraphy occurred in zones where S2.2 fault zone deformation was intense enough to create significant displacements (up to 7km; plate 1, V38, locality 331). Late D2.3 conjugate shear zones are oriented with the obtuse angle (interpreted sigma one bisector) normal to the regional compressive stress direction. This may have been a late stage of D2 deformation that followed a period of increased metamorphic differentiation and related accentuation of rock anisotropy (Cosgrove, 1976). This type of deformation geometry is best preserved and developed within the central and eastern domains. Deformation of this magnitude apparently did not occur within the western domain. This also suggests that D2 deformation was time transgressive from east to west.

6. Following D2 strain softening was a period of strain hardening. This was associated with the development of a third generation of synmetamorphic upright, open to slightly westward overturned Snc+1, Sne+1 folds (S2.3) which fold the dominant Sn foliation of the eastern and central belts (S2.2). The early development of these folds did not have an associated axial planar schistosity, even though axial planes developed coplanar to earlier S2.2. This suggests that S2.3 was part of a strain continuum, in a consistently

oriented stress state. These type of folds are most common in the central part of the central domain. With continued time and strain, S2.3 developed as a weak foliation, axial planar to progressively evolving F2.3. These folds have hinges which progressively migrated toward the elongation direction with increasing strain. The fact that the majority of the F2.3 folds did not fully rotate into the elongation direction suggests that the rotational strain began to locally subside during the development of this phase. The subsequent development of locally present shear zones (T 2.3) is the consequence of strain softening (=higher strain rate) rather than increased rotational strain. The development of S2.3, F2.3, and T2.3 deformation is best displayed in Thatcher Brook, within the eastern domain (plate 1, T35).

7. During a later, obliquely oriented deformation event, another set of folds (F3; Snc+2), with northeast plunging axes and associated axial planar crenulation or pressure solution cleavage developed within the eastern part of the area. The similarity of orientation with the first cleavage found within Silurian-Devonian rocks to the east, suggests that this deformation occurred during the Acadian orogeny (Hatch, 1987; Hatch, pers. comm., 1989).

8. Post metamorphic fault zones (T3) reactivated earlier high strain S2.2 and S2.3 shear zones. These cataclastic, sometimes slickensided zones, appear to post date F3 folds and may either be Acadian or Mesozoic in age.

CHAPTER 6
CONCLUSIONS

Based upon the stratigraphic, metamorphic, and structural information presented in the previous chapters, a general model can be developed for the tectonic evolution of the pre-Silurian units of central Vermont. In order to constrain the model, however, several critical relationships regarding the relative timing of metamorphism and deformation must be addressed.

1. Progression of metamorphism and deformation

The peak of Taconian metamorphism within this belt reached at least garnet grade conditions regionally, and was locally as high as kyanite-chloritoid grade. The baric type facies (ie: medium and medium-high pressure facies series) are not defined by the early thrust sheets discussed in chapter 4. Pressure facies increase from west to east and suggest that the easternmost rocks were progressively tectonically loaded by the preceding pre-peak metamorphic (pre-Dn-1; T1) thrust slices (figure 5.1). This progression in baric type indicates that thrusting led to the progressive east to west tectonic shortening (and loading) of rock units. Rocks to the east were probably extensively loaded by rocks previously incorporated into a westward advancing accretionary wedge that was subsequently eroded during continued orogenesis. Metamorphism postdates thrusting due to the lag time between strain and thermally related processes (England and Richardson, 1977; Thompson and England, 1984). Peak temperatures and associated pressures for M1 metamorphism in the Granville - Hancock area are approximately 530°C (+/- 35) and pressures of 6.1-7.2 Kb (figure 4.10). In general temperatures for M1, where available, also show an east to west decrease. Temperatures for M2 retrograde metamorphism within the eastern part of the belt are nonexistent due to the lack of proper lower greenschist facies geothermometers and barometers.

2. Evolution of the dominant schistosity (S2)

The dominant deformation within this belt, Snw+1, Snc, and Sne (comprising the regional S2) are all coeval with M2 retrograde metamorphism, which progressively decreases in intensity from east to west (figure 5.1). The upright attitude of this fabric, and coplanar later phases, is roughly consistent across the belt (where not deformed by later deformation phases S2.5 and S3). There is no evidence to indicate, or reason to believe, that this fabric was rotated into this position following development. It is possible that this deformation event occurred as a response to continued east-west directed compression following the development of the Lincoln massif into a large anticlinal structure (figure 5.1). The massif would have acted as a large buttress, forcing the eastern cover rocks to shorten in a subhorizontal attitude. Warren (1989) has portrayed development of the Lincoln massif through upright, internal deformation of basement and passive cover within a fault bend fold situated above the Champlain Thrust and possibly located along a significant Iaptean rift-related, steep east dipping normal fault. This interpretation implies that movement on the active decollment (Champlain Thrust) predated the D2 deformation within the eastern cover sequence. Presently there is no evidence to substantiate or refute this interpretation other than the need for a tectonic buttress in order to develop the upright D2 of the eastern cover.

According to DelloRusso (1986), Stanley and others (1989), and Warren (1989), the eastern Lincoln massif developed ductile shear zones and folds following initial development of the Lincoln anticline. The lack of a true basement-cover decollment on the eastern flank of the massif has been interpreted by these workers as evidence for deformation at relatively high temperatures under which strength contrasts between basement gneiss and cover units would be minimal (Byrne and Lucas 1988). This deformation may have occurred earlier when the decollment was initially subhorizontal and at high metamorphic grade during M1. The presence of garnet grade assemblages, albeit retrograded, within these shear zones does indicate that M1 (and thus tectonic loading) did progress sufficiently far west.

3. Western extent of pre-Silurian allochthons in central Vermont

Carbonate units within the Middlebury Synclinorium show mineral recrystallization following the development of significant thrust and fold related deformation (Washington, 1987; Sheppard and Schwarcz 1970; Garlick and Epstein, 1967). These marbles equilibrated at temperatures ranging from ~290°C in the east to ~210°C in the west (Stanley, 1989). Washington (1987) reported a steep metamorphic field gradient within the eastern part of the carbonate sequence, which he attributed to tectonic juxtaposition with greenschist facies rocks of the central Vermont hinterland. The rocks responsible for such thermal conditions could not be those presently adjacent to the carbonate platform, since they also show relatively low temperatures of metamorphism, and were supposedly never in a geometry which could have caused sufficient heating of the carbonate platform sequence (figure 5.1; rocks on the western limb of the Lincoln anticline). The metamorphism of the carbonate sequence was most likely related to the tectonic loading of the platform by allochthonous rocks originating from the east side of the

Lincoln massif. This loading and related metamorphism would represent the leading edge of allochthonous units that led to the M1 metamorphism within the eastern cover. The thrust and fold deformation within the carbonate valley could be related to allochthon emplacement as suggested by Washington (1987). This metamorphic evidence is crucial since it appears to be the only substantive evidence that indicates a northern extension of the Taconic allochthon at this latitude.

Tectonic model

Based upon the preceding information, a general model for the tectonic evolution of central Vermont can be proposed. This exercise is best accomplished progressing back in time from youngest to oldest deformation and metamorphism. The following discussion will emphasize relationships evident in the regional and study area cross sections (Stanley et al., 1989; figure 5.1 and plate 2, respectively).

1. (youngest) The youngest deformation in the belt appears to be the late stage folds (F3) and cleavage (S3) within the eastern part of the field area that deform the dominant and secondary schistosities (S2 and S2.5). This northeast trending fabric intensifies eastward and is similar in style and orientation to the earliest Acadian fabric within the Silurian-Devonian sequence (Hatch, 1987; pers. comm., 1989). Restoration to a time prior to D3 does not significantly alter the geometry of the belt.

2. Based upon structural analysis presented in chapter 5, the next youngest deformation would include some the few post- M2 faults with brittle fabrics, including that recognized along the Pinney Hollow - Ottauquechee contact. The magnitude of displacement along this and the other brittle faults is unclear but not significant enough to produce any new recognizable displacement geometries that would have to be explained in cross section. Some of these faults in the eastern part of the field area are folded by F3. Some of these faults, however, are not deformed and may be younger than F3, and possibly of Mesozoic age. Restoration of these faults does not significantly alter the geometry of either the regional or study area cross sections.

3. The next phase of restoration groups all D2 deformation and related M2 retrograde metamorphism. This includes the coplanar phases of schistosity (S2, S2.5), secondary folds (F2.5), and mylonitic shear zones (T2), and the earlier phases of F2 sheath and inclined folds. All of this deformation occurred in a relatively similar strain state during M2. The abundance of hydrous mineral phases, sulfides, and quartz veins indicates that D2 deformation involved large volumes of fluid, probably derived by dehydration reactions at greater depths (chapter 3). The upright attitude of D2 may be related to simultaneous (or previous) development of the Lincoln massif anticline and fault zones. This produced a western "backstop" upon which the eastern cover rocks were compressed during continued Taconian collision (figure 5.1). The relative decrease in M2 and D2 into the western part of the study area eastern cover rocks suggests that shortening within the belt was accommodated before strain was thoroughly transmitted from east to west across the entire cover. Development of such a steep fabric also suggests that a large amount of flattening may also have been accrued to these rocks, including significant volume loss. The lack of ideal strain markers precludes determination as to the role of flattening in this belt. It should be note that quartz vein development might also have been related to quartz pressure solution. This still requires the need for tremendous volumes of water-rich fluid from some other part of the belt (Thompson and England, 1984). Restoration of D2 and M2 has been attempted in numerous exercises during this study. Although difficult to quantify, palinspastic analysis of D2 produces a subhorizontal sequence of stacked lithotectonic units, from bottom to top, Monastery/Battell, overlain by Mt. Abraham/Pinney Hollow/Stowe (along the Child's Mtn. Thrust), Granville (either along another fault or in depositional contact), and structurally highest Ottauquechee (along the Ottauquechee Thrust; plate 5). These thrust are pre- to syn M1 in relative age.

4. Deformation within the Middle Proterozoic basement, related to the development of the Lincoln massif, would predate D2/M2 within the eastern cover rocks (figure 5.1). According to Warren (1989), evolution of the Lincoln massif began at deep crustal levels with development of numerous ductile shear zones as part of a decollment, which initially included the basement cover contact. Progressive metamorphism at depth may have led to shearing of basement below the cover contact and ultimate development of the Champlain Thrust, as portrayed by Stanley and others (1989; figure 5.1). Simultaneously, at higher levels, more brittle behavior may have led to development of the Lincoln anticline as a fault propagation structure (Suppe, 1983; Stanley et al., 1988). Subsequent breakthrough faults included the Hinesburg and Cobb Hill, and East Middlebury thrust zones. These thrust, which display ductile fabrics, may have originated at deeper crustal levels. This deformation, along with thrusting and folding within the carbonate foreland to the west may have been related to D1 and M1 deformation and prograde metamorphism in the east. This relationship would most likely have been as the last stage of eastern cover thrust emplacement. Basement related thrusting would represent migration of the pre M1 decollment to lower crustal levels following cessation of allochthon transport. This transition has also been recognized within other orogenic belts defining the change from subduction related (B-type) deformation to basement related (A-type) collision (Hodges et al., 1982).

M1 garnet grade assemblages within the eastern part of the massif attest to the tectonic loading of this area during the ductile deformation. The development of the new decollment and the Lincoln massif must have postdated allochthon emplacement in a period of time long enough to allow heating of the lower plate basement and autochthonous cover to peak conditions. Thermal effects within the lower plate, brought upon by the tectonic loading of hot thrust sheets, are not attained for substantial periods of time (England and Richardson, 1977). Thermal models show that the thermal lag time from tectonic loading to peak metamorphic temperatures may be on the order of several million years (Thompson and England, 1984). A substantial period of time may have punctuated allochthon emplacement and subsequent metamorphism of the lower plate rocks of the Lincoln massif. Restoration of the Lincoln massif basement rocks and associated cover creates a subhorizontal, possibly undeformed, geometry. Continuity with rocks of the eastern cover sequence (within this study) can be established by restoration along the Underhill Thrust Zone (plate 3). This thrust may have been one of the progressively developing decollments following allochthon emplacement over the "proto" Lincoln massif.

5. Following restoration of elements in step 4, the eastern cover is essentially composed horizontal pre M1 thrust slices. As discussed before, M1 metamorphism within the eastern cover may have been a direct result of thrust slice emplacement, including tectonic loading by accretionary wedge units. The Snw schistosity of the western belt appears to be both synchronous with, and older than the peak of M1 (chapters 3 and 4). Since the early thrust slice contacts appear to be folded by Fnw folds, Snw is probably younger than allochthon development. This is true of the early schistosities and related folding in the central and eastern domains as well. Because these early fabrics are not well preserved, either in outcrop or map geometry, their restoration has little effect on the regional cross sections.

6. Pre-M1 (T1) thrusting is believed to make up a significant part of overall shortening within the Taconian belt (Stanley et al. 1988, 1989). At least two major thrust surfaces were mapped during this study; the Child's Mtn. and Ottauquechee Thrusts (plate 3). Metamorphic evidence suggests that thrusting during this phase transported eastern cover units at least as far west as the eastern part of the carbonate platform (figure 5.1). Since the erosional level is below the allochthon thrust surface at this latitude the lithologic type(s) of units is unclear. The lithologic reconstruction of units does suggest, however, that aluminous schists and phyllite, and possibly more coarse grained types, similar to those found within the Taconic allochthon, existed in this thrust package. Regardless, this phase of thrusting is believed to be the coeval (and continuous) with Taconic allochthon emplacement further south. This phase of thrusting, leading to M1 metamorphism, has been interpreted as the result of Early to Medial Ordovician subduction of the eastern edge of the North American passive margin (Rowley and Kidd, 1982; Stanley and Ratcliffe, 1985).

7. Numerous lensoidal bodies of ultramafic rocks are regionally found within the Ottauquechee Formation. Within the study area, most of these ultramafics are found either internal to the black phyllite unit and along the T1 thrust contacts, overprinted by the D2 schistosities. Emplacement of these interpreted vestiges of oceanic crustal material may have occurred during accretionary wedge development and early stages of the T1 thrusting event. Specifically, this thrusting event may have had its beginnings within the oceanic subduction phase at the onset of the Taconian orogeny. The duration of thrusting, from subduction onset to allochthon emplacement is presently unclear but may have been of substantial length; Hames and others (1989) report the age of M1 within both Taconic allochthon and autochthon, as ~445 Ma. Rocklandian fossils (~457-452 Ma) occur within the autochthonous Taconian flysch sequence, and thereby bracket allochthon emplacement (in Massachusetts) at post 457 Ma and pre 445 Ma (Hames et al., 1989). $^{40}Ar/^{39}Ar$ and $^{40}K/^{39}Ar$ ages on M1 assemblages within central Vermont yield ages as old as 471 Ma (Laird et al., 1984). If isotopic systematics are correct for these samples then allochthon thrusting (pre-M1) must have occurred before 471 Ma in the eastern cover lithologies. If the additional component of south to north diachroneity is not taken into account T1 thrusting must have lasted at 20 million years. This amount of time seems unreasonable when compared to B-type episodes in modern orogenic belts, usually lasting no longer than 5 to 10 million years (Hamilton, 1988). More data for Ar/Ar isotope correlation analysis needs to be collected from the eastern cover before any conclusions can be drawn from this matter.

Restoration of the pre-D1 thrust slices is directly dependent upon the number of thrust slices one thinks are present. This study could conclusively, only recognize two, the Child's Mtn. and Ottauquechee thrusts. Other studies have interpreted others but appear to be based upon either inconclusive criteria or complex geometric reasoning that precludes a feasible restoration (see Prewitt, 1989; Haydock, 1988). Further evaluation awaits a regional compilation of the detailed central Vermont studies. Restoration of units within this study allows (not demonstrates) for the development of a reasonable pre-orogenic passive margin (plate 4).

8. Restoration to a coherent passive margin diagram necessitates the replacement of allochthonous sequences that are no longer exposed within the central Vermont transect (figure 5.1). The exact nature of the allochthonous rock-types is dependent on the nature of sedimentary transitions from south to north within the restored passive margin (Armstrong and Colpron, 1989b). The projection of rock-types identical to the Taconic sequence into the central Vermont assemblage assumes no drastic facies transitions. This is quite tenuous especially for the rift-clastic section which shows tremendous sedimentary variation in modern environments (Kinsman,1977). The interpreted passive margin diagram, although internally consistent with available data, should be viewed with these assumptions in mind. Although the Taconic and Oak Hill sections (south and north, respectively do show similar lithologic and stratigraphic relationships, these are mainly within what has been interpreted as post-rift phases which are known to be regionally consistent (Dewey, 1982; Kinsman, 1977; chapter 2).

Future work

Based upon the information made available during this, and previous studies in central Vermont, the following is a list of projects that may be useful in solving some of the present arguments, contradictions, and uncertainties regarding Taconian geology in New England:

1. The detailed structural analyses developed during many mapping projects within central and northern Vermont, must be supplemented by similarly well detailed metamorphic studies. The structural models that have been presented for this belt have yet to be tested by thermobarometric constraints, and are therefore inconclusive. Work must be done on separating different mineral assemblages from polymetamorphosed rocks, and relating them chronologically to each other and to the various deformational fabrics. Quantitative as well as qualitative estimates of evolving pressure and temperature conditions must be established.
2. Additional isotopic analyses of both M1 and M2 are needed to constrain estimates of absolute times of metamorphism and deformation. In addition, systematic studies regarding the use of Na-rich amphiboles and actinolite are needed to demonstrate their validity in use as geochronometers.
3. Emphasis should be placed on well constrained structural analyses of detailed problems, regarding the roles of rotational vs flattening strains in the development of the sheath fold structures and shear zones. Studies should be attempted on depicting the effects of metamorphism on deformational processes within shear zone evolution. An understanding of mechanical processes and the pressure-temperature and strain conditions under which they from would be of great use in development of structural evolutionary models for this complex belt.
4. A literature review and comparative analysis of lithologic types and formation names should be conducted by the major investigators within this belt. This should include

numerous field trips and discussions concerning the usage of different formation names for different belts of rocks. In particular attention should be focused on the type localities of particular formations and the rock-types either present or described under that type name. Following a thorough review of rock-types and names various belts of rocks should be given formal names where valid criteria suggest direct correlation with proper type localities. Improper type localities, either repetitive, impractical, or nontypical should be disbanded under the bylaws established by the U.S.G.S. rules of stratigraphic nomenclature. This type of analysis would ultimately come as a benefit to all geologists long burdened and confused by the myriad of stratigraphic names, many of which describe identical lithologies and stratigraphic relationships. Because of the severe tectonization within this belt, this task will not be easy, and may rely on interpretations in many instances. Different belts of rocks should only be given same unit status when a general agreement is reached regarding the validity of a particular across strike correlation.

REFERENCES

Atherton, M.P., and Edmunds, W.M., 1966, An electron-microprobe study of some zoned garnets from metamorphic rocks: Earth and Planetary Science Letters, v. 1,p.185-193.

Albee, A.L., 1957, Geology of the Hyde Park quadrangle,Vermont U.S.G.S. quadrangle map GQ-102, scale 1:62,500.

Albee, A.L., 1965a, A petrogenetic grid for the Fe-Mg silicates of pelitic schists: American Journal of Science, v.263, p.512-536.

Albee, A.L., 1965b, Phase equilibria in three assemblages of kyanite-zone pelitic schists, Lincoln Mountain quadrangle, central Vermont: Journal of Petrology, v.6, P.246-301.

Albee, A.L., 1968, Metamorphic zones in northern Vermont: *in Zen, Ean, White, W.S.,Hadley, J.J., Thompson, J.B., Jr., eds.,* Studies in Appalachian geology: northern and maritime: New York, Interscience, p.329-341.

Albee, A.L., 1972, Stratigraphic and structural relationships across the Green Mountain anticlinorium in north central Vermont: *in Doolan, B.L. and Stanley, R.S., eds.,* Annual Meeting, New England Intercollegiate Geological Conference, 64th, Burlington, University of Vermont, Guidebook to field trips in Vermont: p.179-183.

Armstrong, T.R., Kraus, J.F., Stanley, R.S., and Kimball, C.V., 1988a, Evolution of thrust fault geometries within the Taconian metamorphosed hinterland of central Vermont: Geological Society of America Abstracts with Programs, v.20, no.1.

Armstrong, T.R., Kraus, J.F., Stanley, R.S., 1988b, Graphitic rock-types in the Taconide zone of central Vermont: subdivision and possible modes of origin: Geological Association of Canada - Mineralogical Association of Canada-Canadian Society of Petroleum Geologists Joint Annual Meeting, St. John's, Newfoundland, v.13, p.A5.

Armstrong, T.R., 1989, The Vermont pre-Silurian cover sequence: regional tectonostratigraphic relationships based on detailed structural and lithologic analysis: *in, Doolan, B.L., and Colpron, M.,eds.,* Proceedings of the Québec-Vermont Appalachian Workshop, University of Vermont, Burlington, Vermont, p.63-66.

Armstrong, T.R., and Colpron, Maurice, 1989a, Taconian tectonic melange in northern New England and southern Quebec: The Humber - Dunnage suture: Geological Society of America, Abstracts with Programs, v.21, no.2.

Armstrong, T.R., and Colpron, Maurice, 1989b, Tectonic geometry of the Taconide zone in the White River Valley, central Vermont: A basis for lithotectonic correlations and passive margin modeling: Geological Society of America, Abstracts with Programs, v.21, no.2.

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

Bally, A.W., 1982, Musings over sedimentary basin evolution: Royal Society of London Philosophical Transactions, ser.A, v.305, p.325-328.

Bell,T.H., 1981, Foliation development-the contribution, geometry and significance of progressive bulk, inhomogeneous shortening: Tectonophysics, v.75, p.273-296.

Bell, T.H., and Johnson, S.E., 1989, Porphyroblast inclusion trails: the key to orogenesis: Journal of Metamorphic Geology, v.7, p.279-310.

Bence, A.E., and McLelland, J.M., 1976, Progressive metamorphism in Dutchess County, New York: New York State Geological Association 48th Annual Meeting, Field Guidebook, p. B-7-1 to B-7-27.

Berry, W.B.N., 1960, Graptolite faunas of the Marathon region, west Texas: Texas University Bureau of Economic Geology Publication 6005, 178p.

Billings, M.P., 1942, Structural Geology: Prentice-Hall, New York, 473p.

Biot, M.A., Theory of folding of stratified viscoelastic media and its implication in tectonites and orogenesis: Geological Society of America Bulletin, v.72, p.1595-1620.

Biot, M.A., 1965, Mechanics of incremental deformation: John Wiley & Sons, New York.

Bird, J.M., and Dewey, J.F., 1970, Lithosphere plate-continental margin tectonics and the evolution of the Appalachian orogen: Geological Society of America Bulletin, v.81, p.1031-1060.

Blake, M.C. Jr., and Landis, C.A., 1973, The Dun Mountain Ultramafic Belt-Permian oceanic crust and upper mantle in New Zealand: Journal Reserve of the U.S.G.S., v.1, no.5, p.529-534.

Bohlen, S.R., Wall, V.J., and Boettcher, A.L., 1983, Geobarometry in granulites, *in, S.K. Saxena, ed.,* Advances in Physical Geochemistry, V. 3, Springer-Verlag, New York, p.141-171.

Bosworth, William, 1987, Off-axis volcanism in the Gregory rift, east Africa: Implications for models of continental rifting: Geology, v.15, p.397-400.

Bonatti, Enrico, 1987, The rifting of continents: Scientific American, v.256, no.3, p.96-103.

Bouma, A.H., 1962, Sedimentology of some flysch deposits; a graphic approach to facies interpretation. Elsevier, Amsterdam, 168p.

Bouma, A.H., and Brouwer, A., eds., 1964, Turbidites: Elsevier, Amsterdam, vi, 264p.

Bowin, Carl, Purdy, G.M., Johnston, Chris, Shor, George, Lawver, Lawrence, Hartono, H.M.S., and Jezek, Peter, 1980, Arc-continent collision in Banda Sea region: American Association of Petroleum Geologists, v.80, p.868-915.

Boyer, S.E., and Elliot, D., 1982, Thrust systems: American Association of Petroleum Geologists Bulletin, v.66, p.1196-1230.

Brace, W.F., 1953, Geology of the Rutland area, Vermont: Vermont Geological Survey Bulletin no.6, 120p.

Burg, J.P., Iglesias, M., Laurent, Ph., Matte, Ph., and Ribeiro, A., 1981, Variscan intracontinental deformation: the Coimbra-Cordoba shear zone (SW Iberian Peninsula): Tectonophysics, v.78, p.161-177.

Burgess, C.F., Rosendahl, B.R., Sander, S., Burgess, C.A., Lambiase, L., Derksen, S., and Meader, N., 1987, The structural and stratigraphic evolution of Lake Tanganyika: A case study of continental rifting: American Association of Petroleum Geologists Memoir 66.

Byrne, Tim, and Lucas, S., 1988, Rheological evolution of the decollment in A- and B-type subduction zones: Geological Society of America, Abstracts with Programs, v.20, no.7.

Cady, W.M., 1945, Stratigraphy and structure of west-central Vermont: Geological Society of America Bulletin, v. 56, p. 515-558.

Cady, W.M., 1956, Bedrock geology of the Montpelier quadrangle, Vermont: U.S.G.S. quadrangle map GQ-79, scale 1:62,500.

Cady, W.M., 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Québec: Geological Society of America Bulletin, v. 71, p.531-576.

Cady, W.M., 1969, Regional tectonic synthesis of northwestern New England and adjacent Quebec: Geological Society of America Memoir 120, 181p.

Cady, W.M., Albee, A.L., Murphy, J.F., 1962, Bedrock geology of the Lincoln Mountain quadrangle, Vermont: U.S.G.S. geologic quadrangle map, GQ-164, scale 1:62,500.

Carmichael, D.M., 1969, On the mechanism of prograde metamorphic reactions in quartz-bearing pelites: Contributions in Mineralogy and Petrology, v. 20, p. 244-267.

Chang, H.C., Ern, E.H., Jr., and Thompson, J.B., Jr., 1965, Bedrock Geology of the Woodstock quadrangle, Vermont: Vermont Geological Survey Bulletin no.29, 65p.

Clark, T.H., 1934, Structure and stratigraphy of southern Quebec: Geological Society of America Bulletin, v.45, p.1-20.

Cobbold, P.R., and Quinquis, H., 1980, Development of sheath folds in shear regimes: Journal of Structural Geology, v.2, p. 119-126.

Coish, R.A., Fleming, F.S., Larsen, M., Poyner, R., and Seibert, J., 1985, Early rift history of the Proto-Atlantic Ocean: Geochemical evidence from metavolcanic rocks in Vermont: American Journal of Science, v.285, p.351-378.

Coish, R.A., Perry, D.A., Anderson, C.D., and Bailey, D., 1986, Metavolcanic rocks from the Stowe Formation, Vermont: remnants of ridge and intraplate volcanism in the Iapetus Ocean: American Journal of Science, v.286, p.1-28.

Coish, R.A., 1987, Regional geochemical variations in greenstones from the central Vermont Appalachians: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - 79th Annual Meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p. 345-350.

Colpron, Maurice, Dowling, W.M., Doolan, B.L., 1987, Stratigraphy and structure of the Sutton area, southern Québec: Construction and destruction of the western margin of the Late Precambrian Iapetus: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p. 443-462.

Colpron, Maurice, and Armstrong, T.R., 1988, A possible origin for the admixture of rift stage and post-rift rocks within the Sutton Schist of southern Quebec and correlative sequences of Vermont: Geological Association of Canada-Mineralogical Association of Canada-Canadian Society of Petroleum Geologists Joint Annual Meeting, St. John's, Newfoundland, v.13, p. A23.

Cooper, A.F., 1972, Progressive metamorphism of metabasic rocks from the Haast Schist group of southern New Zealand: Journal of Petrology, v.13, p.457-492.

Cosgrove, J.W., 1976, The formation of crenulation cleavage: Journal of the Geologic Society of London, v.132, p.155-178.

Crawford, M.L., 1966, Composition of plagioclase and associated minerals in some schists from Vermont, U.S.A., and South Westland, New Zealand, with inferences about the peristerite solvus: Contributions in Mineralogy and Petrology, v.13, p.269-294.

Dale, T.N., 1915, The calcite marble and dolomite of eastern Vermont: U.S.G.S. Bulletin no.589, 67 p.

Dallmeyer, R.D., 1975, Ar^{39}/Ar^{40} incremental-release ages of biotite and hornblende from the Cortlandt and Rosetown plutons, New York, and their regional implications: Journal of Geology, v.83, p.629-643.

Dallmeyer, R.D., and Sutter, J.F., 1976, $Ar^{39}/Ar^{40}Ar$ incremental release ages of biotite and hornblende from variably retrograded basement gneisses of the northeasternmost Reading Prong, New York: Their bearing of early Paleozoic metamorphic history: American Journal of Science, v.276, p.731-747.

Davis, D., Suppe, J., and Dahlen, F.A., 1983, Mechanics of fold-and-thrust belts and accretionary wedges: Journal of Geophysical Research, v.88, p.1153-1172.

De Raaf, J.F.M., Reading, H.G., and Walker, R.G., 1965, Cyclic sedimentation in the Upper Carboniferous of North Devon, England: Sedimentology, v.4.

DelloRusso, Vincent, 1986, Geology of the eastern part of the Lincoln massif, central Vermont: [M.S. thesis], University of Vermont, Burlington, Vermont, 255p.

DelloRusso, Vincent, and Stanley, R.S., 1986, Geology of the northern part of the Lincoln massif, central Vermont: Vermont Geological Survey Special Bulletin no. 8, 35p. and 4 plates.

Dewey, J.F., 1982, Plate tectonics and the evolution of the British Isles: Journal of the Geological Society of London, v.139, p. 1-41.

DiPietro, J.A., 1983, Geology of the Starksboro area, Vermont: Vermont Geological Survey Special Bulletin, no. 4, 14 p.

Dodson, M.H., 1973, Closure temperature in geochronological and petrological systems: Contributions to Mineralogy and Petrology, v.40, p.259-274.

Dodson, M.H., 1979, Theory of cooling ages, *in, Jager, E, and others, eds.,* Lectures in Isotope Geology, Springer-Verlag, New York, p. 194-202.

Doll, C.G., Cady, W.M., Thompson, J.B.,Jr., and Billings, M.P., 1961, Centennial Geologic Map of Vermont: Montpelier, Vermont, Vermont Geological Survey, 1:250000.

Doolan, B.L., Gale, M.H., Gale, P.N., Hoar, R.S., 1982, Geology of the Québec reentrant: Possible constraints from early rifts and the Vermont-Québec serpentine belt: *in, St.Julien, P., and Beland, J., eds.,* Major structural zones and faults of the northern Appalachians: Geological Association of Canada Special Paper no.7, p.87-115.

Doolan, B.L., Mock, Timothy, and McBean, Alan, 1988, Stratigraphy and structure of the Camels Hump Group along the Lamoille River transect, northern Vermont: *in, Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th Annual Meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.152-191.

Dowling, W.M., 1987, Depositional environment of the Lower Oak Hill Group, southern Québec: Implications for the Late Precambrian breakup of North America in the Québec reentrant: [M.S. thesis], University of Vermont, Burlington, Vermont, 132p.

England, P.C., and Richardson, S.W., 1977, The influence of erosion upon the mineral facies of rocks from different metamorphic environments: Journal of the Geological Society of London, v.134, p.201-213.

Essene, E., 1982, Geologic thermometry and barometry: Reviews in Mineralogy, v.10, p.153-206.

Folk, R.L., 1968, Petrology of sedimentary rocks, Austin Texas, 170p.

Folk, R.L., and Land, L.S., 1975, Mg/Ca ratio salinity: two controls over crystallization of dolomite: American Association of Petroleum Geologists, v.59, p.60-68.

Frey, M., 1970, The step from diagenesis to metamorphism in pelitic rocks during alpine orogenesis: Sedimentology, v.15, p.261-279.

Garlick, G.D., and Epstein, S., 1967, Oxygen isotope ratios in coexisting minerals of regionally metamorphosed rocks: Geochim. et Cosmochim. Acta, v.31, p.181-214.

Graham, C.M., 1974, Metabasite amphiboles of the Scottish Dalradian: Contributions to Mineralogy and Petrology, v.47, p.165-185.

Hall, L.M., and Robinson, P., 1982, Stratigraphic-tectonic subdivisions of southern New England: *in, St. Julien, P., and Beland, J., eds.,* Major Structural Zones and Faults of the Northern Appalachians: Geological Association of Canada Special Paper 24, p.15-42.

Hames, W.E., Tracy, R.J., and Bodnar, R.J., 1989, Postmetamorphic unroofing history deduced from petrology, fluid inclusions, thermochronometry, and thermal modeling: an example from southwestern New England: Geology, v.17, p.727-730.

Hamilton, W.B., 1988, Plate tectonics and island arcs: Geological Society of America Bulletin, v.100, p.1503-1527.

Harper, C.T., 1968, Isotopic ages from the Appalachians and their tectonic significance: Canadian Journal of Earth Sciences, v.5, p.49-59.

Harte, B., and Graham, C.M., 1975, The graphical analysis of greenschist to amphibolite facies mineral assemblages in metabasites: Journal of Petrology, v.16, p.347-370.

Harwood, D.S., 1972, Tectonic events in the southwestern part of the Berkshire anticlinorium, Massachusetts and Connecticut: Geological Society of America, Abstracts with Programs, v.4, no.1, p.19.

Harwood, D.S., 1975, Fold-thrust tectonics in the southern Berkshire massif, Connecticut and Massachusetts: *in*N.E.I.G.C., 67th Annual Meeting, Guidebook for fieldtrips in western Massachusetts, northern Connecticut, and adjacent areas of New York: New York, City College of C.U.N.Y., Dept. of Earth and Planetary Sciences, p.122-143.

Hatch N.L., 1982, The Taconian Line in western New England and its implications to Paleozoic tectonic history: *in, St. Julien, P., and Béland, J.,eds.,*Major Structural Zones and Faults of the Northern Appalachians: Geological Association of Canada Special Paper 24, p67-86.

Hatch, N.L., 1987, Lithofacies, stratigraphy, and structure in the rocks of the Connecticut Valley trough, eastern Vermont: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont-Volume 2, 79th Annual Meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.192-212.

Hatch, N.L., Jr., and Stanley, R.S., 1973, Some suggested stratigraphic relations in part of southwestern New England: U.S.G.S. Bulletin 1380, 83 p.

Haydock, S.R., 1988, Bedrock geology of the Waitsfield area, central Vermont [M.S.thesis]: University of Vermont, Burlington, Vermont, 245p.

Helmold, K.P, and van de Kamp, P.C., 1984, Diagenetic mineralogy and controls on albitization and laumontite formation in Paleogene arkoses, Santa Ynez Mountains, California: *in, McDonald, D.A., and Surdam, R.C., eds.,* Clastic Diagenesis, American Association of Petroleum Geologists Memoir 37, p.239-276.

Herz, N.L., 1958, Bedrock geology of the Cheshire quadrangle, Massachusetts: U.S.G.S. quadrangle map GQ-108.

Hitchcock, E., Hitchcock, E. Jr., Hager, A.D., Hitchcock, C.H., 1861, Report on the geology of Vermont, Claremont, N.H., 982p.

Hobbs, B.E., Means, W.D., and Williams, P.F., 1976, Structural geology: New York, John Wiley and Sons, 571 p.

Hodges, K.V., Bartley, J.M., and Burchfield, B.C., 1982, Structural evolution of an A-type subduction zone, Lofoten-Rombak area, northern Scandinavian Caledonides: Tectonics, v.1, p.441-462.

Hollister, L.S., 1966, Garnet zoning: an interpretation based on the Rayleigh fractionation model: Science, v.154, p.1647-1651.

Illing, L.V., Wells, A.J., and Taylor, J.C.M., 1965, Penecontemporary dolomite in the Persian Gulf, *in, Pray, L.C., and Murray, R.C. (eds.),* Dolomitization and limestone genesis - a symposium, Society of Economic Paleontologists and Mineralogists Special Publication no.13, p.89-111.

Karabinos, Paul, 1987, Tectonic setting of the northern part of the Green Mountain massif, Vermont: *in Westerman, D.S., ed.,*Guidebook for Field Trips in Vermont, Volume 2, 79th Annual Meeting of the New England Intercollegiate GeologicalConference, Montpelier, Vermont, p.464-491.

Karabinos, Paul, 1988, Tectonic significance of Basement-cover relationships in the Green Mountain Massif, Vermont: Journal of Geology, v.96, p.445-454.

Karabinos, Paul, and Laird, Jo, 1988, Structure and metamorphism from Jamaica to theAthens dome: *in, Bothner, W.A., ed.,*80th Annual Meeting, N.E.I.G.C., Keene St. College, Keene, New Hampshire, p.281-292.

Kay, M., 1947, Geological nomenclature and the craton: American Association of Petroleum Geologists Bulletin, v.31, p.1289-1293.

Keith, Arthur, 1932, Stratigraphy and structure of northwestern Vermont: Washington Academy Journal of Science, v.22, p.357-379, 393-406.

Kinsman, D.D.J., 1975, Rift valley basins and sedimentary history of trailing continental margins: *in, Fischer, G.A., and Judson, S.,eds.,* Petroleum and Global Tectonics: Princeton University Press, p.83-126.

Knipe, R.J., and Wintsch, R.P., 1985; Heterogeneous deformation, foliation development, and metamorphic processes in a polyphase mylonite: *in, Thompson, A.B., and Rubie, D.C.,eds.,* Metamorphic Reactions: Advances in Physical Geochemistry, v.4, Springer-Verlag, New York, p. 180-210.

Kraus, J.F., Armstrong, T.R., Kimball, C.V., Stanley, R.S., 1988, Origin of the Pinney Hollow, Ottauquechee, and Stowe Formations of central Vermont: Geological Society of America Abstracts with Programs, v.20. no.1, p.35.

Kraus, J.F., 1989, Geology of the Northfield Mountains, central Vermont [M.S. thesis]: University of Vermont, Burlington, Vermont, 178 p.

Krumbein, W. C., and Garrels, R. M., 1952, Origin and classification of chemical sediments in terms of Ph and oxidation-reduction potentials: Journal of Geology, v.60, p.1-33.

Kumarapelli, P.S., Dunning, G.R., Pintson, H., and Shaver, J., 1989, Geochemistry and U-Pb zircon age of comenditic metafelsites of the Tibbit Hill Formation, Québec Appalachians: Canadian Journal of Earth Sciences.

Laird, J., and Albee, A.L., 1981, Pressure, temperature and time indicators in mafic schist: their application to reconstructing the polymetamorphic history of Vermont: American Journal of Science, v.287, p.127-175.

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

Laird, J., and Bothner, W.A., 1986, Metamorphism in the northern Green Mountains of Vermont: *in, Robinson, P. and Elbert, D.C., eds.,* Field Trip Guidebook: Regional Metamorphism and Metamorphic Phase Relations in Northwestern and Central New England: Contribution no. 59, Department of Geology and University of Massachusetts, Amherst, Massachusetts, p.11-36.

Laird, J., 1987, Metamorphism of the pre-Silurian rocks, central Vermont: *in, Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont-Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.339-344.

Lanphere, M.A., and Albee, A.L., 1974, Ar^{39}/Ar^{40} age measurements in the Worcester Mountains: evidence of Ordovician and Devonian metamorphic events in northern Vermont: American Journal of Science, v.274, p 545-555.

Lanphere, M.A., Laird, J., and Albee, M.A., 1983, Interpretation of Ar^{39}/Ar^{40} ages of polymetamorphic mafic and pelitic schist in northern Vermont: Geological Society of America Abstracts with Programs, v.15, p.147.

Lapp, E.T., and O'Loughlin, 1986, Tectonic and depositional sequences in polydeformed, polymetamorphosed rocks of the Camel's Hump Group, Lincoln gap area, central Vermont: Geological Society of America, Abstracts with Programs, v.18, no.1.

Lapp, E.T., and Stanley, R.S., 1987, Geology of the Mt. Grant-South Lincoln area, central Vermont: Vermont Geological Survey Special Bulletin No.7, 20 p. and 3 plates.

Law, R.D., Knipe, R.J., and Dayan, H., 1984, Strain path partitioning within thrust sheets: microstructural and petrofabric evidence from the Moine Thrust zone at Loch Eriboll, northwest Scotland: Journal of Structural. Geol., v.6, p.477-497.

Le Pichon, X., and Sibuet, J.C., 1981, Passive margins: a model of formation: Journal of Geophysical research, v.86, p.3708-3720.

Lister, G.S., and Snoke, A.W., 1984, S-C mylonites: Journal of Structural Geology, v.6, p.617-638.

Lister, G. S., Etheridge, M.A., and Symonds, P.A., 1986, Detachment faulting and the evolution of passive continental margins: Geology, v.13, p.246-250.

Lowe, D.R., 1982, Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents: Journal of Sedimentary Petrology, v.52, no.1, p.279-291.

Means, W.D., 1977, Experimental contributions to the study of foliations in rocks: a review of research since 1960: Tectonophysics, v.100, p.413-447.

McKenzie, D.P., 1978, Some remarks on the development of sedimentary basins: Earth and Planetary Science Letters, v.40, p.25-32.

McDougall, I., and Harrison, T.M., 1988, Geochronology and thermochronometry by the Ar^{39}/Ar^{40}method: Oxford University Press, New York, Monographs on Geology and Geophysics, v.9, 212 p.

Milliman, J. D., 1974, Marine carbonates, Springer-Verlag, New York, 375 p.

Miyashiro, A., 1961, Evolution of metamorphic belts: Journal of Petrology, v.2, p.277-311.

Miyashiro, A., 1973, Metamorphism and Metamorphic Belts: George Allen & Unwin, London.

Nicolas, A., and Poirier, J.P., 1976, Crystalline plasticity and solid state flow in metamorphic rocks: New York, John Wiley and Sons, 444 p.

O'Loughlin, S.B., and Stanley, R.S., 1986, Geology of the Mt. Abraham-Lincoln Gap area: Vermont Geological Survey Special Bulletin No.6, 29 p. and 3 plates.

Osberg, P.H., 1952, The Green Mountain Anticlinorium in the vicinity of Rochester and East Middlebury, Vermont: Vermont Geological Survey, Bulletin no.5, 127p.

Papike, J.J., and Cameron, M., 1976, Crystal chemistry of silicate minerals of geophysical interest: Reviews of Geophysics and Space Physics, v.14, no.1, p.37-80.

Passchier, C.W., and Simpson, C., 1986, Porphyroclast systems as kinematic indicators: Journal of Structural Geology, v.8, p.831-843.

Pearce, J.A., and Cann, J.R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analyses: Earth and Planetary Science Letters, v.19, p.290-300.

Perry R.L., 1928, The geology of Bridgewater and Plymouth Townships, Vermont: Vermont State Geologist's 16th Report, p.1-64.

Poirier, J.P., 1980, Shear localization and shear instability in materials in the ductile field: Journal of Structural Geology, v.2, p.135-142.

Potter, D.B., 1979, A traverse across the central part of the Taconic allochthon: *in Skehan, J.W., and Osberg, P.H., eds.,* International Geological Correlation Project 27: The Caledonides in the United States of America. Geological excursions in the northern Appalachians: Weston, Ma, Weston Observatory, p.225-250.

Powell, Roger, and Holland, Tim, 1990, Calculated mineral equilibria in the pelite system, KFMASH (K2O-FeO-MgO-Al2O3-SiO2-H2O): Am. Mineral., v.75, p.367-380.

Prewitt, J.M., 1989, Bedrock geology of the Granville Gulf-Warren area, central Vermont [M.S. thesis]: University of Vermont, Burlington, Vermont, 143p.

Prindle, L.M., and Knopf, E.B., 1932, Geology of the Taconic quadrangle: American Journal of Science, 29, p. 257-302.

Pumpelly, R., Wolff, J.E., and Dale, T.N., 1894, Geology of the Green Mountains, Part III, Mount Greylock: its areal and structural geology: U.S.G.S. Mon., v.22, 158 p.

Quinlan, Garry, 1988, The thermal signature of tectonic environments: in, Nisbet, E.G., Fowler, C.M.R., and Ghent, E.D., (eds.), Short course on heat, metamorphism, and tectonics: Mineralogical Association of Canada, v.14, p.213-257.

Ramsay, J.G, 1967, Folding and fracturing of rocks: McGraw Hill, New York.

Ramsay, J.G., and Huber, M.I., 1983, The techniques of modern structural geology, Vol. 1: Academic Press, New York.

Ratcliffe, N.M., 1975, Cross section of the Berkshire Massif at 42N: Profile of a basement reactivation zone, *in* N.E.I.G.C., 67th Annual Meeting, Guidebook for fieldtrips in western Massachusetts, northern Connecticut, and adjacent areas of New York: New York, City College of C.U.N.Y., Dept. of Earth and Planetary Sciences, p.186-216.

Ratcliffe, N.M., 1979, Field guide to the Chatham and Greylock slices of the Taconic allochthon, and their relationship to the Hoosac-Rowe sequence, *in, Friedman, G.M., ed.,* Annual Meeting, N.Y. State Geological Association and the N.E.I.G.C.,71st Annual Meeting, Guidebook to Field Trips: Troy, New York, R.P.I., p.388-425.

Ratcliffe, N.M., 1987, Basaltic rocks in the Rensselaer Plateau and Chatham slices of the Taconic allochthon: Chemistry and tectonic setting: Geological Society of America Bulletin, v.99, p. 511-528.

Ratcliffe, N.M., Burton, W.C., Sutter, J.F., and Mukasa, S.B., 1988, Stratigraphy, structural geology and thermochronology of the northern Berkshire massif and southern Green Mountains: *in, Bothner, W.A., ed.,* 80th Annual Meeting, N.E.I.G.C., Keene St. College, Keene, New Hampshire, p.1-24, 126-135.

Reineck, H. E., and Singh, I. B., 1975, Depositional sedimentary environments, Springer-Verlag, New York, 439 p.

Ribbe, P.H., 1983, The chemistry, structure, and nomenclature of feldspars: *in, Ribbe, P.H., ed.,* Feldspar Mineralogy, Reviews in Mineralogy, v.2, p.1-19.

Richardson, C.H., 1924, The terranes of Bethel, Vermont: Vermont State Geologist 13th Report, p.77-103.

Robinson, Peter, Tucker, R.D., and Hollocher, K.T., 1989, The nature of the Bronson Hill Zone during the Taconian orogeny: *in, Doolan, B.L., and Colpron, M.eds.,* Proceedings of the Québec-Vermont Appalachian Workshop, University of Vermont, Burlington, Vermont, p.26-28.

Rodgers, John, 1968, The eastern edge of the North American continent during the Cambrian and Early Ordovician, *in, Zen, E-an, White, W.S., Hadley, J.J., Thompson, J.B., Jr., eds.,* Studies in appalachian geology: northern and maritime: New York, Interscience, p.141-149.

Rowley, D.B., 1982, New methods for estimating displacements of thrust faults affecting Atlantic-type shelf sequences: With applications to the Champlain thrust, Vermont: Tectonics, v.1, no.4, p.369-388.

Rowley, D.B. and Kidd, W.S.F., 1981, Stratigraphic relationships and detrital composition of the Medial Ordovician flysch of western New England: Implications for the tectonic evolution of the Taconic orogeny: Journal of Geology, v.89, p.199-218.

Rowley, D.B., Kidd, W.S.F., and Delano, L.L., 1979, Detailed stratigraphic and structural features of the Giddings Brook slice of the Taconic allochthon in the Granville area: *in, Friedman, G.M., ed.,* Annual Meeting, N.Y. State Geological Association and the N.E.I.G.C., 71st Annual Meeting, Guidebook to Field Trips: Troy, New York, R.P.I., p.388-425.

Roy, D.L., 1982, Structure and stratigraphy in the Cambrian-Ordovician sequence in the ultramafic belt near Troy, Vermont: [M.S. Thesis], University of Vermont, Burlington, Vermont, 122 p.

Royden, Leigh, Sclater, J.G., and Von Herzen, R.P., 1980, Continental margin subsidence and heat flow: Important parameters in formation of petroleum hydrocarbons: American Association of Petroleum Geologists Bulletin, v.64, no.2, p.173-187.

Schmid, S.M., 1982a, An introduction to the analysis of textures of deformed rocks: Structural Geology and Tectonics Division of the Geological Society of America (unpubl. shortcourse notes).

Schmid, S.M., 1982b, Microfabric studies as indicators of deformation mechanisms and flow laws operative in mountain building, in Hsu, K.J., ed., Mountain building processes: New York, Academic Press, p.95-110.

Sheppard, S.M.F., and Schwarcz, H.P., 1970, Fractionation of carbon and oxygen isotopes and magnesium between coexisting metamorphic calcite and dolomite: Contributions to Mineralogy and Petrology, v.26, p.161-198.

Simpson, C., and Schmid, S.M., 1983, An evaluation of criteria to deduce the sense of movement in sheared rocks: Geological Society of America Bulletin, v.94, p.1281-1288.

Skehan, J.W., 1961, The Green Mountain anticlinorium in the vicinity of Wilmington and Woodford Vermont: Vermont Geological Survey Bulletin no.17, 159 p.

Slivitzsky A., and St. Julien, P., 1987, Compilation géologique de la région de l'Estrie Beauce: Ministére de l'Energie et des Resources du Québec, MM 85-04, 40 pages, 1 map scale: 1:250,000

Smith, J.V., 1983, Phase equilibria of plagioclase: *in, Ribbe, P.H., ed.,* Feldspar Mineralogy, Reviews in Mineralogy, Volume 2 (second ed.), Mineralogical Society of America, p.223-240.

Spear, F.S., and Cheney, J.T., 1989, A petrogenetic grid for pelitic schists in the system SiO2-Al2O3-FeO-MgO-K2O-H2O: Contributions to Mineralogy and Petrology, v.101, p.149-164.

Stanley, R.S., Roy, D.L., Hatch, N.L., and Knapp, D.A., 1984, Evidence for tectonic emplacement of ultramafic and associated rocks in the pre-Silurian eugeoclinal belt of western New England - vestiges of an ancient accretionary wedge: American Journal of Science, v.284, p.559-595.

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

Stanley, R.S., Haydock, S.R., and Prewitt, J.M., 1986, Tectonic geology of the Waitsfield-Granville Gulf area, central Vermont: Geological Society of America Abstracts with Programs, v.18, no.1, p.69.

Stanley, R.S., DelloRusso, Vincent, O'Loughlin, S.B., Lapp, E.T., Armstrong, T.R., Prewitt, Jeffery, Kraus, J.F., Walsh, Gregory, 1987a, A transect through the pre-Silurian of central Vermont: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.272-295.

Stanley, R.S., Armstrong, T.R., Kraus, J.F., Walsh, Gregory, Prewitt, Jeffery, Kimball, C.V., Cua, A.K., 1987b, The pre-Silurian hinterland along the valleys of the White and Mad Rivers: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.314-338.

Stanley, R.S., and Armstrong, T.R., 1988a, Palinspastic analysis of the Taconian Hinterland as shown in the central Vermont transect: Symposium on Arc-Continent Collision and Orogenic Sedimentation, Eastern Taiwan and Ancient Analogues: Princeton University, Abstracts and Program, p.34.

Stanley, R.S., DelloRusso, Vincent, Armstrong, T.R., Kraus, J.F., Walsh, G.J., O'Loughlin, S.B., Lapp, E.T., Prewitt, J.P., Kimball, C.V., and Cua, A.K., 1988b, The tectonics of the pre-Silurian rocks of central Vermont: Geological Society of America Abstracts with Programs, v.20, no.1.

Stanley, R.S., Armstrong, T.R., and Coish, R.A., 1988c, Palinspastic analysis of the Taconian Hinterland as shown in the central Vermont Transect: Geological Society of America Abstracts with Programs, v.20, no.7, p.395.

Stanley, R.S., Armstrong, T.R., Coish, R.A., and Laird, Jo, 1989, Palinspastic analysis of the Taconian Hinterland as shown in the central Vermont Transect: Geological Society of America, Abstracts with Programs, v.21, no.2.

Suppe, John, 1983, Geometry and kinematics of fault-bend folding: American Journal of Science, v.283, p.684-721.

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

Tauvers, P.R., 1982, Basement-cover relationships in the Lincoln area, Vermont [M.S.thesis]: Burlington, 177 p.

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

Thompson, A.B., and England, P.C., 1984, Pressure-temperature-time paths of regional metamorphism. II. Their inference and interpretation using mineral assemblages in metamorphic rocks: Journal of Petrology, v.25, p.929-955.

Thompson, J.B. Jr., 1957, The graphical analysis of mineral assemblages in pelitic schists: American Mineralogist, v.42, p.842-858.

Thompson, J.B., Jr., 1972, Lower Paleozoic rocks flanking the Green Mountain anticlinorium: *in Doolan, B.L. and Stanley, R.S., eds.,* Annual Meeting, New England Intercollegiate Geological Conference, 64th, Burlington, University of Vermont, Guidebook to field trips in Vermont: p.215-229.

Thompson, J.B. Jr., Rosenfeld, J.L., and Downie, E.A., 1986, Basement-cover relations in the Chester and Athens Domes and adjacent terranes, Vermont: Geological Society of America, Abstracts with Programs, v.18, no.1.

Thompson, P.J., and Thompson, T.B., 1987, Winooski River transect: Refolded folds and thrust faults in the Green Mountain anticlinorium: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.492-504.

Tracy, R.J., Robinson, P., and Wolff, R.A., 1984, Metamorphosed ultramafic rocks in the Bronson Hill anticlinorium, central Massachusetts: American Journal of Science, v.284, p.530-558.

Tracy, R.J., 1982, Compositional zoning and inclusions in metamorphic minerals: *in, Ferry, J.F., ed.,* Characterization of Metamorphism through Mineral Equilibria: Reviews in Mineralogy, Volume 10, Mineralogical Society of America, p.355-397.

Tracy, R.J., Miller, S.J., and Armstrong, T.R., 1990, Selection of garnet and biotite compositions for garnet-biotite thermometry in amphibolite facies metapelites: Geological Society of America, Abstracts with Programs, v.22, no.1.

Von Der Borch, C.C., 1978, Continent-Island arc collision in the Banda Arc: Tectonophysics, v.54, p.169-193.

Walcott, C.D., 1888, The Taconic system of Emmons, and the name in geologic nomenclature: American Journal of Science, v.35, p.229-242, 307-327, 394-401.

Walker, R.G., 1965, The origin and significance of the internal sedimentary structures of turbidites: Proceedings of the Yorkshire Geological Society, v.35, pt.1, no.1, p.1-32.

Walker, R.G., 1978, Deep-water sandstone facies and ancient submarine fans: Models for exploration for stratigraphic traps: American Association of Petroleum Geologists Bulletin, v.62, p.932-966.

Walker, T. R., Waugh, B., and Grone, A. J., 1978, Diagenesis in first cycle desert alluvium of Cenozoic age, southwestern United States and northwestern Mexico: Geological Society of America Bulletin, v.89, p.19-32.

Walsh, G.J., and Stanley, R.S., 1987, Tectonic lithostratigraphic units of the Camel's Hump Group in the vicinity of Fayston, Vermont: Geological Society of America Abstracts with Programs, v.20, no.1, p.76.

Walsh, G.J., 1989, Tectonic geology of the Fayston area, central Vermont [M.S. thesis]: University of Vermont, Burlington, Vermont.

Warren, M.J., 1989, Tectonic evolution of the southern Lincoln massif and rift clastic cover sequence, central Vermont: *in, Doolan, B.L., and Colpron, M.,eds.,*Proceedings of the Québec-Vermont Appalachian Workshop, University of Vermont, Burlington, Vermont, p.58-62.

Washington, P.A., 1987, Cleavage vs folding vs. thrusting: relative timing of structural events in the central Champlain valley: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.351-368.

Westerman, D.S., 1987, Structures in the Dog River Fault Zone between Northfield and Montpelier, Vermont: *in Westerman, D.S., ed.,* Guidebook for Field Trips in Vermont - Volume 2, 79th annual meeting of the New England Intercollegiate Geological Conference, Montpelier, Vermont, p.109-132.

White, S., Burrows, S.E., Carreras, J., Shaw, N.D., and Humphreys, F.J., 1980, On mylonites in ductile shear zones: Journal of Structural Geology, v.2, p.175-188.

Wilde, P., and Berry, W.B.N., 1987, The role of oceanographic factors in the generation of global bio-events: *in, Walliser, O.H., ed.,* Global Bio-events: A Critical Approach: Springer-Verlag, Berlin, p.75-91.

Williams, P.F., 1984, Multiply deformed terranes - problems of correlation: Journal of Structural Geology, v.7, p.269-280.

Wise, D.U., Dunn, D.E., Engelder, J.T., Geiser, P.A., Hatcher, R.D., Kish, S.A., Odom, A.L., and Schamel, S., 1984, Fault-related rocks: suggestions for terminology: Geology, v.12, p.391-394.

Wojtal, S., and Mitra, G., 1986, Strain hardening and strain softening in fault zones from foreland thrusts: Geological Society of America Bulletin, v.57, p.674-687.

Wybrecht, E., Duplay, J., Piqué, A., and Weber, F., 1985, Mineralogical and chemical evolution of white micas and chlorites, from diagenesis to low grade metamorphism; data from various size fractions of greywackes (Middle Cambrian, Morocco): Mineralogical Magazine, v.49, p.401-411.

Yund, R.A., and Tullis, J., 1983, Subsolidus phase relations in the alkali feldspars with emphasis on coherent phases: *in, Ribbe, P.H., ed.,* Feldspar Mineralogy, Reviews in Mineralogy, Volume 2 (second ed.), Mineralogical Society of America, p.141-170.

Zartman, R.E., and others, 1970, A Permian disturbance of K-Ar radiometric ages in New England: Its occurrence and cause: Geological Society of America Bulletin, v.81, p.3359-3374.

Zen, E-an, 1961, Stratigraphy and structure at the north end of the Taconic Range in west-central Vermont: Geological Society of America Bulletin, v.72, p.293-338.

Zen, E-an, 1964, Taconic stratigraphic names: Definitions and synonymies: U.S.G.S. Bulletin 1174, 20 p.

Zen, E-an, 1967, Time and space relationships of the Taconic allochthon and autochthon: Geological Society of America Special Paper no.97, 107 p.

Zen, E-an, ed., Goldsmith, Richard, Ratcliffe, N.M., Robinson, Peter, and Stanley, R.S., compilers, 1983, Bedrock geologic map of Massachusetts: U.S.G.S. and the Commonwealth of Massachusetts, Dept. of Public Works, and Sinnott, J.A., State Geologist, scale 1:250,000.

APPENDIX I
FAULT ZONE CRITERIA

	Pre-metamorphic	Syn-metamorphic	Postmetamorphic
1. Upper and lower plate truncations:	X	X	X
2. Mylonitic fabric:		X	
3. Asymmetric fabric/folds:		X	X
4. Cataclastic fabric:			X
5. "Exotic" fault sliver(s):	X	X	X
6. Slickenlines:	X	X	X

Terminology:

Pre-metamorphic: Development prior to M1 and M2 peak metamorphism and mineral growth

Syn-metamorphic: Fabric development coeval with M1 and M2 peak thermal conditions

Post-metamorphic: Brittle fabric development subsequent to M1 and M2 peak thermal conditions

APPENDIX II
Linear and Planar Fabric Element Nomenclature

The various foliations, bedding surfaces, shear bands, fold hinges, and lineations observed within the study area were identified and subsequently labeled chronologically with regard to the locally dominant planar fabric, arbitrarily called **Sn**. In this way all planar and linear elements are chronologically listed as relatively younger (n+) or older (n-) than the dominant (n) fabric. Planar elements are given the designator "S" whereas linear elements are labeled "L". For example, a secondary foliation that locally cross cuts the dominant foliation in a particular outcrop would be labeled Sn+1 (if it was the next youngest planar element relative to Sn). Any linear element(s) associated with Sn+1 would be labeled Ln+1. It may be desirable to separate different linear elements such as fold hinges and mineral lineations. Therefore, labels such as Mln+1 and Fln+1 could be used for the mineral lineation and fold hinge data respectively. An element older than the Sn foliation would either be called Sn-1 or Ln-1.

The benefit of using this relative chronologic system rather than absolute numerical values (such as F1, F2, etc.) is that it allows for objective systematizing of structural data found in a single outcrop or well controlled locality. This is particularly important in orogenic belts where deformation is multi-phased and of similar orientation. It also allows for continuous and progressive addition of younger and older fabrics as they are subsequently observed. The potential problem of this system is that it is based on localcross cutting relationships; assignment of Sn schistosity in one locality does not imply cogeneity with Sn from a different area. Correlation of fabrics from different localities requires careful evaluation of different structural and metamorphic factors, such as progressive changes in fold geometry, hinge orientations, foliation development, and the relationship between distinct metamorphic growth phases, mineral assemblages, and deformation elements. This type of evaluation requires a large data base and comparison of the different fabric elements from different areas. Further labeling of the fabric elements based on their areal distribution may be necessary. In this study data has been labeled according to its position in one of three domains; western, central, and eastern. Designators are therefore modified from "Sn+1" to "Snw+1" to show their occurrence within the western domain. Domain boundaries may be arbitrary lines, follow topographic changes, or marked structural boundaries. The latter is not favored since it may impose subjectivity that may become ingrained in the fabric comparison and correlation. Following a thorough structural analysis and regional fabric correlation the relative age designators may be converted to absolute numerical values, starting with the oldest fabric observed. Caution should be taken in doing this in that any following studies may find other ages of fabric that will require full modification of the absolute order scheme.

APPENDIX III

Correlation of Deformational Fabric Nomenclature

OLD	NEW
1. WESTERN DOMAIN	
Foliations:	
Snw	S2.1
Snw+1	S2.2
Folds:	
Fnw	F2.1
Fnw+1	F2.2
Fnw+2	F3
Lineations:	
Lnw+1	L2.2
2. CENTRAL DOMAIN	
Foliations:	
Snc-1	S2.1
Snc	S2.2
Snc+1	S2.3
Snc+2 (shear bands)	S2.3s
Snc+3 (postmetamorphic)	S3
Folds:	
Fnc-1	F2.1
Fnc (sheath folds)	F2.2
Fnc+1	F2.3
Fnc+2(crenulate folds)	F3
Lineations:	
Lnc-1	L2.1
Lnc	L2.2
3. EASTERN DOMAIN	
Foliations:	
Sne-1	S2.1
Sne	S2.2
Sne+1	S2.3
Sne+2 (crenulation cleavage)	S3
Folds:	
Fne-2 (folded greenstone)	F1
Fne-1	F2.1
Fne	F2.2
Fne+1	F2.3
Fne+2 (NE trending folds)	F3

Lineations:

Lne-1	L2.1
Lne	L2.2
Lne+1	L2.3
Lne+2	L3

4. PRE-METAMORPHIC FAULTS

Pre Dn-1	T1 (S1) (or Pre-M1 thrusts)

APPENDIX IV

EXPLANATION TO THE LITHOLOGIC SYMBOLS

SILICICLASTIC PLATFORM

Ct - Ticonderoga dolomite
Cp - Potsdam sandstone

WEST CENTRAL VERMONT
(western Lincoln massif)

Ccs - Clarendon Springs dolomite
Cda - Danby Formation; quartzite and dolomite
Cw - Winooski dolomite
Cm - Monkton quartzite (with minor dolomite near top)
Cd - Dunham dolomite
Cc - Cheshire quartzite (lower argillaceous and upper massive members)

Transition to thermal-dominated subsidence

CZfp - Fairfield Pond Formation; green and gray phyllite

Rift-dominated subsidence

CZfd - Forestdale Marble; massive dolomite and dolarenite
CZp - Pinnacle Formation; metawacke and minor laminated, albitic schist
CZpbc - Basal member (also; Tyson Formation, CZt); polymict conglomerate with minor quartzite, dolomite and metawacke

unconformity

Ymh - Mount Holly complex; gneiss, schist, amphibolite, and quartzite of the Lincoln massif

(eastern Lincoln massif)

CZhg - Hoosac Formation; green aluminous chloritoid, garnet schist

Rift-dominated subsidence

CZhga - Hoosac Formation; albite, magnetite, garnet schist

CZhbc - Basal member; polymict conglomerate, schist, quartzite, and metawacke

unconformity

Ymh - Mount Holly complex

UNDERHILL SLICE

Rift-dominated subsidence

CZua - Underhill Formation (western sequence); albitic metawacke and biotite schist
CZuj - Jerusalem slice Member; quartz laminated schist and metawacke

TACONIC ALLOCHTHONS

Group 3 Slices

CZg - Greylock Formation; green chloritoid phyllite

Rift-dominated subsidence

CZgb - black to dark gray albitic chloritoid and albitic schist, metawacke, and feldspathic quartzite
CZga - green and gray albite, chlorite schist

Group 1 slices

Chh - Hatch Hill Formation; dark gray slates with minor interbeds of carbonate sands, chert horizons, and rare massive dolomite (100 ft. max.)
Cheb - Eagle Bridge quartzite; dark gray to blue/black quartzite
Cwc - West Castleton Formation; dark gray slates with turbiditic layers of dolarenite, limestone conglomerate, massive white quartzite, and chert

Transition to thermal-dominated subsidence

Cnm - Metawee Formation; green, purple chloritoid phyllite interlayered with Cnbp
Cnbp - Browns Pond Formation; dark gray slates (identical to Cwc) interlayered with carbonate conglomerates and grits,carbonate sands,and massive white quartzite (Cnmp and Cnzh)
Cnmp - Mudd Pond quartzite; massive white quartzite
Cnzh - Zion Hill quartzite; identical to Cnmp
CZnm -Truthville slate (Metawee); green, purple phyllite with minor massive white quartzite (Cnmp/Cnzh)
CZnb - Bomoseen graywacke; fine grained metawacke with detrital mica and quartz, with interlayered minor quartzite (Cnmp/Cnzh) in the upper section

Rift-dominated subsidence

CZnm - Nassau Formation; green, purple chloritoid phyllite (Rensselaer slice)
CZnr - Rensselaer greywacke member; metawacke with metabasalt horizons
CZnp - Green, gray, and purple chloritoid phyllite of the Chatham slice (similar to CZnm) with metabasalt horizons
CZev - Green, gray, and purple chloritoid phyllite of the Everett slice

CENTRAL VERMONT

Green Mountain anticlinorium

Cb - Battell Formation; carbonaceous albitic schist with pervasive dolarenite layers, dark quartzite, and rare massive dolomite horizons (100 ft. max.)

Transition to thermal-dominated subsidence

CZa - Mount Abraham Formation; chloritoid, paragonite, quartzose schist with rare metawacke and mafic schist; locally interlayered with CZhn and interpretively interlayered with Cg

Rift-dominated subsidence

CZm- Monastery Formation (including eastern Underhill, CZu and Fayston Fms.); white albitic schist with interlayered mafic (in CZhn) schist/gneiss, metawacke, and thin white
quartzite; appears to be gradational with CZpha

White and Mad River valleys

Cg - Granville Formation; carbonaceous albitic schist with dark and white quartzite (identical to Cotb); gradational with Cobp/Cobq

Transition to thermal-dominated subsidence

Cza - Mount Abraham Formation (in depositional contact with CZph and interpretively with Cg)

Rift-dominated subsidence

CZph -Pinney Hollow Formation; silver-green schist with associated mafic schist/gneiss, metawacke, and thin white quartzite; gradational with CZs

Northfield Mountains

Cobp - Ottauquechee Formation; graphitic black phyllite with minor beds of sandy quartz, feldspar wacke, tectonically incorporated greenstone (ocean crust ?), serpentine, and talc; gradational with Cg and Cotb

Cobq- Dark quartzite member; fine grained graphitic quartzite with detrital quartz and feldspar

Coqs - Sandy quartz schist member; intercalated black phyllite and sandy quartz, feldspar wacke with carbonate matrix and minor detrital blue quartz; interlayered with Cobp

Cotb - Thatcher Brook Formation;"blocks" of white albitic schist, greenstone, serpentine, talc schist, and dolomite in a heterogeneous tectonic matrix composed of graphitic, albitic schist (Cg ?) and black phyllite (Cobp) with minor albitic schist (CZph/ CZs ?) and dark quartzite (Cobq). This unit is tentatively assigned a Middle Ordovician age.

Transition to thermal-dominated subsidence

Cg - Granville Formation; interlayered with Cobp/Cobq

CZa - Mount Abraham Formation; interpretively in depositional contact with Cg and underlying CZs

Rift-dominated subsidence

CZs - Stowe Formation; Quart, chlorite phyllite with associated mafic schist/gneiss; identical to, and gradational with, CZph

72°54'
72°45'
44°00'30"
43°56'
WESTERN DOMAIN
CENTRAL DOMAIN
EASTERN DOMAIN
BEDROCK GEOLOGIC MAP
OF THE
GRANVILLE - HANCOCK AREA
CENTRAL VERMONT
by
Thomas R. Armstrong
PLATE 1
WESTERN SEQUENCE
BATTELL Fm
Cb Gray graphitic schist with locally pervasive marble segregations
Cbd Buff to light gray dolomite and marble layers within gray graphitic schist of Cb
Cbw WHITE RIVER MBR: Intercalated graphitic schist and non graphitic albitic schist and granofels, with discontinuous lenses of black marble.
MONASTERY Fm
CZm White to steel gray albitic schist, quartz-feldspar granofels, chloritoid phyllite, quartzite, and dolostone
HOOSAC Fm
CZh Light gray to white quartz-feldspar-biotite granofels and gray quartzose albitic schist
CZhbc polymict conglomerate, quartzose albitic schist, and chloritoid schist
CENTRAL SEQUENCE
GRANVILLE Fm
Cgtm TECTONIC MBR: Heterogeneous graphitic schist, lenses of dark and white quartzite, and lenses of greenstone
Cgl LINCOLN GAP MBR: Very graphitic chlorite-rich schist
Cg Graphitic albitic schist with locally interlayered dark and white quartzite
MT. ABRAHAM Fm
PINNEY HOLLOW Fm
WEST CZa
CZph EAST
Silver-green chlorite schist, silvery albitic schist, and white quartzite
to well foliated greenstone member
White to light gray chloritoid schist with abundant garnet in the western part of the area.
CZag Greenstone member
EASTERN SEQUENCE
OTTAUQUECHEE Fm
Cotb THATCHER BROOK MBR: Heterogeneous, admixed, graphitic albitic schist, very graphitic phyllite, non-graphitic silver-green albitic schist, dark quartzite, greenstone, talc-serpentine, and quartzose schist
Coqs Light gray ankerite-bearing quartz-feldspar-muscovite granofels and schist
Cobq Thin to thick layers of dark blue, gray, and black quartzite with distinctive white quartz veins
Cobp Very graphitic pyritiferous quartz-muscovite-albite phyllite
STOWE Fm
CZs Green to silver-green quartz-chlorite-sericite phyllite and schist, commonly containing magnetite and Fe stained vein quartz
CZsg Greenstone member
SYMBOLS
FOLIATION
Sn - 1 relict schistosity (axial planar to Sn - 1 schistosity)
Sn dominant schistosity (axial planar to Sn)
Sn +1 secondary schistosity (axial planar to Sn + 1)
Sn + 2 slip / pressure solution cleavage (axial planar to Sn + 2)
AXIAL SURFACES
Sn - 1
Sn
Sn + 1
Sn + 2
FOLD HINGES
Fn - 1
Fn
Fn + 1
Fn + 2
LETTER SYMBOLS
(letter on high T side)
S Serpentinite / talc schist body
A - A' cross section register points
LINEATIONS
Ln mineral lineation (including chlorite, sericite streaks)
Ln quartz rodding (predominantly quartz vein Fn hinges)
Ln intersection upon bedding
Ln + 1 mineral / intersection lineations
CHILD'S MTN. THRUST
UNDERHILL THRUST ZONE
OTTAUQUECHEE THRUST
A
A'
B
B'
MAGNETIC NORTH
TRUE NORTH
1:12000
MILES
FEET
KILOMETERS
INDEX MAP OF VERMONT
LOCATION OF PLATE

PLATE 2A
CROSS SECTION A - A'
V.E. = 1X
WEST
A
EAST
A'
Granville / Lincoln Gap Thrust Slice
MT ABRAHAM / PINNEY HOLLOW THRUST SLICE
Ottauquechee Thrust Slice
Underhill Thrust Zone
Granville Thrust Zone
Pinney Hollow Thrust Zone
Ottauquechee Thrust Zone
Battell Mtn
Texas Gap
Childs Mtn
Gulf Brook
Granville
Northfield Mtns.
Єg
Єotb
Єgl
ЄZa
Єzph
ЄZphg
Єo
ЄZs
ЄZm
ЄZh
Cb
Cbd
Cbw
CZa
CZm
Ymh
0
2000
FEET
20000
18000
16000
14000
12000
10000
8000
6000
4000
2000
SEA LEVEL
-2000
-4000
-6000
-10000
VERTICAL SCALE IN FEET

PLATE 2B


0 2500
FEET
0 600
METERS
V.E.= 1x
16000
14000
12000
10000
8000
6000
4000
2000
SEA LEVEL
-2000
-4000
-6000
-8000
-10000
-12000
B
B'
White River
Northfield Mtns.
Ꞓobp
ꞒZs
Ꞓo
ꞒZph
Ꞓobq
Ꞓotb
Ꞓoqs
ꞒZphg
ꞒZsg
ꞒZs
S

LITHOTECTONIC MAP OF SOUTHERN AND CENTRAL VERMONT

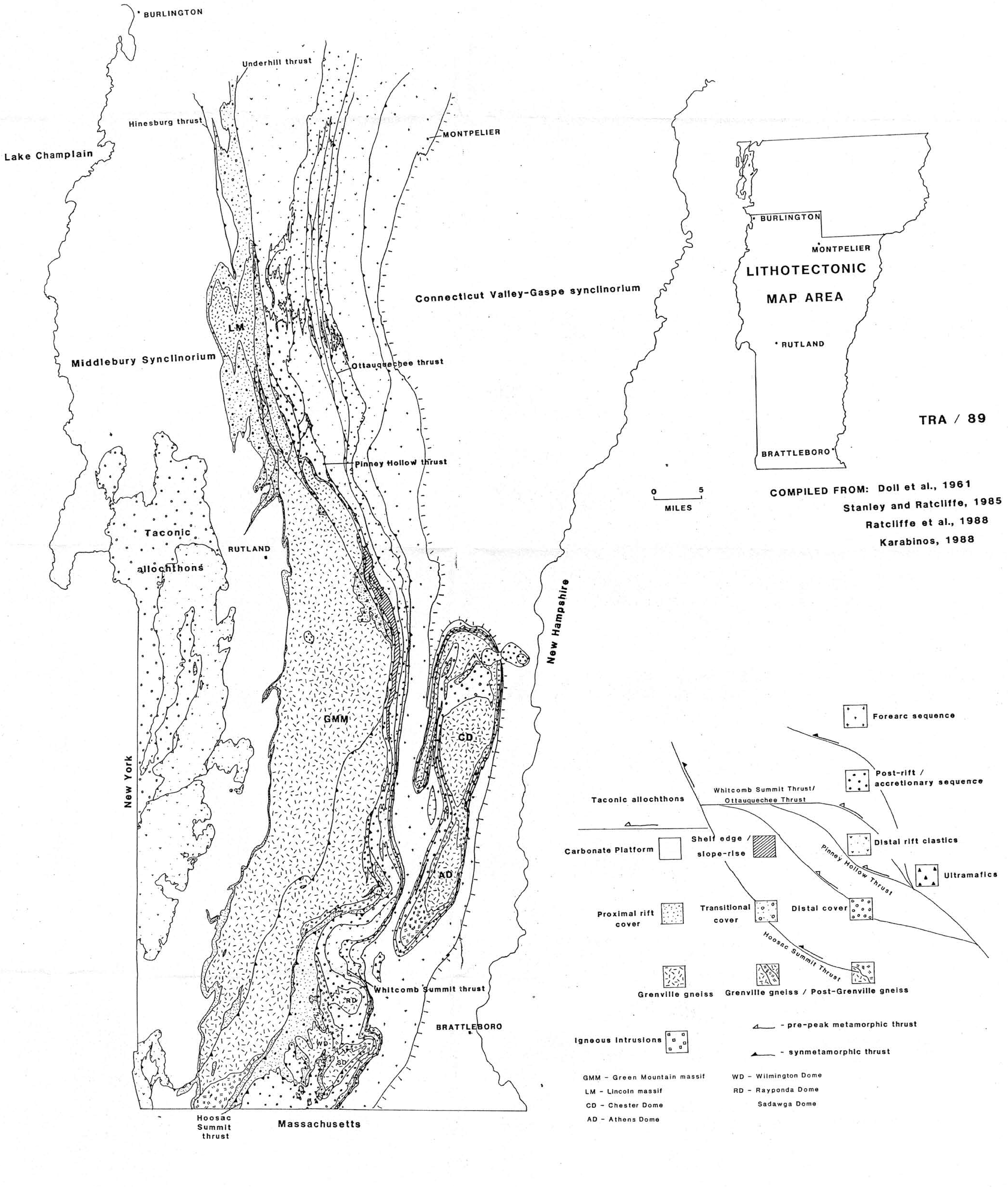


PLATE 4

INTERPRETIVE LATE PRECAMBRIAN - LOWER PALEOZOIC IAPTEAN PASSIVE MARGIN

SEE CHAPTER 2, PASSIVE MARGIN DEVELOPMENT AND CONCLUSION SECTIONS FOR DISCUSSION, AND APPENDIX 3 FOR FORMATION DESIGNATOR DESCRIPTIONS.

ARMSTRONG and STANLEY, 1988

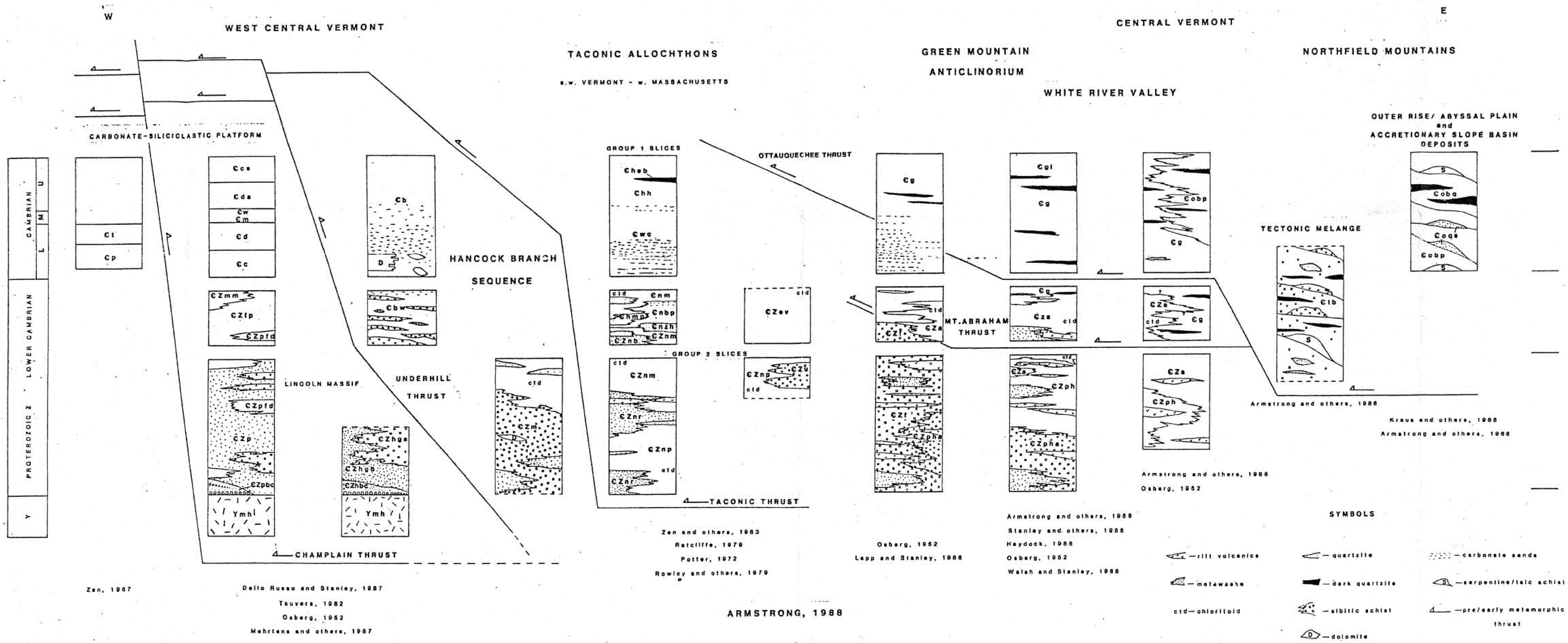
(Crustal geometry adapted from Lister and others, 198

PLATE 5

SEE CHAPTER 5, CONCLUSIONS FOR DISCUSSION OF THRUST FAULT CHRONOLOGY AND LITHOTECTONIC EMPLACEMENT SEQUENCE. SEE APPENDIX 3 FOR FORMATION DESIGNATOR EXPLANATIONS.

LITHOTECTONIC CORRELATION OF THE IAPTEAN PASSIVE MARGIN SEQUENCES

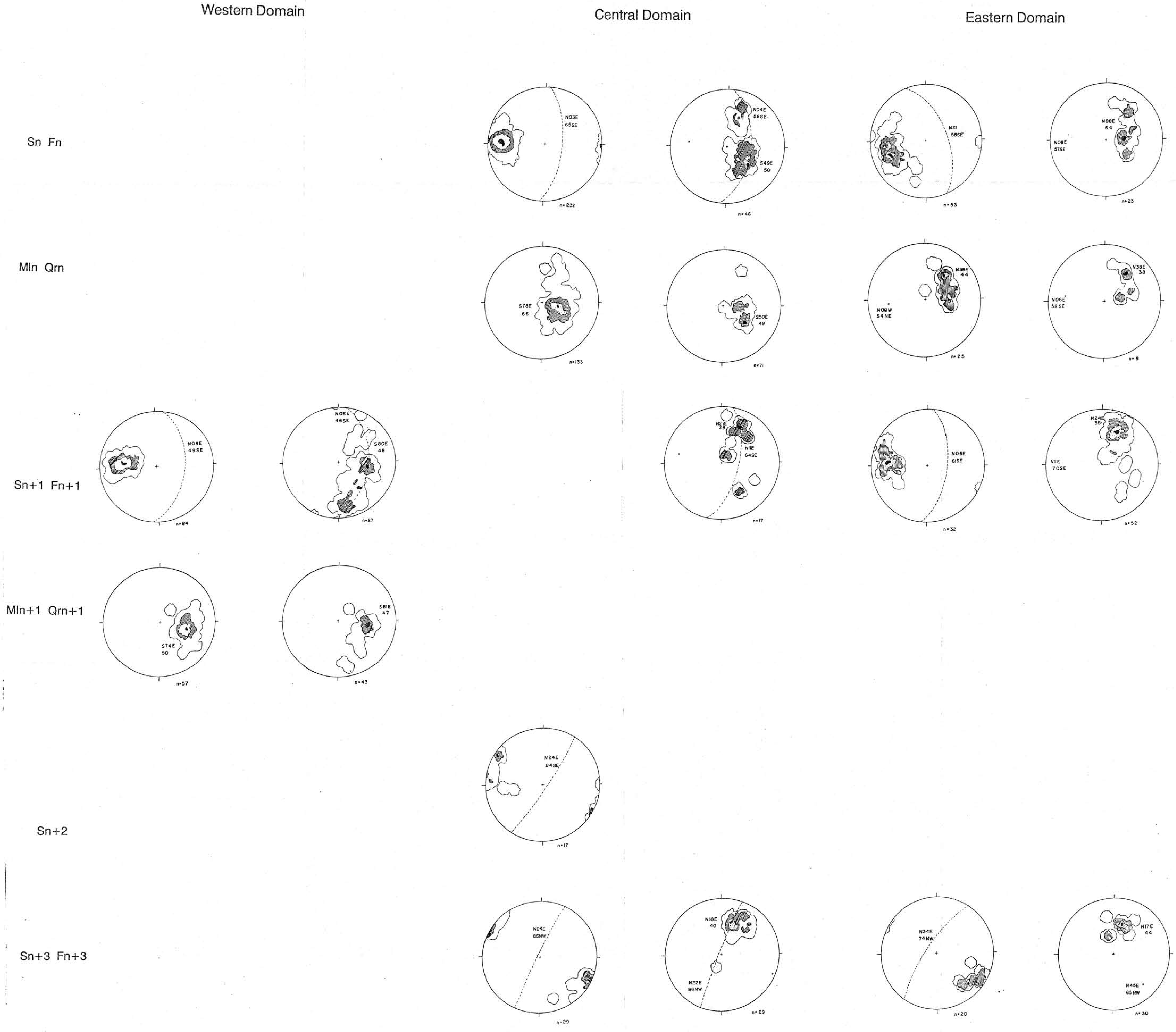
WESTERN NEW ENGLAND



GRANVILLE-HANCOCK
EQUAL AREA NET DATA

PLATE 6

SEE CHAPTER 4, STRUCTURAL GEOLOGY, FOR DISCUSSION OF ALL FABRIC DATA.



All nets contoured in 1, 10, 17, and 25% data per 1% net area