1-1-1987

A Transect Through the Pre-Silurian Rocks of Central Vermont; The Lincoln Massic and Its Immediate Cover; The Pre-Silurian Hinterland Along the Valleys of the White and Mad Rivers, Central Vermont; Metamorphism of Pre-Silurian Rocks, Central Vermont; Regional Geochemical Variations in Greenstones from the Central Vermont Appalachians

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A TRANSECT THROUGH THE PRE-SILURIAN ROCKS OF CENTRAL VERMONT

by

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INTRODUCTION

The central Vermont region is ideally suited to reassess the stratigraphy and structure of the pre-Silurian hinterland because many of the units mapped by previous workers and shown on the Geological Map of Vermont are well exposed in a relatively small region of approximately 200 sq. mi. Furthermore, the transect extends from the carbonate platform through the Middle Proterozoic rocks of the Lincoln massif to the basal Silurian-Devonian unconformity of eastern Vermont (fig. 1). The earlier work by Osberg (1952) and Cady, Albee, and Murphy (1962) is very important because it highlighted many important problems and delineated many of the important formations of eastern Vermont. Subsequent work by Albee (1965, 1968), Laird and Albee (1981), and Laird and others (1984) has placed important constraints on the metamorphic petrology and history of the eastern cover sequence. Isotopic analysis by Laird and Albee (1981) and Laird and others (1984) and discussions by Sutter and others (1985) indicate that much of the metamorphism is associated with the Taconian orogeny although the degree of subsequent cooling and/or Acadian metamorphism is still very much in debate. Important geochemical work by Coish (1985, 1986) and his students at Middlebury College indicates that the mafic schists become more oceanic as they are traced eastward through much of the eastern pre-Silurian cover sequence.

The ultimate goal of our current work is a better understanding of the tectonic evolution of the Taconian orogeny for western New England. Four important questions must be addressed in achieving this goal. They are: 1) What is the dominant mechanism of shortening in the hinterland, folding or faulting?, 2) What is the age of the various structures?, 3) Does the hinterland shorten continuously during orogeny or is it only deformed early in the mountain building process and subsequently acts as a rigid plunger transmitting stress to the imbricating foreland?, and 4) How do the rocks in the pre-Silurian hinterland correlate with those in the Taconic allochthons. Answers to these questions are not only important for the Appalachians in western New England but are important in understanding the processes of compressional tectonics in other mountains belts of the world where small-scale mapping is only available at best.

Recent mapping by Stanley and his graduate students (Tauvers, 1982; DiPietro, 1983; Strehle and Stanley, 1986; O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986; DelloRusso and Stanley, 1986;

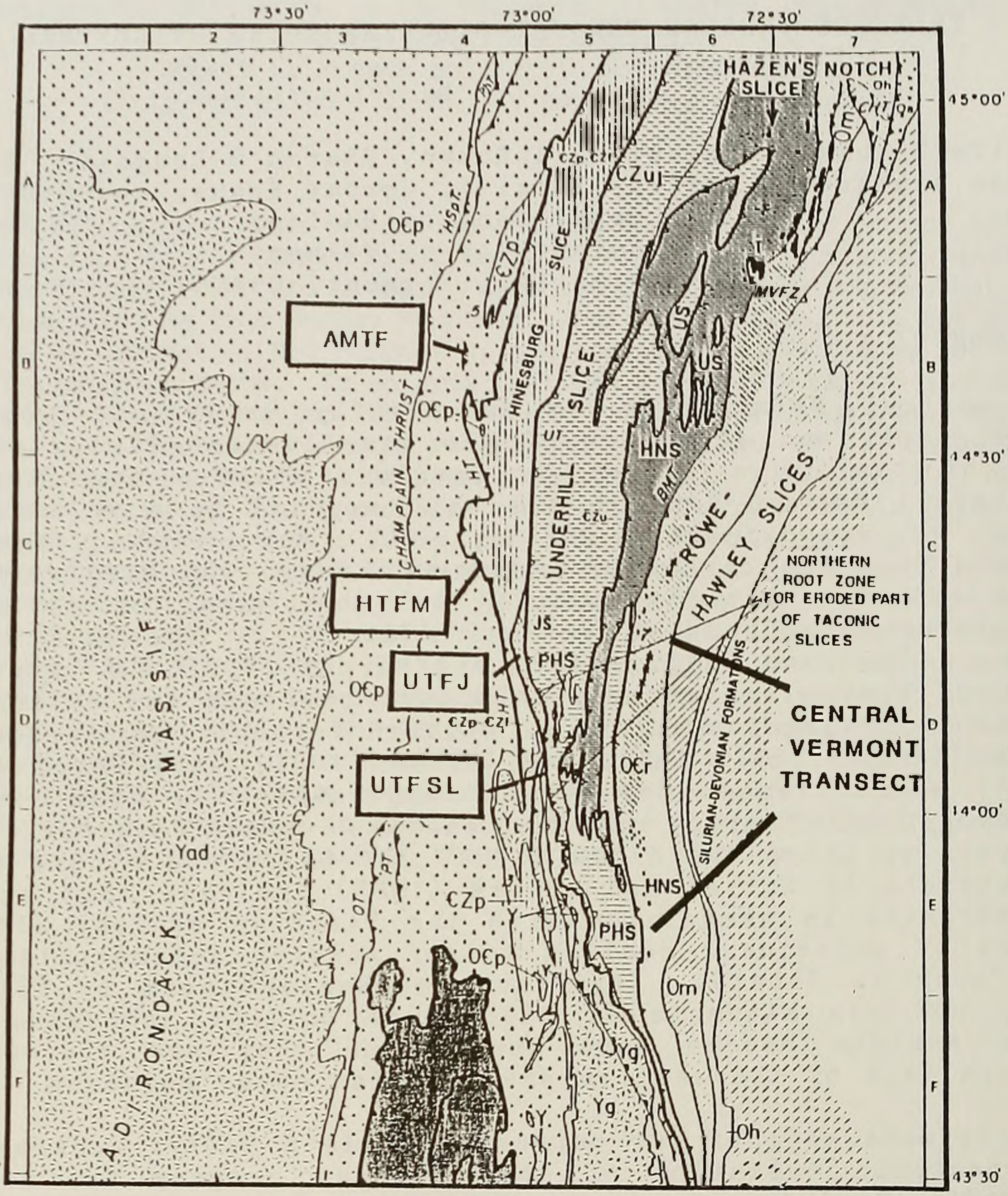


Figure 1 Interpretative Tectonic Map of Vermont and eastern New York showing the general location of the Arrowhead Mountain thrust fault (AMTF), the Hinesburg thrust fault at Mechanicsville (HTFM), and the Underhill thrust fault at Jerusalem (UTFJ), and South Lincoln (UTFSL). The geological map is taken from Stanley and Ratcliffe (1985, Pl. 1, figure 2a). Symbol T in A6 is a glaucophane locality at Tilliston Peak. Short line with x's (Worcester Mountains) and line with rhombs (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones of Albee (1968). Widely-spaced diagonal lines in northcentral Vermont outline the region that contains medium-high pressure amphibolites described by Laird and Albee (1981b). Irregular black marks are ultramafic bodies. Open teeth of thrust fault symbols mark speculative thrust zones. The following symbols are generally listed from west to east. Yad, Middle Proterozoic of the Adirondack massif; Yg, Middle Proterozoic of the Green Mountain massif; YL, Middle Proterozoic of the Lincoln massif; Y, Middle Proterozoic between the Green Mountain massif and the Taconic slices, Vermont; OЄp, Cambrian and Ordovician rocks of the carbonate-siliciclastic platform; rift-clastic sequence of the Pinnacle (CZp) and Fairfield Pond Formations (CZf) and their equivalents on the east side of the Lincoln and Green Mountain massifs, PhT, Philipsburg thrust; HSpT, Highgate Springs thrust; PT, Pinnacle thrust; OT, Orwell thrust, UT, Underhill thrust ; HT, Hinesburg thrust; U, ultramafic rocks; CZu, Underhill Formation; CZuj, Jay Peak Member of Underhill Formation; OЄr, Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation and its equivalents in Vermont; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens Notch slice; MVFZ, Missisquoi Valley fault zone; PHS, Pinney Hollow slice; BMT, Belvidere Mountain thrust; CHT, Coburn Hill thrust; Oa, Ascot-Weedon sequence in grid location 7A.

Prewitt, Haydock, Armstrong, Kraus, Walsh, Cua, and Kimball, all in progress) has shown the following important relations: 1) The Middle Proterozoic rocks of the Lincoln massif are progressively deformed to the east by east-over-west folds and synmetamorphic faults of post-Grenvillian age (Prahl, 1985; DelloRusso, 1986). 2. The eastern boundary of the massif is deformed by several synmetamorphic thrust faults (South Lincoln, Jerusalem, and Underhill thrust faults) which contain slivers of Middle Proterozoic rocks (Tauvers, 1982; DelloRusso, 1986; Strehle and Stanley, 1986). 3. The Hoosac (Pinnacle), Underhill, Mt. Abraham, Hazens-Notch, Pinney Hollow, Ottauquechee, and Stowe Formations in the eastern cover sequence are deformed by major pre- or early -metamorphic thrust faults that have been tightly folded and refolded into sheath-like folds and cut by synmetamorphic and late- to post-metamorphic thrust faults (O'Loughlin and Stanley, 1986; Lapp and Stanley, 1986; Armstrong, Kraus and Walsh, in progress). 4. Fabrics along synmetamorphic faults indicate a complex history of east-over-west movement followed in many places by later flattening which commonly obscure the earlier direction of fault movement. 5. The rocks within these formations have been divided into at least 7 thrust slices that record a complex history of deformation caused by later folding and renewed faulting. Thrust emplacement in the hinterland does not follow a simple east-to west progression toward the foreland. 6. This complex structural history requires that several of the earlier formations (Hazens Notch, Underhill) in the pre-Silurian hinterland be abandoned or redefined. Furthermore, the Ottauquechee Formation and its equivalents (Granville Formation, Battell Member of the Underhill Formation, and carbonaceous units in the Stowe, Pinney Hollow, Hazens Notch Formations) appear to be far more extensive across strike than previous shown by Doll and others (1961). The resulting story is one of an original, relatively simple stratigraphy that was subsequently involved in a very complex structural history. The result is the distribution of units as we see them today. 7. Correlation of mineral growth and structural sequence indicate that synmetamorphic faulting occurred during peak metamorphism and continued during retrogression to produce strongly elongated chlorite clusters in the Underhill, Hazens Notch, Mt, Abraham, Pinney Hollow and Stowe Formations (Lapp, 1986; O'Loughlin, 1986; Armstrong, in progress). 8. Older thrust faults, which are largely responsible for the distribution of thrust slices, occurred before and during peak metamorphism.

Analysis of mineral assemblages by Albee (1965, 1968) and Laird and others (1984) and amphiboles by Laird and Albee (1981, 1984) suggest that peak metamorphic conditions occurred at temperatures of 500^{0}C - 530^{0}C and confining pressures of 4 to 5 kbars (13 to 16 km). Recent information by Laird (this guidebook) on amphiboles from central Vermont indicates that all the mafic schist west of the Green Mountain crest record medium-facies series metamorphism whereas those to the east in the pre-Silurian rocks show medium-high pressure series metamorphsim. Based on a continuous and progressive sequence of structural deformation and metamorphic fabric a Taconian age (Middle Ordovician) is preferred for most of the pre-Silurian geology of central Vermont. The young 385 m.y. 40Ar/39Ar age reported by Laird and others (1984) for muscovite and biotite at Mt. Grant and the 382 m.y. 40Ar/39Ar age for amphibole just east of the Lincoln massif (Lanphere and others (1983) may suggest prolonged recrystallization and late uplift, perhaps over a deep ramp, for this part of Vermont.

THE LINCOLN MASSIF

The Lincoln massif represents the northern most exposure of Middle Proterozoic rocks in the Appalachian Mountains in the United States (fig. 1). Although small compared to other external massifs of the Appalachians (about 50 sq. miles), it is important because 1) it contains mafic dikes associated with Late Proterozoic rifting, 2) the relations between the Middle Proterozoic and its overlying Late Proterozoic cover are reasonably well displayed, and 3) it preserves a record of the progressive deformation of the massif during the Taconian orogeny. The massif is divided into an eastern and western part by a syncline of the lower part of the Pinnacle Formation.

All of the early work on the northern part of the massif was done by Cady, Albee, and Murphy (1962) for the U. S. G. S. Lincoln Mountain 15 minute quadrangle. The southern part of the massif was mapped by Osberg (1952). Our current understanding of the massif is based on detailed mapping at a scale of 1:24,000 or larger of the northern half by Tauvers (1982a), Prahl (1985), DelloRusso (1986)and DelloRusso and Stanley (1986).

Middle Proterozoic Basement Rocks

The Lincoln massif is largely made up of massive to well-foliated, granitic orthogneiss that becomes progressively more overprinted by lower Paleozoic deformation and metamorphism from west to east. The gneisses of the Western Lincoln massif consist of a massive, pink to light - gray weathered, medium grained biotite - quartz - feldspar paragneiss; a finer grained, strongly foliated, biotite gneiss with conspicuous magnetite octahedra and large biotite grains; and a relative thin, but continuous coarse-grained, augen gneiss containing plagioclase, quartz and microcline. The gneiss of the Eastern Lincoln massif is largely made up of fine-grained, massive, light-gray to white, quartz-plagioclase gneiss with microcline and perthite common locally. Biotite is noticeably absent or rare compared to its abundance in the gneisses of the Western Lincoln massif. Epidote, calcite and sericite are abundant particularly where the gneiss is severely altered by Paleozoic deformation. This unit is compositionally homogeneous throughout the Eastern Lincoln massif and it is not interbedded with any of the other rock types suggesting that it is of meta-igneous origin compared to the paragneiss of the Western Lincoln massif.

Metasedimentary sequences consisting of quartzite, aluminous schist, mafic schist, or augen gneiss are commonly found in isolated patches along the margin of the Lincoln massif. In the

Western Lincoln massif 1 to 5 m thick beds of massive, gray to white quartzite occur near the microcline augen gneiss that outlines a series of Paleozoic folds which extend across this part of the massif. Isolated patches of quartzite are also found near the basement-cover contact. These quartzites may correlate with similar quartzites mapped along the western part of the Eastern Lincoln massif where they are associated with light-colored tourmaline - bearing, white mica schist, and mafic schist. The quartzite commonly contains minor amounts of sericite, tourmaline, biotite, magnetite and rutile. The white mica schist is distinctive in that it contains large porphyroblasts of chloritioid and abundant needles of tourmaline.

Along the eastern boundary of the Eastern Lincoln massif, a coarse-grained, plagioclase augen gneiss is interlayered with chloritic and conglomeratic quartzite. This unit contains coarse plagioclase augen up to 4cm across. The biotite-rich matrix of the gneiss contains chlorite, epidote, and small clasts of blue quartz. This unit may represent a metamorphosed arkose in which the source area consisted of tonalite and mafic rocks. To the west within this unit, the augen gneiss becomes coarser grained and more biotite-rich near its contact with the granitic orthogneiss whereas, to the east, it becomes finer grained and more feldspathic as the basement-cover contact is approached.

Mafic schist occurs in a variety of settings that range from dikes, to fault zones, and conformable beds in metasedimentary sequences. Amphibolite layers with pyroxene, hornblende, biotite, epidote and plagioclase are found as isolated outcrops where it is interlayered with the medium grained, biotite gneiss of the Western Lincoln massif. Garnet amphibolite is associated with the bedded quartzite and tourmaline sericite schist along the western part of the Eastern Lincoln massif and may well represent mafic flows deposited on that sequence. Tauvers (1982a) mapped biotite-rich amphibolite as interlayered belts within the light-colored, granitic gneiss in the very northern part of the Eastern Lincoln massif. These belts may well represent highly-deformed Grenvillian dikes.

The remaining mafics schists are thin, highly foliated, biotite-rich schists that occurs as linear bodies with sharp contacts within the granitic gneiss of the Eastern Lincoln massif. DelloRusso (1986) has shown that these bodies are spatially related to Paleozoic fault zones, although they may have had such different origins as intrusive dikes or metasomatically altered amphibolites within fault zones.

One of the most important mafic rocks in the Eastern Lincoln massif are found as linear, metamorphosed dikes that truncate the Grenvillian foliation in the orthogneiss and in turn are truncated by the basal contact of the Late Proterzoic Pinnacle Formation. These rocks are composed of quartz, chlorite, epidote, biotite, and plagioclase. Primary amphibole is absent. Large plagioclase phenocrysts are characteristic and locally abundant although the grains are saussuritized. Dike contacts with the granitic gneiss are sharp and xenoliths of gneiss are common in the dikes. Many of the dike margins are altered to biotite-rich schist that contain larger grains of randomly-oriented biotite. The map and contact relations of these dikes indicate that they formed after the Grenvillian orogeny, but before the deposition of the basal units of the Pinnacle Formation. They probably represent mafic dikes that were associated with the rifting of the North American continental crust during the Late Proterozoic. Current geochemical work by Raymond Coish and his students at Middlebury College suggest that they are chemically similar to the Tibbit Hill Volcanic Member of the Pinnacle Formation, a major rift-clast facies of western New England.

Late Proterozoic Cover Rocks (fig. 2)

The Late Proterozoic cover-sequence north and west of the Lincoln massif consists largely of magnetite-bearing, quartz-feldspar metawacke with matrix-supported conglomerate interbeds. Discontinuous beds of pink to salmon-colored dolostone and associated chloritic metawacke and schist are locally important. Discontinuous quartz cobble conglomerates define the contact between the Pinnacle Formation and the underlying Middle Proterozoic basement of the Lincoln massif. The conglomerates range in thickness from 1 to 20 m. Some of the thicker conglomerates are rich in large cobbles and boulders of plagioclase gneiss which suggests rapid deposition, in contrast to the thinner quartz-rich conglomerates which indicate more prolonged erosion and winnowing. These relations indicate a varied topographic relief to the basement. Based on this evidence Tauvers (1982a) suggested that these rocks were deposited in rift basins which rapidly filled and were replaced by a relatively shallow, broad, featureless basin represented by the dark shales and thin siltstones of the Fairfield Pond Phyllite. This unit in turn grades upward into siliciclastic and carbonate rocks of the stable platform.

In contrast to the depositional contact observed along the western and northern margins of the Lincoln massif, the eastern basement-contact appears to be locally tectonic, although it is poorly exposed (fig. 3). Directly east of the inferred contact, fault zones with slivers of Middle Proterozoic granitic gneiss have been mapped within the basal 500 m of the cover sequence. Conglomerate is rare along the basal part of the eastern cover sequence. One outcrop at South Lincoln, Vermont, exposes a 40 m section of highly-deformed, matrix-supported conglomerate made up of pebbles and cobbles of quartzite and gneiss (Tauvers, 1982; Strehle and Stanley, 1986). The overlying rocks of the Hoosac Formation are made up of quartz-feldspar metawacke interlayered with biotite schist and muscovite-biotite schist some of which contain garnet. Albitic schist, which is typical of the Hoosac to the south along the eastern side of the Green Mountain massif, is present but minor compared to the total section at Lincoln township. These relations suggest that the cover sequence

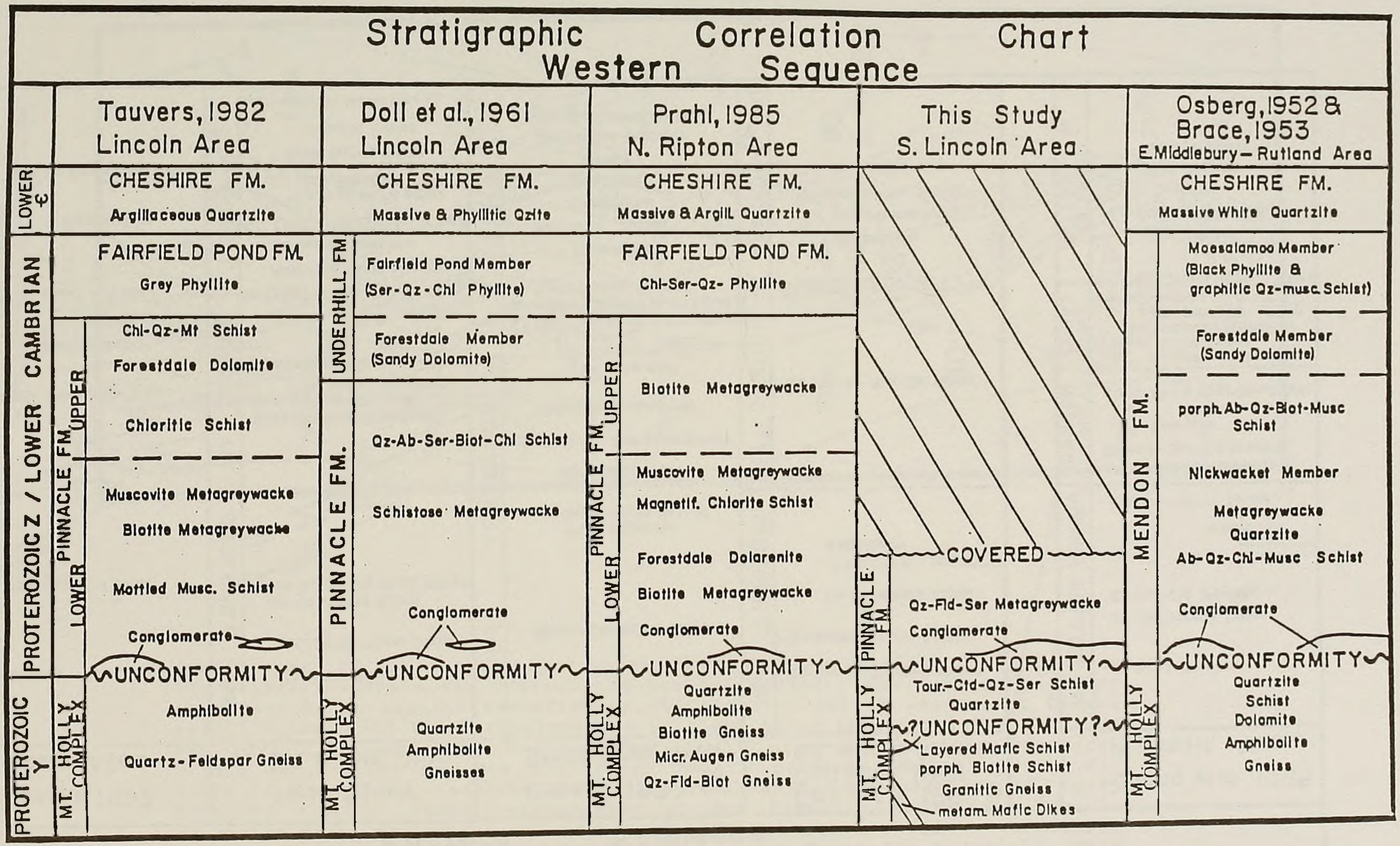
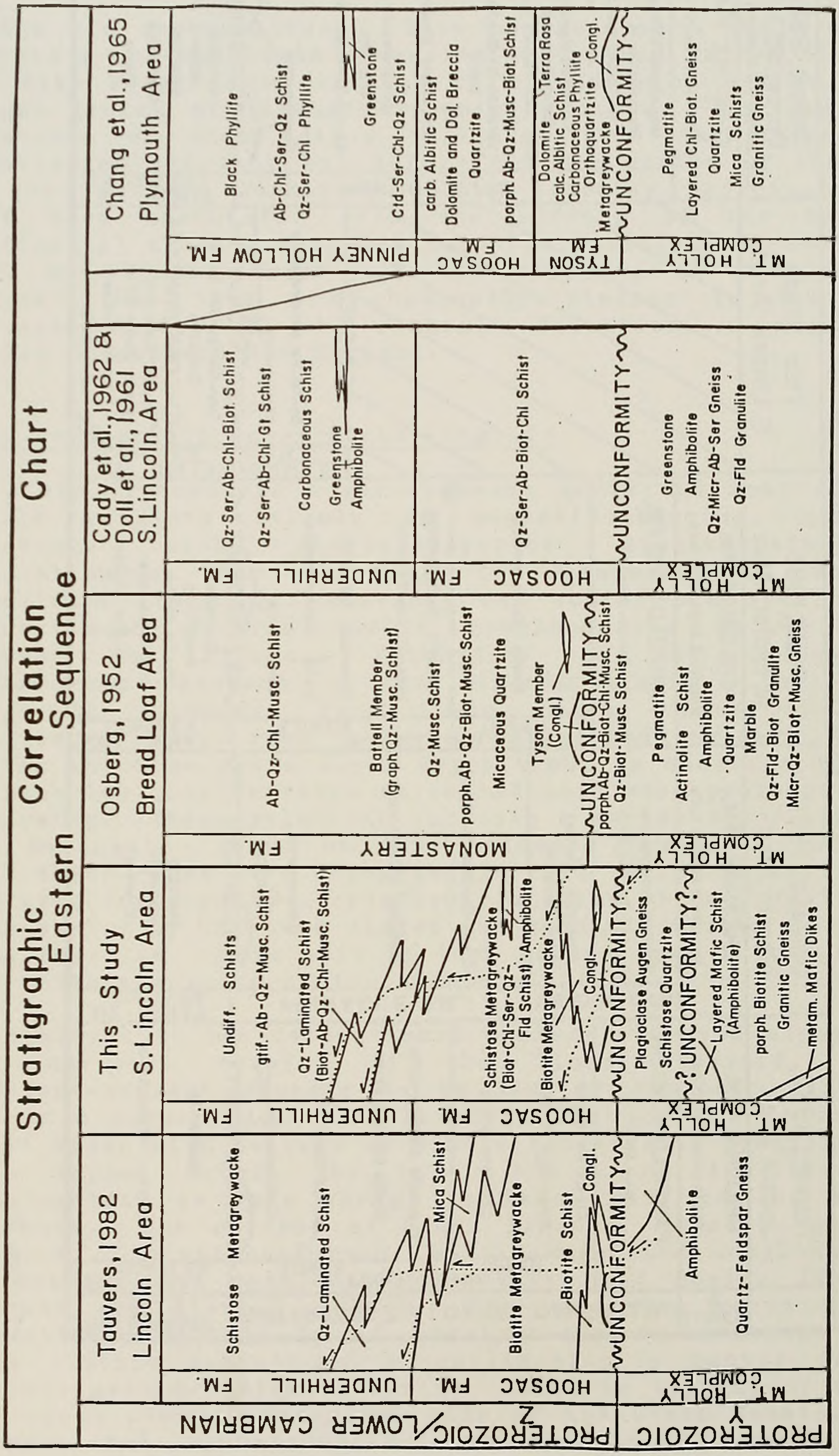

Stratigraphic Correlation Chart
Western Sequence
Tauvers, 1982
Lincoln Area
Doll et al., 1961
Lincoln Area
Prahl, 1985
N. Ripton Area
This Study
S. Lincoln Area
Osberg, 1952 & Brace, 1953
E.Middlebury — Rutland Area
LOWER Ꞓ
PROTEROZOIC Z / LOWER CAMBRIAN
PROTEROZOIC Y
CHESHIRE FM.
Argillaceous Quartzite
Massive & Phyllitic Qzite
Massive & Argill. Quartzite
Massive White Quartzite
FAIRFIELD POND FM.
Grey Phyllite
UNDERHILL FM
Fairfield Pond Member
(Ser-Qz-Chl Phyllite)
Forestdale Member
(Sandy Dolomite)
Chl-Ser-Qz- Phyllite
PINNACLE FM.
UPPER
LOWER
Chl-Qz-Mt Schist
Forestdale Dolomite
Chloritic Schist
Muscovite Metagreywacke
Biotite Metagreywacke
Mottled Musc. Schist
Conglomerate
Qz-Ab-Ser-Biot-Chl Schist
Schistose Metagreywacke
Biotite Metagreywacke
Magnetif. Chlorite Schist
Forestdale Dolarenite
COVERED
Qz-Fld-Ser Metagreywacke
MENDON FM.
Moosalamoo Member
(Black Phyllite & graphitic Qz-musc. Schist)
porph. Ab-Qz-Biot-Musc Schist
Nickwacket Member
Metagreywacke
Quartzite
Ab-Qz-Chl-Musc Schist
UNCONFORMITY
MT. HOLLY COMPLEX
Amphibolite
Quartz-Feldspar Gneiss
Quartzite
Amphibolite
Gneisses
Biotite Gneiss
Micr. Augen Gneiss
Qz-Fld-Biot Gneiss
Tour.-Ctd-Qz-Ser Schist
?UNCONFORMITY?
Layered Mafic Schist
porph. Biotite Schist
Granitic Gneiss
metam. Mafic Dikes
Schist
Dolomite
Gneiss

Figure 2



Stratigraphic Correlation Chart
Eastern Sequence
Tauvers, 1982
Lincoln Area
This Study
S. Lincoln Area
Osberg, 1952
Bread Loaf Area
Cady et al., 1962 &
Doll et al., 1961
S. Lincoln Area
Chang et al., 1965
Plymouth Area
PROTEROZOIC Z / LOWER CAMBRIAN
PROTEROZOIC Y
UNDERHILL FM.
HOOSAC FM.
MT. HOLLY COMPLEX
Schistose Metagreywacke
Qz-Laminated Schist
Mica Schist
Biotite Metagreywacke
Biotite Schist
Congl.
UNCONFORMITY
Amphibolite
Quartz-Feldspar Gneiss
Undiff. Schists
gtif.-Ab-Qz-Musc. Schist
Qz-Laminated Schist
(Biot-Ab-Qz-Chl-Musc. Schist)
Schistose Metagreywacke
(Biot-Chl-Ser-Qz-Fld Schist)
Amphibolite
Biotite Metagreywacke
Congl.
UNCONFORMITY
Plagioclase Augen Gneiss
Schistose Quartzite
? UNCONFORMITY ?
Layered Mafic Schist
(Amphibolite)
porph. Biotite Schist
Granitic Gneiss
metam. Mafic Dikes
MONASTERY FM.
Ab-Qz-Chl-Musc. Schist
Battell Member
(graph Qz-Musc. Schist)
Qz-Musc. Schist
porph. Ab-Qz-Biot-Musc. Schist
Micaceous Quartzite
Tyson Member
(Congl.)
UNCONFORMITY
porph Ab-Qz-Biot-Chl-Musc. Schist
Qz-Biot-Musc. Schist
UNCONFORMITY
Pegmatite
Actinolite Schist
Amphibolite
Quartzite
Marble
Qz-Fld-Biot Granulite
Micr-Qz-Biot-Musc. Gneiss
Qz-Ser-Ab-Chl-Biot. Schist
Qz-Ser-Ab-Chl-Gt Schist
Carbonaceous Schist
Greenstone
Amphibolite
Qz-Ser-Ab-Biot-Chl Schist
UNCONFORMITY
Greenstone
Amphibolite
Qz-Micr-Ab-Ser Gneiss
Qz-Fld Granulite
PINNEY HOLLOW FM.
HOOSAC FM
TYSON FM
Black Phyllite
Ab-Chl-Ser-Qz Schist
Qz-Ser-Chl Phyllite
Greenstone
Ctd-Ser-Chl-Qz Schist
carb. Albitic Schist
Dolomite and Dol. Breccia
Quartzite
porph. Ab-Qz-Musc-Biot. Schist
Dolomite
Terra Rosa
calc. Albitic Schist
Carbonaceous Phyllite
Orthoquartzite
Congl.
Metagreywacke
UNCONFORMITY
Pegmatite
Layered Chl-Biot. Gneiss
Quartzite
Mica Schists
Granitic Gneiss

Figure 2

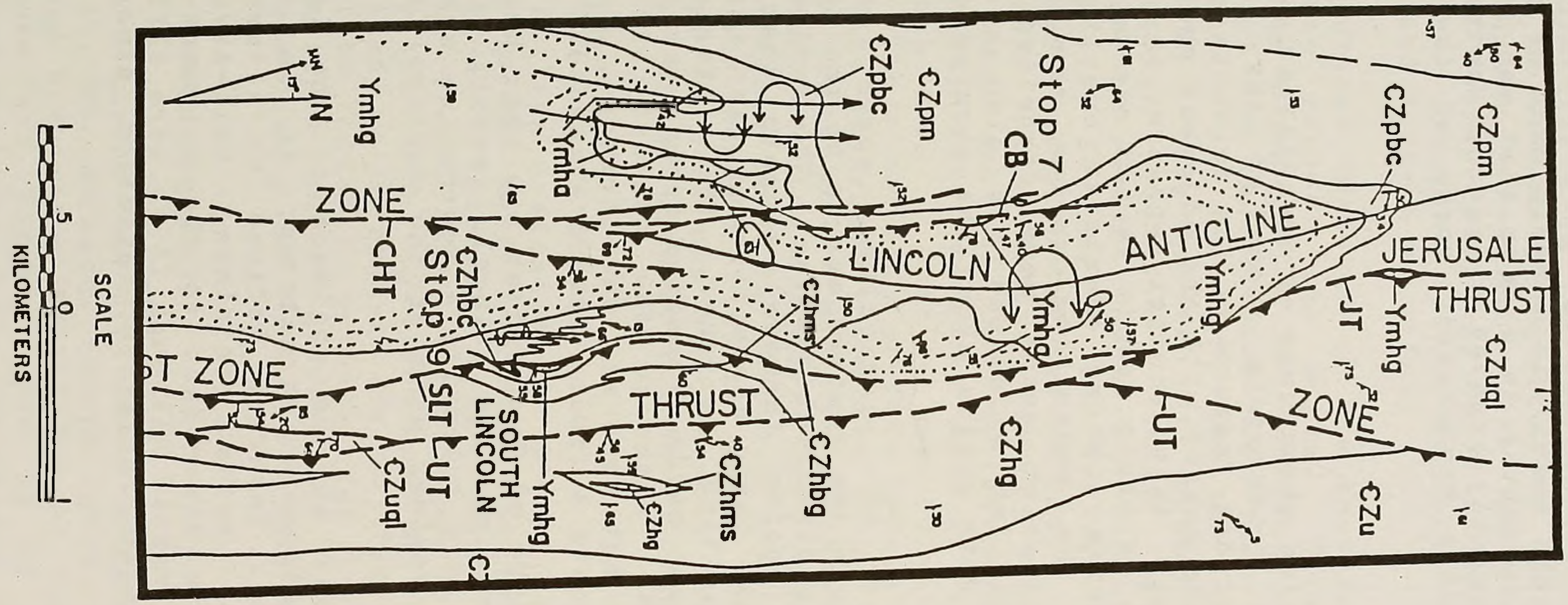


Figure 3

Geologic map of the northern end of the Eastern Lincoln massif illustrating the contrast between the western and eastern border. The western boundary is marked by an angular unconformity that is vertical or overturned to the east. A prominent basal conglomerate is found along this contact and faults are rare. In contrast, major thrust fault mark the eastern boundary. Lithic designators are the following: Ymhg - granitic gneiss of the Mt. Holly Complex. Ymha - amphibolite or amphibolite-rich parts of the Mt. Holly Complex, CZpbc - basal conglomerate in the Pinnaole Formation, CZpm - muscovite-rich metawacke of the Pinnacle Formation, CZu - Underhill Formation, CZuql - quartz laminated schist of the Underhill Formation, CZhg - schistose metawacke of the Hoosac Formation, CZhbg - biotite metawacke of the Hoosac Formation, CZhms - mafic schist in the Hoosac Formation. CB - Crash Bridge, CHT - Cobb Hill thrust zone, JT - Jerusalen thrust fault, UT - Underhill thrust fault, SLT - South Lincoln thrust fault.

immediately overlying the Lincoln massif is largely made up of the rift clastic rocks that were deposited in a site that was located between the basal sections presently exposed along the western and eastern boundaries of the Middle Proterozoic rocks of the Green Mountain massif.

Structural Relations

Grenvillian Structures Few major structures of Grenvillian age can be recognized in the Middle Proterozoic rocks of the Lincoln massif. Perhaps the most important feature is the contacts of the metasedimentary sequences in the massif with the adjacent gneisses. In the Eastern Lincoln massif this contact appears to be an unconformity since the contact is sharp and it is not cut by dikes or sills of the granitic gneiss. Where quartzite overlies the gneiss, however, clasts of the underlying gneiss are absent indicating that mature, quartz-rich sands were deposited on a gentle erosional surface. Similar relations are also observed where quartzite is in contact with gneiss along the margins of the Western Lincoln massif.

The metasedimentary sequence of plagioclase augen gneiss, quartzite, and conglomerate along the eastern margin of the Eastern Lincoln massif suggests quite a different setting. The sedimentary interpretation of the augen gneiss (arkose) suggsts that this sequence of rocks represents the rapid deposition of very immature material from an older tonalitic terrane. The interlayered quartzites could represent mature sands deposited during periods of relatively slow sedimentation. Although the contact with the underlying granitic gneiss is not exposed, the increase in grain size towards this contact suggests a fining-upward sequence.

The only major tectonic structure of Grenvillian age that has been mapped in the northern part of the Lincoln massif is an east-plunging fold delineated in a quartzite - tourmaline schist - mafic schist sequence on the western side of the Eastern Lincoln massif.

Although the Grenvillian fabric in the gneisses of the Lincoln massif is overprinted by Paleozoic structures, it is easily identified by a coarse mineral layering defined by alternating quartz-rich and feldspar-rich bands containing well-developed quartz rods. This layering is discordant to the basement-cover contact throughout the Lincoln massif although it is more difficult to distinquish from the younger Paleozoic foliation along the eastern margin of the massif.

Taconian (?) Structures

A Taconian(?) foliation marked by aligned micas in the gneiss becomes progressively more pervasive as it is traced eastward across the Lincoln massif. Chlorite and biotite are present within this foliation where the host rocks are richer in iron and magnesium than in the sericite - bearing granitic rocks which underlie much of the Eastern Lincoln massif. Available isotopic age information from the surrounding area suggests a Taconian age for this foliation (summarized by Sutter and others, 1985). This foliation was formed when the Lincoln massif was folded into a series of doubly-plunging, north-trending anticlines separated by a intervening syncline. This syncline divides the massif into the Western and Eastern massifs.

The structural features of the Western and Eastern massifs differ in degree rather than style. Broad folds outlined by the microcline augen gneiss and quartzite dominate the Western Lincoln massif. A few faults have been mapped, but they are of minor in importance. In contrast, faults dominate the Eastern Lincoln massif where they become more numerous and prevasive toward the eastern-cover sequence. These faults are of two kinds, east-directed reverse faults and normal faults associated with the flexural-slip folding of the western part of the Eastern Lincoln massif and west-directed thrust fault associated with the Cobb Hill thrust zone in the Eastern Lincoln massif and the imbricate slices of the basal part of the eastern-cover sequence. Well preserved asymmetrical fabrics along all of the fault zones record the consistent east - over - west direction of motion across the massif. The structural profile across the Lincoln massif therefore records the progressive failure of a representative part of North American crust as it was incorporated into the collisional zone during the Taconian orogeny. This detailed mapping has led to the interpretation that the Lincoln massif developed as a series of large-scale basement folds, which were progressively imbricated from east to west and subsequently transported westward along the Champlain thrust zone during the Taconian orogeny.

Metamorphism

Few definitive indicators of Grenvillian metamorphism are present in the Lincoln massif because the rocks are largely granitic in composition. Taconian metamorphism has severely altered the original Grenvillian assemblages. The presence of garnet amphibolites and pyroxene-bearing amphibolites suggest that Grenvillian metamorphism was at least at the conditions of the amphibolite facies or perhaps even higher. These assemblages were recrystallized under greenschist to epidote-amphibolite facies conditions during the Taconian orogeny.

EASTERN COVER SEQUENCE

The pre-Silurian eastern cover sequence consists of the Hoosac, Underhill, Mt. Abraham, Hazens Notch, Pinney Hollow, Ottauquechee, Stowe and Missisquoi Formations of Doll and others (1961). These formations are largely made up of gray to green, non rusty-weathering schists; rusty-weathering, brown to dark gray to black carbonaceous schist; aluminous, silvery chloritoid-bearing schist; metawacke; quartzite; granofels; and abundant mafic schist (greenstone). Albite porphyroblasts are common in many of the

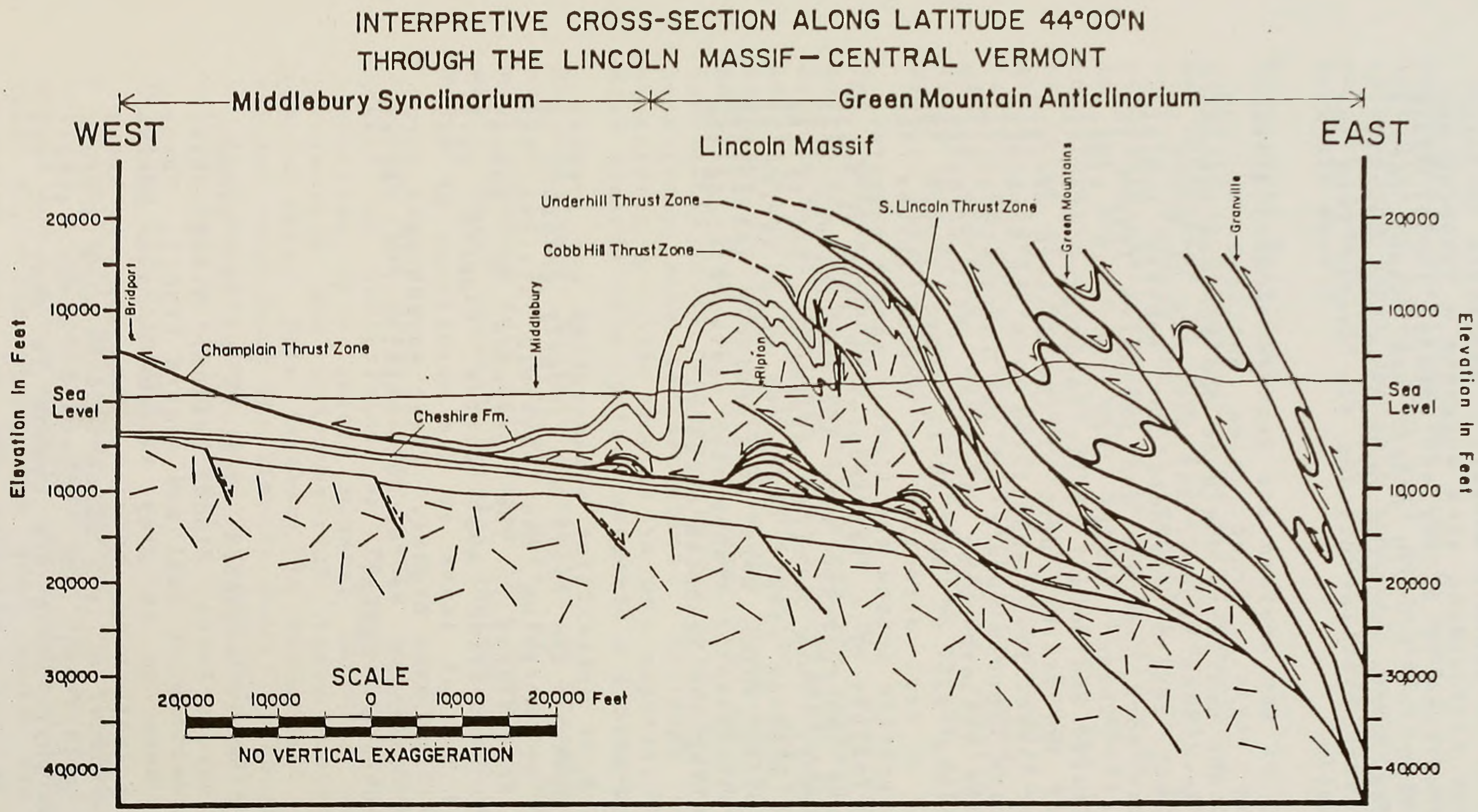


Figure 4

Interpretative east-west cross section along latitude $43^{0}00'$ N through the Lincoln massif, central Vermont. This cross section shows the Champlain thrust fault projected eastward beneath the Lincoln massif where it roots to the east of the exposed basement. Part of the autochthonous platform (CP) and rift-clastic (RC) sequence is shown to exist beneath the Champlain thrust fault. Anticlinal stacks or duplexes shown above the Champlain thrust fault are speculative, but structures of this type are thought to be an important element in the development of large scale basement folds. This cross section shows that Lincoln massif as a series of basement folds that were progressively imbricated from east to west and subsequently transported westward on the Champlain thrust fault during the latter part of the Taconian orogeny (Stanley and Ratcliffe, 1985). Renewed movement on the Champlain thrust fault or a deeper thrust fault during the Acadian orogeny can not be ruled out and may in fact be suggested by the 385 m.y. isotopic age from muscovite and biotite from the crest of the Green Mountains (Laird and others, 1984). Although the fault geometry in the eastern cover sequence is diagrammatic, it is based on the fact that premetamorphic (represented by the folded lines) and synmetamorphic (gently curved lines) thrust faults dominant this sequence.

units. Serpentinite and related rocks are largely found in the Ottauquechee Formation, other carbonaceous schists, and the Moretown Member of the Missisquoi Formation. Many of these rocks are found along faults (Stanley and others, 1984). Although the ultramafic rocks could have originally been olistoliths from a large slice of oceanic crust comparable to those mapped in Quebec and Newfoundland, subsequent deformation has totally obscured such an origin. Blue quartz is found in the metawackes of the Hoosac, Pinney Hollow, and Ottauquechee Formations.

The mafic schists are important because they are easy to map and provide valuable geochemical, petrologic and isotopic age data that helps to constrain the tectonic evolution for western New England (Coish and others, 1985, 1986; Laird and others, 1984, for example). Those in the Pinney Hollow and Stowe are particularly helpful in locating premetamorphic and synmetamorphic faults. Many of the mafic schists that we have mapped west of the Green Mountain crest in the Hazens Notch and Underhill Formations are slices or slivers on faults.

Stratigraphy

The rocks in the pre-Silurian eastern cover sequence have been viewed as a coherent although highly deformed sequence of units that become younger to the east (Doll and others, 1961). Thus the Hoosac Formation to the west was considered the oldest (Late Precambrian to Lower Cambrian) and the Missisquoi Formation was considered the youngest (Ordovican). Fossils are absent from the sequence and the supposed ages are based on correlation to similar fossiliferous rocks either in Quebec or the Giddings Brook slice of the Taconic allochthons. These correlations are highly speculative since none of the pre-Silurian units in eastern Vermont can be traced uninterrupted to the fossilerous localities.

Recent detailed mapping in a number of areas in Vermont have clearly demonstrated that the pre-Silurian section is marked by major faults that are largely premetamorphic (or early metamorphic) and synmetamorphic in age (Stanley and others, 1984; Tauvers, 1982; DiPietro, 1983; Loughlin and Stanley, 1986; Lapp and Stanley, 1986; DelloRusso and Stanley, 1986; Prewitt, Armstrong, Kraus, in progress). Similar findings have been reported by Zen and others (1983) in western Massachusetts and Karabinos (1984) for the basal part of the sequence along the eastern border of the Green Mountain massif in southern Vermont. Thus the sequence is not coherent and the age assignments, stratigraphic sequence and geological evolution are all in doubt and must be reinterpreted in light of these findings. Several authors have suggested that these rocks represent the defomed remains of an accretionary wedge that developed during the Taconian orogeny (Rowley and Kidd, 1981; Stanley and others, 1984; Stanley and Ratcliffe, 1985). If true, we might expect a very complex tectonic history superposed upon an orignal sedimentary sequence that formed along the continental margin of ancient North America.

During the past 4 years our work in central Vermont has shown that many of the mappable units occur in discontinuous lenses and belts which are repeated across strike. The contacts between most of these units truncate smaller units such as greenstone, metawacke, and schist that we map in previously recognized formations. The carbonaceous schists form an excellent example of this relationship. Carbonaceous rocks are found in the Underhill, Granville, Hazens Notch, Pinney Hollow, Stowe and are characteristic of the Ottauquechee Formation. Each of these units are rusty weathering, are marked by a certain amount of sulphidic stain, and contain varying amounts of graphite. Black albite is found in all the units even the Ottauquechee where albite is certainly subordinate compared to the other carbonaceous units. Black, gray, and white quartzite is present in all the units although it is most abundant in the Ottauquechee. Interestingly enough, rocks with all of these characteristics were earlier mapped in the Ottauquechee Formation (Osberg, 1952; Cady and others, 1962). Osberg (1952) recognized the similarity among some of these belts and on his map showed synclines of Ottauquechee throughout the eastern part of the Pinney Hollow. On the Bedrock Geological Map of Vermont (Doll and others, 1961), however, they are shown as carbonaceous schists in the Pinney Hollow. We believe this was a unfortunate choice because the map shows the Ottauquechee as a single linear belt separating the Pinney Hollow from the Stowe rather than several belts which can be recognized across strike. We would go even farther than Osberg did and suggest that the Ottauquechee or rocks that are lithic correlative to it occur as far west as the Underhill (Battell Member) and as far east as the carbonaceous schists in the Stowe. Furthermore, we have been able to demonstrate at the scale of 1:12,000, or larger, that the carbonaceous schists truncate units in the adjacent formations. This relationship, combined with fabric information, indicates that the contacts are either premetamorphic, early metamorphic, or synmetamorphic faults. We propose, therefore, that the Battell Member of the Underhill, the Granville Formation, the carbonaceous schists in the Hazens Notch, Pinney Hollow, and Stowe were originally part of one continuous deposit that extended eastward from the ancient continental crust of North America to oceanic crust. Changes in sedimentary facies did exist but its basic carbonaceous, sulphidic character persisted. Their present distribution is the result of a very complex history of deformation.

Another example is provided by the silvery green to gray green to light gray schists of the Hazens Notch, Pinney Hollow and Stowe. Albite porphyroblasts and deformed quartz veins are characteristic of each of these units. Magnetite is common. Greenstones that look identical in outcrop are common in the Pinney Hollow and Stowe. They are different than those in the Hazens Notch where the greenstones are darker in color and commonly albitic. Osberg (1952) recognized this similarity in the Pinney Hollow and Stowe, but showed them as separate formations because of their differences and the fact that the Ottauquechee consistently separated the two. We suggest, however, that the Pinney Hollow and Stowe were once part of the same deposit. We further suggest that the light gray, magnetite-bearing, albitic schist of the Hazens Notch represents a western facies of the Pinney Hollow. Similar rocks also are found in large tracks of the region mapped as Underhill (Thompson and Thompson, this guidebook; Walsh, personal communications). This proposal requires therefore that the Ottauquechee overlie the Pinney Hollow and Stowe.

Figure 5 is a stratigraphic correlation chart for the pre-Silurian of eastern Vermont. The position of the respective columns in central Vermont is based on recent mapping and the discussion in the foregoing paragraphs. The correlations with the sequence in the Taconic allochthons and the eastern limb of the Berkshire massif is based on lithic similarities and sequence among the respective areas. The correlation between the rocks of Group 3 slices of Stanley and Ratcliffe (1985) and the eastern Vermont sequence has been proposed by a number of earlier workers (Keith, 1934; Thompson, 1967). The stratigraphic arrangement presented here is therefore tentative and is undergoing modification during each field season.

A number of important aspects of this correlation chart require explanation because they differ from the more tradiational view of the pre-Silurian stratigraphy in Vermont. They are the following:

1. Major thrust zones separate each of the columns. Thrust faults are also present within many of the columns. For most of the units shown in this manner it is possible to locate their respective root zone. For example, the Mt. Abraham Schist roots along the western boundary of the Pinney Hollow and the carbonaceous schist of the Hazens Notch roots along the western boundary of the Ottauquechee Formation. The serpentinites and related rocks are presently found along faults. The only exception to this rule are the serpentinites found in the eastern part of the Moretown near Roxbury. These bodies are largely found within greenstone and direct evidence of throughgoing faults is absent. Evidence concerning their original mode of emplacement for all the ultramafic rocks is uncertain.

2. The albitic rocks in the noncarbonaceous rocks of the Hazens Notch, Pinney Hollow and Stowe are considered to be equivalent. Similar rocks also exist in the Underhill. These rocks are correlated to parts of the Greylock Schist (CZga) and Hoosac Formation (CZhab, CZh) of western Massachusetts and the Netop Formation of southwestern Vermont.

3. The Battell Member of the Underhill (not shown) and the carbonaceous schists in the Pinney Hollow are assigned to the carbonaceous schist of the Hazens Notch Formation (O'Loughlin and Stanley 1986; Lapp and Stanley, 1986). These units are a facies of and equivalent to the Ottauquechee Formation. It is identical and traceable into the Granville Formation of Osberg (1952). Although we assigned these carbonaceous schist to the Hazens Notch, we now believe that the term "Granville" would possibly be a better

EXPLANATION FOR FIGURE 5- The lithic symbols shown in each column of figure 5 are explained by geographic locality starting with the oldest unit.

West Central Vermont - Ymh- granitic gneiss of the Mt. Holly Complex, Y - mylonitic gneiss at South Lincoln, CZpbc - basal conglomerate in the Pinnacle Formation, CZpbg - biotite wacke, CZpfd - Forestdale Marble, CZpm - muscovite wacke, CZpcl - chlorite wacke and chlorite schist, CZhbc - basal conglomerate in the Hoosac Formation, CZhbg - biotite wacke, CZhg - schistose wacke, CZugl - quartzite laminated schist of the Underhill Formation, CZufg, well foliated wacke, CZu - mixed unit consisting of garnet, chlorite, muscovite schists, thin quartzite, CZuga - garnetiferous schist, CZfp - gray, finely laminated phyllite of the Fairfield Pond Formation, Cca - argillaceous quartzite of the Cheshire Formation, Ccm - massive, bedding quartzite of the Cheshire Formation, Cd - Dunham Dolomite, Cm - Monkton Quartzite, Cw - Winooski Dolomite, Cda - quartzite and dolostone of the Danby Formation, Ccs - Clarendon Springs Dolomite, Lower Ordovician Beekmantown Group consists of limestone, dolostone and minor quartzite and shale of the Shelburne Formation (Os), Cutting Dolomite (Oc), Bascom Formation (Ob), and Chipman Formation (Ocb), Omm - Middlebury Limestone, Oo - Orwell Limestone, Ogf - Glens Falls Formation, Owl - fossiliferous limestone lenses near the base of the Walloomsac Formation, Oh - black carbonaceous slate and phyllite of the Hortonville Formation, Ow - dark gray, graphitic schist and phyllite of the Walloomsac Formation, Oww - Whipstock Breccia Member of the Walloomsac Formation.

Taconic Allochthons - Group 1 and 2 - CZnr, Rensselaer Graywacke Member of the Nassau Formation, CZnv - metabasalt and basaltic tuff, CZnm - lustrous, yellowish-green, purple laminated chloritoid - chlorite phyllite of the Mettawee Member, CZnt - Truthville Slate, CZnb - Bomoseen Graywacke Member, CZnzh - Zion Hill Quartzite, CZnmp Mud Pond Quartzite, Cwc - dark gray to black, pyritiferous and calcareous slate with thin sandy laminae of the West Castleton Formation, Chh - bluish-gray weathering black sulphidic slate and chert, Opo - white weathering, well laminated gray slate and chert, On - Normanskill Formation, Onmm - black slate and thin bedded cherts of the Mount Merino Member, Onir - red and green slate of the Indian River Member, Onag - Austin Glen Graywacke Member

Group 3- CZga, Light green to gray, white albitic schist with some magnetite, chlorite granofels of the Greylock Schist, CZne - Netop Formation in the Dorset Mt. slice, CZgb - black of dark gray chloritoid or stilpnomelane albite, quartz knotted schist, lenses of feldspathic quartzite, conglomerate, and pink dolostone, CZg - light green, lustrous chloritoid phyllite and minor beds of white albitic schist, CZsc - St. Catherine Formation in the Dorset Mt. slice.

East Limb of the Berkshire massif - CZhsb (CZh) - gray to white albite spotted schist of the Hoosac Formation, CZhgsb (CZhga) - green albite magnetite schist, CZhg - green aluminous chloritoid schist. The symbols in parenthesis refer to designations of equivalent units on the Geologic Map of Massachusetts (Zen and others , 1983).

Green Mountain anticlinorium - Northfield Mountains (Many of the symbols are repeated from column to column. These symbols are explained in the first column that they appear in as the columns are read from left to right)

Lincoln Gap-Mt. Abraham-Hazens Notch slices - CZhn silvery white to dark green to black schist spotted with white albite, minor white to gray laminated quartzite of the Hazens Notch Formation. Mafic schist abundant in the eastern part,CZhnc - rusty weathering, black albitic schist with widespread graphite and minor thin black or gray quartzite. CZhnc has been called the Battell Member of the Underhill (Doll and others, 1961) and the Granville Formation (Osberg, 1952). CZa - Mt. Abraham Schist. Silvery colored paragonite - muscovite - chloritoid - chlorite (garnet) schist with a distinctive pearly sheen on the schistosity, CZamg - similar to the main body of Mt. Abraham Schist but with abundant magnetite and chloritized garnet, CZags - Muscovite - chlorite schist of the Mt. Abraham Schist with large porphyroblasts of garnet and minor chloritoid.

Lincoln Gap-Pinney Hollow slices - Additional symbols are: CZhnca - rusty weathering, dark gray to black albite schist with discontinuous patches of graphite. Traceable into the Granville Formation. CZph - silvery green muscovite - chlorite - quartz schist of the Pinney Hollow Formation, CZpha light gray muscovite - chlorite -quartz schist with albite porphyroblasts, CZphg - mafic schists of the Pinney Hollow, CZphw - gray wacke with minor blue quartz, Co - black, pyritiferous and graphitc schist of the Ottauquechee Formation, Cobq - black and gray quartzite of the Ottauquechee Formation, Coqs - sandy quartzose schist of the Ottauquechee Formation, s - serpentinite and talc-carbonate rock.

Ottauquechee - Stowe slices - Additional symbols are: CZs - Stowe Formation - silvery green, muscovite - chlorite - quartz schist identical to CZph, some schists are richer in chlorite, CZsg - mafic schist in the Stowe Formation.

Stowe-Moretown slices - Om, pinstriped schist and mafic schist of the Moretown Formation, Och - black, graphitic schist and thin quartzite of the Cram Hill Formation.

LITHOTECTONIC CORRELATION CHART FOR CENTRAL VERMONT

Figure 5

Stanley and Armstrong 1987

choice. Much of the carbonaceous schist within the Hazens Notch proper is considered to be a highly deformed thrust slice although some of it may well be a facies of the noncarbonaceous Hazens Notch. According to our present thinking the "allochthonous" part of the carbonaceous schist would be reassigned to the Granville Formation. The stratigraphic column for the Lincoln Gap-Pinney Hollow slices does show some carbonaceous schist as part of the Hazens Notch. Resolution of this problem awaits further work.

4. The Ottauquechee Formation is shown to overlie the Pinney Hollow and Stowe. This contact is a major premetamorphic or early metamorphic fault zone. The root zone for the Ottuaquechee slice has yet to be found but it is thought to be located somewhere in the eastern part of the belt mapped as Stowe by Doll and others (1961) or along the western part of the Moretown.

5. The Ottauquechee Formation is an enigma. Because it contains ultramafic rocks it is thought to have been deposited far to the east on oceanic crust. It contains, however, much quartzite and quartz-rich schist which would suggest a nearby continental source. The presence of rare blue quartz supports this view. How do rocks of this type end up with ultramafic rocks? Did the quartz come from a continental source to the east or north and not to the west as traditional is thought to be the case for rocks with blue quartz? Could some of the rocks in the Ottauquechee represent trench clastics eroded from an older accretionary wedge?

6. The Mt. Abraham Schist is definitely allochthonous and roots along the western part of the Pinney Hollow where similar rocks are found in the Rochester quadrangle. These rocks are correlated to the chloritoid-bearing schists in the Group 3 slices of Stanley and Ratcliffe (1985) and the Saint Catherine Formation of Thompson (1967).

It should be evident from the foregoing discussion that many of the units in the pre-Silurian sequence west of the Moretown are thought to be laterial equivalents of each other rather that a simple eastward-younging sequence in the earlier view shown on the Geologic Map of Vermont (Doll and others, 1961). We do believe, however, that the sequence originally became younger to the east because the average age of the sediment cover becomes younger toward the ridge in an expanding ocean (fig. 6). This sequence was then deformed into a complex arrangement of units in a setting that is similar to an accretionary wedge between two converging plates. Our mapping only begins to show the complexity of this deformation. Much of the earlier history is destroyed by the process itself.

Structure - One of the important questions posed earlier in this paper was the nature of the shortening mechanism in the metamorphosed pre-Silurian hinterland. Was folding the major mechanism or was faulting? To what degree are both mechanisms important? And finally, how are they related, if evidence for both exist? Answers to these questions depend on scale, the

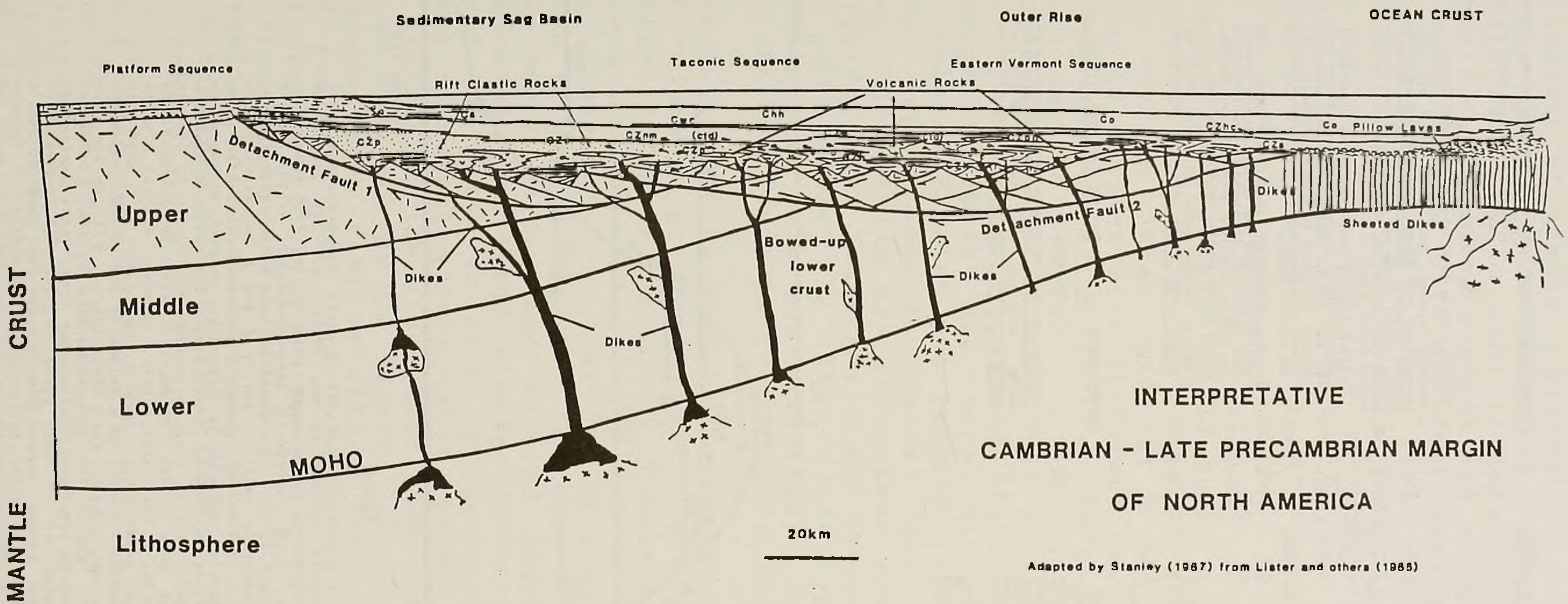


Figure 6 Speculative diagram showing the inferred stratigraphic relations among the Late Precambrian and Cambrian rocks along the ancient North American margin. The rerift model is taken from Lister and others (1986). Evidence is not available to support this particular rift model rather than others for the New England Appalachians. It is used, however, simply as a template to show the inferred stratigraphic relations of the overlying cover rcoks. Lithic designators are the following: CZp - Pinnacle Formation, CZu - Underhill Formation, CZn - Nassau Formation, CZnm - Metawee Member of the Nassau Formation, CZa - Mt. Abraham Schist, CZph - Pinney Hollow Schist, CZs - Stowe Formation, CZhc, carbonaceous schist of the Hazens Notch Formation and its equivalents (Granville Formation), Cs - Sweetsburg Formation, Cwc - West Castleton Formation, Chh - Hatch Hill Formation, Co - Ottauquechee Formation. Other symbols are: ctd - chloritoid, a - albitic rocks, vv - mafic volcanic rocks, ++ gabbroic rocks. Equivalent carbonate and siliciclastic rocks are shown on the left (western) side of the diagram.

Table 1

Criteria for Pre-, Syn-, and Post- Metamorphic Faults

PRE-METAMORPHIC FAULTS — Regional metamorphism occurs after movement. Earlier mineral assemblages are absent.

Criteria

1. Mappable lithic units are truncated on the hanging wall and foot wall of the fault zone.

2. Slivers of rocks foreign to either the hanging wall or foot wall (i.e. ultramafic rocks, felsic plutonic rocks, Proterozoic Y gneiss, exotic metasedimentary rocks).

3. Preserved fault zone fabrics are rare or very difficult to recognize.

SYN-METAMORPHIC FAULTS — Metamorphic recrystallization occurs during faulting.

Criteria

1. Criteria 1 and 2 listed for pre-metamorphic faults.

2. Ductile fabrics preserved along fault zones.
   a. Sharp lithic contacts that truncate internal fabric (i.e. older foliations, compositional layering, quartz veins).
   b. Prominent anastomosing foliation is parallel to or is at a low angle to the fault surface.
   c. Prominent lineation formed by elongate minerals or mineral clusters.
   d. C-S fabrics.
   e. Asymmetric folds, sheath folds, and/or reclined folds with sheared off limbs.
   f. Mylonitic fabrics marked by fine grained, recrystallized minerals compared to larger grains in the surrounding rocks. Porphyroclasts may be present. Evidence for ductile flow is best seen in quartz and feldspar where undulose extinction, deformation twins, deformation lamellae, deformation bands, subgrains, polygonization and sutured grain contacts suggest dislocation glide, dislocation creep and grain boundary migration as important deformation mechanisms. Continued strain during recrystallization is marked by elongate new grains with evidence of internal deformation.
   g. Absence of cataclastic fabrics.

POST-METAMORPHIC FAULTS — Movement occurs after regional metamorphism, but retrograde metamorphism may occur along the fault zone.

Criteria

1. Criteria 1 and 2 listed for pre-metamorphic faults.

2. Strongly oriented, fault zone fabrics are well preserved and will depend upon the depth at which movement occurred. Brittle fabrics reflect movement within several kilometers of the surface whereas fabrics typical of the brittle-ductile transition reflect deeper levels.

   a. Brittle-ductile fabrics consist of a mixture of minerals such as quartz and feldspar in which some of the minerals deform by ductile mechanisms (quartz) and others by fracture (feldspar). The overall mechanical behavior of the rock depends on the dominant mineral.

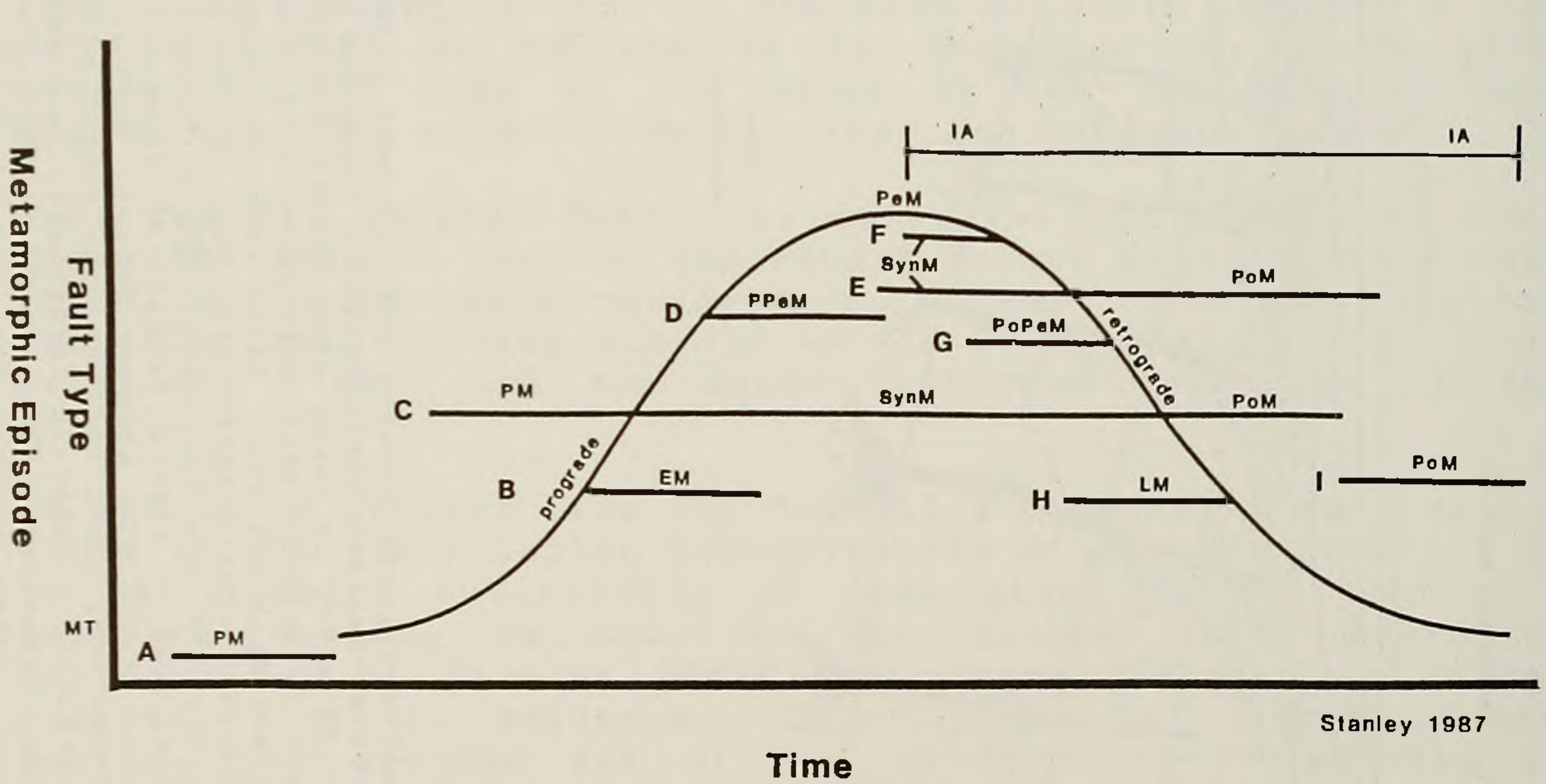
   b. Retrograde minerals along fault zones. In pelitic rocks this may consist of fine grained mica (phyllonite). In gneiss the mica-rich layers are rich in white mica, chlorite, and/or biotite. Finely crushed feldspar may also be present.

   c. Brittle fabrics consist of grains which are angular with random size and orientation. Microfractures are abundant.

COMPOUND FAULTS — Repeated movement occurs under different metamorphic conditions.

Criteria

1. Evidence for post-metamorphic faults superposed on older syn-metamorphic faults are readily preserved and easily recognized. All other examples of fault-superposition are unlikely to be recognized because of subsequent metamorphism.



Stanley 1987

A simplified graph showing the schematic relations of eight representative fault types and a single metamorphic episode in the hinterland. PM represents pre-metamorphic faults, EM represents early metamorphic faults, PPeM represents pre-peak metamorphic faults, SynM represents syn-metamorphic faults, PoPeM represents post-peak metamorphic faults, LM represents late metamorphic faults, and PoM represents post-metamorphic faults. Faults C and E represent faults with a composite history. The distinction between "premetamorphic" and "early metamorphic" faults is commonly diffcult to determine because metamorphism obliterates much, if not all, of the fabric information. The criteria that are listed under "pre-metamorphic faults" would apply therefore to "early metamorphic" faults although some fabric information might be preserved as inclusions in porphyroblasts. The distinction between between "early metamorphic" and "pre-peak metamorphic" faults as well as "post-peak metamorphic" and "late metamorphic" faults is a matter of degree and subjective evaluation of the fault-zone fabric and mineralogy. All faults are distinguished by criteria 1 and 2 for pre-metamorphic faults. Oriented, fault-zone fabrics are preserved for those faults that were active during and after the peak of metamorphism and may well apply to "pre-peak metamorphic" faults. Ductile fabrics are formed during the metamorphic episode, while brittle fabrics develop before and after metamorphism. Transitional fabrics (brittle-ductile) are formed in the region where the fault crosses the metamorphic curve after the metamorphic peak has been achieved. PeM represents the peak of metamorphic intensity. MT represents the "metamorphic threshold" IA represents the range of possible isotopic ages associated with the metamorphic episode.

availability of easily - recognized units, and the degree to which critical fabrics are preserved. Metamorphism, despite producing minerals that can be used to determine the physical conditions of deformation, commonly anneals and simplifies preexisitng fabrics so that much of the deformational history is erased. Thus, we are left with lithic units whose field characteristics are sufficiently distinctive that they can be recognized and mapped from place to place at a scale suitable to show their mutual relations. For example, if we can show that a series of metasedimentary units can be traced consistently through a complex fold pattern and can be used to predict larger structures, then a fair chance exists that they are part of an original depositional sequence. This conclusion is strengthened if their contacts are gradational, even over a small distance, and show no signs of intense, concentrated shear. On the other hand, if a metasedimentary or metavolcanic unit is in contact with different units then the contact is either an unconformity or a fault. Here again the fabric along the contact may be helpful. If the contact is a synmetamorphic or postmetamorphic fault, then well oriented fabrics with displacement criteria are commonly present (table 1). Premetamorphic or early metamorphic faults are more difficult to recognize because fault-related fabrics are largely destroyed by metamorphism. Here we must depend upon the mutual relations of four or more lithic units. Such faults are recognized when two or more units on both the footwall and hangingwall are truncated along the contact (i.e. the fault, table 1). This criterion applies equally well with synmetamorphic and postmetamorphic faults, but in these cases additional evidence is available in the fault fabric.

Additional problems must be solved when analyzing the simple-shear fabric along a proposed fault. The presence of a well developed schistosity and a prominent lineation is really not sufficient since these types of fabrics can be formed equally well by pure shear (flattening) as they can by simple shear. Such asymmetrical fabrics as drag folds, C-S fabrics, mica fish, porphyroclasts with asymmetrical tails, and rotated porphyroblasts are the type of evidence that indicate simple shear rather than flattening. If a contact has features such as these, it is quite likely that the contact in question is a fault, particularly if it truncates other rock units. On the other hand, the presence of symmetrical fabrics such as overgrowths or "beards" on porphyroclasts and the absence of asymmetrical fabrics with a consistent sense across a contact would indicate that the contact is simply a product of severe flattening (pure shear). An offset of an older marker can occur in situations like this but here it is due to volume loss during flattening rather than simple shear displacement along a fault. In many situations not all the evidence is available to determine if a given contact is a fault or not.

Table 1 lists some of the criteria that have been used to recognize faults in the pre-Silurian hinterland of central Vermont. With the possible exception of some of the old faults that we map as "pre- or early metamorphic" faults, most of the faults that we have mapped in the central Vermont transect were formed under elevated temperatures (300^0C to 530^0C) and pressures (2 Kbar to 5 Kbar or approximately a depth of 6 to 16 km) and thus are considered to have formed by ductile or semiductile processes. These faults are divided into eight groups depending on when they occurred relative to the dominant metamorphism represented by the mineral assemblages in the surrounding rocks. The simplified graph in Table 1 shows the schematic relations of these eight groups relative to a single metamorphic episode. The faults are classified as "premetamorphic", " early metamorphic", "pre-peak metamorphic", "peak metamorphic", post-peak metamorphic", "late metamorphic", or "postmetamorphic" depending on when they occurred relative to the peak of metamorphism when the highest grade assemblage was formed in the rocks. "Pre-peak", "peak", and "post-peak" metamorphic faults are considered to be "synmetamorphic". Compound faults are those where repeated movement occurs under different metamorphic conditions. The age of synmetmorphic and postmetamorpic faults can be determined by isotopic analysis of minerals with radioactive elements that formed during the history of the fault. For all the premetamorphic and early metamorphic faults these ages simply represent an upper limit - the faults are older than the calculated age. The criteria and concepts listed in Table 1 have been used with some success (O'Loughlin and Stanley, 1986; Lapp and Stanley,1986; DelloRusso and Stanley,1986) but are constantly undergoing revision as a result of current field work and petrographic analyses.

Let us return to the early question of the relative importance of folding and faulting. There is little doubt that folding has been an important component in the total shortening in the pre-Silurian hinterland. Minor folds of at least five generations can be seen in many parts of the area. Major isoclinal folds are outlined by many mappable units. The hinges of these folds plunge to the east and southeast nearly parallel to the dip of their associated axial surface schistosity. Their geometry therefore decribes complex sheath folds whose shape is a product of severe flattening and profound elongation parallel to the long dimension of the sheath (Armstrong and Prewitt, in progress). It is this generation of folds (Fn) that accounts for most of the recorded fold - related shortening according to our present studies. The schistosity (Sn) that is parallel to the axial surface of these folds forms the dominant schistosity throughout the region. This generation of structures deforms and severely transposes an older generation of folds (Fn-1) and is, in turn, folded by several younger generations (Fn+1 and Fn+2) as it is traced westward across the axis of the Green Mountain anticlinorium (O'Louglhin and Stanley. 1986; Lapp and Stanley, 1986). A even younger fold generation (Fn+3) is found to the east along parts of the Northfield - Braintree Mountains.

The major part of the shortening across the pre-Silurian hinterland, however, is accounted for by the pre- or early metamorphic and synmetamorphic faults. Much of this shortening here appears to be associated with the older pre- or early metamorphic faults because they largely control the distribution of similar map units across the region (for example, the carbonaceous schist of the Hazens Notch Formation or the Mt. Abraham Schist). It is these faults that are subsequently deformed into large sheath folds. The sheath folds are in turn cut by the more planar synmetamorphic faults which parallel the dominant Sn schistosity. The older Fn-1 folds may have developed with the older faults in which case the faults would be "early" rather than "pre" - metamorphic.

Metamorphism and Age of Deformation The metamorphism in central Vermont has received considerable attention by Albee (1965, 1968), Laird and Albee (1981), Laird and others (1984). This work and new findings are summarized by Laird in this guidebook. Studies on the relation between structure and mineral growth are reported in O'Loughlin and Stanley (1986) and Lapp and Stanley (1986). The results of this work along the western side of the Green Mountains shows that there is a continuous overlapping relation between mineral growth and the structural sequence Fn-1 through Fn+2. For example Sn-1 is preserved in isolated hinges between Sn or as inclusion trails of graphite, white mica, quartz, and opaque in porphyroblasts of albite and garnet. Some of these trails can be traced into the Sn schistosity. All the peak metamorphic minerals such as chloritoid, garnet, kyanite, and plagioclase either are oriented parallel to or grow across the Sn schistosity. Chlorite occurs as fine and coarse grains. The fine - grained chlorite is oriented parallel to Sn. The coarser - grained chlorite developed from garnet and is found in pressure shadows that are oriented parallel to Sn. Prominent chlorite streaks, which also grew from garnet, are well developed on the dominant Sn schistosity in the Mt. Abraham Schist and the Pinney Hollow. Chlorite is also deformed by Sn+1, and is oriented parallel to Sn+1 along the west side of the Green Mountains. The only minerals which clearly represent a separate period of growth are large muscovite and biotite grains that cut across Sn+1 and Sn+2 with random orientation in the Underhill and Hoosac Formations. These cross micas are not the ones that gives the 385 m.y. age at Mt. Grant. Those grains are associated with Sn and Sn+1.

The Sn schistosity on the west side of the Green Mountains can be traced eastward to Granville Gulf where it forms the dominant schistosity defined by chlorite, epidote, and actinolite in greenstones. Barroisitic hornblende surrounding actinolite from this locality gives 39Ar/40Ar ages of 448-471 m.y. (sample V145 of Laird and Albee, 1981, Laird and others, 1984). The barroisitic hornblendes may have formed during Sn-1 although there is no evidence for this relation. Laird (this guidebook) states that these rocks have undergone an earlier medium-high pressure series metamorphism represented by the barroisitic hornblende. This was then followed by a lower grade greenschist facies metamorphism. Similar relations are preserved in the pelitic schists to the west where kyanite and garnet are retrograded to sericite and chlorite. At present the relation between the isotopically - dated minerals and the respective schistosities that have been recognized across transect are still not clear. Are the younger ages simply a result of cooling or were metamorphic reactions still going on west of the

Green Mountain axis until 385 m.y. ?

ACKNOWLEDGEMENTS

Funding for this study has been provided by the U.S. Forest Service (U.S. Department of Agriculture Contract 524526) and the National Science Foundation (EAR 8516879). Publication of maps and bulletins for the central Vermont regions has been supported by the National Science Foundation (EAR 8516879) and the U. S. Department of Energy (DE-FG02-81WM46642 awarded to the Vermont Geological Survey).

THE LINCOLN MASSIF AND ITS IMMEDIATE COVER

By

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ITINERARY

The transect across central Vermont is divided into two days. The Saturday trip will begin along the western boundary of the eastern Lincoln massif and progress eastward to the crest of the Green Mountains where we will only see the Hoosac, Underhill, Mt. Abraham and Hazens Notch Formations. The remaining part of the pre-Silurian eastern Vermont sequence will be seen on Sunday when the trip will concentrate on the Hazens Notch, Pinney Hollow, Ottauquechee, and Stowe Formations in the valleys of the White and Mad Rivers. The introductory part of this paper entitled "A Transect through the Pre-Silurian rocks of Central Vermont" applies to both days. The figures are numbered sequentially as if the two days were a single trip. The "References Cited" also apply to both days.

Meeting place for Saturday trip - Lincoln General Store, Lincoln, Vermont. Time 8:30 A.M.

Geological maps, cross sections and texts for all the stops for Part I in the Lincoln massif and the eastern cover sequence are available through the Vermont Geoogical Survey. They are Tauvers (1982a), DiPietro (1983), Strehle and Stanley (1986), DelloRusso and Stanley (1986), O'Loughlin and Stanley (1986), and Lapp and Stanley (1986). The following stop descriptions, therefore, depend heavily on this information.

Stops 1 through 6 - The western boundary of the Eastern Lincoln massif (fig. 7; plate 2, DelloRusso and Stanley, 1986) (DelloRusso) - These stops show the basement cover contact on the western limb of the massif, a small, newly discovered syncline of cover within the massif, and Late Proterozoic mafic dikes. A discussion of the chemistry of these dikes will be presented by Ray Coish of Middlebury College. These stops will also demonstrate the difficulty in recognizing the precise boundary between the Middle Proterozoic basement and its immediate cover even in areas where the deformation is not intense.

Stop 1 - This is a 12 ft high exposure of coarse cobble conglomerate containing abundant, matrix-supported cobbles of quartzite and gneiss. The cobbles are derived from the nearby Mt. Holly Complex. The composition of the matrix is similar to the finer-grained metawacke found elsewhere in the basal part of the Pinnacle Formation. Sedimentary features such as bedding and

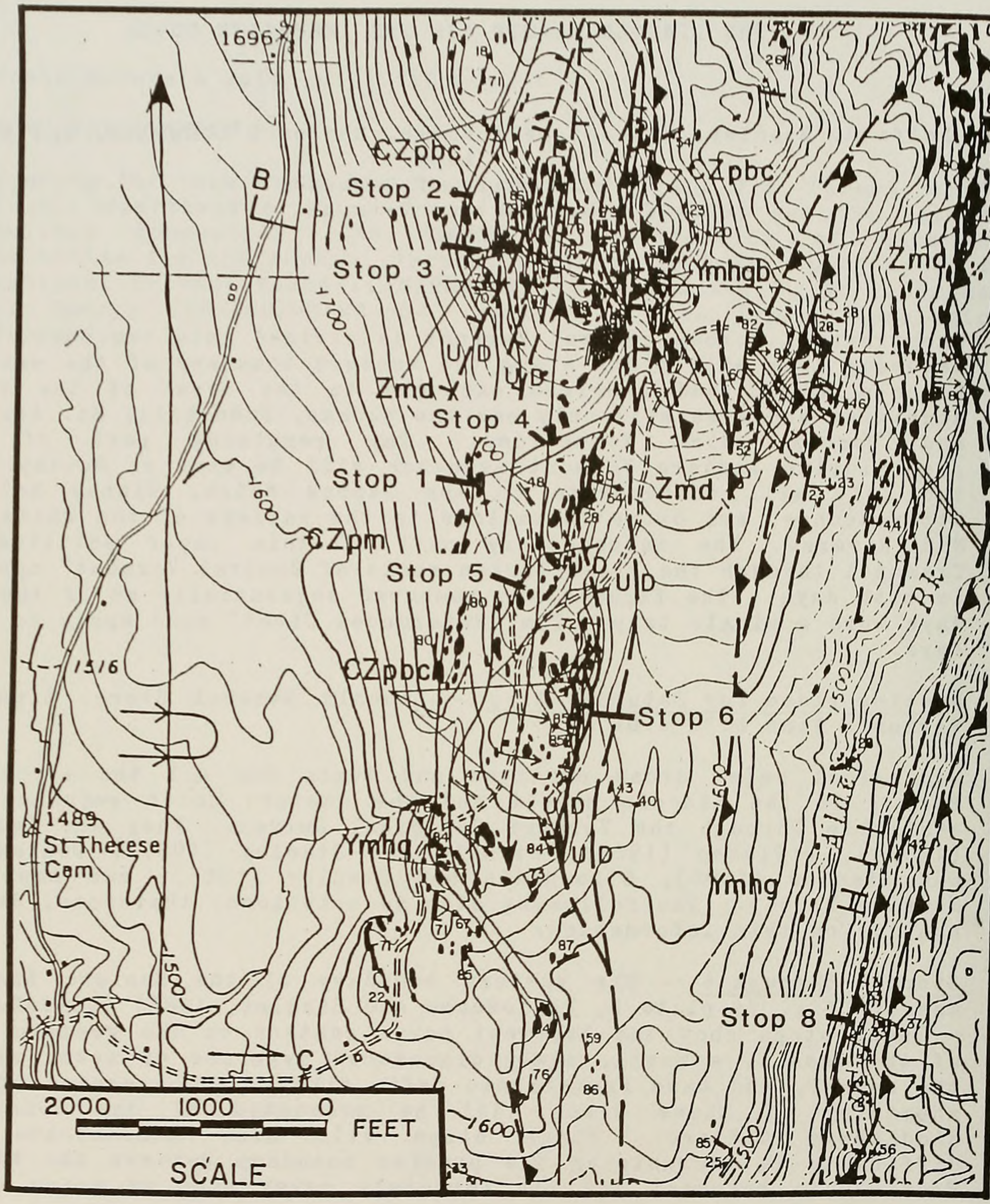


Figure 7

Geological map taken from DelloRusso and Stanley (1986, pl. 2) showing stop locations 1 through 6 for the western limb of the Eastern Lincoln massif. The long arrows parallel to the roads show the route. Lithic symbols are the following: Ymhg - granitic gneiss of the Mt. Holly Complex, Ymhgb - biotite schist zones in the granitic gneiss of the Mt. Holly Complex, Zmd - mafic dikes in the Mt. Holly Complex, CZpbc - basal conglomerate of the Pinnacle Formation, CZpm - schistose metawacke of the Pinnacle Formation. High angle faults are shown by thick dashed lines with "U" and "D" symbols. Thrust faults are shown by thick dashed lines with solid barbs.

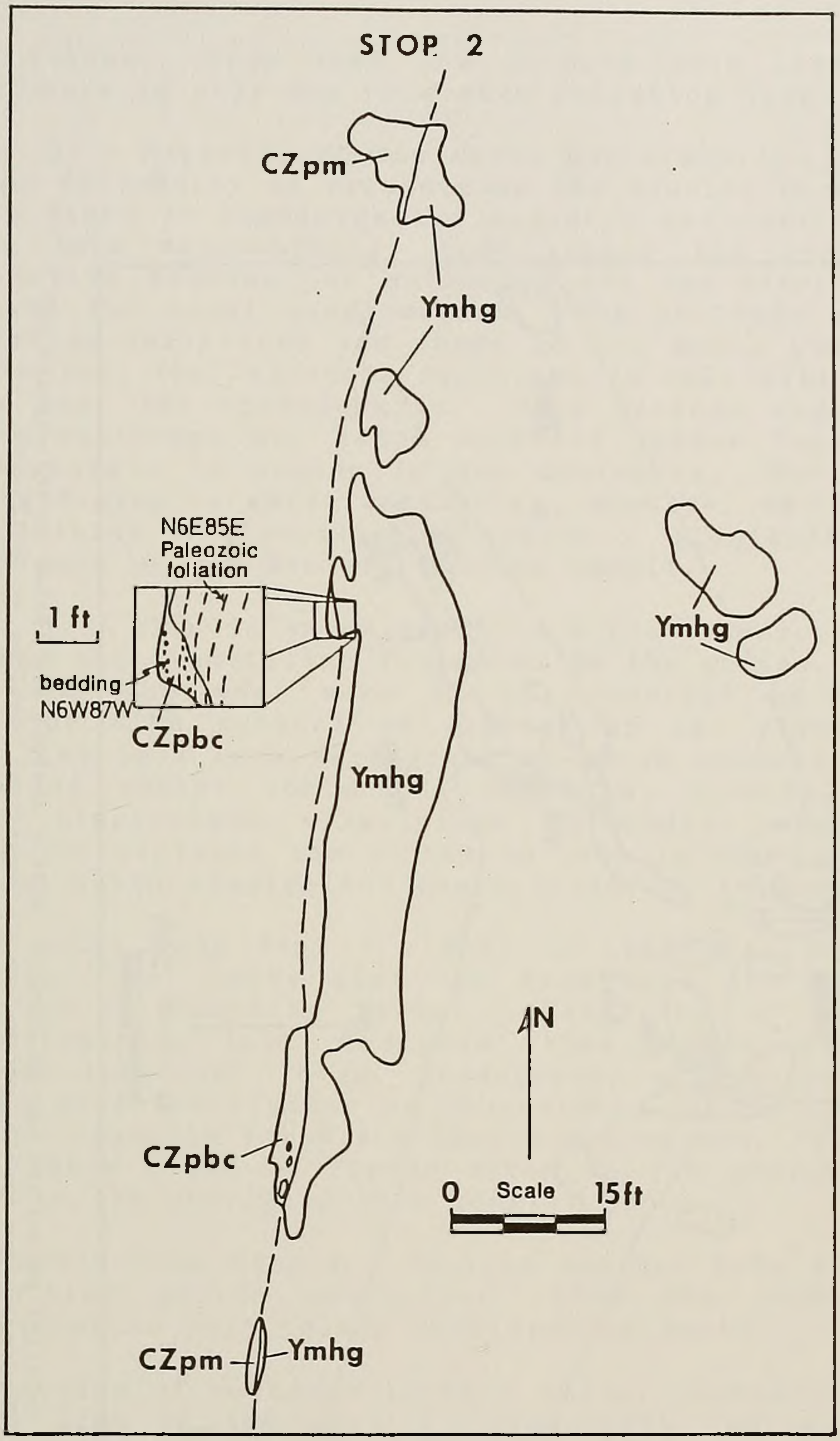


Figure 8

Detailed geological map of Stop 2. Lithic symbols are listed in the caption for figure 7. Geology by DelloRusso.

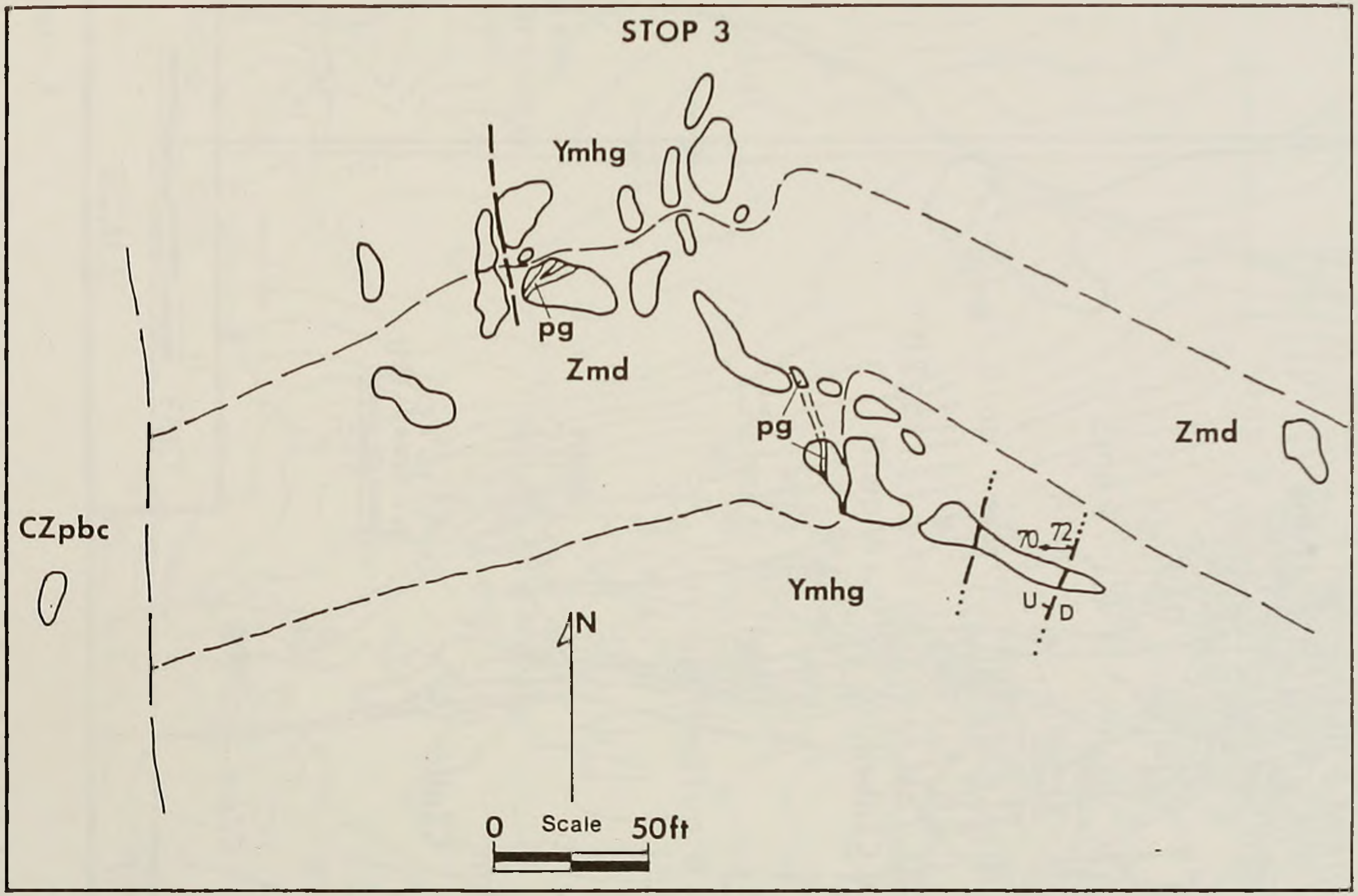


Figure 9

Detailed geological map of Stop 3. Lithic symbols are listed in the caption for figure 7. The symbol "pg" refers to pegmatite. Geology by DelloRusso.

channels are absent. Note that the cobbles show little or no deformation. There is only one Paleozoic foliation here.

Stop 2 - (fig. 8) - Exposure of the basal unconformity. This stop illustrates the difficulty in recognizing the precise unconformity. As a guide try first to recognize the boulders and cobbles. After you have done this successfully, then locate the unconformity. Note the similarity between the metawacke and the deformed gneiss directly beneath the basal conglomerate. The attitude of bedding and the respective foliations are shown on the small inset map on figure 8. Note that the Paleozoic foliation is only well developed in the gneiss near the unconformity. This feature suggests that the gneiss was weathered and hence weakened before deposition of the cover. Magnetite is common in the metawacke. The gneiss is the typical light-gray to white weathering, massive, medium-grained muscovite - perthite - microcline - quartz - plagioclase gneiss that underlies much of the Eastern Lincoln massif.

Stop 3 - (fig. 9) - This is an exposure of a Proterozoic mafic dike that cuts across the Grenvillian foliation in the gneiss. The dike can be traced to the west were it is truncated by the basal conglomerate. The dike contact is exposed in two places in the outcrop. The dike rock is a dark-green to brown weathering, fine- to medium-grained schist containing sericite, epidote, biotite, chlorite, and plagioclase with minor carbonate, opaques, and sphene. Along the contacts the schist is rich in biotite. Large, relict, euhedral white plagioclase phenocrysts are common.

Stop 4 - Walk south from Stop 3. This is another exposure of a Late Precambiran dike. Note that the dike cuts the Grenvillian foliation in the surrounding gneiss. Xenoliths of gneiss are present. Geochemical data indicate that these dikes were originally high-titanium, high phosphorous, alkaline basalts (Filosof, 1986) that are similar in composition to the Tibbit Hill Volcanics of the Pinnacle Formation (Coish and others, 1986). Thus dikes such as these could be feeder dikes to the overlying mafic volcanic rocks in the overlying rift clastic rocks.

Stop 5 - Walk south from Stop 4. This is another Late Proterozoic dike. It contains gneiss xenoliths. Note the contact. The Paleozoic foliation is only poorly developed at best.

Stop 6 - This series of outcrops is in a small, parasitic syncline on the western limb of the massif. High-angle faults developed during the formation of the syncline and cut the basement as well as the overlying cover. As a result it is often difficult to separate deformed basement gneiss along the faults from the overlying basal cover. This is further complicated by the fact that the gneiss directly beneath the cover appears to have been weathered before the basal conglomerate was deposited. The presence of gneiss boulders and quartz cobbles in several places, however, helps delineate the unconformity and resolve some of these problems. In one exposure a pegmatite in the gneiss is truncated by the unconformity. Along the faults, the gneiss and overlying

Pinnacle metawacke are deformed into a tan-colored schist that contains quartz, feldspar, and sericite.

Stop 7 - "Crash Bridge" - (fig. 10 and 11; DelloRusso and Stanley, 1986) (Stanley) - This outcrop is located along the New Haven River at the second bridge to the east of the general store in Lincoln. Here the Middle Proterozoic granitic gneiss and amphibolite are truncated by the basal conglomerate of the Pinnacle Formation which is strongly overturned to the east. Unlike stop 1 the contact here is a thrust fault marked by sheared biotite schist and muscovite schist with gneiss fragments (fig. 10). Again these rocks are difficult to distinquish from some of the overlying metawackes because muscovite schist looks much like the overlying metawacke. In thin section, however, the two differ in the degree of grain orientation and the percentage of biotite. The muscovite schist has no biotite and the muscovite and quartz are strongly oriented compared to the metawacke where the grains are more random and biotite is fairly abundant. It is therefore not surprising that the contact between the fault zone and the basal conglomerate is indistinct.

The basal conglomerate at "Crash Bridge" is quite different than the conglomerate at Stop 1 which is more typical of the basal unit. Here the conglomerate is largely made up of clast-supported cobbles and boulders of granitic gneiss that are arranged in lensoid bodies separated by magnetite-bearing metawacke. Although these rocks are tightly folded and cut by a pervasive Paleozoic schistosity, their shapes and mutual relations suggest that they are paleochannels in which the basal part of the channel faces toward the Middle Proterozoic rocks beneath and east of the bridge. Tauvers (1982a) who was the first to recognize the nature of this deposit suggested that the clast-supported channels formed during periods of very rapid flow whereas the matrix formed during period of reduce flow. One such environment where deposits such as this can develop is on an alluvial fan in arid to semi-arid climates (Tauvers, 1982b). A submarine fan, however, might be equally plausible.

As shown on the geological map and profile section the layering in the Pinnacle Formation has been deformed into a series of reclined folds that become tighter as the fault contact is approached. These folds and the associated axial surface schistosity are coeval with gently-plunging folds along the unfaulted, western boundary of the massif. Their counterclockwise asymmetry indicates that they developed on the western, overturned limb of the Eastern Lincoln massif. Furthermore, we suggest that these folds have been rotated from their original low plunge to a steeper plunge as the thrust fault was developed on the overturned limb of the massif. The transport lineation on thrust faults throughout the massif plunges to the east or slightly south of east (DelloRusso and Stanley, 1986, pl. 4).

Deformation and metamorphism at "Crash Bridge" is Taconian as suggested from isotopic age analysis at this and nearby localities. A biotite K/Ar age of 410 m.y. was obtained from a biotite schist

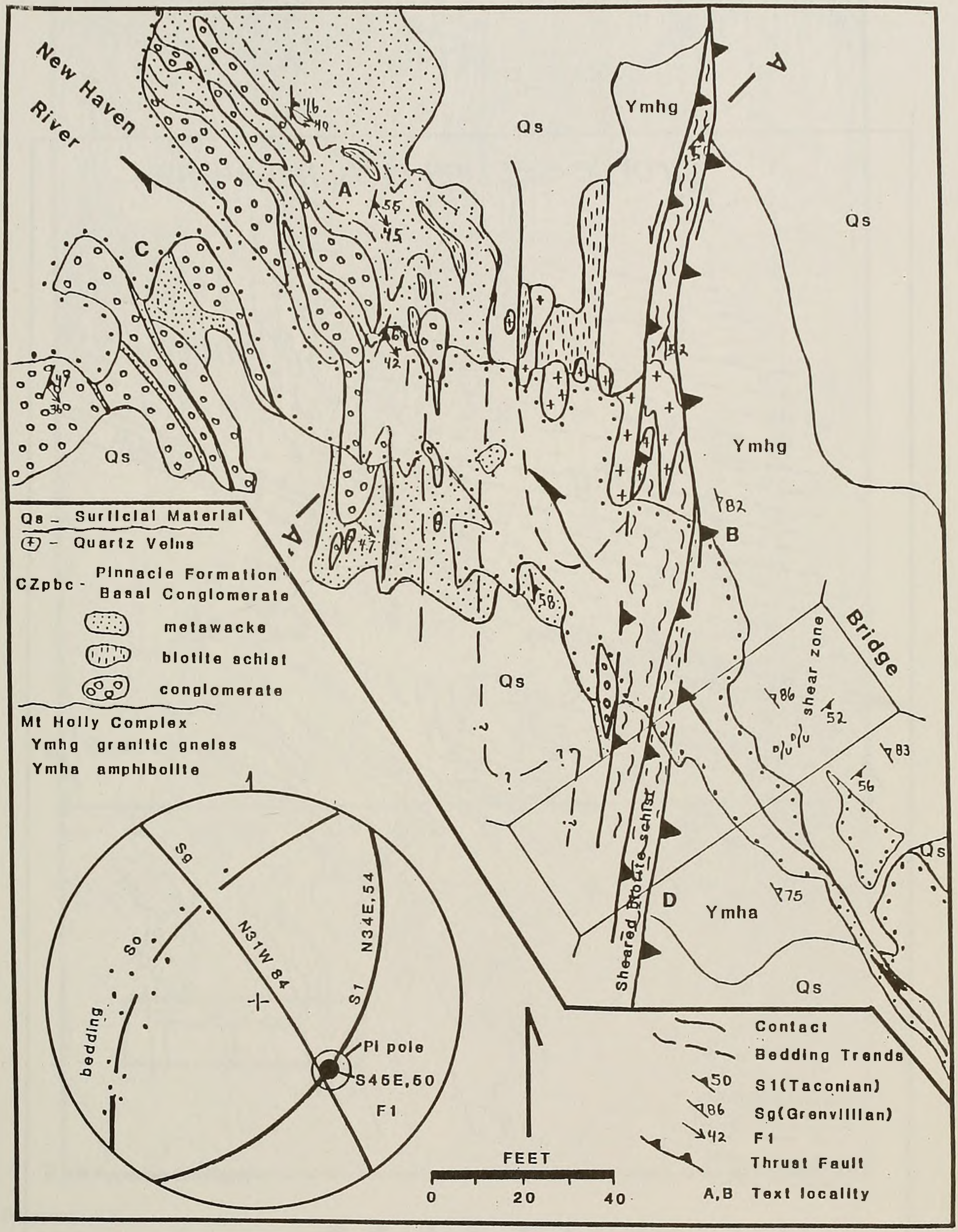
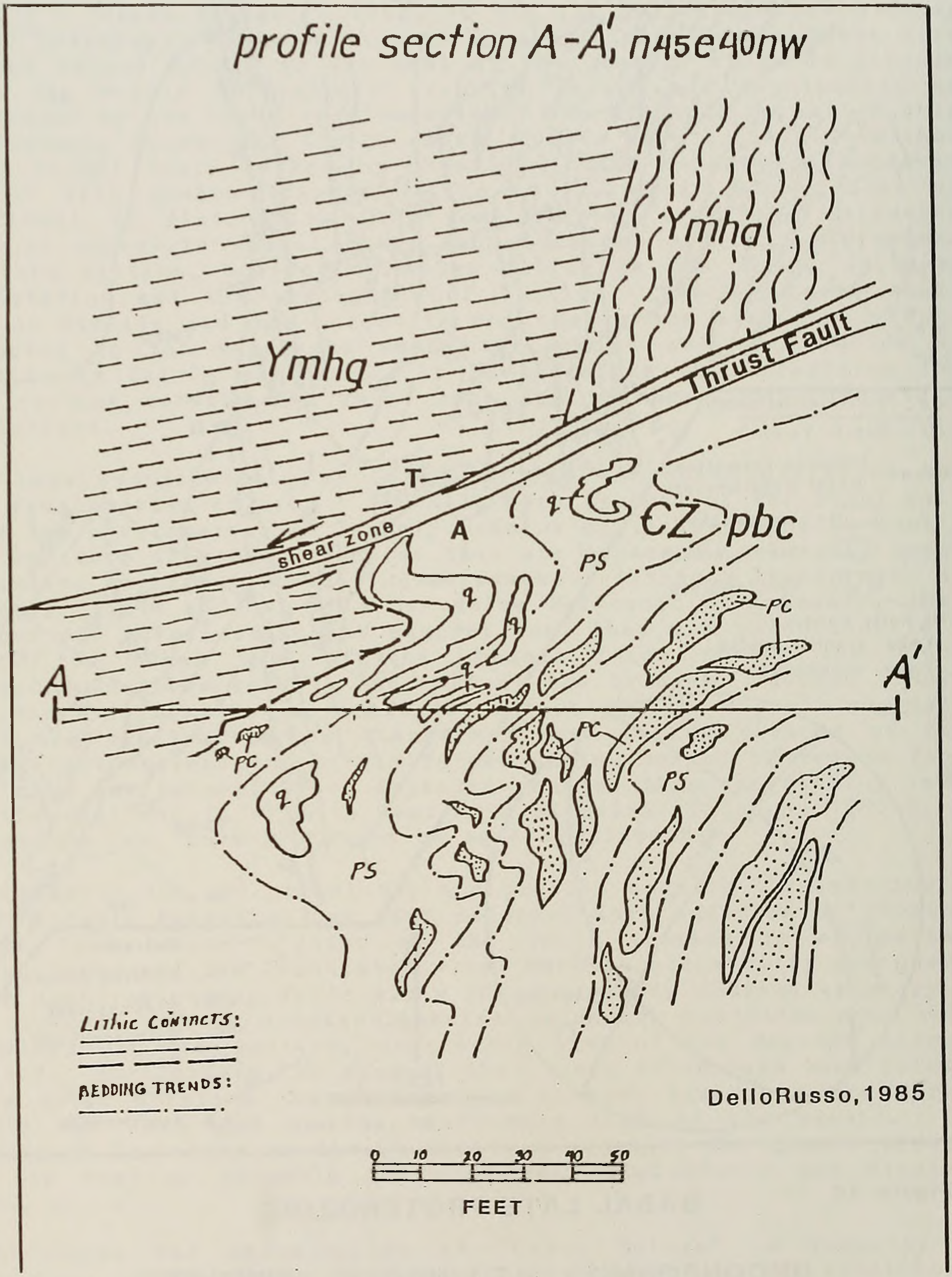


Figure 10

BASAL LATE PROTEROZOIC

UNCONFORMITY AT LINCOLN, VERMONT

Tauvers, 1982 Stanley and DelloRusso 1985



FAULTED UNCONFORMITY AT LINCOLN, VERMONT

Figure 11

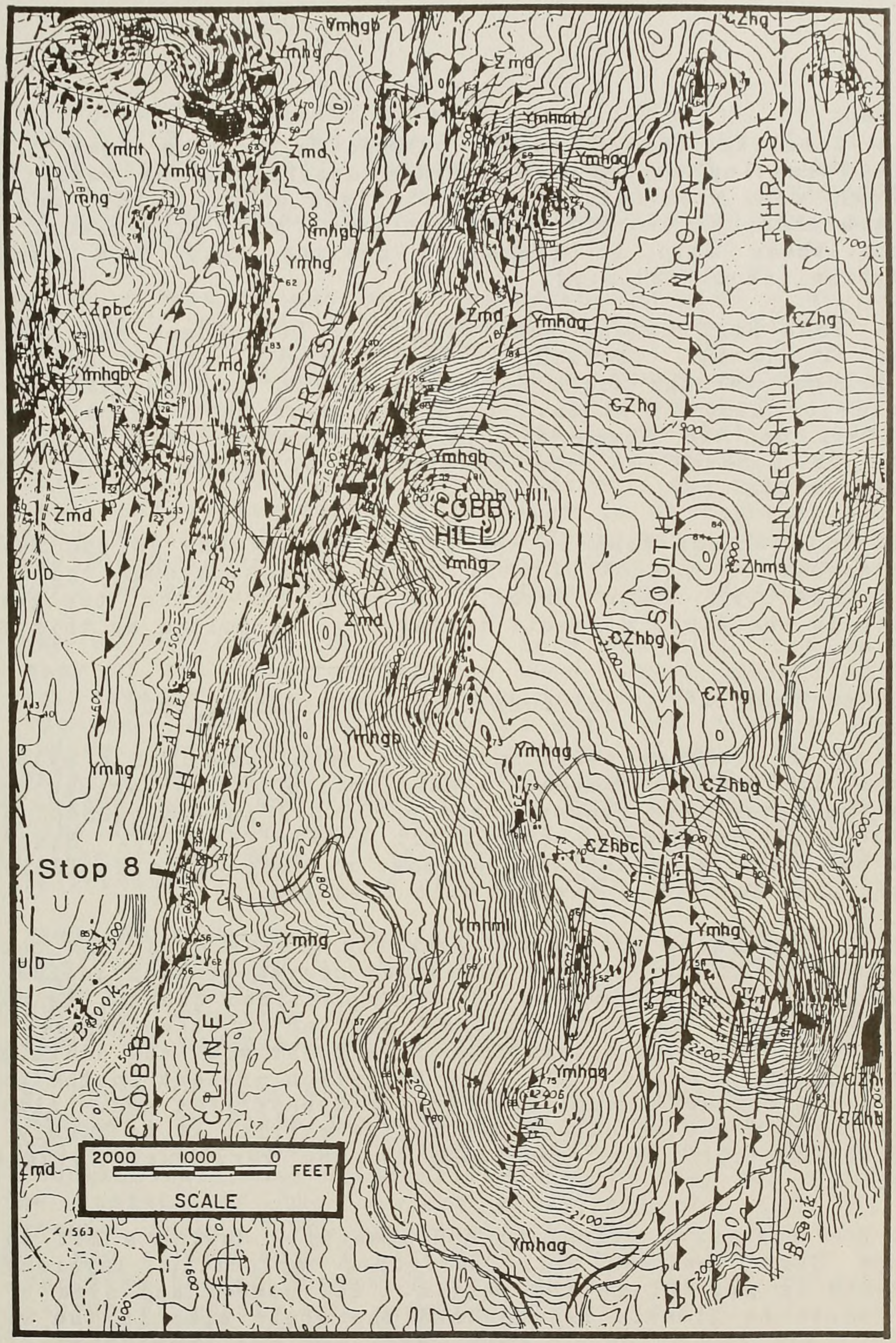


Figure 12

Geological map taken from DelloRusso and Stanley (1986, pl. 2) showing the location of Stop 8. Lithic symbols are listed in the captions for figures 3 and 7. Other symbols can be found on plate 2 of DelloRusso and Stanley (1986). Long arrows parallel to the roads show the route to Stop 8. This is private property and you must ask for permission to visit these outcrops.

at this locality (Cady, 1969). Directly to the west in the Pinnacle Formation other isotopic age determinations reported by Cady (1969) were recalculated by Sutter and others (1985) and fall within the range 370 to 397 m.y. with about a 20 m.y. error. They interrupt these data to represent cooling ages from Taconian metamorphism (about 465 m.y).

Stop 8 - Cobb Hill Thrust Zone (fig. 12; DelloRusso and Stanley, 1986, pl. 2 and 4). (DelloRusso and Stanley) - Drive south from "Crash Bridge" through South Lincoln (about 4 or 5 houses) along a dirt road for about 7 miles. Turn northwest on a Forest Service road. Park at the house at the very end of this road. The locality is located on the east side of Alder Brook and is difficult to find.

The Cobb Hill thrust zone is the major thrust zone in the Eastern Lincoln massif. A branch of this thrust passes through the "Crash Bridge" locality. Alder Brook, which is a major topographic lineament in the Lincoln massif is largley controlled by this fault zone.

NO HAMMERS PLEASE - YOU MAY COLLECT LOOSE MATERIAL ONLY.

The series of outcrops illustrates the development of a mylonitic fault zone in granitic gneiss. A description on the various phases and the supplemental discussion are published in DelloRusso (1986) and DelloRusso and Stanley (1986, pl. 4 specifically). Study the outcrop beginning with the southernmost outcrop. The evolution of the fault zone fabric is divided into 4 phases.

Phase 1 - Fragmentation of the granitic gneiss - In this outcrop we find pebble- to cobble-size clasts of coarse-grained gneiss in a finer grained, foliated matrix of recrystallized quartz, feldspar, and sericite that is compositional similar to the gneiss. The sericite forms from the alteration of feldspar with the addition of water and aluminum oxide. The only foliation present in the gneiss clasts is the coarse mineral layering of Grenvillian age. Although there is no evidence of cataclasis in the matrix, it is likely that fracturing may have been important in the very early stages of deformation.

Phase 2 - Pervasive mylonitic schistosity - Moving a bit to the north we find that a well developed mylonitic layering or schistosity is present in the matrix. The number and size of the gneiss clasts is greatly reduced and the grain size in the matrix is finer. The foliation is well developed and quartz grains and clusters are elongated into a pronounced lineation. We suggest that recrystallization and recovery has continued to reduce the grain size and eliminate the original gneiss clasts.

Phase 3 - Folded mylonitic schistosity - Moving still farther north we find that the mylonitic schistosity is now folded into tight, reclined folds. These folds are asymmetrical with a clockwise or north-over-south asymmetry. Like the folds at "Crash

Bridge" these folds plunge slightly south (S65E,35) of the mineral lineation (S71E, 30). We suggest that the mylonitic fabric at this stage has become unstable because the grain size has been reduced to such a size that it is easier to deform by folding than by further grain-size reduction. In essence, the mylonite has strain hardened.

Phase 4 - Shear bands and fragmented mylonite - Moving still farther north we find that shear bands are present on the attenuated short limbs of the reclined folds. The shear bands are very planar, thin, micaceous layers that clearly truncate the older mylonitic foliation at a low angle. Continued deformation and shear movement along the bands offsets and rotates this older foliation and results in a new "breccia" consisting of mylonitic fragments. The development of sericite along the shear bands indicates that retrogession of feldspar is an important softening mechanism during this phase of deformation. The absence of recognizable deformation of the shear bands indicates that they represent the youngest phase in the development of the fault zone.

Continue to the north for 50 ft. Here you will find an outcrop with many of the features that we have already described. Here, however, several large quartz veins intrude the zone parallel to the layering. Near the quartz veins are sericite-rich schists or phyllonites. We have seen the occurrence of sericite at a number of stages in the evolution of the fault zone. Here they are formed on a larger scale and represent the final product of alteration along the fault zone where siica-rich solutions invade the zone. It is likely that much of the strain along the Cobb Hill thrust zone is concentrated in these sericite-rich zones.

This fault is a postmetamorphic thrust zone in that it overprints the older Grenvillian metamorphism. We estimate that it formed under "ductile" conditions with temperatures in the order of $400^{\circ}C$ to $450^{\circ}C$ and pressures of 4 or 5 kbars (10-15 km), although the indicators are certainly not precise.

Stop 9 South Lincoln Bridge (fig.13) (Stanley) - Return north to the "village" of South Lincoln. Try not to block traffic across the bridge. This outcrop, which consists of mylonitic gneiss and basal conglomerate, was originally mapped by Cady and others (1962) as the contact of the eastern cover sequence and the Middle Proterozoic rocks. The presence of chlorite schist, biotite schist, and garnet schist identical in composition and aspect to the overlying Hoosac Formation to the west and structurally below the outcrop of gneiss and conglomerate indicates, however, that a major thrust fault occurs along the lower contact of the gneiss. This relation was first recognized by Tauvers (1982a, 1982b) who considered it to be part of the Underhill thrust zone. DelloRusso and Stanley (1986) have recognized similar relations along the eastern border of the Lincoln massif. This particular fault they named the South Lincoln thrust. Petrofabric analysis has shown that the gneiss is mylonitic with distinct layers of quartz and feldspar (Strehle and Stanley, 1986). A very prominent mineral

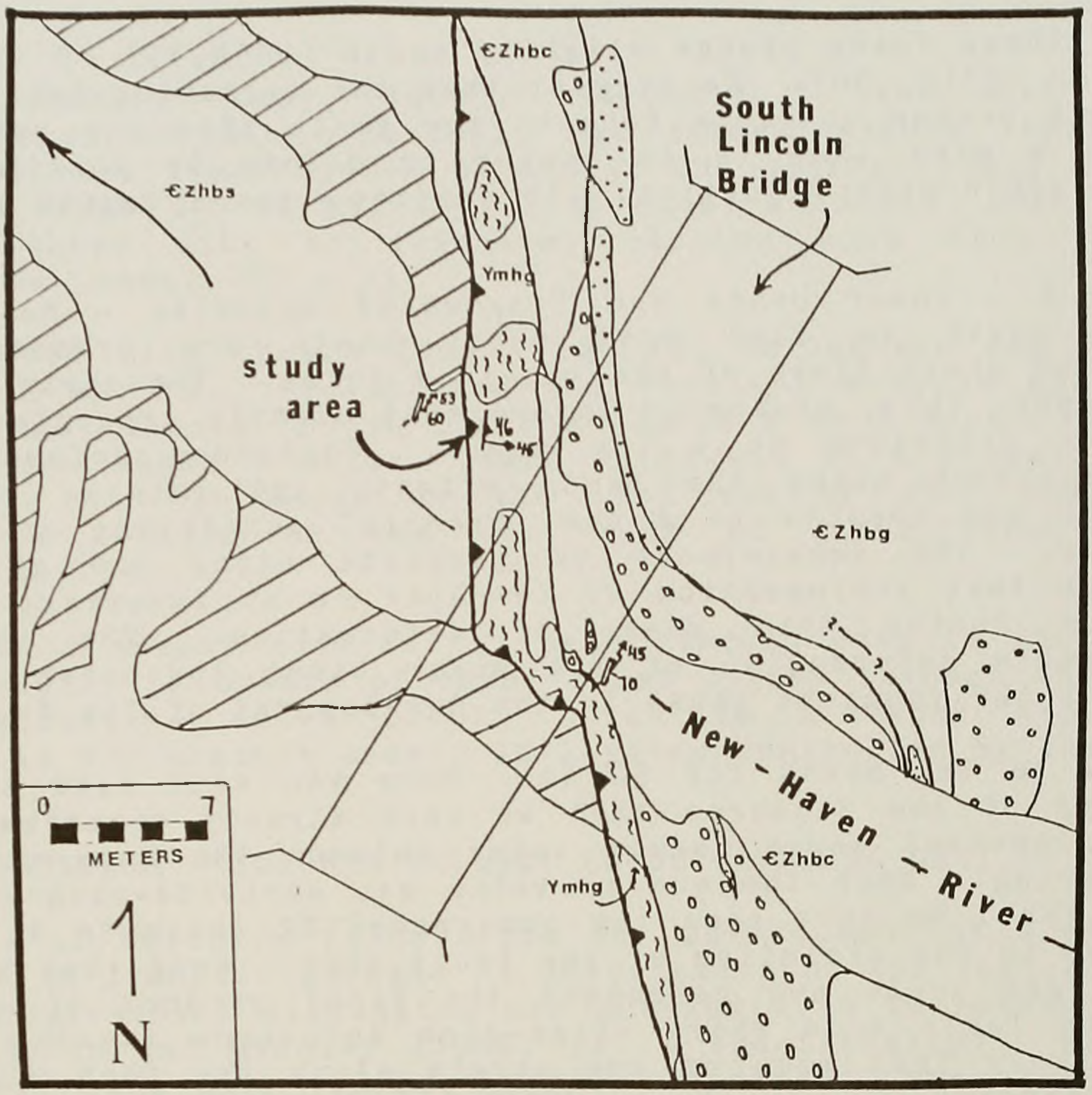
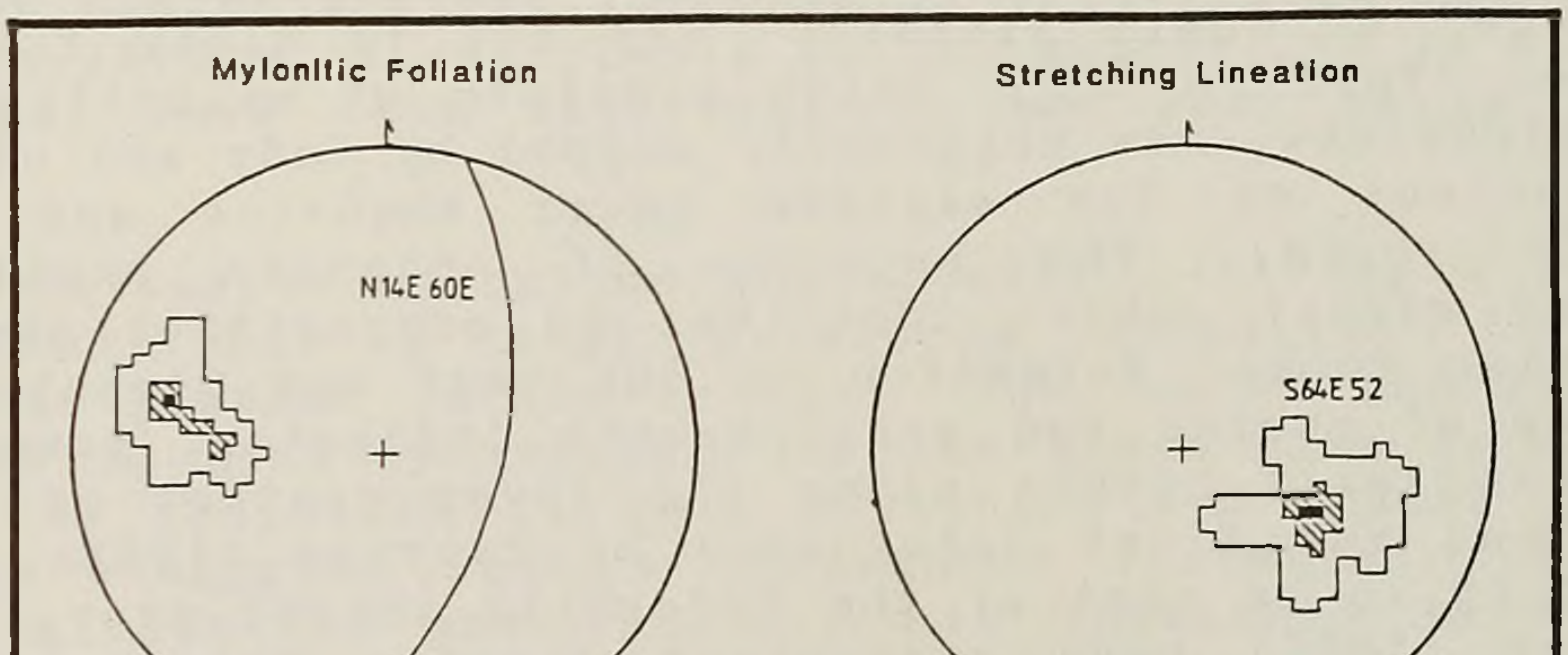


Figure 13

At the bridge at South Lincoln, the mylonitic gneiss of the Mt. Holly Complex (Ymhg) is in thrust contact with the Hoosac Formation (CZhos). A pervasive, platy mylonitic foliation (Sm) and penetrative mineral lineation dominates the gneiss (represented by symbols in gneiss sliver, Ymhg, and shown by the contoured data in Figure 33). The fault is deformed by younger folds (F2) represented by the symbols to the east and west of the sliver. In Jerusalem, the fault-related deformation along the Underhill thrust fault (Utf) is distributed over a much wider zone and numerous tectonic slivers are juxtaposed in the fault zone (fig. 45, Plate 4). Geology mapped by Tavers (1982) and modified by Stanley.



Lower hemisphere equal area projections of n poles to the mylonitic foliation and n stretching lineations. Tick marks north. Contours are given as the percentage of points per 1% area. The stretching lineation is defined by elongation of quartz and feldspar grains.

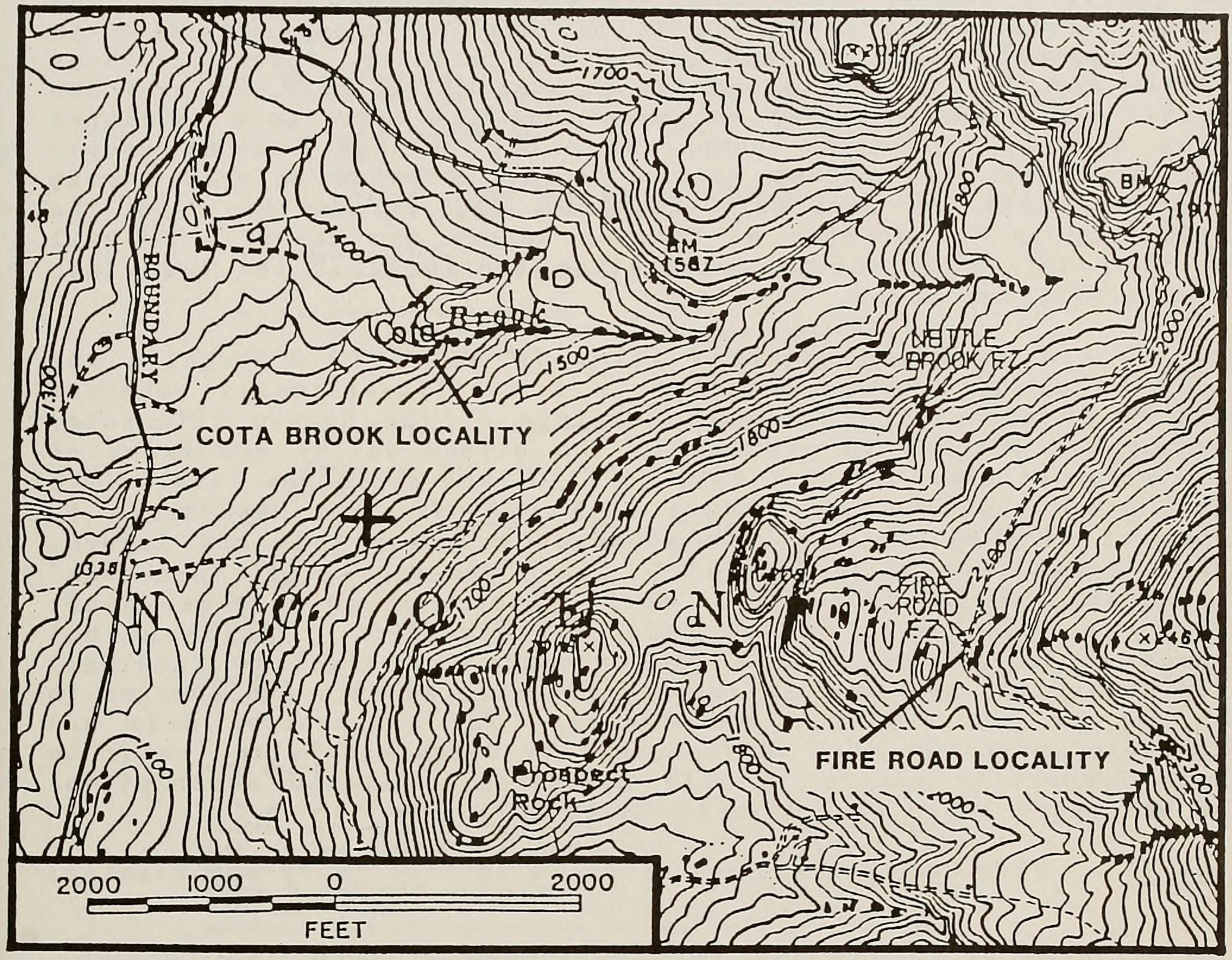


Figure 14

Location map for the Cota Brook Sequence and the Fire Road Fault Zone. Topographic base is from the Lincoln 7.5 minute topographic quadrangle. Elevations are in feet above sea level. Cross mark located below the label for Cota Brook locality is at longitude 72.57.30 west and latitude 44.05.00 north. The road in the northern (upper) part of the map is the Lincoln Gap Road. Lincoln Gap is off the map to the east. A part of Gerry Road is seen along the western boundary of the map. Black spots represent observed outcrop.

elongation plunges to the southeast (S64E, 52) in the plane of the mylonitic foliation. The orientation of recrystallized grains and the quartz C axis fabrics indicate east-over-west movement. The dominant deformation mechanism is dislocation creep resulting in the recrystallization of both quartz and feldspar. Based on these fabrics and the presence of garnet in some of the surrounding rocks Strehle and Stanley (1986) estimate that the fault zone developed under ductile conditions comparable, but perhaps somehat deeper than the Cobb Hill thrust zone. Again, Taconian deformation is inferred based on available isotopic age information and the similarity in orientation of fabrics across the massif. We will see that this same fabric orientation continues to the east into the pre-Silurian cover.

Stop 10 - Cota Brook Sequence (figs 14 and 15) (O'Loughlin and Lapp) - The Cota Brook sequence is a stratigraphically coherent section of metawackes and mafic schists of the Hoosac (Pinnacle) Formation. Outcrop is exposed in the stream valley and banks of Cota Brook at elevations of 1350 to 1390 feet above sea level. The Cota Brook sequence is bounded both upstream and downstream by zones of fissile schists with abundant graphite pods and layers. These zones are interpreted to be "pre-peak" or "syn-peak" metamorphic faults (fig. 15). Rock units surrounding these fault zones are lithologies in the Underhill Formation. Faults do not appear to be present within the sequence. A tight to isoclinal fold within the sequence is defined by mafic schists and changes in the dip of the dominant schistosity. The geochemical data and amphibole petrology will be discussed by Coish and Laird (refer to their respective papers in this guidebook). Lithic descriptions and estimated modes are given in Table 2.

Stop 11 - Fire Road Fault Zone (figs 14 and 16) (Lapp and Stanley) -

This group of outcrops shows the typically complex contact relationships of the area and is easily accessible from a forest service road 2 km southwest of Lincoln Gap (see location map). This area is distinctly marked by a 20 m x 1-2 m milky quartz vein (see outcrop sketch). Black carbonaceous schist of the Hazens Notch Formation (€Zhnc) occurs along the lower (western) part of this outcrop, while chloritoid-white mica Mt. Abraham Schist (€Za) occurs above. Outcrops of the latter unit are, here, anomalously rusty weathering and contain flattened garnets with chlorite pressure shadows and kyanite psuedomorphs composed of white-mica.

Contacts between the two schists are extremely sharp and highly "interfingered". One exposure shows folds in the rusty schist truncated by the carbonaceous schist along a pre- to syn-metamorphic fault. This contact was later folded and intruded by the large quartz vein. Reactivated faulting further displaced the schists to the west, over the quartz vein along a corrugated surface. The east-west trend of the hinge line of the corrugations is parallel to quartz lineations on the carbonaceous schist (€Zhnc), which is also found as lineated slivers within the quartz vein.

Table 2

Cota Brook Sequence

METAWACKES

In general, the metawackes are light tan to brownish in color and are less resistant to stream erosion than the mafic schists. Four lithic types have been distinguished.

lm laminated metawacke
well laminated or foliated, quartz rich, little plagioclase, white mica rich, with "quartzite" beds and layers of coarse biotite porphyroblasts.

m mica poor metawacke
quartz and feldspar rich, with coarse biotite porphyroblasts, less mica than other lithic types.

q massive "quartzite"
primarily quartz and feldspar, little or no mica.

gm grey green laminated metawacke
well laminated or foliated, grey green in color, with coarse biotite porphyroblasts.

Estimated Modes of Metawackes

Location	216	216	219	219Q
Sample	a	b		
Minerals				
Quartz	15	30	30	45
White Mica	60	55	10	-
Chlorite	7	5	tr	1
Plagioclase	4	2	45	45
Garnet	1	1	3	tr
Biotite	2	-	7	2
Graphite	-	5	-	-
Opaques	1	2	1	tr
Carbonate	-	-	-	tr
Epidote	-	-	-	tr

Modes are mineral percentages.
tr = trace percentage observed
dash (-) = mineral not observed

MAFIC SCHISTS

In general, the mafic schists are green to dark green in color and are more resistant to stream erosion than the metawackes. Geochemical work done on some of these mafic schists (Gavigan 1986) has identified them as Type A metabasalts (Coish and others 1985) which were formed during the initial stages of continental rifting prior to the formation of Iapetus (proto-Atlantic). The exposures of mafic schist are each relatively distinct in appearance; important characteristics are noted below by location (Loc.) number.

Loc. 216Z fine grained.
Loc. 217 coarse to very coarse grained, amphibole rich layer at upper contact.
Loc. 218A well laminated or foliated. Similar to mafic schist at Loc. 221.
Loc. 218B coarse grained, massive, grain size fines toward lower contact.
Loc. 220 coarse grained, with a 2-3 inch thick "quartzite" bed, very amphibole rich.
Loc. 221 very coarse grained, well laminated or foliated, very amphibole rich, with small pods of carbonate(?) material. Similar to mafic schist at Loc. 218A.
Loc. 222 "lumpy" texture, with small pods of carbonate(?) material.

Estimated Modes of Mafic Schists

Location	217	217	218A	218A	218A	218A	220
Sample	a	b	a	b	c	d	
Mineral							
Quartz	40	30	10	15	27	27	20
Plagioclase	3	-	10	35	1	28	5
Epidote	5	10	2	1	5	3	-
Garnet	-	-	tr	tr	-	tr	-
Amphibole	-	-	60	2	60	25	70
Chlorite	40	35	10	15	3	10	tr
Biotite	-	-	tr	-	-	-	-
Carbonate	10	25	5	30	3	2	-
Opaques	2	tr	3	tr	1	1	3

Modes are mineral percentages. tr = trace percentage observed
dash (-) = mineral not observed

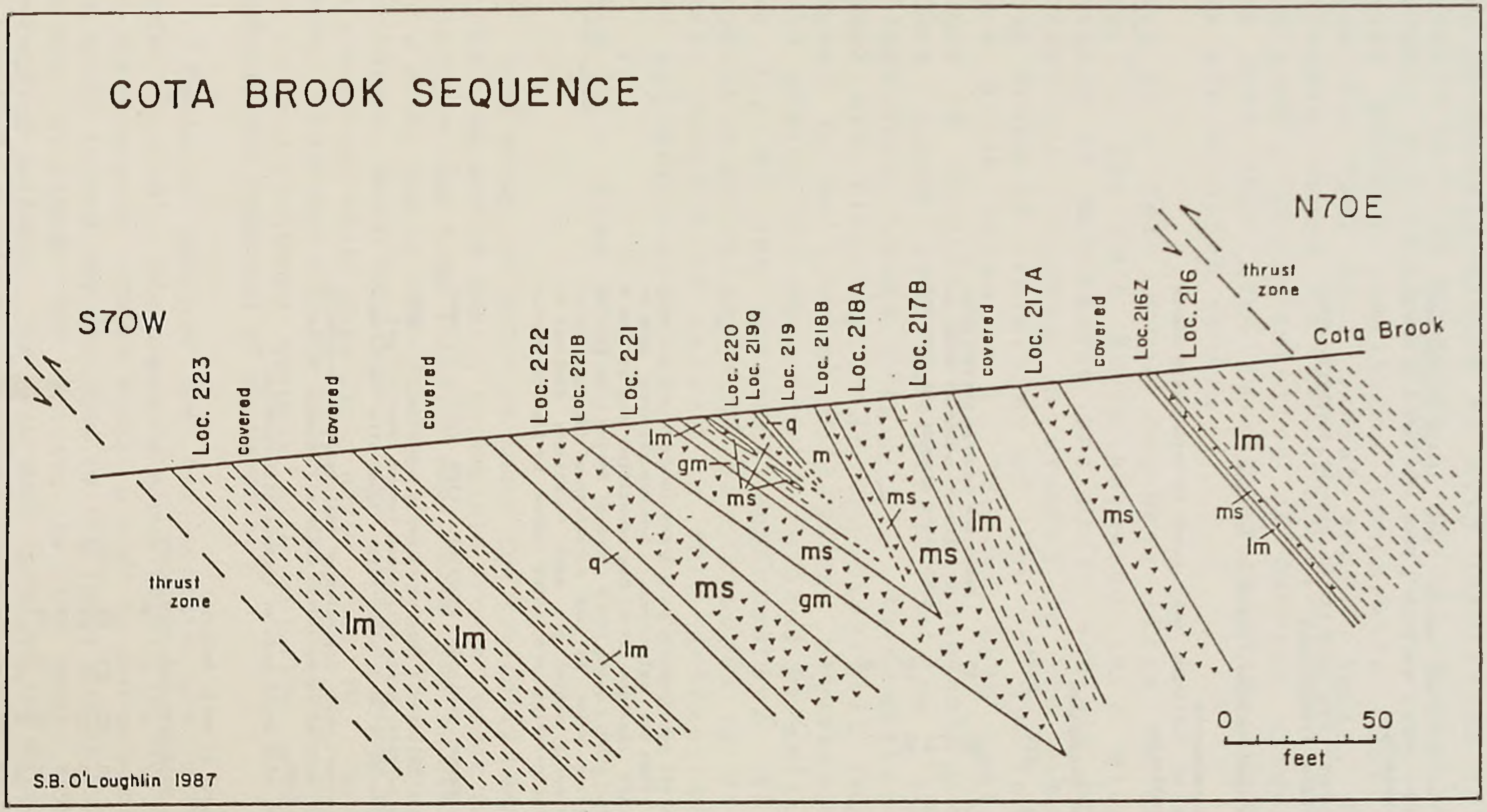


Figure 15

Cross section of the Cota Brook sequence looking north. The sequence consists of stratigraphically coherent metawackes and mafic schists of the Hoosac (Pinnacle) Formation which are bounded by thrust zones (pre-peak metamorphism, locations approximate) and surrounded by lithologies of the Underhill Formation (not shown). Contacts, where exposed, are sharp, generally parallel the foliation, and may be isoclinally folded. The synform (Fn+1 age) shown in the sequence is based upon changes in the orientation of the dominant schistosity (Sn). The geometry of the bounding fault zones is unclear. Three possibilities for this geometry are:
1) Both the eastern and western thrust zones are structurally upright and merge above and below the Cota Brook sequence to form a sliver of the Hoosac in the Underhill. 2) The western thrust zone is overturned and continuous with the eastern thrust zone. The thrust zone is isoclinally folded (Fn age) and closes above the Cota Brook sequence at the hinge of an antiform. An erosional window through the Underhill exposes the Hoosac. 3) The eastern thrust zone is overturned and continuous with the western thrust zone. The thrust zone is isoclinally folded (Fn age) and closes below the Cota Brook sequence at the hinge of a synform. A klippe of the Hoosac rests on the Underhill. In all three variations, the thrust movement sense is east over west for both the eastern and western thrust zones. Elevation of location (Loc.) 216 is 1390 feet above sea level; Loc. 223 is at 1350 feet. No vertical exaggeration. Lithic designators: lm - laminated metawacke, m - mica poor metawacke, q - massive "quartzite", gm - grey green metawacke, ms - mafic schist. For more information, see text.

THE FIRE ROAD FAULT ZONE

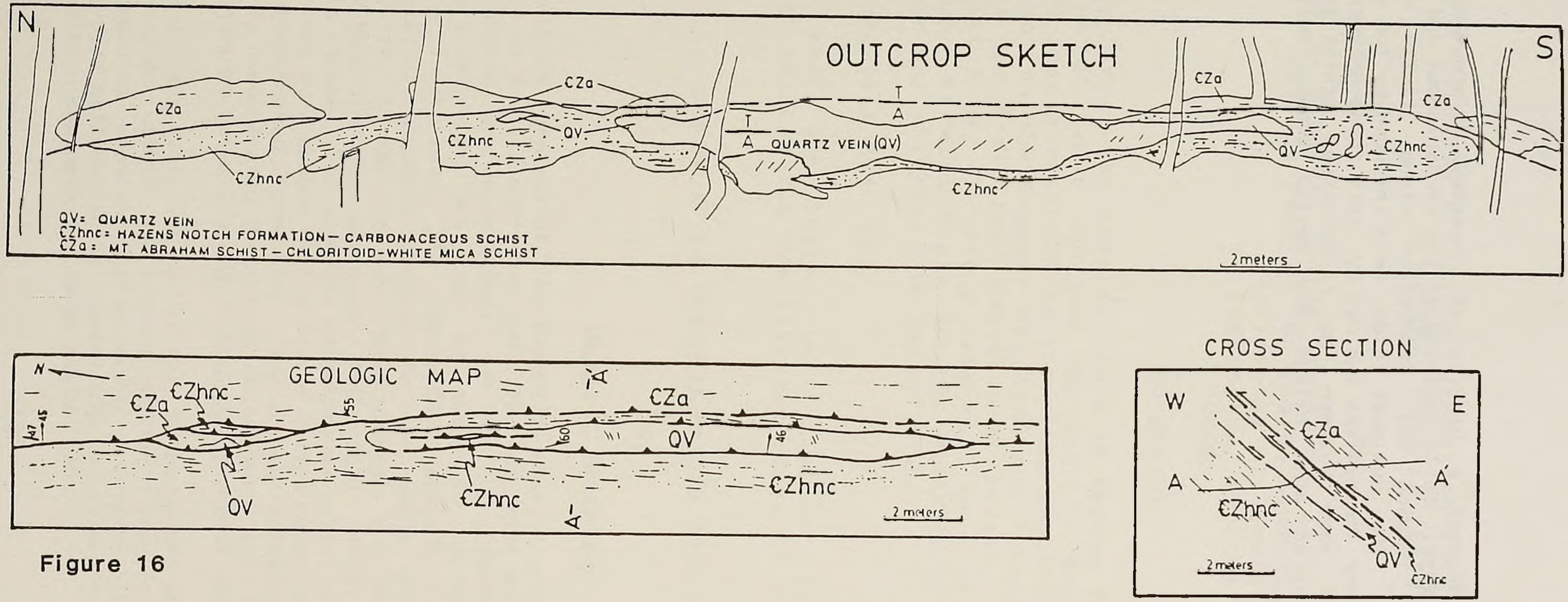


Figure 16

Outcrop sketch, geologic map and cross section of the Fire Road Fault Zone. Longitudinal profile view in outcrop sketch is looking east. Long, heavy line represents an interpreted syn- to post-metamorphic fault contact between black carbonaceous schist of Hazens Notch Formation (€Zhnc) and chloritoid-white mica Mt. Abraham Schist (€Za) (at this location, rusty weathering). T = toward; A = away. The large, milky quartz vein postdates the dominant folding but is cut by discontinous post-metamorphic faults which bear slivers of the carbonaceous schist. The moderately east-dipping, dominant schistosity and east-west trending mineral lineations are symbolized on the geologic map. Cross section shows interpretation of east over west sense of motion for this reactivated fault zone.

These observations suggest a history of repeated deformation and metamorphism at this fault zone. Early faulting of the carbonaceous and rusty schists was followed by folding and truncation along faults during kyanite-grade metamorphism. The dominant folds were intruded by the quartz vein, which was displaced over the schists and cut by late (post-metamorphic) faults. A similar fault zone between the carbonaceous schist and the Underhill Formation occurs in an area called the Nettle Brook Fault Zone (shown on the location map).

THE PRE-SILURIAN HINTERLAND ALONG THE VALLEYS OF THE WHITE AND MAD RIVERS, CENTRAL VERMONT

By

Rolfe S. Stanley, Thomas Armstrong, Jerome Kraus, Gregory Walsh, Jeffery Prewitt, Christine Kimball, and Athene Cua
University of Vermont
Burlington, Vermont
05401

ITINERARY

Meeting Place: The General Store in Warren Village just east of the junction of Route 100 and the Lincoln Gap Road. Warren can be reached from Northfield by traveling south on Route 12a to Roxbury. Turn west on the Roxbury Gap Road and follow the signs on the west side of the Northfield Mountain to Warren. For those that happen to arrive late, the field trip will be along Route 100 for most of the day. We will begin to the south near Hancock and gradually work our way northward to Waitsfield.

Departure time 8:30 AM.

Please read the introductory part of this paper entitled "A Transect through the Pre-Silurian rocks of Central Vermont".

GEOLOGY OF THE GRANVILLE-HANCOCK AREA, CENTRAL VERMONT

by

Thomas R. Armstrong

The bedrock geology of the Granville-Hancock area was first mapped by Osberg (1952) at the scale of 1:62,500. The contacts between the major units were interpreted as depositional due to the intercalation of lithologies along contact zones, and the lack of sharp boundaries between many formations. Osberg interpreted the eastwardly dipping stratigraphy as a coherent depositional sequence younging to the east, consisting of the Monastery, Granville, Pinney Hollow, Ottauquechee, and Stowe Formations. Several of these formations contain minor lithologies which are similar or identical across strike. For example, carbonaceous schist which is shown on the Vermont state map as a member of the Pinney Hollow Formation, was mapped by Osberg as the Ottauquechee Formation. The present detailed mapping of this region has clearly demonstrated the similarity of this unit to the black schist of the Ottauquechee Formation and is thus in agreement with Osberg's interpretation.

Based on the distribution of the Ottauquechee and Stowe Formations, and the greenstones within the Pinney Hollow Formation, Osberg interpreted the structure as consisting of a series of west-verging, overturned folds. These folds were believed to be responsible for deforming the stratigraphy into the map pattern dominated by large scale folds.

New mapping at the scale of 1:12,000 has demonstrated the tectonic nature of contacts between the Mt. Abraham, Hazens Notch, Pinney Hollow, Ottauquechee, and Stowe Formations. Many of the gradational contacts mapped by Osberg are now interpreted as early-metamorphic thrust faults (table 1). These thrusts always appear to juxtapose discreet lithologies of different formations. Fault zone fabrics are absent due to annealing during Taconian peak metamorphism. The tectonic nature of these surfaces, however, can be demonstrated by the lithic truncation of various units along these contacts. The greenstones (The Hancock and Brackett Members of Osberg) are shown truncated along early-metamorphic contacts in numerous localities. Without detailed mapping of this geometry, it would be impossible to distinguish tectonic contacts from depositional boundaries.

The silvery-green quartz-sericite-chlorite-albite schist and albite-epidote-chlorite-calcite greenstone of the Pinney Hollow Formation are demonstrably overlain by the carbonaceous albitic schists of the Hazens Notch Formation (more correctly called "Granville") which in turn, are overlain by the graphitic black phyllite, black quartzite, and quartzose schist of the Ottauquechee Formation. Both of these contacts are early-metamorphic thrusts, shown by the map scale truncation of upper and lower plate lithologies. In some instances, the Ottauquechee thrust slice lies directly above the structurally lowest Pinney Hollow Formation, with the Hazens Notch thrust slice absent. This relationship suggests that after emplacement the carbonaceous schist of the Hazens Notch (Granville) and Pinney Hollow slices were deformed prior to the emplacement of the Ottauquechee slice. This geometry occurs in both the eastern and western parts of this area, and thus rules out a simple stratigraphic climb of the Ottauquechee slice along a ramp through the Hazens Notch and Pinney Hollow slices. Serpentinite bodies, found within the carbonaceous Hazens Notch and Ottauquechee Formations, were emplaced either during this phase of thrusting or an earlier phase associated with the development of an accretionary wedge (Stanley and Ratcliffe, 1985). Many of these bodies were also incorporated in subsequent Taconian syn-metamorphic thrusts.

Following the development of the early-metamorphic thrusts and synchronous with Taconian metamorphism was the formation of the first phase of syn-metamorphic folds (Fn-1) and associated axial plane schistosity (Sn-1). This phase is preserved exclusively as relict fold hinges and schistosity, transposed and sheared along a second syn-metamorphic schistosity (Sn). Infrequently, Fn-1 are exposed in structures refolded by the second phase of syn-metamorphic folds (Fn) and associated axial plane schistosity (Sn) (see Stop 17). The transposition and refolding of Fn-1 and Sn-1 is frequently seen at the microscopic scale.

The second phase syn-metamorphic schistosity (Sn) is the dominant fabric within this region, with an average trend of N05E, 65SE. This schistosity is axial planar to the second phase of syn-metamorphic folds (Fn). The hinges of Fn within the schists are oriented parallel, and sub parallel, to the dominant mineral lineation (Ln) which rakes steeply down the plane of Sn (about 85 from both the south and north strike lines). Some hinges are shallower, and at a considerable angle to the mineral lineation. The arcuate nature of these Fn hinges suggests that Fn are sheath folds, although only a few demonstrable closures can be seen in the field. Hinge orientations within the more competent greenstones and quartzites are consistently at high angles to the mineral lineation. Fn folds developed within these units were more resistent to rotation during this high shear strain event, and thus never formed into sheath configuration, but rather into inclined, isoclinal folds (Armstrong, 1987; Kraus, 1987). This phase of deformation is primarily responsible for the map pattern geometry which shows the early-metamorphic thrusts being deformed into isolated, infolded klippe. This interpretation is very similar to the infold model of Osberg (1952) except that the contacts are now interpreted as faults.

The first phase of syn-metamorphic thrust faults (STn) developed coaxial to Sn and sheared out both Fn-1 and Fn structures, producing a pervasive mylonitic fabric. Kinematic indicators, such as asymmetric porphyroclasts, pressure shadows, rotated porphyroblasts, and C-S fabric consistently display an east-over-west sense of motion. These faults displace the sheath fold klippe, juxtaposing both similar and different lithologies. The magnitude of displacement along STn was much less than that along the early-metamorphic faults which always juxtapose different lithologies. STn faults are further demonstrated by map scale lithic truncations, many of which include the greenstones within the Pinney Hollow and Stowe Formations. Many of these faults also contain tectonic slivers of rock-types foreign to either of the adjacent plates.

A third phase of syn-metamorphic deformation occurred coaxial to the second phase, and includes east-over-west asymmetric folds, a local axial plane schistosity, and several coplanar thrust faults. A progression in the development of this phase of schistosity (Sn+1) can be recognized where the associated asymmetric folds (Fn+1) are pervasive. In these locations Sn+1 is the dominant fabric, shearing out Fn folds and folding Sn into Fn+1. In a few locations, Sn+1 strain produces a second phase of syn-metamorphic faults (STn+1). Kinematic indicators, identical to those found along STn, consistently show east-over-west motion along STn+1.

This phase of deformation is considered a continuum of the second phase, to which it is coplanar. The development of Fn+1, (and possibly Fn) are most likely related to strain hardening during the development of Sn and Sn+1. STn and STn+1 may represent periods of strain softening, the result of mineralogical alteration by a fluid phase during contemporaneous retrograde metamorphism.

Late stage, late-metamorphic upright, open folds (Fn+2) deform the syn-metamorphic fabrics and may be related to either late Taconian or Acadian deformation. Late and post-metamorphic faults, producing pervasive sheared, brittle fabric, probably developed during, and subsequent to Fn+2. Sense of motion along these faults has not been determined. It appears, however, that they occur as reactivated faults along older syn-metamorphic thrusts.

A current working hypothesis suggests that the Pinney Hollow, Stowe, and Mt. Abraham Formations, along with the white albitic schist member of the Hazens Notch Formation, developed as a continuous clastic sequence within a rift basin during the opening of Iapetus (fig. 6). The Ottauquechee Formation, separating the westerly Pinney Hollow Formation from the easterly Stowe Formation, is interpreted as a large pre- or early-metamorphic klippe structurally overlying the continuous former units. The Stowe Formation is therefore interpreted as an eastward continuation of the Pinney Hollow Formation. This is based upon the similarity between the quartz-sericite-chlorite-albite schists and albite-epidote-chlorite-calcite meta-igneous units in both formations. This is also in agreement with the rare earth and trace element geochemical data of Coish and others (1985, 1986) which places the Stowe tholeitic greenstones in an island arc/ MORB environment, and the Pinney Hollow alkaline basalts within an intraplate setting. The present interpretation would depict the Stowe greenstones forming as mafic flows and tuffs within highly attenuated continental crust (beta ~ 5.0), and the Pinney Hollow meta-igneous units within thicker, more westerly, continental crust (beta~ 2.5) where assimilation would occur more readily.

The present interpretation depicts the Ottauquechee Formation as forming within an accretionary wedge environment. The numerous quartzose, albitic, and sandy lithologies could be incorporated following subduction of various clastic sequences, and their subsequent accretion into the wedge. Slivers of ocean crust (altered to serpentinite) would also be incorporated into the wedge. The black phyllite of the Ottauquechee Formation, comprising the majority of the wedge, would be the actual correlative sequence for the Taconic lithologies. The black quartzites may have been tectonically thickened during their accretion, represent part of the clastic sequence incorporated into the wedge, or a deposit shed from an entirely different source, to the north or possibly the east.

The carbonaceous schists of the Hazens Notch Formation would represent a sequence derived from the accretionary prism. Much of the sandy albitic schist within the Ottauquechee Formation is identical to the albitic matrix within the Hazens Notch carbonaceous units. These units, therefore, may be an amalgamation of the black phyllite and the incorporated sandy schists, derived during subaerial exposure of the wedge. This interpretation is in close agreement with that described by Stanley and Ratcliffe (1985). The Hazens Notch white albitic schist would be more properly depicted as an intermediate facies between the albitic schists and metawacke of the Hoosac Formation and the more easterly lithologies of the Pinney Hollow, Mt. Abraham, and Stowe Formations.

STOP 12 - DEPOSITIONAL CONTACT BETWEEN THE SCHIST AND GREENSTONE MEMBERS OF THE PINNEY HOLLOW FORMATION (Armstrong)- This outcrop affords one of the finest examples of the contact relationship between the quartz-chlorite-muscovite-albite schist and albite-epidote-chlorite-calcite greenstone members of the Pinney Hollow Formation. Due to the lack of a pervasive, fault-related schistosity, and the lack of lithic truncations, this contact is interpreted as depositional. Note the zone of alteration of the schist along the contact. This effect is most likely a result of fluid alteration during peak metamorphism, rather than a contact metamorphic effect during deposition of the meta-igneous greenstone. Since many of the greenstones within this area display planar compositional layers and laminae of albite-quartz and epidote-chlorite, their precursor may be ash, or water-lain tuffs. The contact between the schist and greenstone would therefore represent a conformable, contemporaneous sequence, possibly deposited in a rift basin. Greenstone is present at both ends of the outcrop, representing the limbs of a large syn-metamorphic fold (Fn) associated with the axial planar dominant schistosity (Sn) seen at this exposure within the schist.

Geochemical anlysis of similar greenstones within the Pinney Hollow Formation by Coish, et. al (1985) yielded both alkaline and tholeitic compositions, produced within intra-plate and MORB environments respectively.

A current working hypotesis places the Pinney Hollow Formation within a rift basin, distal from its western cratonic source region, upon very thin continental crust (Armstrong, 1987). Rift related igneous rocks (greenstones) of both tholeitic and alkaline affinity could be produced in such an environment, where assimilation with continental crust would be minimum.

STOP 13 - Thatcher Brook Thrust (figs. 17, 18) (Armstrong and Kimball) - A 100 ft. thick sequence of carbonaceous albitic schist of the Hazens Notch Formation is tectonically juxtaposed between albite-chlorite-quartz-sericite schist of the Pinney Hollow Formation (fig. 17).

The eastern contact is a syn-metamorphic fault, displaying a well developed mylonitic fabric, and the truncation of schist sequences in the Pinney Hollow Formation. Asymmetric porphyroclasts, oriented within the dominant syn-metamorphic schistosity (Sn), display east-over-west motion. Microfabric analysis of C-S fabric, rotated porphyroblasts, asymmetric pressure shadows and porphyroclasts, and rotated schistosity, within syn-metamorphic fault zones, consistently display the same sense of motion, and are believed to be localized zones of intense strain development during Sn deformation. The fault zone fabric, at this locality, is oriented N07E, 65SE; similar to the regional Sn. The fault zone is characterized by the progressive intensification of Sn within the Pinney Hollow schist toward the contact, and by a tan weathering mylonitic quartzose schist within the fault zone. This fault (the Thatcher Brook Thrust) can be traced both north and south along strike where similar fault fabrics and tectonic slivers of foreign rock-types can be found.

The Thatcher Brook Thrust is deformed by a secondary phase of syn-metamorphic folds (Fn+1), the largest order of which consistently display counter-clockwise rotation. The average orientation of Fn+1 is ~ N24E, 30° (fig.18). Axial plane orientations for this phase are similar to the undeformed orientations of Sn within this locality. This secondary fold phase is therefore believed to be a product of one syn-metamorphic deformational event, which produced the dominant schistosity and the primary phase of folds (Fn). Several syn-metamorphic faults within this region truncate Fn+1, having a fault zone fabric oriented similar to the Fn+1 axial plane schistosity (Sn+1). The pervasiveness of Sn+1 increases toward the fault zone where only limbs of earlier Fn+1 can be recognized within a very planar fabric. This progression of fabric intensity demonstrates the relationship of Sn+1 and many syn-metamorphic faults within this region. In many areas Sn and Sn+1 form a composite schistosity, with Sn+1 actually being the continuation of Sn deformation.

The western contact is not marked by a strong fault fabric and does not exhibit any mylonitization. Some small Fn folds within thin quartzites of the Hazens Notch Formation are sheared out along Sn; this is a common feature within these highly strained rocks, but is not indicative of a major syn-metamorphic fault zone. The contact, interpreted as pre-metamorphic, has been highly altered during subsequent peak Taconian metamorphism. Kinematic indicators are absent. Fault motion is generally interpreted as being east to

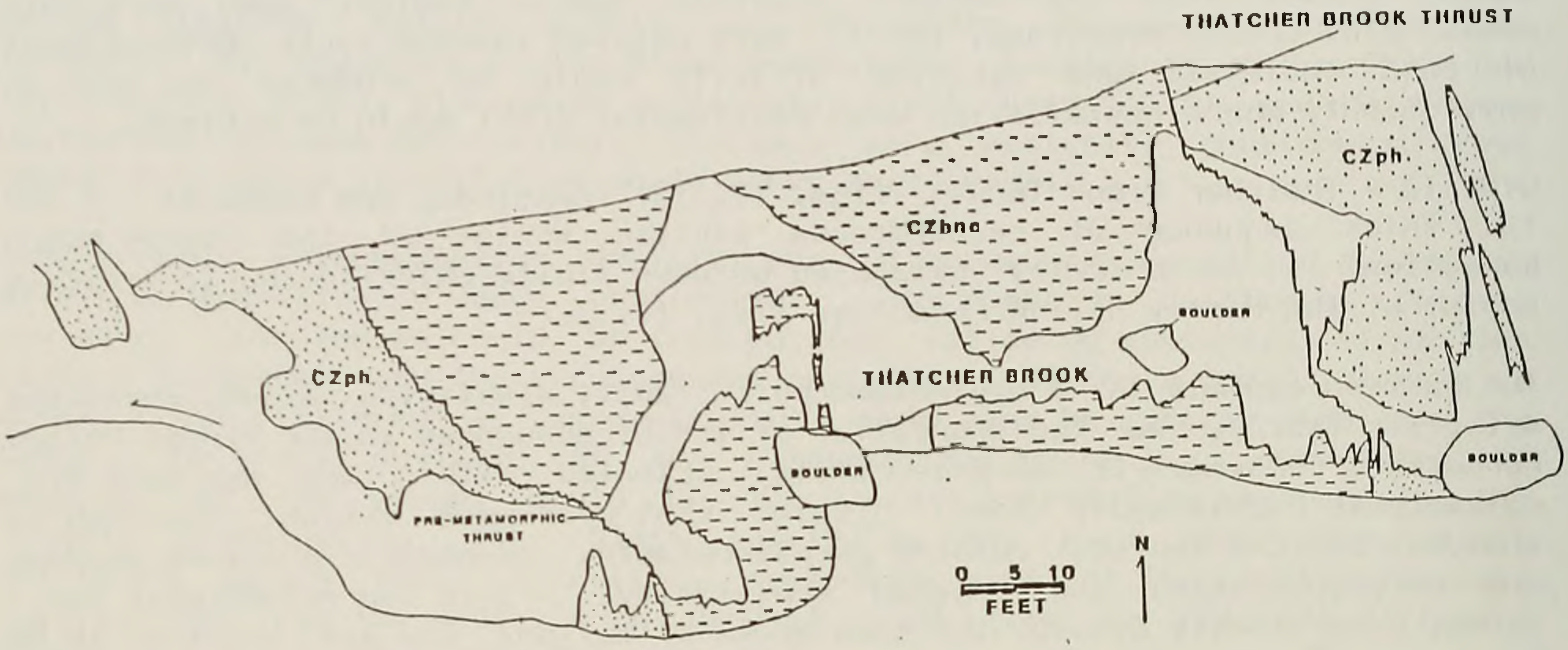
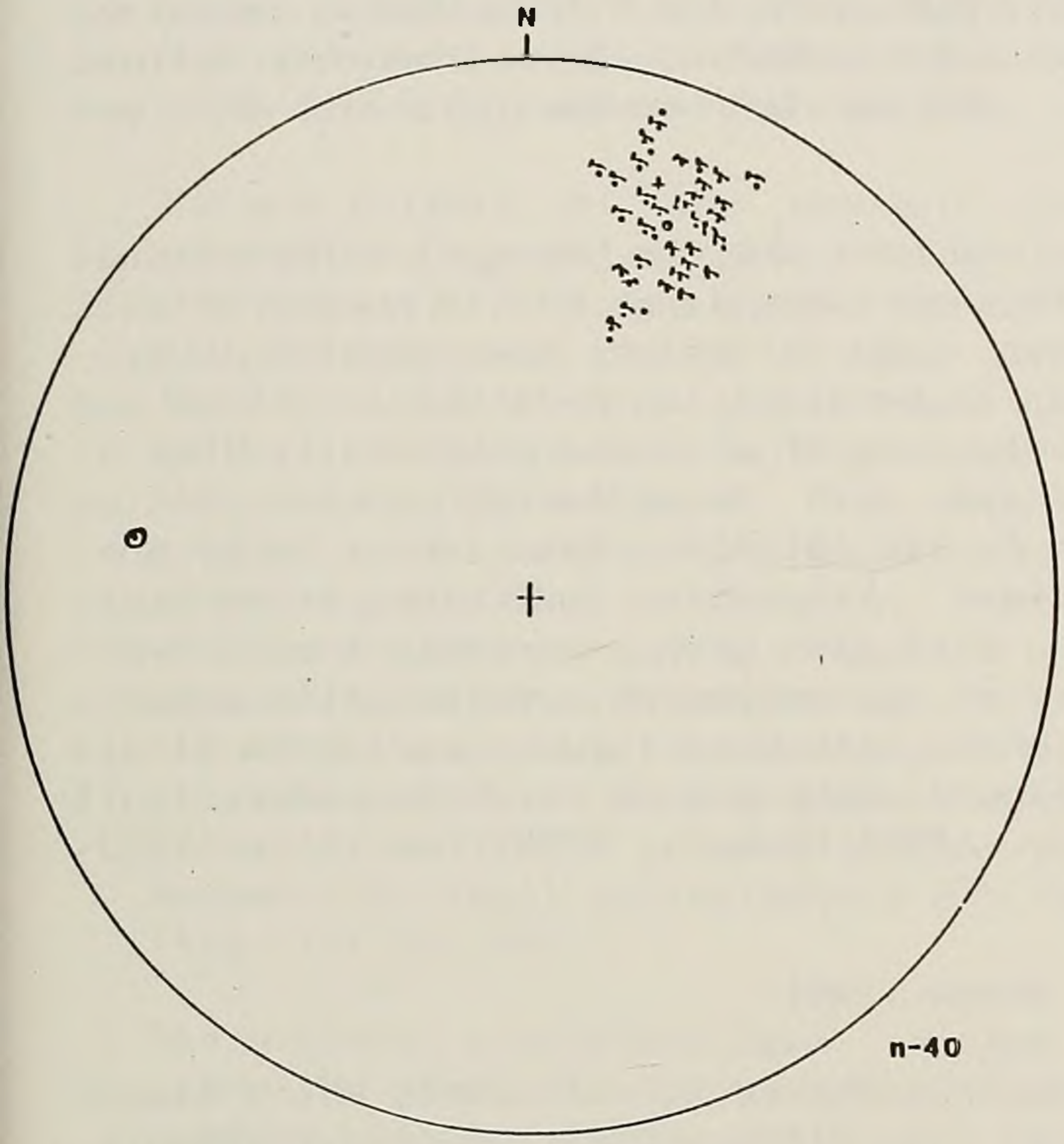


Figure 17

Detailed geologic map of the Thatcher Brook Thrust, a syn-metamorphic fault (eastern contact), and an earlier pre-metamorphic fault (western contact). CZph - quartz-chlorite-sericite-albite schist of the Pinney Hollow Formation; CZhnc - graphitic quartz-sericite-albite-chlorite schist of the Hazens Notch Formation. The pre-metamorphic contact does not have a fault-related fabric due to annealing during subsequent Taconian metamorphism. This contact is interpreted as a thrust fault from relationships found along this same contact sequence, where the truncation of several greenstones of the Pinney Hollow Formation occur. The syn-metamorphic thrust displays a well developed mylonitic fault zone fabric. Asymmetric porphyroclasts along this contact show east-over-west motion. A small sequence of Pinney Hollow schist is truncated along this contact, evidenced by the mergence of a small greenstone body (within the Pinney Hollow) and the fault contact towards the southern end of the outcrop. Both faults are folded by the third phase of syn-metamorphic folds (Fn+1), pervasive throughout this exposure.

THATCHER BROOK
Fn 1 FOLDS



THATCHER BROOK
SN (POLES TO SCHISTOSITY)

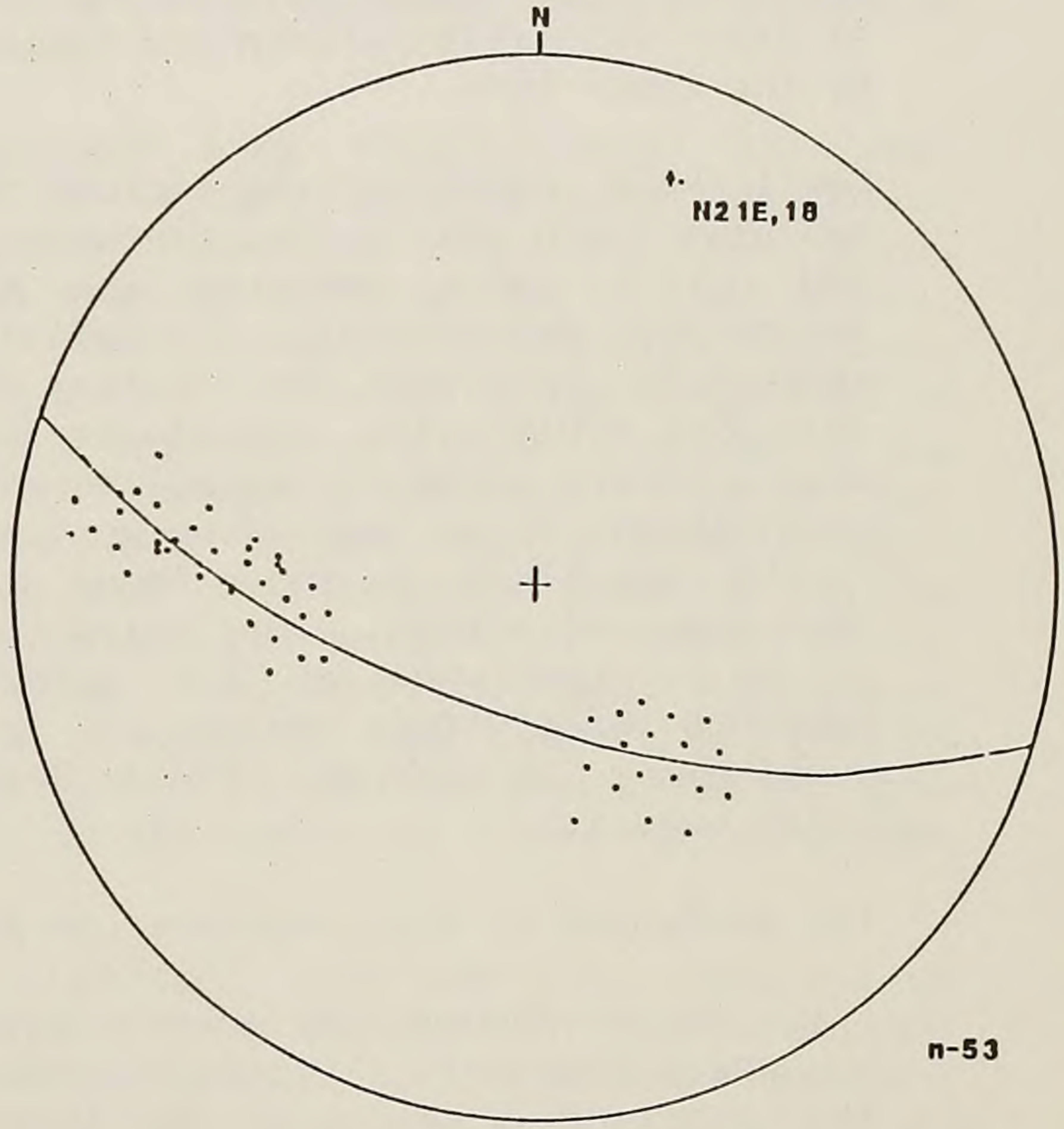


Figure 18

Equal area net projection of poles to the dominant schistosity (Sn) and Fn+1 folds at Thatcher Brook. Sn is folded by Fn+1 at this locality, forming a girdle around a best-fit great circle the pole to which is oriented N21E, 18. This orientation is shown as the cross within the Fn+1 fold hinge net. The open circle denotes the average orientation of Fn+1 at this locality. Arrows indicate the sense of asymmetry. The circle with the dot indicates the average orientation of undeformed Fn+1 axial planes (Sn+1) oriented N06E, 67 SE.

west (Stanley and Ratcliffe, 1985). Distinctive quartzite beds within the carbonaceous schist are parallel to Sn. The distance between these quartzites and the pre-metamorphic contact is variable along strike, demonstrating the early tectonic nature of this contact. The tectonic character of this contact is also recognized to the north and south where lithic truncations, primarily of greenstones within the Pinney Hollow Formation, are present without fault zone fabric. The associated fault zone fabric of pre-metamorphic faults was obliterated during peak Taconian metamorphism, coeval with the development of the dominant schistosity (Sn). The carbonaceous schist of the Hazens Notch Formation is interpreted as structurally overlying the Pinney Hollow Formation above this pre-metamorphic fault. This interpretation stems from observations at other localities within the region. The western contact is also deformed by the later Fn+1.

Small shear bands cutting across Sn and Fn+1 are the youngest deformational features found within this exposure. In some cases large albite porphyroblasts and pyrite cubes overgrow the shear band foliation, demonstrating a syn-metamorphic development. The foliation is parallel to Fn+1 axial planes and appears to represent the initial development of a second schistosity (Sn+1). This Sn+1 schistosity is exclusively found as a local fabric in association with a second phase of syn-metamorphic faults (STn+1). Rotation of Sn by Sn+1 consistently shows east-over-west motion. Kinematic indicators along STn+1 faults (identical to those found along STn) also give a consistent east-over-west sense of motion. This third phase of syn-metamorphic deformation appears to be a continuum of the second phase with Fn+1 folds occuring in strain hardened zones within regions of continued compression and secondary fault development (Armstrong, 1987; Kraus, 1987; Kimball, 1987; Ratcliffe, 1987, pers. comm.).

The evolution of this exposure can be summarized;

1. The development of pre-retrograde metamorphic thrust slices which always juxtapose discreet lithologies, usually from different tectonic environments. The carbonaceous schist of the Hazens Notch Formation thrust east to west over the Pinney Hollow Formation. The western contact was formed during this event.

2. The onset of Taconian retrograde metamorphism with associated deformational events producing an early schistosity (Sn-1) and associated folds (Fn-1) present as relict fold hinges and transposed foliation within the dominant schistosity (Sn) and associated sheath folds.

3. Development of the first phase of syn-metamorphic faults along zones of intense shearing within Sn. Movement was consistently east-over-west, incurring the juxtaposition of both discreet and similar lithologies. This phase produced the eastern contact which truncated the earlier pre-metamorphic structure

4. Continued deformation producing secondary folds (Fn+1), and associated Sn+1 in many localities where deformation was intense. This event is responsible for the folding of both contacts at this locality.

5. Development of the second phase of syn-metamorphic faults along zones of intense Sn+1 development. Movement was consistently east-over-west in a

manner similar to the first phase of faulting. Although not present at this locality, this event is observed throughout the region.

STOP 14 - SYN-METAMORPHIC THRUST BETWEEN THE PINNEY HOLLOW SCHIST AND GREENSTONE MEMBERS (fig. 19) (Armstrong) - Two greenstone bodies are separated by a thin band of quartz-chlorite-muscovite-albite schist (fig. 19). The schist displays a well developed fault zone schistosity and has been altered to a tan, quartz-feldspar mylonite in many localities. This schistose mylonite is diagnostic of many syn-metamorphic fault zones throughout this region. Kinematic indicators, such as asymmetric porphyroclasts, C-S fabric, and rotated porphyroblasts, consistently show east over west displacement along these syn-metamorphic thrusts.

The map pattern of the eastern greenstone body displays east over west asymmetric, syn-metamorphic folds (Fn) which have an associated axial plane schistosity. Fn folds within the schists of the study area are sheath-like, with hinges that plunge at steep angles down the plane of the Sn schistosity. The folds within the greenstones vary from reclined to inclined, isoclinal folds. Variation in fold geometry is directly related to the competency of the lithologies involved. The competent greenstones are more resistent to the simple shear mechanism responsible for sheath development in the pelitic assemblages within this region. Fn fold axes within the greenstones have undergone less physical rotation and are not oriented parallel to the elongation mineral lineation as are the hinges of the pelitic sheath folds. The east over west asymmetry of the minor folds indicates that the eastern greenstone body is on the western limb of a synform. The closure of this Fn fold can be traced toward the northwest where it is truncated along the syn-metamorphic fault approximately 275 feet to the north of Allbee Brook bridge (fig. 19; pt. A).

The western greenstone body is seen in contact with the fault zone schist along the road, directly south of the bridge (fig. 19; pt. B). The greenstone appears to be continuous, and has been mapped in a large isoclinal fold directly west of this locality. The greenstone does not appear to be truncated, but only mildly deformed by the adjacent thrust fault.

The two greenstones are separated by a thin band of schist that displays a pervasive fault zone fabric. A quartzose, tan weathering mylonitic schist is found within this zone at many of the schist outcrops (pt. C, for example). Microfabric analyses of C-S fabric, and asymmetric porphroclasts and pressure shadows, along syn-metamorphic faults consistently show east to west displacement. The displacement along the syn-metamorphic faults must be much less than that along the pre-metamorphic thrusts since they juxtapose similar, as well as discreet lithologies.

The fault zone schist is continuous over the entire length of this suite of outcrops, and can be traced to the north where the western greenstone is abruptly truncated (~2500 ft north), and on into Granville Gulf.

A tan weathering schist is also found immediately southeast of the bridge along the folded eastern edge of the eastern greenstone (fig. 19; pt. D). This schist, although not a fault zone, does display a pervasive fabric which cross-cuts the Fn fold geometry. This fabric is most likely the result of large localized strain, which would likely occur along this contrast in

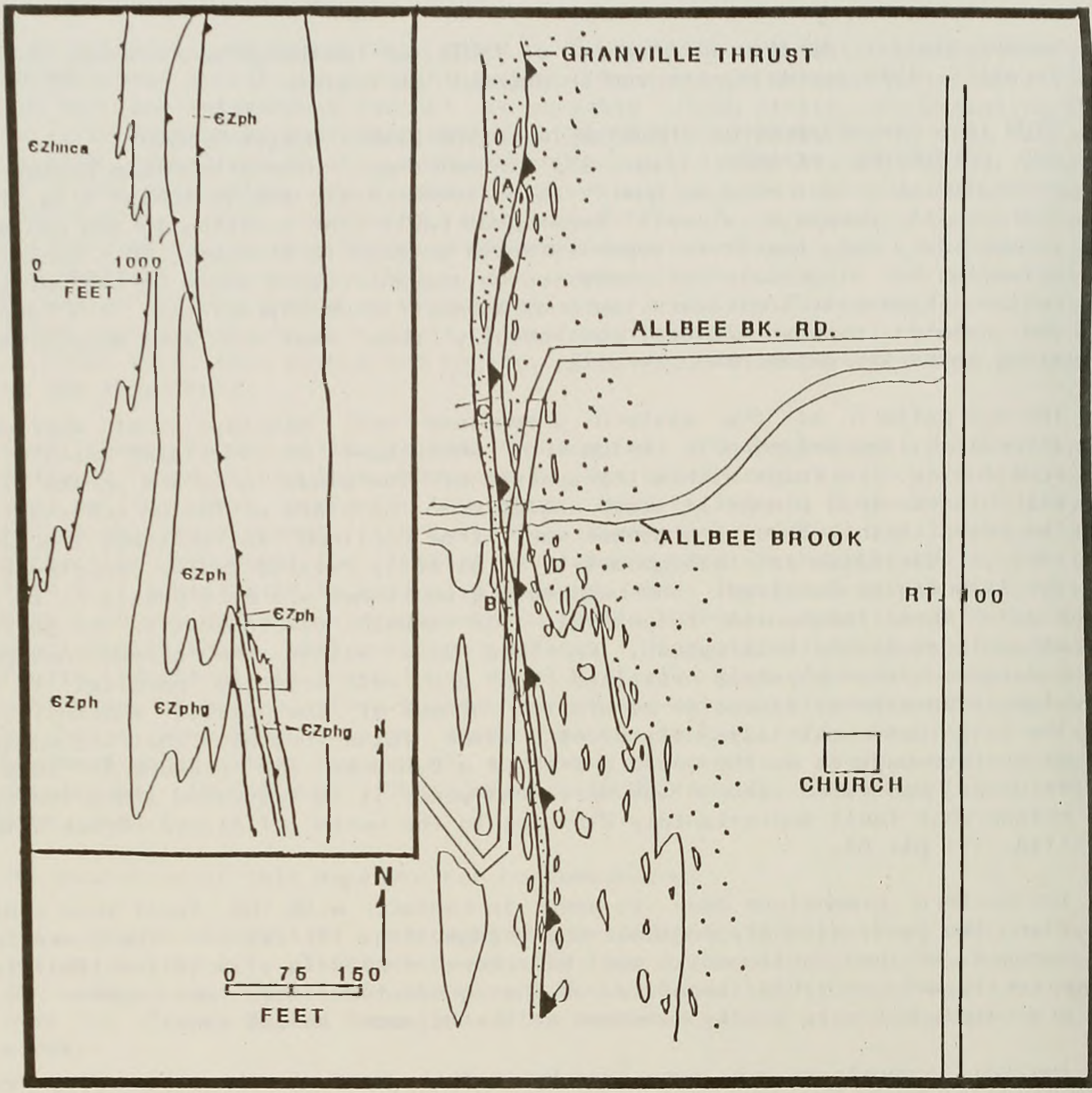


Figure 19

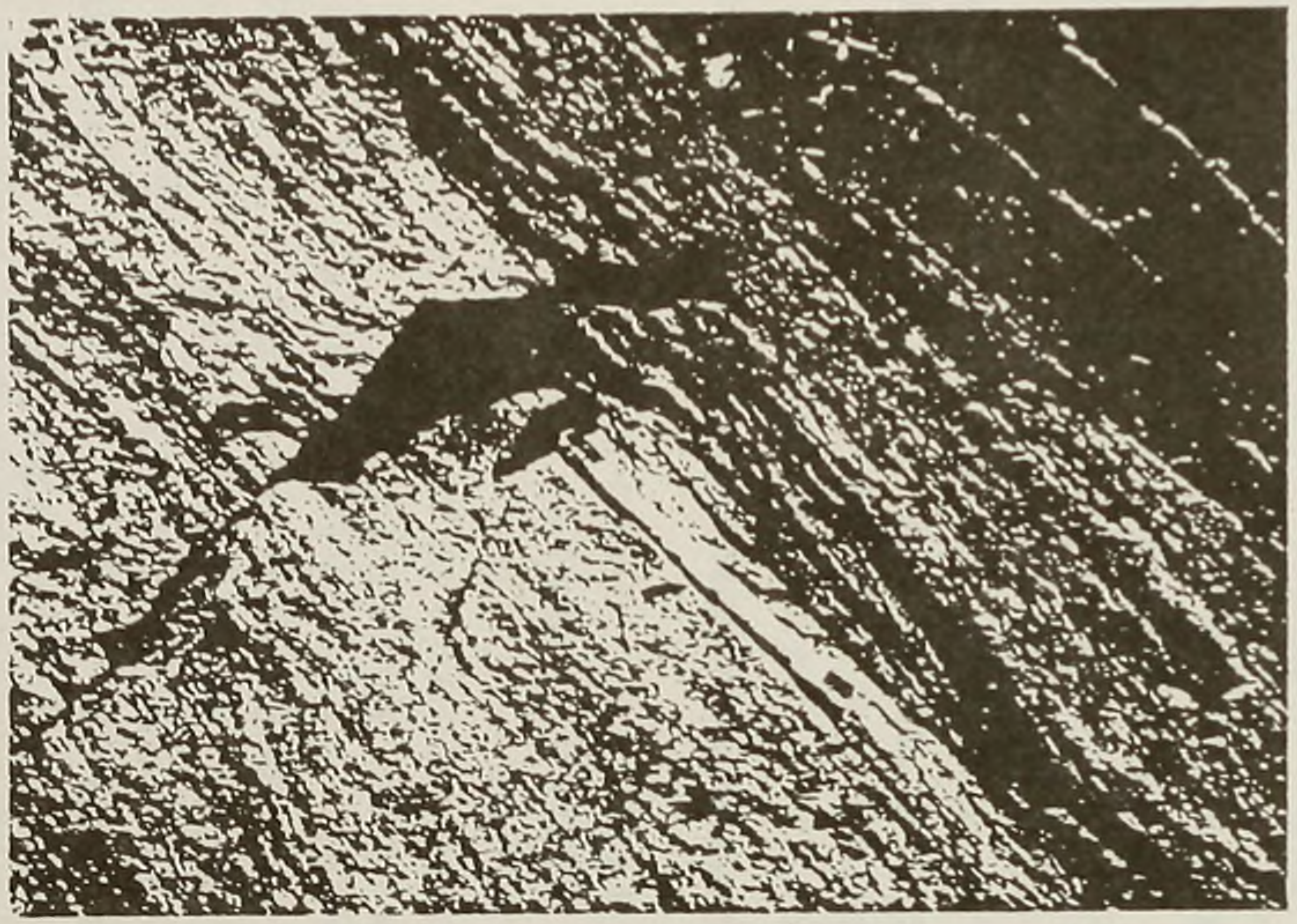
Detailed and regional geologic maps of the Granville Thrust, located in Lower Granville, central Vermont. The Pinney Hollow sliver-green quartz-chlorite-sericite-albite schist is stippled in the detailed map. The Pinney Hollow albite-epidote-chlorite-calcite greenstones are unpatterned. The eastern greenstone body is separated from the larger western greenstone by the fault zone schist contained within the Granville thrust. The map pattern east-over-west folds within the eastern greenstone plunge to the north, and are parasitic folds along the eastern limb of a large antiform. The folded greenstone is truncated along the fault over its entire length and is terminated to the north at pt. A. The western greenstone body is also truncated, to the north, along this fault as seen in the regional inset.

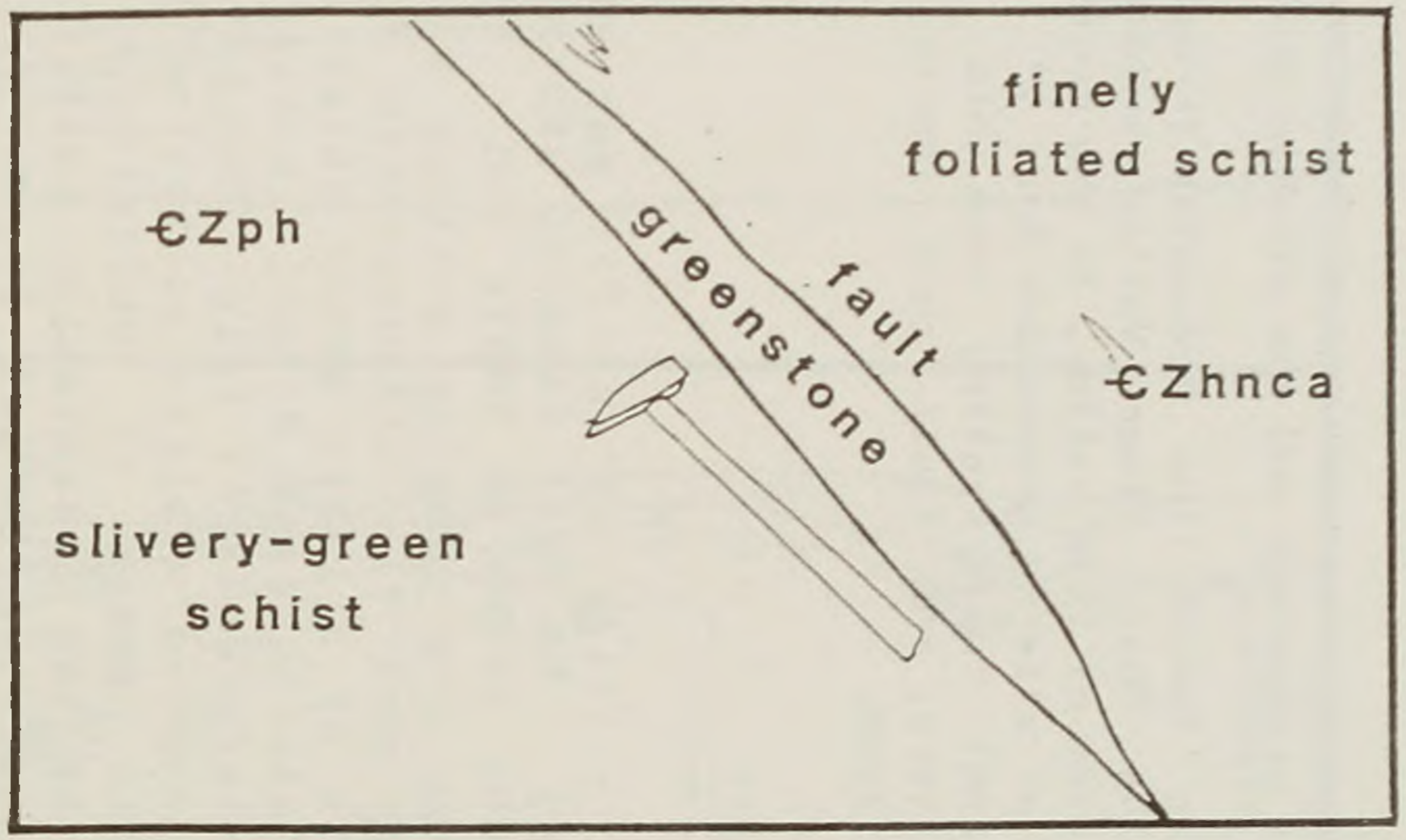
competency between the resistant greenstone and the pliable schist.

This large scale syn-metamorphic thrust (the Granville Thrust Zone) is the primary feature that controls the White River Valley and Granville Gulf. Ongoing work along the eastern side of this valley, to the south near Hancock, has demonstrated the truncation of a large greenstone body (~ 3 km long x 300 m wide) along an ancillary thrust fault rooting from this zone. Numerous lithic truncations, tectonic slivers, and fault zone fabrics may be found along the entire length of this zone.

Stop 15 - Granville Gulf - (fig. 20 and 21; map will be available at conference) - (Stanley) - BE VERY CAREFUL AND STAY OFF THE ROAD BECAUSE THIS IS A DANGEROUS CURVE ON ROUTE 100. This stop is very important because it shows the complex geology along the boundary between the Pinney Hollow and the Ottauquechee Formations. Here the Pinney Hollow consists of silvery green chlorite - muscovite - albite - quartz schist interlayered depositionally with two belts of mafic schist. The eastern contact (fig. 20) is a synmetamorphic fault in which the carbonaceous albitic schist of the Hazens Notch (CZhca) truncates a thin lense of chlorite-rich schist with brown-weathered material on the schistosity (mafic schist ?). The schistosity in the carbonaceous schist is so well developed and closely spaced that it weathers very readily. In thin section the foliation wraps around albite porphyroblasts. This "papery-schist" fabric is characteristic of late metamorphic or post metamorphic faults and has been observed at several localities along the western boundary of the Ottauquechee Formation. Note that the foliation in the carbonaceous schist next to the slivery green schist of the Pinney Hollow is more widely spaced and the rock is more coherent. This contact is a synmetamorphic fault because the albite porphyroblasts grew across the foliation. Thus this zone is one that has a compound history (table 1) - an earlier synmetamorphic fault that was reactivated as a late or postmetamorphic fault.

Continuing eastward along the outcrop you will find a tan - weathering schist with brown spots. This rock is found along several belts in this zone. In thin section is consists of fine-grained, well oriented muscovite and quartz. In sections cut parallel to the lineation the foliation is planar. Small grains of muscovite and quartz, which are oriented at a small angle to the dominant foliation, form a C-S fabric and indicate east-over-west movement. In sections cut perpendicular to the lineation the foliation anastomoses. I consider these rocks to be mylonitic schists which developed from the Pinney Hollow schist along synmetamorphic faults. Continuing eastward you will find another mafic schist, then another belt of carbonaceous schist which passes into black carbonaceous schist typical of the Ottauquechee Formation. The fabric throughout this belt is well represented by figure 21.





<u>The Ottauquechee thrust fault zone at Granville Gulf, central Vermont</u> A close-up photograph of the fault contact showing a thin greenstone lense (above the hammer) of the Pinney Hollow Formation (lower left) in contact with the well foliated, rusty-weathering schist (CZhnca) of the Hazens Notch Formation. Greenstones are common in the Pinney Hollow Formation but are absent in the carbonaceous, albitic schist (CZhnca) of the Hazens Notch Formation.

As you can see on the 1:12,000 scale geologic map there are 3 mappable greenstone boies in the Granville Notch area. Laird (1977) and Laird and Albee (1981) found barroisitic amphibolites with actinolitic rims in these greenstones and report $^{40}Ar/^{39}Ar$ ages ranging from 471 m.y. to 448 m.y.. The analyzed grains occur near quartz veins and in the matrix of the greenstone beds. The quartz veins throughout the Pinney Hollow are deformed and flattened parallel to the dominant foliation Sn, but cut across Sn-1 where it can be observed. Although the barroisitic amphiboles probably formed during the developement of Sn-1, since they are rimmed by actinolite, many of the grains are oriented in the dominant foliation Sn and appear to be oriented parallel to the lineation as viewed in oriented thin sections. These relations indicate, therefore, that the synmetamorphic faults formed during the Taconian Orogeny of Middle Ordovician age. This conclusion would suggest that all the structures that are displayed on the accompanying maps and photographs were formed during this time.

Figure 20

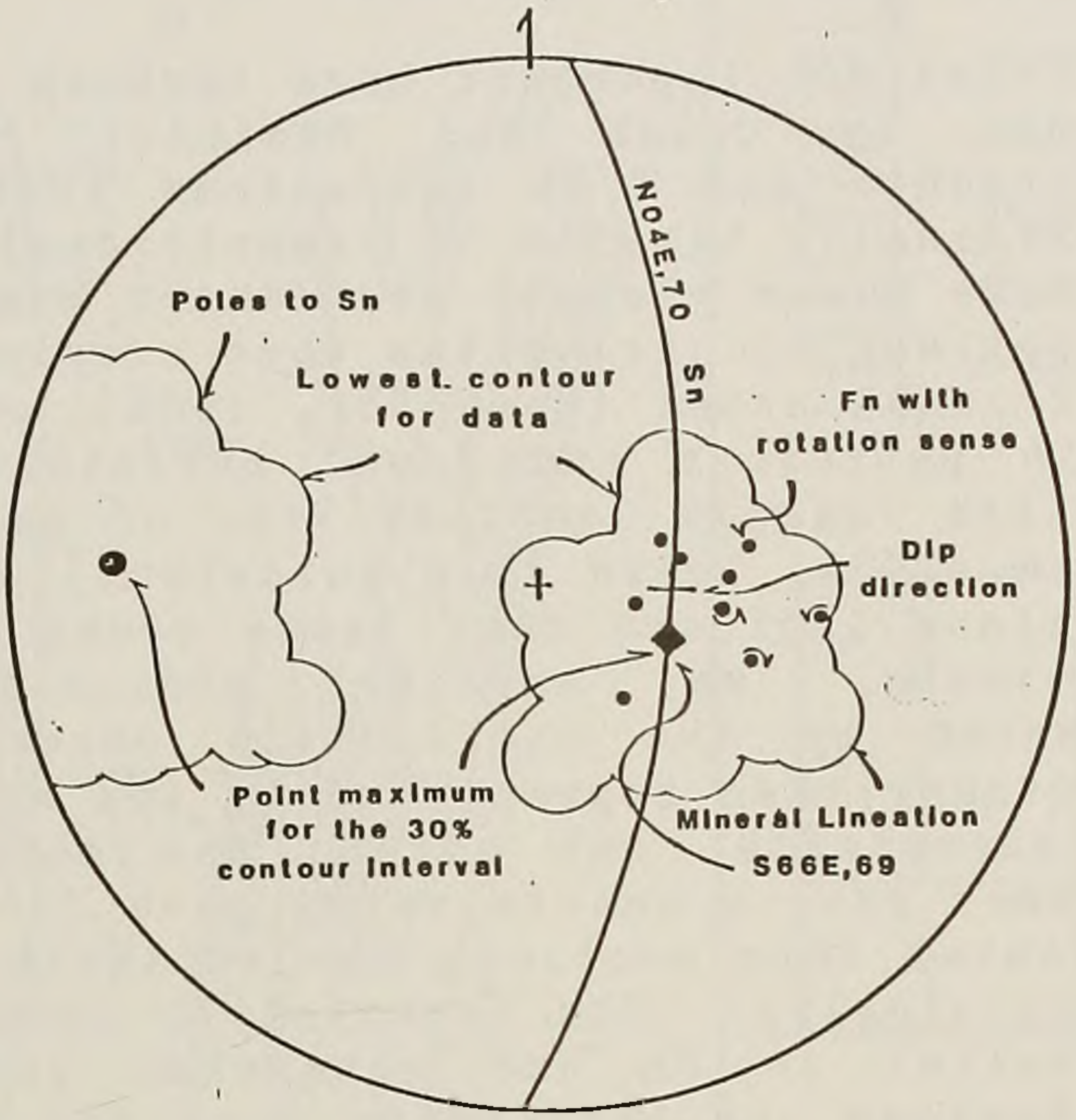


Figure 21

STRUCTURAL FABRIC FROM THE GRANVILLE GULF AREA

Lower hemisphere equal area net showing the orientation of 179 poles to the dominant schistosity (Sn), the orientation of 90 elongate grains, grain clusters, intersection lineations and quartz rods, and the orientation of 9 hinges of Fn folds. The center of the point maximum for the 30% contour interval corresponds to a surface (Sn) striking N04E and dipping to the east at 70 degrees. The corresponding point maximum for the mineral lineation plunges 69 degrees along a trend of S66E. This orientation represents the transport direction and it closely parallels the dip direction for the dominant Sn surface. Some thin sections cut parallel to this direction show asymmetrical fabrics that indicate an east-over-west sense of movement. See the photographs of thin sections from the mylonitic schist from the Ottauquechee fault zone at Granville Gulf. Note that all the observed Fn folds are oriented close to this direction and therefore are reclined. None of these folds have a sheath-fold shape. Compare this fabric with the fabric from Waitsfield where the Fn folds have a more varied orientation. The Fn folds in these two diagrams suggest that Fn hinges are rotated toward the transport direction as the Ottauquechee thrust fault is approached.

We will discuss the map and several cross sections that offer alternative interpretations of the geology.

The belts of mafic schist are important here because they have been geochemically analyzed by Coish and Masinter (1986). Their analyses of major elements and Y/Nb variations indicate that the mafic schists were originally basalts of transitional to tholeiitic character. The immobile trace element abundances suggest that they probably represent continental tholeiites that erupted through thin continental or transitional crust (Masinter, 1986). Laird has also studied the amphibole petrology and found barroisitic hornblende surrounded by actinolite (sample locality V14J of Laird and Albee, 1981; Laird and others, 1984; Laird this guidebook). The amphibole textures and compositions indicate that these rocks have undergone two metamorphic events, an earlier medium-high pressure metamorphism represented by the barroisitic cores and a lower greenschist facies metamorphism represented by the actinolite rims. Several samples of barroisite, one within the main part of the mafic schist and another near a quartz vein, gave 39Ar/40Ar ages of 471 and 448. In oriented thin sections the actinolite is oriented parallel to the Sn schistosity. The barroisitic hornblende appears to be flattened parallel to Sn and elongated parallel to the mineral lineation. Because the barroistic core has been partially altered to actinolite, however, this relations may be deceptive. There is a distinct possibility that the barroisite grew during the formation of the older Sn-1 schistosity. U-stage analysis has not been done on these rocks. We will discuss these implications as it pertains to the age of deformation and the fault classification shown on Table 1.

<u>Stop 16</u> - Structural fabric in the Ottauquechee Formation (figs. 22 and 23) (Kraus) - NO HAMMERS, PLEASE - This stop is located just east of Route 100 south of the junction of Plunkton Road and Route 100 between Warren and Granville Gulf. The outcrop is on the south side of a stream that forms a small ravine and is approximately 30 m east of Route 100 (fig. 22).

The outcrop is located in the quartzose schist member of the Ottauquechee Formation. Other units in this formation are the black phyllite and the black quartzite members. The black phyllite is graphitic and is composed of quartz, muscovite, chlorite and pyrite. The black quartzite is graphitic, commonly contains veins of white quartz, and is found in layers 2 cm to 30 m thick. The white quartzose schist is composed of sand-size quartz, muscovite, chlorite and albite. It places it contains thin interlayers of graphitic phyllite. A few blue quartz grains can be found at this locality. All three members can be seen in the outcrops along the stream.

The main feature at this stop is a spectacular fold-structure exposed in the quartzose schist. Differential weathering has left a remarkable display of tightly-folded layers, three foliations, intersection lineations, slickengrooves, mineral lineations, and associated minor faults (fig. 23). PLEASE BE VERY CAREFUL AND DO

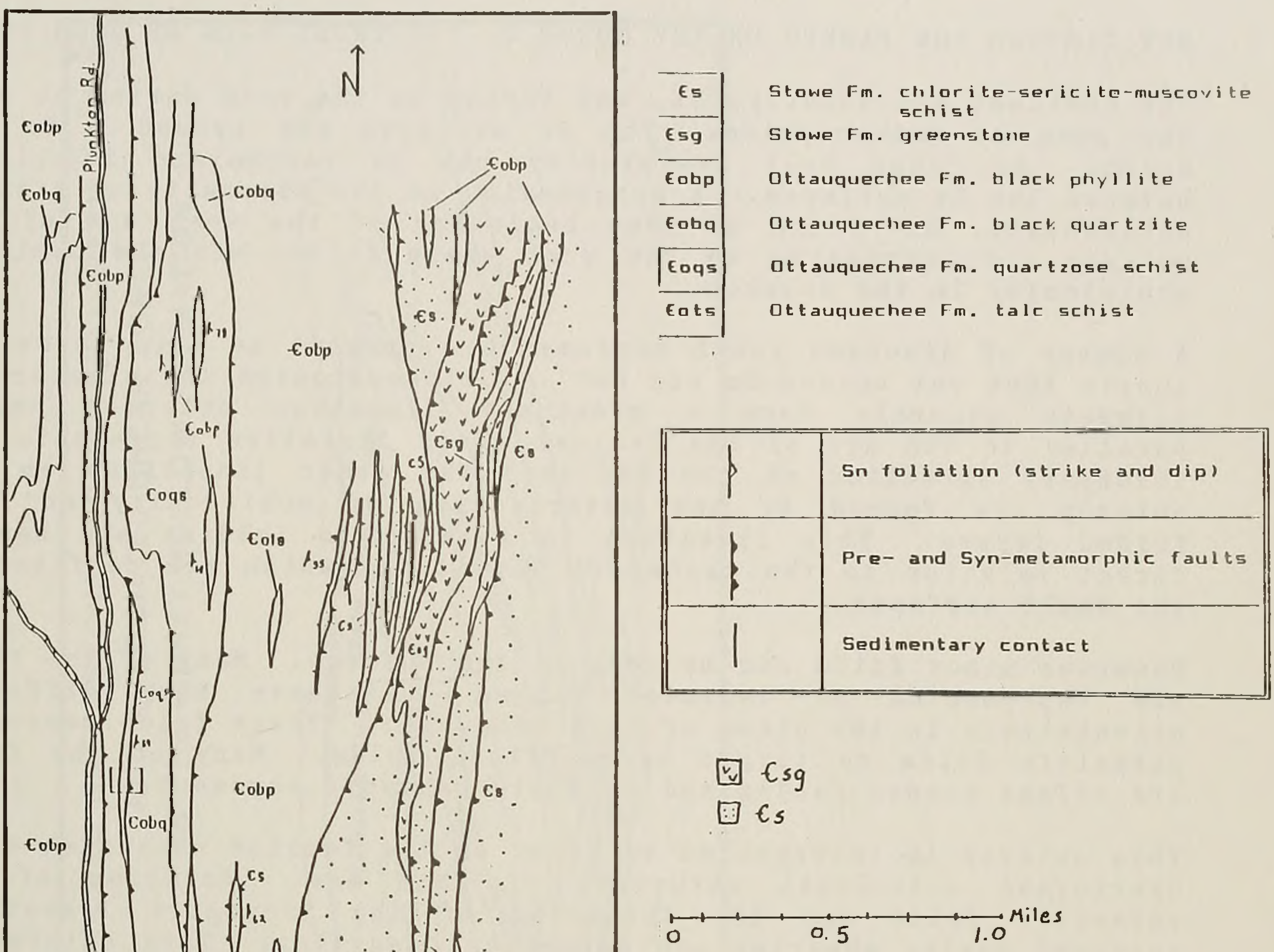


Figure 22

Location and local geology map. Stop location is indicated by the small square just off Rt. 100.

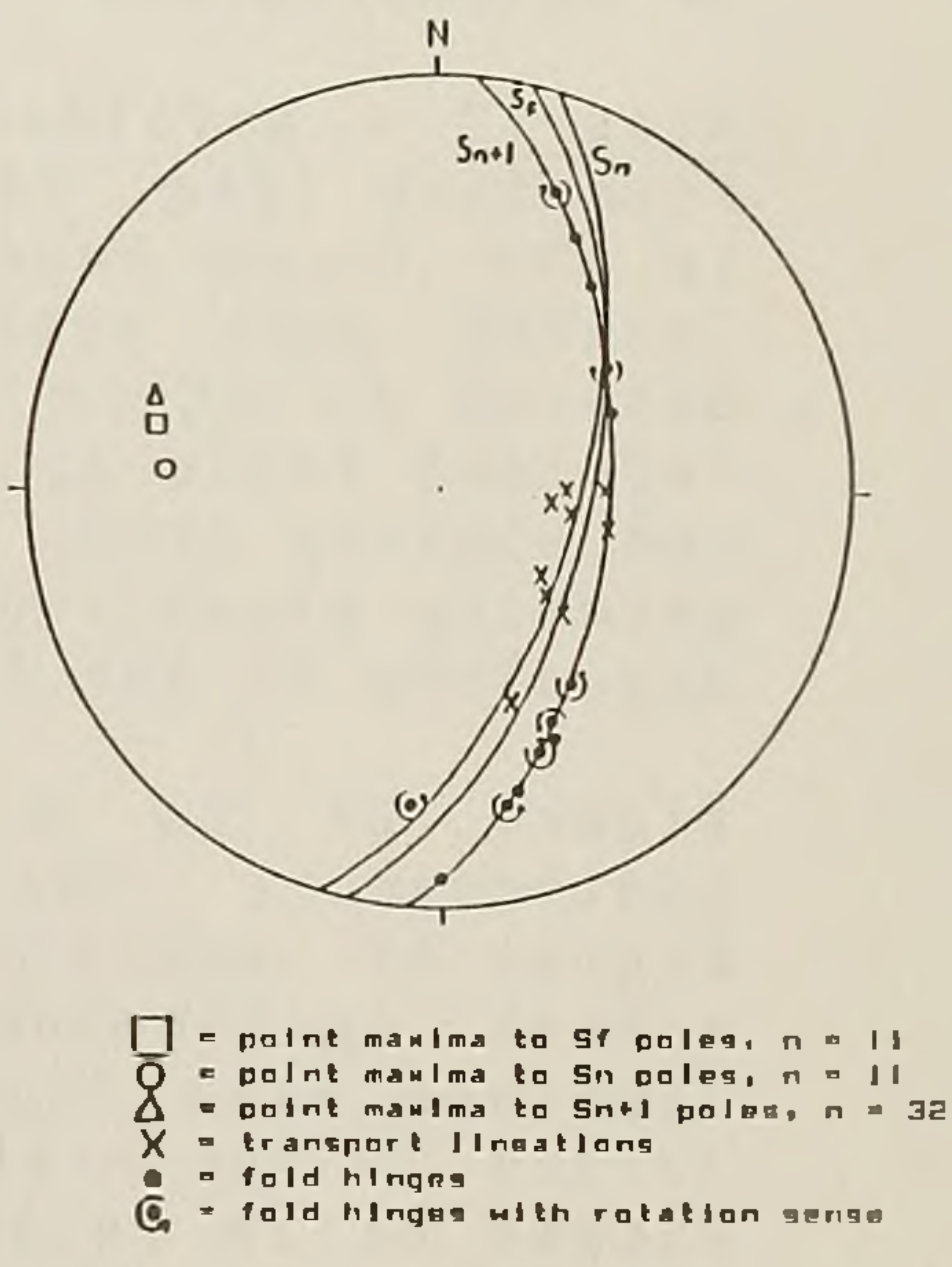


Figure 23

Lower hemisphere equal area projection of structural elements. No separation arc is implied due to uncertain fold to limb relationships.

NOT DISTURB THE FABRIC OR THE CURSE OF THE TROLL WILL BE UPON YOU!

The dominant schistosity, Sn, was formed in the rock during or near the peak of metamorphism. The Sn surfaces are spaced 1 to 5mm apart. An older Sn-1 schistosity can be recognized in places between the Sn surfaces. A progression in the intensity of a third schistosity, Sn+1, can be seen beginning at the west end of the outcrop and increasing to the west where it becomes the dominant schistosity in the outcrop.

A number of discreet fault surfaces are present as thin layers of quartz that cut across Sn and Sn+1. Slickengrooves and clusters of elongate minerals form a prominent lineation oriented nearly parallel to the dip of the faults. This lineation represents the transport direction on the faults. The other lineations in the outcrop are formed by the intersection of schistosity and the folded layers. This lineation is related to folding and has no direct relation to the transport lineations which are confined to the fault surfaces.

Numerous minor folds can be seen in the outcrop. Many of the folds are represented as isolated hinges that have many different orientations in the plane of Sn+1 (fig. 23). These folds represent parasitic folds on larger asymmetrical folds. Many of the folds are offset across faults and indicate eastward movement.

This outcrop is interpreted to occur on the faulted west limb of an overturned anticlinal structure of Fn+1 age. Rotation of the parasitic folds and the formation of the transport lineations occurred during shearing and subsequent faulting. This outcrop is an exposure of a synmetamorphic fault that developed after the peak of metamorphism.

Stop 17 - Refolded fold of Fn-1 and Fn age in the Pinney Hollow Formation (fig. 24) (Stanley, Armstrong) - This outcrop is located in the front yard of a private home in Alpine Village which is located just west of Plunkton Road. PLEASE, NO HAMMERS. This outcrop is quite rare in that it is one of the few places where refolded folds of Fn-1 and Fn age are preserved in the mafic schist and slivery green schist of the Pinney Hollow. We believe that the geometry preserved in this outcrop is representative of the larger structure in the Pinney Hollow Formation.

Figure 24 is a geologic map of the outcrop drawn from a photomosaic. The relations are therefore portrayed with a fair degree of accuracy. Carefully trace the contact between the mafic schist (greenstone) and the silvery schist and compare your findings with figure 24. The axial surface (Sn) of the Fn fold trends to the north. It folds Sn-1 which is well preserved in the hinges of Fn folds in the greenstone but is largely transposed in the adjacent schist. After you have recognized the structure see if the folds continue to the east and west across the outcrop. You will find that they are more difficult to follow in these directions. We suggest that the refolded fold in the center of the

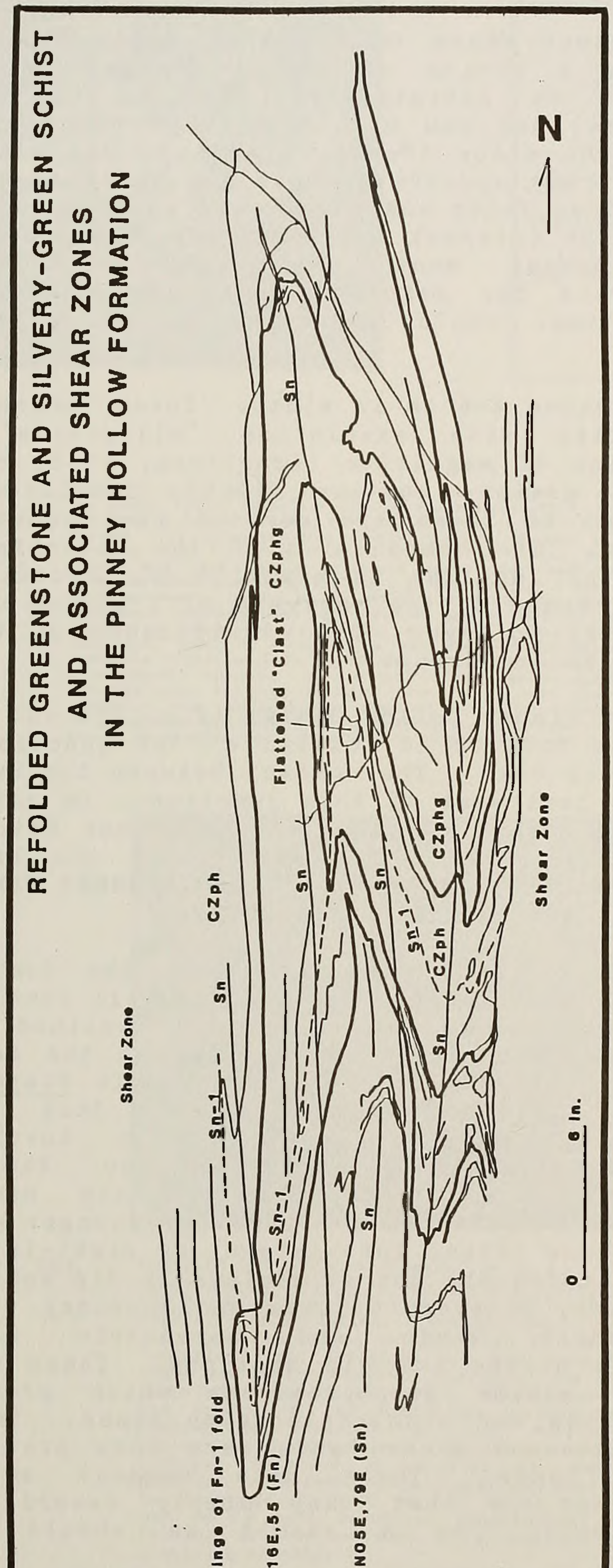

REFOLDED GREENSTONE AND SILVERY-GREEN SCHIST
AND ASSOCIATED SHEAR ZONES
IN THE PINNEY HOLLOW FORMATION
N
Shear Zone
Sn
CZph
Sn-1
Flattened 'Clast' CZphg
CZphg
CZph
Hinge of Fn-1 fold
S16E,55 (Fn)
N05E,79E (Sn)
0
6 in.
Shear Zone

Figure 24

outcrop is truncated by shear zones to the west and east. The eastern one is marked by a string of quartz lenses that are oriented parallel to the Sn schistosity. Follow this zone northward (toward the house) and you will find that the zone is oriented parallel to the the older (Sn-1) layering. The western shear zone is less distinct but appears to have the same geometry. We believe that these refolded folds and associated shear zones are representative of part of the internal fabric in the Pinney Hollow Formation. The shear zones would represent the linear symetamorphic faults whereas the refolded fold, which here is nearly reclined in position, would represent the sheared-out "sheath" folds.

Also note the small dark lenses 4-6 cm in width. These lenses are composed of titanohematite with exsolution "ellipsoids" of ilmenite and separate grains of magnetite (Armstrong, 1986, class paper). The magnetite grains contain (111) lamellae of ilmeno-hematite. Pods such as these are derived from hematitic shales during metamorphism. The composition of the titaniferous hematite host and associated lamellae is hematite 80 percent and ilmenite 20 percent and is based on the analysis of 7 grains using the microprobe at Middlebury College. The significance of these analyses will be discussed by Armstrong.

Stop 18 - Fabric within the Pinney Hollow Schist (fig. 25a and 25b) (Stanley and Coleman) This outcrop is located at the junction of Route 100 and the Lincoln Gap Road. The contact between the Hazens Notch Formation is located just west of this junction. An outcrop of the carbonaceous albitic schist (CZhnca) of the Hazens Notch is located just west of the junction. The outcrop of the Pinney Hollow schist is on the east side of the road. NO HAMMERS PLEASE. JUST OBSERVE AND APPRECIATE WHAT NATURE HAS PRESERVED!

The fabric in the schist is very complex. Here the dominant schistosity (Sn), as we see it in strike section, is well developed and anastomoses. Prominent mineral lineations and reclined fold hinges plunge down the dip. This fabric is similar to the fabric found in many places in the Pinney Hollow (for example Stops 13, 14, and 15). Coleman and Journeay (1987) suggest that these fabrics record a history of rotational strain involving layer-parallel shear, flattening perpendicular to the Sn schistosity and extreme elongation parallel to the mineral lineation. In many places this fabric is offset by younger shear zones (C surfaces) that record either left-lateral or right-lateral displacement. These shear zones are not as obvious in dip sections in outcrop. In thin section, however, Coleman and Journeay (1986) observed microscopic shear bands and asymmetric albite porphyroblasts in both the strike and dip sections. These shear bands wrap around older albite porphyroblasts which preserve inclusion trails of clockwise and counterclockwise sense. In the dip section the down-to-the-east shear bands were more prevalent than the up-to-the-west bands. These data suggest several interpretations. The first is that they simply record late flattening across the belt. As a result we should find

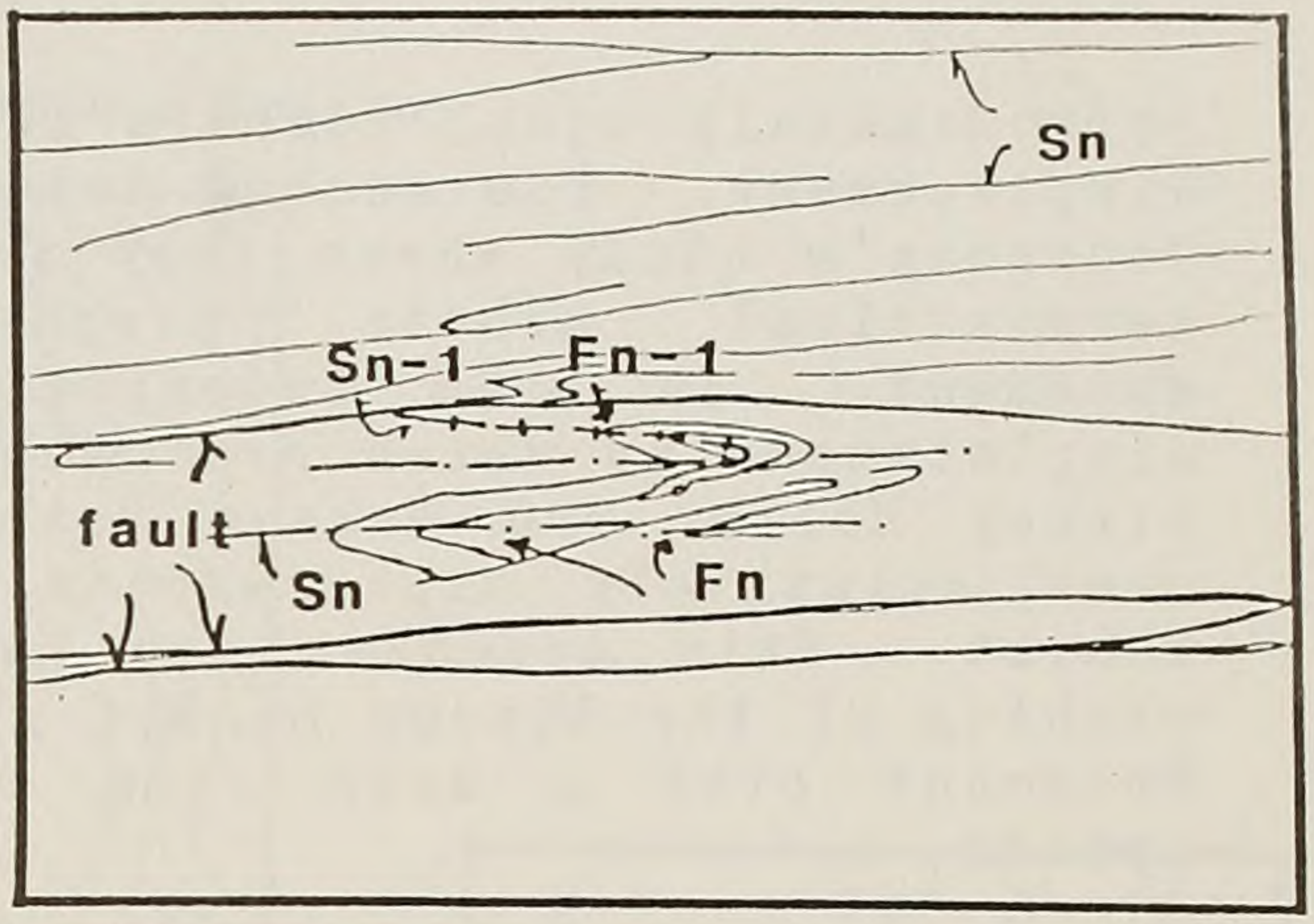


Figure 25a Origin of synmetamorphic faults Fault zone fabrics in the silvery schist of the Pinney Hollow Formation showing the sheared off limb of an Fn fold which has in turn folded an older Fn-1 fold (center of the diagram). These relations suggest that synmetamorphic faults develope through a process of intense shearing along the limbs of tight to isoclinal folds. Thus the fold developed before the fault unlike foreland folds that develope as a result of irregularities (ramps etc.) along a propagating fault surface. The difference in mechanism may be due to the absence of strong contrasts in ductility in the hinterland compared to the foreland where beds of distinctly different mechanical properties form major surfaces of weakness. The anastomosing foliation, commonly measured as the dominant foliation Sn, is really a composite foliation made up of Sn and the transposed remains of Sn-1 and compositonal layering. The schist in many synmetamorphic fault zones is commonly more strongly foliated than it is away from the faults. This indicates that movement has occurred not only during but slightly after metamorphism. Although original bedding is rarely preserved in these rocks, it can be found along contacts with beds of greenstone and quartzite.

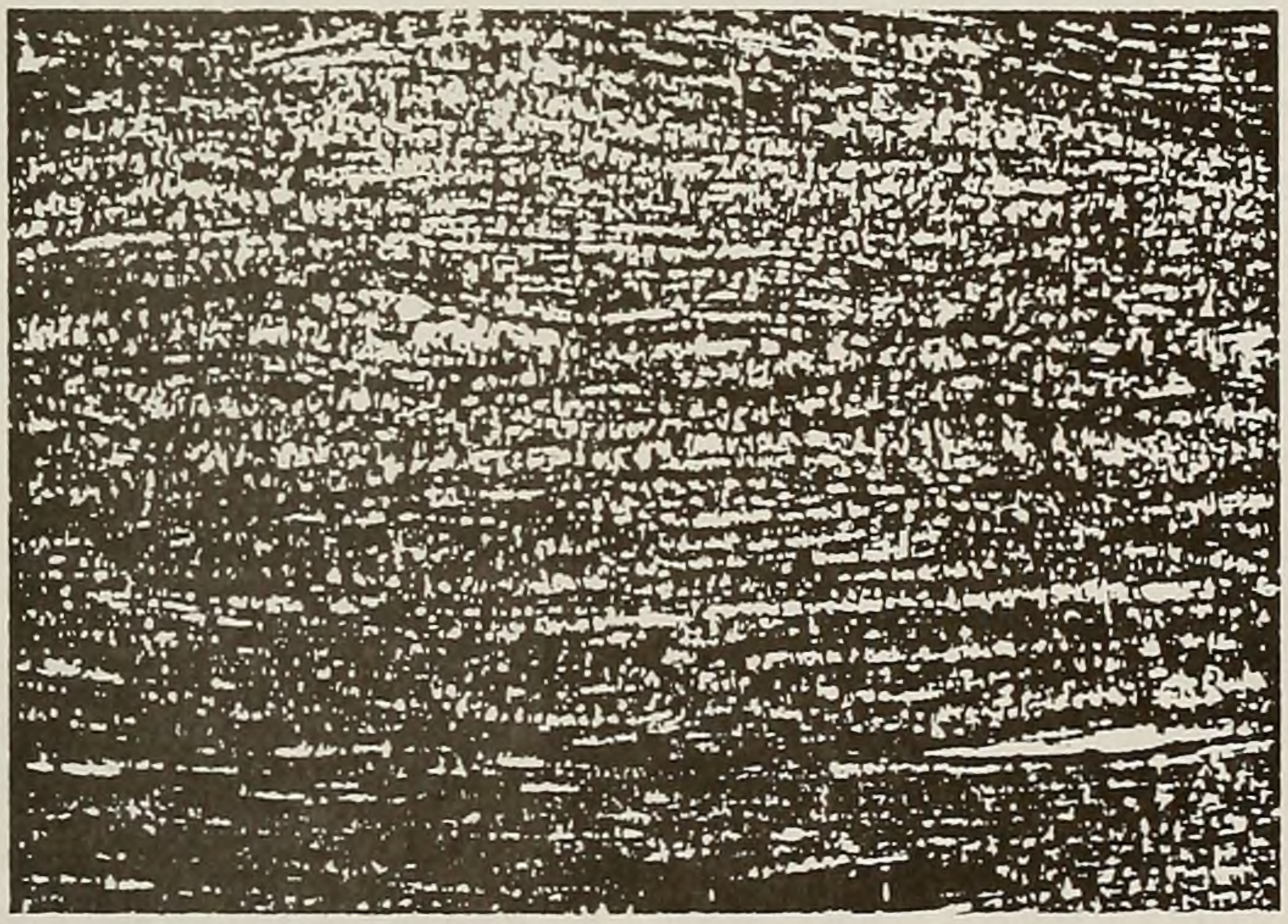


Figure 25b Foliated and fragmented pegmatite in a synmetamorphic fault in the slivery green schist of the Pinney Hollow Formation The foliation is the pervasive Sn foliation. Isolated quartz stringers and veins are strung out parallel to the dominant foliation. The relations are best seen directly above and to the right of the pen. The schist is composed of quartz, muscovite, chlorite and albite. The fabric in the pegmatite indicate that Fn deformation occurred at the ductile-brittle transition where quartz deformed ductily and feldspar deformed by brittle to semi-brittle fracture. Suggested temperatures and pressures are in the range of 350^0C to 400^0 C and 3 or 4 Kbars (10 to 13 km.) based on experimental work on feldspar (Tullis and Yund, 1977) and the estimated stability of chlorite.

approximately equal numbers of shear bands with opposite senses of displacement. The second interpretation is based on Coleman's and Journeay's study where they found that 85% of the shear bands and asymmetrical albite porphyroblasts record down-to-the-east movement. In this interpretation these structures record late displacement in which greenschist facies hanging-wall rocks of the Pinney Hollow and Hazens Notch Formations moved down to the east over relatively higher-grade footwall rocks of the Mt. Abraham Schist. This movement may have been a result of the uplift and arching of the Lincon massif and its cover immediately to the east. Movement over a deep ramp may have been responsible for this uplift.

In one place in the outcrop an older Fn-1 fold is preserved between the Sn schistosity (fig. 25a). In another place an older pegmatite is disrupted by Sn and the younger shear bands (fig. 25b).

Perhaps the most important lesson that can be learned from this outcrop is that the fabric is clearly the result of very complex and intense strain. Originally, we thought that the fabric was largely the result of simple shear associated with faulting. Pure shear (flattening) certainly has played a role in the development of the shear bands. The earlier fabric may well have been the result of a large component of pure shear since asymmetrical structures with a consistent sense of displacment are absent. The rotated trails in the albite certainly suggest this, although more work should be done on these structures. Our present position is that one should be cautious in concluding that outcrops like this one are simply the result of synmetamorphic faulting. Other such criteria as the geometry of map units should be used in conjunction with the fabric.

Stop 19 - Fault zones in the Ottauquechee Formation north of Waitsfield (fig. 26) (Walsh) - These series of outcrops are found on either side of Route 100.

Station A (fig. 26) - A folded contact between the Ottauquechee black phyllite (Cobp) and the Ottauquechee greenstone (Cobg) is well exposed at this locality. The black phyllite can be seen in the cores of several of the folds along the contact with the greenstone. The dominant schistosity (Sn) is axial planar to the folds, thus indiciating that they are Fn in age. The geometry of the Fn folds is best seen in the greenstone where the folds are isoclinal and have hinges that plunge steeply to the south and the north. The geometry of these folds is consistent throughout the area and is most likely related to larger map-scale folds.

The contact between the black phyllite and the greenstone is interpreted as depositional. From the relationship between the contact, the Fn folds, and the dominant schistosity it is evident that the contact predates the later structures. No evidence, however, is seen for a pre- or early metamorphic fault along this contact.

Bedrock Geologic Map of the Northern Fayston Area in the Vicinity of Route 100

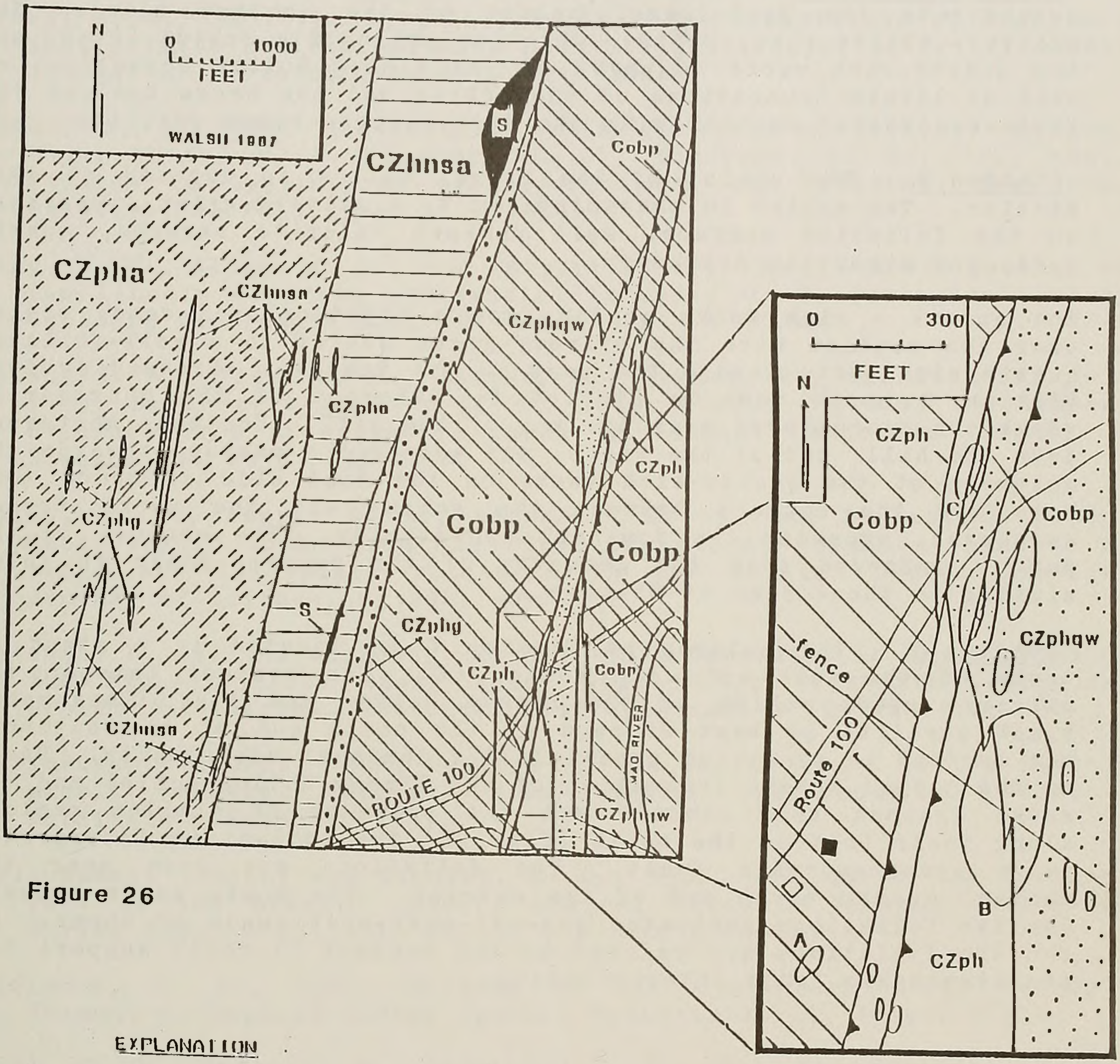


Figure 26

EXPLANATION

Lithologies:

ꞒZpha: albite-quartz-chlorite-muscovite schist
ꞒZph: quartz-albite-chlorite-muscovite-magnetite schist
ꞒZphqw: quartz-rich wacke/schist
ꞒZphg: albite-epidote-chlorite-calcite-actinolite schist/greenstone
ꞒZhnsa: interlayered quartz-chlorite-muscovite-albite schist and albitic carbonaceous schist
Ꞓobp: graphitic pyritiferous quartz-muscovite-chlorite schist
S: undifferentiated serpentinite and talc carbonate

Symbols:

Thrust fault--teeth on upper plate
Outcrop
House
Barn
A,B,C Station location

Stations B and C - The following series of outcrops will demonstrate the gradational nature of the contact between the quartz - albite - muscovite - chlorite -magnetite schist (CZph) and the quartz-rich wacke (CZphqw) of the Pinney Hollow Formation, as well as lithic truncations of the schist and the wacke against the fault-controlled contact with the Ottauquechee black phyllite.

Station B - The schist of the Pinney Hollow is exposed at this station. The schist is characterized by a silvery-green appearance on the foliation surface, well defined chlorite streaks, quartz rods, and magnetite octahedra.

The quartz - rich wacke grades from a light-colored sandy schist near the contact with the silvery-green schist to a well-foliated quartz-rich wacke toward the east in the pasture. Proceeding from Station B north towards Station C, outcrops of the quartz-rich wacke are encountered near the top of the hill. Continue northwest down the hill, out of the woods, and across the pasture towards the outcrops of the quartz-rich wacke on the east side of Route 100. Note how the contact between the schist and the wacke trends northwest, crossing the dominant schistosity. The contact on the map is inferred from the geometry of the outcrop-scale Fn folds similar to those seen at Station A.

Station C - The outcrop at Station C is located in a stand of trees on the west side of Route 100. The contact between the silvery- green schist of the Pinney Hollow and the Ottauquechee black phyllite is best exposed at the north end of the outcrop. The contact is sharp and parallels the dominant schistosity. Based on the mapped lithic truncation of the Pinney Hollow schist and the wacke against this contact and the presence of similar geology along their belt to the north (fig. 26) the contact is interpreted as a synmetamorphic fault. The foliations are seen near the contact at the north end of the outcrop. The acute angle between the two foliations indicates an east-over-west sense of shear. If the two foliations are related to the contact it would support the synmetamorphic fault interpretation.

REFERENCES CITED

Albee, A, L. 1965, 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, E-an, White, W. S., Hadley, J. J., Thompson, J. B., Jr., eds., Studies in Appalachian Geology-Northern and Maritime: New York, Interscience Publisher, p. 329-341.

Cady, W. M., Albee, A. L., and Murphy, J. F., 1962, Bedrock geology of the Lincoln Mountain quadrangle, Vermont: U. S. Geological Survey Geologic Quadrangle Map, GQ - 164, scale 1:62,500.

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

Coleman, M. E., and Journeay, J. M., 1987, Kinematic interpretation of the Pinney Hollow shear zone, west-central Vermont: Geological Society of America, Abstracts with Programs, v 19, Nr. 1, p. 9.

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

DiPietro, J. A., 1983, Geology of the Starksboro area, Vermont: Vermont Geological Survey Special Bulletin No. 4, 14 p., 2 pl.

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, Scale 1: 250, 000.

Filosof, Anette, 1986, Geochemistry of Precambrian dikes near Ripton, Vermont: Senior thesis, Middlebury College, Middlebury, Vermont, 15p.

Gavigan, T. H., 19986, Geochemistry of six South Lincoln greenstones: Senior thesis, Middlebury College, Middlebury, Vermont, 20p.

Karabinos, Paul, 1984, Deformation and metamorphism on the east side of the Green Mountain massif in southern Vermont: Geological Society of America Bulletin, v. 95, p. 584-593.

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

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

Laird, Jo, 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.

Lapp, E. T., 1986, Detailed bedrock geology of the Mt. Grant - South Lincoln area, Central Vermont, M. S thesis, University of Vermont, Burlington, Vermont, 114p.

Lapp, E. T., and Stanley, R. S., 1986, Geology of the Mt. Grant - South Lincoln area, central Vermont; Vermont Geological Survey Special Bulletin No. 7, 20 p. and 3 plates.

Masinter, R. A., 1986, Petrology of the Hazens Notch and Pinney Hollow greenstones: Green Mountain Geologists, Spring Meeting of the Vermont Geological Society.

O'Loughlin, S. B, 1986, Geology of the Mt. Abraham - Lincoln Gap area, Central Vermont, M.S. thesis, University of Vermont, Burlington, Vermont, 164p.

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, Bull. no. 5, 127p.

Prahl, C.J., 1985, The geology of the Middle Proterozoic rocks of the Lincoln massif of west-central Vermont and their suitability for long term storage of high level nuclear waste: Senior thesis, University of Vermont, Burlington, Vermont: 42p.

Rowley, D. B., and Kidd, W. S. F., 1981, Stratigraphic relationships and detritial 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.

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

Stanley, R. S., Roy, D. L., Hatch, N. L., Jr., 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.

Strehle, B. A. and Stanley, R. S., 1986, A comparison of fault zone fabrics in northwestern Vermont: Vermont Geological Survey, Studies in Vermont Geology No. 3, 36 p. and 4 plates.

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

Tauvers, P. R., 1982a, Bedrock geology of the Lincoln area, Vermont: Vermont Geological Survey Special Bulletin No. 2, 8 p., 1 pl.

Tauvers, P. R., 1982b, Basement-cover relations in the Lincoln area, Vermont: M. S. Thesis, University of Vermont, Burlington, Vermont, 177 p.

Thompson, J. B., Jr., 1967, Bedrock geology of the Pawlet quadrangle, Vermont, part II, Eastern portion: Vermont Geological Survey Bulletin No. 30, p. 61-98.

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

METAMORPHISM OF PRE-SILURIAN ROCKS, CENTRAL VERMONT

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(This note is meant to accompany field trip B-8 by Stanley and others, this volume. Please refer to that paper for geologic units, structural features, and field trip stops.)

Our understanding of the metamorphism in central Vermont is based on petrologic and isotopic studies of intercalated pelitic, mafic, and calcareous rocks. West of the Green Mountain anticlinorium (GMA) metamorphic grade increases from chlorite to kyanite zone with the highest metamorphic grade east of the Lincoln Massif (lm, fig. 1). Biotite to kyanite zone rocks also crop out east of the GMA, with the highest metamorphic grade along the Northfield, Worcester, and Elmore mountains (nm, wm, em, fig. 1).

Weakly metamorphosed limestones west of the Lincoln Massif record calcite-dolomite, oxygen, and carbon isotope temperatures between 210 and 295 C (Sheppard & Schwarcz, 1970), while biotite grade dolomite and marble give temperatures between 255 and 400 C. North of the Lincoln Massif biotite grade pelitic schist gives an oxygen isotope temperature of 435 C (Garlick and Epstein, 1967; all stable isotope temperatures have been recalculated using the fractionation curves compiled by Friedman & O'Neil, 1977). Garnet grade mafic schist gives calcite-dolomite and amphibole-plagioclase temperatures of 489 to 450 C and an amphibole-plagioclase pressure of 7 to >8 Kbar. At Mount Grant (m, fig. 1) oxygen isotopes indicate kyanite grade metamorphism at 420 to 465 C (Garlick and Epstein, 1967).

Pelitic assemblages reported by Albee (1965, 1968) indicate that metamorphism reached a higher temperature east of the GMA than farther west (Worcester Mountains vs. Green Mountains, figure 2). Amphibole and plagioclase compositions in mafic schist do not support this hypothesis, perhaps because of retrograde metamorphism. At Elmore Mountain, for example, the earlier pelitic assemblage kyanite + garnet + biotite can be observed even though all of these minerals are pseudomorphed (Albee, 1972, Stop I-2). Hornblende is pseudomorphed by actinolite (Laird and Albee, 1981) and has similar composition to hornblende in garnet grade mafic schist near Lincoln. If the highest grade hornblende is not preserved, there is no way to tell from electron microprobe analyses how high grade the mafic rock was before retrogradation. Plagioclase in mafic rocks is albite both east and west of the GMA.

One cannot distinguish the pressure of metamorphism east and west of the GMA from the pelitic samples alone. The composition of amphibole in amphibole + chlorite + epidote + plagioclase + quartz

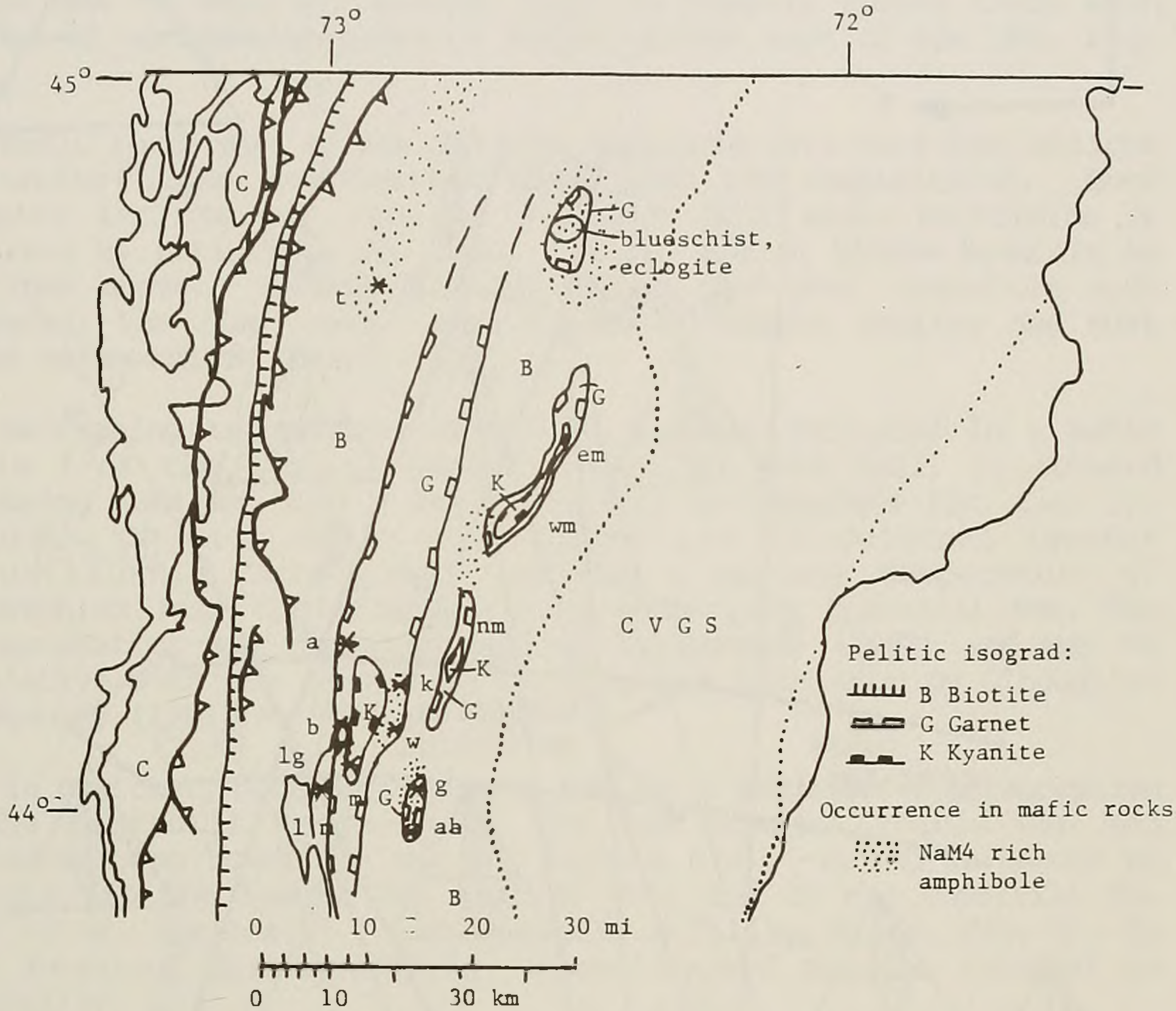

Figure 1: Metamorphism in north-central Vermont after Doll et al. (1961) and Laird et al. (1984). Stars refer to localities discussed in text: t(Tibbit Hill volcanics), a (Appalachian Gap), b (Battell greenstone), lg (Lincoln Gap road and south), lm (Lincoln Massif), m (Mount Grant), k (Kew Hill), w (Warren), g (Granville Gulf), ab (Allbee Brook). Other abbreviations: em (Elmore Mountain), wm (Worcester Mountains), nm (Northfield Mountains), CVGS (Connecticut Valley Gaspe Synclinorium). The blueschist and eclogite locality is discussed by Bothner and Laird (this volume).

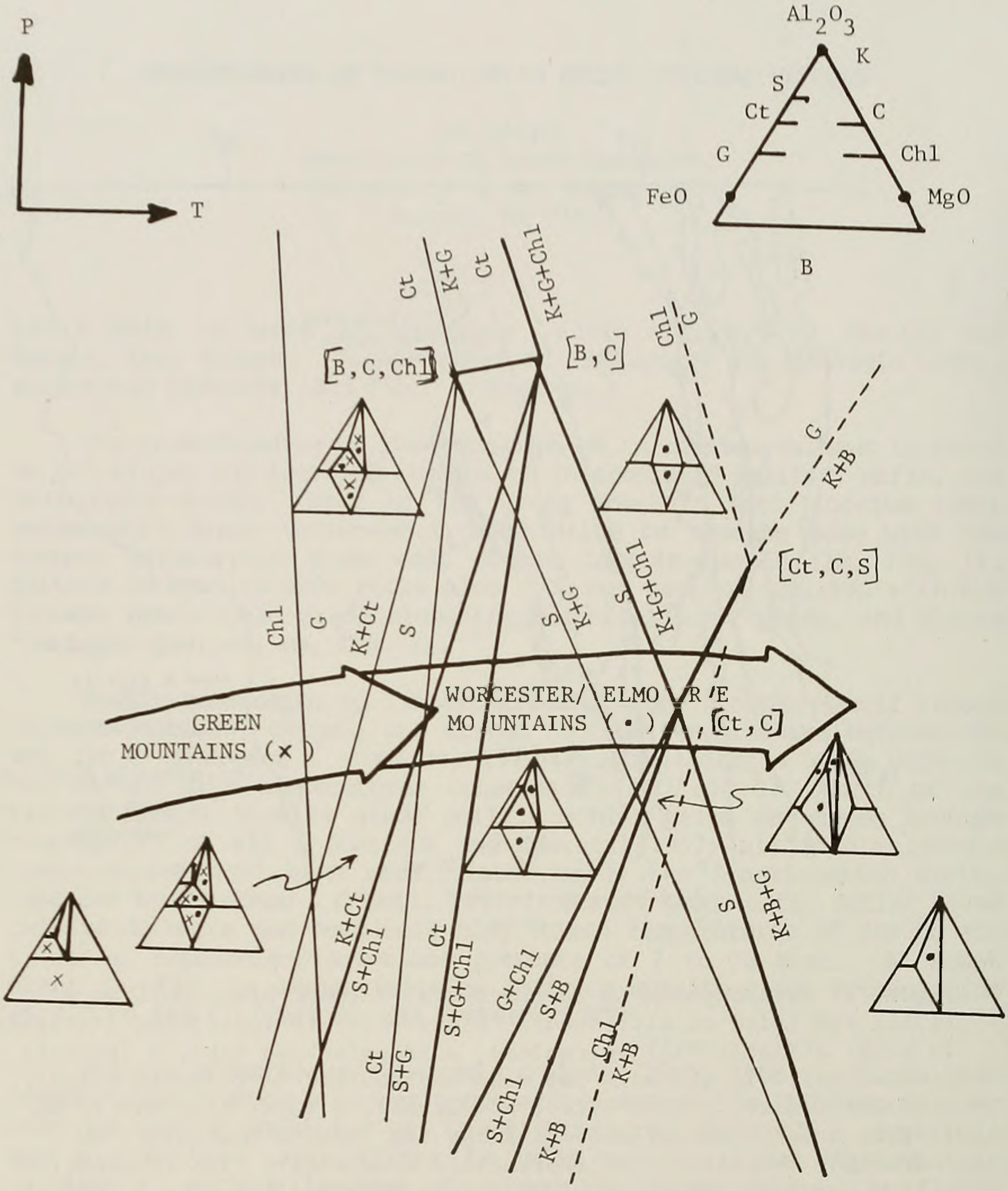

Figure 2: Mineral assemblages in pelitic schist from pre-Silurian rocks, north-central Vermont. Pressure-temperature grid is from Harte and Hudson (1979). Slopes of metamorphic reactions are generalized. The maximum grade of metamorphism in the Green Mountains west of the Green Mountain anticlinorium and in the Worcester Mountains (wm, fig. 1) is represented by the end of the appropriate arrow. Assemblages are shown on a projection from quartz and muscovite (Thompson projection). Abbreviations: B (biotite), C (cordierite), Chl (chlorite), Ct (chloritoid), G (garnet), K (kyanite), S (staurolite).

schist, however, shows very nicely that in central Vermont metamorphism was higher pressure east than west of the GMA (fig. 3). In contrast, high NaM4 in amphibole indicates that medium-high pressure facies series metamorphism is identified both east and west of the GMA in northern Vermont (fig. 1; compare sample LA426 with samples of medium-high-pressure facies series east of the GMA, fig. 3).

Mafic rocks east of the GMA have amphibole with core compositions indicating higher metamorphic grade than rim compositions. Good examples include our stop at Granville Gulf where barroisite is overgrown by actinolite (g, figs. 1 & 3) and at Elmore Mountain as described above. Because the actinolite rims grew (amphibole mode increased) the rims cannot have formed by simple cooling and must record metamorphism (Laird, 1986).

Decreasing temperature with time is also indicated in a mafic sample from the Lincoln Massif (south of Cobb Hill) by outward decreasing NaM4 and AlVI + Fe3 + 2Ti + Cr in amphibole (lm, fig. 3), although Ti rich cores are suggestive of original igneous compositions. This sample records a maximum temperature of metamorphism less than the overlying Hoosac and Underhill fms. The retrogradation was reported earlier by Tauvers (1982) and may be correlative with the retrograde metamorphism documented by O'Loughlin and Stanley (1986).

In contrast, continuous zoning indicates progressive metamorphism in the Underhill Fm. west of the GMA along Lincoln Gap and Appalachian Gap roads (lg and a , figs. 1 & 3). Zoning is harder to interpret in the Hoosac Fm. (lg, EL 475) and in the Underhill Fm. south of Mt. Abraham (b) and at Castle Hill (lg, EL 66, fig. 3), in part because the zoning is irregular and may be related to miscibility.

40Ar/39Ar age data on amphibole obtained by Marvin Lanphere (USGS, Menlo Park) show that Taconian age metamorphism at stations a and g (fig. 1) is 471 Ma (Laird et al., 1984). Laird et al. (1984) report 376 to 386 muscovite and biotite 40Ar/39Ar total fusion ages from the kyanite grade pelitic schist east of the Lincoln Massif studied by Albee (1965). Lanphere et al. (1983) also reported Devonian biotite (387 Ma) and amphibole (382 Ma) 40Ar/39Ar ages from the Lincoln Gap road (VJL340). Do these data mean that Acadian metamorphism is recorded east of the Lincoln Massif while Taconian metamorphism is recorded farther north and east?

Structural data presented by Stanley and his colleagues on this trip indicate a similar age of deformation and metamorphism east and west of the GMA. Sutter et al. (1985) interpret Lanphere's Devonian ages as cooling ages from a Taconian metamorphic event. Further isotopic data are needed along with our combined structural, geochemical, and petrologic studies to address this problem.

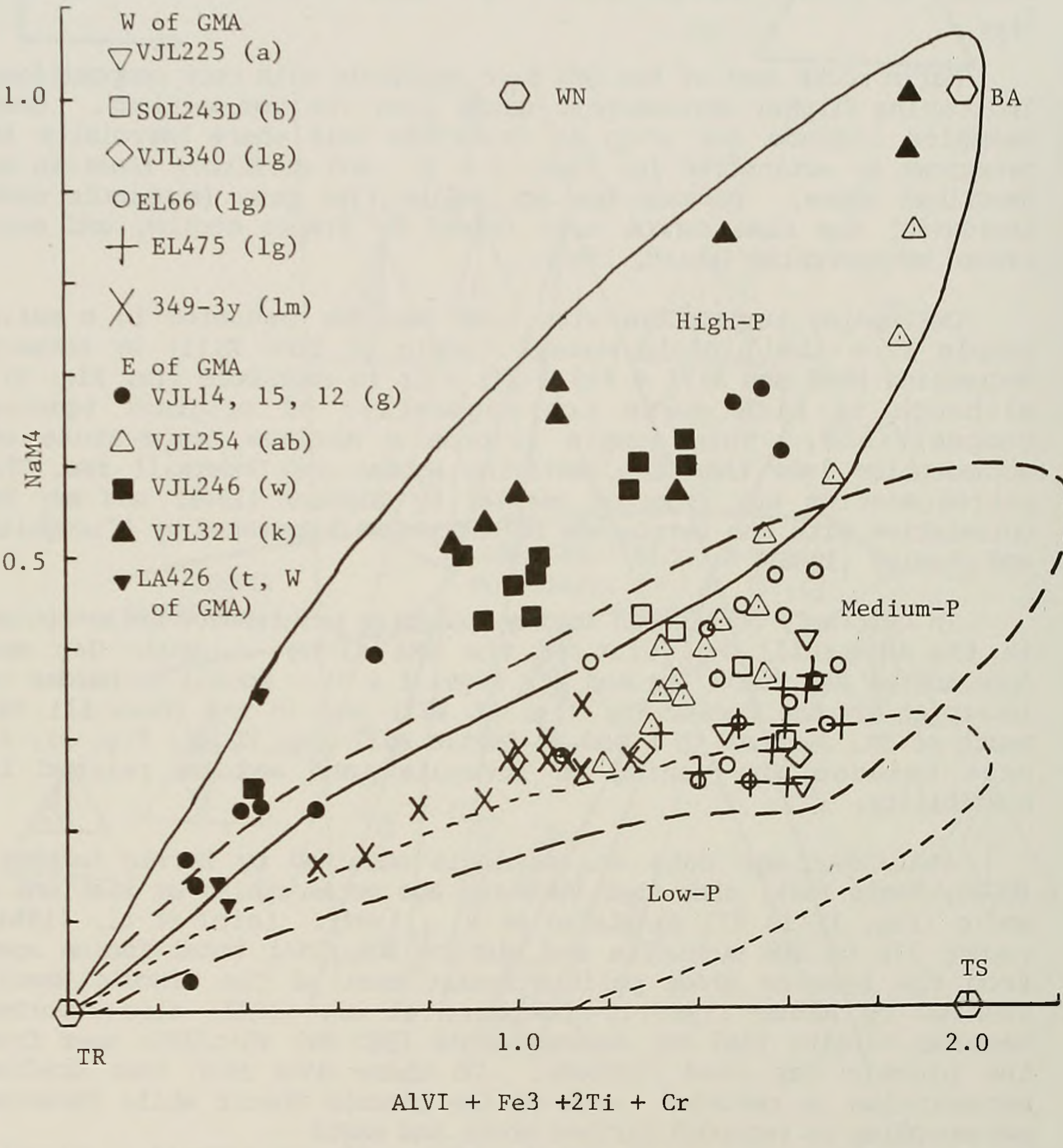

Figure 3: Formula proportion NaM4 versus (AlVI + Fe3 +2Ti + Cr) for amphibole in amphibole + chlorite + epidote + plagioclase + quartz schist, central Vermont. Sample localities of VJL numbers from Laird et al. (1984). Localities in ( ) from fig. 1. Envelopes for high-, medium-, and low-pressure facies series metamorphism are from Laird et al. (1984). Increasing temperature of metamorphism results in increasing advancement along both the X and Y axes. TR (tremolite), TS (tschermakite), WN (winchite), BA (barroisite). Analyses are electron microprobe data and normalized to total cations - (Na + K) = 13

ACKNOWLEDGEMENTS

This research was funded by NSF grants DES 69-0064 and EAR 75-03416 to Arden Albee and by NSF grant EAR-83-19383 to JL and W. A. Bothner. Rolfe Stanley, Sharon O'Loughlin, Eric Lapp, and Vincent DelloRusso provided several samples of mafic schist and stimulating discussions concerning the geology of the transect covered by this field trip. I am especially indebted to Arden Albee, Alan Thompson, and James B. Thompson, Jr. for helping me grapple with the petrology of mafic schist and to Marvin Lanphere for doing the 40Ar/39Ar age dating.

REFERENCES

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

__________ (1968) Metamorphic zones in northern Vermont: in E-an Zen, W.S. White, J.B. Hadley, and J.B. Thompson, Jr. (eds.), Studies of Appalachian Geology: Northern and Maritime: New York, Wiley Interscience, p. 329-341.

__________ (1972) Stratigraphic and structural relationships across the Green Mountain anticlinorium: in B.L. Doolan and R.S. Stanley (eds.), Guidebook for Field Trips in Vermont, NEIGC 64th Annual Meeting, p.179-194.

Doll, C.G., Cady, W.M., Thompson, J.B., Jr., and Billings, M.P. (1961) Centennial geologic map of Vermont: Vermont Geological Survey.

Friedman, L., and O'Neil, J.R. (1977) Compilation of stable isotope fractionation factors of geochemical interest: in M. Fleischer (ed.), Data of Geochemistry, Sixth Edition: United States Geological Survey Professional Paper 440-kk, p. 1-12.

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

Harte, Ben, and Hudson, N.F.C. (1979) Pelitic facies series and the temperature and pressures of Dalradian metamorphism in E. Scotland: in A.L. Harris, C.H. Holland, and B.E. Leake, The Caledonides of the British Isles — Reviewed: Geological Society of London, p. 323-337.

Laird, Jo (1986) Mafic schist: Pursuit of a P,T calibration using whole rock experiments and rocks from Vermont: Geological Society of America Abstracts with Programs, v. 18, p. 29.

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

_________, Lanphere, M.A., and Albee, A.L. (1984) Distribution of Ordovician and Devonian metamorphism in mafic and pelitic schists from northern Vermont: in K.C. Misra and H.Y. McSween, Jr., eds., Mafic and Ultramafic Rocks of the Appalachian Orogen: American Journal of Science, v. 284, p. 376-413.

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

O'Loughlin, S.B., and Stanley, R.S. (1986) Bedrock geology of the Mt. Abraham-Lincoln Gap area Central Vermont: Vermont Geological Survey Special Bulletin No. 6.

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.

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

Tauvers, P.R. (1982) Bedrock beology of the Lincoln area, Vermont: Vermont Geological Survey Special Bulletin No. 2.

REGIONAL GEOCHEMICAL VARIATIONS IN GREENSTONES FROM THE CENTRAL VERMONT APPALACHIANS

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INTRODUCTION

Greenstones from the Mount Holly Complex, the Pinnacle, Underhill, Hazens Notch, Pinney Hollow, Ottauquechee and Stowe formations have been analyzed for their major, trace and rare earth element abundances.This is part of a long term project by students at Middlebury College begun seven years ago. I present here a summary of regional geochemical trends and discuss briefly their tectonic implications. Some of the ideas presented here are from Coish et al. (1985; 1986).

GREENSTONE GEOCHEMISTRY

Mount Holly Complex, Lincoln Massif

Three mafic dikes cut the Lincoln massif, just north of Ripton, Vermont (Delorusso and Stanley, 1986; Filosof, 1986; Stanley et al., this volume, **Stop 3**). The dikes are intrusive into felsic gneiss; one dike contains inclusions of the gneiss, and also has fine-grained apophyses chilled against the gneiss. The age of the dikes is deduced from the stratigraphy. The dikes cut the gneiss but do not intrude the overlying Pinnacle Formation, so they are pre-Pinnacle in age (Dellorusso and Stanely, 1986). The dikes are near vertical and their strike ranges from N60E to about East. The dikes are 5 to 8 meters thick and all have a well-developed foliation. All three dikes contain plagioclase feldspar, biotite, actinolite, chlorite and magnetite. Compared to other Vermont greenstones, the dikes have a higher abundance of biotite. One dike has phenocrysts of plagioclase whereas another is biotite-phyric.

Geochemically, the dikes are basalts. They have high Ti, Zr, Y, P and rare earth element (REE) concentrations (Figs. 1, 2 & 3), and can be further classified as transitional basalts. Compared to greenstones in overlying formations, they have many chemical similarities; however, they do have higher K, and lower Ti/P and Ti/Y ratios. These differences may indicate that the dikes have undergone some contamination with continental crust. The dikes near Ripton are similar to basaltic rocks of Precambrian age in the Catoctin Formation in Virginia , and the Lighthouse Cove Formation in Newfoundland. The chemistry of the Ripton dikes, like the basalts from Virginia and Newfoundland, is consistent with their formation during rifting of the proto-North American continent during the late Precambrian.

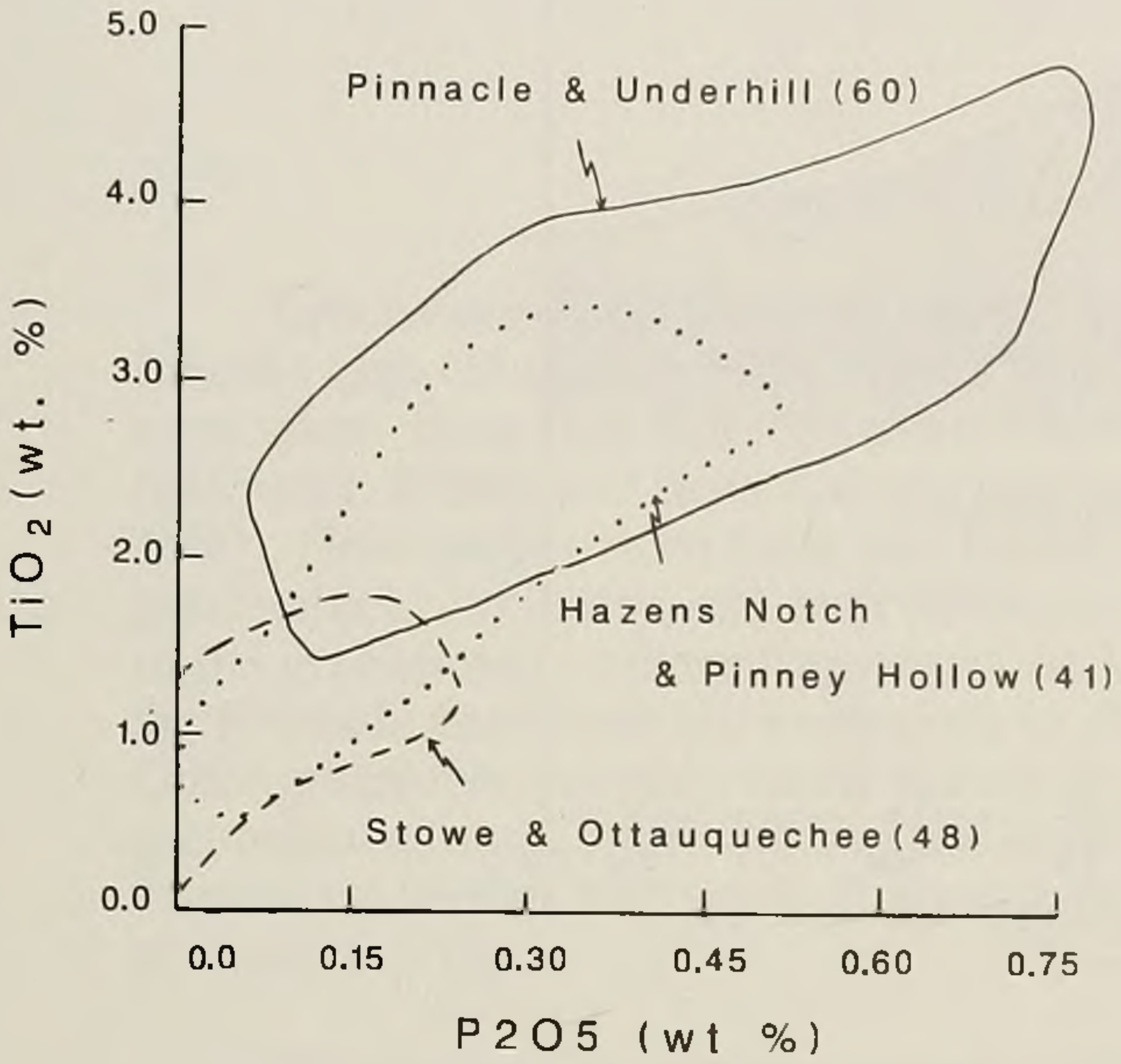

Figure 1. Variation of TiO_2 vs P_2O_5 in Vermont greenstones. # analyses in brackets.

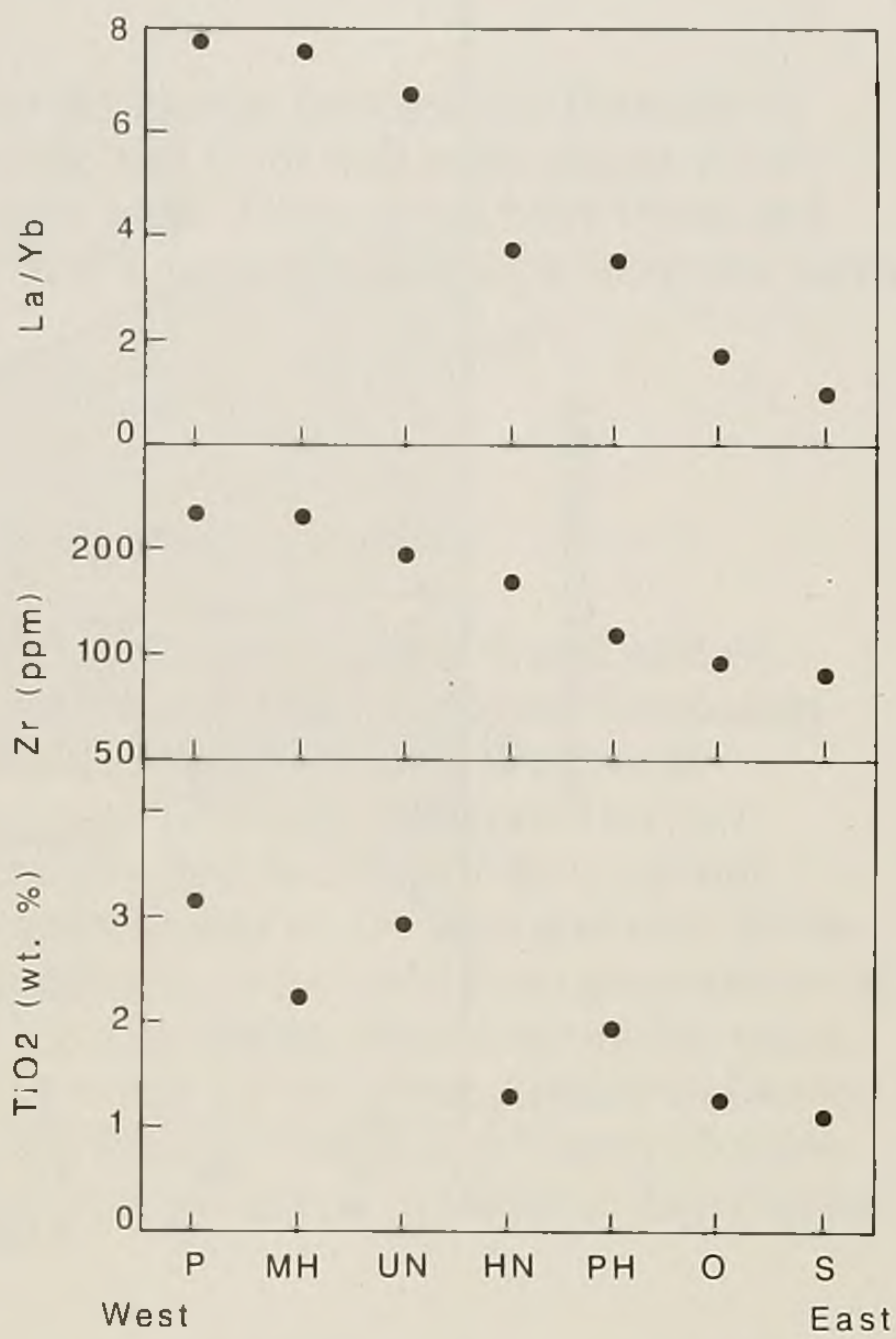

Figure 2. Average concentration of Ti Zr & La/Yb in all units.

Pinnacle and Underhill Formations

Greenstones from the Pinnacle and Underhill Formations are grouped because they are geochemically very similar, and thus, probably have the same origin. Greestones from the Pinnacle Formation have been sampled from the Tibbit Hill metavolcanics in northern Vermont (Fleming, 1980). Greenstones from the Underhill Formation are from two localities near Huntington, Vermont (Larsen, 1984; Seibert, 1984), and from six localities in the vicinity of South Lincoln (Gavigan, 1986). The Tibbit Hill outcrops are part of a more extensive unit in southern Quebec; pillow lavas are common, and relict volcanic textures are seen in thin section. These samples are interbedded with metagraywacke, including conglomeratic units, arkose and phyllite. The Huntington greenstones are coarse-grained, plagioclase-amphibole rocks that may be meta-gabbros. Greenstones near Lincoln are mostly plagioclase, chlorite, actinolite, epidote rocks. One outcrop is a 3-meter wide dike cutting pelitic schist. All greenstones from the Pinnacle and Underhill formations are in the greenschist metamorphic facies, and all contain abundant magnetite.

The chemistry of greenstones from the Pinnacle and Underhill formations indicates the samples were transitional basalts with very high concentrations of Ti, Zr, Y, and the rare earth elements (Figs. 1, 2 & 3). Furthermore, the rare earth element patterns are fractionated such that the light rare earth elements (LREE) are enriched relative to the heavy rare earth elements (HREE) (Fig. 3). These geochemical features are typical of basalts produced in early stages of continental rifting. Although there is variation in chemical

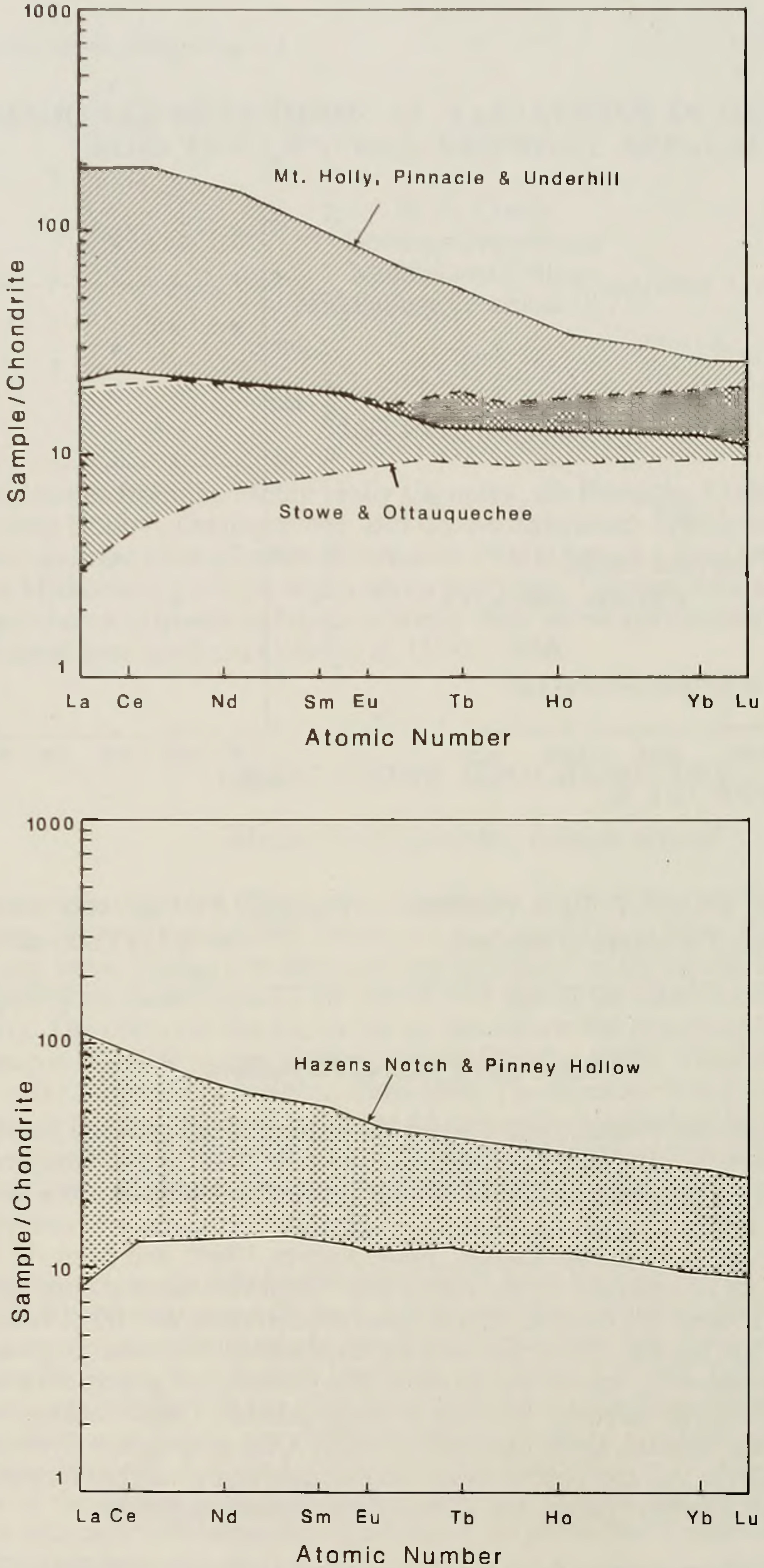

Figure 3. Rare earth element concentrations in Vermont greenstones. Stippled areas represent the range of values for formations indicated.

composition among all the samples, there is no clear difference between the Pinnacle & Underhill greenstones. This indicates that the Pinnacle and Underhill greenstones were basalts generated from the same type of mantle source rock. There could have been, and probably were, many different magmatic episodes over a wide range of time from this same mantle source.

Hazens Notch and Pinney Hollow Formations

Greenstones from the Hazens Notch Formation were sampled north and east of Lincoln Gap (Masinter, 1986). Greenstone samples from the Pinney Hollow Formation were taken from four outcrops along route 100, between Granville and Waitsfield (Masinter, 1986), and from a single outcrop along route 125, near Hancock (Poyner, 1980). Greenstones from these two formations were grouped because of their overall geochemical similarity, and their differences from greenstones to the west and east. Some of the greenstones in these formations are indistinguishable in the field from greenstones in the Pinnacle and Underhill formations in that they are dark green, chorite-actinolite rocks. Other greenstones contain much epidote and thus are much lighter green, similar in fact to greenstones found in the Stowe Formation. Thus, the Hazens Notch and Pinney Hollow formations contain two types of greenstones. At present, we see no systematic distribution of these two types.

Geochemically, the greenstones are high Ti, Zr, Y and REE basalts; concentrations of these elements overlap with the low end of the envelope for the Pinnacle and Underhill greenstones and extend to lower values (Fig. 1 & 3). The greenstones with lower values overlap greenstones from the Stowe Formation. Samples from one particular body exhibit nearly the entire chemical range shown by the whole group. This suggests either 1) magmas erupted in a single place were generated at different depths in the mantle or from different parts of a single magma chamber, or 2) the different basalts were generated in disparate regions and intimately mixed later by faulting. Whatever the reason, it is clear that the Hazens Notch and Pinney Hollow formations contain basaltic greenstones with a wide range of chemical compositions.

Ottauquechee and Stowe Formations

Greenstones from the Ottauquechee Formation were sampled near route 100 between Waitsfield and Waterbury (Doolittle, 1987). Greenstone samples of the Stowe Formation were taken from three large bodies: near Braintree, Vermont (Bailey, 1981); near Moretown (Anderson, 1983); and near Waterbury Center (Perry, 1983). All greenstones are fine-grained, light-green rocks distinct from the darker greenstones to the west. The greenstones are in the greenschist facies; a high abundance of epidote results in the light green color. Many of the greenstones exhibit a distinct mineralogical layering with bands of albite & quartz alternating with bands of epidote, chlorite & actinolite.

Greenstones from the Ottauquechee Formation are indistinguishable chemically from greenstones from the Stowe Formation. Samples from both formations are basaltic in composition and have lower concentrations of Ti, Zr, Y and REE than most greenstones from more westerly formations (Fig. 1, 2 & 3). Also, the rare earth element patterns show a slight to moderate depletion in the LREE relative to the HREE (Fig. 3), contrasting with the LREE-enriched patterns of the western formations. In almost all geochemical features, the Stowe and Ottauquechee greenstones are similar to modern mid-ocean ridge basalts.

The depleted LREE pattern indicates that the greenstones must have been derived from a different mantle source than the western greenstones; in particular, the mantle source must have been depleted mantle. On all counts, the Stowe and Ottauquechee greenstones are different from **all** greenstones in the Pinnacle and Underhill formations and **many** greenstones in the Pinney Hollow and Hazens Notch formations.

TECTONIC IMPLICATIONS

Geochemical trends in modern basaltic rocks have been used to place volcanic rocks in a particular tectonic environment, i.e, mid-ocean ridge, subduction zone, or within-plate environment. Appyling the same technique to ancient rocks, it can be shown that greenstones from the Mount Holly Complex, Pinnacle and Underhill formations formed in a within-plate environment. Furthermore, these greenstones appear to have been formed within a **continental** plate. Most modern within-continental plate volcanism occurs in regions where a plate is being stretched and pulled apart to form a rift valley. The tectonic environment of formation of the Ottauquechee and Stowe greenstones appears to be an ocean ridge. Greenstones from the Hazens Notch and Pinney Hollow formations show chemical characteristics of both within-plate **and** ocean ridge environments.

Before an attempt is made to describe tectonic events in the late Precambrian, it is instructive to look at the geographical distribution of some important elements in the greenstones (Fig. 2). In figure 2, the various formations are arranged in their current spatial distribution from west to east, and the average concentrations of TiO_2, Zr and La/Yb (a measure of LREE enrichment) are plotted. There is an obvious decreasing trend from west to east. This trend can be attributed to a change in the environment of formation of the volcanic rocks either with time if the rocks young toward the east or with distance from a rift axis if the greenstones are all approximately the same age. Whatever the interpretation, it seems that the present arrangement of the formations roughly reflects the arrangement at the time of formation; amazingly, things are not completely scrambled. It is certainly possible and perhaps likely that considerable lateral shortening has occurred, but the geochemistry of the greenstones tells us that the formations have **not** been well shuffled.

The geochemistry of greenstones in central Vermont can be explained by a fairly simple model of splitting of a continent and formation of an ocean basin. During the late Precambrian (~ 650? m.y.a), the Adirondack continent began to stretch and pull apart presumably due to a thermal plume in the mantle below. Volcanism accompanied this stretching resulting in the formation of basalts in the Mount Holly, Pinnacle and Underhill formations. The continental crust at this time was distended but still fairly thick; this can be called an early stage of rifting. Slightly later (~600? m.y.a.), volcanism continues to produce basalts further east as the zone of magmatism migrates. The continental crust is now more stretched and an ocean basin is starting to develop. This is a transitional stage in the development of ocean crust; the basalts produced here will have transitional features between true within-plate basalts and ocean ridge basalts. The basalts erupted at this time become part of the Hazens Notch and Pinney Hollow formations. Still later (~550? m.y.a.), the continental crust has been completely stretched and true sub-oceanic mantle established; ocean ridge type basalts are erupted farther east and will later become part of the Ottauquechee and Stowe formations. The ocean basin at this time could have been very small. Eventual closure of the ocean basin during the mid-Ordovician results in metamorphism of all the basalts to greenstones.

REFERENCES

Anderson, C.D., 1983, Major and rare earth element geochemistry of the Stowe Formation metavolcanics near Waitsfield, Vermont: unpub. undergraduate thesis, Middlebury College, 65 p.

Bailey, D.J., 1981, Major, trace and rare earth element geochemistry of the Brackett member greenstones: unpub. undergraduate thesis, Middlebury College, 63 p.

Coish, R.A., F.S. Fleming, M. Larsen, R. Poyner and J. Seibert, 1985, Early rift history of the proto-Atlantic ocean: geochemical evidence from metavolcanic rocks in Vermont: Amer. Jour. Sci., v. 285. p. 351-378.

Coish, R.A., D.A. Perry, C.D. Anderson and D. Bailey, 1985, Metavolcanic rocks from the Stowe Formation, Vermont: remnants of ridge and intraplate volcanism in the Iapetus ocean: Amer. Jour. Sci., v. 286, p. 1-28.

Dellorusso, V. and R.S. Stanley, 1986, Bedrock geology of the northern part of the Lincoln Massif, central Vermont: Vermont Geol. Surv., Spec. Bull. 8.

Doolittle, J., 1987, A geochemical analysis of greenstone and serpentinite in the Ottauquechee Formation: unpub. undergraduate thesis, Middlebury College, 54 p.

Filosof, A., 1986, The geochemistry of Precambrian dikes near Ripton, Vermont and implications for the tectonic evolution of Vermont: unpub. undergraduate thesis, 39p.

Fleming, F.S., 1980, Major, trace and rare earth element geochemistry of the Tibbit Hill metavolcanics, Mount Mansfield Quadrangle, Vermont: unpub undergraduate thesis, Middlebury College, 62 p.

Gavigan, T.H., 1986, Geochemistry of six South Lincoln greenstones: unpub. undergraduate thesis, Middlebury College, 62 p.

Larsen, M., 1984, Geochemistry of a greenstone in the Underhill Formation: unpub. undergraduate thesis, Middlebury College, 45 p.

Masinter, R.A., 1986, Geochemistry of the Hazens Notch and Pinney Hollow greenstones: late stages of continental rifting: unpub. undergraduate thesis, Middlebury College, 51 p.

Perry, D.A., 1983, A geochemical study of greenstones in the Stowe Formation, north of Waterbury, Vermont: unpub. undergraduate thesis, Middlebury College, 61 p.

Poyner, R.E., 1980, Geochemistry and tectonic implications of the Hancock metabasalts, central Vermont: unpub. undergraduate thesis, Middlebury College, 62 p.

Seibert, J.S., 1984, Rare earth and major elements of a greenstone body near Huntington, Vermont: unpub. undergraduate thesis, Middlebury College, 61 p.