BEDROCK GEOLOGY OF THE FAYSTON - BUELS GORE AREA CENTRAL VERMONT

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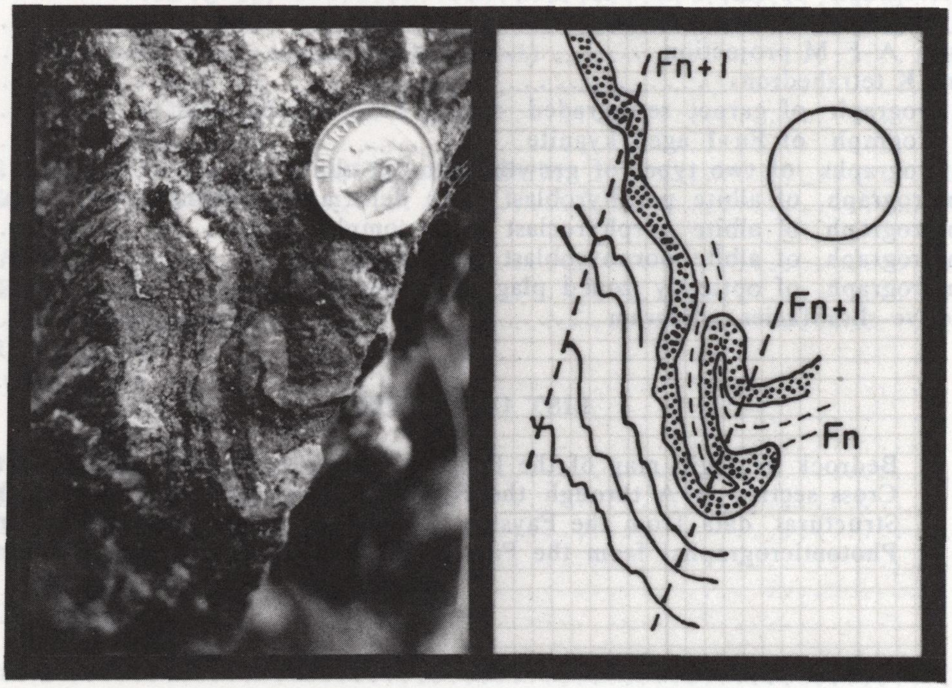

Cover photograph: Refolded quartz segregation (Fn deformed by Fn+1) in the Underhill quartz laminated schist (CZuql) from the northwestern corner of Buels Gore (E-9 on Plate 1). View is to the south (downplunge). Axial surface of Fn+1 dips to the east (left). Dime for scale.

ABSTRACT

The nine formations recognized in the Fayston - Buels Gore area by detailed bedrock geologic mapping (scale 1:12000) are, from west to east, the Underhill, Monastery, Battell, Fayston, Granville, Mount Abraham, Pinney Hollow, Ottauquechee, and Stowe. The non-carbonaceous quartz-muscovite-albite-chlorite schists of the Underhill, Fayston, Pinney Hollow, and Stowe represent a fining-eastward clastic facies change. Deposition occurred between Late Proterozoic and Cambrian times. Schists of the Mount Abraham represent a discontinuous sequence of aluminous pelagic shales atop the non-carbonaceous rocks. The carbonaceous schists and phyllites of the Battell, Granville, and Ottauquechee represent a sequence of black shales deposited from the shelf edge to outer rise. Deposition occurred during a period of anoxic ocean conditions, perhaps as the result of decreased continental glaciation and subsequently poor oceanic ventilation. Deposition occurred during the Cambrian. The breakdown of calcic plagioclase to albite and a Ca-phase in the quartz-feldspar clastic rocks is a probable origin for the abundant albite seen in the rocks today.

The geochemistry of the greenstones and amphibolites from the non-carbonaceous clastic rocks agrees with the west-to-east rift-clastic facies change model. Metabasites from the Underhill plot as within-plate and alkali basalts. The greenstones from the Fayston, Mount Abraham, and Pinney Hollow plot as transitional from within-plate to tholeiitic basalts. Stowe greenstones plot as tholeiitic basalts.

At least four generations of foliations occur in the area: A relict schistosity (Sn-1 or S0), a pervasive schistosity (Sn or S1), a crenulation cleavage (Sn+1 or S2), and kink bands (S3). Two types of Fn folds are associated with the dominant schistosity--isoclinal reclined folds and sheath folds. Regional Fn+1 folds associated with the Green Mountain anticlinorium (GMA) deform the dominant schistosity (western Sn or S1) in the western half of the area, but are transposed in the eastern half of the area by Sn+1 (S2). Open Fn+2 folds deform the regional Fn+1 folds, and are confined to the western part of the research area. At least three generations of faults occur in the area: Pre-peak metamorphic (Tn-1), syn-metamorphic (Tn), and post-peak metamorphic (Tn+1). Four Tn-1 faults are recognized: 1) Fayston over Battell and Monastery; 2) Mount Abraham (CZa2 and CZa4) over Fayston and Granville; 3) Granville over Pinney Hollow and Mount Abraham; and 4) Ottauquechee over Pinney Hollow. Many Tn faults occur in the area, although they are most common in the east. Of nine samples of Tn fault fabrics analyzed petrographically, two contained conclusive east-over-west kinematic indicators. Samples from three Tn+1 fault zones were analyzed petrographically, two contained conclusive east-over-west kinematic indicators. The Green Mountain anticlinorium is defined by large Fn+1 folds. Tn+1 (S2) faults cut the anticlinorium and are believed to be related to a later stage in the development of the GMA; the Fn+2 folds may be associated with the later phase of development. Fractures, dikes, and normal faults are associated with a minor, post-metamorphic, Mesozoic deformation event. Appalachian Gap is a Mesozoic fracture zone across the Green Mountains.

Prograde garnet and kyanite assemblages are preserved in the western part, and chlorite grade assemblages overprint the high grade assemblages in the eastern part of the area. The prograde assemblages are associated with the western Sn (S1) and the Fn+1 regional folds. The retrograde assemblages are associated with the eastern Sn (S2) which is correlative with the western Sn+1 (S2) crenulation cleavage. Analyses of porphyroclastic albite with lamellar growth twins from coarse schist and metawacke suggest that the clasts preserve relict detrital textures by epitaxial inheritance from original detrital feldspars. This supports the idea that the albitic rocks originally contained detrital quartz and feldspar.

INTRODUCTION

This bulletin includes a detailed study of the stratigraphy, geochemistry, structure, and metamorphism of the rocks across the Green Mountain anticlinorium in the vicinity of Fayston and Buels Gore, Vermont (Plate 1). The area extends south from Burnt Rock Mountain to Lincoln Mountain and east from the Huntington River to the Mad River within the towns Fayston, Waitsfield, Lincoln, Starksboro, Warren, and Buels Gore, and encompasses approximately 50 square miles.

Detailed bedrock geologic mapping was conducted at a scale of 1:12000 and compiled at 1:24000 on the U.S. Geological Survey's Waitsfield, Mt. Ellen, Huntington, and Waterbury 7.5' quadrangles. Field methods included the use of pace and compass techniques assisted by an altimeter. Lithologic descriptions are based on outcrop and hand sample appearances. Petrographic characteristics are based on the study of 142 thin sections. Various chemical and analytical procedures involved the use of an ICAP spectrometer, electron microprobe, and single crystal and powder camera x-ray diffractometers. Further details on the work presented in this report are located in an unpublished Master of Science thesis (Walsh, 1989).

Previous work in the Fayston - Buels Gore area includes mapping by Cady et al. (1962) of the 15 minute Lincoln Mountain quadrangle. Recent mapping in adjacent areas includes work to the south (Tauvers, 1982; Lapp and Stanley, 1986; O'Loughlin and Stanley, 1986; Haydock, 1988; Kraus, 1989; and Prewitt, 1989), to the east (Cua, 1989), and to the west (DiPietro, 1983).

The geology exposed in western Vermont is the remains of the ancient continental margin of North America (Bird and Dewey, 1970; and Williams, 1978). The rocks were deposited from Late Proterozoic to Middle Ordovician times over the 1.1 Ga Grenvillian crystalline basement of North America as rift-clastic, shelf, slope, and rise sediments in the proto-Atlantic or Iapetus ocean basin. During the late Early to early Middle Ordovician the Iapetus basin began to close. As the orogenic event progressed, rift to slope-rise rocks of the Taconic sequence were transported westward and emplaced on the western-Vermont eastern-New York autochthonous Cambro-Ordovician sequence (Zen, 1967, 1968; Ratcliffe, 1979; Rowley and Kidd, 1981; and Stanley and Ratcliffe, 1985). The emplacement of allochthons is believed to be the result of a westward advancing accretionary wedge during the subduction of Iapetus ocean basin and continental margin rocks of North America under the colliding Bronson Hill volcanic arc (Stanley and Ratcliffe, 1985). The occurrence of a narrow belt of ultramafic rocks in the eastern part of the pre-Silurian section is believed to be obducted fragments of Iapetan ocean crust upon which the slope-rise carbonaceous rocks were deposited (Doolan et al., 1982; Stanley et al., 1984). The Green Mountain / Sutton Mountain anticlinorium developed after tectonic emplacement of the allochthons. The anticlinorium is likely the result of duplexing or ramping of Middle Proterozoic basement off of a basal décollement (Champlain thrust fault) as the buoyant continental crust resisted transport down the subduction zone.

LITHOLOGIES

The rock units that crop out in the Fayston - Buels Gore area are described in order of their distribution from west to east. Each formation is introduced by a brief discussion on its name, geographic distribution, and members. Petrographic analyses are based on 142 oriented thin sections. For detailed petrographic descriptions refer to Walsh (1989). The section on rock types is followed by discussions on the correlation, age, and interpreted depositional environments of the units in the Fayston - Buels Gore area.

Four types of contacts are present in the research area--depositional contacts, pre-peak metamorphic faults, syn-peak metamorphic faults, and post-peak metamorphic faults. The characteristics of the contacts are briefly described below:

1. Depositional contacts: The contacts are recognized by an intercalation of two adjacent rock types. The intercalation is continuous not lensoidal. The zone of intercalation ranges in thickness from several meters to many tens of meters. The trend of the contact is commonly oblique to the trend of the dominant schistosity.

2. Pre-peak metamorphic faults: The contacts are recognized as sharp contacts that separate discrete rock units. The observation of both upper and lower plate truncations is a necessary criterion for the identification of these contacts. In places the contact is overprinted by the growth of metamorphic minerals--such a contact is referred to as metamorphically gradational as opposed to lithologically gradational. The zone of metamorphic overgrowth rarely exceeds one meter in thickness. The trend of the contact is commonly oblique to the trend of the dominant schistosity.

3. Syn-peak metamorphic faults: The contacts are always planar and parallel to the dominant schistosity, unlike the earlier faults which are folded during subsequent phases of deformation. These faults possess a mylonitic or ductile fault fabric, and are found separating discrete rock units as well as within the units themselves.

4. Post-peak metamorphic faults: The contacts are parallel to the dominant schistosity and are recognized by a zone of well-developed, closely spaced, "papery" foliation. Post-peak metamorphic faults appear to be reactivated syn-peak metamorphic faults.

Of the four types of contacts the latter two are the easiest to recognize. Depositional and pre-peak metamorphic fault contacts are the most cryptic, and their identification requires significant outcrop control. These early contacts are considered depositional until proven otherwise.

UNDERHILL FORMATION

The three members recognized within the Underhill Formation include the quartz laminated schist (CZuql), schist and metawacke (CZu), and greenstone and amphibolite (CZug). Modal analyses of the members are presented in table 1.

Quartz Laminated Schist (CZuql)

This unit is a light gray to tan and rusty, fine grained, very well foliated, quartz-muscovite-chlorite-albite$\pm$dolomite schist. Quartz and albite laminations are separated by rusty laminations that may or may not contain highly disseminated dolomite.

Schist and Metawacke (CZu)

This unit forms much of the Underhill Formation in the area. The unit contains a light gray to silvery green, fine grained, occassionally laminated, quartz-albite-muscovite-biotite-chlorite schist with abundant quartz segregations. The laminations consist of quartz and albite which commonly give the rock a pinstripe appearance. Magnetite and pyrite occur sporadically. The metawacke is a coarse grained, light grayish green quartz-albite-muscovite-biotite-chlorite schist/gneiss. The poorly developed foliation wraps around pebbles of white quartz, albite, and blue quartz up to 5 mm in diameter. Quartz segregations are not common. Porphyroblasts are absent. Cross muscovite, biotite, and chlorite are rare, but present. Magnetite and pyrite occur sporadically.

The schist grades by intercalation in and out of the metawacke which ranges from a 1 to 15 meters thick. The schist and metawacke also grade by intercalation with the greenstone and amphibolite (CZug). These gradational changes represent remnant depositional contacts.

Table 1
Average Modal Compositions of the Underhill Formation

Mineral: / Member:	Schist	Metawacke	Greenstone	Amphibolite
Quartz	50	42	1	5
Albite	14	21	22	18
Muscovite	12	18	--	--
Chlorite	6	2	20	7
Biotite	13	12	11	12
Garnet	2	--	--	--
Epidote	1	1	7	9
Clinozoisite	2	3	6	7
Tourmaline	1	Tr	--	--
Opaques *	2	Tr	5	4
Sphene	Tr	--	2	7
Calcite	--	--	1	1
Apatite	--	--	Tr	1
Hornblende	--	--	25	29
	n=1	n=1	n=2	n=2

* Opaque minerals include magnetite, ilmenite, pyrite, and chalcopyrite.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace. No petrographic analyses of the quartz laminated schist are available.

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Greenstone and Ampbibolite (CZug)

The greenstone is a fine to medium grained, light to dark green, well foliated amphibole-albite-chlorite schist. Biotite may or may not be visible in hand sample. Epidote is visible as small yellowish green blotches and small veins parallel to the schistosity. Magnetite and pyrite may or may not be visible in hand sample. Minor calcite forms weathered pits and a brown rind on some outcrops. Calcite and quartz segregations are not very abundant. The amphibolite is a very coarse to medium grained, dark green, poorly foliated amphibole-albite-biotite-epidote granulite. The amphibolite is very resistant to erosion and crops out as irregularly weathered, rounded and bulbous masses. Coarse (up to 1 cm) dark green amphiboles are randomly oriented in a finer grained plagioclase-biotite-epidote matrix. Chalcopyrite, pyrite and magnetite are readily identifiable in hand sample. Pistachio green epidote veins spotted with opaques are found up to 3 cm wide, and cut randomly across outcrops.

The Underhill greenstone is in gradational contact by intercalation with the amphibolite, metawacke, and less often with the quartz-albite-muscovite-biotite schist. All contacts appear to be depositional. The amphibolite is in gradational contact with the greenstone and the metawacke, although the contact with the metawacke is not as clearly exposed. All contacts appear depositional.

The eastern boundary of the Underhill Formation occurs where the schist and metawacke to the west are in sharp contact with the rocks of the Monastery Formation. This sharp contact is marked by syn- and post-peak metamorphic fault fabrics.

BATTELL FORMATION

The name Battell is tentatively assigned to a distinct group of graphitic rocks with limited occurrence in the study area. The name Battell was first used by Osberg (1952) to describe a black graphitic quartz-muscovite schist that contained minor lenses of white to gray dolomite and dolomitic marble found near Battell Mountain in the northern part of the Rochester 15 minute quadrangle. Osberg considered the Battell as a member of the Monastery Formation. Doll et al. (1961) describe the Battell as member of the Underhill Formation that contains carbonaceous sericite-quartz-albite-chlorite schist, schistose quartzite, carbonaceous and non-carbonaceous limestone, and quartz-sericite-chlorite-albite schist.

Armstrong (in progress) raises the Battell to formation status to describe graphitic schists with carbonates that depositionally overlie the Monastery Formation in the Granville-Hancock area of central Vermont. Armstrong describes an upper and lower sequnce in the Battell. The upper part of the Battell is carbonaceous albitic schist with local lenses and blocks of tan to gray dolomite and limestone. The schist grades stratigraphically downward into the lower part of the Battell--the White River Member. The top of the White River Member includes carbonaceous schist with black dolomitic marble. The White River Member progresses stratigraphically downward as a gradational sequence of intercalated Monastery schist, graphitic schist, and minor zones rich in tourmaline.

Using Armstrong's criteria to define the Battell, only the White River Member (Cbw) was seen in the Fayston - Buels Gore area. The upper part of the Battell with carbonaceous schist and tan and gray dolomitic marbles is not observed in the research area. The White River Member (Cbw) contains the carbonaceous schist with black dolomitic marble. Modal analyses of the White River Member of the Battell Formation are given in table 2.

The White River Member consists of dark gray to rusty carbonaceous muscovite-quartz-albite schist, white to black quartzites, and black dolomitic marble. Quartz segregations are common. The black dolomitic marble occurs as discontinuous lenses up to 20 cm thick and 1 m long. It is brecciated and cross cut by veins of dolomite, quartz, and calcite and always enclosed in carbonaceous schist.

The White River Member (Cbw) appears to be in fault contact with the Underhill Formation along the eastern boundary of the Underhill in Buels Gore. This contact has mylonites and paper schists along it and appears to have a complex history. The White River Member appears to be in depositional contact with the Monastery Formation at all observed locations. The White River Member always occurs as small bodies within the schists of the Monastery Formation.

MONASTERY FORMATION

The name Monastery is tentatively assigned to a group of distinctive, primarily non-graphitic rocks in Domain 4 and the western part of Domain 3. Osberg (1952) uses the name Monastery to describe rocks that occur between the Mount Holly complex below, and the Granville Formation above. Osberg's Monastery contains three distinct units: A pale green quartz-chlorite-muscovite schist (Monastery schist), a basal conglomerate and schistose sandstone (Tyson Member), and graphitic schist (Battell Member). Locally the Monastery schist contains 1-2 mm porphyroblasts of albite, sporadic garnet and chloritoid schists, schistose quartzite, and arenaceous porphyroblastic albite-quartz-muscovite schist. Doll et al. (1961) include Osberg's Monastery in the Underhill Formation. Armstrong (in progress) uses the name Monastery Formation to describe albitic schist with fine grained metawacke and very minor conglomerate, quartz-muscovite-chlorite-albite-biotite schist, local lenses of dolarenitic schist, and local lenses of aluminous schist with garnet and chloritoid. Armstrong resurrects the name Monastery because the rocks are distinguishable from rocks in the Underhill and Fayston Formations. Rocks mapped as Monastery Formation in Fayston and Buels Gore are similar to

Table 2

Average Modal Compositions of the White River Member of the Battell Formation and Schists of the Monastery Formation

	CZm		Cbw	
Mineral:	Muscovite Garnet Schist	Muscovite Chlorite Quartz Schist	Graphitic Schist	Black Marble
Quartz	9	17	28	9
Albite	4	9	6	--
Muscovite	55	46	54	13
Chlorite	6	18	5	--
Biotite	--	4	2	--
Garnet	23	6	--	--
Epidote	--	Tr	--	--
Tourmaline	Tr	Tr	--	--
Opaques *	2	Tr	1	--
Sphene	--	--	Tr	--
Dolomite	--	Tr	1	70
Calcite	--	--	--	Tr
Apatite	1	--	--	--
Graphite	--	--	3	8
	n=2	n=2	n=1	n=1

* Opaque minerals include ilmenite and magnetite.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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those mapped as Monastery by Armstrong in the Granville-Hancock area.

Schist (CZm)

The Monastery schist is a heterogeneous unit consisting of a variety of rock types. The rock types include tan sandy quartz-muscovite-biotite-albite±garnet±graphite schist; silvery tan, fine grained, pearly sericite schist; silvery green muscovite-chlorite-quartz-albite schist; tan garnetiferous muscovite schist, thin (up to 20 cm) light gray and tan quartzite; and tan to rusty weathering pods and disseminated grains of dolomite and ankerite. Muscovite and biotite micas oriented across the plane of the schistosity are quite abundant; cross chlorite, although present, is not as abundant. Garnet porphyroblasts up to 1 cm in diameter comprise up to 30% of the tan garnetiferous muscovite schist. The unaltered garnets are resistant to weathering and stick out of the rock like the "chips" in a chocolate chip cookie. Magnetite and pyrite occur sporadically. The schist contains intercalated dark gray to black, rusty, carbonaceous muscovite-quartz-albite schist with white to black quartzites near contacts with Cg and Cbw.

The different rock types were not subdivided and mapped as separate units because the schists are intercalated on a scale too small to be mapped at 1:12000. Changes in lithology occur gradually from anywhere within 10 m to 10 cm. These changes in lithology are interpreted to represent sedimentological changes across relict depositional contacts. The Monastery is in gradational contact by intercalation with the White River Member of the Battell Formation. The contact between the Monastery schists and the white albitic schist of the Fayston Formation (CZf) is the most gradational contact observed anywhere in the Fayston - Buels Gore area. The Monastery and Fayston schists are intercalated along the contact in a zone that ranges from less than one meter to many tens of meters wide. On the west side of the Monasterty Formation a planar contact with the Underhill Formation has both paper schists and mylonites and is interpreted to be a reactivated fault with a complex history.

Greenstone (CZmg)

The greenstone is a fine grained, dark green, rusty weathering, well foliated chlorite-albite-calcite schist. Abundant chlorite gives the rock its dark green color. Albite porphyroblasts up to 2 mm in diameter are easily identified. Pyrite cubes are often visible. Carbonate and quartz segregations are common, but not ubiquitous. A brown rusty rind is visible on weathered samples. The greenstone is in gradational contact by intercalation with the Monastery schists.

FAYSTON FORMATION

A current problem in the stratigraphic nomenclature is the use of both the names Underhill and Hazens Notch to describe the same coarse muscovite-quartz-albite-chlorite schist. The name Fayston Formation is proposed by Walsh (1989) to describe a group of rocks in which the primary unit is the muscovite-quartz-albite-chlorite schist that contains abundant porphyroblasts of white to light gray albite. This schist is commonly referred to as the white albitic schist. Christman and Secor (1961) included this rock type within the Underhill Formation as part of their eastern sequence, but they did not delineate the difference between the eastern and western sequence on their map. In this study the name Underhill is applied only to rocks similar to Christman and Secor's western sequence. Cady et al. (1962) included the white albitic schist in both the Underhill and Hazens Notch Formations as a non-carbonaceous schist. The assignment of the white albitic schist to both formations was based on the interpretation that the Mount Abraham Schist was stratigraphically between the Underhill and Hazens Notch Formations. According to the interpretation, the upward stratigraphic succession was from Underhill into the Mount Abraham Schist, and from Mount Abraham into the Hazens Notch. Work by O'Loughlin (1986), Lapp (1986), Haydock (1988), and Prewitt (1989) indicates that the sequence is, in part, tectonic and that the simple stratigraphic succession does not exist.

The name Underhill applies to rocks exposed in the western part of the study area. Thompson and Thompson (1987) use the name Underhill to refer to the white albitic schist with magnetite. More recently the Thompsons suggested the name Huntington Member for the western sequence, and Camels Hump Member for the eastern sequence of the Underhill Formation (Thompson and Thompson, 1989). The name Camels Hump Member is inappropriate because the name Camels Hump Group is well established in the literature as referring to the Underhill, Hazens Notch, Mount Abraham, and Pinney Hollow Formations. Previous workers to the south (O'Loughlin, 1986; Lapp, 1986; Haydock, 1988; Prewitt, 1989) assigned the white albitic schist to the Hazens Notch Formation (CZhn) along with carbonaceous albitic schists (CZhnc and CZhnca). The name Hazens Notch, however, is currently one of the big problems in stratigraphic nomenclature for Vermont geologists. The Hazens Notch carbonaceous and non-carbonaceous schist are associated with ultramafics, mafic schists, and blueschists in northern Vermont (Chidester et al., 1978; Doolan et al., 1982; Stanley et al., 1984; and Bothner and Laird, 1987). In the Camels Hump 15 minute quadrangle Thompson and Thompson (1987) consider the Hazens Notch to be strictly a carbonaceous albitic schist with associated mafic schist. The use of the name Hazens Notch, therefore, is not recommended for the white albitic schist of the Fayston - Buels Gore area.

To avoid the problems associated with the use of current names the white albitic schist has been renamed the Fayston Formation. The name Fayston Formation has never been used, thus any ambiguities associated with the names Underhill and Hazens Notch are avoided.

The white albitic schist (CZf) constitutes approximately 90% of all bedrock exposure of the Fayston Formation (type locality: French Brook elev. 1120 feet, Plate 1, R-8). Other units include the quartzo-feldspathic granofels (CZfq), quartz-biotite gneiss (CZfb), quartz-muscovite-tourmaline schist (CZft), and greenstone (CZfg). The occurrence of these units is minimal, yet significant. Because of their significance they are tentatively assigned member status. Modal analyses of the members of the Fayston Formation are presented in table 3.

White Albitic Schist (CZf)

The white albitic schist is a silvery green, medium grained muscovite-quartz-albite-chlorite schist. Large white to light gray albite porphyroblasts up to 7.5 mm in diameter are easily recognizable in hand sample and outcrop. Coarse mica and chlorite define the schistosity. Quartz is granular and also occurs as numerous segregations. Chlorite streaks and quartz rods define a pronounced lineation on the foliation surface. Small (up to 3 mm) euhedral garnet porphyroblasts and large (up to 7.5 mm) anhedral magnetite porphyroblasts may or may not be present. Cross biotite and muscovite grains are common to the west in domain 3. Thin (5 to 15 cm) white to light gray quartzites are found within the white albitic schist.

The white albitic schist (CZf) is in depositional contact with all of its members (CZfq, CZfb, CZft, and CZfg). The white albitic schist is in very sharp contact with the kyanite-bearing Mount Abraham schist (CZa4) along a pre-peak metamorphic fault. In places albite porphyroblasts overgrow the contact in a zone one meter wide. To the east in Domain 2 the white albitic schist is in sharp and gradational contact with the Granville carbonaceous albitic schist (CZg), and in sharp contact with the Mount Abraham schist (CZa2). The sharp contacts are well defined pre-, syn-, and post-peak metamorphic faults. The gradational contacts are intercalated in zones that varies from approximately 1 to 10 meters wide, and are believed to be depositional.

To the west in Domain 3 the relationship between the white albitic schist and the other lithologies is quite complex. At Mad River Glen Ski Area the white albitic schist has graphitic segregations close to contacts with the carbonaceous albitic schist of the Granville. The contacts with the carbonaceous albitic schist are gradational by intercalation, and small, isolated lenses of Cg within CZf are common in the center of Domain 3. The presence of a black quartzite within the white albitic schist near a gradational contact with the carbonaceous albitic schist and schists of the Monastery Formation (M-15) strongly suggests a depositional relationship between these units, because such black quartzites are normally confined to the graphitic rocks. Within Domain 3 the white albitic schist is in contact with the allanite-bearing Mount Abraham schist (CZa3) along a gradational contact (K-15). The contact is marked by intercalation of the two rock types over a distance of 3 or more meters. The only occurrence of allanite within the white albitic schist is near this contact. The allanite grains are anhedral and pre-date the dominant schistosity. Perhaps the occurrence of allanite in the white albitic schist near this contact is due to the original mixing of the sediments. If this is the case it would support the idea that this contact is depositional.

Quartzo-Feldspathic Granofels (CZfq)

The quartzo-feldspathic granofels is a coarse grained, light gray albite-quartz-muscovite-chlorite granofels. The rock is massive with a poorly defined foliation. The coarse texture is due to abundant ($\approx$50 %) interlocking albite porphyroblasts up to 3 mm in diameter.

Table 3

Average Modal Compositions of the Fayston Formation

Member:	CZf Schist	CZf Quartzite	CZfq	CZfb	CZft	CZfg
Mineral:						
Quartz	24	75	27	47	31	6
Albite	23	3	46	29	20	36
Muscovite	35	16	14	4	29	--
Chlorite	12	2	9	--	7	41
Biotite	1	1	1	17	--	--
Garnet	1	1	Tr	Tr	--	--
Epidote	Tr	Tr	1	1	--	4
Clinozoisite	Tr	Tr	Tr	Tr	--	2
Tourmaline	1	1	Tr	Tr	9	--
Opaques *	3	1	2	1	4	2
Sphene	Tr	--	--	Tr	--	2
Carbonate **	Tr	1	--	--	Tr	7
Apatite	Tr	--	Tr	--	--	--
Actinolite	--	--	--	--	--	Tr
Graphite	Tr	--	--	--	--	--
Allanite	Tr	--	--	--	--	--
Zircon	Tr	--	--	--	--	--
Alteration #	Tr	--	--	--	--	Tr
	n=18	n=2	n=3	n=2	n=2	n=2

* Opaque minerals include magnetite, ilmenite, and pyrite.
** Carbonate includes calcite and dolomite.
\# Alteration includes limonite or rutile.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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Quartz, muscovite, and chlorite fill the spaces between the porphyroblasts. Magnetite may or may not be present. The quartzo-feldspathic granofels appears very homogeneous in outcrop and hand sample. Quartz segregations are absent. The rock is resistant to weathering and has a blocky appearance.

The quartzo-feldspathic granofels is in depositional contact with the white albitic schist, and to a lesser extent the quartz-biotite gneiss. A large mappable unit of the quartzo-feldspathic granofels also occurs in proximity to a gradational contact between the white albitic schist and the allanite-bearing Mount Abraham schist (CZa3; K-15). The granofels also appears to be in depositional contact with rocks of both the Monastery and Fayston Formations just south of Burnt Rock Mountain (N-4), thus supporting the idea that the two formations are in depositional contact.

Quartz-Biotite Gneiss (CZfb)

The quartz-biotite gneiss is a medium to coarse grained quartz-albite-biotite gneiss. The term gneiss is preferred over schist because the rock usually is very massive, coarse grained, and has a pronounced foliation. The foliation of the rock varies from a moderately developed schistosity to a poorly developed gneissosity. The color of the rock is a mixture of white and black that resembles salt and pepper. Quartz and biotite are easily identified in hand sample and outcrop, but albite is not as easily identified because it does not occur as porphyroblasts. Pyrite and magnetite may or may not be present. Quartz segregations are rare or absent. The rock is resistant to weathering and has a blocky appearance.

The quartz-biotite gneiss is in depositional contact with the white albitic schist, and to a lesser extent the quartzo-feldspathic granofels. The quartz-biotite gneiss with associated white albitic schist also occurs in close proximity to the allanite-bearing Mount Abraham schist. A single exposure of the quartz-biotite gneiss occurs within a pre-peak metamorphic thrust slice of the Mount Abraham schist (CZa2) in a tributary of French Brook. The contact between the two units is not exposed, and the gneiss is interpreted to be a window through the thrust slice. It appears that the quartz-biotite gneiss occurs near transitional contacts between the white albitic schist and the allanite-bearing Mount Abraham schist (CZa3) in Domain 3.

Quartz-Muscovite-Tourmaline Schist (CZft)

The tourmaline-bearing unit is a medium grained, dark silvery gray to rusty colored, quartz-muscovite-albite-tourmaline-chlorite schist. The rock has a well developed schistosity. Quartz occurs as granules, thin laminations, and quartz vein segregations. Muscovite is silvery tan to rusty colored. Tourmaline occurs as black euhedral crystals up to 1 mm in diameter and 1.5 cm in length. Tourmaline crystals are scattered randomly in the plane of the schistosity as well as across the schistosity--the crystals have no preferred orientation. Albite and chlorite are not easily identified in hand sample and outcrop. The quartz-muscovite-tourmaline schist is in gradational contact by intercalation with the white albitic schist.

Greenstone (CZfg)

The greenstone is a fine grained, dark green, rusty weathering, well foliated chlorite-albite-calcite schist. Abundant chlorite gives the rock its dark green color. Albite porphyroblasts up to 2 mm in diameter are easily identified. Pyrite cubes are often visible. Carbonate and quartz segregations are common, but not ubiquitous. A brown rusty rind is visible on weathered samples. The Fayston greenstone is in depositional contact with the white albitic schist.

GRANVILLE FORMATION

Osberg (1952) proposed the name Granville Formation to describe a graphitic quartz-muscovite schist with occasional interbeds of blue-gray quartzite and buff colored dolomite. Armstrong (in progress) includes the graphitic schist and dolomite in the Battell Formation, and uses the name Granville for carbonaceous albitic schists with quartzites but no carbonates. Cady et al. (1962) mapped carbonaceous and non-carbonaceous quartz-sericite-chlorite-albite schist with quartzite greenstone, and calcareous marble as the Hazens Notch Formation. O'Loughlin (1986), Lapp (1986), and Haydock (1988) subdivided Cady and other's Hazens Notch by mapping each rock type separately. Haydock (1988) demonstrates that the greenstone and calcite marble (actually calcareous greenstone) occur solely within the white albitic schist member. Due to the aforementioned problems with the name Hazens Notch, O'Loughlin's, Lapp's, and Haydock's carbonaceous albitic schist (CZhnca) and super carbonaceous schist (CZhnc) members of the Hazens Notch are now included with the Granville Formation. The super carbonaceous schist is given the name Lincoln Gap Member, a name first proposed by Gordon (1927). The Lincoln Gap area is where Gordon (1927), O'loughlin (1986) and Haydock (1988) recognized the super carbonaceous schist.

The carbonaceous albitic schist with quartzite (Cg) and the super carbonaceous schist of the Lincoln Gap Member (Cgl) are the two members of the Granville Formation. Modal analyses of the Granville Formation are given in table 4.

Table 4

Average Modal Compositions of the Granville Formation

Mineral:	Cg Schist	Cg Quartzite	Cgl Schist	Cgl Quartzite
Quartz	33	91	47	51
Albite	16	2	11	6
Muscovite	38	7	24	14
Chlorite	7	Tr	4	--
Biotite	--	--	4	--
Garnet	1	--	--	--
Epidote	Tr	Tr	--	--
Tourmaline	Tr	--	--	--
Opaques *	1	Tr	Tr	--
Sphene	Tr	--	Tr	--
Calcite	--	--	Tr	--
Apatite	Tr	--	Tr	--
Graphite	4	Tr	10	29
Limonite	Tr	--	Tr	--
	n=5	n=4	n=4	n=2

* Opaque minerals include ilmenite and pyrite.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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Carbonaceous Albitic Schist (Cg)

The carbonaceous albitic schist is a dark to rusty colored, muscovite-quartz-albite-chlorite-graphite schist. The schist is characterized by discontinuous patches of graphite on foliation surfaces. Gray to black albite porphyroblasts up to 5 mm in diameter, and quartz segregations are ubiquitous. Discontinuous white, black and white, gray, and black quartzites ranging from 5 to 30 cm in thickness and up to 100 m in length occur throughout the schist, but are not present at every outcrop. Pyrite may or may not be present as very small weathered cubic crystals or crystal aggregates.

The carbonaceous albitic schist is in depositional contact with its associated quartzites. Two important contact relationships exist between the carbonaceous albitic schist (Cg) and the other rock units in the Fayston - Buels Gore area. The large body of carbonaceous albitic schist in Domain 2 appears to be a pre-peak metamorphic thrust slice. Greenstones in the lower plate of Mount Abraham and Pinney Hollow, and quartzites in the upper plate of Granville are truncated along sharp pre-metamorphic contacts. This large unit of carbonaceous albitic schist is bound by syn- and post-peak metamorphic faults as evidenced by the occurrence of numerous mylonites and paper schists. The evidence suggests that this body is a pre-peak metamorphic thrust slice that has undergone subsequent shearing.

Occurrences of the carbonaceous albitic schist in the western part of Domain 2 do not always have clearly defined contact relations. Where the contacts are observed in this area the schist is primarily in gradational contact with the white albitic schist. Due to the paucity of marker horizons such as greenstones and quartzites, truncations can not be demonstrated. The gradational contacts are such by intercalation and are believed to be depositional. All observed contacts between the Mount Abraham (CZa2) and the carbonaceous albitic schist are very sharp, although truncations are not observed. No gradational contacts of any sort were observed between the Mount Abraham schist (CZa2) and the carbonaceous albitic schist in Domain 2. Gradational contacts between the Mount Abraham (CZa3) and the carbonaceous albitic schist do occur in Domain 3 in similar locations that the graphitic schist is intercalated with the white albitic schist (CZf) and the schists of the Monastery Formation.

Lincoln Gap Member (Cgl)

The Lincoln Gap Member consists of black, rusty weathered quartz-muscovite-albite-graphite schist, and up to 5 cm thick quartz-graphite-muscovite-albite quartzite. The schist is referred to as the super carbonaceous schist because of the abundance of graphite. Graphite occurs throughout the rock unlike the patchy occurrence in the carbonaceous albitic schist. The quartzite is interlayered with the schist and often looks like a black quartzose schist or slate because the foliation is not as closely spaced as in the super carbonaceous schist. Light colored quartzites are not present. Small (up to 2 mm) albite porphyroblasts are rare in hand sample and outcrop. Quartz vein segregations are common. Pyrite may or may not be present. The super carbonaceous schist is not very resistant and large continuous outcrops are not very common.

The three mapped bodies of the Lincoln Gap Member are in contact with the Mount Abraham Schist (CZa2 and CZa4). The Lincoln Gap Member on Stark Mountain appears to be gradational by intercalation with the Mount Abraham Schist (CZa4). This relationship is most evident on the Paradise ski trail at Mad River Glen Ski Area and at the junction of the Jerusalem and Long Trails. The Lincoln Gap Member is on top of the Mount Abraham at the summit of Stark Mountain. The contact relationships at Fayston Farms subdivision (Q-14) are not clear. The closest outcrops of super carbonaceous schist and Mount Abraham schist (CZa2) are 50 meters apart. Fn fold data indicate, however, that the Lincoln Gap Member is a synform atop the Mount Abraham thrust slice.

MOUNT ABRAHAM SCHIST

Four distinctive units of Mount Abraham Schist (CZa1-4) occur within the Fayston - Buels Gore area. The numbers 1 through 4 are suffixes assigned to the lithic designator CZa in order of the units' distribution from east to west. The units are recognized on the basis mineralogy, contact relationships, and geographic distribution.

Cza1 is white mica (muscovite and paragonite)-quartz-chlorite-chloritoid schist that contains intraformational greenstone (CZag) and metawacke (CZaw). CZa2 is similar to CZa1, but contains no greenstone or metawacke. CZa3 is similar to CZa2, but also contains an appreciable amount of allanite. CZa4 contains kyanite. Modal analyses for the units of the Mount Abraham Schist are given in table 5.

Mount Abraham Schist (CZa1)

CZa1 is a fine grained, silvery gray-green to silvery tan, white mica (muscovite and paragonite)-quartz-chlorite-chloritoid-garnet schist. White mica is silvery to tan and very fine grained. The fine grained nature of the mica gives the foliation surfaces a characteristic pearly sheen in which individual grains of mica are not discernable. The sheen is due to the presence of intergrown paragonite with the muscovite, and along with the absence of albite, is the best field criteria for determining the possible presence of paragonite. Quartz occurs as granular laminations and abundant segregations. Chlorite and white mica are finely interlayered. Chlorite streaks are the common mineral lineation, especially where garnet is altered. Chloritoid occurs as very small (up to 1 mm) porphyroblasts that may or may not be visible in hand sample. Fine grained magnetite and ilmenite can be confused with chloritoid.

Table 5

Average Modal Compositions of the Mount Abraham Schist

Mineral:	CZa1	CZaw	CZag	CZa2	CZa3	CZa4
Quartz	34	39	2	30	18	22
Albite	--	27	10	--	--	--
White Mica *	46	15	--	51	58	58
Chlorite	13	Tr	18	11	14	7
Chloritoid	4	--	--	5	6	6
Biotite	--	14	--	--	--	--
Garnet	1	--	--	1	2	1
Kyanite	--	--	--	--	--	2
Opaques **	2	Tr	Tr	2	2	4
Epidote	Tr	5	26	Tr	Tr	Tr
Clinozoisite	Tr	Tr	12	Tr	Tr	Tr
Allanite	Tr#	--	--	Tr#	Tr	Tr#
Tourmaline	Tr	--	--	Tr	Tr	Tr
Sphene	--	Tr	Tr	Tr	--	--
Calcite	--	--	9	--	--	--
Apatite	Tr	Tr	--	Tr	Tr	Tr
Actinolite	--	--	23	--	--	--
Graphite	--	--	--	--	--	Tr
Rutile	Tr	--	--	Tr	Tr	Tr
	n=5	n=2	n=1	n=11	n=3	n=4

* White mica includes muscovite and paragonite in the schists (CZa1-4), and muscovite only in the metawacke (CZaw).
** Opaque minerals include magnetite, ilmenite, hematite and pyrite.
# Allanite is present in only one sample.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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Garnet porphyroblasts up to 3 mm in diameter are only rarely pristine and euhedral, and are usually partially altered to chlorite and recognized as clots or bumps on the foliation surface. Garnet is not ubiquitous. Small (up to 2 mm) magnetite octahedra or anhedral porphyroblasts are often present. A schist which is transitional in composition and location between CZa1 and CZph is a fine grained, tannish silvery-green, quartz-muscovite-chlorite-albite-garnet-chloritoid schist. Quartz, muscovite, and chlorite are distinct. Albite and chloritoid are rarely visible in hand sample. Garnet is ubiquitous, but is most often recognized by chlorite clots or bumps on foliation surfaces. Quartz segregations are abundant.

CZa1 is in depositional contact with its greenstone (CZag) and metawacke (CZaw) members. The schist is also in depositional contact with the fine grained muscovite-quartz-albite-chlorite schist of the Pinney Hollow Formation. The contact between the Mount Abraham and Pinney Hollow is quite gradational and the schists are intercalated in a zone that ranges from 10 to 25 meters in width. In this zone the transitional schist appears as a combination between typical Mount Abraham and fine grained Pinney Hollow. Small greenstone bodies also occur within the transitional schist.

Greenstone (CZag)

The greenstone is a fine grained light green to light yellow epidote-actinolite-chlorite schist. The greenstone often has a green and yellow banded appearance due to compositional layering of actinolite and chlorite with epidote minerals. Albite porphyroblasts up to 1 mm in diameter are rare. In places calcite veins are present. The rock is not well foliated and is more resistant to erosion than the surrounding schist. These greenstones mark the first reported occurrence of metaigneous rocks in the Mount Abraham Schist (Walsh and Stanley, 1988). The greenstone is in contact by intercalation over a distance of 0.5 to 2 meters with the Mount Abraham Schist (CZa1) and the metawacke (CZaw). The contact is believed to be depositional.

Metawacke (CZaw)

The metawacke is a homogeneous, coarse grained, weakly foliated, greenish gray quartz-albite-muscovite-biotite granofels. Granular quartz is visible in hand sample. Albite porphyroblasts are small (up to 1 mm) and ubiquitous. Micaceous partings of muscovite and biotite are silvery and greenish brown respectively. Quartz segregations are characteristically absent. Magnetite may or may not be present. The rock is very resistant to erosion and breaks off in blocks not slabs.

The Mount Abraham metawacke (CZaw) is intercalated with the schist (CZa1) and the greenstone (CZag). The metawacke always occurs in proximity to the greenstone. The contacts are believed to be depositional.

Mount Abraham Schist (CZa2)

CZa2 is very similar to CZa1. Outcrops of CZa2 have a steel blue-gray color. This characteristic color is more evident in CZa2 than in CZa1. The white mica sheen is more consistent from outcrop to outcrop than it is in CZa1. Anhedral magnetite porphyroblasts up to 2 cm in diameter are present, but smaller porphyroblasts are most common. Magnetite is not ubiquitous. Anhedral aggregates of pyrite are rare. Albite is characteristically absent. CZa2 is more homogeneous than CZa1 because it is not transitional with any units. Greenstone and metawacke are absent.

The Mount Abraham schist (CZa2) is in sharp pre-peak metamorphic contact atop the white albitic schist of the Fayston Formation (CZf) and the carbonaceous albitic schist of the Granville Formation (CZg). CZa2 is a pre-peak metamorphic thrust slice that has subsequently been faulted by syn- and post-peak metamorphic faults. The Lincoln Gap Member of the Granville Formation (CZgl) is above the Mount Abraham schist.

Mount Abraham Schist (CZa3)

The Mount Abraham Schist CZa3 is similar to CZa1 and CZa2, for this reason only the differences are described. The color of the white mica sheen is generally a silvery white rather than a tan. Tan foliation surfaces are present, but they are not as common as in CZa1 and CZa2. Garnet porphyroblasts up to 2 cm in diameter occur in CZa3--the largest garnets seen in the Mount Abraham Schist in the Fayston -Buels Gore area. Garnets are almost always pristine and are only slightly altered to chlorite. Anhedral magnetite porphyroblasts are common, but they do not reach the size (2 cm) of those found in CZa2.

The major petrographic distinction between CZa3 and the rest of the Mount Abraham schists is the presence of allanite. Allanite occurs in all three samples of CZa3 analyzed, and in one sample (590 GW) it comprises 1 % of the rock. Allanite grains are small (<0.3 mm), and generally anhedral except when they occur as euhedral inclusions in garnet. Many grains are partially altered to clinozoisite or clinozoisite and epidote. Alteration is in the form of cross cutting veins, and as reaction rims around partially eroded euhedral crystals. Because the alteration products are oriented parallel to the dominant schistosity the allanite grew before the development of the dominant schistosity. Some garnets in CZa3 contain allanite inclusions with sigmoidal inclusion trails that record an older Sn-1 schistosity.

Allanite has the following observed optical properties: Allanite is yellow to dark brown and often metamict. Pleochroism is weak to moderate in non-metamict grains, and absent in metamict grains. Allanite causes pleochroic haloes in neighboring chlorite due the decay of thorium. Relief is moderate and similar to epidote, but higher than chloritoid. Birefringence ranges from upper first to lower second order colors, but the colors are usually poorly developed due to the partially metamict nature of most crystals. Allanite is biaxial positive with a high 2V angle, but good interference figures are rare. Diamond shaped sections {010} yield poor optic axis figures. Cleavage on {001} or {100} is only seen on sections viewed perpendicular to the Y direction.

The Mount Abraham Schist (CZa3) is in depositional contact with the white albitic schist of the Fayston Formation (CZf). The contact is gradational over a distance of 1-3 meters where the two rock types are intercalated. CZa3 is also intercalated with the carbonaceous albitic schist (Cg) (M-16). CZa3 is apparently in depositional contact with the quartz-biotite gneiss (CZfb) and the quartz-muscovite-tourmaline schist (CZft) of the Fayston Formation, although the contacts are never clearly exposed. The map pattern and distribution of the rock types, however, indicates that the rocks are closely associated. Another possibility is that the discordant map pattern of CZa3 in this area is due to an an early fault, although no evidence of faulting is observed. These relationships are seen in the eastern part of Domain 3 where the complicated map pattern is due, at least in part, to complex sedimentary changes. CZa3 is also in syn-peak metamorphic fault contact with rocks of the White River Member of the Battell Formation and rocks of the Fayston and Granville Formations in Domain 3. In Domain 4 CZa3 is in depositional contact with both the schists of the Monastery and Fayston Formations.

Mount Abraham Schist (CZa4)

Outcrops of kyanite-bearing Mount Abraham schist (CZa4) have a characteristic silvery, steel blue-gray appearance. White mica again gives a characteristic pearly sheen to foliation surfaces. Tan colored foliation surfaces are not as common as in CZa1 and CZa2. Kyanite is generally visible in hand sample where the blades are usually 0.5 to 1.0 cm in length and up to 0.5 mm in diameter. Unaltered kyanite is white, light blue, and light green, but rusty kyanite blades are more common. The kyanite blades are randomly oriented on the foliation. Chloritoid porphyroblasts are usually similar to those in the other schists of the Mount Abraham. One sample was found, however, with large 1.0 cm chloritoid porphyroblasts. Lapp (1986) reported a similar occurrence in the Mount Abraham Schist in the Mount Grant area. The chloritoid is dark blue-green and the {001} cleavage is easily discernible. Garnet porphyroblasts are generally small (up to 3 mm) and usually unaltered. Small (up to 2 mm) anhedral magnetite porphyroblasts are relatively rare. Quartz segregations are abundant.

Petrographically, kyanite porphyroblasts are colorless to light blue with weak pleochroism. Birefringence is upper first order and usually mottled or uneven. Kyanite is commonly euhedral, and has moderately high relief and discernible {001} and {010} cleavages. Kyanite is usually unaltered or only slightly altered to white mica. Garnet in CZa4 is quite different from garnet in the other Mount Abraham schists. The garnet that is present with chloritoid, but without kyanite (1 of 4 samples), occurs as pristine, euhedral porphyroblasts with essentially no alteration, and is similar to garnet in CZa3. The garnet with kyanite, however, has a drastically different appearance in that it is not euhedral, and it has numerous inclusions of quartz that give the porphyroblast a sieve texture. Only minor amounts of chlorite, muscovite, and opaques occur around such garnets. Kyanite in the vicinity of such garnets appears relatively fresh. It appears as if the garnet in the samples with kyanite is breaking down during a prograde reaction, and not retrograded like garnet in samples of the Mount Abraham farther to the east.

The kyanite-bearing Mount Abraham Schist is a pre-peak metamorphic thrust slice. CZa4 is in sharp contact with the underlying rocks of the Fayston and Granville Formations. The contact predates the dominant schistosity and cuts across depositional contacts in the lower plate rocks. The rock types on either side of the fault are not intercalated, but the contact is metamorphically gradational as indicated by the overgrowth of albite porphyroblasts in a zone that is less than one meter wide. The Lincoln Gap Member of the Granville Formation appears to be in depositional contact atop the Mount Abraham (CZa4) thrust slice. No evidence of upper or lower plate truncations exists.

PINNEY HOLLOW FORMATION

The Pinney Hollow Formation contains four major members: silvery green schist (CZph), greenstone (CZphg), metagraywacke (CZphw), and quartzose schist (CZphq). Modal analyses of the Pinney Hollow Formation are presented in table 6.

Silvery Green Schist (CZph)

The majority of the Pinney Hollow Formation is a silvery green muscovite-quartz-albite-chlorite schist that is commonly called the "Pinney Hollow schist." Distinct quartz segregations, and elongate streaks of chlorite as mineral lineations on the foliation are characteristic of the silvery green schist. Muscovite is silvery and coarse grained. The schist ranges from fine-grained to the more common coarse grained schist characterized by albite porphyroblasts no greater than 4 mm in diameter. Rare chlorite clots or bumps (often streaked out) represent completely retrograded garnet. Magnetite octahedra 1 to 3 mm in size may or may not be present. Thin (less than 10 cm) gray quartzites are a minor rock type of the silvery green schist. Thin (less than 20 cm) weakly foliated quartzo-feldspathic layers also occur as a minor rock type in the schist. Neither the quartzites nor the quartzo-feldspathic layers occur on a mappable scale. The quartzites are virtually identical to those in the white albitic schist of the Fayston Formation (CZf). The quartzo-feldspathic layers are very similar to the quartzo-feldspathic granofels of the Fayston Formation (CZfq).

The silvery green schist (CZph) is in depositional contact with the greenstone (CZphg), metawacke (CZphw), and quartzose schist (CZphq). The silvery green schist is also in depositional contact with the Mount Abraham Schist (CZa1), and a transitional sequence of rocks is often associated with the contact. The schist is in pre- and syn-peak metamorphic fault contact with the carbonaceous albitic schist of the Granville Formation (CZg) along the western boundary of Domain 1. The silvery green schist is in syn-peak metamorphic fault contact with the Ottauquechee Formation along its eastern boundary.

Greenstone (CZphg)

The greenstone is a fairly homogenous, light green chlorite-albite-actinolite-epidote schist. The foliation varies from moderate to weakly developed. The greenstone is usually fine grained and occasional light greenish yellow compositional banding may be present. Patches of calcite are present in places as well as finely disseminated calcite in the matrix. Quartz segregations are rare, but do occur.

The greenstone is in gradational contact by intercalation with the silvery green schist. The greenstone often serves as an excellent marker horizon in the schist because it is more resistant.

Table 6

Average Modal Compositions of the Pinney Hollow Formation

Mineral:	CZph	CZphw	CZphg	CZphq	CZph-a	CZph-ag
Quartz	38	42	5	45	30	2
Albite	21	22	22	27	10	19
Muscovite	32	14	--	26	43	--
Chlorite	7	11	31	--	14	20
Biotite	1	3	--	--	--	--
Garnet	Tr	--	--	--	2	--
Chloritoid	--	--	--	--	Tr	--
Opaques *	1	3	2	2	1	Tr
Epidote	Tr	3	12	--	Tr	23
Clinozoisite	Tr	Tr	8	--	Tr	8
Allanite	--	--	--	--	Tr	--
Tourmaline	Tr	Tr	--	--	Tr	--
Sphene	Tr	Tr	1	--	--	2
Carbonate **	Tr	2	3	--	--	18
Apatite	Tr	Tr	Tr	--	Tr	--
Actinolite	--	--	16	--	--	8
Alteration #	Tr	Tr	Tr	--	--	--
	n=8	n=4	n=3	n=3	n=5	n=3

* Opaque minerals include magnetite, ilmenite, and pyrite.
** Carbonate includes calcite and dolomite.
# Alteration includes limonite or rutile.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace. CZph-a and CZph-ag are the transitional schist and greenstone between the Pinney Hollow (CZph) and Mount Abraham (CZa1).

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Metawacke (CZphw)

The metawacke is a homogeneous, coarse grained, weakly foliated, greenish gray quartz-albite-muscovite-chlorite granofels. Granular quartz is visible in hand sample. Albite porphyroblasts are small (up to 1 mm) and ubiquitous. Micaceous partings of muscovite and biotite are silvery and greenish brown respectively. Quartz segregations are characteristically absent. Magnetite may or may not be present. The metawacke is identical to the metawacke in the Mount Abraham schist (CZaw). The metawacke is intercalated with the silvery green schist. The metawacke was found close to a greenstone in one location (V-15).

Quartzose Schist (CZphq)

CZphq is a white, to light gray green, well foliated quartz-albite-muscovite schist. The quartzose schist is usually well foliated, but in places where the foliation is less developed the rock resembles an albitic quartzite. Fine grained quartz and albite give the rock a sugary or sandy appearance. Muscovite defines the schistosity. Quartz segregations are rare. Petrographically, the quartzose schist has a lepidoblastic texture that becomes proto-mylonitic to mylonitic near syn-metamorphic fault zones.

The quartzose schist is intercalated with the silvery green schist over a distance of 10 to 15 m at the Valleymeade Farm (AA-13), although the contact is often transposed by syn-peak metamorphic faults. The zone of intercalation is at an angle to the trend of the dominant schistosity indicating that it predates the foliation.

OTTAUQUECHEE FORMATION

The Ottauquechee Formation contains two major units: The black phyllite (Cobp) and the Thatcher Brook carbonaceous albitic schist Member (Cotb). The black phyllite contains a previously unreported sub-unit called the gray carbonate schist (Cog). Modal analyses of the Ottauquechee Formation are presented in table 7.

Black Phyllite (Cobp)

The main body of the Ottauquechee Formation consists of a quartz-muscovite-graphite phyllite. The phyllite is black, and in places rusty weathered, and foliation surfaces have a sheen due to the abundance of fine-grained muscovite and graphite. Quartz occurs as thin laminations and segregations. The phyllite is characterized by 1-2 cm pyrite molds and a lack of albite.

Uncharacteristic of the Ottauquechee black phyllite in the Fayston - Buels Gore area is a lack of dark gray to black quartzites. Although numerous workers immediately to the south (Haydock, 1988; Armstrong, in progress; Kimball, in progress; and Kraus, 1989) report the presence of the dark quartzites in the black phyllite, none were found in the Fayston - Buels Gore area.

The black phyllite is intercalated with the gray carbonate schist over a distance of 20 cm. No other depositional relationships are observed in the Ottauquechee Formation. Numerous syn-peak metamorphic faults cut through the black phyllite. The phyllite is in syn-peak metamorphic fault contact with the Thatcher Brook Member of the Ottauquechee, and the Pinney Hollow and Stowe Formations. The black phyllite is also in pre-peak metamorphic fault contact with the rocks of the Pinney Hollow Formation. Several windows of the Pinney Hollow occur within the center of the main body of the black phyllite (AA-11, AA-13).

Gray Carbonate Schist (Cog)

The gray carbonate schist was described for the first time in Walsh (1989). The rock is a well foliated, light greenish gray, quartz-chlorite-calcite-muscovite-graphite schist. Disseminated graphite and abundant chlorite give the rock its characteristic gray color. Granular quartz and quartz laminations are common, but segregations are not. Calcite is highly disseminated. Weathering of calcite gives the rock a characteristic rusty rind. Cross muscovite crystals (up to 2 mm) are often visible in hand sample.

Petrographically, the gray carbonate schist has a lepidoblastic texture. The dominant foliation is defined by chlorite, muscovite, and graphite. As well as occurring parallel to the schistosity, muscovite commonly occurs as randomly oriented cross micas. The cross muscovites sharply truncate the dominant schistosity and commonly contain graphite inclusion trails which parallel the schistosity indicating post-tectonic crystallization. Rutilated quartz grains are seen in one sample. Calcite occurs as anhedral granular aggregates and as post-tectonic euhedral rhombic crystals.

The gray carbonate schist is in depositional contact with the black phyllite. Contacts are gradational by intercalation within a distance of 5 to 20 cm. The gray carbonate schist occurs in small bodies less than 10 meters across, and is not an extensive unit in the Ottauquechee Formation.

Table 7

Average Modal Compositions of the Ottauquechee Formation

Mineral:	Cobp	Cog	Cotb	Cotb Quartzite
Quartz	46	25	33	92
Albite	Tr	8	17	2
Muscovite	38	19	39	5
Chlorite	1	24	5	Tr
Graphite	15	2	6	Tr
Calcite	--	21	--	--
Opaques *	Tr	1	Tr	Tr
Epidote	--	--	Tr	Tr
Tourmaline	--	Tr	Tr	Tr
Apatite	--	--	Tr	--
Limonite	Tr	Tr	Tr	Tr
Rutile	--	Tr	--	--
	n=2	n=6	n=3	n=2

* Opaque minerals include pyrite and ilmenite,
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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Thatcher Brook Member (Cotb)

The name Thatcher Brook Member of the Ottauquechee is tentatively assigned to a heterogeneous group of rocks that is quite distinct from the typical black phyllite and quartzite of the Ottauquechee. This group of rocks includes a black, rusty, carbonaceous albitic schist, greenstones, and ultramafics. Cady et al. (1962) included these rocks in the Ottauquechee and did not differentiate them from the black phyllite. The name Thatcher Brook Member is proposed by Armstrong et al. (1988).

The Thatcher Brook Member consists of a rusty weathered muscovite-quartz-albite-chlorite-graphite schist with dark gray to black, commonly banded, quartzite. The graphitic schist contains distinct discontinuous layers, or segregations, of graphite as opposed to the patches of graphite in the carbonaceous albitic schist of the Granville Formation. Black albite porphyroblast, which are small (<3mm) to absent, are most abundant in schist segregations that contain less graphite.

The graphitic schist is more heterogeneous than the schists of the Granville. Heterogeneities include the distribution of graphite, discontinuous chlorite-rich, and albite-rich graphite-poor lenses, and the close association of the graphitic schist with greenstones, serpentinites, and talc schists. The carbonaceous albitic schist of the Granville is compositionally similar to the Thatcher Brook Member. The major difference is the distribution of graphite. Graphite in the carbonaceous albitic schist occurs as discontinuous patches that are evenly dispersed throughout the rock, whereas graphite in the Thatcher Brook occurs as discontinuous lensoidal layers parallel to the foliation. Although albite is present in the graphitic segregations it is not as abundant as in the carbonaceous albitic schist. The Thatcher Brook does not resemble the Lincoln Gap Member as much as it does the carbonaceous albitic schist due to the greater abundance of graphite and lesser abundance of albite in the latter. The Lincoln Gap Member more closely resembles the Ottauquechee black phyllite.

Depositional contacts, apart from the dark quartzites in the graphitic schist, do not occur in the Thatcher Brook Member. Numerous syn-metamorphic faults cut through the Thatcher Brook member. Generally, the Thatcher Brook is in syn-metamorphic fault contact with the silvery green schist of the Pinney Hollow to the west and the black phyllite of the Ottauquechee to the east. All greenstones, serpentinites, and talc schists in the Thatcher Brook Member occur as tectonic slivers along syn-metamorphic faults. The contacts are always sharp, always parallel to the dominant schistosity, and are characterized by mylonitic fabrics. No transitional contacts between the exotic rock types and the graphitic schist are ever observed.

STOWE FORMATION

The Stowe Formation contains two members: The silvery green schist, and the Stowe greenstone (CZsg). Modal analyses of the Stowe Formation are given in table 8.

Silvery Green Schist (CZs)

The Stowe schist is a fine grained, silvery to dark green quartz-muscovite-albite-chlorite schist. Quartz is present as thin granular laminations and abundant discontinuous segregations. Albite occurs as small porphyroblasts up to 2 mm in diameter. Chlorite is fine grained, and thick laminations are often dark green to bluish black. Trace amounts of graphite occur as small, thin, discontinuous patches in the schist found to the west near the contact with the black phyllite of the Ottauquechee Formation. In places dark gray albite occurs in the schist at locations that contain patchy graphite.

The silvery green schist is in depositional contact with the Stowe greenstone. The contact is gradational over a distance of a few meters to less than 10 cm, and is often characterized by an intercalation of the two rock types.

The contact between the silvery green schist and the Ottauquechee Formation is a syn-metamorphic fault contact. The contact parallels the dominant schistosity, and Stowe greenstones are truncated against it. A mesoscopic mylonitic or tectonite fabric occurs along the contact where it exposed in Shepard Brook (BB-9). The fabric is characterized by thin (<5 cm) discreet zones of alternating and anastomosing carbonaceous and non-carbonaceous rock types.

Greenstone (CZsg)

The greenstone is a homogenous, fine grained, light green actinolite-albite-epidote-calcite-chlorite schist. Finely disseminated calcite grains and discontinuous calcite veins parallel to the schistosity are common. Thin discontinuous quartz segregations are present in places. Thin (<5 cm) compositional layering of epidote, chlorite, and calcite-rich layers occurs in the greenstone, but is not common. The greenstone is resistant to erosion and large outcrops are common. The greenstone is in depositional contact with the silvery green schist.

ULTRAMAFICS (S)

The ultramafic rocks include serpentinite and talc-magnesite schist. The serpentinite is a dense, massive, weakly foliated, dark green serpentine-magnetite rock. Magnetite is visible as small black specks in the rock. The serpentinites are cut by numerous fractures and veins that are filled with magnesite and limonite. Weathered outcrops of serpentinite have a light gray to tan alteration rind that penetrates up to 2 cm into the rock. Talc-magnesite schists are soft, well-foliated, light gray to tan schists. Magnesite occurs as irregular patches and spots in the schist, and is often recognized by rusty alteration to limonite. Modal analyses of the

Table 8

Average Modal Compositions of the Stowe Formation

Mineral:	CZs	CZsg
Quartz	45	2
Albite	11	24
Muscovite	28	--
Chlorite	6	6
Opaques *	Tr	Tr
Epidote	Tr	17
Clinozoisite	--	11
Sphene	2	1
Calcite	8	14
Actinolite	--	25
Graphite	Tr	--
	n=1	n=1

* Opaque minerals include magnetite and ilmenite.
The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

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ultramafic rocks are presented in table 9.

All ultramafic bodies are in fault contact with rocks of the Ottauquechee Formation, especially the Thatcher Brook Member. The ultramafics occur as elongate lens-shaped bodies parallel to the dominant schistosity. All three of the ultramafic bodies occur along strike of each other, and the zone is characterized by mylonitic fabrics.

CRETACEOUS DIKES (Kd)

The dikes are massive, well-jointed, vertically oriented, black, aphanitic, mafic intrusives that cut the metamorphic foliation. Small (up to 3mm) light gray to blue amygdules of dolomite occur in some dikes. The dikes range in thickness from 0.5 to 1.5 m and display thin (up to 5 cm) chill margins and in places rusty and spheroidal weathering. Contact aureoles in the surrounding country rock are not present.

A single albite epidosite dike (Kds) occurs in the carbonaceous albitic schist of the Granville Formation (703, K-14), and is an exception to the common dike rock. The albite epidosite is a tannish white to light greenish gray rock that possesses a weak, irregular foliation that does not appear to be related to the surrounding country rock. Albite, quartz, and dolomite are the only minerals identifiable in hand sample. The albite epidosite is considered a Cretaceous dike because it cuts the schistosity of the country rock, has the same orientation as the other dikes, and is 0.5 m wide.

Thin sections from 2 of the 20 dikes in the study area indicate that the dikes are diabases. Both samples have idiomorphic diabasic textures characterized by interlocking laths of plagioclase with clinopyroxene in the interstices. Optical properties of the clinopyroxene resemble those of augite and salite. Opaques of magnetite and ilmenite are abundant. The

Table 9

Average Modal Compositions of Ultramafics (S)

Mineral:	Serpentinite	Talc-Magnesite Schist
Serpentine	90	--
Talc	6	94
Magnesite	3	5
Magnetite	1	--
Limonite	Tr	Tr
	n=2	n=1

The "n" indicates the number of thin sections analyzed.
Average modal analyses are volume percent estimates based upon a minimum of 100 point counts.
Tr=trace.

***

opaques occur both as euhedral polygonal crystals, and as a network, boxwork or Windmanstätten texture. Dark brown amphibole is kaersutite or barkevikite. Biotite, olivine, and alkali feldspar occur in small amounts. Dolomite is restricted to amygdules. A pale green to yellow alteration product identified as uralite occurs around clinopyroxene and as rims around amygdules.

The albite epidosite is an unusual rock. Epidote occurs as fine grained granular aggregates in cross-cutting veins and fractures. Albite appears to be relict porphyroblasts that have undergone subsequent fracturing. The porphyroblasts have inclusions of graphite, quartz, and epidote. Porphyroblasts have either no twins or a single albite twin. Fractures through the albite are filled with epidote and dolomite, and to a lesser extent muscovite. Quartz, dolomite, and chlorite occur in the matrix with epidote. Modal analyses of the dikes are presented in table 10.

The albite epidosite is interpreted to be a hydrothermally altered, or sausseritized version of the carbonaceous albitic schist. This alteration may have occurred above a rising dike. If this interpretation is true, the nature of the hydrothermal fluids must include the following conditions: 1) The fluids must be calcium-rich in order to produce the abundant epidote and dolomite; 2) The oxidation state must be high enough to cause graphite to combine with oxygen to form CO_2, which in turn would help produce dolomite. Evidence for this is found in the fact that graphite is present in the matrix of the country rock, but only occurs as preserved inclusions in albite in the altered rock; and 3) The fluids must have the capacity to remove K and Si because muscovite and quartz are less abundant in the dike than in the country rock.

CORRELATIONS

The correlation of lithologies in the Fayston - Buels Gore area is presented on a local and regional scale.

Local Correlations

Within the area there is a noticable similarity between the rock units, and a discernible variation of the units from west to east.

Table 10

Modal Compositions of Cretaceous Dikes

Mineral:	Diabase (Kd) 185GW	Diabase (Kd) 414GW	Mineral:	Albite epidosite (Kds) 703GW
Plagioclase	53	56	Epidote	38
Clinopyroxene*	3	15	Albite	22
Amphibole**	5	--	Quartz	14
Olivine	1	3	Chlorite	12
Alkali Feldspar	1	4	Dolomite	10
Biotite	5	--	Muscovite	2
Dolomite	2	13	Opaques#	2
Opaques#	12	8	Graphite	Tr
Uralite	18	Tr		

* Clinopyroxene is augite and/or salite.
** Amphibole is kaersutite and/or barkevikite.
\# Opaques include magnetite and ilmenite.

Modal analyses are volume percent estimates based upon a minimum of 100 point counts. Tr=trace.

***

The abundance of muscovite-quartz-albite-chlorite schists across the research area is striking. Such rocks are the predominant lithology in the Underhill, Fayston, Pinney Hollow, and Stowe. In the Underhill this schist is in depositional contact with relatively abundant coarse-grained metawacke, greenstone, and amphibolite. The schist is commonly "pinstriped" in appearance due to the abundance of quartz and albite. Relict detrital textures are common in the feldspars, and detrital blue quartz occurs in both the schist and metawacke. The Fayston white albitic schist is coarser grained than the schist of the Underhill due to the abundance and size of albite porphyroblasts, but the two schists have essentially the same assemblage. Coarse quartzo-feldspathic rocks occur at a mappable scale as the granofels (CZfq) and quartz-biotite gneiss (CZfb). Of the two, CZfb has feldpsars with relict detrital textures. The granofels is a quartz and feldpar-rich, and mica-poor rock, and although it lacks relict detrital textures its present composition suggests a quartz-feldspar clastic protolith. Greenstones and thin white quartzites are also present in the white albitic schist. The Pinney Hollow silvery green schist is very similar to the white albitic schist. The silvery green schist has both coarse-grained and fine-grained muscovite-quartz-albite-chlorite schist with the former being slightly more abundant. Thin quartzites and quartzo-feldspathic granofels occur in the Pinney Hollow, but to a lesser extent than in the Fayston. Greenstones and metawackes also occur in the silvery green schist. The silvery green schist of the Stowe is virtuallly identical to the fine-grained silvery green schist of the Pinney Hollow. Greenstones are present, but metawackes and quartzites are not observed. Sedimentologically, and stratigraphically the muscovite-quartz-albite-chlorite schists appear to represent a continuous sequence of clastic rocks that becomes finer to the east. The sequence is interpreted as a west to east facies change (Walsh and Stanley, 1988). Aluminous schists of the Mount Abraham, and to a lesser extent the Monastery, generally do not have the coarse clastics and greenstones found in the aforementioned rocks. The Mount Abraham Schists CZa1 and CZa3 appear to be in depositional contact with the muscovite-quartz-albite-chlorite schists. CZa1 contains greenstones and metawackes similar to those in the Pinney Hollow. The aluminous rocks are believed to be a discontinuous sequence of finer grained rocks deposited above the coarser quartz-feldspar clastics.

The graphitic rocks in the area are characterized by a lack of depositional greenstones. The graphitic albitic rocks of the Granville in the western part of the research area appear to be in depositional contact with the quartz-feldspar clastics, and to a lesser extent the aluminous schists. The presence of graphite suggests that the environmental conditions had changed to a more anoxic state. Coarse clastics, except for minor quartzites, are absent from the rocks. Carbonates are present in the White River Member of the Battell, but absent from the rest of the graphitic rocks. Ultramafics occur in the Ottauquechee Formation suggesting an earlier association with oceanic crust. This evidence suggests that the graphitic rocks were deposited as a relatively fine-grained, coherent sequence over the quartz-feldspar clastics and, in places, the aluminous schists. The sequence is believed to span the depositional basin from a minimal carbonate environment to one over ocean crust.

Table 11 summarizes the local correlations between the units of this study with those of previous workers in the Lincoln Mountain quadrangle, and with those of the most recent workers to the west, south, and north.

Regional Correlations

Regional correlations between the rocks of the pre-Silurian sequence of central Vermont with those of the Taconics were made as early as the 1930's (Keith, 1932; and Prindle and Knopf, 1932). The best regional correlations concern the Ottauquechee, Pinney Hollow, and Mount Abraham. The Ottauquechee is correlated with the Hatch Hill and West Castleton in the Taconics (Potter, 1972, 1979; Rowley et al., 1979; and Stanley and Ratcliffe, 1985). The Taconic correlatives, however, contain significant interbedded carbonates which are not present in the Ottauquechee. Kraus (1989) compares samples of the Taconic correlatives with the Ottauquechee black phyllite and notes the abundance of interstitial carbonate in the Taconic rocks. Kraus (1989) concludes that the Ottauquechee was deposited below the calcium carbonate compensation depth in an an environment more distal than the that of the Taconic rocks. The Ottauquechee therefore, is considered a distal correlative of the Taconic rocks.

The silvery green schist of the Pinney Hollow is correlated with the green slates and phyllites of the Nassau from the Group 1 and 2 Taconic slices (Potter, 1979; Stanley and Ratcliffe, 1985). The greater abundance of interbedded metawackes in the Taconics implies that the Pinney Hollow is a more distal correlative.

The four different units in the Mount Abraham are correlated separately and not as a whole, coherent unit. CZa1 is similar to chloritoid schists in the Pinney Hollow in southern Vermont (Karabinos, 1987). Similar chloritoid units also occur in the Stowe from central to northern Vermont (Albee, 1957, 1972; Cua, 1989; Doll et al., 1961). In the Group 2 Chatham and Rensselaer slices of the Taconics, green and purple chloritoid-bearing slates and phyllites are depositionally intercalated with minor metawackes, mafic schists, and green slates of the Nassau Formation (Potter, 1979; Ratcliffe, 1987). CZa3 is similar to chloritoid schists in the Hoosac of southern Vermont and Massachusetts (Ratcliffe, 1979; Zen, 1983; Karabinos, 1987). In northern Vermont and southern Quêbec, correlatives include the chloritoid units in the Underhill (Thompson and Thompson, 1989) and West Sutton (Colpron et al., 1987; Colpron, 1990), respectively. In the Taconics CZa3 is correlated with chloritoid units in the Group 2 and 3 Everett and Greylock slices (Potter, 1979; Ratcliffe, 1979), and the Metawee facies of the Nassau (Potter, 1972, 1979).

The allochthonous Mount Abraham CZa2 and CZa4 appears to be overlain by a depositional sequence of black shales (Cgl). Perhaps this sequence is a distal correlative of the upper Metawee facies and lower Hatch Hill sequence of Potter (1972) and Rowley et al. (1979). Carbonates are recognized in the Taconic sequence, but not in the sequence from central Vermont. Albee (1957) reports a sequence of aluminous chloritoid bearing schist overlain by

Table 11
Local Correlation Chart

THIS STUDY	WEST (2)	ADJACENT AREAS SOUTH (3 & 4)	NORTH (5)	PREVIOUS WORK (1)
STOWE				
CZs	East of Area	East of Area	CZs	OCs
CZsg			CZsg	OCsg
OTTAUQUECHEE				
Cobp		Co	Co	Co
Cog		Not Reported	Not Reported	Co
Cotb		Hazens Notch CZhnca	Granville CZg	Co
PINNEY HOLLOW				
CZph		CZph	CZph	Cph
CZphg		CZphg	CZphg	Cpg
CZphw		CZphw	Not Reported	Cph
CZphq		Not Reported	Not Reported	Co
MOUNT ABRAHAM				
CZa1		Not Reported	Not Reported	Cph
CZag		Not Reported	Not Reported	Cph
CZaw		Not Reported	Not Reported	Cph
CZa2		CZa	Not Reported	Ca
CZa3		CZa	Pg-Ctd schist in Eastern Underhill	CZu
CZa4		CZa	Not Reported	Ca
GRANVILLE				
Cg		Hazens Notch CZhnca	Granville CZg	Ch & Cu
Cgl		Hazens Notch CZhnc Lincoln Gap Member	Lincoln Gap Member	Ch & Cu
FAYSTON				
CZf		Hazens Notch CZhn & Underhill CZu	Eastern Underhill	Cu
CZfg		Hazens Notch CZhng	Not Reported	Ch
CZfb		Not Reported	Not Reported	Cu
CZfq		Not Reported	Not Reported	Cu
CZft		Not Reported	Not Reported	Cu
MONASTERY				
CZm		CZu, Prospect Rock Belt	Not Reported	Cu
BATTELL				
Cbw		CZuc ?	Not Reported	Cu
UNDERHILL				
CZu	CZu	CZufg	Western Underhill	Cu
CZuql	CZuql	CZufg		Cu
CZug	CZua	CZums		Cu

References: (1) Cady et al., 1962; (2) DiPietro, 1983; (3) Haydock, 1988; (4) O'Loughlin and Stanley, 1986; and (5) Thompson and Thompson, 1989. The correlation with the previous work indicates how the units of this study were mapped by Cady et al. (1962).

graphitic phyllite in the Foot Brook syncline near Hyde Park, Vermont. Albee originally called the lower unit the Stowe and the upper unit the Ottauquechee. Doll et al. (1961) later lumped the lower and upper units into the Underhill as the Foot Brook Member (Cufb) and graphitic phyllite member (Cug), respectively, but gave Cufb the same color and symbol as the Mount Abraham Schist Member of the Underhill.

Table 12 summarizes the regional correlations between the units of this study with those of workers in northern Vermont and southern Québec, southern Vermont and Massachusetts, and the Taconics.

DEPOSITIONAL ENVIRONMENT

Due to the degree of metamorphism and deformation within the study area, analysis of sedimentary features is not possible. Aside from relict depositional contacts, sedimentary structures such as bedding are not preserved in the rocks. The determination of protoliths, therefore, is a task that involves considerable speculation and interpretation. The most important means for deciphering the early history of these rocks involves the recognition of depositional contacts identified by intercalations of rock types, and by a detailed recognition and separation of different metamorphic rock types. The latter can be accomplished by recognizing subtle differences in the present metamorphic assemblages.

The discussion concentrates on the three general rock types in the area: 1) Noncarbonaceous quartz-feldpar clastic rocks; 2) Metavolcanic mafic schists interlayered with the former; and 3) Carbonaceous schists and phyllites.

Noncarbonaceous Rocks

The noncarbonaceous rocks fall into two catagories: 1) Medium to coarse grained muscovite-quartz-albite-chlorite schists with interlayered metawackes, gneisses, granofels, metaquartzites, and metavolcanics; and 2) The aluminous Mount Abraham schists.

The albitic schists are a very extensive clastic rock type in central Vermont. They vary in grain size partially as a function of metamorphic grade, but also as a function of their original composition. The coarseness of the rock is directly related to the amount and size of quartz and albite. The presence of blue quartz in some metawackes, and relict detrital fragments of feldspar suggests that the source for these rocks is the Middle Proterozoic basement of the North American craton. Presumably the original sedimentary rocks were subarkosic to arkosic sandstones, siltstones, and mudstones with interbedded graywackes and quartz arenites. The monomineralic nature of quartz arenites aids in the preservation of their original character as present day metaquartzites. The coarse grained quartz and feldspar-rich nature of the graywackes aids in their preservation as well, although original variations in these coarse rocks along with subsequent deformation and metamorphism results in the formation of metawackes, gneisses, and granofels. The fine grained sandstones, siltstones, and mudstones, however, have undergone the greatest change to psammitic albitic schists.

The abundance of albite in the schists does not directly reflect the original composition of detrital feldspars because both K-feldspar and calcium plagioclase would not be in equilibrium with the greenschist facies conditions recorded in the rocks. Potassic feldspars contribute their potassium to the formation of micas. Calcic plagioclase alters by albitization as early as diagenesis (Helmold and van de Kamp, 1984).

Moody and Jenkins (1981) determined experimentally that labradorite in the presence of Na^+ and SiO_2 breaks down to form both albite and sodic plagioclase. In the reaction excess Ca^{2+} and Al^{3+} from labradorite go to produce clinozoisite. Helmold and van de Kamp (1984, p.259) document the process of diagenetic albitization of oligoclase and andesine in Paleogene calcite-free arkosic sandstones from the Santa Ynez Mountains in California. The reaction

Table 12
Regional Correlation Chart

THIS STUDY	SOUTHERN VERMONT & MASSACHUSETTS	NORTHERN VERMONT & SOUTHERN QUEBEC	TACONICS
STOWE	Rowe (Y) Stowe (I)	Stowe (I,J,W) Upper Mansonville (C) Caldwell (J)	No Correlation
OTTAUQUECHEE	Ottauquechee (I) Rowe (Y)	Sweetsburg (J,K,G) Ottauquechee (I) Lower Mansonville (C,E)	Groups 1 & 2 West Castleton & Hatch Hill (Q,R,U,V)
PINNEY HOLLOW	Pinney Hollow (L,S)	Hyde Park Member of Hazens Notch ?(A,I) Terminates in Camels Hump 15' Quad. (D,I,X)	Groups 1 & 2 Nassau (R,V)
MOUNT ABRAHAM			
CZa1	Ctd units in Pinney Hollow (L)	Ctd units in Stowe (A,H,I)	Ctd units in Group 2 Rensselaer & Chatham Slices (R,T)
CZa3	Ctd units in Hoosac (L,S,Y)	Ctd units in Underhill (X) & West Sutton (F,G)	Ctd units in Groups 2 & 3 Everett & Greylock Slices (R,S) Metawee facies of Nassau (Q,R)
CZa2 & CZa4	Mount Abraham (B,M,O)	Foot Brook Syncline (A,I)	Groups 1 & 2 upper Metawee facies (Q,R)
GRANVILLE	Granville (B,P) Hazens Notch (I)	Hazens Notch (I,K,W) Sutton Schists (F,G)	Groups 1 & 2 West Castelton & Hatch Hill (Q,R,U)
FAYSTON	Hoosac (L,S) Allochthonous Hoosac (S,V,Y)	Sutton Schists (C,F,G) Hyde Park Member of Hazens Notch (A,I) Underhill (I,K)	Greylock and Dorsett Slices (S,V)
MONASTERY	Monastery (B,P)	No Correlation	No Correlation
BATTELL	Battell (B,I,P)	Sweetsburg with black limestone (N) Mansville suite (F,G)	No Correlation
UNDERHILL CZug	No Correlation	Lower Pinnacle (F,G,J,K) Tibbit Hill (K)	No Correlation

Table 12 Continued
References for Table 12:

(A) Albee, 1957, 1972
(B) Armstrong, 1989a, 1989b
(C) Cady, 1960
(D) Christman and Secor, 1961
(E) Clark, 1934
(F) Colpron, 1989
(G) Colpron et. al, 1987
(H) Cua, 1989
(I) Doll et. al, 1961
(J) Doolan et. al, 1982
(K) Doolan et. al, 1987
(L) Karabinos, 1987
(M) Lapp, 1986
(N) Mock, 1989a, 1989b
(O) O'Loughlin, 1986
(P) Osberg, 1952
(Q) Potter, 1972
(R) Potter, 1979
(S) Ratcliffe, 1979
(T) Ratcliffe, 1987
(U) Rowley et. al, 1979
(V) Stanley and Ratcliffe, 1985
(W) Stanley et al., 1984
(X) Thompson and Thompson, 1989
(Y) Zen, 1983

involves plagioclase, quartz, water, and sodium ions to form albite, laumontite, and calcium ions as follows:

$$NaAlSi_3O_8{*}CaAl_2Si_3O_8 + 3\ SiO_2 + 2\ H_2O + Na^+ =$$
$$2\ NaAlSi_3O_8 + 0.5\ CaAl_2Si_4O_{12}{*}4H_2O + 0.5\ Ca^{2+}$$

For the reaction to occur significant amounts of sodium and silica are needed. The authors also state that connate water and sea water combined lack sufficient sodium, unless abnormally large volumes of such water move through the rock. Hower et al. (1976, p. 773) indicate that the needed sodium and silica can come from the conversion of smectite to illite according to the following simplified reaction:

$$\text{smectite} + Al^{3+} + K^+ =$$
$$\text{illite} + Si^{4+} + Na^+ + Ca^{2+} + Mg^{2+} + Fe^{2+,3+} + H_2O$$

Correlative rocks of the Nassau Formation, especially the feldspathic and lithic Rensselaer and Bomoseen graywackes, contain detrital oligoclase, plagioclase, and microcline (Potter, 1972). If the graywackes contain detrital feldspar it is likely that the finer grained sandstones, siltstones, and mudstones also contained detrital feldspar at one time. Potter (1972) and Ratcliffe (1987) interpret the graywacke facies in the Taconics as turbidite deposits and note that they thin both to the northeast and southwest, especially in the Chatham slice.

The albitic schists of the Fayston, Pinney Hollow, and Stowe Formations are likely to represent more distal equivalents of the Group 1 and 2 rocks. The quartzo-feldspathic granofels of the Fayston (CZfq) is a significant unit in the western part of the study area. Towards the east thin layers of similar lithology occur within the Pinney Hollow schist. The decrease in abundance of such quartz and feldspar-rich rocks from west to east is one of the contributing factors to the interpretation that the albitic schists represent a fining eastward facies change (Walsh and Stanley, 1988).

The aluminous schists of the Mount Abraham can be generally divided into two catagories. The first includes the relatively smaller bodies of interlayered schists in the Fayston and Pinney Hollow (CZa3 and CZa1, respectively). The second includes the dismembered thrust sheet of CZa4 and CZa2. Schists of the first group represent metamorphosed aluminous pelagic muds interlayered with protolithic equivalents of the albitic schists, metawackes, and metavolcanics. Schists of the second group represent similar pelagic muds that lack the interlayered rock types of the first group. Perhaps CZa2 and CZa4 were deposited higher up and farther east in the sequence.

The next question regarding the aluminous schists involves the origin of aluminum enrichment. The Metawee facies in the Taconics is an Al-Fe rich sequence that contains hematite at lower grades and chloritoid and paragonite at higher grades (Zen, 1960). Remembering that laumontite is produced by diagenetic albitization of Ca-plagioclase, the following reactions show the diagenetic breakdown of laumontite (Helmold and van de Kamp, 1984, p. 269):

$$\text{laumontite} + CO_2 = \text{calcite} + \text{kaolinite} + 2\ \text{quartz} + 2\ H_2O$$
and
$$\text{laumontite} + 2\ H^+ = \text{kaolinite} + 2\ \text{quartz} + Ca^{2+} + 3\ H_2O$$

The utilization of the kaolinite in the Al-Fe rich rocks of the Taconics was shown by Zen (1960, p. 152) in the following reaction:

$$\text{albite} + \text{kaolinite} = \text{paragonite} + 2\ \text{quartz} + H_2O$$

a reaction that takes place from chlorite to biotite grade metamorphism. In the aluminous rocks kaolinite may also go to pyrophyllite, and at higher grades alumino-silicates (Miyashiro, 1973).

The kaolinite reactions as well as the presence of hematite should be kept in mind when considering the formation of chloritoid as follows (Miyashiro, 1973, p. 205):

$$5 \text{ hematite} + \text{chlorite} =$$
$$5 \text{ magnetite} + 2 \text{ chloritoid} + 2 \text{ quartz} + 2\ H_2O$$
$$\text{OR}$$
$$5 \text{ kaolinite} + \text{chlorite} = 7 \text{ chloritoid} + 7 \text{ quartz} + 2\ H_2O$$

Both of the reactions are conceivable in light of the present day assemblages, and indicate a likely source for the abundant quartz segregations in the Mount Abraham Schist.

The initial origin of the Al-Fe rich sediments is less clear. Ratcliffe (1987) introduces the idea that the chloritic, iron-rich facies of the Nassau may have been derived from mafic volcanics that once capped the North American basement, but are now eroded away. A likely protolith for the aluminous rocks must meet the requirement of chemical maturity. The relative abundance of aluminum may be the result of depletion of the more mobile elements during chemical weathering. Slack and Bitar (1984) suggest that the Al-Fe rich nature of the Mount Abraham is due to extensive pre-metamorphic hydrothermal alteration. This interpretation, however, is not consistent because such extensive hydrothermal alteration should have associated sulfide and related ore deposits, yet such deposits are rare.

The tourmaline-bearing schist of the Fayston Formation (CZft) represents one of the more anomalous rocks in the study area. The rock contains up to 9% metamorphic tourmaline. Trace amounts of tourmaline occur in most rocks in the study area, but such high concentrations are confined to CZft. The abundance of tourmaline represents boron enrichment either during deposition or metamorphism. Significant amounts of boron can occur in sea water, but amounts in fresh water are much less (Selley, 1978). Potter et al. (1963) state that boron is a good marine vs. fresh water environmental discriminant in argillaceous sediments. Degens (1965) states that boron attaches readily to clay minerals in marine sediments, and that it is difficult to detach the boron from the clays even by leaching with hot mineral acids. He also points out that even greater boron enrichment can occur in marine sediments if the erosion of terrestrial volcanics supplies the added boric acid. Another possibility for the abundance of metamorphic tourmaline is significant detrital tourmaline in the original sediments. This evidence suggests that the environment was marine, and perhaps volcanics were being eroded in the source area.

Mafic Schists

The mafic schists are altered volcanic or hypabyssal intrusive rocks and their origin is more fully discussed in the section on geochemistry. Depositional relationships, however, are presented here.

No original structures are present in the mafic schists. Potter (1972) and Ratcliffe (1987) describe the metavolcanic rocks in the Nassau Formation as sills, flows, pillow basalts, and basaltic tuffs. Since original features are absent in the central Vermont sequence, contact relationships provide the only insight into the volcanic or intrusive origin of the metabasic rocks. Intercalated contacts between metasedimentary rocks and mafic schists may indicate tuffaceous deposits or the margins of flows. Sharper contacts may represent flows or sills. The very coarse grained nature of the Underhill amphibolites suggests an intrusive origin.

Carbonaceous Schists

The carbonaceous schists and phyllites are believed to have been deposited upon the ancient slope-rise or abyssal plain (Doolan et al., 1982; Stanley and Ratcliffe, 1985). Graphitic schists which contain exotic serpentinite and greenstone (Ottauquechee black phyllite and

Thatcher Brook Member) are interpreted to be tectonic melanges (Armstrong et. al, 1988). The striking similarity between the carbonaceous albitic schists with and without exotic rock types (Thatcher Brook vs. Granville and Battell) suggests that the rocks originated in a similar fashion, but that later tectonic activity lead to the incorporation of ultramafic rocks and greenstones into the Thatcher Brook. The Ottauquechee black phyllite also contains ultramafic rocks. The presence of ultramafics in some of the graphitic schists leads to the conclusion that such rocks were deposited on or near oceanic crust. Graphitic schists which lack ultramafics were probably deposited nearer to continental crust, and atop rift clastics (i.e. Granville over Fayston and Lincoln Gap Member over Mount Abraham).

The original deposition of graphitic schists and phyllites as black shales is most likely a function of paleogeography and paleoclimatology. Middle Cambrian to Middle Ordovician black shales have a worldwide distribution. Berry and Wilde (1978) report that the distribution is due to extensive areas of anoxic marine water produced by poor ventilation of the oceans. The authors note that during this time continental glaciation was at a minimum. Reduced glaciation decreases the flow of deep oxygenated polar waters towards the equatorial region. At non-glacial times aeration of the oceans only takes place at shallow depths via atmospheric mixing by wind and waves. Furthermore, with a decrease in supply of cold polar water oxygen solubility decreases as a whole as the ocean's mean temperature rises. In sub-tropical and tropical regions of the ocean the effects of reduced ventilation are more pronounced due to the warmer water temperatures, and the consumption of oxygen by progressive mixing of deep ventilated currents as they proceed toward the equator. In accordance with the ideas on ventilation of the oceans is the paleomagnetic evidence that central Vermont was less than 30° from the equator from Late Cambrian to Middle Ordovician times. The ancient margin of North America was primed for widespread anoxic conditions and the deposition of black shales.

Of all the graphitic units in the study area, the White River Member of the Battell Formation appears to be the most proximal unit. The thin discontinuous black dolomitic marbles are a unique characteristic of the White River Member. Mock (1989a,b) and Colpron (1990) report very similar black limestones in the Sweetsburg and Mansville suite, respectively. Berry and Wilde (1978) report that thinly bedded black limestones occur in lower Paleozoic black shale facies deposited on the outer parts of continental shelves worldwide. The Hatch Hill and West Castleton represent proximal turbidites that contain thinly bedded and laminated dolostone (Keith and Friedman, 1977). The lack of carbonates in either the Granville or Ottauquechee implies that they were deposited basinward of the Taconic correlatives.

In light of the modern theories on passive margin and basin evolution from rift to drift stage (McKenzie, 1978; Royden et al., 1980; and Dewey, 1982) the deposition of the rocks in the central Vermont sequence can be viewed in a more current model. The non-carbonaceous rocks represent rift-stage clastic rocks which pre-date the development of oceanic crust. Mafic schists in the non-carbonaceous rocks are rift-related igneous rocks erupted through the thinned continental lithosphere. As rifting slowed and the margin changed from rapid initial subsidence due to lithospheric stretching to slower thermal subsidence due to the lithospheric cooling, carbonaceous rocks were deposited with a noticable lack of coarse clastics and volcanics. In light of the decreased ventilation of the oceans hypothesis, it seems rather fortuitous that the change from rift to drift-stage development of the passive margin coincides with the paleoclimatological event.

AGE OF UNITS

Based on the discussion of correlations and depositional environment, general age assignments are proposed.

Fossils are not present in any of the rocks of the polymetamorphosed eastern cover sequence. Age determination, therefore, must be conducted by correlation with less deformed, fossil-bearing units in accordance with stratigraphic and sedimentological evidence.

The North American basement upon which the cover sequence was deposited is represented by the 1.1 Ga Middle Proterozoic granitic gneisses of the Green Mountain and Lincoln massifs. A profound unconformity separates the base of the cover sequence (Pinnacle and Hoosac Formations in central Vermont) and the Middle Proterozoic basement. Late Proterozoic mafic dikes cut across the Grenvillian fabric, but in turn are cut by the basal unconformity (DelloRusso, 1986). The dikes are interpreted as rift-related feeder dikes to the mafic volcanics in the cover sequence (DelloRusso, 1986; DelloRusso and Stanley, 1986; and Coish, 1987). This suggests that rift-clastic rocks that contain rift-related volcanics can be as old as Late Proterozoic, but no older than the Middle Proterozoic rocks upon which they lie.

The Lower Cambrian trilobite *Olenellus thompsoni* was named by Hall (1860) who discovered this trilobite in Georgia, Vermont in rocks that would later be called the Dunham. Walcott (1886) found *Olenellus* and *Hyolithes* in the Cheshire near Bennington, Vermont. The term "*Olenellus* Zone" now applies to all Lower Cambrian strata (Resser and Howell, 1938). Immediately to the west and south of the study area DiPietro (1983) and Tauvers (1982) show that the Pinnacle is conformably overlain by the Fairfield Pond, Cheshire, and Dunham Formations. The Pinnacle Formation, therefore, is clearly pre-*Olenellus* and post Middle Proterozoic, and is assigned the age of Late Proterozoic to early Cambrian (CZ). Following suit, all the metavolcanic-bearing rift-related rocks of the eastern cover sequence are assigned the same age. Correlation of the graphitic schists with fossil-bearing rocks is the only way to make a reasonable estimate of their age. The correlative black shales and slates in the Taconic sequence (West Castleton and Hatch Hill) contain various fossils of late Early Cambrian to Late Cambrian age (Zen, 1967; Potter, 1972; Rowley et al., 1979). Berry and Bird (1963) report the occurrence of Late Cambrian graptolites in the Hatch Hill and Early Ordovician graptolites in the overling Poultney. This places a Late Cambrian age constraint on the youngest graphitic rocks in the area if one assumes that deposition was time correlative.

Berry and Wilde (1978) report the worldwide deposition of lower Paleozoic black shales occurred during three stages: Middle Cambrian to Middle Ordovician, Middle Silurian, and Late Devonian. Berry (1974) and Churkin (1974) state that Cambrian to Ordovician black shale facies were deposited from continental slopes seaward. On the basis of these arguments, all of the graphitic schists and phyllites within the study area are given a Cambrian age.

The diabasic dikes in the study area correlate with similar dikes in the region. McHone (1987) summarizes radiometric age dates determined from various dikes in the Champlain Valley, Green Mountains, and Adirondacks, and reports early Cretaceous ages ranging from 146-105 Ma.

GEOCHEMISTRY

Metaigneous Rocks

Whole rock geochemical analysis of 13 greenstones was conducted on the Thermo Jarrel Ash 61 inductively coupled argon plasma spectrometer (ICAP) at Middlebury College, Middlebury, Vermont. Sample locations are indicated on the geologic map (Plate 1). The major and trace element chemistry is presented in table 13. The data are plotted on four geochemical diagrams in figure 1. Figure 1 was originally published by Walsh and Kimball (1989, figure 2) and includes 6 analyses from Pinney Hollow greenstones and 10 analyses from Stowe greenstones from the Rochester area. Kimball (1991) performed the geochemistry on the 16 samples, and the complete results are listed in her work.

Geochemical signatures of the metabasic rocks from the research area are separated into the zone classification of Coish (1989). Geochemical signatures of greenstones and

Table 13

Major and Trace Element Geochemistry of Greenstones and Amphibolites

UNIT	CZug	CZug	CZug	CZuv	CZuv	CZuv	CZfg	CZfg
SAMPLE	726BGW	726CGW	727GW	726DGW	729GW	730GW	156GW	392GW
LOCATION	E-12	E-12	E-12	E-12	E-12	E-12	S-11	R-13
Major (%)								
SiO_2	44.89	46.97	45.14	49.76	47.19	45.56	46.04	47.42
TiO_2	4.66	3.55	3.41	4.18	3.57	4.51	1.31	2.21
Al_2O_3	16.21	15.48	14.74	13.00	13.64	12.33	17.14	14.22
Fe_2O_3 *	17.65	15.18	14.44	16.85	13.77	16.30	12.94	21.25
MnO	0.26	0.17	0.18	0.24	0.21	0.24	0.18	0.13
MgO	6.17	5.33	6.00	3.60	3.23	3.70	9.02	6.20
CaO	5.44	6.89	8.74	9.96	12.23	10.04	9.26	1.13
Na_2O	4.15	3.70	2.99	3.37	2.74	3.03	3.93	4.41
K_2O	0.70	1.46	0.90	1.76	1.01	1.99	0.00	0.00
P_2O_5	0.82	0.57	0.53	1.56	0.68	1.03	0.22	0.35
TOTAL	100.95	99.30	97.03	104.28	98.27	98.73	100.04	97.32
LOI%	2.83	1.87	1.96	0.90	1.49	1.13	10.30	3.24
Trace (ppm)								
Sc	30	27	22	31	23	31	48	44
V	330	341	263	263	307	223	274	405
Cr	107	87	175	49	45	36	206	38
Co	118	96	95	95	86	92	78	96
Ni	116	76	116	51	46	31	126	63
Cu	16	0	63	42	98	103	28	0
Sr	225	241	482	350	924	312	188	41
Y	47	37	31	64	39	54	23	24
Zr	96	293	219	34	302	76	71	136
Ba	102	181	246	375	141	560	2	4

UNIT	CZag	CZag	CZag	CZag	CZphg	CZphg	CZphg	CZsg
SAMPLE	347GW	488GW	493AGW	493BGW	192GW	238GW	341GW	290GW
LOCATION	U-10	S-14	S-14	S-14	U-12	Y-14	W-10	AA-11
Major (%)								
SiO_2	46.12	47.20	45.27	47.43	54.95	49.60	47.68	46.11
TiO_2	0.76	2.50	0.93	0.88	2.33	1.24	2.01	1.13
Al_2O_3	18.13	14.97	16.80	15.75	13.91	13.88	16.44	16.41
Fe_2O_3 *	8.74	14.51	11.46	9.67	14.66	1.79	15.25	11.41
MnO	0.11	0.20	0.14	0.14	0.26	0.17	0.26	0.20
MgO	6.81	5.33	9.00	7.91	6.89	7.81	9.17	6.69
CaO	13.89	11.73	12.17	13.06	6.29	10.72	7.51	14.07
Na_2O	1.88	1.98	2.91	3.44	3.03	3.04	2.12	2.00
K_2O	0.95	0.26	0.03	0.04	0.01	0.09	0.07	0.41
P_2O_5	0.12	0.36	0.14	0.14	0.31	0.20	0.30	0.21
TOTAL	97.51	99.04	98.85	98.46	102.64	97.54	100.81	98.64
LOI%	3.36	3.89	6.83	6.70	7.07	2.39	4.37	6.74
Trace (ppm)								
Sc	37	47	46	43	38	42	40	50
V	212	339	251	234	328	333	282	337
Cr	482	208	844	473	145	369	238	373
Co	67	80	89	73	81	68	96	68
Ni	298	87	432	297	118	167	165	140
Cu	38	27	68	99	65	20	39	23
Sr	155	367	266	334	180	135	257	156
Y	21	37	26	25	29	32	28	31
Zr	33	201	42	39	115	65	91	55
Ba	348	274	3	3	3	8	8	43

* Total FeO as Fe_2O_3 CZug = greenstone CZuv = amphibolite
LOI% = Loss on ignition
Analyses by ICAP at Middlebury College, Middlebury, Vermont.

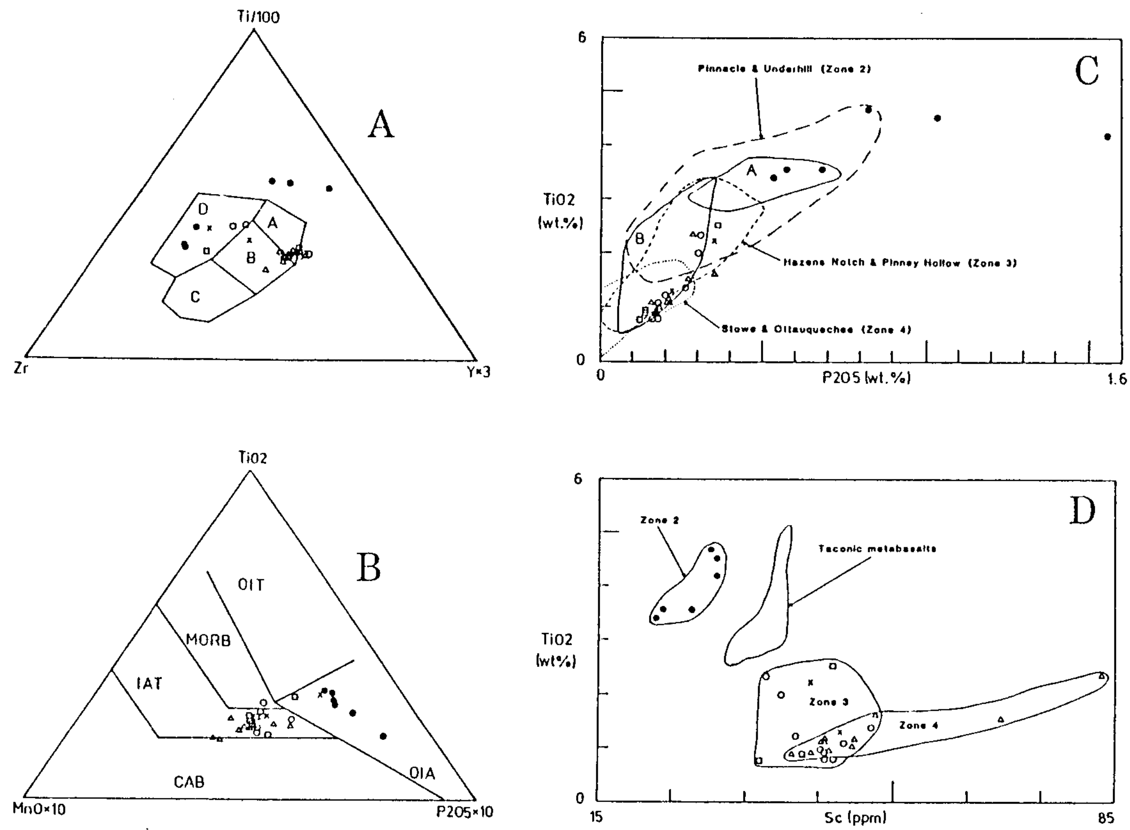

Figure 1. Geochemical diagrams: Closed circle = Underhill (6 analyses), open square = Mount Abraham (4 analyses), x = Fayston (2 analyses), open circle = Pinney Hollow (8 analyses), open triangle = Stowe (11 analyses). **A.** Zr-Ti-Y diagram. Tectonic fields from Pearce and Cann (1973): A and B = ocean floor basalts and low K tholeiites, C and B = calc-alkaline basalts, and D = within-plate basalts. **B.** TiO_2-MnO-P_2O_5 diagram. Tectonic fields from Mullen (1983): OIT = ocean island tholeiites, MORB = mid-ocean ridge basalts, IAT = island arc tholeiites, CAB = calc-alkaline basalts, OIA = ocean island alkali basalts. **C.** TiO_2-P_2O_5 diagram. Fields A and B from Coish et al. (1985). Pinnacle and Underhill (Zone 2), Hazens Notch and Pinney Hollow (Zone 3), and Stowe and Ottauquechee (Zone 4) fields from Coish (1987, 1989). **D.** TiO_2-Sc diagram. Fields for Zones 2, 3, and 4 are from this study. Field of Taconic metabasalts is from Ratcliffe (1987). Reproduced from Walsh and Kimball (1989, fig. 2).

amphibolites from the Underhill Formation (Zone 2 of Coish) are distinct from those of the Fayston Formation, Mount Abraham Schist and Pinney Hollow Formation (Zone 3 of Coish), and Stowe Formation (Zone 4 of Coish; figure 1). The metabasic rocks of Zone 2 have very high TiO_2, P_2O_5, Zr, and Y and plot as within-plate and alkali basalts (figure 1 a & b). The geochemistry of Zone 2 is similar to Group A of Coish et al. (1985) which includes Huntington (Underhill Formation) and Tibbit Hill (Pinnacle Formation) greenstones and amphibolites (figure 1 c). Zone 3 rocks have intermediate TiO_2, P_2O_5, Zr, and Y and plot as transitional from within-plate basalts to tholeiitic basalts (figure 1 a & b). Zone 3 geochemistry is similar to Group B greenstones of Coish et al. (1985) from the Underhill and Pinney Hollow Formations, and to Type 1 greenstones from the Stowe Formation (Coish et al., 1986). Zone 4 rocks have intermediate to low TiO_2, P_2O_5, Zr, and Y and plot as tholeiitic basalts (figure 1 a & b). In the absence of REE data, the TiO_2-Sc diagram shows the distinction between the three zones better than any other trace element diagram (figure 1 d). Zone 2's high TiO_2 versus low Sc values are similar to Hawaiian alkalic basalts and transitional basalts from the Kenyan rift (Basaltic Volcanism Project, 1981), a similarity first recognized for Group A by Coish et al. (1985). Zone 3's intermediate values are again similar to Group B of Coish et al. (1985) which are in the MORB range. Zone 4's lower TiO_2 and higher Sc also plot within the MORB range and overlap with the Zone 3 field. This overlap coincides with detailed mapping that indicates the Pinney Hollow and Stowe Formations are depositionally related and are part of an east-west rift-clastic facies change (Stanley et al., 1987a. and 1987b.; and Kraus et al., 1988). The geochemical overlap between the Fayston, Mount Abraham, and Pinney Hollow greenstones within Zone 3 supports evidence that all three units are part of the east-west rift-clastic facies change (Walsh and Stanley, 1988), and that their current position west of the Stowe Formation is a reflection of their previous position in the rift-clastic sequence. The Underhill Formation includes the western most equivalent of the predominant rock type in the rift-clastic sequence of central Vermont--a noncarbonaceous, quartz-muscovite-albite-chlorite schist. A depositional transition between the Underhill Formation and the Fayston Schist has not been recognized in central Vermont, and mapping indicates a tectonic boundary with a complex history between the two units.

With the correlation of the lithologies from the pre-Silurian sequence of central Vermont with those of the Taconic allochthons well established, it is logical that a geochemical correlation should follow. Metabasalts from the Rensselaer Plateau and Chatham slices (Ratcliffe, 1987) occur in rock types that are similar to those of the central Vermont rift-clastic sequence. All characteristics of the Taconic metabasalts indicate that they are tholeiitic to transitional alkalic basalts and basaltic tuffs. The geochemistry of the metabasalts is similar to that of the Tibbit Hill volcanics. The Taconic metabasalts have a high TiO_2 content similar to Group A of Coish et al. (1985), but the Sc values are higher. The geochemical comparison is most striking in the TiO_2-Sc diagram where Taconic metabasalts plot between Zones 2 and 3, thus filling in the geochemical void in the central Vermont rift-related metabasic rocks (figure 1 d).

The agreement of the geochemical data with the recognition of a significant tectonic boundary between the Underhill and Monastery Formation could imply that the Taconic root zone of Stanley and Ratcliffe (1985) is located between the two units in central Vermont. There is, however, no geochemical data from the greenstones in the western part of the Fayston Formation, nor is there any data from the Monastery greenstones. It is possible that these greenstones have intermediate geochemical signatures. North of the Winooski River, current work by Peter and Thelma Thompson (Thompson and Thompson, 1987, 1989) shows a depositional transition from Fayston-like to Underhill-like lithologies. If this relationship is correct it suggests that the tectonic boundary observed in the vicinity of Appalachian Gap terminates to the north. It is well known that no allochthonous masses lie to the west of the central Vermont pre-Silurian sequence. The absence of allochthons is the result of either of the following scenarios:

1. The Taconic-correlatives from central Vermont were transported westward and subsequently eroded (Stanley and Ratcliffe, 1985).

OR

2. The Taconic-correlatives never left, and they currently lie in (or under) the central Vermont sequence.

The true answer may be a combination of the two possibilities in that the area from Appalachian Gap to the Winooski River marks the transition from a tectonic root zone to a coherent depositional sequence because westward displacement of the allochthons decreases towards the north, and one would expect more and more of the allochthons to appear in the root zone. Once geochemical data from the remaining Monastery and Fayston greenstones is available it will be possible to determine whether the rocks in the Fayston - Buels Gore area represent the Taconic correlatives or if the geochemical and tectonic gap still coincide.

Gray Carbonate Schist (Cog)

Prior to petrographic analysis it was thought that the gray carbonate schist of the Ottauquechee Formation might be a metaigneous rock due to the abundance of chlorite and carbonate identifiable in hand sample. The schist has a rather unique and unusual assemblage of minerals--one not characteristic of a metamorphosed basaltic rock. The rock has a significant amount of quartz (average of 25%), and some of the quartz grains contain needles of rutile. If the schist is not a metabasaltic rock perhaps it is a volcaniclastic or metafelsic rock. Two samples of the gray carbonate schist were analyzed by ICAPS to determine their whole rock major and trace element geochemistries (table 14). The major element chemistry looks very similar to that of the metabasaltic rocks, therefore ruling out a possible felsic origin. The protolith may have been a volcaniclastic rock.

The loss-on-ignition values are extremely high indicating the abundance of hydrous phases. When the values of Ti, Y, Zr, and P are used in the geochemical diagrams the schist plots as within plate basalt similar to Zone 2. Perhaps this anomalous data can be explained in a way that does not directly concern the protolith. Both samples were collected close to prominent Tn fault zones. The limited occurrence of the gray carbonate schist does not provide the luxury of choosing samples otherwise. Sample 254 has a mylonitic texture and sample 300X can be classified as a protomylonite. O'Hara and Blackburn (1989) report the enrichment of the minor and trace elements Ti, P, Zr, Y, and V in mylonites that exhibit significant volume loss. During mylonitization of felsic gneiss country rock the high-field-strength elements were immobile while significant volume loss depleted the major, more mobile elements resulting in a relative enrichment of the trace and minor elements. It is quite possible that similar mechanisms produced the odd geochemistry of the gray carbonate schist.

STRUCTURE

The structures observed in the Fayston - Buels Gore area are characteristic of a polydeformed hinterland terrane. Typical structures of such a terrane include multiple generations of foliations, folds, and faults. Structural fabrics within the research area suggest a minimum of at least four foliations, four fold generations, and three generations of faults. The foliations include a pervasive schistosity (Sn), a relict schistosity (Sn-1), a crenulation cleavage (Sn+1), and kink bands. The folds include isoclinal reclined folds and sheath folds (Fn), relict isoclinal folds (Fn-1), tight folds and chevron folds (Fn+1), and open folds (Fn+2). The faults include early or pre-peak metamorphic faults (Tn-1), syn-metamorphic faults (Tn), and late or post-peak metamorphic faults (Tn+1). The Tn and Tn+1 faults have associated kinematic indicators. A post-metamorphic period of deformation is characterized by fractures, dikes, and normal faults.

All of the structures discussed are represented on the geologic map (Plate 1) and geologic cross section (Plate 2). Measured structural data are plotted on lower hemisphere equal area projections on Plate 3.

Table 14

Major and Trace Element Geochemsitry of the Gray Carbonate Schist

UNIT	Cog	Cog
SAMPLE	254GW	300XGW
LOCATION	Z-13	AA-11
Major (%)		
SiO_2	46.39	46.48
TiO_2	3.07	2.63
Al_2O_3	12.88	11.61
Fe_2O_3 *	12.15	12.98
MnO	0.40	0.17
MgO	8.10	11.56
CaO	11.99	9.79
Na_2O	2.03	0.06
K_2O	0.00	0.63
P_2O_5	0.45	0.41
TOTAL	97.46	96.32
LOI%	12.39	15.16
Trace (ppm)		
Sc	25	26
V	286	287
Cr	245	370
Co	81	86
Ni	175	294
Cu	0	68
Sr	546	499
Y	28	30
Zr	216	172
Ba	0	24

* Total FeO as Fe_2O_3
LOI% = Loss on ignition
Analyses by ICAP at Middlebury College, Middlebury, Vermont.

***

The structural data from the Fayston - Buels Gore area are divided into four domains. Domain 4 is located on the west flank of the Green Mountain anticlinorium (GMA) and is characterized by overturned west-verging folds (Fn+1), and by both a north-northeast trending and south-southeast moderately dipping dominant schistosity (Sn) and crenulation cleavage (Sn+1). Domain 3 is located at the crest of the GMA and is characterized by upright folds (Fn+1) that deform the dominant schistosity (Sn) into a relatively horizontal position. Domain 2 is located on the east flank of the GMA and is characterized by a north-northeast trending and south-southeast moderately dipping dominant schistosity that is deformed by east-over-west anticlinorium-related folds. Domain 1 is located in the Mad River Valley and is characterized by a north-northeast trending and south-southeast steeply dipping dominant schistosity. Folds of the GMA are not readily apparent in Domain 1.

FOLIATIONS

Dominant Schistosity (Sn)

The term dominant schistosity (Sn) is a relative expression that is applied to the most pervasive foliation seen at a specific outcrop. Sn is used as a datum to which all other structural fabrics at an outcrop are compared. The use of Sn across the research area does not imply that the dominant schistosity is all the same age. The ages of the structural fabrics are discussed at the end of this chapter and in Chapter 5. The assignment of the terms n-1, n, n+1, and n+2 is based on cross cutting and superposed relationships observed at the outcrop scale.

The dominant schistosity is defined by the parallel arrangement of phyllosilicates and quartz ribbons. In the coarsest rocks (CZfb, CZfq, CZaw, CZphw) the foliation is often a gneissosity. In Domain 1 Sn is the most pervasive fabric recognized. The dip of Sn varies from ≈70-80° SE in the east, and ≈60-70° SE in the west of Domain 1, with an average strike and dip of N13°E,70°SE. In Domain 2 Sn is still the most pervasive fabric, but later folds (Fn+1) deform Sn and become more common to the west. The dip of Sn varies from ≈60-70°SE in the east, and ≈40-50°SE in the west of Domain 2, with an average strike and dip of N17°E, 49°SE. In Domain 3 Sn does not have a consistent orientation. Poles to Sn in Domain 3 (Plate 3) describe an east-west trending, north dipping girdle. The π pole to the best-fit great circle (S7°E,11°) approximates the point maximum to Fn+1 fold hinges (S6°E,15°) indicating that Sn is deformed by Fn+1 into a south-plunging fold. Poles to Sn in Domain 4 (Plate 3) generally define a statistical plane with an orientation of N6°E,42°SE. The poles also describe a poorly developed east-west trending, north dipping girdle. The π pole to the best-fit great circle (S1°E,6°) also approximates the point maximum to Fn+1 fold hinges (S9°E,14°) indicating that Sn is only moderately deformed by Fn+1 into a south-plunging fold.

Relict Schistosity (Sn-1)

A relict schistosity (Sn-1) is only rarely observed mesoscopically in the hinge regions of Fn folds, and microscopically as inclusion trails in porphyroblasts. In the field Sn-1 is observed only in the more competent greenstones, quartzites, and folded quartz veins. The less competent schists rarely preserve the older schistosity. On the limbs of Fn folds that contain Sn-1 the older schistosity is parallel to the younger Sn, but at Fn hinges Sn and Sn-1 schistosities are near perpendicular. In the quartzites Sn-1 exactly parallels compositional layering defined by alternating black and white, graphite-rich and graphite-poor layers. Sn-1 observed in greenstones and quartz veins is usually marked by folded chlorite laminations. In thin section Sn-1 is preserved as folded inclusion trails in albite porphyroblasts.

Because Sn-1 is only observed as a relict schistosity in the most competent rock types, its initial distribution is not completely known. Sn-1, however, occurs throughout the area, and is most evident in Domain 4. Since Sn-1 occurs most often as a schistosity parallel to compositional layering, especially in Domain 4, and no older schistosities are observed, it is believed to be the remnants of an early bed-parallel schistosity. This interpretation is consistent with observations of an early bed-parallel schistosity immediately to the west in Starksboro by DiPietro (1983). If this is the case, the Sn-1 seen in the large Fn fold of CZfq south of Burnt Rock Mountain may represent remnant stratigraphy. In Domain 1 the compositional layering (Sn-1) seen in the western domains is only preserved in more competent greenstones and quartzites and never in the schists.

Sn-1 appears to represent a regional schistosity that predates the development of Sn. Sn-1 may also represent an initial stage of Sn development in which the transition from older schistosity to younger schistosity represents a continuum in time and deformation, and not two events separated by a discreet time interval. In this sense it is possible that Sn-1, which is not associated with compositional layering, as seen in the schists of Domain 1 represents the Sn of the western domains.

Crenulation Cleavage (Sn+1)

A crenulation cleavage (Sn+1) is observed from the center of Domain 2 westward. Sn+1 is only observed in the fine grained, less competent schists, especially the Mount Abraham.

Sn+1 does not occur in the coarse grained granofels and metawackes, and is rare in the coarse albitic schists. Sn+1 is axial planar to Fn+1 folds and, in its most recognizable form, varies from a discrete to zonal cleavage. Borradaile et al. (1982, p. 152) define a discrete crenulation cleavage as one with distinct boundaries recognizable in both hand sample and thin section, whereas a zonal crenulation cleavage lacks distinct boundaries. In its most pervasive form Sn+1 varies from discrete cleavage planes with a spacing of 0.5 cm or less to a penetrative schistosity that transposes Fn+1 folds. Sn+1 as a penetrative schistosity is best-recognized in Domain 2 and 4 in the vicinity of Tn+1 fault zones. Crenulations on the surface of Sn caused by the intersection of the cleavage (Sn+1) and the schistosity (Sn) are sub-parallel to Fn+1 fold hinges. The orientation of Sn+1 is parallel to the measured axial surfaces of Fn+1 folds shown on Plate 3, and the measured crenulation cleavages were not plotted on a separate net.

Kink Bands

Kink bands occur only in the eastern part of Domain 1. The kink bands are observed in the Ottauquechee black phyllite and the silvery green schists of the Pinney Hollow and Stowe Formations. The kink bands are not ubiquitous in Domain 1 and only occur locally at individual outcrops. All the kink bands observed strike northwards and dip steeply around 70° - 80° to the west. The spacing between kink bands is consistently between 3 and 10 cm. The relative displacement sense for the kink bands is always west side down.

LINEATIONS

Mineral and grain cluster lineations form the common linear fabrics associated with the dominant schistosity. Mineral lineations consist largely of chlorite streaks, but streaks of biotite, magnetite, and graphite also occur. Quartz rods are the common type of grain cluster lineation.

The chlorite streaks commonly occur as stretching lineations due to the preferential elongation of tabular faces of the chlorite. In Domain 3 the chlorite streaks are usually primary (i.e. the chlorite appears to have nucleated on its own during metamorphism and not at the expense of another mineral). In Domains 2, 4, and to a greater extent 1, secondary chlorite streaks are more abundant than primary ones. Secondary chlorite streaks are characterized by the nucleation and growth of chlorite around garnet during retrograde metamorphism. Secondary chlorite streaks vary from simple chlorite pressure shadows around garnet to elongate streaks that contain clots of chlorite and fragmented garnet at their origin.

In the eastern and western-most parts of the area where the dominant schistosity dips consistently eastward, mineral lineations consistently trend steeply down the dip of the schistosity with rakes ranging from 70° to 90° both north and south. The average trend and plunge of mineral lineations in Domains 1, 2, and 4 is S70°E, 69°, S78°E, 51°, and S68°E, 53°, respectively. Mineral lineations in Domain 3 are not as prominent as in the other domains and for this reason, sufficient data are lacking to make a conclusion on their statistical orientation.

Quartz rod lineations are as equally abundant as mineral lineations. Quartz rods vary in width from a few millimeters up to 5 cm, and can reach lengths up to 1 meter when viewed on the plane of the dominant schistosity. When viewed down the dip of the schistosity quartz rods appear as discontinuous lens shaped pods of white quartz.

Quartz rods have essentially the same orientation as other mineral lineations. Quartz rod measurements from Domain 1 are separated into those from schist versus those from greenstone and quartzite to determine if there is a difference in orientation from rocks of different competence. The average trend and plunge from the schist is S70°E, 66°, and N90°E, 69° from the greenstone and the quartzite. A difference is discernible, but the significance can not be stated due to the limited measurements from the greenstone and quartzite. Quartz rods from Domain 2 are also quite similar to mineral lineations from Domain 2. The quartz rods cluster and have an average trend and plunge of S71°E, 49°. Quartz rod lineation measurements from Domain 3 do not have a statistical cluster, but appear concentrated in the southeast and northwest--as expected when the east-dipping schistosity is folded to west-dipping orientations. Quartz rods from Domain 4 have an average orientation of S44°E, 35° which is similar to, but sligthly shallower in plunge than those of Domains 1 and 2.

FOLDS

Fn Folds

By definition, Fn folds are folds in which the axial surface coincides with the plane of the dominant schistosity Sn. All Fn folds are isoclinal with their limbs parallel to Sn. Fn folds play a major role in the distribution of the various rock types. Virtually all mesoscopic folds are intrafolial, and are commonly rootless. The majority of structural data on Fn folds comes from the eastern half of the research area where the schistosity has a consitently planar orientation. The overprinting of Fn folds by Fn+1 folds in Domain 3 often inhibits identification of the earlier folds. Isoclinal reclined folds and sheath folds are the two types of Fn folds in the Fayston - Buels Gore area.

Isoclinal Reclined Folds

The majority of Fn folds observed at outcrop scale are isoclinal reclined folds. Such folds occur as folded quartz veins, folded lithologic contacts, and folded compositional layering in greenstones and quartzites. Quartz veins deformed into Fn folds generally display hinges which coincide with quartz rod aggregate lineations. In many places the axes of the Fn folds are essentially parallel to the stretching lineations.

Reclined Fn folds vary in size from small centimeter scale folded quartz veins to folds many meters across (N-4, V-10). At one location an Fn fold in a black and white banded quartzite within the Thatcher Brook Member is approximately 15 cm from limb to limb, yet each limb is on the order of 2 meters in length. Isoclinal reclined Fn folds are commonly referred to as "tube structures" due to their outcrop appearance. In such instances the tube is an intrafolial, rootless folded quartz vein that is both cored and surrounded by matrix material.

The orientation of Fn fold hinges is shown on Plate 3. In Domain 1 Fn fold hinges are separated into those that occur in the more competent greenstones and quartzites versus those that occur in the less competent schists. In general most Fn fold hinges plunge steeply down the surface of Sn in accordance with the reclined nature of such folds. Fn folds measured in the schists have an average trend and plunge of S33°E, 80° and show a tight cluster, very similar in fact to the measured quartz rod and mineral lineations from Domain 1. Fn folds measured from greenstones and quartzites in Domain 1 have an average trend and plunge of N83°E, 72°. Measured data points are distributed along a great circle which coincides with the average plane of Sn. The data show that the orientation of Fn fold hinges in the more competent lithologies are not as clustered as they are in the schists, probably as a result of a smaller degree of rotation.

Fn fold hinges in Domains 2 and 4 have an average orientation of S56°E, 38° and S43°E, 36°, respectively. These orientations show that the fold hinges have shallower rakes on the surface of Sn than the rakes of hinges in Domain 1, suggesting that the folds have not gone through as great a degree of rotation. Fn fold hinge data from Domain 3 are similar to the quartz rod data in that the data are more common in the southeast and northwest.

Sheath Folds

Aside from the evidence for isoclinal reclined folds there is also evidence that suggests the existence of sheath folds within the study area. The evidence, however, is limited to a single example at the mesoscopic scale, and interpreted structural data at the macroscopic scale. All evidence for sheath folds appears to be confined to the eastern half of the area.

A single outcrop example of a sheath fold is located at station 498 (W-14). The exposure shows half of a synformal sheath fold that is approximately 20 cm long and 5 cm wide. The western half of the fold is the only half remaining and, it is concave on the eastern side. The fold occurs in the silvery green schist of the Pinney Hollow Formation and is primarily a folded quartz vein with a schist core. The mean trend and plunge of the fold axis is parallel to the mineral lineations and measures S68°E, 63°.

At map scale, it is the orientation of mesoscopic Fn folds and the distribution pattern of the lithologies that indicates the presence of macroscopic sheath folds. The entire body of CZa1 (V-10) appears to be an antiformal sheath fold. Fn fold hinges from the north end of the CZa1 body plunge to the north while those at the south end plunge to the south, and the entire body is elliptically shaped.

Similar structural geometry is evident at several other locations in the study area. At location S-14 small (100-300 meters along the long dimension of the lens shaped bodies) antiforms of CZa1 are apparently poking up through the Granville (Cg) as windows. The lithologies present at this location are similar to those around CZa1 mentioned in the above paragraph, but in the previous case the mass of CZa1 was outside of, and to the east of the main mass of Granville. This relationship supports the idea that Mount Abraham (CZa1) and the Pinney Hollow along the boundary between Domains 1 and 2 structurally underlie the main mass of Granville. Within the Ottauquechee belt two masses of Pinney Hollow also appear to have antiformal sheath fold geometries (Z-13 and AA-11). Again the existence of antiformal folds of Pinney Hollow as windows in the Ottauquechee belt indicates that the carbonaceous lithologies structurally overly the non-carbonaceous lithologies.

Origin of Reclined and Sheath Folds

Reclined and sheath folds can form by the same or similar mechanisms. The two types of folds are separated for descriptive purposes based on their geometries. The major question concerning the origin of these folds involves the role of simple shear and pure shear mechanisms in their formation.

Ramsay (1979) states that fold hinges inclined to the principal strain axes undergo rotation within the X-Y plane towards the X-axis. The amount of rotation is due to the initial orientation with respect to X or Y, the magnitude of strain, and the nature of the strain. Folds that are initially parallel or close to Y may show little or no rotation even under large strains. Sub-horizontal Fn fold hinges are never observed in the research area, and this may indicate that the initial orientation of the Fn fold axes was never parallel to the Y-direction. The variation in orientation between Fn folds in quartzites and greenstones versus schists suggests that the rheological properties of the various lithologies plays an important role in the final orientation of Fn fold hinges.

Bell (1978) states that fold axes within mylonite zones are rotated into parallelism with the X-axis. This rotation is due to both inhomogeneous bulk flattening during pure shear and inhomogeneous simple shear. Ghosh and Sengupta (1987) report the rotation of sub-horizontal folds to down dip orientations by a combination of first simple shear and then pure shear in the Singhbhum shear zone in India.

Another possible way to produce Fn fold hinges with axes parallel to the X-direction is by constriction, a common mechanism in the formation of salt domes (Nicolas, 1987). In the case of constriction, layers initially oriented at low angles to the X-direction are deformed until fold axes parallel the X-direction. Another possible method of producing fold hinges parallel to the X-direction is by sheath folding.

Sheath folds form in zones of high shear strain such as the base of nappes (Cobbold and Quinquis, 1980; Henderson, 1981; and Nicolas, 1987) and in regional shear zones (Cobbold and Quinquis, 1980; Ghosh and Sengupta, 1987; and Park, 1988).

Experimental studies on the development of sheath folds were conducted by Cobbold and Quinquis (1980). In their study they developed three models for the development of sheath folds: Homogeneous simple shear, bulk simple shear perturbed by a resistant layer with initial deflections (i.e. nappes), and bulk simple shear with the development of boudinage in resistant layers at the onset of deformation (most applicable to layered metamorphic rocks). Cobbold and Quinquis conclude that sheath folds develop passively during bulk simple shearing often due to the amplification that developed prior to, or as the result of deformation. In each of the experiments shear strain values of $\Gamma>10$ were used.

Henderson (1981) reports the development of sheath folds with horizontal fold axes parallel to the principal X-axis of the finite strain ellipsoid. Henderson (1981, p. 203) states that "outcrop patterns of truncated isoclinal sheath folds resemble cylindrical folds except in relatively small areas around paraboloidal caps." Henderson studied horizontal sheath folds and emphasized that exposures of sheath fold caps may be quite rare for field geologists conducting bedrock mapping. If caps of horizontal sheath folds are a rarity then the caps of reclined sheath folds should be equally rare, if not more of a rarity, due to their inclined orientation with respect to the earth's surface.

Unlike Henderson (1981) who reports natural examples of horizontal sheath folds at the base of nappes, Ghosh and Sengupta (1987) and Park (1988) report natural examples of thrust-related sheath folds from steeply dipping shear zones. The existence of shear zones within the pre-Silurian hinterland of central Vermont is well documented (Stanley et al., 1986; Stanley et al., 1987a; Stanley et al., 1987b; Armstrong et al., 1988; Stanley et al., 1988; Stanley and Armstrong, 1989; and Stanley, 1989). Furthermore, Stanley et al. (1989) suggest the existence of sheath folds within the central Vermont hinterland. It is becoming apparent that sheath folds do exist in central Vermont, and that evidence of their existence has previously been overlooked.

Fn-1

A relict generation of folds is rarely preserved in the research area. Fn-1 is recognized as an earlier generation of folds refolded by Fn. The relict schistosity Sn-1 is axial planar to all Fn-1 folds, although the recognition of Sn-1 does not always coincide with an Fn-1 fold.

Fn-1 folds in the western part of the area are characterized by a folded Sn-1 schistosity that is parallel to a compositional layering. Such Fn-1 folds are recognized by a refolded pattern of discreet compositional layers. The Fn-1 folds are deformed by Fn folds and are transposed by Sn.

Fn-1 folds observed in Domain 1 appear as refolded Fn folds in the schists as well as refolded patterns of discreet compositional layering in greenstones. The folds in the schists occur as refolded quartz segrgeations that contain an earlier Sn-1 schistosity which is transposed by Sn. Such refolded quartz segregations are rare in Domain 2 and have not been observed in Domains 3 and 4. Fn-1 folds in Domain 1, therefore, appear to be an earlier phase of deformation in a continuum of Sn development.

Fn+1

Two types of Fn+1 folds occur in the study area: Regional folds and local fault-related folds.

Regional Fn+1

The regional Fn+1 are the predominant Fn+1 fold type, and are the most readily recognized mesoscopic fold type in the study area. Fn+1 folds are ubiquitous in the western half of the study area (from the center of Domain 2 westward). The occurrence of regional Fn+1 folds becomes more sporadic and limited from the center of Domain 2 eastward.

Fn+1 vary in size from microscopic thin-section scale folds to the large macroscopic antiforms and synforms of Domains 3 and 4. The inter-limb angle of Fn+1 ranges from open to tight with an average range from 90°-30°. Fn+1 folds predominantly posses a similar fold geometry. In the finer grained rock types, especially the Mount Abraham Schist, Fn+1 commonly occurs as chevron folds.

The axial surfaces of Fn+1 folds in Domain 2 are sub-parallel to slightly steeper than Sn with an average strike and dip of N37°E, 55°SE. This average, however, is not thoroughly conclusive because the data do not plot as a single point maximum, but rather as a partial great circle. The axial surfaces of Fn+1 in Domain 3 are more upright and near perpendicular to Sn with an average strike and dip of N18°E, 79°SE, and the data cluster better than in Domain 2 although a partial great circle is described. The great circle distribution may in part be due to later folding by Fn+2, however, the π pole (S10°E, 24°) to the great circle does not correspond to the average trend and plunge of Fn+2 fold hinges (S5°W, 16°). Also, the Fn+1 fold hinges do not clearly fall on a great circle that approximates the average orientation of Fn+1 axial surfaces. If they did it would support the idea that the hinges have been rotated during the development of Tn+1 shear zones. But because the π pole to the Fn+1 poles is close to the average Fn+1 hinge line orientation (S6°E, 15°) and not Fn+2 (S5°W, 16°) the best explanation appears to be that Fn+1 folds are progressively deformed during Sn+1, and therefore Tn+1, development. A similar relationship holds true in Domain 4, although the average axial surface of Fn+1 dips at shallower angle due to the overturned nature of some large Fn+1 folds.

Fault-related Fn+1

The western half of Domain 1 has the unique characteristic in that Fn+1 folds are generally absent. The dominant schistosity (Sn) is consistently planar and uniform in this area. In the eastern half of Domain 1, however, more Fn+1 folds are recognized, but only locally. These local, fault-related Fn+1 folds are not a common fold generation in the area.

The fault-related Fn+1 folds are similar to the regional Fn+1 folds only in that they deform a dominant schistosity--there the similarity ends. Fault-related Fn+1 folds are most commonly tight to near isoclinal with interlimb angles from 45°-10°. These Fn+1 folds are observed only along shear zones, are often intrafolial, and are commonly transposed by a fault zone schistosity. The axial surfaces of the fault-related Fn+1 folds are much steeper than those of the regional Fn+1 folds seen to the west. The average strike and dip of six measured axial surfaces of fault-related Fn+1 folds is N15°E, 85°SE. Hinges of fault-related Fn+1 folds also differ substantially from their regional counterparts. The hinges have moderate to steep plunges down the surface of the fault zone schistosity. The average trend and plunge for six measured hinges is S05°W, 58°.

The sparse occurrence of such Fn+1 folds and the meager amount of data collected prohibits a complete investigation and understanding of these folds. Two possibilities as to the nature and origin of the folds exist. The first is that the folds originated as regional Fn+1, similar to those found to the west, but underwent subsequent transposition and rotation due to the development of shear zones. The second is that the folds are a result of strain hardening in shear zones, and therefore are a unique phenomenon to their respective shear zones.

Kraus (1989, "Troll outcrop") describes fault-related Fn+1 folds that are similar to those in the Fayston - Buels Gore area. Armstrong (1992) cites evidence for both possibilities. At this time substantial evidence is lacking to unequivocally support either possibility in the area. It is likely, and most probable that both scenarios are applicable to some degree.

Fn+2

Fn+2, the youngest generation of folds in the Fayston - Buels Gore area is restricted to the western half of the research area--from the west side of Domain 2 westward. Fn+2 folds are upright, open folds with inter-limb angles that range from 130°-150°. Fn+2 folds are not common structures, in fact they are relatively rare. Fn+2 folds are easy to recognize because they deform the axial surfaces of Fn+1 folds by gently warping them, and by changing the orientation of the surfaces from upright to sub-horizontal. Fn+2 folds are confined to the mesoscopic and microscopic scale as no macroscopic Fn+2 folds have been recognized.

The orientation of Fn+2 folds is quite consistent. Fn+2 axial surfaces strike north-south and dip steeply east to near vertical. The average strike and dip of Fn+2 axial surfaces is N16°E, 80°SE in Domain 2, N12°E, 84°SE in Domain 3, and N26°E, 77°SE in Domain 4. Fn+2 hinges plunge shallowly to the south and have an average trend and plunge of S09°W, 20° in Domain 2, S08°W, 19° in Domain 3, and S4°W, 14° in Domain 4. Due to the open nature of Fn+2 folds they are most often symmetrical. Asymmetric Fn+2 folds with east-over-west, or clockwise, rotation senses do occur. Asymmetric folds with west-over-east rotation senses were not observed in the area.

FAULTS

Faults in the Fayston - Buels Gore area are categorized by their timing with respect to peak metamorphic conditions. Faults that predate the peak metamorphic conditions and the development of the current penetrative schistosity, are called pre-peak or early metamorphic faults (Tn-1). Faults that occur during the peak of metamorphic recrystallization, and are generally parallel to the associated dominant schistosity are called syn-metamorphic faults (Tn). Faults that occur significantly after the peak metamorphic recrystallization episode are termed post-peak or late metamorphic faults (Tn+1). An important note is that the term "peak metamorphic conditions" in the description of these faults refers primarly to the textural development of a pervasive metamorphic foliation during ductile conditions.

Pre-Peak Metamorphic Faults (Tn-1)

Pre-peak metamorphic faults are the most cryptic structures in the area because any fault fabric initially associated with the fault is subsequently annealed by peak metamorphic conditions. The principal criterion for the recognition of Tn+1 faults is the presence of upper and lower plate truncations.

At least four Tn-1 faults occur in the Fayston - Buels Gore area. These Tn-1 faults are the lower boundaries of pre-peak metamorphic thrust slices. The four Tn-1 thrust slices that occur in the research area are, from west to east, as follows:

1. Fayston over Battell and Monastery
2. Mount Abraham (CZa2 and CZa4) over Fayston and Granville
3. Granville (Cg) over Pinney Hollow, Mount Abraham (CZa1), and Fayston
4. Ottauquechee over Pinney Hollow

The first Tn-1 fault is exposed in Domain 3. In the area of the eastern-most antiform of the GMA (M-13) lithologies in the upper plate (CZfb and CZa3) that are depositional with the Fayston schist (CZf) are truncated along the Tn-1 contact. The Battell (Cbw) and Monastery (CZm) lithologies of the lower plate are also truncated along this contact. CZm is considered depositional with Cb below the fault and with CZf and Cg above the fault. At the fault, however, CZm is depositional with Cb of the lower plate, but juxtaposed with CZa3 of the upper plate. Furthermore, no Cbw occurs above the fault anywhere in Domain 3. Cbw does not reappear until the western side of the area in Domain 4 where it is in depositional contact with the Monastery. The occurrence of the graphitic schist with black marble is a crucial marker in the recognition of this Tn-1 fault. Where the contact is exposed it is sharp and not parallel to the dominant schistosity. In the region of the GMA, therefore, the Battell is exposed as the lowest structural level due to the formation of a tectonic window in the core of a GMA antiform. In Domain 4 the Fayston is structurally above but depositionally below the Monastery on the overturned limb of a major antiform. The contact is quite dissimilar to the one in Domain 3 and it is possible that the Tn-1 fault was truncated eastward of the current CZm and CZf contact in Domain 4.

The second Tn-1 fault is at the base of the Mount Abraham thrust sheet. The Mount Abraham thrust sheet is made up of CZa2 and CZa4 in Domains 2 and 3, respectively. At the Mad River Glen ski area on the east side of Stark Mountain CZa4 cuts down through the Fayston (CZf) and into a large depositional layer of Granville (Cg) within CZf. The Granville is truncated along the Tn-1 fault at the base of the thrust sheet. In Domain 2 the thrust sheet has been disarticulated by subsequent Tn and Tn+1 faults. As CZa2 and CZa4 are traced southward into the area mapped by Haydock (1988) the lithologies converge and the thrust sheet is continuous. The convergence is due to the southward plunge of the eastern-most GMA antiform.

The third Tn-1 fault is at the base of the Granville thrust sheet. The Granville thrust sheet consists entirely of the carbonaceous albitic schist (Cg). Near the summit of an unnamed hill behind South Fayston Cemetery (T-14) quartzites in the upper plate of Granville carbonaceous albitic schist and greenstones in the lower plate of Mount Abraham schist (CZa1) are truncated along a sharp pre-peak metamorphic contact. The lower plate rocks are exposed in three tectonic windows near the summit of the hill. A similar relationship exists 2.7 km to the north near the summit of another unnamed hill (V-10). In this instance, however, the Mount Abraham Schist (CZa1) is not exposed in a window, but is located to the east of the Granville thrust slice.

The fourth Tn-1 fault is located in the Ottauquechee belt of Domain 1. The silvery green schist and quartzose schist of the Pinney Hollow are exposed in two antiformal windows within the Ottauquechee black phyllite (Z-13, AA-11). Mapping demonstrates that the silvery green schist and quartzose schist are in depositional contact indicated by intercalation over a distance of approximately ten meters. Both units, however, end abruptly against a pre-metamorphic contact with the black phyllite. The gray carbonate schist of the Ottauquechee (Cog) is found close to the Pinney Hollow windows, but is never observed within the Pinney Hollow lithologies. To the south Kraus (1989) has evidence that the Ottauquechee is tectonically over both the Pinney Hollow and Stowe.

Syn-Metamorphic Faults (Tn)

Unlike pre-peak metamorphic faults, syn-metamorphic faults are easily recognized by pervasive fault fabrics. Mapped truncations also assist in the identification of Tn faults. Tn faults are the most numerous fault type in the research area.

Because the central Vermont hinterland is a polydeformed, polymetamorphosed terrane the use of the term syn-metamorphic can be misleading and ambiguous. Stanley et al. (1987b) devise a graph for separating syn-metamorphic faults in time with respect to the metamorphic conditions and come up with at least six different possibilities for Tn faults. Such faults span the metamorphic episode from prograde to retrograde conditions. All syn-metamorphic faults encountered in the research area have well developed mylonitic fabrics. It is the presence of such fabrics, therefore, that characterizes any syn-metamorphic fault regardless of whether it occurred during prograde, retrograde, or even composite conditions--a single fault zone could then have a long history from prograde to retrograde conditions. Some faults in the area have both mylonitic and "papery" schist fabrics (the latter is characteristic of Tn+1 faults) and hence have been given a composite symbol on the bedrock map consisting of a solid barb with an open square.

Nine examples of Tn faults are described below. Although more Tn faults occur in the area, the discussion is limited to those faults studied both in the field and petrographically. The faults are referred to by station number and grid location as follows:

250 (Z-12)

This Tn fault is exposed at an outcrop first described in Stanley et al. (1987a, station 19C by Walsh). The fault occurs within the Ottauquechee belt where an antiformal window of Pinney Hollow schist is exposed. At the fault exposure the silvery green schist of the Pinney Hollow is in contact with the Ottauquechee black phyllite. The contact is very sharp and planar, parallel to the trend of the dominant schistosity, and marked by two foliations. The two foliations are restricted to the fault zone, for away from the fault only one foliation--the regional foliation--is observed. The two foliations represent a crude C-S fabric with an east over west sense of asymmetry. Also present near the contact is a boudinaged quartz vein that is transposed into an en echelon array with east over west asymmetry.

The mylonitic fabric at the outcrop is also quite evident in thin section (Plate 4, figure 2). Two samples were collected from the vicinity of the fault. One sample was collected approximately 4 meters from the fault and shows a common assemblage of quartz, muscovite, and chlorite with porphyroblasts of albite and magnetite. The other sample contains only microcrystalline quartz and mica with no trace of albite or chlorite, and only scant amounts of opaques.

313 (Z-11)

Location 313 is a silvery sericite mylonite with small (1mm) patches of rusty material (altered pyrite, limonite?) that crops out in the Ottauquechee black phyllite. The mylonite is characterized by transposed fault-related Fn+1 folds with reclined hinges parallel to the stretching lineation. The rock consists primarily of fine grained quartz and muscovite (80%) with roughly equal proportions of porphyroclastic albite and disseminated opaques (rusty patches). Minor amounts of chlorite, graphite, and tourmaline are present. Textbook examples of mica fish yield an east-over-west sense of shear (Plate 4, figure 3).

429 (R-14)

Sample 429 was collected from a drainage ditch on the north side of the western extension of Bragg Hill Road at an elevation of 1750 feet. The rock is a silvery tan mylonite from a fault between the Mount Abraham (CZa2) and Fayston schists (CZf). The mylonite consists primarily of muscovite (62%) and quartz (26%) with lesser amounts of chlorite, porphyroclastic albite, ilmenite, tourmaline, and apatite. A broken and displaced hard grain (ilmenite) kinematic indicator from the mylonite yields an east-over west sense of shear (Plate 4, figure 4).

413 (Q-16)

Station 413 is located in an unnamed stream at an elevation of 1170 feet. The Tn fault at 413 is located between the Granville (Cg) and the Fayston (CZf). The Tn fault at 413 is unique due to the presence of a very coarse grained tan albitic schist found in between two mylonite zones. The mylonites are a tannish-white muscovite (42%), quartz (29%), and ankerite

(23%) rich rock with minor amounts of chlorite, albite, ilmenite, and apatite. The mylonite zone is approximately 5 m thick including the albitic schist in the center. The schist is a very coarse light tannish-green albitic schist. The principal constituents are albite (41%), muscovite (25%), quartz (20%), chlorite (10%), and ilmenite (4%). Garnet, tourmaline, epidote, and apatite are present as albite inclusions, and the garnet occurs as 0.1 mm euhedral and unaltered crystals. The albite occurs as very large porphyroblasts up to .75 cm in diameter and is twinned on the albite law. The coarse albitic schist is possibly the result of excess Na^+ being transported from the fault zone fluids into the surrounding rock. Phillips and Griffen (1981) report that ankerite is a common hypothermal vein mineral, and it is possible that there is some link between the abundance of albite and ankerite, and the fault zone fluids.

235 (Y-13)

Sample 235 crops out in a stream at an elevation of 770 feet. The Tn fault is located between the Pinney Hollow (CZph) and the Thatcher Brook (Cotb) lithologies. The mylonite zone is approximately 2 meters wide and is characterized by a well-foliated, tan to greenish-white mylonitic quartzite. The mylonite consists mainly of quartz (46%) and muscovite (27%). Similar to location 413, 235 also contains a significant amount of fine-grained ankerite (19%). Small porphyroclastic albite (3%), fragmented euhedral tourmaline (2%), opaques and epidote constitute the remainder of the mylonite. No asymmetric fabrics are recognized in 235, and this is possibly due to the fact that pure shear (bulk flattening strain) predominated over simple shear during the development of the fault zone.

Mylonitic fabrics were also studied at four other locations: 245 (O-41), 254 (P-43), 256 (Q-43), and 273 (O-44). Sample 245 occurs in the Thatcher Brook Member near a small serpentinite lens. Sample 254 is from the Ottauquechee gray carbonate schist (Cog) within the black phyllite. Samples 256 and 273 both occur in the quartzose schist of the Pinney Hollow (CZphq). At all four locations well developed mylonitic fabrics are present. No asymmetric kinematic indicators, however, were identified in thin section. It is possible that pure shear mechanisms predominated over simple shear during the formation of the four mylonites, or that pure shear overprinted simple shear and that continued recrystallization obliterated the earlier simple shear fabric.

Numerous mylonitic fabrics occur along the western boundary of Domain 4. The contacts between CZm and CZu, and CZu and CZuql are tectonic contacts marked by mylonites that are parallel to both Sn and Sn+1. Fn+1 fold hinges are transposed and rotated to down dip orientations at locations where the mylonites are parallel to Sn+1 (E-9 and E-27). This zone of syn-metamorphic faulting marks a significant tectonic boundary that coincides with the garnet isograd originally identified by Cady et al. (1962).

Post-Peak Metamorphic Faults (Tn+1)

Post-peak metamorphic faults are characterized by a well developed and closely spaced foliation that makes the rock highly friable. The pervasiveness of the Tn+1 fault fabric lead to the use of the term "paper schist" to describe Tn+1 fault zones. Three of the occurrences are described in detail below.

108 (T-12)

Sample 108 was collected from a roadcut on the west side of Kew Road, located on the west side of the Granville thrust sheet. The roadcut displays an excellent paper-schist fabric. The constituents of the paper schist are mostly quartz (52%) and muscovite (37%); lesser amounts of chlorite (6%) and graphite (5%) are present and only trace amounts of albite, opaques, limonite, and apatite occur. In general the quartz displays a prominent fabric of alternating bands of mylonitized microcrystalline quartz and larger quartz ribbons. The majority of muscovite is fine-grained and parallel to the pervasive foliation, yet some cross micas occur indicating that although metamorphic conditions are past their peak they are still conducive to the growth of new muscovite. No kinematic indicators were observed; however, a distinct parallel arrangement of deformation bands in quartz inclined at 45° to the foliation warrants further study.

712 (H-12)

Sample 712 is conspicuously located at a roadcut on the west side of the Appalachian Gap Road (Route 17) at an elevation of 2220 feet. The Tn+1 fault zone is approximately five meters wide and is characterized by a paper schist fabric that anastomoses around less deformed microlithons of CZf. Location 712 also contains the unique occurrence of fault-related Fn+1 folds in Domain 3. Minor, intrafolial folds with an east over west sense of asymmetry deform the paper schist, and are in turn transposed by it.

In thin section the rock has alternating bands of deformed and undeformed country rock. The rock consists of quartz (50%), muscovite (23%), albite (16%), chlorite (7%), and biotite (2%), with trace amounts of graphite, apatite, opaques, limonite, tourmaline, and epidote. Quartz displays grain size reduction, recrystallized grains distinctly flattened normal to the foliation, subgrains around larger grains, undulose extinction, and deformation bands. The grain size reduction is a result of dynamic recrystallization of original grains that deformed by dislocation flow (Etheridge, 1982). The flattened recrystallized grains indicate recrystallization at low temperatures (Tullis and Schmid, 1982). The subgrains around larger grains indicate crystal plasticity (Etheridge, 1982). From this evidence it is clear that quartz has behaved in a ductile fashion in the Tn+1 fault zone. Albite in thin section exists as small porphyroclasts. The porphyroclasts have minor amounts of recrystallization along their boundaries. Microcracks and deformation twins are also common in the porphyroclasts. The albite has, therefore, behaved in both a ductile and brittle manner. From the microcrystalline properties of quartz and albite it is concluded that the paper schist fabric developed in a transitional brittle to ductile regime, but perhaps slightly more to the ductile side.

Kinematic indicators at 712 include mesoscopic and microscopic east-over-west intrafolial folds, and mica fish with east-over-west senses of shear.

186 (N-33)

Sample 186 occurs within the silvery green schist of the Pinney Hollow near the CZal schist approximately 10 meters from a highly friable paper schist. Sample 186GW contains several garnet porphyroclasts that have been completely to near completely replaced by chlorite and magnetite. The pseudomorphed porphyroclasts are σ_a-type equidimensional porphyroclasts according to Passchier and Simpson (1986, figure 2b&c). σ-type porphyroclasts indicate that the rate of crystallization was greater than the rate of shear during the formation of the porphyroclast. The sense of shear can be determined from the position of the deflected schistosity as it wraps around the porphyroclast--the deflected schistosity is on the trailing side of the porphyroclast.

THE GREEN MOUNTAIN ANTICLINORIUM

All mesoscopic Fn+1 folds appear to be parasitic to the macroscopic folds that define the Green Mountain anticlinorium (GMA). All Fn+1 folds in Domain 2 are on the east flank of the eastern-most GMA antiform and have an east-over-west or, due to their southward plunge, clockwise rotation sense. West-over-east or counterclockwise rotation sense folds are rare in Domain 2, occurring only as parasitic folds on larger east-over-west folds. West-over-east Fn+1 folds are quite common, however, at the axial region of the GMA. In general the rotation senses of Fn+1 in Domain 3 agree with the dip direction of the dominant schistosity. Schistosity that dips to the east has east-over-west parasitic Fn+1 folds, and vice versa for west dipping Sn. These relationships are most apparent in the southern part of Domain 3 where several macroscopic antiforms and synforms occur.

The GMA is the major regional structure in the area where it occurs from the center of Domain 2 westward. The GMA is characterized by the rotation of the dominant schistosity at antiformal crests and synformal troughs to a near-horizontal position that coincides with the physiographic crest of the Green Mountains. The along-axis culmination and bifurcation of fold hinges at map scale is identical to smaller versions at outcrop scale, and this noncylindricity is a characteristic of passive folds in a metamorphic environment (Donath and Parker, 1964).

Doolan et al. (1987) and Mock (1989a,b) refer to GMA-related folds as F3 folds on the west flank of the GMA in northern Vermont. In the Winooski River area Thompson and Thompson (1987) use the same F3 terminology to describe GMA folds. To the south O'Loughlin (1986), Lapp (1986), Haydock (1988), and Prewitt (1989) use the Fn+1 terminology. To the west Tauvers (1982a,b) and DiPietro (1983a,b) refer to the folds on the western flank of the GMA as F2 folds.

The majority of the GMA folds are relatively upright or west verging structures with the exception of a westward-overturned antiform in Domain 4. The overturned nature of the antiform is recognized by a change in asymmetry of parasitic Fn+1 folds observed on the west flank of Stark Mountain, and also by the abundance of Fn+1 folds with sub-horizontal axial surfaces in the hinge region. In the regions where Fn+1 folds have sub-horizontal orientations Fn+2 folds are quite prevalent.

The correlation of Fn+2 folds is limited to the south, for workers to the north do not report broad, open F4 folds that deform F3. O'Loughlin (1986) and Lapp (1986) both report Fn+2 folds and note how they gently warp Fn+1 axial surfaces, and in places deform Fn+1 axial surfaces to horizontal orientations. Tauvers (1982a,b) and DiPietro (1983a,b) refer to such folds as F3, and point out their progressive scarcity to the west away from the GMA.

Faults are also an integral part of the GMA. Most recognizable are the east dipping Tn+1 faults on the west side of the GMA. The faults bound major folds of the GMA on the west side, yet the folds in the center of the GMA are unfaulted. From this evidence, as well as from the geometry of the folds and faults, it appears that the GMA is a fault-related structure--perhaps related to duplexing of basement at depth. The geometry of structures presented in cross-section A-A' looks remarkably similar to that of a fault-propagated fold with a synformal breakthrough. The breakthrough could be the Tn+1 fault at station 712. There are numerous other folds and faults, however, and it is more than likely that the GMA is a complicated fault-related structure not easily represented by a single, simple fault-propagated fold model.

The east flank of the GMA is characterized by east dipping Sn with east-over-west parasitic folds. The parasitic (Fn+1) folds, however, dissipate towards the east until regional Fn+1 folds are no longer observed. Interesting questions are whether or not such Fn+1 folds ever developed in the east (i.e. Domain 1), and how far to the east were the effects of the formation of the GMA felt? Perhaps a clue to these questions is present in the Ottauquechee belt.

Unique quartz extension fractures (or gash fractures) occur in sample 247 (Y-12) from the Ottauquechee black phyllite. The gash fractures are the oldest structures present in the rock as they are cross-cut and mylonitized during the development of the dominant schistosity. The gash fractures record a period of down to the east motion that predates Sn in the Ottauquechee belt. The down to the east motion recorded by the fractures is precisely what one would expect on the limbs of folds produced by passive slip or shear folding. Far to the east, therefore, away from the crest of the GMA, down to the east extension fractures may record deformation in the region of the GMA. If the gash fractures do not have any regional significance they could be the the result of a local stress field.

MESOZOIC FEATURES

A post-Paleozoic period of deformation is recorded in the Fayston - Buels Gore area by the presence of dikes, fractures, and faults.

Twenty diabasic dikes and one albite epidosite dike were recognized in the area. The dikes are parallel to a regional conjugate fracture trend with the majority of the dikes having a slightly north of east trend. Eleven of the dikes occur in the vicinity of Appalachian Gap. A pronounced lineament is evident in the gap area on topographic maps and aerial photographs. The lineament trends N65°W. Private homeowners living on Route 17 near the top of the gap (elevation 2377 feet) have drilled bedrock water wells with high yields. A normal fault with pseudotachylyte slickenlines occurs at a Route 17 roadcut just west of the pond in Chamberlain Glen (H-12). Given the Cretaceous ages of the dikes, it is apparent that Appalachian Gap is a Mesozoic fracture zone.

METAMORPHISM

The rocks in the Fayston - Buels Gore area have a polymetamorphic history that ranges from kyanite and garnet grade of medium to medium-high pressure facies series metamorphism to chlorite grade of medium pressure facies series metamorphism. Metamorphic events associated with the Taconic and Acadian orogenies are well documented in Vermont (Laird et al., 1984; Sutter et al., 1985), however, considerable debate still exists over the timing and the structural and metamorphic overlap of the two orogenic episodes.

ASSEMBLAGES

Twelve metamorphic assemblages are present in the rocks of the Fayston - Buels Gore area. The twelve assemblages are as follows:

-Pelitic and Psammitic Rocks-

With albite:	Without Albite:
1. Qz-Mu-Ab-Ch	7. Qz-Mu-Pa-Ch-Cd-Gt
2. Qz-Mu-Ab-Bi	8. Qz-Mu-Pa-Ch-Cd-Ky-Gt
3. Qz-Mu-Ab-Ch-Bi	9. Qz-Mu-Pa-Ch-Ky-Gt
4. Qz-Mu-Ab-Ch-Gt	
5. Qz-Mu-Ab-Ch-Bi-Gt	
6. Qz-Mu-Ab-Ch-Cd-Gt	

-Mafic Rocks-

10. Ep-Ab-Ch-Ca-Qz
11. Ho-Ab-Bi-Ch-Ep-Ca-Qz
12. Ep-Ac-Ab-Ch-Ca-Qz

Abbreviations: Ab-albite, Ac-actinolite, Bi-biotite, Ca-calcite, Cd-chloritoid, Ch-chlorite, Ep-epidote & clinozoisite, Gt-garnet, Gr-graphite, Ho-hornblende, Ky-kyanite, Mu-muscovite, Pa-paragonite, Qz-quartz, and Tm-tourmaline.

From the above list it is evident that there are three groups of assemblages. The pelitic and psammitic schists, metawackes, granofels, and gneisses have assemblages 1-6, the aluminous pelitic schists have assemblages 7-9, and the metaigneous greenstones and amphibolites have assemblages 10-12. A fourth group, and thirteenth assemblage is the serpentine-talc-magnesite assemblage present in the ultramafic rocks. The ultramafic rocks were not studied in detail.

BULK COMPOSITIONS

The metamorphic assemblages present in the rocks of the Fayston - Buels Gore area can be schematically represented by ternary and tetrahedral bulk composition diagrams (figures 5 and 6). The diagrams are schematic because, aside from chemical analyses of albite, muscovite, and paragonite discussed later, detailed chemistry of the minerals is not available.

Figure 5 illustrates the bulk compositions of the pelitic and psammitic rocks. Figure 5 is modified after Albee (1965) who indicates that the assemblages Qz-Mu-Ab-Ch-Bi-Gt(5), Qz-Mu-Pa-Ch-Cd-Gt (7), and Qz-Mu-Pa-Ch-Cd-Ky are equilibrium assemblages for medium to medium-high pressure facies series metamorphism in the Mount Grant area. The different assemblages are in equilibrium due to the different bulk compositions of the original rocks. The location of all the assemblages from the Fayston - Buels Gore area onto the diagram is not metamorphically correct because some of the assemblages represent conditions at grades other than garnet-kyanite. Schematically, assemblage 1 plots in the chlorite zone as quartz, muscovite, and albite are additional phases in this field. Assemblage 1 is one of the more common assemblages in the eastern part of the research area where retrograde metamorphism is most substantial. It is likely, therefore, that assemblage 1 is a retrograde product of any of assemblages 2-5, and that placement of assemblage 1 in the chlorite field of figure 5 is not a true representation of the bulk composition because it ignores additional Fe phases such as magnetite, and because it represents chlorite grade metamorphism. Assemblage 2 plots on the biotite line to the right of the Gt-Bi-Ch three phase field. Assemblage 2 occurs only in the Fayston biotite gneiss and because of its association with the other high grade assemblages in Domain 3, perhaps it represents another equilibrium assemblage. Assemblage 3 plots on a two phase tie line between the chlorite and biotite fields to the right of the Gt-Bi-Ch three phase field, and could well be a retrograde product of assemblage 2 or 5. Assemblage 4 plots on a two phase tie line between chlorite and garnet to the left of assemblage 1. Assemblage 5 plots in the three phase field of Gt-Bi-Ch. Assemblage 6 is anomalous and occurs only in the transitional Pinney Hollow - Mount Abraham schist. Schematically, assemblage 6 must plot above assemblage 4 to contain chloritoid, but at the same time it must be close to the garnet-chlorite join, on and below which albite is an additional phase. The presence of paragonite coexisting with albite in assemblage 6 is a possibility, but it has yet to be determined. Due to stratigraphic evidence it is likely that assemblage 6 represents a transitional bulk composition between the more aluminous rocks of the Mount Abraham and the albitic schists of the Pinney Hollow. Assemblage 6 is also likely to occur in transitional schists between CZa3 and CZf although it has not been recognized in this area. Albee (1965) did not report this assemblage, but Lapp (1986), O'Loughlin (1986), and Prewitt (1989) have recognized this assemblage between the Mount Abraham and the white albitic schist of the Hazens Notch.
Assemblage 6 could conceivably be in equilibrium with the higher grade assemblages. Above the garnet-chlorite join albite and especially biotite are characteristically absent, and paragonite becomes the Na-bearing phase. Assemblage 7 is one of the most common Mount Abraham assemblages and plots within the Gt-Cd-Ch three phase field. Assemblages 8 and 9 are anomalous due to the coexistence of kyanite and garnet which leads to the crossing of the chloritoid-chlorite and kyanite-garnet joins. As pointed previously, garnet that coexists with kyanite in CZa4 appears to be metastable. Albee (1965) reports the occurrence of assemblage 9, but not assemblage 8 with chloritoid. Albee explains that the assemblage Ky-Gt-Cd is not compatible with the diagram of figure 5. Chemical analyses of garnets in this assemblage (see La-167 of Albee, 1965) indicates that the garnet is 51% spessartine and only 33% almandine whereas the garnet in the other assemblages is 55-77% almandine, 6-20% spessartine, 13-22% grossular-andradite, and 4-8% pyrope. Thus the garnet contains two to eight times more MnO than garnet of other assemblages. The chlorite analyzed by Albee has a composition that plots in the kyanite-chlorite two phase field. Albee (1965, p. 271) points out that assemblage 9, "is actually Ky-Ch-Qz-Mu with an additional phase, garnet, introduced by the high Mn content of the rock." Assemblage 9, therefore, plots in the kyanite-chlorite two phase field of figure 5, and similarly assemblage 8 should plot in the kyanite-chloritoid-chlorite three phase field with garnet as an additional phase.

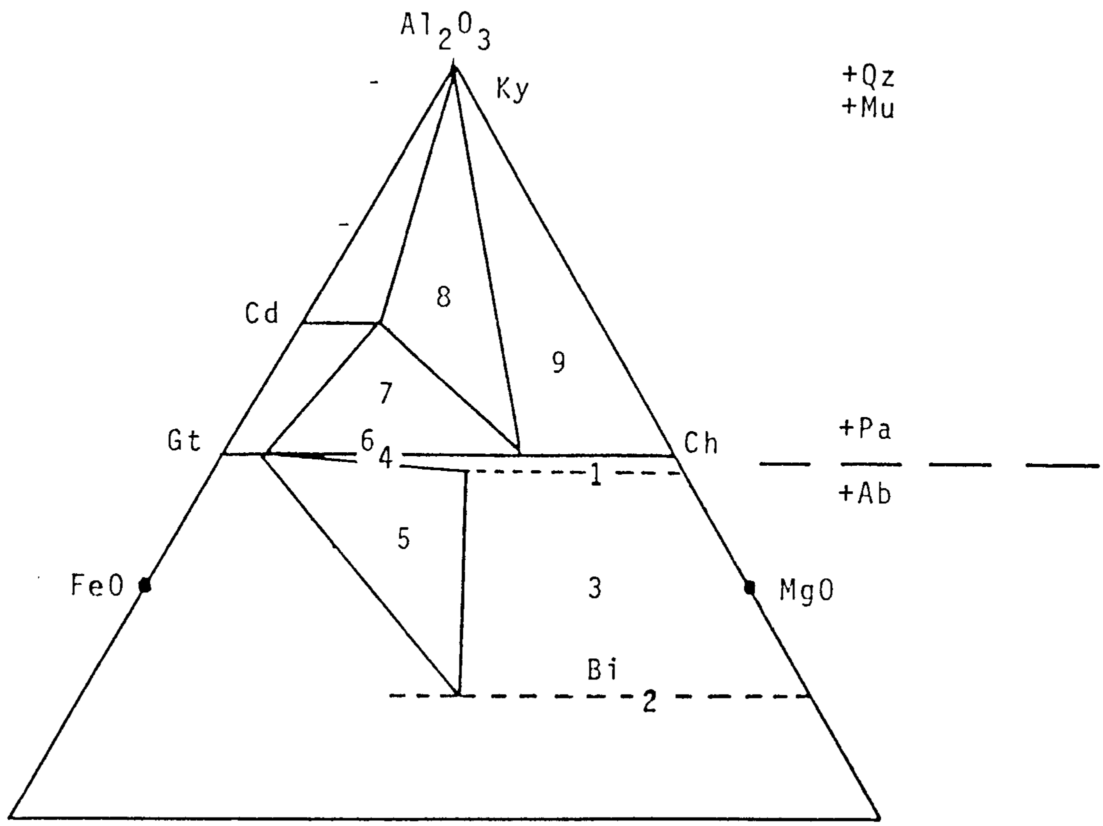

Figure 5. Thompson A-F-M projection from muscovite onto the Al_2O_3-FeO-MgO plane showing schematic bulk compositions for pelitic and psammitic units. Numbers correspond to assemblages described in text. Quartz and muscovite are additional phases throughout. Paragonite is an additional phase above the Gt-Ch join. Albite is an additional phase below the Gt-Ch join. Quartz and water are saturating phases. Abbreviations for phases are indicated in text. Diagram is modified after Albee (1965; fig. 5, p. 271).

The three assemblages present in the mafic rocks are illustrated on the A-C-Fm-K tetrahedron in figure 6. Assemblage 10 plots in the epidote-calcite-chlorite three phase field on the A-C-Fmface of the tetrahedron. Assemblage 11 plots in the epidote-hornblende-biotite (dotted) three phase field that is internal to the tetrahedron. Assemblage 12 plots on the epidote-actinolite-chlorite three phase field on the A-C-Fmface. In both assemblages 11 and 12 calcite is an additional phase possibly introduced by an increase in P-CO_2 causing a Ca-amphibole first, and then epidote to break down to a carbonate and chlorite (Miyashiro, 1973). Perhaps an increase in P-CO_2 took place during retrograde metamorphism, if this is the case then assemblage 10 represents a possible retrograde product of assemblages 11 and 12. Epidote and hornblende give evidence for prograde metamorphic reactions in the metabasic rocks of the Underhill. The presence of zoned epidote with pistacite cores and clinozoisite rims corresponds to a decrease in Fe content towards the rim which is indicative of progressive metamorphism (Miyashiro, 1973). From the biotite to garnet zone amphibole changes color from green to blue-green (Miyashiro, 1973; Laird and Albee, 1981). The blue-green hornblende in the Underhill amphibolite has ferrohornblende cores and ferrotschermakitic hornblende and ferrotschermakite rims (see V225B in Laird et al., 1984).

CHRONOLOGY

In order to discuss the relative timing of the growth of various minerals it is important to define the frame of reference. Recent workers in the pre-Silurian sequence (Lapp, 1986; O'Loughlin, 1986; Haydock, 1988; and Prewitt, 1989) use the standard n-1, n, and n+1 terminology to compare various stages of mineral growth, and conclude that each generation of "n" was correlative across the belt. As will be pointed out, Sn no longer appears to be the same Sn everywhere. For this reason, Sn in this section is referred to as either a western Sn (in Domains 3 & 4) or an eastern Sn (in Domain 1).

The minerals discussed in the previous section are the most significant to the discussion of mineral growth chronology. Quartz recrystallizes and grows during every observed deformation episode with the possible exception of Tn+1 "papery" schist fault zones where only grain-size reduction is observed. Quartz, therefore, is not a very useful mineral for unraveling the metamorphic and deformational history.

The history of albite is similar to that of quartz. Detrital plagioclase exists in the western half of the research area, but is not observed in the east. Albite porphyroblasts grew at many different times. In places throughout all three domains the porphyroblasts contain n-1 inclusion trails. In Domain 3 albite growth is synchronous with Sn and often post-dates n+1 as evidenced by Fn+1 inclusion trails in porphyroblasts. Albite porphyroblasts also overgrow the eastern Sn. The porphyroblasts did not grow within Tn+1 paper schist fault zones, but did develop in some Tn fault zones (see 413, p. 46). Because albite porphyroblasts grew at conditions ranging from garnet to chlorite grade, as well as during the development of the many structural fabrics they are not characteristic of any specific conditions or deformational event. The optically zoned plagioclase porphyroblasts (probably albite cores and oligoclase rims) occur only in the western part of the research area and record prograde conditions during the development of the western Sn.

White mica (predominantly muscovite) and chlorite span the entire range of mineral growth episodes. The oldest white mica and chlorite are parallel to the n-1 schistosity. The youngest mica and chlorite occur as cross mica and chlorite in Domains 1, 3, and 4. Cross micas in Domains 3 and 4 appear to be synchronous with the n+1 or even n+2 deformation. The youngest cross muscovite in Domain 1 occurs in the gray carbonate schist of the Ottauquechee (Cog) where it post-dates the eastern Sn.

Biotite occurs both syn and post western Sn in Domains 3 and 4. Cross biotites are of the same age (n+1 to n+2) as cross muscovites. Biotite generally predates the eastern Sn in Domain 1 and in the eastern half of Domain 2. Biotite is often pseudomorphed and

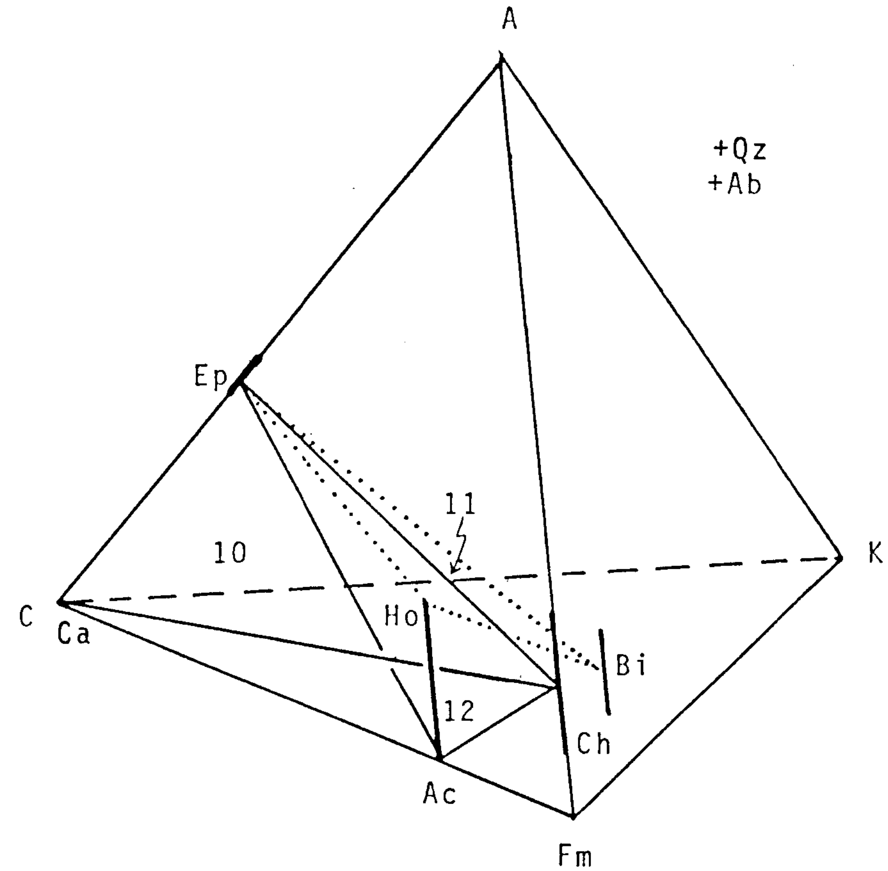

Figure 6. A-C-Fm-K tetrahedron showing schematic bulk compositions for mafic units. Numbers correspond to assemblages described in text. Quartz is an additional phase throughout. Quartz and water are saturating phases. Abbreviations for phases are indicated in text. Diagram is modified after Hyndman (1985; fig. 12.6 , p. 500).

retrograded during the development of the eastern Sn. Biotite is preserved only in the more competent wackes in Domain 1, yet it is quite abundant in many lithologies in Domains 3 and 4.

Chloritoid grew during the development of the western Sn, regional Fn+1 folds, and up to, but not including, the Sn+1 crenulation cleavage. In Domain 3 large porphyroblasts of chloritoid are present in the Mount Abraham. In the east, however, chloritoid pre-dates the eastern Sn and is often very small and retrograded.

Garnet is ubiquitous in Domain 3 and grew during the development of the western Sn and up to, but not including the development of the regional Fn+1 folds. In contrast garnet in Domain 1 is a pre-eastern Sn mineral. Garnet in Domain 1 that is not in albite as inclusions is always partially or completely retrograded and replaced by chlorite, and to a lesser extent magnetite. The distribution of the unaltered prograde garnets and altered retrograde garnets is clearly visible on the geologic map (Plate 1) where the symbols Gtp and Gtr are used for prograde and retrograde garnet, respectively. Retrograded garnets are observed in Domain 4, but their occurrence can be attributed to their proximity to later faults, a point that was first noted by Lapp (1986) and O'Loughlin (1986). In Domain 2 retrograded garnet is observed where the Sn+1 crenulation cleavage becomes quite pronounced and develops into a penetrative schistosity (figure 7). Anderson (1987) recognizes that garnet is retrograded during the development of the "Sc" schistosity--a direct correlative with Sn+1.

Kyanite porphyroblasts are restricted to Domain 3. The growth of kyanite is synchronous to the development of the western Sn, and in a few places it is found to overgrow Fn+1 folds (figure 8). Albee (1965, p. 253) reports that planar inclusion trails in kyanite and albite indicate that they, "grew prior to the development of the present crinkled texture." Albee did not recognize the evidence for Fn+1 albite and kyanite, and the presence of Fn+1 kyanite is very significant due to the limited conditions at which it grew. The growth of kyanite during Fn+1 indicates that the peak of metamorphism extended through the development of the western Sn and *continued into* the development of the regional Fn+1 folds. Considering that the peak metamorphic minerals in the area (garnet and kyanite) grew during the development of the western schistosity and the folding of that schistosity, and that garnet is substantially retrograded during the development of both Sn+1 and the eastern Sn it becomes apparent that the western Sn is a prograde schistosity whereas the eastern Sn is a retrograde schistosity.

Epidote, including clinozoisite, is commonly a euhedral mineral that grew during the development of the western Sn in Domains 3 and 4. In Domain 1, syn-eastern Sn epidote is most often fine grained and granular, and occasionally euhedral.

Actinolite is not recognized in Domain 4, but occurs parallel to the eastern Sn (Sn+1) in Domains 2 and 3. Hornblende is present only in Domain 4 and grew during the development of the western Sn.

Carbonates occur commonly as anhedral interstitial masses. Euhedral rhombs of calcite and magnesite postdate the eastern Sn in Domain 1 in the gray carbonate schist and the talc magnesite schist, respectively.

PLAGIOCLASE

Plagioclase, primarily albite, is one of the most conspicuous and abundant minerals in the Fayston - Buels Gore area. The origin of the abundant albite in the area was discussed in the lithologies section. In this section the metamorphic aspects of albite are discussed.

Twinning

The majority of albite is untwinned, a phenomenon which is in agreement with observations of metamorphic plagioclase by Turner (1951). Growth and deformation twins are the two main groups of twins observed in metamorphic albite from the area. Tullis (1983) describes evidence for growth versus mechanical (or deformation) twins as follows:

Growth Twins	Mechanical Twins
A. broader and fewer twins per grain	A. very fine scale
B. straight and parallel sided	B. tapered and/or curved
C. cut across entire grain	C. restricted to small areas of a grain
D. no relation to bending or fracturing	D. more numerous at grain boundaries and internal boundaries

Generally, two types of growth twins occur in albite from the research area. In some porphyroclasts from such coarse grained rocks as metawackes the grains contain numerous lamellar twins with many subindividuals. In some porphyroblasts from the schists the twinned grains contain only a single polysynthetic twin that separates two subindividuals (figure 9). In detailed studies of twinning in metamorphic plagioclase Turner (1951) and Smith (1974) report that lamellar twins from sodic plagioclase (An_0-An_7) are predominantly of the albite law with some cases of twins on the pericline law. They both point out that the lamellar twins are exceptions to the rule in greenschist facies albite where the albite is usually untwinned, but may occasionally have a simple albite twin. Turner (1951, p. 584) points out that although it is difficult to distinguish a Carlsbad from an albite twin it is possible even without the use of a universal stage. In the case of an albite twin,

"...the twin axis of suitably oriented albite porphyroblasts may be identified microscopically as perpendicular to {010} from the fact that the transverse {001} cleavage continues across both subindividuals with but slight deflection where it crosses the composition plane."

In a Carlsbad twin, however, if the {001} cleavage is visible in one half of the grain it is invisible in the adjacent half. In the Fayston - Buels Gore area all the twinned albite porphyroblasts obey the albite twin law and not the Carlbad law. Carlsbad twins are virtually absent from greenschist facies albite porphyroblasts (Smith, 1974). According to Smith such simple albite twins in albite porphyroblasts are the result of nucleation error during growth in a solid medium.

Lamellar growth twins in albite occur in the following lithologies: Underhill schist, metawacke, and amphibolite, the Fayston biotite gneiss (CZfb), and the virtually identical metawackes from the Pinney Hollow and Mount Abraham (CZphw & CZaw). All of these lithologies are coarser grained, and have a less well-developed foliation than the majority of schists in the area.

The occurrence of growth twins in the porphyroclastic albite is an uncommon phenomena in metamorphic rocks. According to the evidence presented by Tullis the twins clearly appear to be growth twins and not deformation twins. Figure 10 shows an albite porphyroclast with deformation twins for comparison to the growth twins of figure 9. Smith (1974) states that growth twinning may occur by epitaxial inheritance--a process by which pre-existing material, either detrital or igneous in origin, acts as a nucleus for crystal growth. Coombs (1954) describes coarse grained sediments with authigenic overgrowths of albite around detrital plagioclase in which there is complete optical continuity of the twin pattern from the detrital grain to the overgrowth. Turner (1951) describes amphibolites of igneous origin that contain coarse plagioclase with relict idiomorphic outlines and highly complex twinning usually confined to igneous rocks. The plagioclase contains twin combinations on the albite-Carlsbad-pericline laws. Turner suggests that the plagioclase twins are good evidence of the igneous parentage of the amphibolite. Smith (1974) cites Turner's example and points to the role of epitaxial inheritance. Coincidentally, similar plagioclase occurs in the Underhill amphibolite. Figure 11 shows an albite from CZug with simple albite and multiple pericline twinning--twinning that is characteristic of igneous and not metamorphic rocks (Smith, 1974).

To further the point, all the observed examples of porphyroclasts with relict twins come from the coarse grained schist, gneiss, amphibolite, and metawacke. The metawackes even contain pebbles. Presumably the coarsest lithologies with the least well-developed foliation are the least recrystallized. If any rocks in the area were to preserve the relict detrital textures, these would be the ones. In the less competent units the relict textures would not be preserved due to the greater degree of recrystallization. In fact there is evidence that the albites with relict textures were themselves nucleation sites for the formation of some albite porphyroblasts (figure 12).

The significance of relict growth twinning in porphyroclasts is highlighted if one recalls the discussion on the origin of the abundant albite in the pre-Silurian section. It was shown that the breakdown of calcic plagioclase can lead to the formation of albite porphyroblasts with accessory Ca phase minerals (i.e. epidote). The angular albite porphyroclasts with growth twins appear to be relict detrital grains. Thus, it is conceivable that the abundance of albite in the rocks of the Fayston - Buels Gore area is a direct function of the amount of plagioclase (detrital or igneous) originally present in the protoliths.

Chemistry

It is necessary to determine the composition of the plagioclase upon identification of the relict textures in order to determine if the grains are detrital or metamorphic. Although detrital plagioclase is recognized in the Taconics it is unlikely that detrital grains with their original chemistry would be preserved in the central Vermont rocks during metamorphism at middle and upper greenschist facies conditions. Bishop (1972) reports that detrital plagioclase in metawackes from South Island, New Zealand retains its original composition only up to the initial stages of prehnite-pumpellyite facies metamorphism, and that the original chemical signature is long gone by the time greenschist facies conditions set in.

Chemical analyses of 29 points from 14 individual porphyroclasts with relict lamellar twins were obtained on the ARL-EMX electron microprobe with KRISEL automation at Middlebury College, Middlebury, Vermont (see Walsh, 1989 for details). The data indicate that the plagioclase is albite with an anorthite content that ranges from An_0 to $An_{1.3}$. The chemical analyses of eight porphyroclasts from the Fayston biotite gneiss (CZfb) and six from the Pinney Hollow metawacke (CZphw) indicates a virtually pure albite composition. From this evidence it is appears that the porphyroclasts are metamorphic and that their shape and twinning is a function of the original detrital grains. The original composition of the grains has changed to a composition of albite which is the stable plagioclase phase in greenschist conditions. The relict textures remain due to epitaxial inheritence.

The chemistry of albite porphyroblasts has not been determined in this project, but has been done by previous workers. Laird and Albee (1981) report that plagioclase from the mafic schists of the Underhill Formation are in biotite and garnet grade assemblages and have compositions that range from $An_{0.1}$ to $An_{1.2}$. Haydock (1988) analyzed four porphyroblasts: One sample from each of the silvery green schist of the Pinney Hollow and carbonaceous albitic schist of the Hazens Notch (Granville in this report), and two samples from the Hazens Notch (Fayston) white albitic schist. Haydock's results indicate no chemical zonation in the four porphyroblasts and only slight variation in anorthite content from An_1 to An_4. The porphyroblasts, however, were not optically zoned. Theoretically, given that the rocks in the pre-Silurian sequence were metamorphosed at or above garnet grade, the albite porphyroblasts should record chemical zonation due to the formation of oligoclase as the peristerite gap is crossed (Crawford, 1966). The occurrence of optically zoned porphyroblasts--the first indication of chemical zonation--has not been reported by earlier workers. Optically zoned porphyroblasts do occur in the Fayston - Buels Gore area (figure 13).

Crawford (1966) reports that plagioclase porphyroblasts from the Gile Mountain Formation have albite cores that are rimmed by oligoclase as a result of peristerite unmixing during progressive metamorphism through the biotite and into the garnet zone. The porphyroblasts have a readily discernible optical discontinuity at the oligoclase-albite boundary. More recently, Anderson (1987) describes optical and compositional zoning in a porphyroblast from the pre-Silurian sequence in the vicinity of the Winooski River transect. The inner core of the porphyroblast ranges from $An_{0.2}$ to $An_{1.1}$ and is surrounded by an outer core of $An_{1.5}$ to $An_{3.9}$. An optical discontinuity occurs at the boundary between the outer core and the rim. The rim has an anorthite content that ranges from $An_{21.5}$ to $An_{30.7}$. The optical zoning is concentric, and parallel to the chemical zoning. Anderson attributes the zoning to the progressive reaction of epidote with plagioclase to produce more Ca-rich plagioclase during increasing metamorphic grade, and states that zoning represents the peristerite gap. Anderson explains that the zoned plagioclase grew during the development of an "Sb" schistosity. Anderson's Sb is directly correlative to the western Sn in the study area. Locations of observed optical zoning in plagioclase porphyroblasts from the area are indicated on the geologic map with the symbol "Abz." All observations of zoned plagioclase are restricted to the western half of the area. In the eastern half of the area the plagioclase might have developed compositional and optical zoning. If the zoning ever developed, it would presumably be recrystallized to pure albite during retrograde metamorphism. This would coincide with the observation that only chlorite pseudomorphs after garnet are observed in the eastern half of the area, and as retrograde metamorphism was only chlorite grade the oligoclase would no longer be stable.

WHITE MICAS

The coexistence of muscovite and paragonite in the Mount Abraham Schist is reported by Cady et al. (1962), Albee (1965), and Albee and Chodos (1965). Walsh (1989) contains a detailed discussion on the structure, chemistry, occurrence, X-ray analyses, microprobe analyses, and geothermometry aspects of the white micas. The results are summarized below.

Muscovite and paragonite coexist in the Mount Abraham white mica schists as evidenced by use of the powder method of X-ray diffraction on three samples from the Fayston - Buels Gore area. A single sample from the Pinney Hollow silvery green schist was analyzed as well, but did not contain paragonite as albite is the Na-phase in the schist.

Microprobe results of 14 separate analyses indicate end-member compositions of coexisting paragonite and muscovite as $Pa_{87}Mu_{13}$ and $Pa_{12}Mu_{88}$, respectively. Mg, Fe, Mn, Cr, and Ti contents are consistently higher in the muscovite analyses and correspond to a decrease in the Al content indicating that these ions are replacement elements for Al in the octahedral site. The presence of such elements, especially Mg and Fe, in the muscovite indicates a definite phengitic component. The Si:Al ratio in the tetrahedral site is between 4.12:1 and 3.87:1, a ratio significantly higher than 3:1, again indicating a phengitic component. Because of the phengitic component in the white micas they can not be used as a geothermometer.

REGIONAL CONDITIONS

Estimates of pressure and temperature of metamorphism for the pre-Silurian rocks of Vermont have been presented by many authors. Laird (1987) indicates that oxygen isotope analyses, originally conducted by Garlick and Epstein (1967), from biotite grade pelitic schist north of the Lincoln massif and kyanite grade pelitic schist from Mount Grant give temperatures of 435° and 420° to 465° C, respectively. Calcite-dolomite and amphibole-plagioclase geothermometry from garnet grade mafic schist yields temperatures of 450° to

489°C. Amphibole-plagioclase geobarometry from the same assemblage gives a pressure of 7 to >8 Kbar (Laird, 1987). Albee (1968) suggests that the high grade rocks of the Worcester-Northfield Mountain anticlinorium (WNMA) record slightly higher metamorphic conditions than the high grade rocks of the GMA. Although trace amounts of staurolite are present at both high grade areas, the major assemblages are garnet-chloritoid-chlorite and kyanite-chloritoid-chlorite in the GMA, and kyanite-garnet-biotite-muscovite in the WNMA. Rocks in the WNMA, however, have seen a greater degree of retrogression. Laird (1987) supports Albee's claim by showing that amphibole in mafic schist from both areas records a higher pressure in the WNMA than in the GMA. Amphibole from the Pinney Hollow greenstone at Granville Gulf has barroisitic cores with actinolite rims (V12, 14 and 15 of Laird and Albee, 1981). This contrasts with ferrohornblende cored and ferrotschermakitic hornblende and ferrotschermakite rimmed amphibole from the Underhill amphibolite at Appalachian Gap (V225 of Laird et al., 1984).

AGE OF METAMORPHISM

Taconian

In regards to the amphiboles mentioned above, Laird et al. (1984) used $^{40}Ar/^{39}Ar$ total fusion techniques and obtained an age of 471 ± 10.6 Ma and 471 ± 5.8 Ma for the Pinney Hollow and Underhill samples, respectively. This implies that the higher metamorphic conditions east of the GMA occurred simultaneously with the lower conditions to the west.

Lanphere and Albee (1974) report a 444 Ma age on coarse grained muscovite from kyanite grade schist of the Stowe in the Worcester Mountains. Sutter et al. (1985) report a $^{40}Ar/^{39}Ar$ plateau age of 449 Ma from similar coarse muscovite in the Stowe. These analyses are from essentially the same rocks that Rickard (1965) obtained a K-Ar date of 430 Ma from the coarse muscovite. Albee (1968) reports that the growth of the coarse grained muscovite, kyanite, and garnet from the Worcester Mountains is associated with the formation of a steep crenulation schistosity that is parallel to the axial planes of major folds in the Worcester Mountain anticline. I suggest that these folds may correspond to the Fn+1 folds of the GMA in the research area. Remembering that kyanite and muscovite in the research area grew during western Sn and Fn+1 development it is logical to assume that the age of growth corresponds to ages reported in the Worcester Mountains. From these ages and relationships it is concluded that the Green Mountain and Worcester-Northfield Mountain anticlinoriums are Taconian structures in accordance with Albee (1968), Laird et al. (1984), and Sutter et al. (1985). The presence of blueschists east of the GMA in northern Vermont suggests that the rocks had travelled further down the subduction zone than those to the west and south (Doolan et al., 1982; and Stanley et al., 1984). In central Vermont, the slightly higher metamorphic conditions recorded east of the GMA (Granville Gulf vs. Appalachian Gap) implies that the eastern rocks were subducted to greater depths than the western rocks, but not to the extent in northern Vermont. Another possibility is that the eastern rocks never actually went down the subduction zone, but were incorporated into the accretionary wedge at greater depths than the western rocks.

It is concluded that the prograde assemblages recorded in the research area are associated with the development of the western Sn and Fn+1, and that these deformational events are related to compression during the Taconic orogeny.

Acadian

Prior to Albee's (1968) suggestion that much of the metamorphism in western Vermont was Taconian most workers believed that all of the metamorphism was Acadian. The previous section pointed out that there is now considerable evidence for significant Taconian metamorphism. The extent of Acadian metamorphism in central and northern Vermont still remains a topic of debate.

Lanphere and Albee (1974) obtained a 363 Ma age from fine grained muscovite pseudomorphs after kyanite from the same rocks that yielded the 444 Ma from coarse, pre-retrogressive muscovite, and interpreted the age of retrograde metamorphism as Acadian. Lanphere et al. (1983) report ages of 387 Ma from biotite and 382 Ma from amphibole by $^{40}Ar/^{39}Ar$ techniques for samples collected on the Lincoln Gap Road. Laird et al. (1984) report ages of 376 to 386 Ma from muscovite and biotite in kyanite grade pelitic schists from the Albee (1965) locality at Mount Grant. Laird et al. (1984) interpret all of these ages as Acadian metamorphic ages. Sutter et al. (1985) consider these ages as Taconian cooling ages and point out that most of the Devonian ages are from muscovite and biotite. The 382 Ma age from amphibole at Lincoln Gap is from actinolitic hornblende and the argon-loss systematics of actinolite are not well constrained (Sutter, pers. comm., UVM short course, 1987). Since muscovite and biotite have argon closure temperatures of 320° and 260° C, respectively (Harrison and McDougall 1980, 1981) age estimates from the micas should be avoided for rocks above biotite-garnet grade unless cooling rates are known to be rapid after mineral formation (Sutter et al., 1985).

Cua (1989) reports that the highest grade on the east flank of the Northfield Mountains in the Stowe Formation is recorded by biotite that is retrograded to chlorite. Sutter et al. (1985) place the Acadian biotite isograd east of the area mapped by Cua. This implies that the biotite reported by Cua grew during the Taconian and was retrograded during the Acadian within the Acadian chlorite zone. If this is true, how could static cross biotite and muscovite of post-GMA age possibly be related to an Acadian metamorphic event if they occur 15 kilometers west of the Acadian biotite isograd? The static overgrowths may be a post-peak Taconian event.

DISCUSSION

The most significant aspect of the structural and metamorphic evolution of the rocks in the Fayston - Buels Gore area pertains to the development of the two pervasive schistosities, western and eastern Sn.

The western Sn developed during prograde metamorphism and is coeval with garnet and kyanite-bearing assemblages. The western Sn is deformed by two generations of folds--Fn+1, the predominant GMA folds, and Fn+2. The occurrence of kyanite with Fn+1 folded inclusion trails is significant because it indicates that prograde conditions persisted up to the development of the Fn+1 folds, and hence the GMA. In Domains 3 and 4, the Sn+1 crenulation cleavage is incipient on Fn+1 folds in the finer grained, less competent rock types. Both high grade indicator minerals, garnet and kyanite, are retrograded during the development of the Sn+1 crenulation cleavage. The Sn+1 cleavage is a structural fabric that is not directly related to the intial development of Fn+1 folds in that it develops subsequent to Fn+1 folding and transposes such folds. Sn+1 becomes more pervasive from west to east. The orientation of regional Fn+1 axial surfaces shown on the cross section (plate 2) corresponds to the orientation of Sn+1. As the axial surafces are traced from Domain 3 to Domain 1 they approach parallelism with the observed eastern Sn. Since Sn+1 and the eastern Sn are both associated with retrograde metamorphism they are metamorphically correlative. Considering that Sn+1 becomes more pervasive to the east it implies that Sn+1 in the west becomes the eastern Sn. The progressive change of Sn+1 from a crenulation cleavage to a pervasive schistosity must occur in Domain 2. Regional Fn+1 folds are observed in greenstones and quartzites from Domain 2, but in the surrounding schists the folds have been transposed by the new, composite schistosity. In Domain 1 regional Fn+1 folds are not readily apparent--they have either been transposed and rotated to appear as Fn folds or they did not develop to the same extent as seen in the west, or a combination of both.

Using the above arguments, it is now possible to assign the time-dependant terminology of S0, S1, S2, etc. to the deformation in the Fayston - Buels Gore area. Sn-1 is parallel to compositional layering in the western part of the area and it is the oldest recognized foliation, therefore it is correlative with S0. Sn in Domains 3 and 4 is the pervasive schistosity associated with peak metamorphic conditions--it is correlative with S1. Sn+1 is a crenulation cleavage in the west and a penatrative schistosity in the east, and it is associated with retrograde conditions--Sn+1 is S2. The most significant point to be made is that the dominant schistosity over the GMA is S1, whereas in the Mad River Valley the dominant schistosity is S2. West dipping kink bands in the easternmost part of the area are correlative with S3. No crenulation cleavage is associated with Fn+2 folds in the western part of the area, and at this time it is not possible to correlate the Fn+2 folds and the kink bands into a single S3 event.

CONCLUSIONS

The non-carbonaceous quartz-muscovite-albite-chlorite schists of the Underhill, Fayston, Pinney Hollow, and Stowe represent a fining-eastward clastic facies change. Deposition occurred between Late Proterozoic and Cambrian times. Rocks of the Mount Abraham represent a discontinuous sequence of aluminous pelagic shales atop the non-carbonaceous rocks. The carbonaceous lithologies of the Battell, Granville, and Ottauqechee represent a sequence of black shales that were deposited during a period of anoxic ocean conditions, perhaps as the result of decreased continental glaciation and subsequently poor oceanic ventilation. Deposition occurred during the Cambrian. The breakdown of calcic plagioclase to albite and a Ca-phase in the quartz-feldspar clastic rocks is a probable origin for the abundant albite seen in the rocks today.

The geochemistry of the greenstones and amphibolites from the non-carbonaceous clastic rocks agrees with the west-to-east rift-clastic facies change model. Metabasites from the Underhill plot as within-plate and alkali basalts. The greenstones from the Fayston, Mount Abraham, Pinney Hollow, and Stowe plot as transitional from within-plate to tholeiitic basalts. Stowe greenstones plot as tholeiitic basalts.

At least four generations of foliations occur in the area: A relict schistosity (Sn-1 or S0), a pervasive schistosity (Sn or S1), a crenulation cleavage (Sn+1 or S2), and kink bands (S3). Two types of Fn folds are associated with the dominant schistosity--isoclinal reclined folds and sheath folds. Regional Fn+1 folds associated with the GMA deform the dominant schistosity (western Sn or S1) in the western half of the area, but are transposed in the eastern half of the area by Sn+1 (S2). Open Fn+2 folds deform the regional Fn+1 folds, and are confined to the western part of the research area. At least three generations of faults occur in the area: Pre-peak metamorphic (Tn-1), syn-metamorphic (Tn), and post-peak metamorphic (Tn+1). Four Tn-1 faults are recognized: 1) Fayston over Battell and Monastery; 2) Mount Abraham (CZa2 and CZa4) over Fayston and Granville; 3) Granville over Pinney Hollow and Mount Abraham; and 4) Ottauquechee over Pinney Hollow. Many Tn faults occur in the area, although they are most common in the east. Of nine samples of Tn fault fabrics analyzed petrographically, two contain conclusive east-over-west kinematic indicators. Samples from three Tn+1 fault zones were analyzed petrographically, two contain conclusive east-over-west kinematic indicators. The Green Mountain anticlinorium (GMA) is defined by large Fn+1 folds. Tn+1 (S2) faults cut the anticlinorium and are believed to be related to a later stage in the development of the GMA. Fn+2 folds may be associated with the later phase of development. Fractures, dikes, and normal faults are associated with a minor, post-metamorphic, Mesozoic deformation event. Appalachian Gap is a Mesozoic fracture zone across the Green Mountains.

Prograde garnet and kyanite assemblages are preserved in the western part, and chlorite grade assemblages overprint the high grade assemblages in the eastern part of the area. The prograde assemblages are associated with the western Sn (S1) and the Fn+1 regional folds.

The retrograde assemblages are associated with the eastern Sn (S2) which is correlative with the western Sn+1 (S2) crenulation cleavage. Analyses of porphyroclastic albite with lamellar growth twins from coarse schist and metawacke suggest that the clasts preserve relict detrital textures by epitaxial inheritance from original detrital feldspars. This supports the idea that the albitic rocks originally contained detrital quartz and feldspar. The white micas, muscovite and paragonite, can not be used as a geologic thermometer.

INTERPRETATIONS

DEPOSITION

Late Proterozoic rifting of the Grenvillian basement results in the deposition of rift-clastic rocks and associated rift-related igneous activity. From west to east the rift-clastic rocks consist of an eastward-fining sequence of quartz and feldspar-rich sediments of the Underhill, possible Taconic-type graywackes, Fayston, Pinney Hollow, and Stowe (figure 14). As rifting slows, igneous rocks become rarer upward in the sequence and continued erosion produces chemically mature Fe-Al rich sediments. From west to east the sequence of rocks includes the Monastery, possible Taconic-type turbidites with aluminous shales, Fayston with CZa3 and CZft, Pinney Hollow with CZa1, Mount Abraham CZa2 and CZa4, and aluminous Stowe shales (figure 14).

Between the Middle Cambrian and Middle Ordovician climatic conditions change, and reduced continental glaciation leads to a depletion of the oxygen content in the oceans. This change leads to the deposition of black shales worldwide. From west to east--shelf edge to outer slope and rise--the sequence of rocks includes the Battell, possible Hatch Hill and West Castleton-type correlatives, Granville, Lincoln Gap Member, and Ottauquechee (figure 14).

DEFORMATION AND METAMORPHISM

Pre-metamorphic thrust faults develop at the onset of deformation. The Ottauquechee with fragments of ocean crust is transported over the Pinney Hollow. The Granville is transported over the Pinney Hollow, Mount Abraham (CZa1), and Fayston. The Mount Abraham (CZa2 and CZa4) with Lincoln Gap Member depositionally on top is transported over Fayston and Granville. If Taconic correlatives left the area as allochthons they were transported over the Underhill and deposited on the carbonate platform. The Fayston is transported over the Battell and Monastery. If deformation proceeded from east to west, the above order of description is a probable order of emplacement. A relict Sn-1 (S0) schistosity may be associated with the development of these early faults.

The ensuing development of a regional prograde metamorphic event produces a dominant schistosity with associated garnet and kyanite grade assemblages. The initial Taconian metamorphism is characterized by medium pressure (5-7 Kbar) in the west and medium-high (7-9 kbar) pressure in the east, perhaps as a function of the degree of subduction. The development of the Green Mountain anticlinorium and associated regional Fn+1 folds marks the peak of metamorphic conditions at garnet-kyanite grade. Basement duplexing or ramping off of a basal décollement as continental crust clogs the subduction zone is a possible mechanism of anticlinorium formation.

Anticlinorium folds on the east flank of the GMA are transposed along with the development of a composite eastern Sn, correlative with the Sn+1 crenulation cleavage seen in the west. Retrograde metamorphism at chlorite grade is associated with this period of deformation. The entire Pinney Hollow / Ottauquechee belt is reactivated as a major shear zone. Higher grade assemblages in the GMA and WNMA are preserved where shear zone development is not as intense.

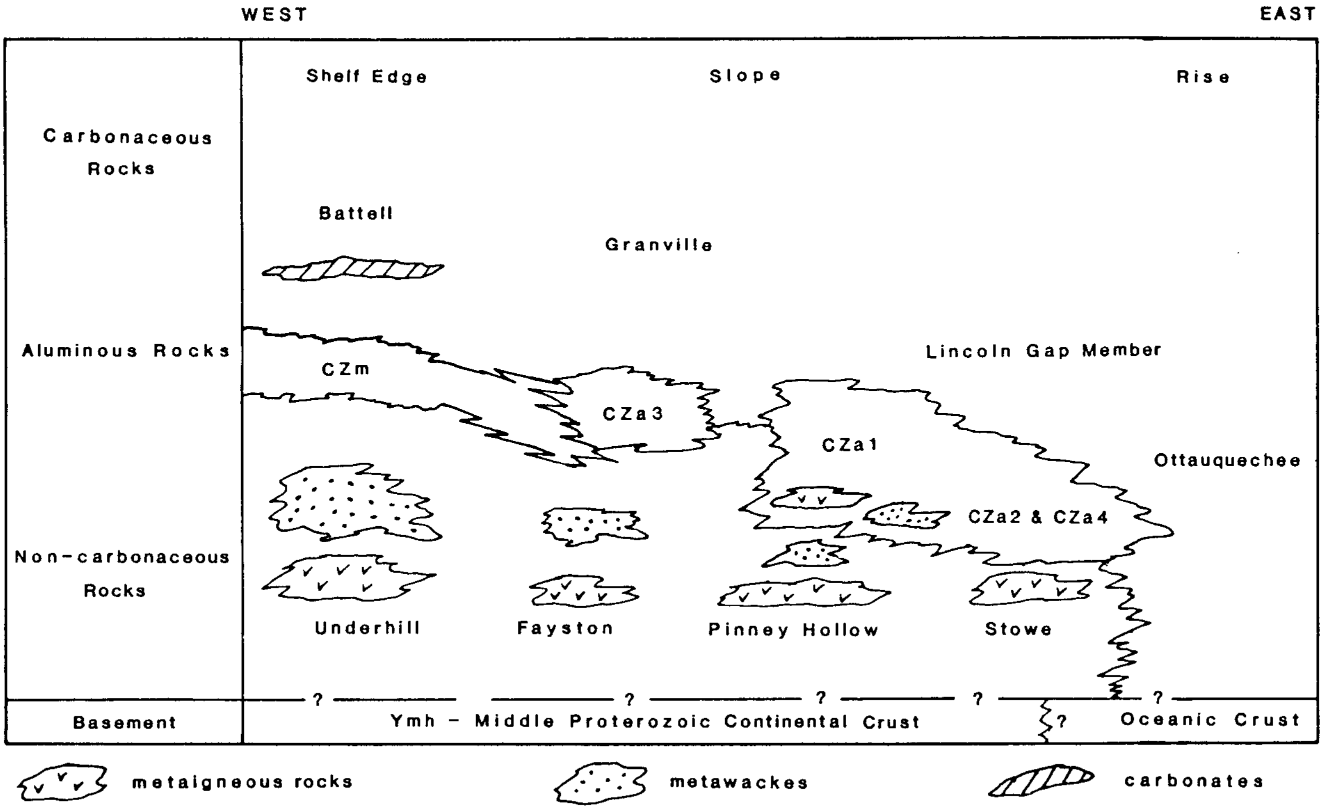

Figure 14. Interpretive lithofacies diagram. Diagram schematically shows the west to east changes observed in the rocks of the Fayston - Buels Gore area.

Continued compression produces Tn+1 faults that cut the GMA. Such faults might be associated with the development of Fn+2 folds from the GMA westward, and the reorientation of earlier Fn+1 folds to horizontal positions on the west side of the GMA. Static overgrowths of biotite and muscovite in the west are produced by tectonic loading as a result of shearing to the east.

Late west-dipping kink bands with down to the east motion develop in Domain 1. The kink bands are correlative with the late west-dipping pressure solution cleavage and associated folds from the Northfield Mountains eastward, which may represent the western limit of Acadian deformation.

Mesozoic tectonics produce fractures, faults, and early Cretaceous calc-alkaline dikes related to the rifting of North America during the creation of the Atlantic basin.

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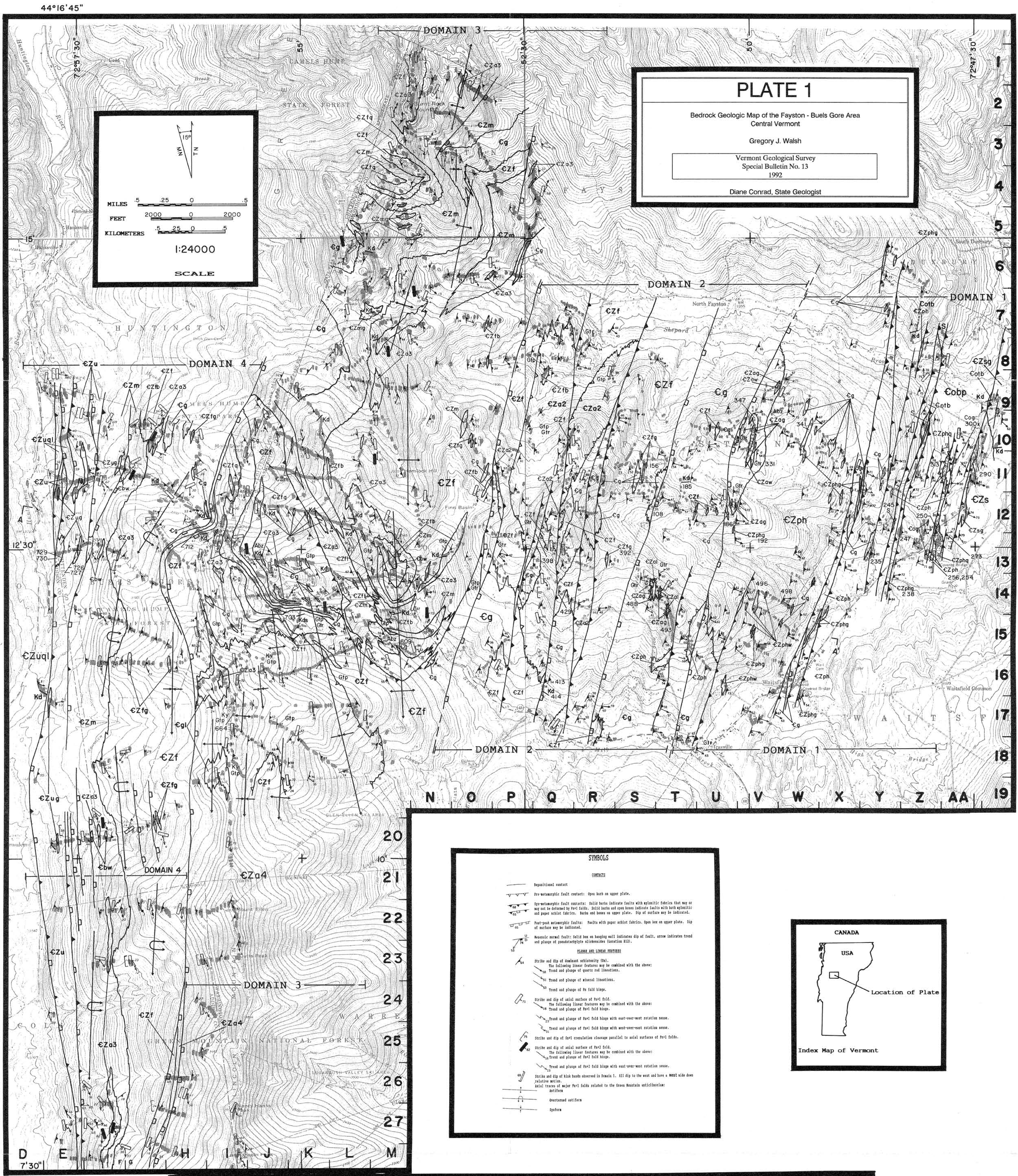

PLATE 1

Bedrock Geologic Map of the Fayston - Buels Gore Area
Central Vermont

Gregory J. Walsh

Vermont Geological Survey
Special Bulletin No. 13
1992

Diane Conrad, State Geologist

SYMBOLS

CONTACTS

Depositional contact

Pre-metamorphic fault contact: Open barb on upper plate.

Syn-metamorphic fault contacts: Solid barbs indicate faults with mylonitic fabrics that may or may not be deformed by Fn+1 folds. Solid barbs and open boxes indicate faults with both mylonitic and paper schist fabrics. Barbs and boxes on upper plate. Dip of surface may be indicated.

Post-peak metamorphic faults: Faults with paper schist fabrics. Open box on upper plate. Dip of surface may be indicated.

Mesozoic normal fault: Solid box on hanging wall indicates dip of fault, arrow indicates trend and plunge of pseudotachylyte slickensides (Location 812).

PLANAR AND LINEAR FEATURES

Strike and dip of dominant schistosity (Sn).
The following linear features may be combined with the above:
Trend and plunge of quartz rod lineations.
Trend and plunge of mineral lineations.
Trend and plunge of Fn fold hinge.

Strike and dip of axial surface of Fn+1 fold.
The following linear features may be combined with the above:
Trend and plunge of Fn+1 fold hinge.
Trend and plunge of Fn+1 fold hinge with east-over-west rotation sense.
Trend and plunge of Fn+1 fold hinge with west-over-east rotation sense.

Strike and dip of Sn+1 crenulation cleavage parallel to axial surfaces of Fn+1 folds.

Strike and dip of axial surface of Fn+2 fold.
The following linear features may be combined with the above:
Trend and plunge of Fn+2 fold hinge.
Trend and plunge of Fn+2 fold hinge with east-over-west rotation sense.

Strike and dip of kink bands observed in Domain 1. All dip to the west and have a west side down relative motion.

Axial traces of major Fn+1 folds related to the Green Mountain anticlinorium:
Antiform
Overturned antiform
Synform

LITHOLOGIES

Cretaceous Dikes:

Kd--diabase dike: Massive, vertically oriented, black, aphanitic, basic dikes ranging from 0.5 to 1.5 m in thickness.

Kds--albite epidosite dike: Tannish white to light greenish gray epidote-albite-quartz-chlorite-dolomite dike.

Underhill Formation:

CZuql--quartz laminated schist: Light gray to tan and rusty, fine grained, very well foliated, quartz-muscovite-chlorite-albite±dolomite schist. Quartz and albite laminations are separated by rusty laminations that may or may not contain highly disseminated dolomite.

CZu--schist and metawacke: Light gray to silvery green, fine grained, occasionally laminated quartz-albite-biotite-muscovite-chlorite schist that contains abundant quartz vein segregations. The laminations consist of quartz and albite and often give the schist a pinstripe appearance. Coarse grained, light grayish green, quartz-albite-biotite-chlorite-magnetite-pyrite schist/gneiss. The poorly developed foliation wraps around occasional pebbles of white quartz, albite, and rarely blue quartz up to 5 mm in diameter.

CZug--greenstone and amphibolite: Fine to medium grained, light to dark green, well foliated, amphibole-albite-chlorite-epidote±biotite±magnetite±pyrite schist. Very coarse to medium grained, dark green, poorly foliated, amphibole-albite-biotite-epidote-magnetite-pyrite amphibolite. Coarse (up to 1 cm) dark green amphiboles are randomly oriented in a finer grained plagioclase-biotite-epidote matrix.

Battell Formation:

Cbw--White River Member: Dark gray to black, rusty, carbonaceous muscovite-quartz-albite schist with white to black quartzites, brecciated black dolomitic marble, quartz vein segregations, and tan weathering pods and disseminated grains of dolomite.

Monastery Formation:

CZm--schist: Intercalated tan sandy quartz-muscovite-biotite-albite±garnet±graphite schist; silvery tan, fine grained, pearly sericite schist; silvery green muscovite-chlorite-quartz-albite schist; thin (up to 20 cm) light gray and tan quartzite; and tan to rusty weathering pods and disseminated grains of dolomite and ankerite. Garnet porphyroblasts measure up to 1 cm in diameter. Magnetite and pyrite occur sporadically. The schist contains intercalated dark gray to black, rusty, carbonaceous muscovite-quartz-albite schist with white to black quartzites near contacts with Cg and Cbw.

CZmg--greenstone: Fine grained, dark green, rusty weathering, well foliated chlorite-albite-calcite-epidote±pyrite schist.

Fayston Formation:

CZf--white albitic schist: Silvery green, medium to coarse grained muscovite-quartz-albite-chlorite±biotite±garnet±magnetite schist with occasional thin (5 to 15 cm) white to light gray quartzites. White to light gray albite porphyroblasts up to 7.5 mm in diameter, and abundant quartz vein segregations are characteristic.

CZfq--quartzo-feldspathic granofels: Coarse grained, light gray albite-quartz-muscovite-chlorite±magnetite granofels. The rock is massive with a poorly defined foliation.

CZfb--quartz-biotite gneiss: Medium to coarse grained quartz-albite-biotite±magnetite±pyrite schist/gneiss. The color of the rock is a mixture of white and black that resembles salt and pepper.

CZft--quartz-muscovite-tourmaline schist: Medium grained, dark silvery gray to rusty colored, quartz-muscovite-albite-tourmaline-chlorite schist. Tourmaline occurs as black, randomly oriented, euhedral crystals up to 1 mm in diameter and 1.5 cm in length.

CZfg--greenstone: Fine grained, dark green, rusty weathering, well foliated chlorite-albite-calcite-epidote±pyrite schist.

Granville Formation:

Cg--carbonaceous albitic schist: Dark colored, rusty weathering, muscovite-quartz-albite-chlorite-graphite±pyrite schist. The schist is characterized by discontinuous patches of graphite, gray to black albite porphyroblasts up to 5 mm in diameter, quartz vein segregations, and discontinuous white, black and white, gray, and black quartzites ranging from 5 to 30 cm in thickness and up to 100 m in length.

Cgl--Lincoln Gap Member: Black, rusty weathering quartz-muscovite-albite-graphite±pyrite schist, and up to 5 cm thick black quartz-graphite-muscovite-albite quartzite. Graphite occurs throughout the rock unlike the patchy occurrence in the carbonaceous albitic schist.

Mount Abraham Schist:

CZa1--Mount Abraham schist: Fine grained, silvery tan to gray-green muscovite-paragonite-quartz-chlorite-chloritoid-garnet±magnetite schist with abundant quartz vein segregations. Fine grained mica gives the foliation surfaces a characteristic pearly sheen in which individual grains of mica are not discernable.

CZag--greenstone: Fine grained light green to light yellow epidote-actinolite-chlorite schist. The greenstone often has a green and yellow banded appearance due to compositional layering of actinolite and chlorite with epidote minerals.

CZaw--metawacke: Coarse grained, weakly foliated, greenish gray quartz-albite-muscovite-biotite±magnetite±pyrite granofels. Quartz vein segregations are characteristically absent.

CZa2--Mount Abraham schist: Fine grained, steel blue-gray to silvery tan, muscovite-paragonite-quartz-chlorite-chloritoid-garnet±magnetite schist with abundant quartz vein segregations. CZa2 is more homogeneous than CZa1, and greenstone and metawacke are absent.

CZa3--Mount Abraham schist: CZa3 is similar to CZa1 and CZa2. The color of the white mica sheen is more often a silvery white than a tan. Tan foliation surfaces are present, but not as common as in CZa1 and CZa2. Garnet porphyroblasts up to 2 cm in diameter are the largest garnets seen in the Mount Abraham Schist. Significant amounts of microscopic allanite are unique to CZa3.

CZa4--Mount Abraham schist: Fine grained, steel blue-gray muscovite-paragonite-quartz-chlorite-chloritoid-kyanite-garnet±magnetite schist with abundant quartz vein segregations. Kyanite blades are usually 0.5 to 1.0 cm in length and up to 0.5 mm in diameter. Chloritoid porphyroblasts can reach up to 1.0 cm.

Pinney Hollow Formation:

CZph--silvery green schist: Fine to medium grained, silvery green muscovite-quartz-albite-chlorite±magnetite schist with distinct quartz vein segregations, elongate streaks of chlorite as mineral lineations on the foliation surfaces, and minor thin (less than 10 cm) gray quartzites and thin (less than 20 cm) quartzo-feldspathic layers.

CZphg--greenstone: Fairly homogenous, fine to medium grained, light green chlorite-albite-actinolite-epidote schist.

CZphw--metawacke: Homogeneous, coarse grained, weakly foliated, greenish gray quartz-albite-muscovite-chlorite-biotite±magnetite±pyrite granofels. Quartz vein segregations are characteristically absent.

CZphq--quartzose schist: Fine grained, white to light gray green, well foliated quartz-albite-muscovite schist to albitic quartzite. Fine grained quartz and albite give the rock a sugary or sandy appearance.

Ottauquechee Formation:

Cobp--black phyllite: Black quartz-muscovite-graphite phyllite. Foliation surfaces have a sheen due to the abundance of graphite. Quartz occurs as thin laminations and vein segregations. The phyllite is characterized by 1-2 cm pyrite molds and a lack of albite.

Cog--gray carbonate schist: Well foliated, fine grained, light greenish gray, quartz-chlorite-calcite-muscovite-graphite schist. Weathering of calcite gives the rock a characteristic rusty rind. Cross muscovite crystals (up to 2 mm) are common.

Cotb--Thatcher Brook Member: Heterogeneous, rusty weathering muscovite-quartz-albite-chlorite-graphite schist with dark gray to black, often banded, quartzite (Q). The graphitic schist contains distinct discontinuous layers, or segregations, of graphite as opposed to the patches of graphite in the carbonaceous albitic schist of the Granville Formation. Heterogeneities include the distribution of graphite, discontinuous chlorite-rich, and albite-rich graphite-poor lenses, and the close association of the graphitic schist with exotic greenstones, serpentinites, and talc schists.

Stowe Formation:

CZs--Stowe schist: Fine grained, silvery to dark green quartz-muscovite-albite-chlorite±magnetite schist with abundant quartz vein segregations. Thick laminations of chlorite are often dark green to bluish black. Trace patches of graphite are found with near the contact with Cobp. Occasional dark gray albite occurs in the schist at locations that contain patchy graphite.

CZsg--greenstone: Homogenous, fine grained, light green actinolite-albite-epidote-calcite-chlorite schist.

Ultramafics:

S--serpentinite and talc-magnesite schist: Serpentinite is a dense, massive, weakly foliated, dark green serpentine-magnetite rock. The serpentinites are cut by numerous fractures and veins that are filled with magnesite and limonite. Talc-magnesite schists are soft, well-foliated, light gray to tan schists. Magnesite occurs as irregular, rusty patches and spots in the schist.

Note: Except for listing the Cretaceous dikes first, the order of description of the other rock units does not imply any relative age, but rather their occurrence in the area from west to east. The age of the these units is believed to range from Late Proterozoic to Middle Ordovician.

PETROGRAPHIC FEATURES

Metamorphic features recognized in thin section. Symbols used on map correspond to samples collected for petrographic analysis, although some of the following features can be recognized in the field.

Gtp = Prograde garnet: Euhedral and unaltered; observed primarily in Domain 3 and 4, and to a lesser extent in Domain 2.
Gtr = Retrograde garnet: Anhedral and altered primarily to chlorite; observed primarily in Domains 1, 2 and 4, and to a lesser extent in Domain 3.
Ky = Prograde kyanite: Euhedral and relatively unaltered; observed only in CZa4 in Domain 3.
Abz = Optically zoned albite porphyroblasts. Plagioclase probably has an albite-oligoclase composition.

MISCELLANEOUS

A - A' Line of cross-section

156 Sample locality referred to in text.

Bedrock exposures

Note: Topographic base compiled from the Mt. Ellen, Waitsfield, Huntington, and Waterbury 7.5' U.S.G.S. quadrangles. Contour Interval = 20 feet.

Vertical Cross Section A-A'

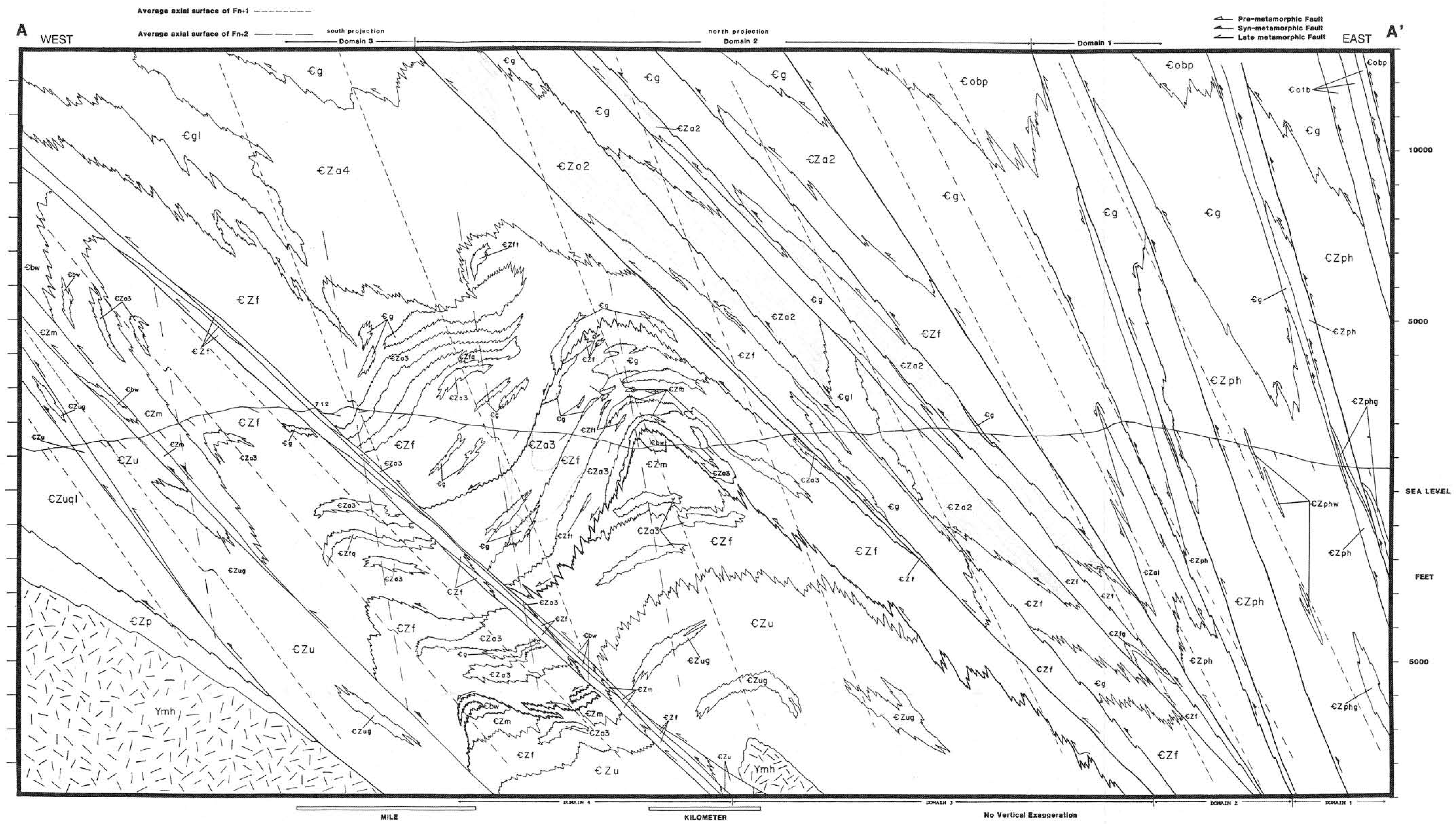

PLATE 2

Cross Section A-A' through the Fayston - Buels Gore Area
Central Vermont

Gregory J. Walsh

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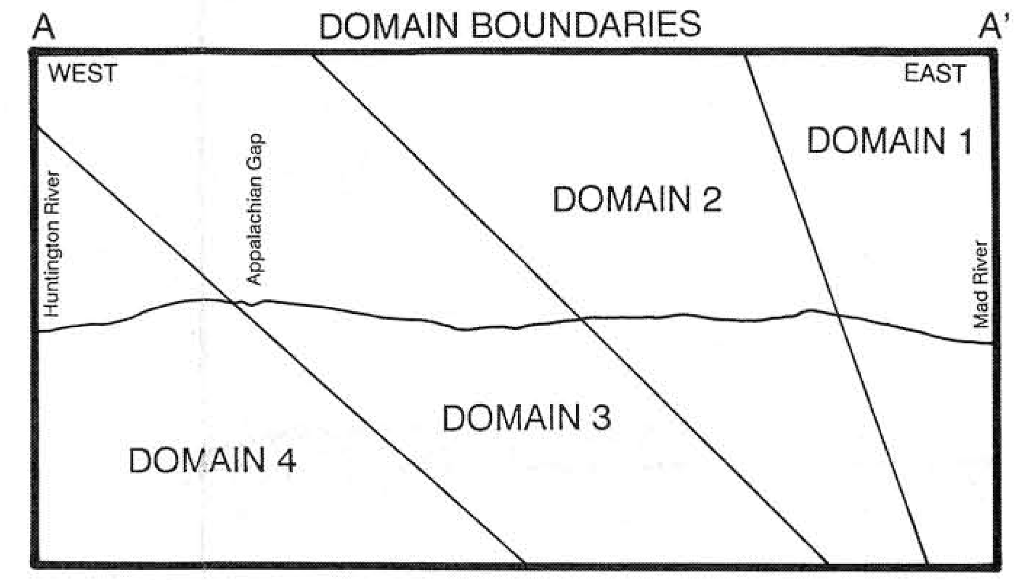
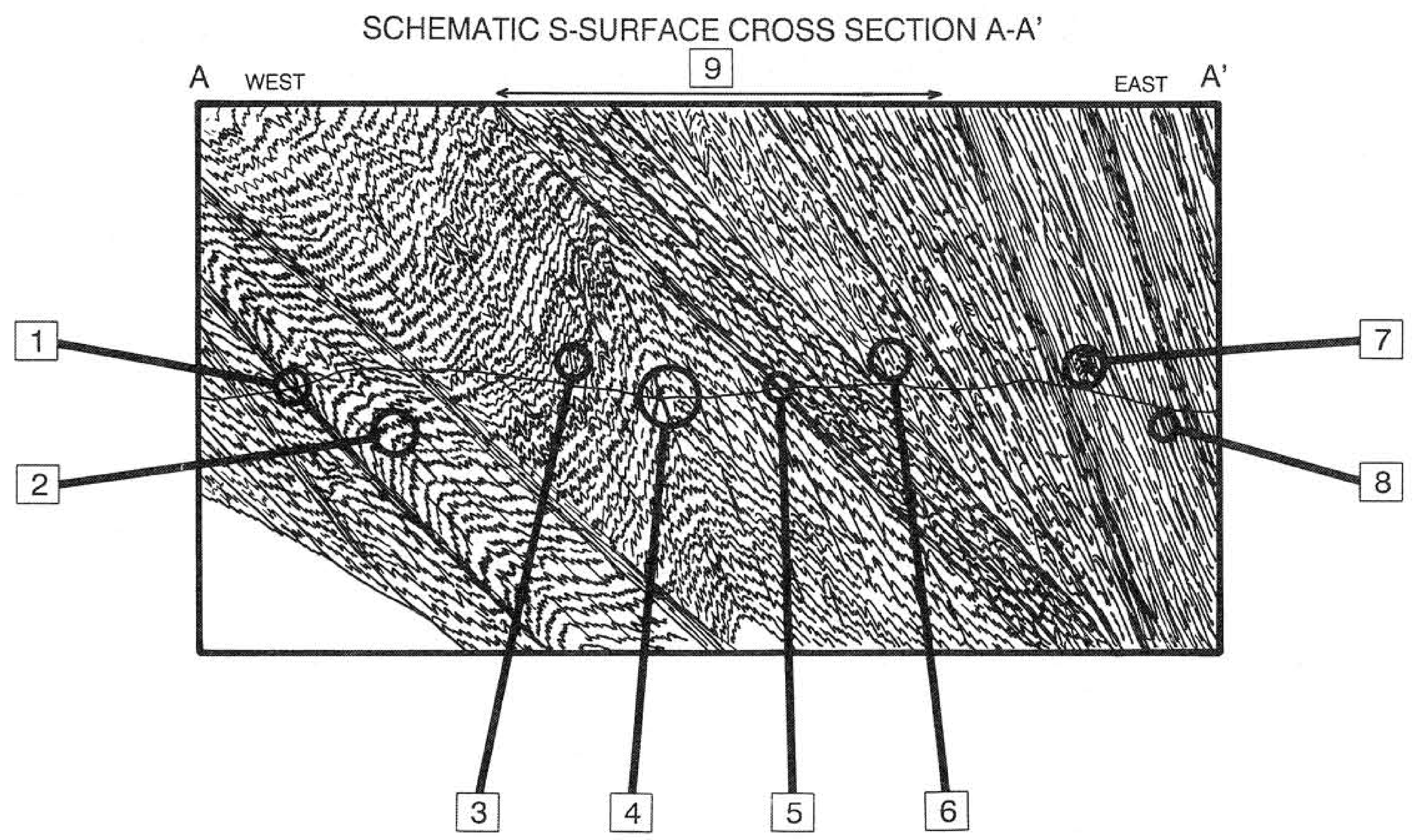
Special Bulletin No. 13
1992

Diane Conrad, State Geologist





S-SURFACE CROSS SECTION EXPLANATION

1. Fault between CZm and CZu. No evidence of garnet grade metamorphism west of the fault. Fault corresponds with Cady and others' (1962) western garnet isograd.
2. Overturned antiform on the west flank of the GMA.
3. Unaltered garnet, kyanite, and zoned plagioclase (albite-oligoclase) record peak metamorphic conditions during the development of western Sn (S1) and up to the development of Fn+1 folds.
4. Crest of the GMA. Sn+1 (S2) crenulation cleavage is present in fine-grained schists, but not coarse-grained rocks, throughout domains 3 and 4. Fn+1 fold hinges have shallow plunges.
5. Ductile mylonitic fault zones with east-over-west kinematic indicators.
6. Garnet and chloritoid are retrograded during the progressive development of Sn+1 (S2) from a crenulation cleavage to a penetrative schistosity. Fn+1 fold hinges record progressive rotation to down dip orientations.
7. Some Fn+1 folds are preserved in more competent metawackes, greenstones, and quartzites. Sn+1 (S2) is the pervasive schistosity. The orientation of Sn+1 (S2) corresponds with the orientation of the crenulation cleavage in domains 2 and 3.
8. The complete replacement of garnet and chloritoid by chlorite records the overprinting of earlier higher grade assemblages during the development of Sn+1 (S2).
9. The transposition of Fn+1 folds in domain 2 is associated with the progressive development of Sn+1 from west to east.

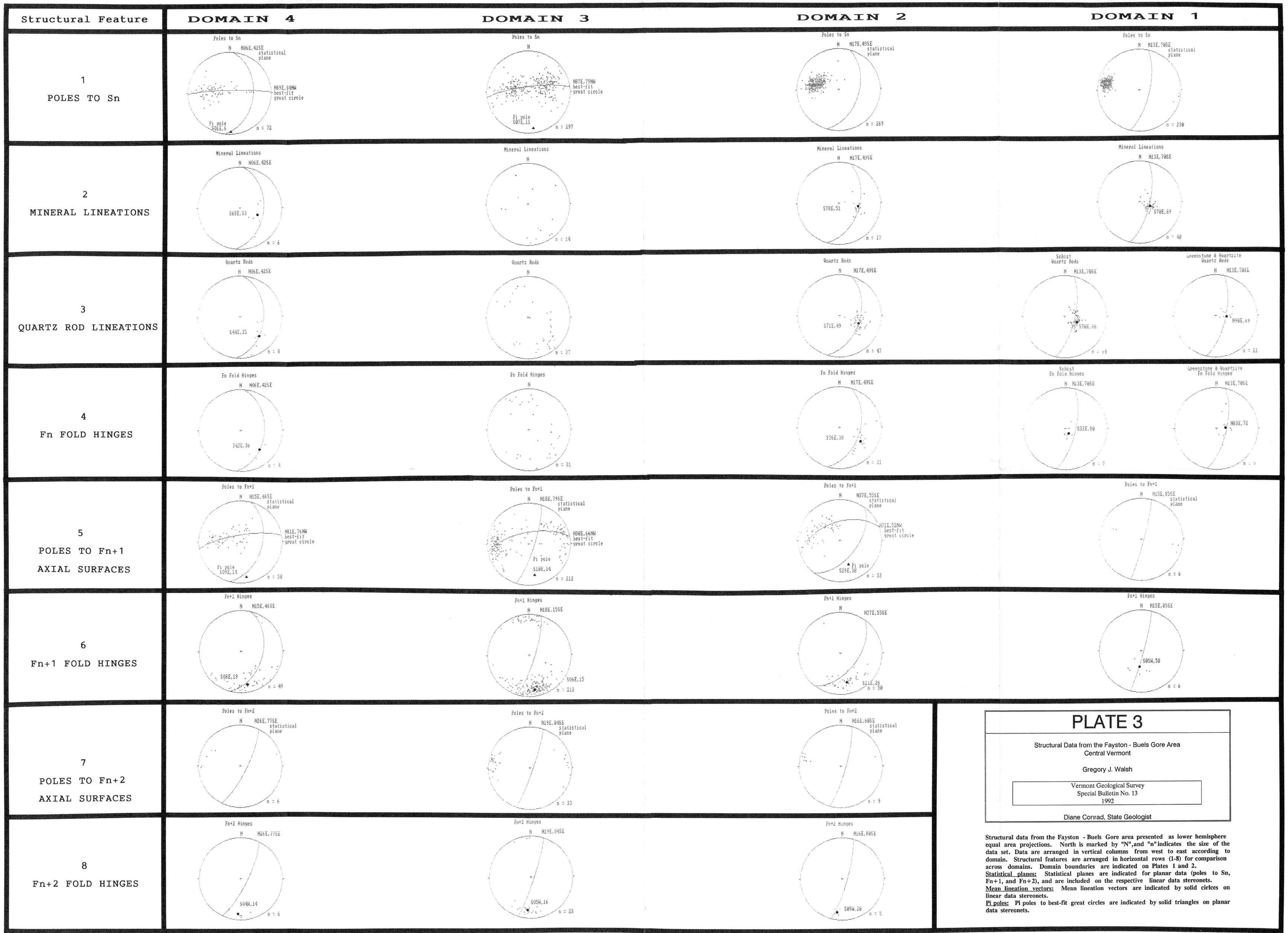


PLATE 3

Structural Data from the Fayston - Buels Gore Area
Central Vermont

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Structural data from the Fayston - Buels Gore area presented as lower hemisphere equal area projections. North is marked by "N", and "n" indicates the size of the data set. Data are arranged in vertical columns from west to east according to domain. Structural features are arranged in horizontal rows (1-8) for comparison across domains. Domain boundaries are indicated on Plates 1 and 2.

<u>Statistical planes:</u> Statistical planes are indicated for planar data (poles to Sn, Fn+1, and Fn+2), and are included on the respective linear data stereonets.

<u>Mean lineation vectors:</u> Mean lineation vectors are indicated by solid cirlces on linear data stereonets.

<u>Pi poles:</u> Pi poles to best-fit great circles are indicated by solid triangles on planar data stereonets.

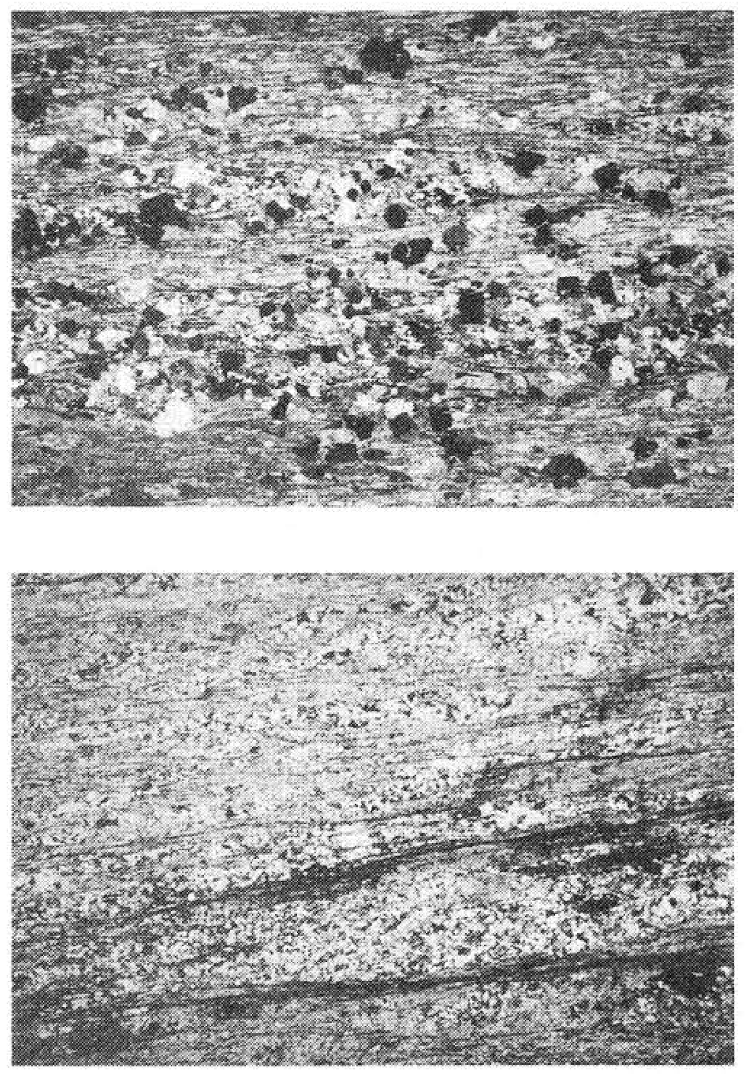

Figure 2. Photomicrographs of unmylonitized and mylonitized silvery schist of the Pinney Hollow (CZph). TOP: The photo shows the texture of the schist as it appears 3 meters west of the fault zone--note abundant porphyroblasts of albite and magnetite. BOTTOM: The photo shows the mylonitic texture--note the absence of the porphyroblasts and the grain size reduction of quartz. Both thin sections are cut perpendicular to the foliation and lineation. Samples 250A1GW (top) and 250GW (Z-12). FOV = 2.0 X 1.2 cm. Polarized light.

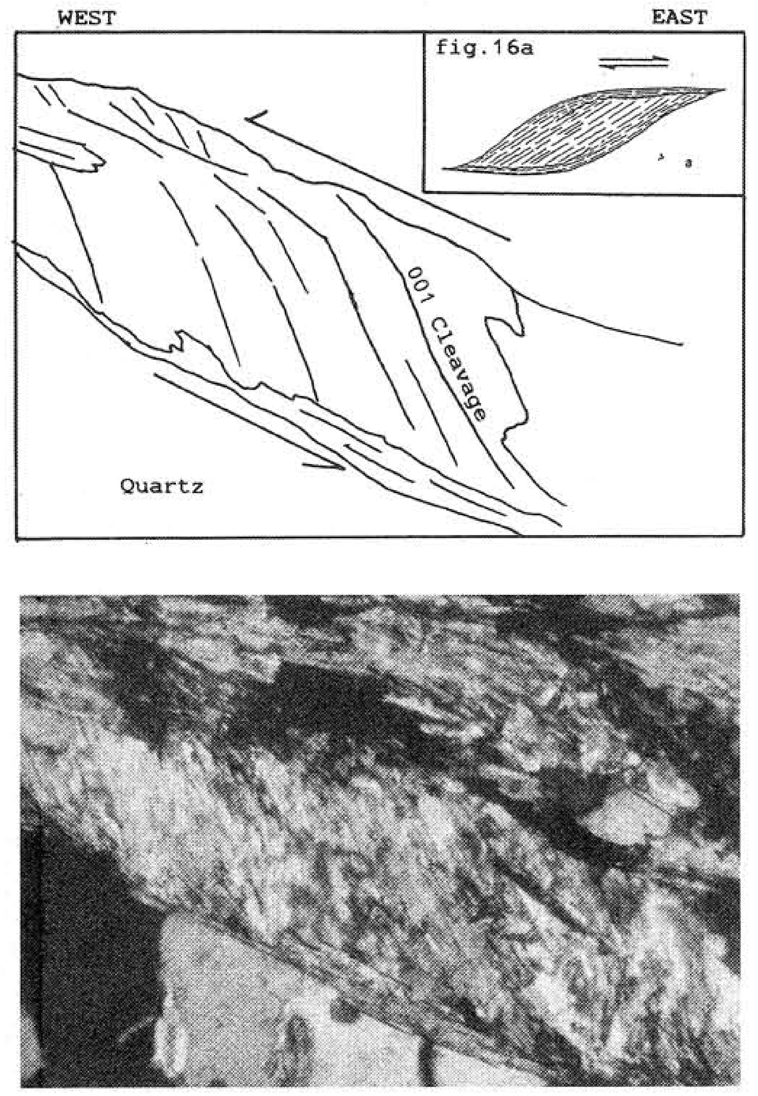

Figure 3. Photomicrograph of east-over-west mica fish from the Ottauquechee sericite mylonite. View is to the north. Mica fish is virtually identical to that of Lister and Snoke (1984, fig. 16a). Sample 313BGW (Z-11). FOV = 0.53 X 0.35 mm. 10X objective in polarized light.

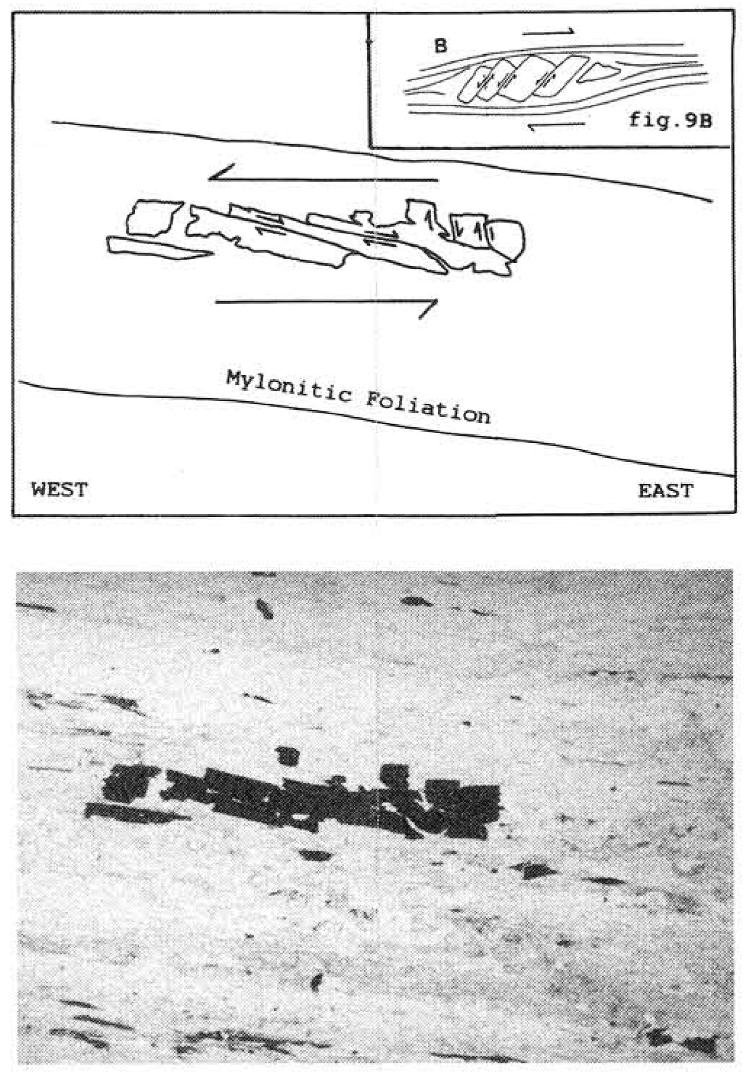

Figure 4. Photomicrograph of broken and displaced hard grain with east-over-west asymmetry. View is to the north. Hard grain is ilmenite. Note how the sense of displacement in the grain is opposite of the overall sense of shear. Example resembles that of Simpson and Schmid (1983; fig. 9B). Sample 429GW (R-14). FOV = 1.2 X 0.7 cm. Plane light.

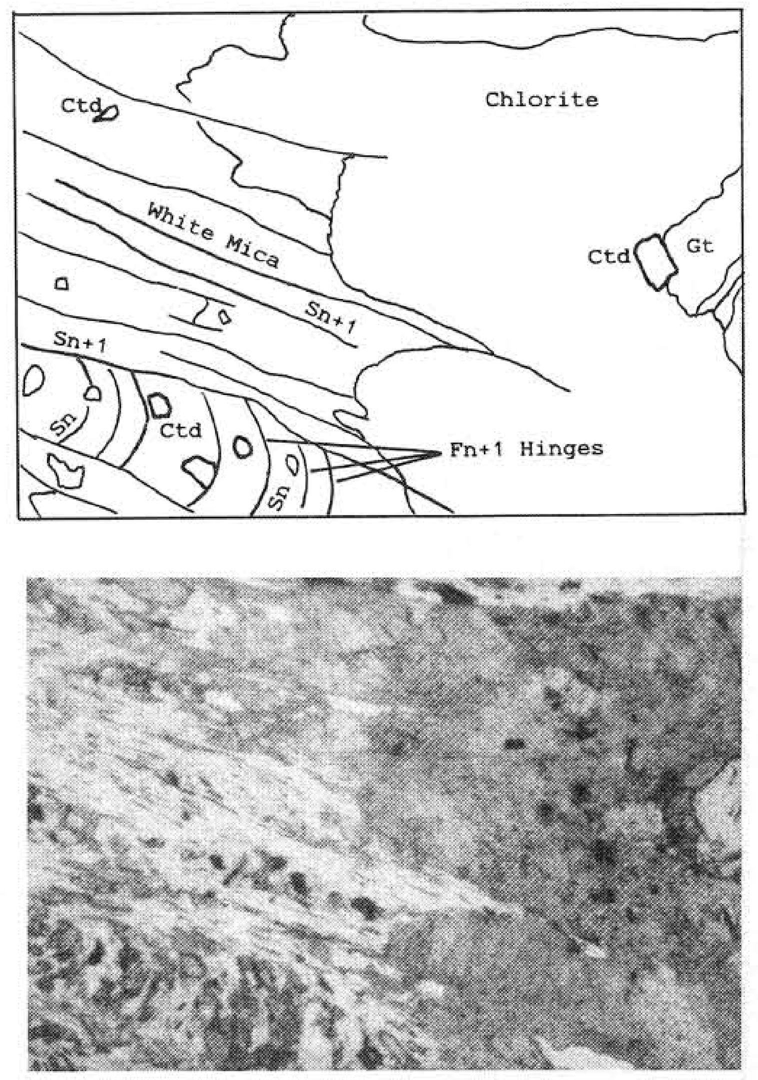

Figure 7. Photomicrograph of garnet retrograded during Sn+1 developmemt. Only a slight remnant of garnet remains surrounded by a large clot of chlorite. Chloritoid is also retrograded during Sn+1 development. Relict Fn+1 fold hinges contain relatively unaltered chloritoid. Sample 331AGW (V-11). FOV = 1.5 X 1.0 mm. 3.5X objective in plane light.

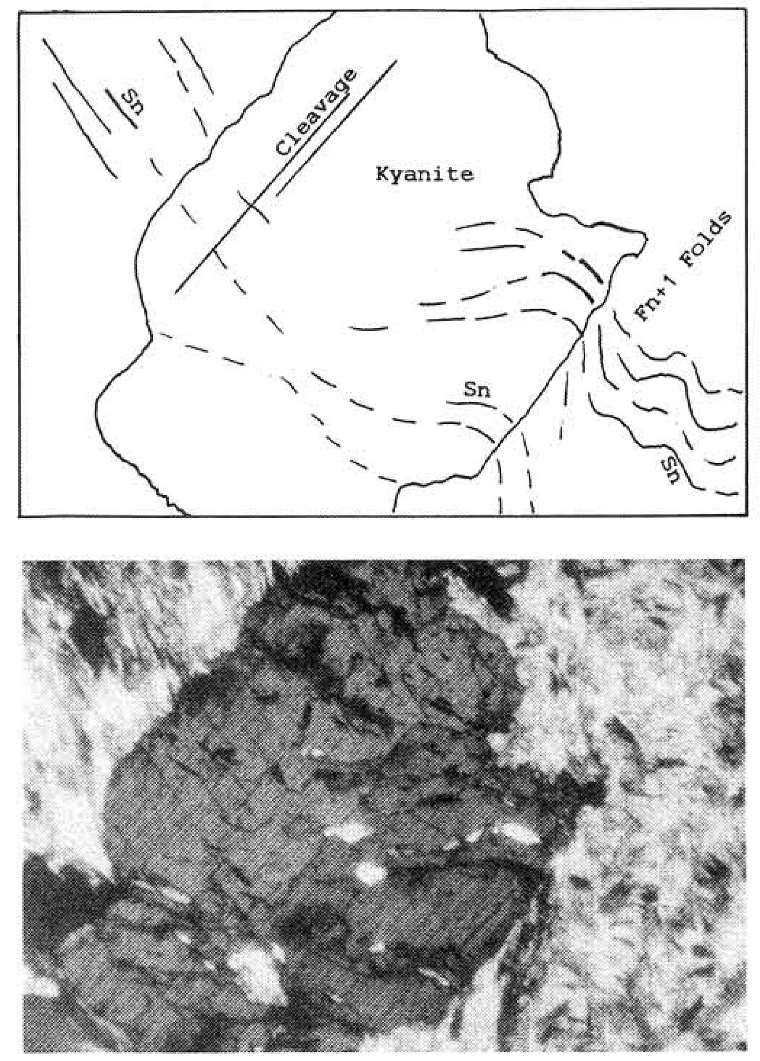

Figure 8. Photomicrograph of Fn+1 aged kyanite in CZa4. Folded inclusion trails in the euhedral kyanite porphyroblast indicate that the growth of the mineral took place during and after the development of some of the folds. Sample 664GW (I-18). FOV = 1.5 X 1.0 mm. 3.5X objective in polarized light.

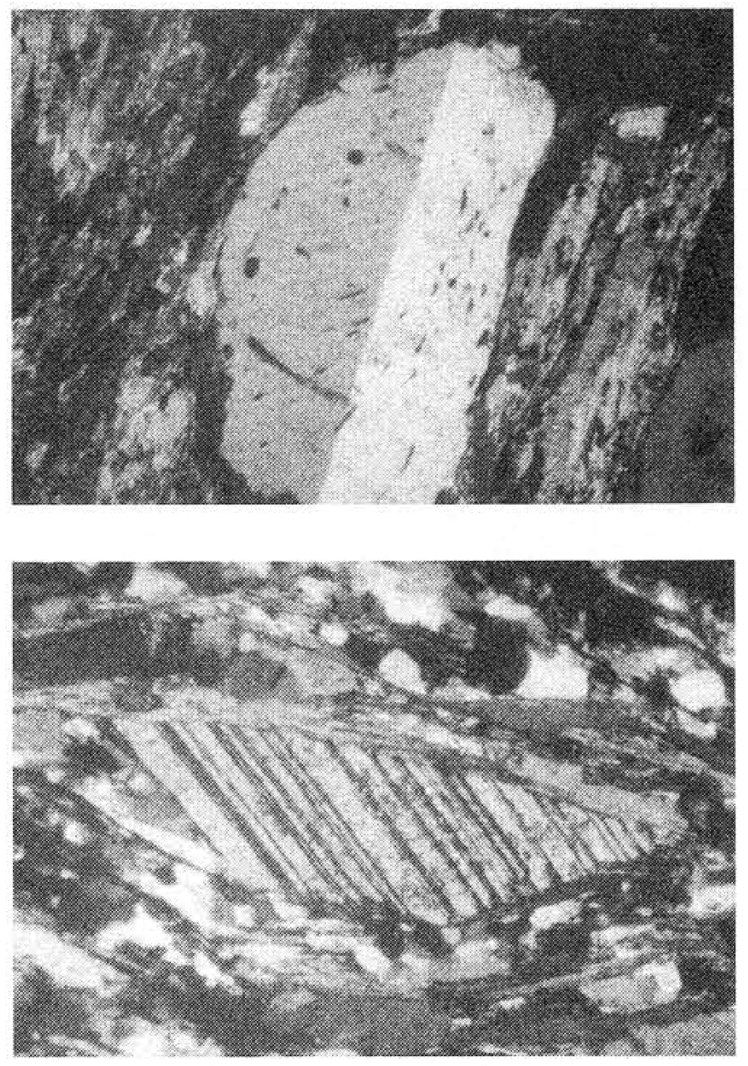

Figure 9. Photomicrographs of two types of growth twins observed in albite. TOP: Albite porphyroblast with single, simple albite twin dividing two subindividuals. Sample 398BGW1 (P-13) from CZf. BOTTOM: Angular albite porphyroclast with multiple polysynthetic growths twins, most likely on the albite or pericline law. Note how twins are parallel, perfectly straight, and cut across the entire grain. Sample 726AGW (E-13) from CZu. Both photos: FOV = 1.5 X 1.0 mm. 3.5X objective in polarized light.

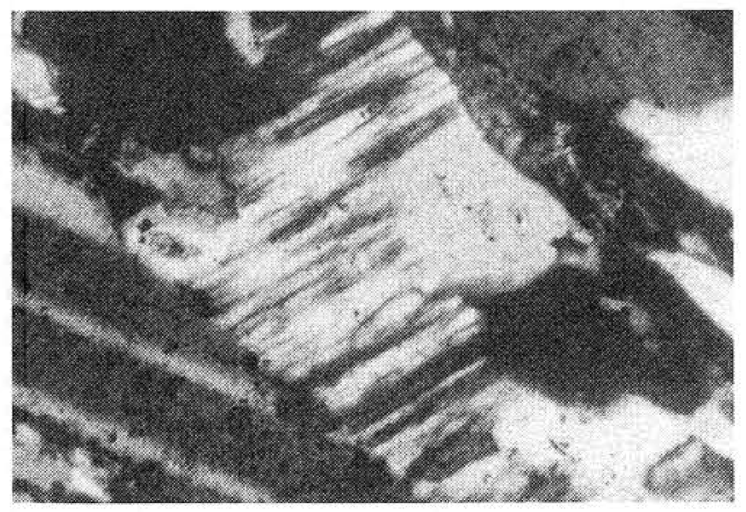

Figure 10. Photomicrograph of albite porphyroclast with deformation twins. Compare with twins in figure 9 and note how the twins are not parallel, not straight, and do not cross the entire grain. Clast is partially recrystallized to "water clear" albite in several spots. Sample 726AGW (E-13) from "pinstripe" CZu. FOV = 0.53 X 0.35 mm. 10X objective in polarized light.

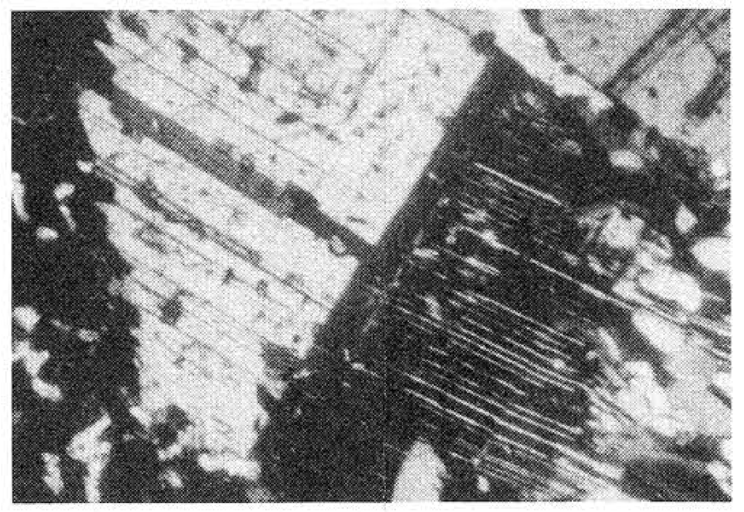

Figure 11. Photomicrograph of albite porphyroclast with complex twins from the Underhill amphibolite. Twins are most likely on the simple albite and multiple pericline laws. Such twins are characteristic of an igneous, not a metamorphic, origin. Sample 726GW (E-13). FOV = 1.5 X 1.0 mm. 3.5X objective in polarized light.

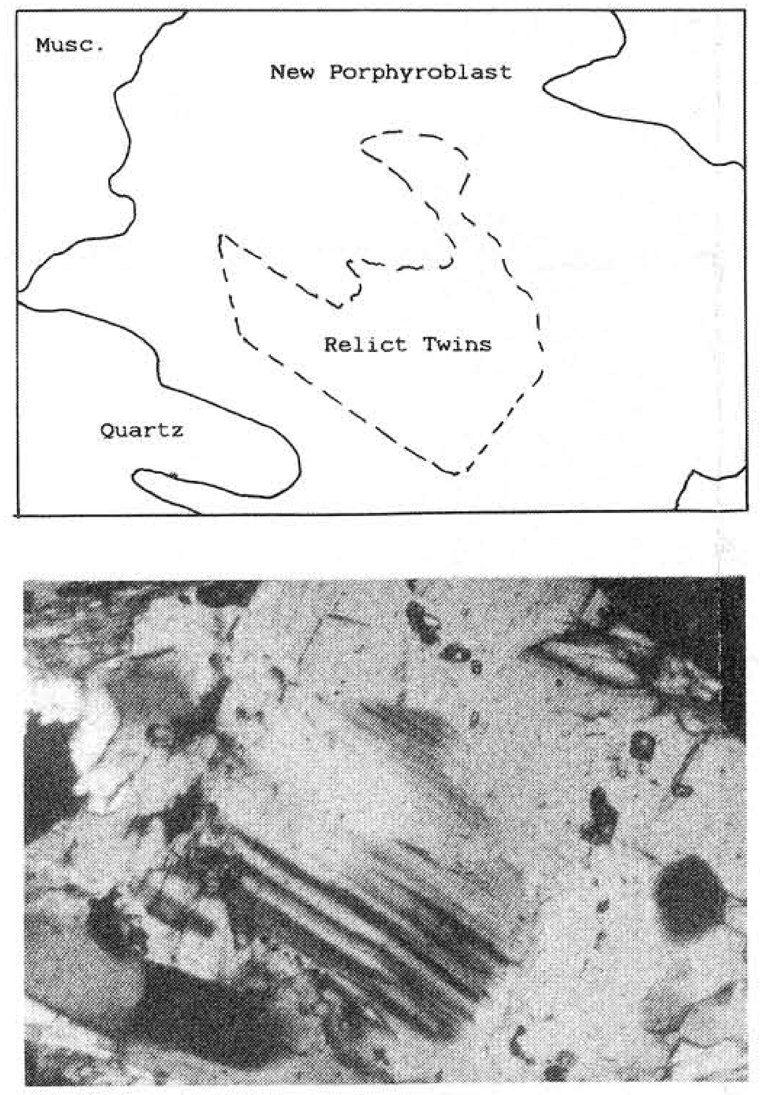

Figure 12. Photomicrograph of albite porphyroblast with relict twins in the center. Fuzzy relict twins are all that remain of an original grain with growth twins indicating that the older porphyroclast acted as a nucleation site for the younger porphyroblast. Sample 496GW (V-14). FOV = 0.53 X 0.35 mm. 10X objective in polarized light.

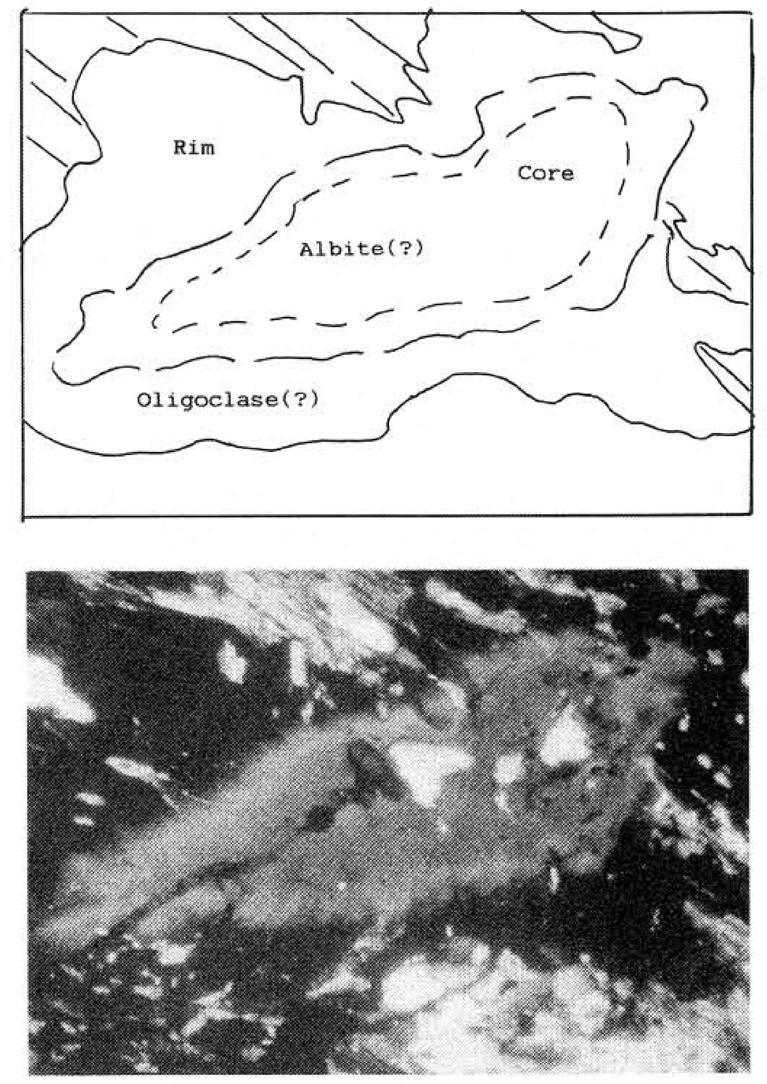

Figure 13. Photomicrograph of optically zoned plagioclase porphyroblast in CZf. Ragged outline of porphyroblast indicates that the plagioclase was in the process of growth. Lighter core is probably albite, and darker rim is most likely oligoclase indicating growth during progressive metamorphism. Sample 609GW (M-15). FOV = 1.5 X 1.0 mm. 3.5X objective in polarized light.

PLATE 4

Photomicrographs from the Fayston - Buels Gore Area
Central Vermont

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