BEDROCK GEOLOGY OF THE BRATTLEBORO QUADRANGLE, VERMONT-NEW HAMPSHIRE

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TABLE OF CONTENTS

ILLUSTRATIONS

PLATES

(in pocket at back)

TABLES

ABSTRACT

The Brattleboro quadrangle of southeastern Vermont and southwestern New Hampshire is underlain principally by metamorphosed Lower to Middle Paleozoic sedimentary and volcanic rocks. The area sits astride portions of three regional tectonic zones, from west to east: the Green Mountain anticlinorium, the Connecticut River-Gaspé synclinorium and the Bronson Hill anticlinorium. The first two of these zones are discussed as the Western Sequence and include strata ranging in age from Precambrian through Silurian. The geology of the Bronson Hill anticlinorium is discussed as the Eastern Sequence covering strata ranging in age from Ordovician to Lower Devonian. Granitic intrusive rocks in the area belong to the Ordovician-aged Oliverian Plutonic Series and the Devonian-aged New Hampshire Plutonic Series.

The structural deformation in the area is divided into two major folding stages: an earlier one during which large recumbent folds or nappes developed; and a later one in which the nappes were arched upward by rising domes. The Guilford dome, with the Waits River Formation in its exposed core, is a prominent structural feature in the central part of the quadrangle. The emplacement of this dome deformed the axial surface of the earlier Prospect Hill recumbent fold. The south end of the Athens dome, with Precambrian rocks in its core, lies in the northwestern part of the quadrangle. Two Oliverian gneiss-cored domes of the Bronson Hill anticlinorium occur in the eastern part of the quadrangle. These two domes, the Vernon dome and part of the Keene dome, deform the earlier Bernardston and Skitchewaug nappes. The Wellington Hill anticline, in the northeast corner of the quadrangle, may represent the position of a third dome, whose Oliverian core is still buried. At least five stages of minor folding occurred prior to, during and after the major deformational events.

Regional metamorphism of the area ranges from the lower greenschist facies to the upper amphibolite facies in a Barrovian-type metamorphic sequence. The principal post-Precambrian metamorphism and structural deformation took place during the Acadian orogeny.

CHAPTER 1
INTRODUCTION

Location

The Brattleboro 15 minute quadrangle (U.S. Geological Survey, 1954) occupies an area of about 202 sq. miles in Windham Co., southeastern Vermont, and about 23 sq. miles in Cheshire Co., southeastern New Hamshire (Figure 1-1). It lies between north latitudes 43°00′ and 42°45′ and west longitudes 72°30′ and 72°45′. Brattleboro, Vermont, on the Connecticut River, is the largest town in the quadrangle, with a population of about 15,000.

Topography, Drainage, and Geologic Exposure

The Brattleboro quadrangle lies along the western margin of the New England upland section and the eastern margin of the Green Mountain section of the New England physiographic province (Fenneman, 1938). The maximum topographic relief in the quadrangle is 1840 feet. The highest point, 2020 feet above sea level is Jolly Mountain, Halifax; the lowest point, 180 feet above sea level, is along the Connecticut River in the southeastern part of the quadrangle. The local topographic relief averages 500 to 1000 feet.

The area lies in the Connecticut River drainage basin. The Connecticut River itself flows in a southerly direction through the eastern part of the quadrangle. The low water mark on its western bank forms the boundary between Vermont and New Hampshire. The West and Ashuelot Rivers are major tributaries to the Connecticut within the area. The major drainages were pre-glacial in origin.

In general, bedrock is moderately well exposed throughout the quadrangle and permits the detailed tracing of geological units. On many of the region's hills, particularly in the western part of the quadrangle or along the Wantastiquet Mountain ridge, large glacially polished areas near the summits provide nearly continuous outcrop. Exposures of bedrock are generally better on south-facing slopes than on north-facing ones. The north-facing slopes are typically more gentle and covered with a veneer of glacial till. Pleistocene kame terrace sand deposits obscure the bedrock on the valley floors of many of the larger streams, particularly the Connecticut River.

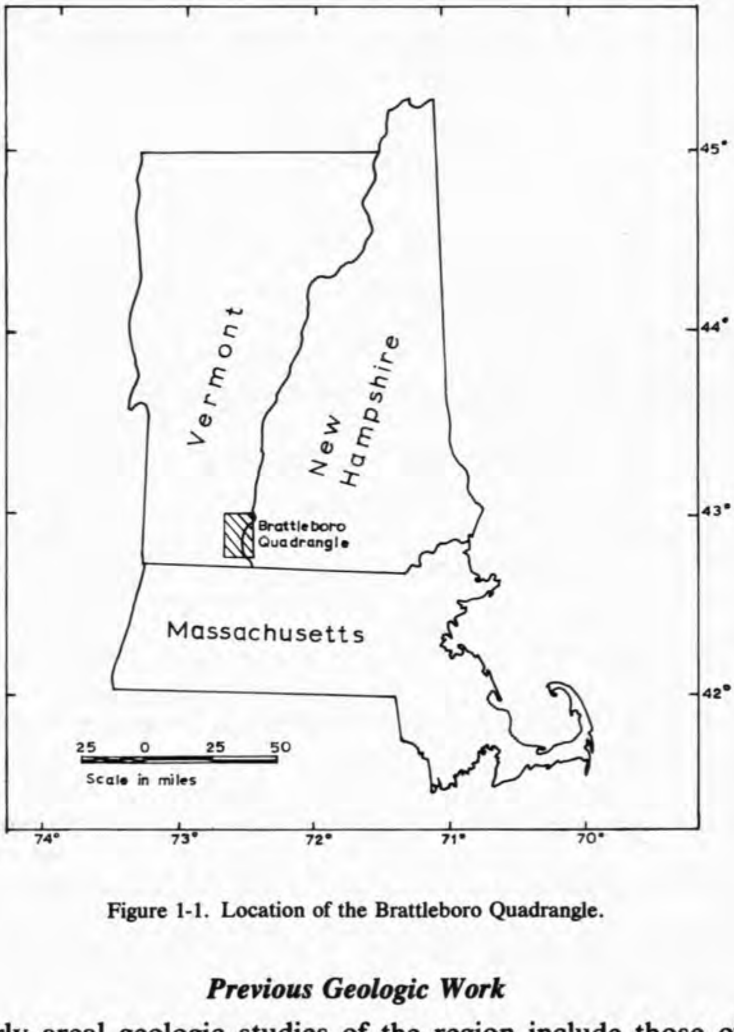


Figure 1-1. Location of the Brattleboro Quadrangle.

Previous Geologic Work

Early areal geologic studies of the region include those of E. Hitchcock (1823), E. Hitchcock et al. (1861), C. H. Hitchcock (1874, 1877, 1878), and C.T. Jackson (1844). B. K. Emerson (1917) incorporated the southern edge of the Brattleboro quadrangle in the geologic map of Massachusetts and Rhode Island. Richardson (1933), Richardson and Maynard (1939), and Richards (1931) studied portions of the area in more detail.

In recent years, a great deal of detailed geological work has been done in Vermont, New Hampshire, and western Massachusetts. This work, begun in New Hampshire by Billings (1937) and in southern Vermont by Thompson (1950) and Rosenfeld (1954), has led to the publication of geologic maps of Vermont (Doll et al., 1961) and New Hampshire (Billings, 1955). Zen (1981, 1983) has recently compiled a new geologic map of Massachusetts. Moore (1949) mapped the eastern third of the Brattleboro quadrangle in his studies of the Keene-Brattleboro area, and many of his conclusions remain unchanged in the present study. Balk compiled an unpublished map of parts of the area during his work in Massachusetts. Other geologic mapping in the areas immediately surrounding the Brattleboro quadrangle in Vermont have been compiled by Skehan (1961; Wilmington quad.), Rosenfeld (1954; Saxtons River quad.), and Kruger (1946; Bellows Falls quad.). Parts of the eastern Brattleboro quadrangle have been included in more regional geological maps by Thompson et al. (1968) and Robinson, Thompson and Rosenfeld (1979). Geologic mapping of quadrangles to the south of the Brattleboro area have been prepared by Balk (1956A, B), Segerstrom (1956), Pferd (1981), Hatch and Hartshorn (1968) and Chidester et al. (1967). For indexes of geologic mapping covering broader areas the reader is referred to the following: Vermont, Doll et al. (1961); New Hampshire, Billings (1955); southern Vermont and northwestern Massachusetts, Skehan (1961) and Norton (1967); and Massachusetts, Zen (1981, 1983). The surficial deposits of the quadrangle have been shown on the surficial geologic map of Vermont (Doll et al., 1970).

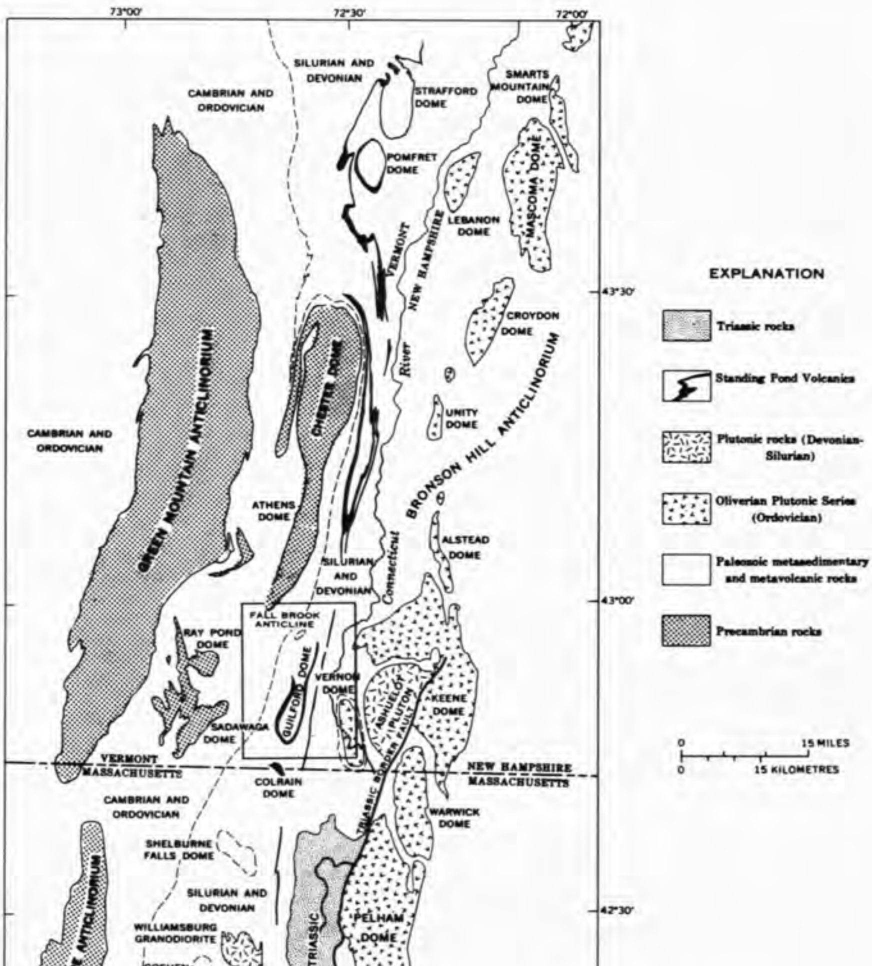
Previous detailed studies on the Black Mountain Granite included those reported by Dale (1909, 1910, 1923) and those of Church (1937). Naylor (1971) has published the most recent radiometric age determinations for this granite.

Most of the Brattleboro quadrangle was mapped as parts of PhD dissertations by Trask (1964) and Hepburn (1972A). Readers interested in more detailed accounts of the geology of the area than could be included here, including thin section modes, are referred to these theses.

Regional Geologic Setting

Two structural features of regional extent pass through the Brattleboro quadrangle, and a third lies just to the west (Figure 1-2). Most of the quadrangle is in the Connecticut River-Gaspé synclinorium (Doll et al., 1961; Cady, 1960). Here Siluro-Devonian rocks are exposed between the Bronson Hill anticlinorium to the east and the Green Mountain anticlinorium to the

Figure 1-2. Map showing the regional geologic features and the location of the Brattleboro quadrangle. The contact between the Ordovician and older rocks to the west and the Silurian-Devonian rocks to the east is shown for Vermont and Massachusetts west of the Connecticut River. East of the river, the Paleozoic metasedimentary rocks are undifferentiated. Modified from Hepburn (1975), Doll et al. (1961), Thompson et al. (1968), and Billings (1956).



west. The Bronson Hill anticlinorium (Billings, 1956; Thompson et al., 1968) is a complex series of roughly *en echelon* domes with cores of granitic gneiss. The Vernon dome in the southeastern part of the quadrangle and the Keene dome in the eastern part are two of about twenty such structures in the anticlinorium, which stretches from Berlin, New Hampshire to Long Island Sound (Thompson et al., 1968). Lower Paleozoic metasediments and metavolcanics in the Eastern Sequence of the Brattleboro area occur as both mantling strata to these domes and in large recumbent folds that pre-date the doming. The axial surfaces of these folds have been arched by the rising of the domes. The Partridge Formation is exposed in the anticlinal core of the regional Bernardston nappe in West Chesterfield (Plate I, Figure 6-1). A small area along Catsbane Brook, Chesterfield, in the very eastern part of the quadrangle contains the Clough Formation in the next structually higher recumbent fold, the Skitchewaug nappe (Thompson et al., 1968).

The rocks in the western portion of the quadrangle are metamorphosed Ordovician sediments and volcanics in a generally eastward topping sequence on the eastern flank of the Green Mountain anticlinorium (Doll et al., 1961). This sequence is disrupted just west of the quadrangle by the Sadawga and Ray Pond domes and by the Chester and Athens domes just north of the quadrangle (Figure 1-2). The southern tip of the Athens dome extends into the northwestern corner of the Brattleboro quadrangle (Plate I). These four domes are mantled gneiss domes with Precambrian rocks exposed in their cores and are part of a belt of domes along or just west of the axial region of the Connecticut River-Gaspé synclinorium. The belt of domes extends southward from east central Vermont to Connecticut, analogous to, but more widely spaced than, the domes of the Bronson Hill anticlinorium east of the Connecticut River. The western belt of domes includes some that have only Paleozoic metamorphic rocks exposed in their central portions, in addition to those with the exposed Precambrian cores. The Guilford dome (Figure 1-2), with the Silurian Waits River Formation exposed in its central portion, is a conspicuous feature in the central part of the quadrangle. Large recumbent folds are also present in the strata mantling the domes of this western belt.

The Standing Pond Volcanics are an important marker unit in eastern Vermont (Doll et al., 1961), where they outline both recumbent folds and some of the later domes. The axial surfaces of the recumbent folds have been arched upward by the doming. The hook-shaped, closed, double band in the Standing Pond Volcanics surrounding the Guilford dome is such a feature.

Metamorphic highs to the staurolite-kyanite zone are associated with the Chester, Athens, Colrain and Guilford domes (Thompson and Norton, 1968; Doll et al., 1961). The grade of metamorphism decreases to the chlorite zone in a narrow belt along the Connecticut River in the eastern part of the quadrangle (Plate I). This belt is part of a regional metamorphic low that extends northward from the northern limit of the Triassic basins near Greenfield, Massachusetts, to northeastern Vermont along or near the Connecticut River Valley (Thompson and Norton, 1968). Eastward from this belt, the metamorphic grade increases sharply to highs in the sillimanite and sillimanite-K-feldspar zone along the Bronson Hill anticlinorium. The western part of the area is in the garnet zone of metamorphism.

Two important series of igneous rocks occur in the quadrangle. Gneisses of the Oliverian Plutonic Series form the cores of the domes in the Bronson Hill anticlinorium. There are also a number of small, late orogenic to post-orogenic granitic bodies assigned to the New Hampshire Plutonic Series (Billings, 1937). The largest of these in the Brattleboro area is the Black Mountain Granite.

Eastern Sequence-Western Sequence Division

As noted previously, the eastern portion of the Brattleboro quadrangle is on the west limb of the Bronson Hill anticlinorium; the central and western portions lie in the Connecticut River-Gaspé synclinorium and on the east limb of the Green Mountain anticlinorium, respectively. The stratigraphy mapped in these last two provinces forms a coherent package that can be correlated along strike into northern Vermont (Doll et al., 1961) and western Massachusetts. This stratigraphy is distinct from that mapped on the Bronson Hill anticlinorium, which has type localities in the Littleton-Moosilauke area of New Hampshire. The question as to the exact correlation of these two roughly contemporaneous stratigraphic sequences has been a long-standing problem. Although there is evidence for a correlation of these two stratigraphies, we feel that the divisions should be maintained. Thus, this bulletin has been divided into two separate parts, reflecting this division.

The Western Sequence or Western Terrane encompasses the geology on the east limb of the Green Mountain anticlinorium and the Connecticut River-Gaspé synclinorium west of the outcrop belt of the Littleton Formation (Plate I). The discussion of the geology of the Eastern Sequence or Eastern Terrane includes those areas east of the Putney Volcanics-Littleton Formation contact. The stratigraphy of the Eastern Sequence is that of the Bronson Hill anticlinorium and is essentially the same as Billings' (1956) "New Hampshire" sequence. The Western Sequence is similiar to Billings' (1956) "Vermont" sequence.

In recent years there has been much speculation about the placement of major paleo-plate tectonic boundaries through southeastern Vermont and the Brattleboro area. The recent COCORP seismic profiles through southeastern Vermont (see Ando et al., in press) may help to shed light on the major structural divisions in the area. However, the present authors have refrained from speculating on the broader implications of the geology of the Brattleboro area in this publication, but have tried to present an accurate account of the geology as presently exposed at the surface. This may, we hope, serve to aid, as well as constrain, future plate tectonic models for the area.

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CHAPTER 2

STRATIGRAPHY OF THE WESTERN SEQUENCE

West Flank of the Connecticut River-Gaspé Synclinorium and East Flank of the Green Mountain Anticlinorium

General Statement

The oldest rocks in the Brattleboro area are exposed at the southern end of the Athens dome in the northwestern corner of the quadrangle. Gneisses, probably Precambrian (Grenvillian) in age, are exposed in the core of the dome. These gneisses are mantled by a series of mapped units that are believed to range in age from the late Precambrian through Cambrian. West of the Brattleboro area these same units form a generally east-dipping sequence on the east flanks of the Ray Pond and Sadawga domes (Skehan and Hepburn, 1972).

The rocks of the outer-mantling sequence of the Athens dome are here correlated with the Tyson, Hoosac, Pinney Hollow, and Ottauquechee Formations as exposed further northwest between Windham and Sherburne, Vermont, on the east flank of the Green Mountain anticlinorium (Thompson, 1950; Rosenfeld, 1954; Chang et al., 1965). The Rowe Formation on the east flank of the Berkshires in Massachusetts is probably correlative with the Pinney Hollow, Ottauquechee, and Stowe Formations of Vermont. The Stowe Formation of the above areas cannot be identified on the Athens dome. The "Cavendish Formation" of Doll et al. (1961) is here abandoned inasmuch as all units in it can be reassigned to other units, mainly the Tyson and Hoosac Formations, as indicated by A. B. Thompson et al. (1977).

LITHOLOGY "WESTERN SEQUENCE"	FEET THICK-NESS	COLUMNAR SECTION	FORMATION	AGE
				L. DEV.
Spv: Buff to light brownish-gray feldspathic phyllite; thin beds of feldspathic granulite; light greenish-gray phyllite; minor gray slate.	0-400	Spv	Putney Volcanics	SILURIAN
Sgm: Interbedded light gray, impure quartzites and mica schist (western band). Gray, fine-grained *phyllite and slate with interbedded*, thin, micaceous quartzite and rare, punky-brown-weathering impure marble (eastern band). **Sgmm**: Black phyllite with interbeds of punky-brown-weathering impure marble.	2500-5000	Sgm Sgmm	Gile Mountain Formation	SILURIAN
Ssp: Dark gray to black, medium-grained amphibolite and epidote amphibolite predominant; very coarse grained garnet-hornblende fasciculite schist; rare, impure quartzite, coticule and schist (western bands). Gray to greenish, massive plagioclase-biotite-quartz and plagioclase-biotite-hornblende-quartz granulite (eastern band).	0-500	Ssp	Standing Pond Volcanics	SILURIAN
Swr: Dark gray mica schist and calcareous mica schist with abundant interbeds of punky-brown-weathering, impure marble; thin interbeds of impure quartzite toward eastern contact. **Swrq**: Gray feldspathic and micaceous quartzite interlayered with muscovite schist.	3000-7500	SWrq Swr	Waits River Formation	SILURIAN
Sn: Gray, commonly crinkled mica schist with conspicuous almandine porphyroblasts; minor impure quartzite and impure punky-brown-weathering marble beds.	1000-2500	Sn	Northfield Formation	SILURIAN
Srm: *Light gray to white quartz pebble conglomerate* and quartzite; coarse-grained hornblende fasciculite schist; medium-grained amphibolite.	0-20	Srm	Russell Mountain Fm.	SILURIAN
OSsa: Rusty-weathering black carbonaceous phyllite and schist; minor gray schist, black amphibolite, quartzite. **OSsaa**: Black medium-grained amphibolite and porphyritic amphibolite predominant.	2000-4000	OSsaa OSsa	Schist-Amphibolite Unit	SILURIAN / ORDOVICIAN
Obv: Dark gray to black, poorly to well-foliated amphibolite and porphyritic amphibolite; gray to light gray quartzo-feldspathic gneiss, schist, *granulite and layered gneiss*; minor black, rusty-weathering mica schist.	4000-8000	Ombv	Barnard Volcanic *Member* (Missisquoi Fm.)	ORDOVICIAN
Omm: Gray to light gray, interbedded quartzite, quartz-mica schist, mica schist and impure quartzite. Distinctive laminated "pinstriped" quartzite locally abundant; 10-20% medium-grained amphibolite and garnet amphibolite.	3000-5500	Omm	Moretown Member (Missisquoi Fm.)	ORDOVICIAN
Єo: Dark gray, graphitic mica schist with large garnet porphyroblasts; epidote amphibolite.	Thin around South End of Athens Dome	Єo	Ottauquechee Fm.	CAMB.
Єph: Light green quartz-muscovite-chlorite schist; *thin amphibolites*.		Єph	Pinney Hollow Fm.	CAMB.
pЄ-Єh: Mica schist with conspicuous albite porphyroblasts; epidote amphibolite. **pЄ-Єhs**: Silvery garnet-muscovite schist with staurolite porphyroblasts.		pЄ-Єh pЄ-Єhs	Hoosac Formation	PЄ or Є?
pЄ-Єbh: Augen gneiss, quartz-plagioclase-biotite gneiss.		pЄ-Єbh	Bull Hill Gneiss	PЄ or Є?
pЄm: Quartz-biotite-plagioclase gneiss.		pЄm	Mt. Holly Complex	PЄ

AGE	FORMATION	COLUMNAR SECTION	FEET THICK-NESS	LITHOLOGY "EASTERN SEQUENCE"
L. DEV.	Littleton Formation	Dl Dlc	5000-6000	**Dl**: Gray to dark gray slate, phyllite and mica schist with interbedded quartzite; biotite, garnet and staurolite porphyroblasts at the appropriate metamorphic grades. **Dlc**: Polymict conglomerate with fine-grained gray slate matrix; pebbles abundant to scarce.
SILURIAN	Fitch Formation	Sf	0-500	**Sf**: Purplish-brown biotitic granulite; greenish-gray calc-silicate granulite; minor quartz-mica schist and quartzite.
SILURIAN	*Clough Formation*	Scs Sc	0-450	**Sc**: Quartz pebble conglomerate and quartzite; quartz-mica schist with porphyroblasts of garnet or staurolite. **Scs**: Gray mica schist.
ORDOVICIAN	Partridge Formation	Opas Opa Op	200-3000	**Op**: Rusty-brown-weathering sulfidic mica schist; gray quartz-mica schist; biotite, garnet and staurolite porphyroblasts at appropriate metamorphic grades; amphibolite; light gray quartz-feldspar-biotite gneiss. **Opas**: Amphibolite predominates over mica schist. **Ops**: Mica schist; quartz-feldspar-biotite gneiss.
ORDOVICIAN	Ammonoosuc Volcanics	Oam	500-2000	**Oam**: Light gray quartz-feldspar-biotite gneiss; thick amphibolites; minor sulfidic mica schist near top.

Figure 2-1. Stratigraphic column for the Brattleboro quadrangle, showing correlations between the Western and Eastern Sequences.

A thick sequence of Siluro-Devonian rocks in the Connecticut River-Gaspé synclinorium underlies the central portion of the Brattleboro area. The majority of these rocks are metasediments in the Northfield, Waits River, and Gile Mountain Formations. Two thin intervals of meta-volcanics occur within this sequence (Figure 2-1): one near its middle (Standing Pond Volcanics), separating the Waits River and Gile Mountain Formations; and one forming the boundary with the Eastern Sequence near its top (Putney Volcanics), separating the Gile Mountain and Littleton Formations.

Following Billings et al. (1952), Ern (1963), and Chang et al. (1965), the Standing Pond Volcanics are designated as a separate formation. This formation forms an excellent marker unit, and its boundaries are considered to represent approximate time-statigraphic horizons. Although the Standing Pond Volcanics here separate the Waits River and Gile Mountain Formations by definition, it must be recognized that in the Brattleboro area the Standing Pond Volcanics does not everywhere exactly separate calcareous rock types (characteristic of the Waits River Formation) from the non-calcareous rocks (characteristic of the Gile Mountain Formation). Calcareous rocks in the Gile Mountain Formation similar to those in the Waits River Formation have been designated the marble member of the Gile Mountain on Plate I. The quartzitic member of the Waits River Formation contains schistose and feldspathic quartzites and mica schists, similar to those in the Gile Mountain Formation. Wherever the Standing Pond Volcanics are not present, as for example along parts of its eastern belt of outcrop, the boundary between the Waits River and Gile Mountain Formations is based on geometric extrapolation from known Standing Pond outcrops and does not everywhere separate units of contrasting rock type.

The metasediments and metavolcanics of the Western Sequence have been metamorphosed under conditions ranging from the lower greenschist to mid-amphibolite facies of regional metamorphism.

Igneous rocks in the area include a late synorogenic to post-orogenic stock and numerous smaller bodies of New Hampshire Series granite, a small ultramafic body in the Moretown Member of the Missisquoi Formation, and a small post-metamorphic mafic dike.

PRECAMBRIAN

Mount Holly Complex

A small area of gneiss, north of Newfane Hill is probably correlative with the gneisses of the Mount Holly Complex, more extensively exposed to the north in the Athens and Chester domes and in the central Green Mountains near Mount Holly, Vermont. Radiometric dating in these regions, as summarized by Naylor (1975), indicates Grenvillian ages of approximately 1,000 million years. The mineralogy of the gneisses exposed in Newfane is dominantly quartz-biotite-plagioclase with minor and sporadic muscovite, microcline, garnet, and epidote or clinozoisite.

LATE PRECAMBRIAN (?) OR CAMBRIAN (?)

Bull Hill Gneiss

The first unit in the mantling sequence of the Athens dome, as here mapped, is the Bull Hill Gneiss. The Bull Hill Gneiss is characterized by conspicuous augen or flaser of microcline that locally may be as much as three inches in diameter. The matrix contains quartz, biotite, and plagioclase, and locally muscovite, garnet, and epidote. Other rock types include well-layered quartz-plagioclase-biotite gneiss. Slabby gneisses containing a greenish phengitic muscovite are also common.

The Bull Hill Gneiss (Doll et al., 1961) forms part of the inner mantle sequence around much of the Athens and Chester domes. Similar augen gneisses crop out on the east flank of the Green Mountains in and near Wardsboro, Vermont, and in extensive areas northeast of Rutland (north of Sherburne Pass). Radiometric dating of similar rocks elsewhere (Naylor, 1975) suggests a possible late Precambrian or Cambrian age, but more work is clearly needed. The structural and stratigraphic setting of the Bull Hill Gneiss is still unsettled. Although similar gneisses have undoubtedly been formed by the shearing and cataclasis of preexisting granites, we suggest here that a plausible alternative interpretation is that these represent metamorphosed rhyolitic ignimbrites, possibly correlative with less metamorphosed felsites of late Precambrian or earliest Cambrian age in the central and southern Appalachians. If so, it would occupy the approximate stratigraphic position of the Tyson Formation as mapped by Chang et al. (1965) in the Woodstock area.

Hoosac Formation

The Bull Hill Gneiss is succeeded, outward from the core of the Athens dome, by rocks assigned to the Hoosac Formation. These consist mainly of albite-rich mica schists and gneisses. The albite forms conspicuous porphyroblasts or small augen in the schists. Biotite and garnet are additional porphyroblasts. The gneisses are characteristically layered with thin laminae of muscovite and biotite separating thicker layers of quartz and feldspar that generally make up 60% to 80% of the rock. Biotitic epidote amphibolite is a subordinate lithology in the Hoosac Formation.

A light gray to silvery garnet-muscovite schist (**Chs**, Plate I) occurs at the base of the Hoosac Formation northwest of Kenny Pond in Newfane. Many outcrops also contain abundant staurolite and a fine-grained paragonitic mica. Kyanite is also present locally, and some outcrops are graphitic. Near its contact with the Bull Hill Gneiss this schist contains quartzitic layers and in places has pits, possibly from the weathering of carbonate minerals. This schist has many mineralogical and textural features in common with the Gassetts Schist as described by A. B. Thompson et al. (1977).

CAMBRIAN

Pinney Hollow Formation

Light green quartz-muscovite-chlorite phyllites and schists are the principal metasedimentary rocks in the Pinney Hollow Formation. Garnet and magnetite porphyroblasts are usually present in these schists, and quartz lenses are abundant. Thinly laminated, ligniform epidote amphibolites are abundant throughout the Pinney Hollow. In its upper part, as exposed on the northwest ridge of Newfane Hill, these are the dominant rock type.

Ottauquechee Formation

The characteristic rock type of the Ottauquechee Formation is a dark gray to silvery mica schist that is typically graphitic and rich in pyrrhotite. Many outcrops are deeply weathered to a rusty, sulfate-stained rind. Large garnets, as much as an inch in diameter, are abundant locally and commonly contain sigmoid inclusion trails. Other rock types include thin-laminar micaceous granulite, minor epidote amphibolite, and, rarely, garnet-quartz granofels (coticule).

SYSTEM	Western Sequence Brattleboro area This Report	Southeastern Vermont Doll et al., 1961	Western Massachusetts Hatch, 1969 Hatch and Stanley, 1973	Emerson Western Mass. 1898, 1917
SILURIAN ?	Putney Volcanics		Waits River Formation	
	Gile Mt. Fm.	Gile Mt. Fm.		Leyden Argillite
	Standing Pond Volc.	Standing Pond Volc.		Conway Amphibolite
	Waits River Formation	Waits River Formation		Conway Schist
SILURIAN	Northfield Formation	Northfield Formation	Goshen Fm.	Goshen Fm.
	Russell Mountain Fm.	Shaw Mt. Fm.	Russell Mountain Fm.	
ORDOVICIAN	Unnamed schist - amphibolite	Missisquoi Fm.: Carboniferous Schist Mbr.	Hawley Fm.	Hawley Schist
	Missisquoi Fm.: Barnard Volc. Mbr.	Missisquoi Fm.: Barnard Volc. Mbr.		
	Missisquoi Fm.: Moretown Mbr.	Missisquoi Fm.: Moretown Mbr.	Moretown Formation	Savoy Schist
CAMBRIAN		Stowe Fm.	Rowe Schist	
	Ottauquechee Fm.	Ottauquechee Fm.		
	Pinney Hollow Fm.	Pinney Hollow Fm.		Chester Amphibolite
				Rowe Schist
P€ or CAMBRIAN ?	Hoosac Formation	Hoosac Formation	Hoosac Formation	Hoosac Schist
	Bull Hill Gneiss	Tyson–Cavendish		
PRECAMBRIAN	Mt. Holly Complex	Mt. Holly Complex		Becket Gneiss

Figure 2-2. Correlation chart showing the Brattleboro area and southeastern Vermont, after Doll et al., 1961, and adjoining areas of western Massachusetts.

ORDOVICIAN

Missisquoi Formation

General Statement.

The Missisquoi Formation of Ordovician age occupies a large tract in the western part of the Brattleboro area. Its eastern boundary is here drawn at the base of an unnamed series of schists and amphibolites formerly assigned to the Cram Hill Member of the Missisquoi (Hepburn, 1972A). This schist-amphibolite unit is now mapped separately. Recent work by J. L. Rosenfeld, and J. B. Thompson, Jr. in the adjoining Saxtons River quadrangle has brought into question the correlation of this unnamed unit with the Cram Hill Member. In the Brattleboro quadrangle, the Missisquoi is subdivided into two members: the lower Moretown Member, and the upper Barnard Volcanics.

The Missisquoi Formation is continuous from the Brattleboro quadrangle to extreme northern Vermont, where it was first defined as the Missisquoi Group by Richardson (1925, 1927) and Richardson and Cabeen (1923). The Missisquoi Formation is continuous in part with rocks mapped as the Savoy and Hawley Formations in western Massachusetts (Figure 2-2).

Moretown Member.

General Statement. The Moretown Member in the Brattleboro quadrangle is continuous with the Moretown Formation first described by Cady (1956) along the Mad River in Moretown, Vermont (Doll et al., 1961). It crops out over much of the western and northern portions of the quadrangle, where excellent exposures are found on most of the hills. The hills southwest of Branch School, Marlboro, and west of the Marlboro Branch of the Rock River have particularly good sections.

The upper contact of the Moretown Member with the Barnard Volcanics is gradational. The lower contract of the Moretown Member is relatively sharp at the south end of the Athens dome, where the quartz-rich rocks of the Moretown are in contact with the dark gray graphitic schists of the Ottauquechee Formation.

Rock Description. The Moretown Member is composed of distinctive quartz-rich rocks: schists, including mica schists and quartz-mica schists; quartzites; impure, micaceous and feldspathic quartzites; and laminated quartz-rich schists with micaceous partings. Quartz and muscovite are the most abundant minerals in these rocks, with the differences between the rock types due to variations in texture or in the absolute amounts of these minerals, plus the presence of biotite. Interlayers of amphibolite generally make up 10 to 20% of the unit.

The quartz-mica schists and mica schists make up the bulk of the member. They weather gray to light gray to tan and are somewhat lighter in color on fresh surfaces. Minor amounts of pyrite cause a slight rusty weathering on some of the schistose surfaces; but the light gray, quartz-rich nature of their fresh surfaces is distinctive. The quartz-mica schists tend to contain somewhat more feldspar in the upper part of the member, as the contact with the Barnard Volcanics is approached. Some of the schists in the Moretown with abundant, fine- to medium-grained muscovite have a silvery appearance. Almandine, biotite, and chlorite are common porphyroblasts in the schists.

A complete range of rocks exists between the mica schists and the quartzites, depending upon the percentage of quartz present. The mica-rich rocks and the quartz-rich rocks form both distinctive individual beds up to several feet thick and interbedded layers on a much smaller scale. Typically, light gray quartzites or quartz-rich schists are interlaminated with light to dark gray muscovite and biotite laminae on a scale of 2 mm to a few centimeters in thickness. Where the lamination is continuous and abundant on a fine scale, the rock has a distinctive "pinstriped" appearance. This texture is the most characteristic in the Moretown Member elsewhere (Skehan, 1961; Chang et al., 1965). However, in the Brattleboro quadrangle, this pinstriped texture is not as common as elsewhere and probably represents no more than 5 to 10 percent of the Moretown Member, although it is locally abundant. At least some of the thin interlayering of the mica bands, which produce the pinstriped texture, is of secondary origin. Figure 3-2 shows pinstriped texture that formed parallel to the axial planes of an early set of isoclinal folds.

The thicker quartzite beds and impure quartzites are fine- to medium-grained, composed mostly of quartz, and have subordinate amounts of feldspar, biotite, hornblende, or epidote. Small almandine porphyroblasts are common, though in some rocks they form mere skeletons between grains of quartz. As the amount of mica increases, the impure quartzites grade into quartz-rich schists and mica schists. The percentage of garnet, biotite, and chlorite porphyroblasts is higher in the schists, with garnets occasionally reaching up to 2 cm in size. Pyrite crystals up to 1.5 cm are a rare accessory.

Distinctive but minor rocks in the Moretown include chlorite-biotite fascicular schist and quartz-pebble conglomerate. Micaceous partings in some areas contain radiating, needle-like mixtures of chlorite and biotite (possibly after an orthoamphibole) up to 15 cms long. A typical outcrop of this rock may be seen in a highway cut on the north side of Vermont Rt. 9, ½ mile east of B.M. 1666 feet, Marlboro. Quartz-pebble conglomerate was found in one locality (Plate I). The pebbles are of vein quartz and range from 3 to 5 cm in size in a matrix of quartz-mica schist.

Small, light-green "pods" as much as 15 cms in length and consisting largely of actinolite and quartz were seen in several outcrops. It is believed that these represent original calcareous concretions.

Amphibolites generally make up 10 to 20 percent of the Moretown, except near the top of the member where they are more prominent. The amphibolites are black to dark green, medium-grained and contain hornblende, plagioclase and epidote, with or without biotite and garnet. They occur as distinct units a few inches to several feet thick. Garnet-bearing amphibolites are not as common, and they usually occur as thin units only a few inches thick. Non-porphyritic amphibolites within the Barnard Volcanics are lithologically identical with those of the Moretown. In only a few instances can the amphibolites be seen to cut bedding.

A rare coarse-grained hornblende-quartz-cummingtonite-garnet schist was found in one locality. The cummingtonite forms radiating crystals in the foliation plane and is intergrown with the hornblende and garnet.

Origin. The Moretown Member was originally deposited as a series of interbedded quartzose sands, argillaceous sands, graywacke sands, and muds. The amphibolites probably represent both intrusive dikes and sills and tuffaceous volcanic rocks.

Thickness. The base of the Moretown Member is exposed only in the northwestern part of the Brattleboro quadrangle, around the south end of the Athens dome, where it is tectonically thinned. South of this region, the full thickness of the Moretown is not exposed in the Brattleboro quadrangle; its base lies to the west in the Wilmington quadrangle (Skehan, 1961). The thickness of the Moretown is estimated from cross sections to be 3,000 to 5,500 feet.

Barnard Volcanic Member.

General Statement and Areal Distribution. Light- to dark-colored, felsic gneisses and amphibolites, occuring to the east of the Moretown Member, are included in the Barnard Volcanic Member (Doll et al., 1961). Richardson (1927) first mapped this unit as the Barnard Gneiss in Barnard, central Vermont. The Barnard Volcanic Member may be traced nearly continuously from the Brattleboro quadrangle to Richardson's type locality (Doll et al., 1961).

The Barnard Volcanic Member occurs in a band 1 to 1½ miles wide that strikes northeasterly through the towns of Halifax, Marlboro, and Newfane, and also in an elliptical area in the northwestern part of the town of Dummerston (Plate I) where it is exposed in the core of the doubly-plunging Fall Brook anticline.

The contact of the Barnard with the Moretown Member is gradational. Where interlayered quartz-feldspar rocks are recognizably feldspathic in the field, they have been included in the Barnard. There is thus a transition zone within the lower part of the Barnard Member where quartz-rich, Moretown-like rocks are interbedded with feldspathic quartz-mica schists and mica-feldspar schists. This transition zone ranges from a few tens of feet wide in southern and south-central Marlboro to as much as 800 feet wide in northern Marlboro and the southern part of Newfane. Directly above this zone typical Barnard occurs as light-colored, feldspar-rich gneisses, granulites, and amphibolites. Although garnets are present in the Barnard Volcanics, a rapid decrease in their overall abundance is noted at the contact compared to the Moretown.

Amphibolites increase in abundance in the upper Moretown near the contact, where they may comprise as much as twenty percent of the exposed rock. They continue to increase through the transition zone.

Rock Description. The Barnard Volcanic Member includes a wide variety of rocks, but three general types are most abundant: non-porphyritic amphibolite; porphyritic amphibolite; and light colored, felsic schist and gneiss. The felsic rocks make up 35 to 50 percent of the member but are generally less well exposed than the two varieties of amphibolite, which occur about equally. Felsic rocks are somewhat more common near the base of the member. Minor amounts of mica schist and phyllite are included in the felsic lithology. Amphibolites are interlayered with the felsic rocks on all scales. In the layered gneisses, the interlayering may be only a few millimeters thick. Commonly the amphibolites are several feet

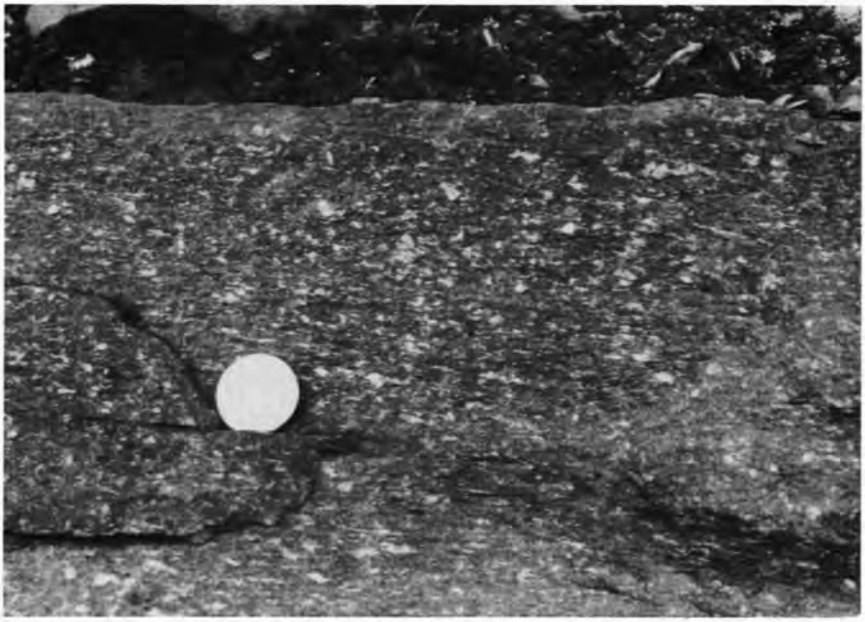

Figure 2-3. Porphyritic amphibolite in the Barnard Volcanic Member with porphyroblasts of plagioclase feldspar. Quarter for scale.

thick. Interlayering on a larger scale of several tens to hundreds of feet is also typical in the Barnard Volcanics.

Non-porphyritic amphibolites are black to dark green, fine- to medium-grained, generally massive rocks composed mostly of hornblende, plagioclase, quartz, and epidote. Some fine-grained amphibolites are well foliated. Thin layers of feldspar up to a few millimeters thick are locally present. Where these layers are abundant the rocks grade into hornblende-gneisses and are distinct from the more massive amphibolites. Biotite, chlorite, and ankerite are common accessory minerals in the amphibolites. The ankerite weathers out to form distinctive, small, brownish pits on weathered surfaces.

Porphyritic amphibolite (Figure 2-3) is distinguished from the non-porphyritic type by the presence of numerous, white to buff-colored feldspar megacrysts. These give the rock a distinctive mottled appearance. The porphyritic amphibolities weather a lighter gray than the non-porphyritic varieties but are similar in their other characteristics and overall mineralogy. In numerous places these two types of amphibolites are interlayered on the scale of an outcrop (for example, in the bed of the Rock River, by B. M. 521′ in Williamsville). The contacts between these two types of amphibolite may be sharp to gradational.

The felsic gneisses, schists, and layered gneisses (Figure 2-4) of the member include a wide variety of rocks formed by varying the amounts of the common minerals: feldspar, quartz, biotite, hornblende, muscovite, and chlorite. The color of these rocks ranges from nearly white or tan in the rocks rich in feldspar and quartz, to gray or greenish as the percentages of biotite, hornblende, and chlorite increase. Likewise, rocks composed of nearly pure feldspar and quartz are granulose in texture and poorly foliated. Those richer in the micas and chlorite are schistose. Gneisses are formed where the quartz-feldspar fraction of the rock is interlayered on a scale of a few millimeters, with thinner layers of the other minerals. These gneisses represent the most common felsic rocks.

In addition to the major minerals previously listed, small almandine crystals commonly occur in some of the felsic rocks. Pyrite to 5 mm, magnetite, and ilmenite are also minor accessory minerals.

A thin, metatuff layer with small rounded or elongated quartz grains has been noted at four localities (Plate I). Skehan (1961) noted silimar outcrops in the Barnard Gneiss of the Wilmington area.

Fine-grained, silvery-green sericite-chlorite schist is also a distinctive but uncommon rock in the Barnard.

Minor amounts of dark gray to black, fine-grained, rusty-weathering phyllite, similar to that of the overlaying unit, occur scattered throughout the Barnard. Similarly, mica schist resembling that found in the underlying Moretown Member is found interlayered with feldspar-quartz granulites near the base of the member.

Origin. The gneisses and amphibolites of the Barnard Member represent original volcanic and weathered volcanic deposits. The amphibolites have a chemical composition which indicates they were initially basaltic (Table 2-1). Some of the "megacrysts" in the

Figure 2-4. Layered gneiss in the Barnard Volcanics in the bed of Fall Brook, Dummerston. Hammer for scale. Dark layers consisting of hornblende alternate with light-colored layers of quartz and feldspar.

porphyritic amphibolites are undoubtedly relict phenocrysts. The more massive amphibolites may represent intrusives or flows. No discordance of the amphibolites with the surrounding bedding was found, which supports the latter interpretation. The felsic gneisses have chemical similarities to dacites or rhyodacites, (Chang et al., 1965). Relicts of plagioclase and quartz phenocrysts are common in the felsic granulites.

Thickness. The Barnard Volcanic Member is estimated to be 4,000 to 8,000 feet thick in the Brattleboro quadrangle.

Sample Number	#888	#B66-6A
SiO_2	49.19	49.25
TiO_2	1.56	1.57
Al_2O_3	18.46	17.79
Fe_2O_3*	10.05	10.25
MnO	0.18	0.22
MgO	6.52	6.31
CaO	10.73	10.92
Na_2O	2.67	2.97
K_2O	0.26	0.42
P_2O_5	0.17	0.19
Total	99.79	99.89
C.I.P.W. NORM		
OR	1.54	2.48
ALBITE	22.59	25.13
ANOR	37.61	33.96
DIOP	11.88	15.56
HYPER	15.46	7.73
OL	5.00	9.19
MT	1.44	1.48
IL	2.96	2.98
AP	0.41	0.45

Table 2-1. Representative major element analyses and C.I.P.W. norms for amphibolites from the Barnard Volcanics and the schist-amphibolite unit. Oxides except Na_2O by X-ray fluorescence, laboratory of Prof. M. Rhodes, Univ. of Massachusetts, Amherst. Total iron as Fe_2O_3*. Na_2O by Instrumental Neutron Activation Analysis, Dept. of Geology and Geophysics, Boston College. C.I.P.W. norm calculation assumes $FeO/Fe_2O_3 = 0.9$. Sample #888, porphyritic amphibolite with feldspar megacrysts, south bank of West River, N.30°E., 0.9 mile from B.M. 386′ north of West Dummerston. Sample #66-6A, massive amphibolite with feldspar porphyroblasts, roadcut, Green River Rd., 0.2 mile west of Reid Hollow. Halifax.

ORDOVICIAN or SILURIAN

Unnamed Schist-Amphibolite Unit

General Statement. A unit of rusty-weathering, black carbonaceous phyllite and gray mica schist with interlayered amphibolites lies between the Barnard Volcanics and the Russell Mountain Formation. Currier and Jahns (1941) first described a rusty phyllite in this stratigraphic position as the Cram Hill Formation in central Vermont. Doll et al. (1961) included the Cram Hill as a member in the Missisquoi Formation but indicated that this unit is missing in the area north of the Chester dome. Thus, the Cram Hill cannot be traced directly from its type locality into the Brattleboro area. Because of this, Doll et al. (1961) made these rocks an informal member of the Missisquoi Formation in southeastern Vermont but correlated them with the Cram Hill Member, as did Hepburn (1972A, 1972B) and Skehan and Hepburn (1972).

Recent work by J. B. Thompson, Jr. and J. L. Rosenfeld in the Saxtons River and Claremont quadrangles (north of the Brattleboro area) has shown that a unit of orthoquartzite, quartz-pebble conglomerate and gray schist occurs just above the top of the Barnard Volcanics, a stratigraphic position well below the base of the lithologically similar Shaw Mountain Formation. Thin gray schists, believed to be continuous with this new unit on Putney and Windmill Mountains in the Saxtons River quadrangle, have been traced into the schist-amphibolite unit in the northern part of the Brattleboro quadrangle. However, as yet, no quartzite or quartz-pebble conglomerate has been found associated with these gray schists in the Brattleboro quadrangle.

The regional correlations of this stratigraphically lower quartzite unit are not yet clear, and there is a question as to whether it is this unit or the Shaw Mountain Formation which represents the basal Silurian above the Taconian unconformity. For these reasons, it was felt that the schist-amphibolite unit should remain informally named and not be correlated with the Cram Hill member of the Missisquoi Formation until the regional correlations and age are better known.

The schist-amphibolite unit occurs in a northeasterly striking band, one to two miles wide, east of the Barnard Volcanics (Plate I). It is overlain by the Russell Mountain and Northfield Formations. A zone in which amphibolite is the predominant rock occurs along part of the eastern boundary of this unit and has been mapped separately (Plate I, **OSsaa**).

The contact between the schist-amphibolite unit and the overlying Russell Mountain and Northfield Formations is sharp and likely represents an unconformity. The lower contact is also relatively sharp and probably also is an unconformity. The characteristic amphibolites, gneisses and schists of the Barnard grade upward into the schists and phyllites of this unit over an interval of a few tens of feet. However minor amounts of rusty-weathering schist and phyllite also occur within the Barnard. The mapped contact of the schist-amphibolite unit indicates where these schists become the predominant rock.

An excellent section of the schist-amphibolite unit may be seen along the Rock River, east of the village of Williamsville, town of Newfane, where the upper contact with the overlying Northfield Formation is exposed.

Rock Description. Fine-grained, well-foliated to fissile carbonaceous black phyllite or schist makes up most of the unit. Finely disseminated sulfides, generally pyrite or pyrrhotite, cause the characteristic rusty-brown weathering. In a few places, the sulfides are so abundant that secondary, yellowish, iron sulphates are formed as efflorescence on outcrop surfaces. Where this rock is not stained brown, it weathers to a sooty-black or dark gray. Biotite, pyrite, and to a lesser extent, garnet, are the common porphyroblastic minerals large enough to be distinguished megascopically in the field. Sericite, formed along foliation surfaces, may give the black phyllites a notable sheen. Numerous small crinkle folds are common in the phyllites.

Thin, black, fine-grained quartzite beds, 5-10 cm thick, are common. Where these quartzite beds are not found, bedding is difficult to distinguish in the phyllites.

Amphibolites are black to dark gray-green, medium-grained, massive to moderately foliated rocks similar to those found in the Barnard Volcanics. Both porphyritic and non-porphyritic varieties are common and form layers, commonly boudinaged, two to tens of feet thick, bounded by phyllites, The amphibolites account for 10 to 20 percent of the member, except near the top where they form 70 to 90 percent of the rock zone shown as **OSsaa** on Plate I. Felsic schist and black phyllite make up the remainder of this thin zone.

Gray mica schist, layered feldspar-quartz schist, and coticule are present in minor amounts in the schist-amphibolite unit. The gray mica schists are medium-grained, commonly contain garnet, and are distinctly gray in contrast to the dark phyllites. These schists generally form thin layers in the phyllites. However, one of the thicker layers of gray schist, some 20 to 100 feet thick in the northern part of the Brattleboro quadrangle, is believed to be continuous with the gray schists associated with quartzites and quartz-pebble conglomerate on the west slope of Windmill Mountain in the Saxtons River quadrangle. The layered schists are composed of thin layers of fine-grained, granular feldspar and quartz, separated by thinner laminae of muscovite with subordinate biotite and chlorite.

Coticule is an early French name used by Omalius-d'Halloy in 1808 for a granular rock composed of fine-grained garnets, rich in the spessartite component, quartz and muscovite. This distinctive rock is locally abundant in thin layers in the phyllite. The layers, usually no more than a few centimeters thick, are a distinctive light pink color but may weather black from manganese staining. Typically, these layers are highly folded on a small scale. Coticule layers are particularly abundant in some of the glacially polished outcrops on the north side of the 1500 foot hill, ½ mile northeast of the three-corner junction of the town lines of Dummerston, Newfane, and Putney.

Origin and Thickness. The schist-amphibolite unit originated as a series of black sulfidic muds and interlayered mafic volcanics. The coticule may represent original manganiferous chert layers.

The thickness of this unit in the Brattleboro quadrangle is estimated to range from 2,000 feet to 4,000 feet.

SILURIAN

Russell Mountain Formation

General Statement and Areal Distribution. Quartzite, quartz-pebble conglomerate and associated rocks of the Russell Mountain Formation crop out in three thin, elongated lenses, in the Brattleboro quadrangle (Plate I). The formation was named by Hatch et al. (1970) for exposures of quartzite and calc-silicate granulate occurring between the Hawley and Goshen Formations on Russell Mountain in the Woronoco quadrangle, western Massachusetts. Previously these rocks in the Brattleboro quadrangle were included in the Shaw Mountain Formation of Silurian age by Hepburn (1972A). Rosenfeld (1954) and Doll et al. (1961) likewise correlated more extensive exposures of these rocks just north of the Brattleboro quadrangle on Putney and Windmill Mountains with the Shaw Mountain Formation, whose type locality is near Northfield, Vermont. However, as previously noted, the recent discovery of a second quartzite unit in the stratigraphic section below this horizon has brought into question this correlation. While it is still possible that these rocks in the Brattleboro quadrangle could correlate with the Shaw Mountain Formation, there is little doubt that they do correlate with the Russell Mountain Formation, which occurs at the same stratigraphic horizon. Thus, this latter name is preferred.

The quartzites and quartz-pebble conglomerates of the Russell Mountain Formation form a very sharp contact with the underlying schist-amphibolite unit.

Rock Description. White- to light brown-weathering, fine-grained quartzite and quartz-pebble conglomerate form the most conspicuous part of the formation. The conglomeratic quartzites are restricted to the lower few feet. The stratigraphically higher quartzites tend to contain more mica, and the formation grades upward into quartz-mica schists. The pebbles in the conglomerate are of clear vein-type quartz. They are small, generally less than 2 cm across (although a few to 10 cm in diameter were noted). The pebbles are now ellipsoidal and stretched in the plane of foliation, the long axes plunging 60 to 80 degres north to northeast.

Distinctive, coarse-grained hornblende fasciculite schist is found in thin beds above the basal quartzites and quartz conglomerates. In places, it is interbedded with the quartzites and quartz-muscovite schists. The hornblende-mica schists weather gray to tan and commonly part along the schistosity, leaving hornblende sprays prominently exposed. Individual crystals of hornblende are up to several inches long.

Fine- to medium-grained, black to dark green, massive amphibolites, garnet amphibolites, and hornblende schists are present near the top of the formation, particularly in the northern lens. These resemble rocks found in the underlying schist-amphibolite unit and the Barnard Volcanics. The amphibolites are commonly boudinaged, with boudin necks typically filled with coarse, black crystals of tourmaline.

It is suspected that the amphibolites and hornblende schists found above the basal quartzites may be more widespread than the quartzites but could not be distinguished with certainty from the underlying schist-amphibolite unit. Thus, the Russell Mountain Formation has not been mapped in the Brattleboro quadrangle where the distinctive quartzites and conglomerates are not present.

In contrast to the northern lens, the uppermost portion of the formation in its southern occurrence is a quartz-muscovite-biotite granulite or schist. Similar quartz-mica schists, weathering gray to light brown, are also interbedded with the quartzites in the middle of the formation in the southern occurrence.

Below is given a typical section seen in the southernmost of the three lenses. The middle lens is similar to this, but quartzite is less common.

Schists of the Northfield Formation

5′ to 10′ covered

0.5′	Quartzite, quartz-muscovite granulite and schist
2.5′	Coarse amphibolite (boudinaged)
1.5′	Quartzite
1′	Coarse hornblende fasciculite schist
2′-4′	Quartzite and quartz conglomerate

Hornblende amphibolite, gneiss of Schist-Amphibolite Unit

Origin. The Russell Mountain Formation originated as a shelf type deposit of the transgressing Silurian seas. The quartz sands and conglomerates formed as a beach or in a near-shore marine environment. The pebbles probably are locally derived. The amphibolites represent pyroclastic volcanics or volcanic material washed into the deposit. Irregular initial deposition, tectonic events, or a post-Russell Mountain period of erosion may explain the lensoid and discontinuous nature of the formation.

Thickness. The southernmost of the three lenses of Russell Mountain is the largest in the Brattleboro quadrangle. It is 500 feet in length and has a maximum exposed thickness of approximately 20 feet.

Northfield Formation

General Statement. A sequence of gray mica schists is mapped as the Northfield Formation of Silurian age. This formation is continuous with similar rocks in the Montpelier area named the Northfield Slate by Currier and Jahns (1941). Richardson and Cabeen (1923) and Richardson (1927, 1929) called these same rocks the Randolph Phyllite. The Northfield Formation is continuous with the Goshen Formation of western Massachusetts.

The Northfield extends as a north-northeasterly band through the western portions of the quadrangle. It is a ridge-former, and excellent exposures are seen on the glacially polished surfaces of many hilltops in this area, particularly Dummerston Hill, Brattleboro, and Bear Hill, Halifax.

The contact at the base of the Northfield is sharp. The gray mica schists of the Northfield usually rest upon the amphibolites or dark phyllites of the schist-amphibolite unit. No amphibolites have been observed in the Northfield. The gray non-rusty color, the abundance of garnets, and the numerous quartz lenses in the Northfield Formation distinguished it from the rusty weathering phyllites below, in which neither garnets nor quartz lenses are particularly abundant.

From the map pattern of the amphibolite-rich zone at the top of the schist-amphibolite unit, (**OSsaa**) it could be interpreted that an unconformity has truncated this zone in the area south of the Rock River, Newfane. The outcrop in the area, however, is not sufficient to prove whether this interpretation or a facies change along strike is responsible for the disappearance of the amphibolite-rich zone.

The lower contact with the Russell Mountain is commonly covered and usually forms a small depression, several feet wide. Where seen, the contact is sharp and appears conformable, with schists of the Northfield Formation resting on the amphibolites and hornblende-mica schists of the upper Russell Mountain Formation.

Rock Description. The Northfield Formation in the Brattleboro quadrangle is a homogeneous gray to dark gray, graphitic muscovite-quartz-almandine schist. The ratio of quartz and muscovite vary considerably, but together they form a constant 70 to 80 percent of these rocks. Almandine garnets are generally very conspicuous, forming porphyroblasts 1 to 10 mm in diameter. Biotite is also commonly present. Staurolite is common in the areas northwest of Richardson Mountain, Brattleboro, and in western Guilford (Plate I). Chlorite occurs as porphyroblasts that cut across the foliation, as alteration products on garnets, and as rare pseudomorphs after staurolite.

In outcrop the schists weather dark gray to somewhat rusty-brown. Quartz pods and lenses are common. The schists are typically highly crinkled on a small scale, with the crinkles grading into a slip cleavage where the micas have been realigned parallel to the axial planes of the folds (Figure 3-13). Generally the wavelength of these microfolds is between 1 and 4 mm.

Locally abundant, particularly in the lower portion of the Northfield, are thin beds of impure, micaceous quartzite and quartz-rich, quartz-mica schist (Figure 3-10). These range in thickness from an inch to several feet, but generally average a few inches thick. Rarely some of these beds show graded bedding. This is in marked contrast to their equivalent unit along strike in Massachusetts, the

Goshen Formation, in which graded beds are extensively developed (Hatch, 1968, 1975).

Scattered thin impure marble beds occur in the Northfield. They are similar to those making up a large percentage of the overlying Waits River Formation. They are typically thin, 6 to 8 inches thick, and weather a distinctive punky brown color. This weathering is due to a rind of iron oxides, quartz, and micaceous minerals left after the carbonate has been leached. The impure marbles are fine-grained and are gray to bluish gray on fresh surfaces. Quartz and calcite make up 70 to 80 percent of these beds, with muscovite and biotite common. Almandine and zoisite are common accessory minerals.

Origin. The Northfield was most likely deposited as a series of deep water muds with a few argillaceous sand and arenaceous lime interbeds. The graded arenaceous bedded units indicate that at least some of the formation was likely deposited by turbidity currents.

Thickness. The general outcrop width of the Northfield is about 7,500 feet. However, in the area east of the Fall Brook anticline (Figure 3-1), it is only about 1,000 feet wide. This fold and the adjacent syncline to the west undoubtedly account for a considerable broadening of the Northfield belt in the Brattleboro quadrangle. Thus, even if the area near the Fall Brook fold is tectonically thinned, it is believed that this area may be more representative of the true thickness of the Northfield Formation than elsewhere. Numerous isoclinal folds are present in the Northfield (Hatch, 1968). Therefore, it is estimated that the thickness of the Northfield Formation in the Brattleboro area ranges from 1,000 to 2,500 feet. Doll et al. (1961) show that the Northfield generally has a narrower outcrop width in the rest of Vermont than it does in the Brattleboro quadrangle.

Waits River Formation

General Statement. A sequence of mica schists, impure marbles, and impure quartzites is correlated with the Waits River Formation of Currier and Jahns (1941). Richardson (1906) originally designated this sequence the Waits River Limestone. This formation is exposed over a large area in the central portion of the Brattleboro quadrangle (Plate I) and may be traced continuously to its type locally in central Vermont.

The Waits River Formation is not a prominent ridge-former, as the impure marbles and calcareous schists are easily weathered. It underlies many of the best remaining farms in the area.

A unit in which interbedded feldspathic quartzites and mica schists predominate, with little or no impure marble present, has been mapped separately (Plate I) as the quartzitic member of the Waits River Formation (**Swrq**). These rocks are not greatly different from those in the overlying Gile Mountain Formation. However, they occur below the Standing Pond Volcanics and are therefore included in the Waits River Formation.

Well-exposed sequences of the Waits River Formation are not common, but the sections along the Green River in Guilford and Halifax and along the north bank of the West River, Brattleboro are representative.

Rock Description. Three broad categories of rocks are interbedded in the Waits River Formation: impure marbles, schists, and impure quartzites. Of these, the most distinctive and characteristic are the impure marbles. These rocks weather to a punky light to dark brown and have a friable rind of non-calcareous minerals, largely quartz, left behind when the carbonate has weathered out. This rind may be up to a few inches thick. Examination of the rind shows that the marble is commonly very impure and may contain 40 percent or more quartz and biotite. The marble on a fresh surface is a steel-gray, hard, fairly massive, fine-grained granulite. Garnets and small "clumps" of actinolite or tremolite, with chlorite, form porphyroblasts in the higher metamorphic grades. These minerals protrude from the weathered surface and give the rock a "knobby" appearance. Quartz and calcite or ankerite make up 60 to 90 percent of the marble, but the individual percentages of these minerals differ considerably from bed to bed. Light brown biotite and muscovite are commonly present and give these rocks a weak foliation.

The impure marble beds range in thickness from a few inches to tens of feet. They may show extensive folding. Individual beds of the marble are not traceable over distances of more than a couple hundred feet. Whether this is due to lack of outcrop or to lensing of the beds is unknown. The percentage of impure marble beds varies greatly throughout the formation. Locally, marble beds may make up greater than 50 percent of the outcrop, particularly where these beds are 10 to 20 feet thick. The area in the center of the Guilford dome is particularly rich in marble beds. The percentage of marble beds as a whole is generally much less than this; and, on the average they represent some 10 to 20 percent of the formation. The marbles do not crop out well, and thus sections taken a few hundred feet apart may show quite different percentages of the impure marbles. The Waits River Formation is not defined on the basis of these impure marbles alone, but on their presence in the sequences with other rock types.

The schistose rocks in the Waits River Formation are varied. In the westernmost outcrops of the formation, the schists are gray muscovite-quartz-almandine rocks similar to those of the underlying Northfield Formation. Except in this area, the pelitic rocks in the Waits River are generally irregular, rusty-brown to dark gray-weathering, black to dark gray, carbonaceous schists and phyllites. Muscovite, quartz and biotite are the principal minerals, with a few almandine porphyroblasts usually present.

The schists have varying amounts of carbonate, which give them a pitted and friable appearance on weathered surfaces. Fine-grained pyrite (and perhaps pyrrhotite) is common as an accessory and undoubtedly causes the rusty-brown weathering color of the schists where present. Zoisite in dark, tabular crystals to several inches in length is common in some of the calcareous schists, particularly in the area near the Standing Pond contacts in the staurolite-kyanite zone. Large, light-colored clinozoisite crystals up to 12 inches in length are found associated with quartz lenses throughout the formation. Hornblende crystals and pseudomorphs after hornblende are common as radiating sheaves along the schistosity surfaces in the calcareous schists. Kyanite and staurolite have been found in only a few of the schists of the Waits River Formation.

Thin beds of impure quartzite are present throughout the formation. They increase in size and abundance toward the Standing Pond Volcanics in both the central part of the area and toward the eastern contact of the Waits River. They are developed to the exclusion of the other rock types in a narrow band adjacent to the Standing Pond Volcanics and have there been subdivided into a separate member (see below). These fine- to medium-grained, feldspathic and micaceous quartzites weather light gray and occur in beds from a few inches to two feet thick. Fine-grained muscovite usually forms a weak schistosity parallel to bedding in these rocks. Transitional rock types between these beds and the impure marbles are locally present.

Distribution of Rocks within the Waits River. Well-developed, impure marbles in black to gray, brown-weathering schists, with few impure quartzite beds, mark the Waits River in the center of the Guilford dome and in the area just outside the contact of the quartzitic member. Almost everywhere else, the overall percentage of the pelitic rocks increases relative to the impure marbles. However, near the eastern contact of the Waits River Formation, the number of thin quartzite beds increases, but they do not account for more than 10 to 20% of any sequence of rocks. Impure marbles also make up about 10-20% of the eastern exposures of the Waits River Formation.

Contact Metamorphosed Rocks. Within a few hundred yards of the Black Mountain Granite, the Waits River Formation has been subjected to contact metamorphism in addition to the regional metamorphism (Plate I). In this area the impure marbles become light green calc-silicate granulites, with extensive development of actinolite, clinozoisite, and diopside. Large, dark green hornblendes are conspicuous in the calcareous schist layers. In general, the schists become brittle and acquire many of the characteristics of a hornfels, but the grain size remains unchanged (i.e., medium-grained). The rocks tend to weather dark gray in this zone in contrast to the rusty-brown weathering seen elsewhere.

Gradational Contact with the Northfield Formation. In the central portion of the quadrangle, the transition between the Waits River and Northfield Formations occurs over an interval of, at most, a few hundred feet. South of the Green River and north of approximately the Putney town line, the transition zone is much broader. In the western part of the transition zone, punky-weathering, fairly thick, impure marbles appear in schists typical of the Northfield. These impure marble beds may be several feet thick but occur very sporadically. Invariably the topography reflects their presence, as they weather easily and form small valleys in the otherwise resistant schists of the Northfield. Several hundred feet of gray schist typical of the Northfield may intervene between these impure marble beds. The contact of the Northfield and the Waits River is drawn where the percentage of impure marbles becomes greater than 1 percent of the exposed rock. This places most of the transition zone within the Waits River Formation. Typically, the contact drawn on this occurs to the west of the area where the schists themselves change character. A particularly extensive sequence, in which the impure marbles amount to 1 to 5% of the rock in an otherwise typical Northfield-like schist, has been separated on Plate I in the southern portion of the area. Progressing eastward across the transition zone, the impure marbles become more numerous. Following this, the rocks enclosing the impure marbles change rather abruptly from the uniformly gray schists of the Northfield Formation to the highly non-uniform calcareous schists of the Waits River Formation.

Quartzitic Member, Waits River Formation. The quartzitic member occurs as a narrow belt adjacent to the outer loop of the Standing Pond Volcanics on the Guilford dome (Plate I). This member consists of feldspathic and micaceous quartzites interlayered with thinner beds of muscovite schist. The quartzites are similar to those described above in the Waits River Formation but occur in thicker beds here, some reaching 10 feet. They weather light gray to tan and are fine- to medium-grained with a granulitic texture. Biotite, almandine, chlorite, and carbonate occur in minor amounts, with the carbonate weathering to small, brownish spots. The mica schists are gray to dark gray, medium-grained muscovite-quartz-biotite schists with or without almandine, chlorite, staurolite, and kyanite. There are essentially no impure marble beds in this member and only small amounts of the rusty brown-weathering, black schists typical of the rest of the Waits River Formation. This member is very similar to the Gile Mountain Formation except for the greater percentage of plagioclase in the impure quartzites. This member would not be easily distinguishable in the field from the Gile Mountian were it not for the intervening Standing Pond Volcanics.

The contact of this member with the rest of the schists and impure marbles of the Waits River is quite definite. In a few tens of feet, the rocks change from the common schists of the Waits River with a relatively high percentage of impure marbles to rocks consisting of quartzite and interbedded muscovite schist with essentially no impure marble beds.

Although the quartzite member resembles the Gile Mountain Formation, there are differences. The bedding in the quartzitic member is generally not as prominent as that in the Gile Mountain Formation, although locally it may be well developed. The amount of feldspar and biotite in the quartzitic member is generally higher, and the schists appear somewhat darker in color and are less aluminous. Small garnetiferous amphibolite beds are present in the quartzitic member while none have been mapped in the Gile Mountain Formation. These amphibolite beds have a maximum thickness of 10 feet but are usually on the order of 2 feet thick. No more than 3 of these amphibolites have been encountered in any one crossing of the member. They were not observed to cut bedding.

The amphibolities are dark gray to black and are generally coarse-grained to fasciculitic, and almandine may make up 15 percent of these rocks. The amphibolites are not associated with the thicker, fine-grained, continuous amphibolites of the Standing Pond and are separated from them by hundreds of feet of impure quartzite and mica schist.

Excellent exposures of this member occur on Bear Hill, Brattleboro.

Origin. The Waits River Formation formed as a series of marine arenaceous limestones, impure sandstones, and black calcareous shales. These probably accumulated as a deep-water, base-of-slope deposit of turbidite-related origin (J. Rehmer, Pers. Comm. 1983). The thin amphibolites and the relative abundance of plagioclase in the quartzitic member may indicate a possible volcanic or reworked volcanic origin for this unit, which could imply close ties between it and the adjacent Standing Pond Volcanics.

Thickness. The thickness of the Waits River Formation is difficult to estimate. At its maximum outcrop width, it is five miles across; at its minimum, no more than 1,000 to 2,000 feet. To know how much these latter figures represent tectonic thinning and how extensively folded the Waits River is in its wider outcrop belts is a problem. The dips in much of the central part of the area (Guilford dome) are shallow. The impure marble beds invariably show extensive isoclinal folding. Therefore, it is believed that a reasonable estimate of the original thickness of the Waits River Formation in the Brattleboro quadrangle is between 3,000 and 7,500 feet.

Standing Pond Volcanics

General Statement. Doll (1944) introduced the name Standing Pond Amphibolite for amphibolites in the Memphremagog, now Waits River Formation, in the Strafford quadrangle, Vermont. Following Billings et al. (1952), Ern (1963), and Chang et al. (1965), the Standing Pond Volcanics are here elevated to full formation status. Doll et al. (1961) show that the eastern outcrop belt (band #3, Figure 2-5) of the Standing Pond Volcanics in the Brattleboro quadrangle is in the same strike belt as the type locality but cannot be traced continuously to the Strafford quadrangle.

The three bands of the Standing Pond Volcanics in the Brattleboro quadrangle have been divided into two distinctive belts. A thin, north-south striking band, (band #3, Figure 2-5), running discontinuously through eastern portions of the towns of Guilford, Brattleboro, Dummerston, and Putney, will be discussed as the eastern belt. The western belt of the Standing Pond Volcanics

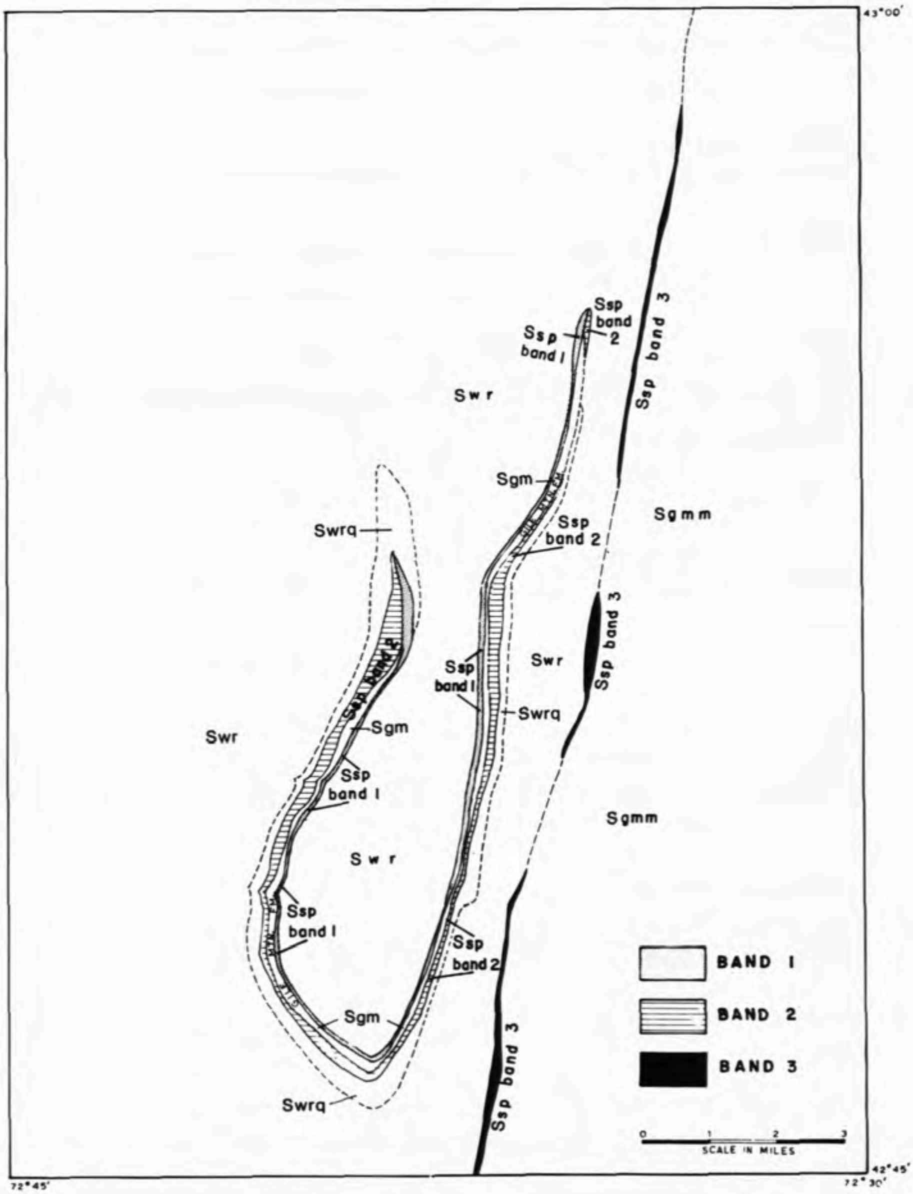

Figure 2-5. Outline map of the Standing Pond Volcanics, Brattleboro quadrangle, Vt. Bands #1 and #2 in the arcuate shaped loop are referred to in the text as the western belt. Band #3 is called the eastern belt. Ssp = Standing Pond Volc., Swr = Waits River Fm., Swrq = Waits River Fm., quartzitic mbr., Sgm = Gile Mt. Fm., Sgmm = Gile Mt. Fm., marble mbr.

forms a double band, closed at both ends, due to multiple folding in the central portion of the quadrangle (Plate I). This belt is subdivided into an inner band (band #1) and an outer band (band #2, Figure 2-5). These two bands are separated by the Gile Mountain Formation. A small, isolated outcrop of amphibolite and coarse-grained garnet amphibolite like those in the Standing Pond occurs on a small hill crossed by the Brattleboro-Guilford town line.

The western belt and the Gile Mountain Formation, which separates the amphibolite bands, are prominent ridge-formers and generally are well exposed. Many of the hills held up by these formations have dip-slopes formed by the Standing Pond Volcanics (for example, Richardson and Round Mountains, Brattleboro). Particularly good exposures are found on Governors Mountain, Guilford. The eastern belt (band #3) is too thin to have any major effects on the topography. Where it is thickest, it may locally crop out well and hold up a bench on the side of a hill.

The contact of the Standing Pond with the Waits River Formation is sharp and forms an excellent marker horizon. In the western belt of Standing Pond (bands #1 and #2) this contact is placed at the occurrence of the first thick, continuous amphibolite. On the side of the double loop (band #1) facing the interior of the Guilford dome, the contact is commonly marked by the local development of very course-grained mica-garnet-hornblende schist.

The contacts of the eastern belt (band #3, Figure 2-5) are also sharp. Where the volcanics are missing along this horizon, it is not possible to distinguish the Waits River Formation from the marble member of the Gile Mountain Formation.

Western Belt, Bands 1 and 2.

Rock Description. Black to dark green, fine- to medium-grained amphibolite is the most prominent rock in the western belt of the Standing Pond Volcanics. The amphibolites are massive to moderately foliated. Layered amphibolite, where hornblende and epidote-plagioclase form separate layers a few millimeters thick, is locally abundant. Hornblende, plagioclase, and quartz are the principal minerals, with minor amounts of chlorite, epidote and biotite present. Light green epidote knots are common, particularly near the two hinge areas of the double loop. On Richardson Mountain, Brattleboro, epidote-rich "knots" (pseudo-pillows?) 3 feet across have been deformed into cylinders up to 5 feet long parallel to the lineation. The size of hornblende crystals in the amphibolites is quite variable and is responsible for most of the textural differences. Chlorite and carbonate usually occur as alterations of the other minerals. Almandine is an additional porphyroblast, locally present.

A thin, course-grained amphibolite occurs near the contacts of some of the finer-grained amphibolites. Along the contact of band #1 of the Standing Pond with the Waits River Formation this is particularly well developed. It commonly occurs along with a distinctive, coarse-grained garnet-hornblende-quartz-mica schist. These schists, with a coarse-grained texture, are usually no more than a few feet thick. However, in a few localities, the grain size is abnormally coarse for several tens of feet adjacent to this contact. The garnets in this rock average 2 inches in diameter and may reach 8 inches or more. The hornblende is characteristically arranged in radiating sheaves on schistosity surfaces and may be up to one foot in length. Course-grained kyanite is associated with the rock in one locality (see Chapter 4).

The detailed stratigraphy of band #1 is somewhat different from that of band #2 in areas away from the closures of the double loop. Figure 2-6 gives two representative columns that are typical for bands #1 and #2. As the closures of the double loop are approached, bands #1 and #2 become indistinguishable and pass directly into one another.

Within the mapped band #1, a brown-weathering, poorly exposed schist usually separates two fine-grained amphibolites. The schist is quite variable in composition and is similar to schists in the Waits River Formation. Gray, felsic schists and thin, coarse-grained garnet amphibolites are also found with this schist. This schist band is continuous, although typically thin (10 to 50 feet thick) along most of band #1 of the Standing Pond Volcanics. It grades laterally into amphibolite near the hinges of the loop.

Band #2 of the Standing Pond Volcanics includes several fairly continuous, thin (3 to 50 feet) zones of well-bedded quartzite and light gray mica schist within the amphibolites. These rocks are similar to those found in the Gile Mountain Formation. Kyanite is abundant in these rocks. Thin, pink coticule beds are commonly found associated with the first such beds in band #2 above the Gile Mountain Formation. Coticule in association with thin, two-inch thick beds of black, fine-grained quartzite occurs in the amphibolite at the contact of the Gile Mountain Formation in the area north of and east of the village of Green River, Guilford. Pyrite and minor chalcopyrite are accessory minerals in the amphibolites here. Some of the quartzite beds become calcareous and grade into dark gray, brown-weathering, calcareous schists near the northeastern hinge

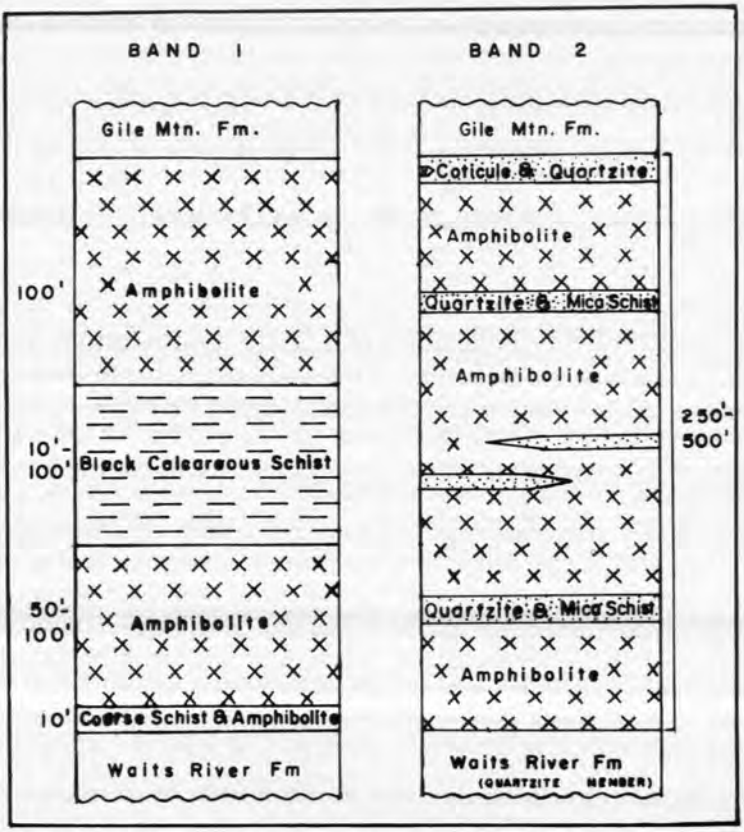

Figure 2-6. Sketch of detailed stratigraphy within the western belt of the Standing Pond Volcanics. Columns are representative of Bands #1 and #2 away from the areas near the closures of the double loop, where the bands become indistinguishable and pass directly into one another. Band #1 is sketched from exposures near the village of Green River in western Guildford township. Band #2 is sketched from exposures on Governors Mtn., Guilford.

of the Standing Pond loop near Dummerston Center.

Eastern Belt, Band #3.

Rock Description. Gray to dark gray, greenish, massive plagioclase - biotite - quartz and plagioclase - biotite - hornblende - quartz granulites, gneisses, and schists make up most of the eastern band of the Standing Pond Volcanics. Characteristically, the rocks are medium grained, with greenish hornblende porphyroblasts. Small altered plagioclase porphyroblasts or relict clasts are common. Carbonate, probably ankerite, is commonly abundant and weathers to small pits or brownish tan spots in the rock. Small almandine porphyroblasts are locally present west of the garnet isograd.

The thickness of this unit varies considerably along strike. Where it is at a maximum of up to 400 feet thick just north of Vermont Rt. 9, Brattleboro, and just north of the quadrangle boundary in Guilford, the granulites are separated by a band of brown-weathering phyllite up to 100 feet thick.

An excellent, fresh exposure of the rocks in the eastern belt of the Standing Pond and the contacts of this belt are visible in a roadcut, and in fields to the north of it, at the approach to Interstate 91 northbound at Exit 2 from Vermont Route 9, Brattleboro.

Origin. The Standing Pond Volcanics originated as a series of largely basaltic lava flows and pyroclastics. Table 2-2 shows representative chemical analyses and C.I.P.W. norms for the Standing Pond in the Brattleboro area. Hall (1959) found pillow structures in the Standing Pond in the St. Johnsbury quadrangle. Crystal tuffs are developed in the Standing Pond in the Saxtons River quadrangle (Rosenfeld, 1954). The bands of phyllite, schist, and quartzite were probably argillaceous and calcareous sands and muds, developed during quiescent times. The thin black quartzites and thin coticule beds were likely chert and manganiferous chert beds.

Thickness. The thickness of the Standing Pond Volcanics in the Brattleboro quadrangle ranges from 0 to 500 feet.

Gile Mountain Formation

General Statement and Areal Distribution. Doll (1944) introduced the term Gile Mountain Schists for quartz-mica schists in the Strafford quadrangle, Vermont. These rocks are continuous with similar rocks between the eastern belt (band #3, Figure 2-5) of the Standing Pond Volcanics and the Putney Volcanics in the eastern Brattleboro quadrangle (Plate I).

A band of rocks containing conspicuous impure marble beds similar to those in the Waits River Formation just east of band #3 of the Standing Pond Volcanics (Plate I), has been separated as the marble member of the Gile Mountain Formation. The marble member and quartzose schists, phyllites, and quartzites of the Gile Mountain Formation in this belt are in the chlorite and biotite zones of metamorphism. They will be discussed separately from the narrow band of quartzites and schists that has been correlated with the Gile Formation and lies between the two bands (bands #1 and #2, Figure 2-5) of the Standing Pond Volcanics in the center of the

Weight % Oxides

Sample Number	B3C	B6A	B8H	B11A	B500H	B4A	B9B
SiO_2	50.85	47.66	48.62	50.65	49.42	47.84	47.83
TiO_2	1.45	0.73	1.48	1.84	1.46	2.24	2.31
Al_2O_3	16.68	19.77	14.45	15.06	14.77	16.76	17.89
Fe_2O_3	11.47	7.62	13.65	11.65	10.62	10.75	9.48
MnO	0.15	0.15	0.23	0.14	0.18	0.16	0.15
MgO	5.88	8.53	8.59	7.22	7.27	9.94	6.94
CaO	8.18	11.79	9.43	9.67	14.16	6.63	9.80
Na_2O	4.96	3.13	3.26	2.97	2.11	3.55	3.84
K_2O	0.34	0.19	0.17	0.34	0.25	1.07	0.90
P_2O_5	0.13	0.06	0.14	0.20	0.16	0.60	0.52
Total	100.09	99.63	100.02	99.74	100.40	99.54	99.66

Table 2-2. Representative major element analyses from the Standing Pond Volcanics. Total iron as Fe_2O_3*. Analyses, except Na_2O, by X-ray Fluorescence in the laboratory of Prof. M. Rhodes, Dept. of Geology, Univ. of Massachusetts, Amherst. Na_2O by Instrumental Neutron Activation Analyses in the laboratory of Prof. F. Frey, Dept. Earth and Planetary Sciences, M.I.T.

B3C — fine-grained amphibolite, Prospect Hill Dummerston.

B6A — fine-grained amphibolite with plagioclase porphyroclasts, Richardson Mt., Brattleboro.

B8H — fine- to medium-grained amphibolite, band #1, (Fig. 2-5), 0.6 miles, SW of 831′ rd. junct., Hinesburg, Guilford.

B111A — fine-grained amphibolite, Richardson Mt., Brattleboro.

B500H — fine-grained amphibolite, band #2 (Fig. 2-5), on hill 1.1 miles NW of West Brattleboro Village, Brattleboro.

B4A — massive gray plagioclase-hornblende-chlorite granulite, band #3 (Fig. 2-5), roadcut on I91 interchange, northbound from Vt. Rt. 9, Brattleboro.

B9B — massive hornblende-plagioclase granulite with plagioclase porphyroclasts, band #3 (Fig. 2-5), on small hill, 0.65 miles south of Weatherhead Hollow School.

quadrangle (Plate I). These will hereafter be referred to as the western band of the Gile Mountain Formation.

The quartzites and quartz-rich schists of the Gile Mountain Formation are excellent ridge-formers and generally outcrop well.

Western Band.

Rock Description. The principal rocks in the western band of the Gile Mountain Formation are micaceous and feldspathic quartzites and mica schists. The impure quartzites weather light gray, gray, or tan. They are a lighter color on fresh surfaces, generally fine-grained, and massive to moderately well foliated, depending upon the percentage of micaceous minerals present. Quartz is the major constituent of these beds, with varying amounts of feldspar, muscovite, and biotite. Small, rusty-weathering pits indicate the presence of a carbonate, probably ankerite. Light to dark green hornblende porphyroblasts are usually found in the carbonate-bearing beds. Small almandine porphyroblasts are also usually present in minor amounts. The quartzites have a granulitic texture, are well-bedded, and occur interlayered with the mica schists in beds a few inches to several feet thick. Bedding is distinct and usually has the major schistosity parallel to it. A few poor examples of graded beds were found in the thinner quartzites.

The mica schists are gray to light gray, medium-grained, and consist of muscovite and quartz with garnet, biotite, staurolite, and kyanite porphyroblasts. The micas forming the schistosity are usually crinkled, but not as extensively as in the Northfield Formation. These schists generally form as thin beds between thicker quartzite beds, or as part of an alternating sequence of quartzites and schists in which the individual beds are only a few inches thick. More rarely, the schists are found in beds up to tens of feet thick.

The northeasternmost outcrops in the western band of the Gile Mountain Formation, south of the village of Dummerston Center, include some rocks similar to those found in the Waits River Formation. Brown-weathering, black, fine-grained, somewhat calcareous mica schists with thin beds of punky, brown-weathering, impure marble occur here. They grade into the impure quartzites and mica schists along strike to the south. The brown-weathering, dark schists are not uncommon as minor rocks elsewhere in this band of the Gile Mountain.

Contacts with the Standing Pond Volcanics. The western band of the Gile Mountain Formation is in sharp contact with amphibolite of the Standing Pond Volcanics on both sides. However, band #2 (Figure 2-5) of the Standing Pond Volcanics may contain two or three thin impure quartzite beds, similar to those in the Gile Mountain Formation, within the amphibolites. The contact of the Gile Mountain Formation is defined as the boundary of the large central belt of quartzites and schists with the first, thick, fine-grained amphibolite. This places all of the fine grained amphibolites in the Standing Pond.

In the small area just east of Black Mountain, Dummerston, where the Standing Pond (band #2) pinches out, the western band of the Gile Mountain Formation is in contact with the quartzitic member of the Waits River Formation. As noted previously, there is little difference between these rock units. The contact here is drawn by extending the known contact from the areas where Standing Pond band #2 is present.

Eastern Band-Marble Member.

General Statement. The marble member occurs along the western portion of the eastern band of the Gile Mountain Formation and separates the phyllites and quartzites of the Gile Mountain Formation from the Waits River and Standing Pond Formations (Plate I).

Rock Description. The marble member of the Gile Mountain Formation is identical to the Waits River Formation, except that the black schists described for the Waits River Formation are black phyllites in the marble member, since the metamorphic grade is lower. This member in the central portion of the quadrangle includes a high percentage of impure quartzite beds, 20 to 25 percent, but also contains an abundance of the impure marble beds, 20 to 30 percent. As the mapping criterion is essentially the presence or absence of the impure marble beds in amounts greater than 1%, these rocks are placed in the marble member.

Eastern Band.

Rock Description. The most common rocks in the eastern band of the Gile Mountain Formation are gray, fine-grained phyllites (slates in the chlorite zone of metamorphism). These rocks weather gray to dark gray but may have a light brown surficial stain. The phyllites are well foliated, commonly crinkled, and secondary cleavage may be present. Porphyroblasts are not common. Biotite is the most common porphyroblast where one can be observed. Pyrite, chlorite, and carbonate, the presence of which is indicated by small weathering pits, are occasionally present. Quartz lenses are numerous.

Thin beds of fine-grained, micaceous quartzite and quartz-rich phyllite are the other major rocks in the eastern band of the Gile Mountain Formation. These weather light gray and range in thickness from beds a few inches thick to rarely as much as three feet thick. Many of the individual beds of the quartzite have paper-thin laminae of muscovite. The micaceous quartzites form very sharp boundaries with the enclosing phyllite and do not appear graded. They are present locally to as much as 40 percent of the rock but overall account for no more than 10 to 20 percent of the formation.

Rare, thin, punky-brown-weathering, impure marbles similar to those described in the Waits River Formation are present. These occur as less than one percent of the rocks.

In the very southeastern part of the eastern band of the Gile Mountain Formation, between East Mountain, Guilford, and the Putney Volcanics, the rocks are largely gray slate and thin interbedded quartzites without impure marble interbeds. These rocks are indistinquishable lithologically from those in the low-grade Littleton Formation. It is possible that some of the rocks in this area do belong to the Littleton Formation in a tight fold such as the one along the northern boundary of the quadrangle, west of Putney Village (Plate I). However, since no additional bands of the Putney Volcanics or the conglomeratic member of the Littleton were observed in the area, these rocks have been included in the Gile Mountain Formation.

Contacts. The western contact of the marble member is sharp where it is in contact with band #3 (Figure 2-5) of the Standing Pond Volcanics. Where the Standing Pond is not present or exposed, the contact is drawn by extending the known contact from areas where Standing Pond band #3 is present.

The eastern contact of the marble member is gradational with the phyllites and impure quartzites of the more typical Gile Mountain. The impure, punky-brown-weathering marble beds become less abundant from west to east. The phyllites of the marble member become lighter in color to the east, weather gray to dark gray instead of brown, and are noncalcareous. The contact is drawn where the impure marbles make up no more that about 1 percent of the exposed rock in any sequence. Thin, impure marbles are found scattered throughout much of the Gile Mountain Formation but are quantitatively minor. Rarely is there much problem in placing the eastern contact of the marble member with more typical Gile Mountain, as the eastern part of the marble member has a high percentage of impure marble beds. There is a marked change between this zone rich in marble beds and the phyllites of the Gile Mountain proper, where there are essentially none. This change is reflected by the topography in some area, as the marbles weather easily.

Origin. The Gile Mountain Formation originated as a series of alternating muds and argillaceous sands. It is probable that some of the alternating sandstone and shale beds were the result of deposition by turbidity currents. The marble member was deposited in the same manner as the Waits River Formation.

Thickness. The outcrop width of the Gile Mountain Formation, including the marble member, ranges from 2,500 feet to 10,000 feet. If the Putney Volcanics are correlated with the Standing Pond Volcanics and if the Gile Mountain Formation overlies these units, then the top of the formation is not present in the Brattleboro quadrangle. If the Putney Volcanics are correlated with the Standing Pond Volcanics and are above the Gile Mountain Formation, the total thickness of the formation would be present and is estimated to be 2,500 to 5,000 feet. If the Putney Volcanics are not correlated with the Standing Pond Volcanics, then the Gile Mountain Formation must lie between these formations and a similar thickness would result.

The maximum observed thickness of the western band of the Gile Mountain Formation is 600 to 1,000 feet. The total section is not present in this band.

Putney Volcanics

General Statement and Areal Distribution. A north-south trending band of greenstones, feldspathic granulites and interbedded slates and phyllites in the eastern Brattleboro quadrangle has been named the Putney Volcanics by Hepburn (1972A,B) and Trask (1980). Previously, these rocks, which separate the Littleton Formation to the east from the Gile Mountain Formation, were correlated with the Standing Pond Volcanics (Doll et al., 1961; Billings, 1956). The name Putney Volcanics has been adopted because these rocks cannot be traced directly to the type locality of the Standing Pond, are of a considerably less mafic composition in the Brattleboro area than the type Standing Pond Volcanics, and because questions exist as to their proper correlation.

The type localities for the Putney Volcanics are exposures of this belt of rocks in the towns of Putney and Dummerston (Plate I). The rocks in this narrow belt are continuous across the eastern part of the quadrangle, except for a short gap in the town of Brattleboro and in Dummerston where they cross a bend of the Connecticut River. This same belt of rocks is shown by Doll et al. (1961) to be continuous northward as far as Springfield, Vermont. Trask (1964) notes that it apparently pinches out southward on Wildcat Mountain, in Bernardston, Massachusetts.

Small lenses of gray slate-matrix, quartz-pebble conglomerate occur along the contact with the Littleton Formation. These conglomeratic lenses were previously included in the Putney Volcanics by Hepburn (1972A) and Trask (1980). It is now felt that because the matrix is similar to that of the Littleton Formation, they should be placed in the Littleton and not the Putney Volcanics. Thus, they have been included in the Littleton Formation in this report and are shown as **D1c** on Plate I.

Rock Description. Poorly foliated, fine-grained, light-to dark gray, quartz-plagioclase-muscovite granulites, phyllites, and interbedded gray slates make up the Putney Volcanics. The granulites and feldspathic phyllites weather buff to light brownish gray, characteristic of feldspar-rich rocks. The rest of the phyllites weather a light greenish gray. Many of the foliation surfaces have a notable silky sheen. Small, brownish pits where carbonate has weathered out are common. Thin beds of fine-grained, more resistant feldspathic granulite a few inches thick are commonly interlayered with beds of the light greenish phyllite and gray slate. Individual beds of granulite may show a fine lamination that locally is cross-bedded. Somewhat thicker beds of the feldspathic granulite (to 2 feet) are irregularly interlayered with gray, well-foliated slate or phyllite near both contacts of the formation. Minor amounts of the gray slate are found scattered throughout the formation. The gray slate is indistinguishable from that in the adjoining Littleton and Gile Mountain Formations but distinct from the phyllites of the Putney Volcanics, as described above.

Stratigraphic Topping Evidence. In the Brattleboro area, stratigraphic topping evidence has been found at several sites within the Putney Volcanics and at its contact with the Littleton Formation. The topping evidence consistently indicates that the stratigraphy gets younger to the east, i.e. that the Littleton overlies the Putney.

The topping evidence is obtained from small cross-bedded sequences in the laminated feldspathic granulite beds in the Putney. Load casts (?) at the base of these granulite beds also support the same topping direction.

The cross-beds (Figure 2-7) are best seen in an abandoned chicken yard (adjacent to Vt. Route 5, 0.45 miles N.34°E. from road junction, 428′ south of Dutton Pines State Forest, Dummerston); on an exposure on the east bank of the Connecticut River (just south of Canoe Meadow, Westmoreland); and on various small islands in the Connecticut River south of Putney Station, Putney

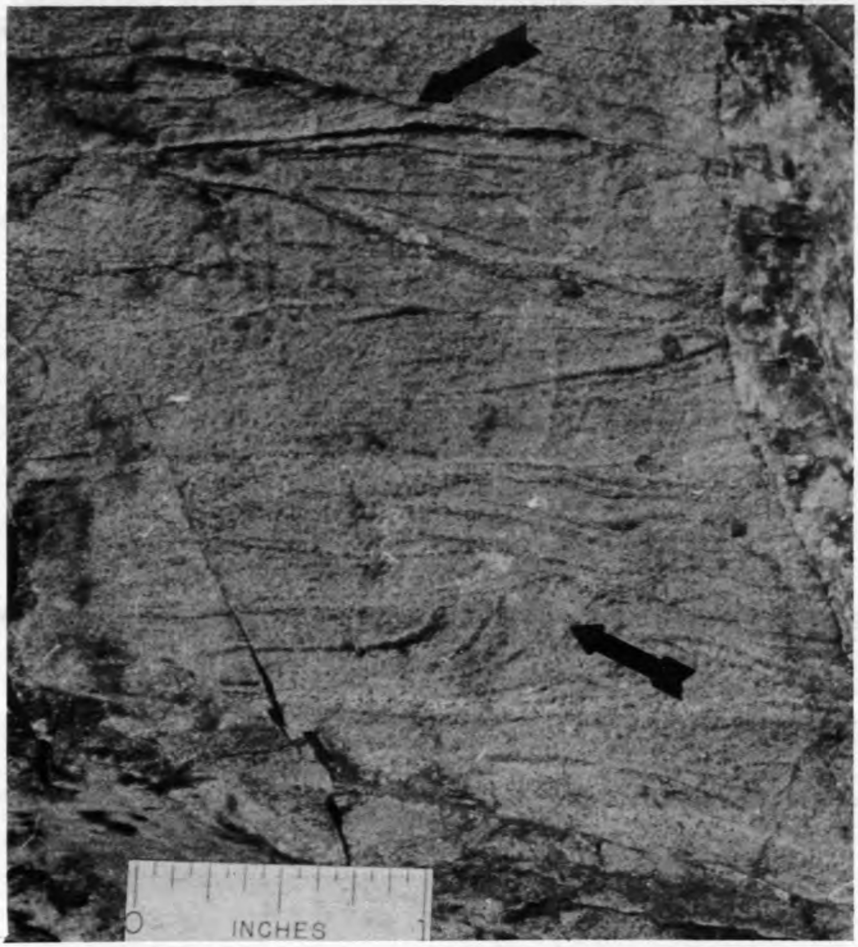

Figure 2-7. Cross-bedding in feldspathic granulite beds in the Putney Volcanics. Arrows point to particularly good examples. Photo looking down on the eastern bank of the Connecticut River, just south of Canoe Meadow, Westmoreland, N.H. Stratigraphic tops are toward the top of the photo (east).

(exposed at low water only). The chicken yard locality, however, is rapidly becoming overgrown since the chickens flew the coop, and the sedimentary structures are increasingly obscured.

The consistency of the topping direction in the Putney Volcanics, and the fact that they are observed very close to its contact with the Littleton Formation, give the authors confidence in their assignment of the stratigraphic tops at this boundary.

Contacts. The Putney has a gradational contact with the Gile Mountain Formation over a few tens of feet. The contact is drawn on the basis of the westernmost feldspathic granulite bed, which is included in the Putney. On fresh surfaces these beds are difficult to distinguish from the fine-grained quartzites of the Gile Mountain, but the characteristic felsic weathering color is distinctive. Thick feldspathic phyllites and greenstones are encountered a few tens of feet east of the contact.

The contact with the Littleton Formation to the east is generally sharp. Where the conglomeratic member of the Littleton is not present, the easternmost outcrops of the Putney Volcanics consist of felsic granulite beds interbedded with gray slate. The rock has been mapped as the Putney where the felsites are present; as Littleton, where they are not. Where the conglomerate is present, the contact is drawn at the base of the conglomerate and is sharp.

Origin. The Putney Volcanics are at least partially of volcanic origin. Trask (1964) notes that these rocks have compositions suggesting that they were derived from dacites. Further north, near Springfield, Vermont, the Putney contains greenstones of a basaltic composition. The thin beds of fine-grained feldspathic granulite suggest that the unit is at least partly pyroclastic. The laminated nature of some of the beds and the interbedded pelitic material indicates they were water-laid or water-reworked.

Thickness. The thickness of the Putney Volcanics ranges from 0 to 400 feet in the Brattleboro quadrangle.

PLUTONIC ROCKS

Ultramafic Rocks

Location. A small altered ultramafic body occurs in the very northwestern part of the Brattleboro quadrangle (Plate I). It forms a teardrop-shaped body about 1,000 feet in length and up to 300 feet in width, elongated parallel to the schistosity.

Description. The ultramafic body is largely serpentinite surrounded by an aureole of talc-carbonate rock. The serpentinite is generally massive, dark green, and weathers to a rusty-orange. The talc-carbonate is gray with prominent, rusty-weathering pits where the carbonate has weathered out. The schistosity in the talc-carbonate rock parallels that in the surrounding rocks of the Ordovician Moretown Member (David Walker, 1970, personal communication). Locally an actinolite, biolite, chlorite, "blackwall" alteration zone has developed between the country rock and the ultramafic.

A revised map of the large East Dover ultramafic (Skehan, 1961) by D. Walker (1969, unpublished) shows that this body does not extend into the Brattleboro quadrangle. Skehan (1961), Chidester (1968), Hess (1933), Bain (1936), and Chidester et al. (1951) give detailed descriptions of similar ultramafic bodies elsewhere in Vermont.

Age. This body was probably emplaced tectonically during the Late Ordovician, if it is contemporaneous with the other ultramafic bodies in western New England (Chidester, 1968).

New Hampshire Plutonic Series

General Statement. In the Brattleboro quadrangle one large body of granodiorite (Black Mountain Granite) and numerous small bodies of binary granodiorite and quartz diorite are assigned to the New Hampshire Plutonic Series (Billings, 1937)

Black Mountain Granite. The Black Mountain Granite occupies an elliptical area about 2 miles long by 1 mile wide, east of the village of West Dummerston. The granite is well exposed on Black Mountain and extends southward across the West River. This granite has been the subject of at least three previous studies (Dale, 1909, 1910, 1923; Maynard, 1934; and Church, 1937) as a result of the quarrying of this rock for building stone. The rock is a very light gray, massive, medium-grained, two-mica granodiorite. Flakes of muscovite form a very weak foliation roughly parallel to the schistosity in the surrounding rocks. A few small pegmatites and aplites cut the main granite body. Sheeting is well developed in the Black Mountain Granite, particularly along the western side. This sheeting dips westward at a moderate angle and was extensively used to aid in the quarrying. Figure 3-17 of the abandoned Black Mountain quarry shows clearly how the sheets become progressively thicker with depth. A well-developed joint set strikes about N.15^{0}E. and is vertical in the western part of the body. Dale (1910) reports a compressive strength of 27,810 pounds per square inch for this rock. Dale (1910, p. 172) gives the following description of the rock:

> it "is of very light gray shade, speckled with bronze-colored mica (muscovite and biotite) and of even-grained medium texture with feldspars up to 0.3 inch and mica to 0.1 inch. Its constituents, in descending order of abundance are: clear to pale smokey quartz, showing effect of

strain, with hairlike crystals of rutile and a few fluidal cavities in sheets; milk-white soda:lime feldspar (oligoclase to oligoclase-albite), some of it with flexed twinning planes, kaolinized and micacized; clear potash feldspar (microcline and orthoclase); muscovite and less biotite apparently intergrown and bent or twisted with fibrous muscovite stringers extending out from them into and between the other particles. Accessory: Apatite, rutile. Secondary: Kaolin, white micas, epidote, zoisite, calcite. There are crush borders about the quartz and feldspar particles."

Geochemically the granite is an "S"-type peraluminous granite likely derived by the partial melting of a sedimentary source (Hepburn, 1982). The peraluminous character of the Black Mountain Granite is shown mineralogically by the presence of muscovite and small garnet phenocrysts and by the presence of corundum in the norm (Table 2-3).

Sills and dikes of granodiorite and aplite are common outward a few hundred feet from the main body (Figure 3-16). An aureole of contact metamorphism extends outward several hundred feet beyond the area where the dikes and sills are prominent (see description under Waits River Formation). The dikes and sills seem to be little deformed macroscopically, except for a faint mica foliation approximately parallel to the regional schistosity. Microscopically the grains in the dikes and sills are somewhat more crushed than in the main body of the granodiorite. Inclusions and blocks of country rock surrounded by granitic apophyses are common near the granite contacts. Stoping seems to have been the most likely method for the emplacement of the Black Mountain Granite.

Naylor (1971) has compiled the most recent radiometric data on the Black Mountain Pluton. He cites difficulties encountered in Rb/Sr studies, because the homogeneity of the rock prevents the establishment of an adequate isochron. Fine-grained muscovite and biotite give mineral ages of 335 million years (m.y.) and 316 m.y., respectively, by the Rb/Sr method. The coarse micas in the Black Mountain Pluton have a Rb/Sr age of 377 m.y. and 383 m.y., which Naylor feels establishes a minimum age for the body as early Middle Devonian. He interprets the ages of the fine-grained micas as having been partially reset by a younger thermal event, perhaps about 250 m.y. ago.

Other Granitic Bodies of the New Hampshire Plutonic Series. Numerous, small bodies of granodiorite and quartz diorite as-

Weight % Oxides

	BLM	BMG1A	BMG1B	BMG2	BMG3
SiO_2	75.07	74.88	74.03	73.57	74.10
TiO_2	0.08	0.18	0.12	0.19	0.22
Al_2O_3	14.94	14.92	14.39	15.03	14.91
Fe_2O_3	0.52	0.58	0.23	0.58	0.70
MnO	0.01	0.02	0.01	0.20	0.02
MgO	0.08	0.13	0.02	0.13	0.17
CaO	1.15	1.07	0.79	0.98	0.98
Na_2O	3.58	4.18	4.13	4.46	4.58
K_2O	4.37	4.48	4.60	4.10	4.24
P_2O_5	0.03	.10	.08	0.09	0.10
Total	99.83	100.54	98.40	99.33	100.02

C.I.P.W. Norm

	BLM	BMG1A	BMG1B	BMG2	BMG3
Quartz	35.03	31.21	30.90	29.89	29.17
Corundum	2.30	1.48	1.36	1.68	1.24
Orthoclase	25.83	26.47	27.18	24.23	25.05
Albite	30.29	35.37	34.94	37.74	38.75
Anorthite	5.51	4.68	3.42	4.29	4.21
Hypersthene	0.20	0.32	0.05	0.32	0.42
Hematite	0.52	0.58	0.23	0.58	0.70
Ilmenite	0.02	0.04	0.03	0.04	0.04
Rutile	0.07	0.16	0.11	0.17	0.19
Apatite	0.07	0.23	0.18	0.21	0.24

Table 2-3. Representative chemical analyses and C.I.P.W. norms for the Black Mountain Granite. Samples from in and around the abandoned Presbrey-Leland quarry, West Dummerston. Analyses by Atomic Absorption and Spectrophotometry, Dept. of Geology and Geophysics, Boston College. Total iron as Fe_2O_3.

signed to the New Hampshire Plutonic Series occur in the eastern portion of the Brattleboro quadrangle. The larger ones have been shown on Plate I. Moore (1949) has described these in detail. They are generally small stocks of light gray, medium-grained granodiorite. Both foliated and non-foliated granodiorite bodies occur, with the smaller bodies more likely to be foliated. In these foliated bodies the quartz and feldspar have been granulated and the plagioclase saussuritized.

Post-Metamorphic Diabase Dike

A small diabase dike occurs in the bed of the Marlboro Branch of the Rock River, ½ mile north of B.M. 1066′, Marlboro. The dike is a massive, black, fine-grained, diabase some three feet thick intruding the Moretown Member of the Missisquoi Formation. The jointing in the dike is perpendicular to the contacts with the surrounding schists. The diabase has a subophitic to intergranular texture. Orthopyroxene phenocrysts as long as 0.6 mm occur in a groundmass of plagioclase laths and clinopyroxene. The grains appear quite fresh.

The dike is clearly post-metamorphic and probably is related to the Triassic-Jurassic diabasic intrusions in Massachusetts.

AGE AND CORRELATIONS

General Statement. No fossils have been found in the Brattleboro area. The ages of the stratified rocks have been established either by direct tracing along strike to areas where fossils have been found or by correlation with fossiliferous strata. Fossils do occur just to the south of the Brattleboro area at Bernardston, Massachusetts in the Clough and Fitch Formations of the Eastern Sequence.

Pre-Missisquoi Rocks. Detailed accounts of the correlations used to establish the ages of these rocks are given by Cady (1960, 1969), Rosenfeld (1954), Osberg (1965), and Chang et al. (1965). The Precambrian Mount Holly complex is exposed in the core of the Athens dome. The Hoosac Formation is Cambrian or Precambrian.

A continuation of the Ottauquechee Formation in Quebec can be traced around the north end of the Sutton Mountains anticlinorium into the Sweetsburg and Scottsmore Formations (Osberg, 1965; Cady, 1960). The Scottsmore has Early Cambrian fossils (Osberg, 1965). The Sweetsburg is at least partially Middle Cambrian and possibly partially Upper Cambrian on the basis of a *Cedaria* fauna in its equivalents in northwestern Vermont (Osberg, 1965; Cady, 1960; Theokritoff, 1968). Chang et al. (1965) summarize evidence which gives the Ottauquechee a possible range in age from Lower to Upper Cambrian by lithological correlations with fossiliferous rocks in the Taconic sequence. This means the Hoosac and Pinney Hollow can be no younger than Early Cambrian.

Missisquoi Formation. The Missisquoi Formation passes northward into the Beauceville Series (MacKay, 1921) in Canada (Cooke, 1950). Graptolites occur in these rocks at several localities near Magog, Quebec. Berry (1962) has restudied the graptolites from the classic Castle Brook and nearby slate localities and has assigned them a late Middle Ordovician (Wilderness and Trenton) age.

Russell Mountain Formation. As noted previously, there are two orthoquartzite-conglomerate horizons in southern Vermont, one at the base of the Northfield Formation (herein called the Russell Mountain) and one apparently on strike with part of the schist-amphibolite unit. Which of these two horizons represents the basal Silurian transgression and correlates with the fossiliferous Shaw Mountain Formation in northern Vermont is at present unclear. In any case, the Russell Mountain Formation can be no older than the Shaw Mountain. Corals and brachipods from Albany, Vermont (Konig and Dennis, 1964), indicate the Shaw Mountain can be late Llandovery to early Gedinnian, Early Silurian to Early Devonian (Boucot and Thompson, 1963). Fossiliferous limestones above a conglomerate correlated with the Shaw Mountain in the Lake Aylmer Group and at Lake Memphremagog in the Eastern Townships of Quebec are dated as Ludlow or Late Silurian in age by Naylor and Boucot (1965) and Boucot and Drapeau (1968). Thus, the Russell Mountain is Silurian in age. The unnamed schist-amphibolite unit below the Russell Mountain is Middle Ordovician or Silurian.

Northfield, Waits River, Standing Pond, Gile Mountain, and Putney Volcanics. A Siluro-Devonian age for this sequence of rocks is now fairly well established. Despite the fact that a few fossils have been found in the Waits River (Doll, 1943A; Cady, 1950) and Gile Mountain (Doll, 1943B) Formations, they are non-definitive; and the age of these rocks is best established by tracing them into fossiliferous rocks in Quebec or correlating them with fossiliferous rocks in New Hampshire. The details of these correlations have been presented elsewhere (see for example, Chang et al., 1965; Billings, 1956; Naylor and Boucot, 1965; Cady, 1960; Boucot and Drapeau, 1968; St.-Julien, 1965, and the Geologic Map of Quebec, 1969). Clearly the Northfield through Putney Volcanics section in the Brattleboro area lies above the Silurian Russell Mountain Formation, but exactly where the Silurian-Devonian boundary occurs is not known. It appears likely that at least the Northfield and Waits River are Silurian. If the interpretation is correct that stratigraphic tops along the Putney Volcanics-Littleton Formation boundary are to the east (i.e., the Putney is below the Littleton), then the majority of this sequence is most probably Silurian and has been so assumed in this report.

Since fossil control and clear evidence for stratigraphic tops has generally been lacking in the Siluro-Devonian rocks of the Connecticut River-Gaspé synclinorium in eastern Vermont, several plausible alternative interpretations of the stratigraphic order and structure are possible (e.g. Billings, 1956; White and Jahns, 1950; Doll et al., 1961; Doll, 1951; Murthy, 1957, 1958, 1959; Ern, 1963; Goodwin, 1963; Rosenfeld, 1968). Basically the problem lies in trying to determine where to draw the axis of the synclinorium. However, recently Fisher and Karabinos (1980) have discovered good stratigraphic topping evidence near Royalton, Vermont that indicates the Gile Mountain is younger than the Waits River Formation. This, coupled with the topping evidence from the Putney Volcanics, strongly supports the stratigraphic column as has been adopted here, (Figure 2-1, Plate I): Russell Mountain, Northfield, Waits River, Standing Pond, Gile Mountain, Putney Volcanics, from oldest to youngest. However, alternative interpretations are clearly possible.

Correlations with Western Massachusetts. Correlation of the rocks in the Brattleboro area with those of western Massachusetts (Figure 2-2) is relatively straightforward, as the rocks are similar along strike. The Hoosac Formation is correlated to the type Hoosac in western Massachusetts. The Pinney Hollow, Ottauquechee and Stowe Formations are equivalent to the Rowe Schist (Hatch, 1969; Hatch and Stanley, 1973). The Moretown Member is equivalent to the upper portion of Emerson's (1917) Savoy Schist or the Moretown of Hatch (1969) and Hatch and Stanley (1973). The Barnard Member of the Missisquoi and the schist-amphibolite unit are together equivalent to the Hawley Formation as defined by Hatch (1967); Hatch, Schnabel, and Norton (1968); and Hatch, Osberg, and Norton (1967). The Russell Mountain Formation is correlative to the type Russell Mountain Formation (Hatch, Stanley, and Clark, 1970). The Northfield is essentially continuous with the redefined Goshen of Hatch (1967), although graded quartzite beds make up a more substantial portion of the Goshen than the Northfield. The Waits River Formation is equivalent to the Conway Schist of Emerson (1917). Emerson's (1898) Leyden Argillite is essentially the same as the Gile Mountain Formation. An amphibolite member in the Conway Formation mapped by Segerstrom (1956) in the Colrain quadrangle is equivalent to the Standing Pond Volcanics.

CHAPTER 3

STRUCTURAL GEOLOGY OF THE WESTERN TERRANE

General Statement

The western portion of the quadrangle lies on the eastern limb of the Green Mountain anticlinorium, which is disrupted by the Sadawga and Ray Pond domes just to the west of the quadrangle (Figure 1-2). The southern tip of the Athens dome is present in the northwestern corner of the area. The Guilford dome occupies much of the central portion of the area (Figure 3-1). It is part of a belt of domes in eastern Vermont that occurs along or just west of the axial region of the Connecticut Valley-Gaspé synclinorium (Doll et al., 1961). The Colrain dome, just to the south of the Brattleboro quadrangle, also lies in this belt. The Waits River Formation forms the exposed core in both of these domes. Large recumbent folds are present in the strata mantling these domes. The Prospect Hill fold is one of these, outlined by the double closed loop of the Standing Pond Volcanics and by the quartzitic member of the Waits River Formation. The axial surface of this fold has been arched by the Guilford dome and depressed into troughs in the areas adjacent to the dome. A metamorphic high (staurolite-kyanite zone) is associated with the Guilford dome. The Black Mountain Granite intrudes the northern part of this dome. The Fall Brook anticline is a major fold northwest of the Guilford dome and is associated with the intense deformation in the belt of rocks underlain by the Northfield Formation.

The area to the east of the Guilford dome is part of a major synclinorium, locally named the Brattleboro syncline (Doll et al., 1961). East of this syncline are the gneiss-cored domes and recumbent folds of the Bronson Hill anticlinorium.

Secondary foliation is well developed in all of the metasedimentary and metavolcanic rocks. Where bedding or compositional layering is present, the major secondary foliation is generally parallel or subparallel to it. Schistosity, defined by the parallel arrangement of micaceous minerals or hornblende, is the most typical secondary foliation. In the chlorite and biotite metamorphic zones the platy minerals are generally too small to distinguish individually and give the rocks a slaty cleavage (Billings, 1954, p. 339). Slip cleavage and fracture cleavage (used descriptively, according to Billings, 1954, p. 339) are additional types of foliation commonly present.

Other structural features studied in the field include lineations

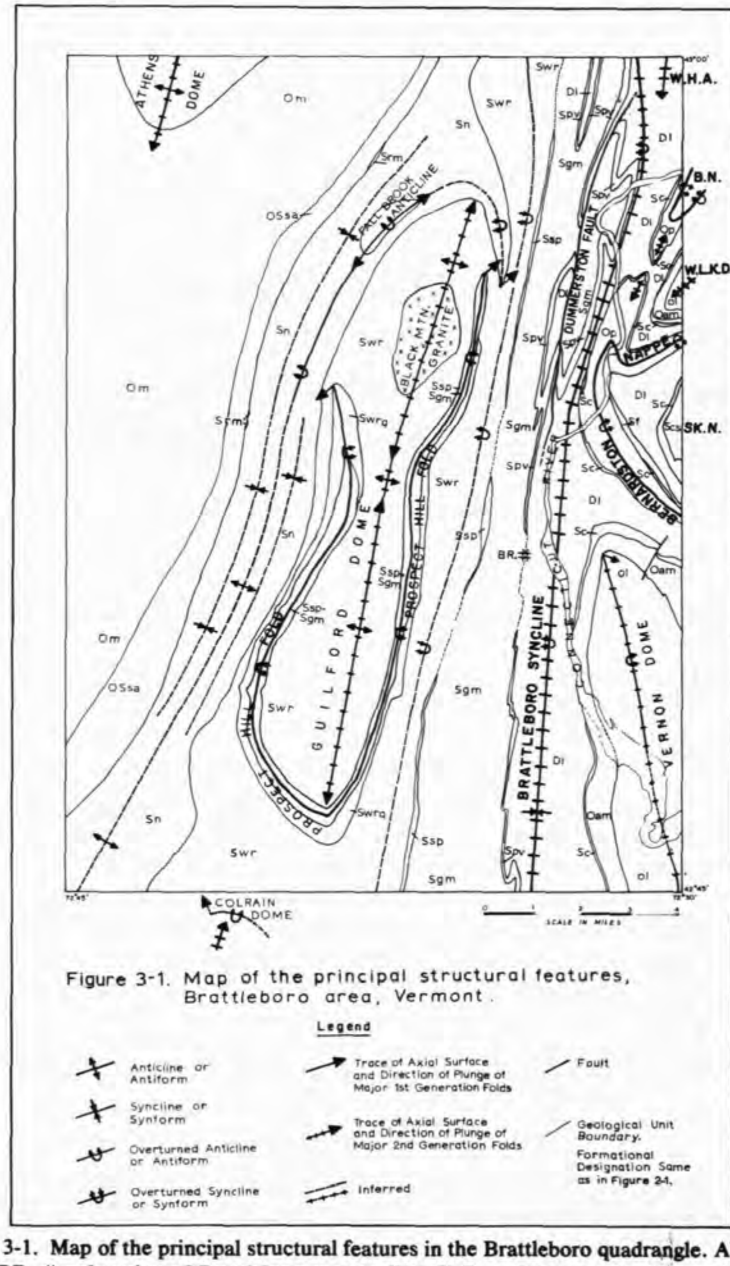

Figure 3-1. Map of the principal structural features, Brattleboro area, Vermont.

Figure 3-1. Map of the principal structural features in the Brattleboro quadrangle. Abreviations: BR. # = location of Brattleboro center, W.L.K.D. = Westmoreland lobe of Keene dome, B.N. = Bernardston Nappe, W.H.A. = Wellington Hill Anticline. The shading is on both limbs of the Standing Pond Volcanics and the Gile Mt. Fm. in the core of the Prospect Hill recumbent fold. For more detail on the eastern portion of the area, see Figure 6-1. Formational abbreviations are the same as Figure 2-1 or Plate I.

and minor folds. The most common lineations are mineral alignments and microcrenulations or crinkles that are well developed in most of the pelitic rocks. Table 3-1 summarizes the stages of minor folding by geological sub-areas. More complete descriptions and justifications for the sequential order of the minor fold generations in each of these sub-areas is given in Hepburn, 1972A.

Pre-Silurian Rocks, East Limb, Green Mountain Anticlinorium

Major Structure.

Most of the pre-Silurian rocks in the western Brattleboro quadrangle form an east-dipping sequence on the east limb of the Green Mountain anticlinorium. The Sadawga and Ray Pond domes (Skehan, 1961) are present on this limb of the anticlinorium just to the west of the Brattleboro quadrangle. The formations trend north-northeasterly with straight to gently curved contacts as a result of the moderate to steeply east-dipping major bedding and schistosity surfaces. The rocks of the Missisquoi Formation underlie most of the area. Older rocks are exposed at the south end of the Athens dome in the northwestern corner of the area. The Athens dome has an upright axial surface and plunges to the southwest at a moderate angle. The doming is later than the major schistosity in this area, as the schistosity has been arched by the dome (Rosenfeld, 1954).

Minor Structures.

While the large-scale structure in these rocks produces a relatively simple map pattern, the minor structures are complex. The "pinstriped" quartzites of the Moretown Member of the Missisquoi Formation show minor structures particularly well. A well-developed schistosity is present parallel to bedding[1] (S_0) or compositional layering throughout all of the more schistose pre-Silurian rocks.

The earliest minor folds, M_1 (Table 3-1) are tight to isoclinal, generally small, fold bedding (S_0), and have a well developed schistosity (S_1) parallel to the axial surface of the folds (Figure 3-2A). Metamorphic differentiation has produced compositional layering parallel to S_1 that can be observed to cut S_0 in the hinge areas of the early folds. At least some of the thin "pinstriped" compositional layering in quartzites of the Moretown Member are tectonically developed by this differentiation parallel to S_1 and the axial sur-

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[1] S_0 is used to denote bedding, S_1 denotes the earliest observable schistosity in an outcrop, S_2 denotes the second . . .

Sequence of Folds	Area 1 pre-Silurian rocks- Missisquoi Formation	Area 2a Northfield Formation	Area 2b Barnard Gneiss in Fall Brook Anticline	Area 3 Guilford Dome Area	Area 4 Brattleboro Syncline
F_6 F_5					B_5 Kink-bands. B_4(?) Large, steeply plunging sinistral folds in the Putney Volcanics.
F_4	M_4 Open buckles or warps in schistosity; crinkles may parallel axis.	N_3 Open buckles or warps in schistosity; may have crinkles parallel to axis.		GD_4 Open buckles or warps in the schistosity.	B_3 Flexure slip folds; axial plane slip-cleavage; crinkles parallel axis; plunge moderately to steeply.
F_3	M_3 Open folds in schistocity and bedding; possibly more than one generation; with or without axial-plane cleavage.	N_2 Open folds; slip-cleavage parallel to axial planes; strike of axial planes more easterly than schistosity; plunge moderately to steeply.	FB_2 More open folds; strike of axes NE-SW; shallow to moderate plunges.	GD_3 Open folds, particularly in area of south end of Guilford Dome; may be more than one generation; may or may not have axial plane slip-cleavage developed.	B_2 Flexure slip folds; moderately open, shallow plunges; crinkles parallel axis.
F_2	M_2 Isoclinal folds; schistosity parallel to axial planes in meta-pelites; not well developed in quartzites; main schistosity whether formed by M_1 or M_2 parallel to axial surfaces; M_2 cuts M_1 only in hinges of folds; axes generally NE at 20-40°; ~~axial surface~~ strikes NE and dips moderately to steeply east.	N_1 Isoclinal folds; various scales; main schistosity developed parallel to axial plane of these folds; plunges are gently NE-SW.	FB_1 Tight to isoclinal folds; fold schistosity; have weak axial plane slip-cleavage; axial planes dip NW; folds plunge gently to moderately NE or SW.	GD_2 Tight to isoclinal folds; these fold schistosity and banding; weak axial plane slip-cleavage; these folds related to recumbent folding; plunge NE or SW moderately.	B_1 Isoclinal folds in thin quartzite beds; main cleavage axial planar to folds; shallow plunges.
F_1	M_1 Small isoclinal folds (only a few observed); these fold bedding and have the principal schistosity parallel to the axial surface; shallow plunges (?); at least some "pin-stripe" laminations in Moretown are developed parallel to the axial planes of these folds.	?? Early schistosity parallel to bedding.	Early schistosity well developed; possibly early minor folds.	GD_1 Small isoclinal folds; principal schistosity axial planar to folds.	? Earlier cleavage parallel to bedding.

Table 3-1. Summary of minor fold stages in the Western Brattleboro quadrangle. Sequence of minor folds in each sub-area is shown by letter designation for that area (i.e., N_1 is earlier than N_2, etc.).

faces of the early folds. Where the M_1 folds are observed, the axes plunge moderately to the north, with the indication that the bedding has a more easterly strike than the later schistosity. Where these folds are not observed, S_1 generally parallels bedding (S_0). M_1 folds are rarely observed in the Barnard Volcanics.

The most common minor folds (M_2) in the pre-Silurian rocks fold both bedding (S_0) and the early schistosity (S_1) (Figure 3-2B). The folds (M_2) are generally isoclinal. A schistosity (S_2) is well developed parallel to the axial surfaces of these folds in the more pelitic rocks and may obliterate the earlier S_1 schistosity. In quartz-rich beds, S_2 is not as well developed, and a slip-cleavage has developed parallel to the M_2 axial surfaces. Since S_1 and S_2 are commonly parallel, the axial surfaces of the M_2 folds are also commonly parallel to the major schistosity in any given outcrop. S_1 and S_2 only diverge near the M_2 fold hinges. In certain areas S_2 is the dominant schistosity, but it is believed that S_1 is the more widespread of the two schistosities.

The axial surfaces of the M_2 folds strike north-northeast and dip moderately to steeply east or steeply northwest. The axes of these folds in the Moretown, except near the south end of the Athens dome, plunge northeasterly from 20 to 40 degrees. Crinkle lineations are developed parallel to the fold axes as the result of the intersection of S_1 and S_2. Hornblende may be aligned parallel to these fold axes. Quartz stringers are commonly folded by the M_2 folds and may form rod-shaped masses parallel to the fold axes (Figure 3-2C). M_2 folds are not as tight in the amphibolites of the Barnard where the axes show more variability than in the Moretown. In the west-central and southwestern part of the area, particularly along the Green River, the M_2 axes plunge east to southeast at moderate angles. These plunges are similar to those for fold axes just to the west in the Wilmington quadrangle (Skehan 1961), his Plate II and Figure 41) and may be related to refolding by the Sadawga dome.

Open folds (M_3) and "warps" or "buckles" in the schistosity (M_4) form younger generations of folds in the area. The M_3 folds are only locally well developed and generally have a slip-cleavage parallel to the axial plane in the hinge area of the folds. Crinkle lineations parallel these fold axes. The open "warps" or "buckles" (M_4) cause changes in the attitude of the schistosity on a scale of a few feet. Crinkles may also be developed locally parallel to the M_4 fold axes.

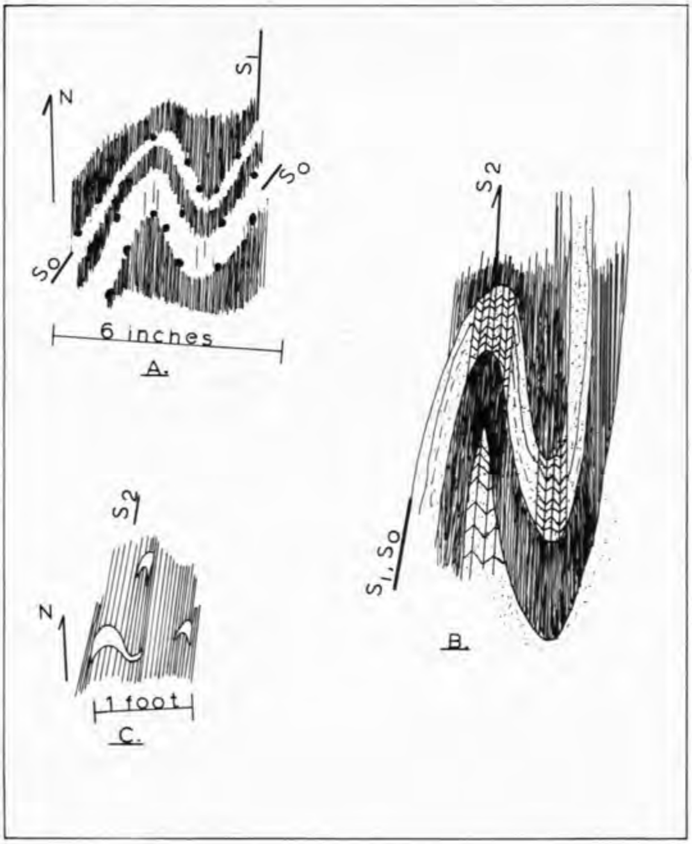


Figure 3-2

A. "Pinstriped" layering is parallel to S_1. S_1 is parallel to the axial surface of the M_1 folds (Table 3-1) shown by the folded bedding, S_0, in a more quartzose bed. Dark spots represent garnets. From field sketch of outcrop in the Moretown Member, 1500′ contour, south of 1717′ peak, SW corner, Newfane.

B. Schematic representation of the different development of S_2 cleavage in schistose and quartzose (dotted pattern) beds in the hinge areas of M_2 folds.

C. Folded quartz lenses in the hinges of M_2 folds, From field sketch of outcrop in the Moretown Member, 1000′ N.20°W. from B.M. 1666′, Marlboro.

Central and East-Central Brattleboro Quadrangle

General Statement.

The central and east-central portions of the Brattleboro quadrangle are underlain by Silurian rocks, with the exception of a small area of the Barnard Volcanics brought up in the core of the Fall

Brook anticline (Plate I). The deformational history of this area includes two major and at least five minor stages of folding. The Prospect Hill fold and Fall Brook anticline formed during the first major deformational stage, while the Guilford dome and Brattleboro syncline formed during the subsequent major stage of deformation. The rise of the Guilford dome arched the axial surface of the earlier Prospect Hill recumbent fold, so that now the hinge plunges to the northeast and southwest away from the dome. Minor folding formed prior to, during and subsequent to the development of the major structures. Tight to isoclinal minor folds occurred during the first two minor folding episodes. More open minor folds were formed by the subsequent generations (Table 3-1). Cleavage surfaces and lineations developed during most of the minor folding events.

Younger Major Structures.

Guilford Dome. The Guilford dome is a large, elliptical, doubly plunging anticline formed during the second major stage of deformation. The Waits River Formation forms the exposed core of the dome (Plate I). The Standing Pond Volcanics and Gile Mountain Formation in the Prospect Hill fold, form a doubly closed loop that outlines the flanks of the dome. As measured by this loop, the dome is about 12 miles long and 2 to 3 miles wide. The Black Mountain Granite intrudes the northern part of the axial region of the dome. The hinge of the dome trends slightly east of north and plunges moderately to the north and south at the ends of the dome. The axial surface of the dome dips very steeply to the west. A small depression in the axis forms a saddle in the center of the dome, east of Round Mountain, Brattleboro, dividing the dome into northern and southern lobes. This saddle is reflected on the geological map as a constriction in the mantling rocks (Plate I). The dome is asymmetrical with the axial trace lying closer to the eastern side of the dome. Foliation has steeper dips on the east flank of the dome than on the west. Bedding with a schistosity parallel to it has been arched by the dome.

Brattleboro Syncline. The Brattleboro syncline (Moore, 1949; Trask, 1964; Doll et al., 1961) is the local name for the eastern part of the regional Connecticut Valley-Gaspé synclinorium between the Guilford dome and the Vernon and Keene domes of the Bronson Hill anticlinorium. The Gile Mountain, Putney, and Littleton Formations occur at relatively low metamorphic grade within this syncline, which developed during the second major stage of deformation. The Brattleboro syncline strikes aproximately N.10°E.

Both limbs and the axial surface are vertical or dip steeply to the east. The axis is essentially horizontal. The exact position of the trace of the synclinal axial surface depends to some extent on the interpretation of the stratigraphic order of the formations involved; in Figure 3-1 this axial surface has been shown as lying within the Littleton Formation.

Early Major Folds.

Prospect Hill Fold. The Prospect Hill fold is well exposed as a doubly closed, narrow loop of the Standing Pond Volcanics that extends around the southern end of the Guilford dome with a hook-shaped map trace (Plate I, Figure 3-1). The Gile Mountain Formation is in the core of the fold. The quartzitic member of the Waits River Formation is adjacent to band #2 of the Standing Pond (Figure 2-5) for most of its length and defines the northwestern hinge of the fold on Bear Hill, Brattleboro, with the Standing Pond closure being just to the south on Richardson Mountain.

The fold is named for Prospect Hill, Dummerston, where excellent exposures of the hinge at the northeastern end of the loop occur in the Standing Pond Volcanics. Except for a short gap in Dummerston, which may be of tectonic origin, the Standing Pond Volcanics may be traced continuously around the doubly closed loop. The thickness of the Standing Pond bands varies considerably, but the detailed stratigraphy within each band is suprisingly consistent and can be traced for long distances (Figure 2-6). Stratigraphic units in the hinges of the fold have been notably thickened relative to the limbs.

The map pattern shown by the Standing Pond Volcanics, Gile Mountain Formation, and quartzitic member of the Waits River Formation (Plate I) is an excellent example of a refolded fold of the Ramsay Type II pattern (Ramsay, 1967, 1962). This accounts for its hook-shaped map pattern. Prior to the doming, the recumbent fold had a subhorizontal axial surface and a hinge striking approximately N.45°E. This has been arched by the doming so that it now plunges moderately to the northeast and southwest away from the north-south axis of the dome. Figure 3-3 shows the cluster of minor fold hinges in the major fold hinge area on Prospect Hill. A plot of poles to foliation in the same area gives the B-axis as N.34°E. at 25°.

The southwestern closure of the fold on Richardson Mountain is not as well exposed as that on Prospect Hill due to glacial drift. However, a thin (20 to 50 feet thick) interbed of quartzite, mica schist, and coticule in amphibolites of the Standing Pond Volcanics can be traced around the hinge on the southern slopes of Richardson Mountain. Moreover, the quartzitic member of the Waits River Formation is well exposed on Bear Hill, just to the north, and its contact with the rest of the Waits River outlines this hinge. The axis of the Prospect Hill fold plunges moderately to the southwest in this locality. The minor folds and the foliation are somewhat less regular near this hinge, as the later refolding at this point took place near the saddle in the axis of the Guilford dome (Plate I).

Figure 3-4 is a schematic representation of the Standing Pond Volcanics, projected above the north end of the north-plunging Guilford dome. It shows the northeasterly plunge of the Prospect Hill fold that resulted from the later doming.

An early, tight, steeply east-dipping synform must lie directly east of the Guilford dome, between Standing Pond bands #2 and #3 (Figure 2-5). The hinge line where the Standing Pond rocks cross the axial surface of this synform is not seen in the Brattleboro area and is presumably buried. It is believed this synform is the overturned upper, or anticlinal, part of the Prospect Hill fold.

A thin band of amphibolite and course-grained schist mapped as the Standing Pond Volcanics crops out along and to the north of a small hill straddling the Brattleboro-Guilford town line, 0.4 mile SW of road junction 932′, Brattleboro (Plate I). This band is nearer to the axis of the dome than Standing Pond band #1 (Figure 2-5), and it can be traced for about 3,000 feet in a north-south direction. It is surrounded on all sides by the Waits River Formation. There is no direct evidence to indicate whether this is a small infold of the Standing Pond or a lens of amphibolite at a slightly different stratigraphic position.

Fall Brook Anticline. The Barnard Volcanics occurs in the core of a tight, nearly isoclinal, doubly plunging anticline on the northwest flank of the Guilford dome, well exposed along Fall Brook in the western part of Dummerston (Plate I) where the West River has cut a particularly deep valley with over 1,000 feet of local topographic relief. Minor amounts of black schist, similar to those of the schist-amphibolite unit and distinct from those of the Northfield, are present with the Barnard in this area but have not been mapped separately. The anticline is overturned, with both limbs dipping to the northwest. The overturned eastern limb dips more steeply (60° to 85°) than the western limb (40° to 60°). The axial surface dips moderately to the northwest and is gently curved, striking NNE in the south and ENE in the north. The anticline plunges 20°, N.45°E.

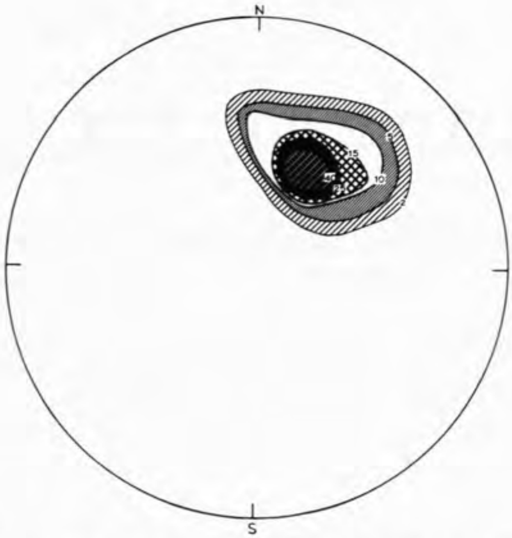


Figure 3-3. Lower hemisphere equal-area stereonet plot of minor fold hinges from the major fold hinge area of the Prospect Hill fold at the northeastern end of the Standing Pond Volcanics loop, Dummerston, Vt. Fifty fold hinges; contours labeled at 5, 10, 15, 25, and 40 %. After Hepburn, 1975.

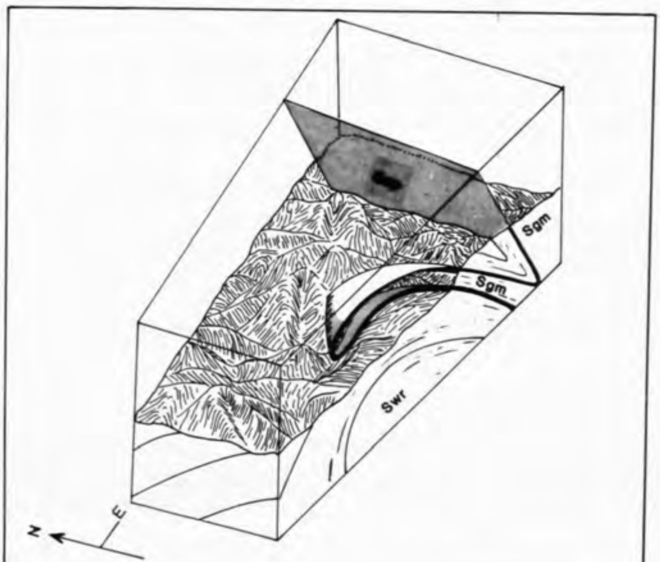


Figure 3-4. Schematic block diagram of the Prospect Hill fold projected above the north end of the north-plunging Guilford dome.

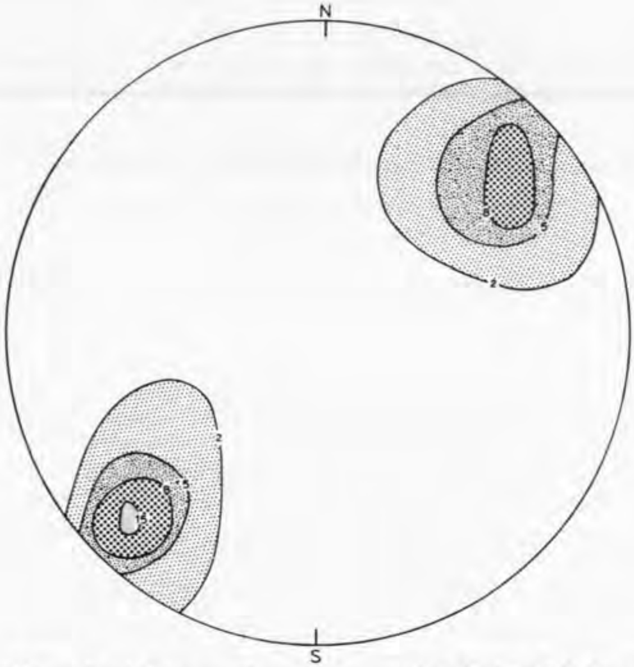


Figure 3-5. Lower hemisphere equal-area stereonet plot of minor fold hinges in the Fall Brook anticline. Northeast-plunging folds are from exposures near the northeastern end of the fold; southwest-plunging folds are from exposures near the southwestern end of the fold. Eighty-six fold hinges. Contours at 2, 5, 8, and 15 %.

and 12°, S.45°W. (Figure 3-5) with a culmination approximately where the West River cuts the fold. The Fall Brook anticline, and the associated tongue of the Northfield Formation extending around the north end of the Guilford dome (Figure 3-1, Plate I), are believed to represent the anticlinal part of Prospect Hill fold. This recumbent anticline lies above the synclinal portion that makes the double loop exposed around the Guilford dome (Plate I).

Minor Folds.

General Statement. The following sequence of deformation has been determined from the interrelations of the various minor structional features (e.g. refolded folds, schistosity, cleavage, and lineations). No more than two or three minor folding stages were observed in any one outcrop, and the evidence for all of the stages is not preserved in any one area of the quadrangle. Deceptively simple minor structural patterns were observed in many localities where one of the deformational stages has been particularly intense. The sequence of the deformations observed in any one sub-area (Table 3-1) of the quadrangle is more certain than correlations between sub-areas.

Figure 3-6. F_1 folds in compositional layering in amphibolites of the Standing Pond Volcanics, Prospect Hill, Dummerston. Schistosity (S_1) is parallel to the axial surface of the F_1 folds. Note refolding by the more open F_2 fold, which has an axial trace subparallel to the pencil. The pencil is approximately 5 inches long. After Hepburn, 1975.

F_1: A well-developed schistosity (S_1) approximately parallels bedding throughout the area. This schistosity is folded by minor folds congruous with the Prospect Hill fold and therefore predates the recumbent folding. A few cases in which S_1 cuts bedding or layering have been observed in the hinges of small isoclinal folds. Figure 3-6 shows an F_1 fold in layered amphibolites of the Standing Pond Volcanics. The principal schistosity is parallel to the axial surfaces of the F_1 folds. Note that the F_1 fold and this schistosity have been refolded by an F_2 minor fold. No large-scale structural features except the prominent schistosity are correlated with the F_1 minor folds.

Figure 3-7. Typical F_2 minor fold in an amphibolite of the Standing Pond Volcanics in the hinge area of the Prospect Hill recumbent fold, Prospect Hill, Dummerston. A weak slip cleavage parallel to the axial surface of the fold has been accentuated by dashed lines.

F_2: The F_2 minor folds formed contemporaneously with the major recumbent folds. These minor folds are isoclinal to moderately open, depending upon the type of rock involved. Amphibolites in the Standing Pond and Barnard Volcanics and quartz-rich rocks in the Gile Mountain and Waits River Formations are folded somewhat more openly than the schists and impure marbles in these same formations. Figures 3-7 and 3-8 show examples of typical F_2 folds. An axial planar cleavage, S_2, has developed parallel to the axial surface of the F_2 folds. It varies from a weak slip-cleavage in the hinge area of the F_2 folds in the more competent rocks to a well-developed schistosity in less competent rocks.

F_2 minor folds increase greatly in abundance in the hinge areas of the Prospect Hill fold, where they show the classic reversal in shearing direction across the axial surface of the larger fold. For example, in the well-exposed hinge area on Prospect Hill, Dummerston, the pattern of minor folds changes from S (left-handed) to M to Z (right-handed) from east to west when viewed to the northeast down the axis of the major fold.

A plot of the poles to the axial planes of the F_2 minor folds (GD_2, Table 3-1), parasitic to the larger recumbent fold in the north end of the Guilford dome, shows they fall on a fairly well defined great circle about the axis of the later Guilford dome (Figure 3-9).

The F_2 minor folds associated with the synform east of the Prospect Hill recumbent fold plunge at moderate angles. The axial surfaces of these folds dip 50-70°E., parallel to the regional schistosity.

Southwest of the Guilford dome near the southern boundary of the quadrangle, the schistosity dips to the north away from the Colrain dome. The F_2 minor folds, congruous with the recumbent fold, plunge to the northwest in this area. The axial surface of the Prospect Hill recumbent fold in this area was arched by the Colrain dome in a manner analogous to the arching by the Guilford dome. This causes the axial surface of the Prospect Hill fold to wrap around the northern end of the Colrain dome and produces the northerly plunges of the F_2 minor folds.

Isoclinal minor folds, tentatively assigned to the F_2 stage, occur in the Northfield Formation west of the Guilford dome. The structure is complex in this area, and the general schistosity changes from east-dipping, through vertical, to west-dipping adjacent to the Guilford dome. These isoclinal folds are observed on a scale of a few inches to as much as 100 feet in places where thin quartz-rich beds occur in the otherwise uniform schists of the Northfield. The lack of well-defined bedding throughout the formation hinders the interpretation of the large-scale fold patterns.

Ten determinations of topping direction were made from graded bedding in outcrops of the Northfield Formation, not visibly isoclinally folded. Six of the determinations indicated stratigraphic tops are to the northwest; four indicated tops to the southeast. Stratigraphic arguments indicate that tops are to the east at the base of the Northfield. Thus, the presence of large isoclinal folds in the Northfield is indicated, as shown schematically in Figure 3-1.

The observed folds in the Northfield are tightly isoclinal, have steeply dipping axial surfaces, and plunge at shallow angles to the

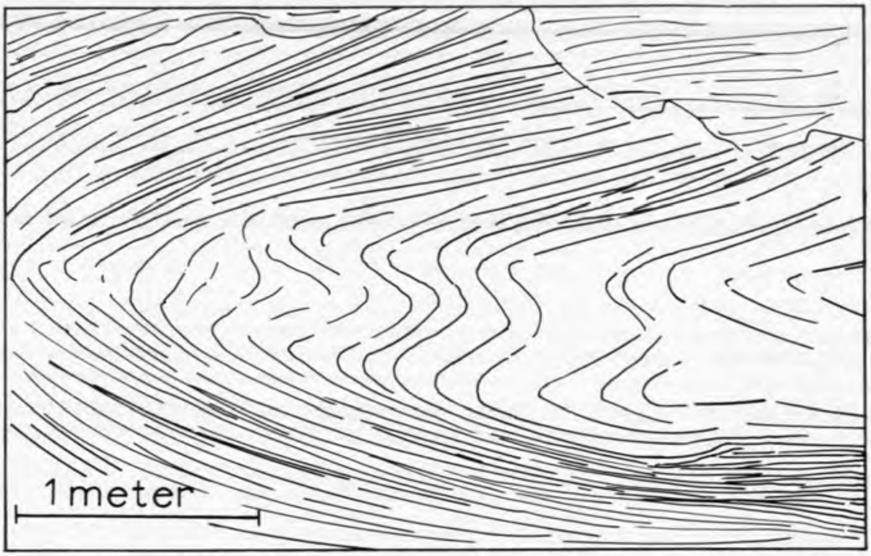


Figure 3-8. Sketch of an F_2 fold in an impure marble bed in the Waits River Fm., north bank of the Green River, 500 feet west of the Halifax-Guilford town line.

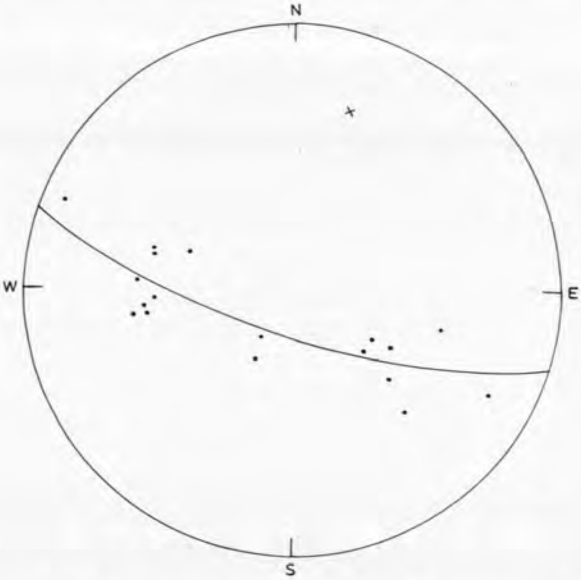


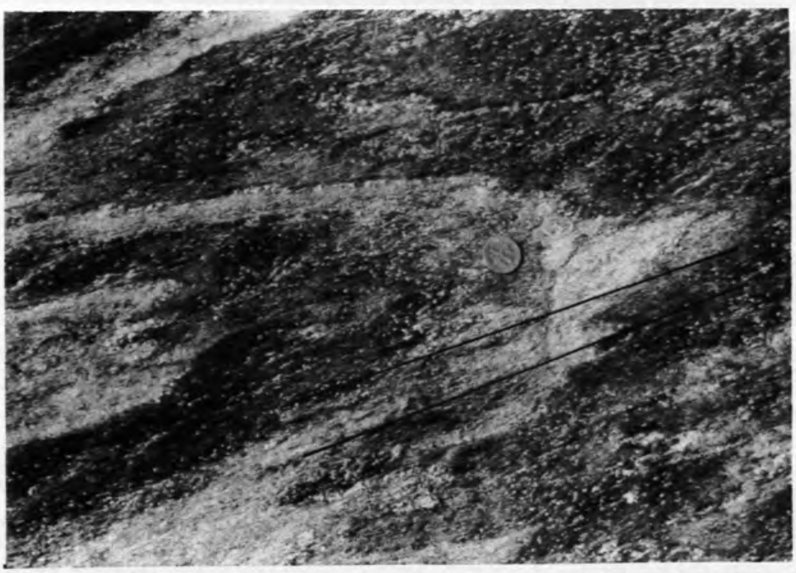
Figure 3-9. Lower hemisphere stereonet plot of the poles to the axial surfaces of minor F_2 folds congruous with the major Prospect Hill recumbent fold in the north end of the Guilford dome. The poles show a fairly well-defined great circle distribution about the axis (X) of the later dome folding. After Hepburn, 1975.

northeast or southwest. A well-developed schistosity parallels the axial surfaces of the folds and forms the principal schistosity in the area (Figure 3-10). The schistosity generally parallels bedding except in the hinges of the folds. In a few cases in particularly quartz-rich beds, it appears that an earlier, now deformed, schistosity parallel to bedding may also be present. The N_1 folds have been assigned to the F_2 stage of folding, largely on the basis of their similarity with the Fall Brook anticline and their northeasterly plunge at the north end of the Guilford dome. The present attitude of the folds, particularly the steeply dipping axial surfaces, is attributed to the later refolding.

In the Brattleboro syncline tight isoclinal folds (B_1, Table 3-1, Figure 3-11) in arenaceous beds are the earliest minor folds observed. A cleavage (S_1) parallel to bedding is folded by these folds, and the presence of an earlier fold stage is inferred. The principal slaty cleavage (S_2) of the area is parallel to the axial surfaces of these folds, striking north and dipping steeply. The axes of the folds generally plunge north to northeast at shallow angles where not disturbed by later folds. The axial plane cleavage (F_2) is weak in the arenaceous beds, but appears to be the prominent foliation in the phyllites and slates. Although there is some thickening in the hinge areas of these folds (Figure 3-11), slip seems to have taken place largely along the very thin micaceous laminae between the thicker (1 to 3 cm) quartzose layers. The individual folds, though isoclinal, die out quickly in a direction perpendicular to the axis in the axial plane, reminiscent of parallel folding.

F_3: Open minor folds with a well-developed slip cleavage (Billings, 1954, p. 339) parallel to their axial surfaces (Figure 3-12) are assigned to the third stage of minor folding. These folds are particularly well developed in the area west and south of the Guilford dome (N_2, GD_3, Table 3-1). They generally plunge moderately northeast and have axial surfaces that usually strike northeast and dip steeply northwest. Bedding (S_0) and the earlier schistosities (S_1 and S_2) are folded by the F_3 folds and transected by the S_3 slip cleavage. S_3 results from the realignment of S_2 and the growth of micas parallel to the axial surfaces of small microcrenulations (Figure 3-13). White (1949) described similar slip cleavage in detail. Excellent crinkle lineations occur at the intersection of slip cleavage and schistosity in the pelitic rocks. S_3 is prominently developed in the area where the major foliation changes from east-dipping, off the Green Mountain anticlinorium, to west-dipping, adjacent to the Guilford dome. In this zone, the slip cleavage generally strikes

Figure 3-10. (A) Minor F_2 folds in thin quartzitic interbeds in the Northfield Fm. in the hinge of a larger isoclinal fold. Graded bedding shows tops. Arrow shows location of Fig. 3-10B.



(B) Schistosity parallel to the axial surfaces of the F_2 folds in Fig. 3-10A cutting bedding (S_0) and schistosity (S_1) in the hinge of the minor fold in Fig. 3-10A. The axial planar schistosity has been accentuated by two black lines. It forms the principal schistosity in the pelitic beds of the Northfield Fm. in this area. Location, along power line ½ mile north of Williamsville Station, Dummerston.

Figure 3-11. Refolded fold in phyllite and thin quartzites in the Gile Mountain Formation; corner of Elliot and Union Streets, Brattleboro. Marking pencil is approximately 5 inches long. View is looking north. Early fold axis, N.15°E. at 8°; axial surface, N.8°E., dip 34°SE. One bedding surface in the early fold closure has been highlighted. Later fold axis, N.20°E. at 15°; axial surface, S.73°E., dip 16°NE. Fold is same one sketched by Moore (1949, p. 1655).

more easterly than the earlier bedding and schistosity. Here, S_3 is well developed and grades into a schistosity. The rocks are highly crinkled in this area. South of the Guilford dome, S_3 is also prominently developed between the Guilford and Colrain domes, where it has a more easterly trend than it has west of the Guilford dome, with dips moderate to steep to the north or south.

The sequence of younger folding generations in the Brattleboro syncline is difficult to establish, since the minor folds show a great range of orientations and, except for the early isoclinal folds, most fold the same S surface. Because of the low metamorphic grade,

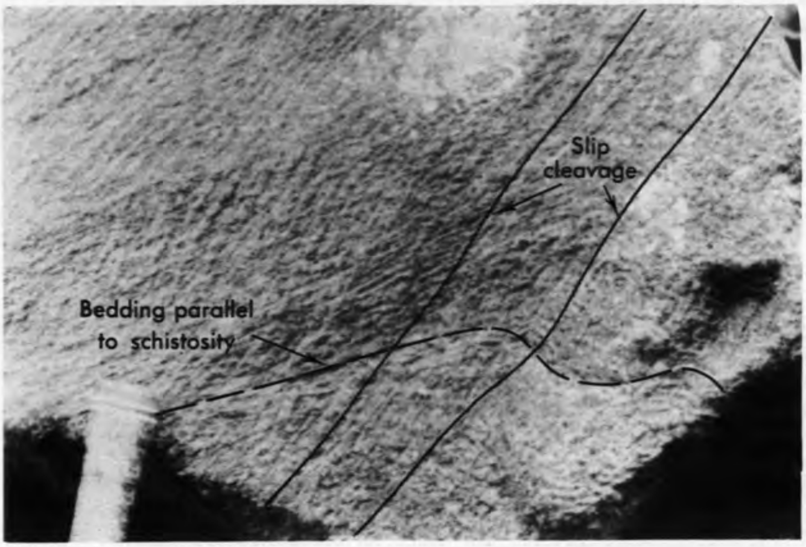


Figure 3-12. Slip cleavage developed parallel to the axial surface of an open F_3 fold, Northfield Fm., Halifax. Schistosity parallel to bedding is folded. After Hepburn, 1975.

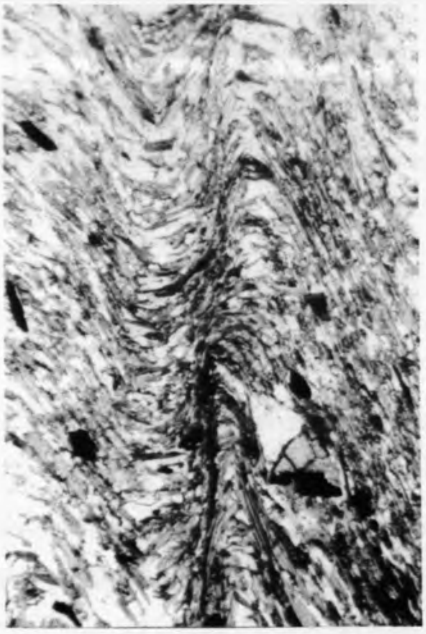


Figure 3-13. Photomicrograph of crinkles and slip cleavage in mica schist from the Northfield Fm. Slip cleavage parallels the axial surfaces of the F_3 folds. Magnification 40X.

new mineral growth and alignment of minerals parallel to the axial surfaces of folds did not take place readily. Trask (1964) believes the rocks in this area were somewhat brittle during the later stages of deformation, causing more localized areas of strain to develop. Because of this, it is difficult to relate the sequence of minor folds in the Brattleboro syncline to that in the western portion of the quadrangle, and the possibility exists that sequences are not equivalent.

In general, minor fold generations younger than the F_2 isoclinal folds in the Brattleboro syncline (Table 3-1) are flexure folds, with slip occurring along the cleavage (S_2) that developed parallel to the axial planes of the F_2 folds. The later folds are fairly open, and many approach a chevron shape. As a general rule more steeply plunging folds offset ones with shallower plunges. Likewise, folds with steep-dipping axial surfaces appear to be younger than those with moderate-to shallow-dipping axial surfaces. Minor folds in the Brattleboro syncline correlated with the F_3 stage are moderately open flexure slip-folds with generally shallow plunging axes (Figure 3-14). Crinkles are commonly developed parallel to the axes of these folds. In some cases (Figure 3-11) these folds are nearly homoaxial with the F_2 folds, although the axial planes commonly form at moderate angles to one another.

F_4: The youngest folds throughout most of the area are small-scale open warps or buckles in the schistosity and slip cleavage. These belong to one or more stages of minor folding and have a variety of orientations, of which the most common are sub-horizontal axes and gently to moderately dipping axial surfaces. Crinkle lineations may be developed in the hinge areas of these folds.

F_5: Only in the Brattleboro syncline were folds younger than F_4 observed (Table 3-1). The folds have moderate to steep plunges, commonly to the northeast, with steeply dipping axial surfaces. Most of the folds have a sinistral map pattern, showing an east-side-north sense of movement direction. The easternmost band of the Putney Volcanics has been offset by a series of these F_5 folds that are small but still large enough to be shown on Plate I.

Kink bands (Figure 3-15) are also relatively common in this area and probably are related to this final period of folding. They strike approximately N.40°E., across the trend of the Brattleboro syncline, and have axial surfaces that dip 40°-70° southeast. The kink bands parallel the axial surfaces of very open steeply plunging minor folds that have east-side-north movement patterns. The kink

Figure 3-14. Open, flexural folds in the Gile Mountain Formation. Grease pencil is 5 inches long. Axis plunges 18°N.20°E., axial surfaces strike N.48°W., dip 20°NE. Location, intersection of Green St., and Western Ave., Brattleboro.

bands and these open folds either are related to the mappable F_5 folds or are younger, possibly related to post-tectonic faulting as has been suggested by Rumble (1969).

Lineations.

Lineations are common throughout the area. Hinge lines of folds, crinkles, quartz-rods, and quartz-feldspar streamings have been observed parallel to the axis of the Prospect Hill fold. North-plunging hornblendes and stretched quartz-pebbles are found in the Russell Mountain Formation. Crinkles are the most common lineations and are nearly ubiquitous in the more pelitic rocks. They parallel the axes of the F_2 recumbent folds and subsequent fold generations. In areas where more than one cleavage is present,

Figure 3-15. Kink-bands in the Putney Volcanics, ¼ mile south of East Dummerston Village. Pen lies approximately north-south. Kink-bands strike N.40°E., dip 47°SE.

crinkles form at the intersection of the cleavage planes, and two or more sets of crinkle lineations are common. Multiple sets of crinkles are particularly well developed in the pelitic rocks of the Northfield Formation west and southwest of the Guilford dome and in the Brattleboro syncline.

Black Mountain Granite.

The Black Mountain Granite, correlated with the New Hampshire Plutonic Series (Billings, 1956), intruded into the Waits River Formation in the axial region near the northern end of the Guilford dome (Plate I). The ovoid outcrop area of the body is 2½ miles long in a north-south direction and 1¼ miles wide at its widest point. The contacts are sharp and steeply dipping, and cut both the bedding and schistosity. Numerous dikes and sills of the granite (Fig-

Figure 3-16. Outcrop showing numerous dikes and sills of granite intruding the Waits River Formation near the main Black Mountain Granite contact, roadcut on Vt. Rt. 30, Dummerston.

ure 3-16) and a relict (cf. retrograde?) contact metamorphic aureole occur within hundreds of feet of the main body of the intrusion. Larger sills to 50 feet thick occur near the body on the east and northeast sides. The dikes and sills have knife-sharp contacts and tend to divide the country rock into large blocks near the granite contact. On the basis of this relationship, piecemeal stoping (Daly, 1933, p. 270) appears to be the principal means of emplacement of the pluton, although xenoliths are not abundant in the main body itself. There is no evidence to suggest the granite was forcefully injected.

Although most of the regional deformation occurred prior to the granite intrusion, the granite body and many of the medium-grained dikes and sills have a weak foliation produced by the alignment of fine-grained mica flakes. The foliation has a steep to vertical dip and roughly parallels the strike of the schistosity in the surrounding rocks. It appears to have developed during the waning stages of the second major stage of deformation (i.e., doming). Irregular flow banding is also present in the granite body, particularly near the contacts. Some of the larger dikes show both flow banding parallel to the contacts of the dike and foliation approximately parallel to the regional schistosity that trends across the dikes.

Late-stage aplite dikes and small pegmatitic lenses cut both the

Figure 3-17. Sheeting well developed in the Black Mountain Granite. Note increased thickness of the sheets with depth. Sheets dip 30° to 40° to the west; abandoned Presbrey-Leland quarry, West Dummerston.

granite and the surrounding rock. The mica foliation was not observed in the aplites, but this could be a function of grain size. Just west of the abandoned Presbrey-Leland Quarry in West Dummerston, small quartz-rich aplite dikes to several inches thick cut the granite body along an apparent joint set that strikes N.35°W. and dips 55°NE., indicating that at least a portion of the granite body had cooled enough to "crack" before these aplites were intruded. Many small irregular sheer zones are present throughout the granite body.

Excellent sheeting is present in the Black Mountain Granite (See Dale, 1923). It is better developed in the northern part of the body than the southern. Along the western side of the body it dips 30° to 40° to the west and northwest. Figure 3-17 shows the excellent development of sheeting in the Presbrey-Leland Quarry (abandoned), West Dummerston, which permitted the relatively easy

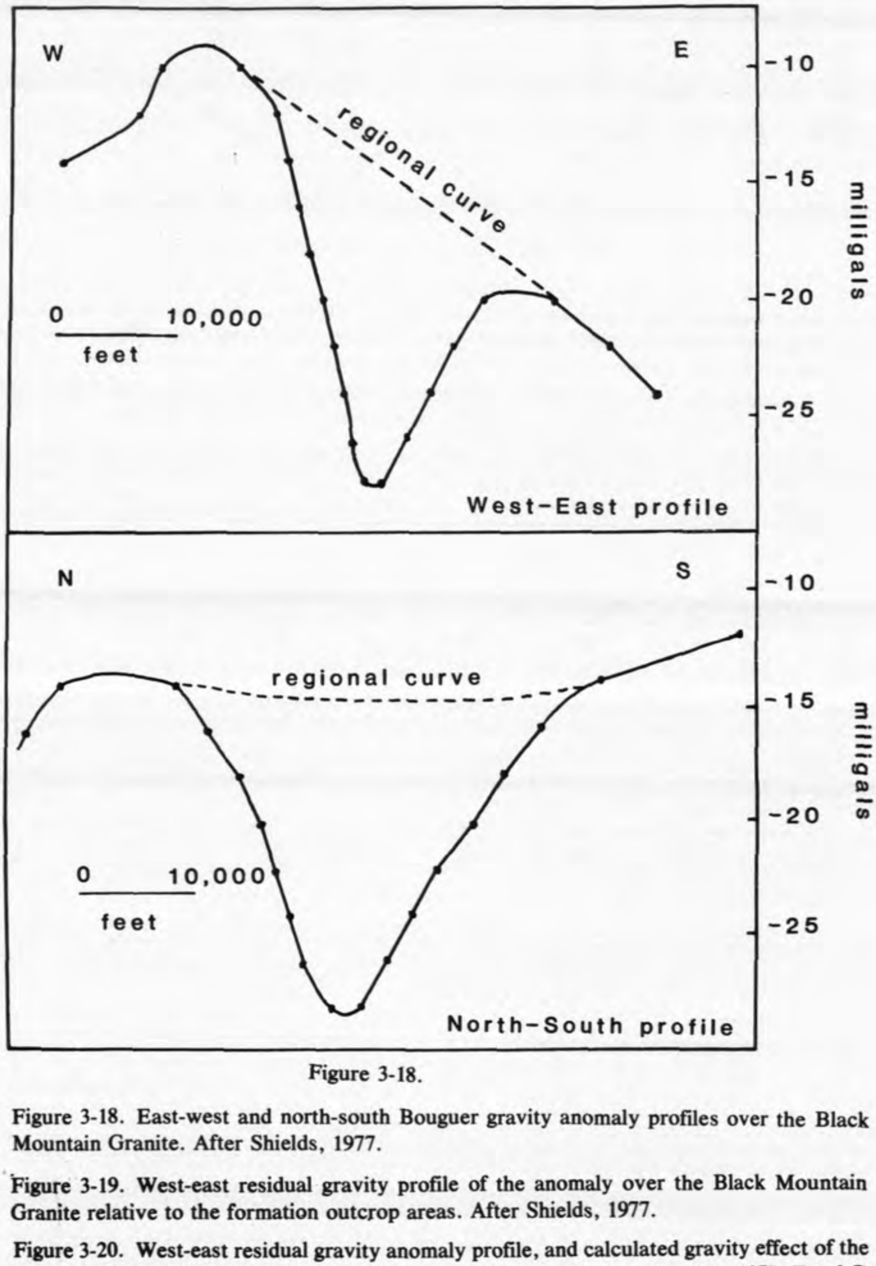


Figure 3-18.

Figure 3-18. East-west and north-south Bouguer gravity anomaly profiles over the Black Mountain Granite. After Shields, 1977.

Figure 3-19. West-east residual gravity profile of the anomaly over the Black Mountain Granite relative to the formation outcrop areas. After Shields, 1977.

Figure 3-20. West-east residual gravity anomaly profile, and calculated gravity effect of the most satisfactory 2-D model for the Black Mountain Granite, from Shields, 1977. For 3-D model refer to Shields, 1977.

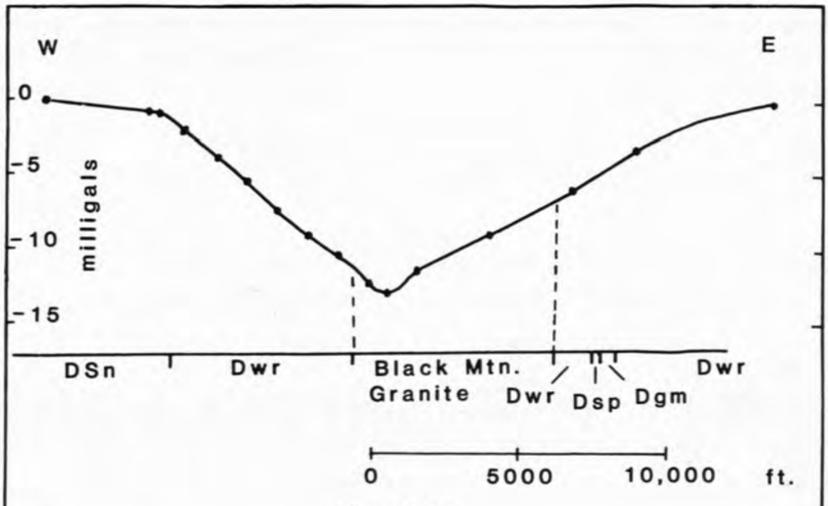


Figure 3-19.

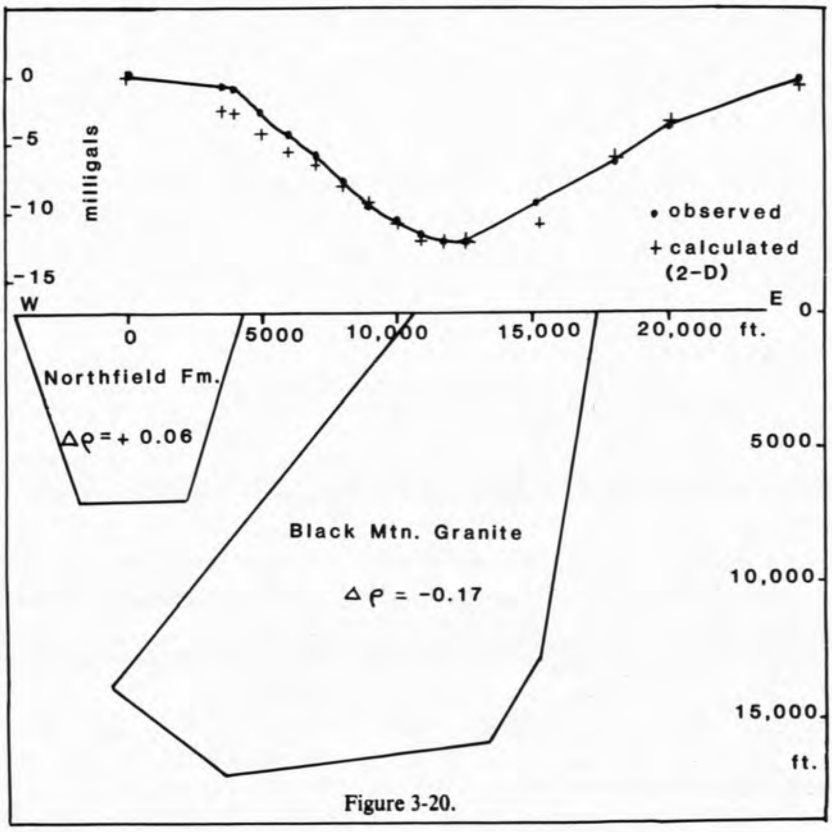


Figure 3-20.

removal of large blocks of granite during quarrying operations. The individual sheets increase notably in thickness with depth.

The Black Mountain Granite has characteristics similar to those that Page (1968) assigns to his early post-tectonic subdivision of the New Hampshire Plutonic Series. Naylor (1971) dated coarse unaligned muscovite flakes from the Black Mountain Granite that appear to cut the earlier foliation-producing micas. He interprets the Rb-Sr dates of 377 and 383±7 m.y. (early Middle Devonian) as establishing a minimum age for the intrusion. This date on the unaligned micas indicates the Black Mountain Granite is a late Acadian synorogenic intrusion and establishes a minimum age for the major deformation in the area.

Several other small granite to granodiorite bodies are mapped in the area (Plate I), particularly in the Brattleboro syncline. They range from unfoliated to moderately foliated. Several granitic dikes, too small to be shown on Plate I, were observed near the crest of the southern lobe of the Guilford dome. They are unfoliated except for flow banding parallel to the dike walls.

Gravity Model of the Black Mountain Granite. Shields' (1977) Bouguer gravity profile (Figure 3-18) and model of the Guilford dome area provides insight into the shape of the Black Mountain Pluton at depth. This study has shown that a −12 milligal anomaly exists over the Black Mountain Granite (Figure 3-19) that can be explained if the body extends beneath the surface for 17,000 feet (5.18 km) (Shields, Johnson, and Hepburn, 1978). Shields' two dimensional (Figure 3-20) and three dimensional gravity models indicate the pluton has a nearly vertical eastern contact while the western contact dips to the west at approximately 55°, i.e., the body has been intruded more or less along the steeply west dipping axial surface of the Guilford dome. The granite body also dips steeply to the north and south.

Shields (1977) also found a small negative gravity anomaly of −3 milligals centered on the sorthern lobe of the Guilford dome. Using the measured density contrast of −0.17 g/cc between the Black Mountain Granite and the surrounding Waits River Formation, the southern anomaly could be explained by a similar granitic body existing some 3,200 to 4,400 feet below the surface with a thickness of less than 15,000 feet. Such an explanation could also account for the increased number of granitic dikes observed in the southern lobe of the Guilford dome. However, Shields (1977) also notes that the rise of felsic rocks, such as those in the Barnard Volcanics, could also explain the anomaly in this part of the dome.

Tectonic Synthesis

Major Structures.

The outcrop pattern of the Prospect Hill fold around the Guilford dome indicates that doming followed emplacement of the recumbent fold (Figure 3-21). The original hinge of the Prospect Hill fold had a strike of approximately N.45°E. in the north-central part of the Brattleboro quadrangle. Interestingly, Trask (1964, p.83) indicates that the hinge of the Bernardston-Skitchewaug nappe, in the adjoining Vernon-Chesterfield area of the Bronson Hill anticlinorium, also locally trends about N.40°E. The Prospect Hill fold is probably a continuation of the Ascutney sigmoid (Rosenfeld, 1968; Doll et al., 1961). If this is true, the hinge of the Prospect Hill fold must turn more northerly a short distance north of Prospect Hill in order to connect with the Ascutney sigmoid in the Saxtons River quadrangle. The Prospect Hill fold also continues south-southwest from the Guilford dome area and reappears in the Colrain dome (Rosenfeld, 1968).

The exact amplitude of the Prospect Hill fold is impossible to determine, but it is at least on the order of several miles in the southern Guilford dome. The direction of tectonic transport of the Prospect Hill fold is related to the problem of the stratigraphic facing between the Gile Mountain and Waits River Formations. If the Gile Mountain is younger than the Waits River, as has been assumed here, then the Prospect Hill fold is the recumbent syncline beneath an east-facing recumbent anticline. The anticline hinge of this recumbent fold would be the early synformal hinge east of the Guilford dome. The Fall Brook anticline and the tongue of Northfield Formation encircling the northern end of the Guilford dome are interpreted to be in the axial region of this recumbent anticline. The anticlinal hinge could possibly parallel the hinge of the Prospect Hill fold, striking approximately N.45°E. and plunging northeast and southeast, away from the hinge of the Guilford dome. Southwest of the dome, the anticlinal hinge would turn more to the south.

In one interpretation based on rotated garnet studies, Rosenfeld (1968) had the Ascutney sigmoid formed by intrastratal westward flow of the Waits River Formation beneath the Standing Pond of the Prospect Hill fold. This would produce the east-facing recumbent fold by having the major tectonic transport direction to the west, similar to that in the nappes of the Bronson Hill anticlinorium.

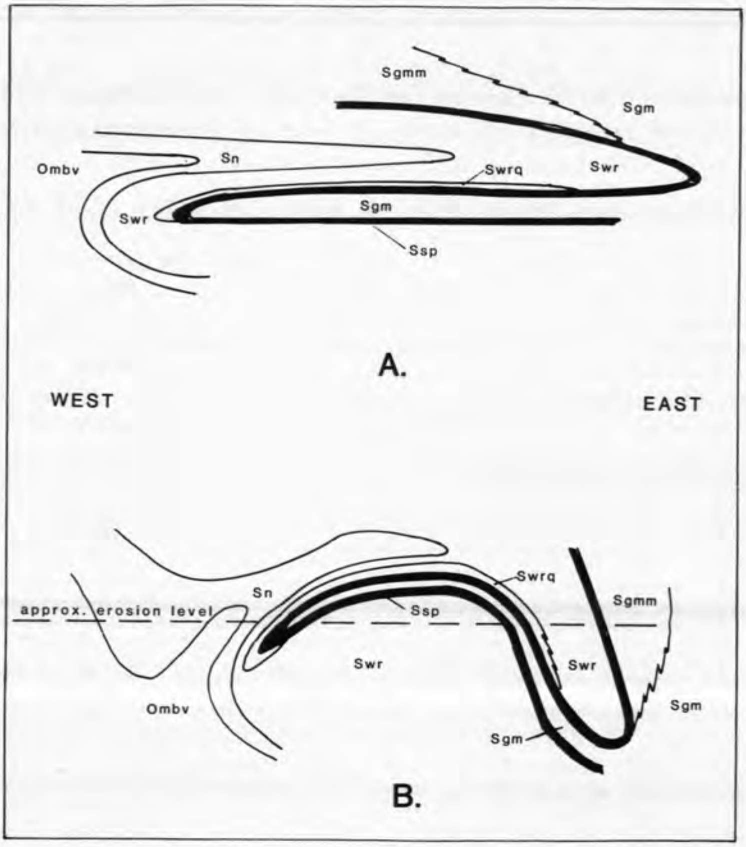


Figure 3-21. Schematic cross sections showing the evolution of the structural features in the two major stages of deformation in the Guilford dome area. The Standing Pond Volcanics are shown in black. (A) The Prospect Hill fold at the end of the first major stage of deformation, before the rise of the Guilford dome. (B) Prospect Hill fold following the second major stage of deformation, after the rise of the Guilford dome. The horizontal line represents the erosion surface. Formations: Sgm = Gile Mt. Fm.; Sgmm = marble mbr., Gile Mt., Fm.; Swr = Waits River Fm.; Swrq = quartzitic mbr., Waits River Fm.; Sn = Northfield Fm.; Ssp = Standing Pond Volc. (in black); Ombv = Barnard Volc. Mbr., Missisquoi Fm. Modified in part from Hepburn, 1975.

The minor structural data support the interpretation that the Fall Brook anticline and Northfield Formation north of the Guilford dome are part of the recumbent anticline. Fold axes and lineations in the Fall Brook anticline strike northeast, parallel to the axis of the Prospect Hill fold. The fold axes and lineations in the Northfield Formation north of the Guilford dome also parallel the fold axes and lineations in the synclinal hinge on Prospect Hill. The minor folds in the amphibolites of the Barnard Volcanic Member of the Missisquoi Formation in the core of the Fall Brook anticline have a style similar to those in the Standing Pond Volcanics in the hinges of the Prospect Hill fold. In both of these localities, the minor folds associated with the major structure fold an earlier schistosity, S_1. In the schists and thin quartzites of the Northfield Formation, the principal schistosity is parallel to the axial planes of the F_2 isoclinal folds. S_1 is only questionably present. This is believed to be the result of the thorough reorientation and recrystallization of the micaceous minerals in the schistose rocks parallel to the axial planes of the F_2 isoclinal folds.

The outcrop width of the Northfield Formation in the north-central Brattleboro quadrangle is at least twice as great as its width in the rest of southern and central Vermont (Doll et al., 1961). The width increases where the trace of the proposed recumbent anticlinal fold joins the main Northfield belt. The Northfield-Goshen belt continues southward into Massachusets at this increased width (Hatch and Hartshorn, 1968; Hatch et al., 1967), suggesting the possible continuation of this fold.

Upright isoclinal folds with wavelengths of 60-600 m (200 to 2,000 ft.) have been mapped by Hatch (1968, 1975) in the Goshen Formation, the Northfield equivalent in western Massachusetts. Here the Goshen has a greater abundance of graded arenaceous beds than does the Northfield Formation in the Guilford dome area. It is believed that these folds (Hatch, 1975, stage II) are the same generation as the upright isoclinal folds in the Northfield Formation of the Guilford dome area, here assigned to the F_2 stage. However, based upon observations in western Massachusetts, Osberg (1975) presents a case for placing these upright isoclinal folds (Hatch's stage II) into a different folding sequence, whereby they follow a recumbent folding but precede F_3. If these upright isoclinal folds in the Northfield-Goshen belt are of the F_2 stage, then the recumbent folding Osberg proposes in Massachusetts is earlier than the recumbent folding (F_2) of the Guilford dome area. However, if these two recumbent folding stages are equivalent, it would be necessary to have an additional folding stage present in the Guilford dome area ($F_{2.5}$) which produced the upright isoclinal fold in the Northfield.

In the Guilford dome area, evidence suggesting an additional isoclinal folding stage, not equivalent to the F_2 recumbent folding has not been found. This however, could be due in part to the lack of well-defined bedding in the Northfield Formation in this area. Also, refolding of the Northfield Formation west of the Guilford dome during the second major stage of deformation has reoriented the isoclinal minor folds associated with F_2 into upright attitudes that would be nearly identical in orientation to the folds Osberg purposes. Thus it would be difficult to distinguish if two isoclinal fold sets were present in this area. In the tongue of the Northfield Formation immediately north and northeast of the Guilford dome, the isoclinal folds plunge northeastward, parallel to the axis of the Prospect Hill recumbent fold. This supports the interpretation that the Northfield isoclinal folds are congruous with major recumbent folding stage (F_2) and are not a later, north-trending fold generation.

The slip cleavage (S_3) that is well developed in the Northfield Formation parallel to the axial surfaces of the F_3 folds is also a prominent minor structural feature in the rest of the Northfield-Goshen belt to the south. In the Guilford dome area, it is believed that this slip cleavage formed contemporaneously with the rising Guilford dome. The slip cleavage is best developed to the south and west of the dome, where the rocks were highly flexed during the doming. Maximum flexure in this area would be expected if uplift occurred simultaneously in the Green Mountain anticlinorium and Guilford and Colrain domes.

Doming.

Thompson (1950), Rosenfeld (1954, 1968), and others have concluded that the Chester and Athens domes were formed by nearly vertical diapiric upward movement of the core rocks, possibly the result of lower specific gravities of these rocks. This is also a likely explanation for the formation of the Guilford dome. The impure marbles of the Waits River Formation in the core of the dome are extensively folded and were undoubtedly more plastic and probably less dense than the surrounding rocks. Daly et al. (1968) indicate the average density of a marble is generally less than that of a metamorphosed pelitic rock, the inverse of the relationship between an unmetamorphosed limestone and a shale. This could have been a factor in the doming (see also Shields, 1977). A metamorphic high centered on the Guilford dome (Plate I) also indicates these rocks were hotter than surrounding rocks presently exposed at the surface. The role of the Black Mountain Granite in the doming is unknown. It cuts the surrounding rocks at the present level of erosion and apparently at depth (Shields, 1977). However, the negative gravity anomaly centered on the southern lobe of the Guilford dome could be caused by granite at depth. Similar negative gravity anomalies also occur in the Pomfret and Strafford domes (Bean, 1953). Granite plutons that intrude the Strafford-Willoughby arch in central and northern Vermont are similar to the Black Mountain Granite in composition, method of intrusion, and general structural setting (Murthy, 1957; Goodwin, 1963). Goodwin also notes the possibility of a relationship between the arching and the granitic intrusions in this area.

Recent work by the COCORP seismic reflection project in New England (Ando et al., in press) traversed the area across the Chester dome some twenty miles north of the Brattleboro quadrangle. Preliminary analysis indicates the possible presence of horizontal reflectors at depth beneath the Green Mountain anticlinorium and Chester dome. These may be interpreted as zones along which large scale horizontal transport has taken place. East-dipping reflectors at depth below the Connecticut River Valley and Bronson Hill anticlinorium may indicate the location of deep-seated tectonic ramps. Clearly, mechanisms for the formation of the geologic structures in the Brattleboro area must be reconciled with these seismic observations when more complete interpretations of the seismic profiles become available.

An Alternative Tectonic Interpretation — Backfolding Accompanying Doming (by J.L.R.).

[The following is an alternative interpretation of the structural development in the area by one of us (J.L.R.) based, in part, on studies of rotated garnets from areas outside the Brattleboro quadrangle. This interpretation has a different stratigraphic sequence for the Silurian than presented elsewhere in this paper (cf. Plate I, Figure 2-1), placing the Gile Mountain Formation stratigraphically below the Waits River Formation. However, the kinematic interpretation given below is not dependent on the stratigraphic facing (cf. Rosenfeld, 1968, p. 200).]

It is possible to arrive at the cross section in Figure 3-21B by an alternative route that does not do violence to known fact and that accords with other facts not taken into account above. Study of rotated garnets in the contiguous area to the north, and integration of that information with other information obtained from both geologic mapping and observations of minor structural evidence demonstrated that the Acadian orogeny in that area consisted of two distinct events, diastrophic events I and II (Rosenfeld, 1968). Those two events, corresponding to nappe formation and doming respectively, are characterized in a cross section of the Chester

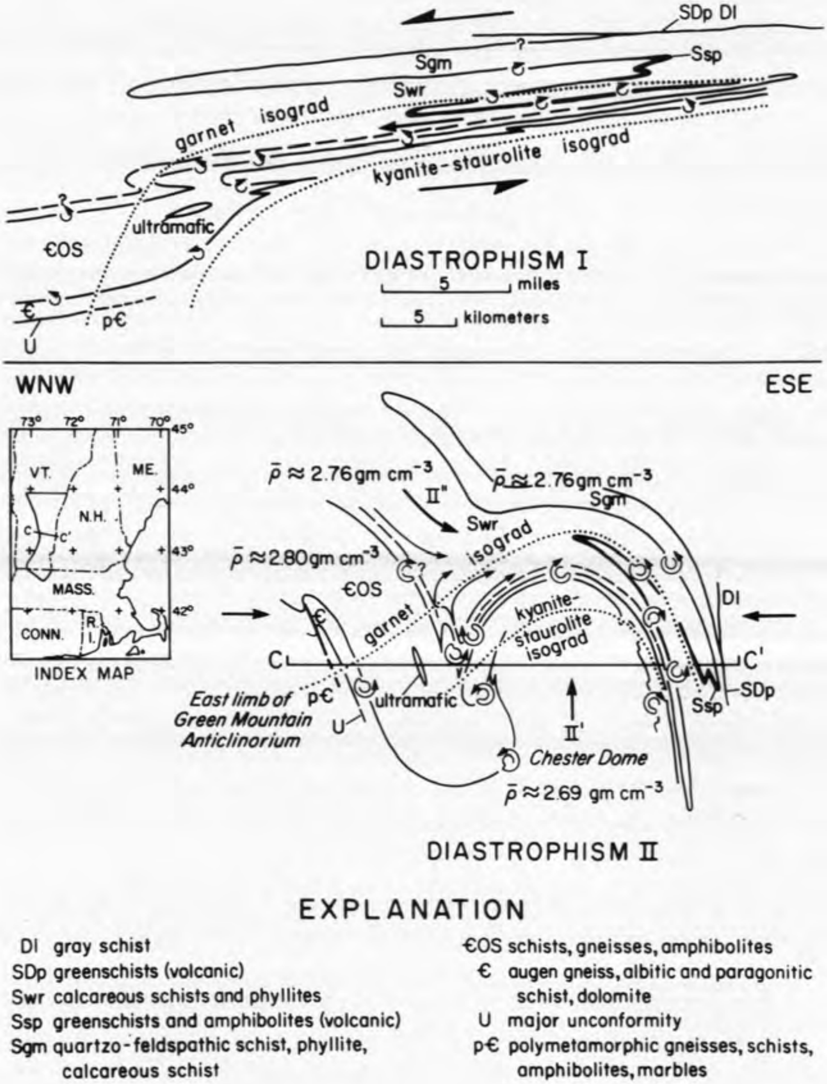


Figure 3-22. Schematic model for Paleozoic diastrophism in southeastern Vermont. Modified from Rosenfeld, 1968, Figures 14-10, 14-11.

Dome in Figure 3-22, updated from Figures 14-10 and 14-11 of Rosenfeld (1968) to accord with new mapping. There was no discontinuity in the metamorphism between the two events. An unpublished part of the earlier study was an extended check of similar evidence along the apparent extension of the Prospect Hill fold both north to the vicinity of Strafford Village, Vermont, and south, through the Brattleboro quadrangle, to the approximate latitude of Greenfield, Massachusetts, a span of about 160 kilometers. While this is not the place to discuss the extended evidence, suffice it to say that the findings were consistent with those determined around the Chester and Athens domes. There was no evidence requiring change of kinematic sequence and style. A principal difference, both north and south of the Chester and Athens domes, was the apparent later rise of the isogradic surfaces relative to stratigraphic surfaces, evident in the findings that the rotated garnets commonly do not show the large internal "snowball" rotation common in garnets around the Chester and Athens domes. In some instances the garnets show little or no rotation, indicating that they grew very late relatively to the Acadian orogeny. In a few instances the garnets may have grown during event I; and event II, the lesser deformation, may have been too weak to express itself.

In the context of the Brattleboro area, two questions arise: (1) Can the sequence of processes implied from the rotated garnets and associated features of the Chester and Athens domes be applied to the interpretation of the structures associated with the Guilford dome in a way compatible with the tectonometamorphic facts and certain non-controversial limited or short-range interpretations derived from the facts? (2) If that sequence can be applied, how can it be applied?

First consider Hepburn's (1975) treatment as a test of the first question. Attention will focus on the schematic cross sections in Figure 42 of Hepburn (1975), slightly modified here as Figure 3-21 (see also Figure 4-8). Based on the field geology, cross section B while certainly complex, would seem to be a necessary proximate interpretation. Hepburn then interpreted its antecedent state, A in Figure 3-21, as that of a large recumbent sigmoid isocline with top side displaced toward the east with the fold incorporating units down to the Barnard Volcanic Member of the Missisquoi Formation. The lower fold of the isocline would have been the pre-doming Prospect Hill fold and the upper fold, the predoming Fall Brook anticline. Thus the interpretation in Figure 3-21A treats similarly the pre-doming kinematics of the Prospect Hill fold, involving the

Standing Pond Volcanics, and the Fall Brook fold, involving the older and lower Barnard Volcanic member. They are both considered complementary parts of a single sigmoid fold with upper rocks translated east relative to lower rocks. While the interpretation in Figure 3-21A accords with the sense of rotation of garnets to the north at the level of the lower limb of the Prospect Hill fold in the Standing Pond Volcanics, that interpretation is discordant with the event I rotational information at the level of the base of the Northfield Formation on the east side of the Chester and Athens domes, including the west end of cross section B in Figure 3-21. At the latter horizon, event I garnet rotations were large and had a rotation sense opposite to that implied in Figure 3-21A (cf. Rosenfeld, 1968, Figure 14-4a). Thus the interpretation implied by Figure 3-21A would seem to be incompatible with the evidence from the rotated garnets, and an answer to the questions raised in the previous paragraph must be derived in another way.

Examination of the sequence of motions derived primarily from the rotated garnets on the Chester and Athens domes, illustrated in Figure 3-22, allows an essentially identical kinematic sequence to lead to a cross section conformable with that in Figure 3-21B. An example of evidence of that sequence from the Brattleboro area appears in Figure 3-23, modified from Rosenfeld (1970, Plate 15). Figure 3-23 is a view of a thin section, oriented to look due south down the 36° dip of the Pinney Hollow Formation where that unit wraps around the south end of the Athens dome. The schistosity in the snowball garnets in Figure 3-23 shows that, during growth, the top strata moved westerly relative to those beneath. The continuation of the schistosity into the exterior shows post-growth rotation with the opposite sense for both the garnets and their attached earlier quartz pressure shadows. The sequence of motions is identical to that observed far to the north, where the Waits River Formation wraps around the north end of the Chester dome.

Figure 3-24 represents an alternative interpretation to that in Figure 3-21, in which the sequence of motions is consistent with that mentioned in the previous paragraph. In this interpretation, the Gile Mountain Formation is older than the Waits River Formation, an inference based on discovery to the north of the Brattleboro quadrangle by one of us (J.L.R.) of an intermittent, thin zone of Standing Pond-type rocks between the extension of the Prospect Hill fold and the base of the Waits River Formation. Event I has the Prospect Hill fold in its early stage satellitic to the Cornish nappe of Thompson et al. (1968) and the early-stage Fall Brook fold as a

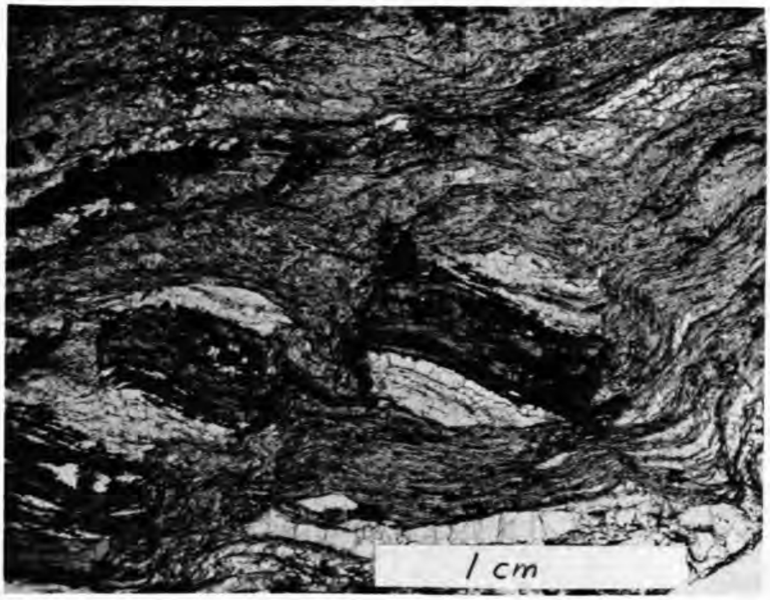


Figure 3-23. Photomicrographic view looking due south down the 36° dip of the Pinney Hollow Fm., south end of Athens dome, northeast side of the brook next to the road 1000 m southwest of the outlet of Kenny Pond.

small nappe overturned to the west (position A in Figure 3-24A). Event II west of the developing Guilford dome appears to include an easterly flow of schists off of the structurally higher Green Mountain anticlinorium, across the Townshend-Brownington synclinorium, and over the Athens dome. The Fall Brook fold in its present placement then becomes a backfold (position A′ in Figure 3-24B), flipped over onto the Guilford dome by the easterly flow.

Backfolding, first described by the famous Swiss geologist, Emile Argand, early in this century in the Central Alps, has considerable interest for students of tectonics. This interest stems from the extension of its causative flow over the conspicuous anticlines and synclines visible at the earth's surface. This overriding indicates operation of a larger scale tectonic process. The causative flow across the surface structures (Rosenfeld, in press) is hypothesized to result from gravitational response to tilting of the lithospheric substrate of the surface structures. That view in turn has the tilting consequent on lithospheric overthrusting, along a nascent subduction zone between colliding continents, that is arrested due to the buoyancy of the low density continental crust that is being underthrust. The asymmetric gravity anomaly across the Taconic and Green Mountains (Diment, 1968), the more recent

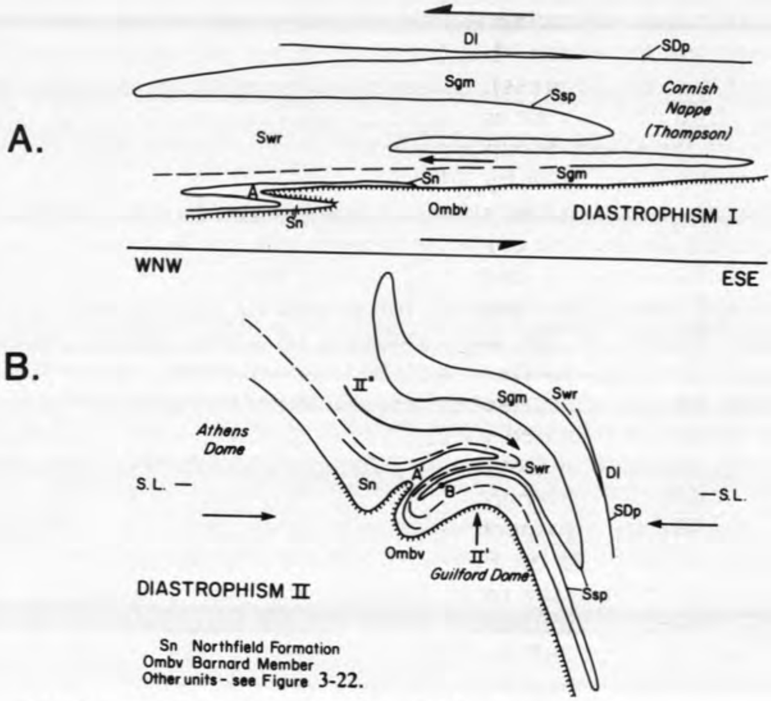


Figure 3-24. Alternative interpretation of Figure 3-21, see text.

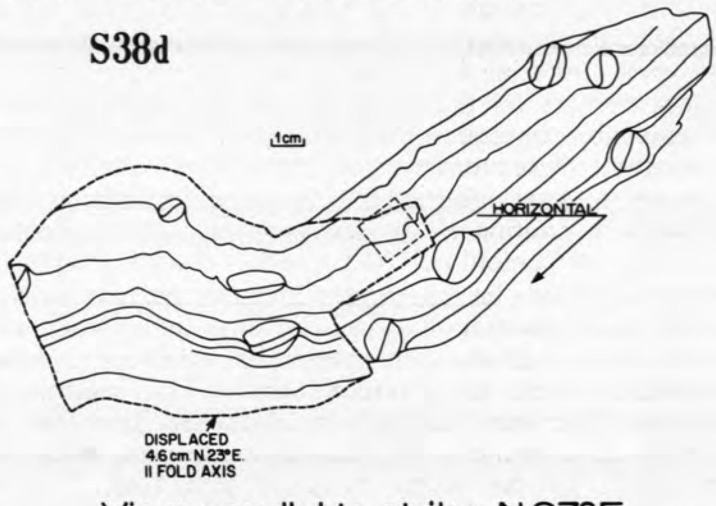


Figure 3-25. Tracing on a sawed surface of representative schistosity traces and rotated garnets from the west side of the road along the east side of Governors Mountain, approximately 70 meters south of the Guilford-Brattleboro town line.

COCORP data indicating easterly dipping seismic reflectors extending from the Green Mountains to a depth of approximately 30 km (Ando et al., in press), the overturning of the axial surface of the Guilford dome to the east, and the well-known easterly dipping thrusts of the Green Mountain region are supportive of or at least consistent with such an hypothesis.

At a less transcendent scale, drag folds and to a lesser extent rotated garnets provide evidence of upthrusting of both the Chester and Athens domes at deeper structural levels. This upthrusting could well result from buoyant forces due to the relatively lower average density of the core gneisses of those domes. Similar kinematic evidence exists within the Guilford dome. Figure 3-25 is a tracing on a sawed surface of a specimen of representative schistosity traces and rotated garnets from the west side of the road along the east side of Governors Mountain, about 70 meters south of the Guilford-Brattleboro town line. The specimen is from the lower limb of the Prospect Hill fold, and its structural position is indicated as point B on Figure 3-24B. For consistency with the model derived from the rotated garnet sequence to the north, we should expect a northward view of the specimen to reveal clockwise rotation followed by counterclockwise rotation. The specimen shows a drag fold with its short limb rotated clockwise and garnets with largely post-growth counterclockwise rotation. The simplest interpretation would seem to be the following. Before the rise of the garnet isograd through that volume of rock, a drag fold developed, reflecting a clockwise rotation. Subsequently, during a quasi-static interval, the garnets formed, perhaps as a result of thermal relaxation consequent on intrusion of the Black Mountain Granite nearby. Subsequently the Guilford dome developed, accompanied by buoyant upthrusting. The latter would have caused counterclockwise rotation of the garnets at this structural position and caused partial "unfolding" of the earlier clockwise drag fold. The area of the square inset in Figure 3-25, shown here as Figure 3-26, confirms the reversal of shear. The small fold, satellitic onto the fold of Figure 3-25 shows a splaying of schistosity, volume-compensated by quartz (cf. pressure shadow). This indicates that the secondary fold was originally a clockwise drag fold later sheared with the opposite sense, thereby causing the initial short limb to splay. In the sharp nose of that fold, bytownite pseudomorphs of muscovites are included in a less calcic plagioclase porphyroblast that overgrows the fold post-tectonically. The reaction resulting in this interesting intracrystalline cation ex-

Figure 3-26. Photomicrograph of square inset in Figure 3-25. See text.

change ($3\ Ca^{+2} - 2K^{+} - 4H^{+}$) supports the belief that a major contributor to the thermal maximum was the Black Mountain Granite, which intruded quite late in the tectonic sequence.

Not all the drag folds at structural positions near the underside of the Prospect Hill fold on the west side of the Guilford dome show the good evidence of shear reversal illustrated in Figures 3-25 and 3-26. Nevertheless the odd shapes of some of the folds observed in that structural position become more understandable in terms of shear reversal. A particularly fine example of such a fold from the east bank of the West River west of Black Mountain, also in calcareous schist of the Waits River Formation, is illustrated in a northward view in Figure 3-27. It is most easily interpreted as an early clockwise drag fold that suffered a subsequent and probably smaller shear in the opposite sense in approximately the same plane.

Figure 3-27. Fold in calcareous schist of the Waits River Fm. showing shear reversal. See text. From east bank of the West River west of Black Mt., Dummerston.

CHAPTER 4

METAMORPHISM OF THE WESTERN TERRANE

Introduction.

The grades of regional metamorphism in the western Brattleboro area range from the greenschist facies to the amphibolite facies in a metamorphic series of the Barrovian type. The pelitic rocks in the area show systematic mineralogical differences with metamorphic grade; and isograds have been drawn (Plate I) on the basis of the first appearance of biotite, almandine, and staurolite or kyanite in the pelitic rocks. The chlorite and biotite zones occur near the center of the Brattleboro syncline and are continuous with a regional metamorphic low along the Connecticut River valley in Vermont and north-central Massachusetts (Doll et al., 1961;

All assemblages contain quartz and muscovite

Chlorite zone

Chlorite
Chlorite-potassium feldspar
Chlorite-spessartine rich garnet

Biotite zone

Biotite
Biotite-chlorite
Biotite-chlorite-potassium feldspar

Garnet zone

Almandine
Chlorite
Almandine-biotite
Chlorite-biotite
Almandine-chlorite-biotite

Staurolite-Kyanite zone

Biotite
Almandine-biotite
Almandine-biotite-potassium feldspar
Almandine-biotite-chlorite
Almandine-biotite-staurolite
Almandine-biotite-staurolite-chlorite
Almandine-biotite-stuarolite-kyanite
Almandine-biotite-staurolite-kyanite-chlorite

Table 4-1. Observed mineral assemblages in the pelitic rocks of the Western Brattleboro area.

Thompson and Norton, 1968). Westward from the Brattleboro syncline, the metamorphic grade increases to the staurolite-kyanite zone in the Guilford dome. Similar metamorphic highs occur in the northwestern corner of the area, at the southern end of the Athens dome and in the Colrain dome to the south. The remainder of the area is in the garnet zone of regional metamorphism, except for the rocks surrounding the Black Mountain Granite that have been additionally altered by contact metamorphism.

Pelitic Rocks.

Table 4-1 summarizes the mineral assemblages observed in the pelitic rocks. Minerals occurring together in a single thin section are considered to be in equilibrium unless otherwise stated. Isograds mapped on the basis of the first appearance of biotite, almandine, kyanite, and/or staurolite serve as a guide to the metamorphic conditions of the area, although it must be realized that the first

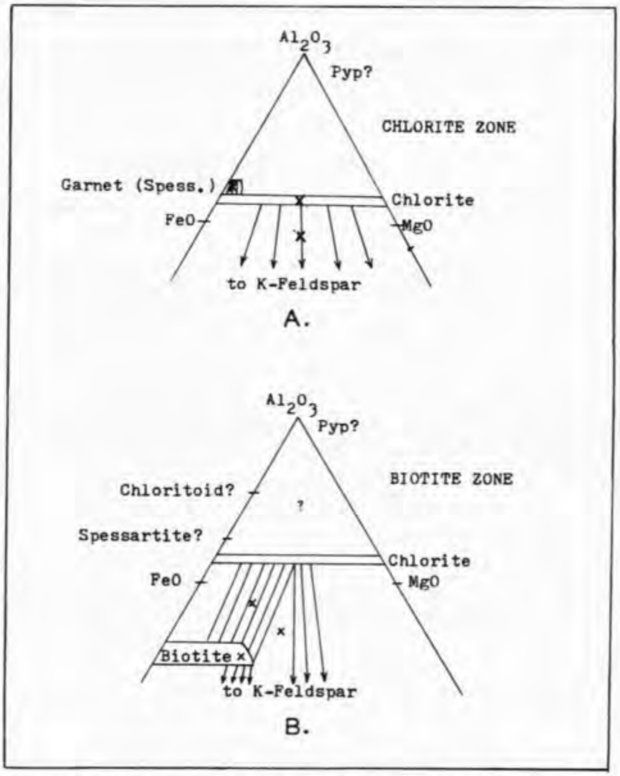


Figure 4-1. Observed mineral assemblages in the pelitic rocks of the Western Brattleboro area for the chlorite and biotite zones plotted on an AFM diagram (Thompson, 1957). X's indicate mineral assemblages present with quartz and muscovite.

appearance of these minerals is dependent on the bulk chemical composition of the rocks as well as on the externally controlled variables.

Chlorite Zone. The slates and phyllites of the chlorite zone (Plate I) have a relatively simple mineralogy (Figure 4-1A). Chlorite, and less commonly chlorite plus K-feldspar, along with quartz, muscovite, and plagioclase form the bulk of the rocks. One outcrop containing small spessartine-rich garnets was found in slates of the conglomeratic member of the Littleton Formation (D1c). Electron microprobe analyses of two of the garnets (Hepburn, 1972A) indicate they contain 35.4 percent and 27.4 percent of the spessartine component. Because of this high MnO percentage in these garnets and the lack of rocks containing normal almandine in the vicinity, these rocks have been included in the chlorite zone on Plate I.

Biotite Zone. The fine-grained nature of the phyllites makes the biotite isograd difficult to place precisely in the field. In thin section, biotite and chlorite, with or without K-feldspar, are found instead of chlorite and K-feldspar. This indicates that the first appearance of biotite in a given pelitic rock is probably the result of a shift in the AFM-projected three-phase field, chlorite-K-feldspar-biotite, across the projected bulk composition of the rock (Figure 4-1B). The reaction would be of the type: chlorite + K-feldspar → biotite + muscovite + quartz + H_2O. The geometry of the AFM projection above the chlorite composition is unknown, as highly aluminous rocks were not found. Chloritoid was not found in the Western Brattleboro quadrangle, although it has been identified in the Littleton Formation in the New Hampshire part of the area, (See Chapter 7).

Garnet Zone. Garnet-biotite and garnet-biotite-chlorite, with quartz and muscovite, are the most common garnet zone assemblages (Figure 4-2). The garnet isograd is drawn (Plate I) on the basis of the first appearance in pelitic rocks of almandine garnets not notably high in the spessartine component. The placement of this isograd is quite easy, as garnets become abundant within a few hundred feet of their first appearance. The garnets appear first in rocks with a high FeO/MgO ratio by a dehydration reaction of the type: chlorite + muscovite + quartz → almandine + biotite + H_2O.

Staurolite-Kyanite Zone. Figure 4-3 shows the AFM projection geometry believed to best represent the most common mineral assemblages observed in the pelitic rocks of the staurolite-kyanite zone (Table 4-1). Many of the rocks in the Waits River Formation near the staurolite-kyanite isograd are low in aluminum and have garnet-biotite as the common AFM assemblage. It is possible that the isograd mapped on the first appearance of staurolite or kyanite (Plate I) would be displaced outward if the rocks in this area were more aluminous. Even in areas "inside" the first appearance of staurolite, the assemblage garnet-biotite-chlorite-quartz-muscovite is still encountered. Staurolite, where found, coexists with biotite. Thus it is believed that the staurolite isograd, as mapped, is based on the reaction of almandine + chlorite + muscovite → staurolite + biotite + quartz + water. The mapped distribution of staurolite and kyanite shows that staurolite may appear slightly before kyanite. However, as noted, this could be a function of the bulk composition of the rocks, and no clear indication of which mineral phase appears first could be mapped consistently.

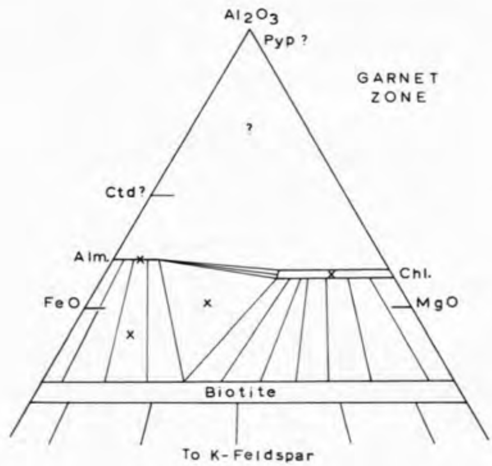
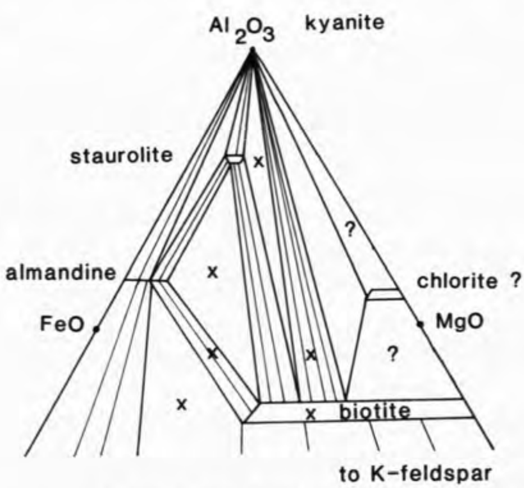


Figure 4-2. AFM projection (Thompson, 1957) for garnet zone mineral assemblages in the pelitic rocks of the Western Brattleboro area. Observed mineral assemblages with quartz and muscovite are indicated by X's. Pyp = pyrophyllite, Ctd = chloritoid, Alm = Almandine, Chl = Chlorite.



STAUROLITE-KYANITE ZONE

Figure 4-3. Observed mineral assemblages in the pelitic rocks of the Western Brattleboro area, Vermont, for the staurolite-kyanite zone of metamorphism plotted on an AFM projection (Thompson, 1957). X's indicate mineral assemblages present with quartz and muscovite. Topology shown for the AFM projection is believed to best represent the observed assemblages. See text.

Assemblages with four or even five coexisting mineral phases in the AFM projection are not uncommon in the staurolite-kyanite zone. Such assemblages are not believed to represent unusual variances in the rocks but may be explained in one of two ways: (a) either additional components, not considered in the projection, are present in one or more of the phases, particularly garnet, or (b) chlorite is a product of a retrograd reaction and did not form in equilibrium with the other minerals. Analyses of six garnets from the staurolite-kyanite zone in this area were given by Hepburn, 1972A. He found CaO and to a lesser extent MnO present in amounts large enough to displace the garnets off the AFM plane and allow an extra phase to be present. In addition, all of these analyzed garnets were compositionally zoned.

The extent to which the chlorite in the staurolite-kyanite zone is a product of retrograde alteration is difficult to determine. Much of it is present in large porphyroblasts without textural evidence to indicate whether it has replaced earlier ferromagnesian minerals. However, it commonly cuts the foliation. The list of assemblages in Table 4-1 does not include chlorite that has clearly formed as an alteration product, typically as rims on garnet or staurolite.

The AFM four phase assemblage garnet-biotite-staurolite-chlorite is most easily explained by having the chlorite formed at a later time than the rest of the assemblage. Kyanite-staurolite-biotite-garnet is a common projected assemblage and is undoubtedly due to the garnets lying slightly off the projection plane due to the presence of additional components such as CaO and MnO. The AFM five projected phase assemblage garnet-biotite-staurolite-kyanite-chlorite is also probably the result of the garnet, and possibly other phases, having additional components, while the chlorite may have formed at a later time. This assemblage has also been reported from northern Vermont (Albee, 1968).

The assemblage garnet-biotite-kyanite presents a problem. It is doubtful that it represents the breakdown of staurolite, since staurolite is widespread in this zone of metamorphism and has been found short distances from the above assemblage. It is believed more likely that this assemblage represents a garnet with additional components beyond those in the AFM projection, coexisting with the two phase projected field kyanite-biotite.

An unusual metamorphic paragenesis is observed locally near the contact of the Standing Pond Volcanics and Waits River Formation where coarse-grained hornblende coexists in an assemblage with kyanite, staurolite, garnet, biotite, quartz, muscovite, plagio-

Weight %

	U101	U103	U111	U141
SiO_2	41.87	41.39	40.81	41.14
TiO_2	0.35	0.34	0.38	0.39
Al_2O_3	17.85	18.13	18.51	19.24
FeO^*	17.87	18.20	18.39	17.70
MgO	6.95	6.56	5.90	6.08
MnO	0.04	0.02	0.05	0.07
CaO	10.59	10.57	10.53	10.79
K_2O	0.36	0.50	0.41	0.48
Na_2O	1.51	1.52	1.43	1.43
Total	97.39	97.23	96.41	97.32

Recalculation on Basis of 23 Oxygens

	U101	U103	U111	U141
Si	6.24	6.20	6.17	6.14
Al	1.76	1.80	1.83	1.86
Al	1.38	1.41	1.47	1.53
Ti	0.04	0.04	0.04	0.04
Fe^{+2}	2.23	2.28	2.23	2.21
Mn	0.01	0.00	0.01	0.01
Mg	1.54	1.47	1.33	1.35
Ca	1.69	1.70	1.71	1.73
Na	0.44	0.44	0.42	0.41
K	0.07	0.10	0.08	0.09
Total Cations	15.40	15.43	15.38	15.37

Recalculation on Basis of 15 Cations without Na, K

	U101	U103	U111	U141
Si	6.29	6.25	6.22	6.20
Al	1.71	1.75	1.78	1.80
Al	1.45	1.48	1.55	1.62
Ti	0.04	0.04	0.04	0.04
Fe^{+2}	2.24	2.30	2.34	2.23
Mn	0.01	0.00	0.01	0.01
Mg	1.56	1.48	1.34	1.36
Ca	1.70	1.71	1.72	1.74
Na	0.44	0.44	0.42	0.42
K	0.07	0.10	0.08	0.09
Total Cations	15.51	15.54	15.50	15.51

Table 4-2. Electron microprobe analyses of hornblendes coexisting with kyanite from the Standing Pond Volcanics, sample location #500 (Hepburn, 1972A), 0.56 mile N.69°E. from B.M. 530′, Brattleboro. OH^- not analyzed. Total iron as Fe^{+2}. Analyses recalculated on the basis of 23 oxygens and 15 cations without Na or K.

clase and ilmenite. The presence of hornblende in a rock containing kyanite is rare and has been found only in a few other metamorphic rocks in the world. The hornblende, kyanite and garnet crystals are all several inches in size in this rock. There is no textural evidence to indicate they were not formed in equilibrium. Table 4-2 gives representative analyses of four hornblendes from this rock. Analyses were made on an ARL microprobe in the Department of Geological Sciences at Harvard University using Bence and Albee (1968) correction factors in a program developed by David Walker. These analyses represent a refinement from those presented in Hepburn, 1972A[1] for these minerals, since different data reduction methods were used. Note, in these analyses OH^- has not been analyzed for, thus the totals do not approach 100%. Also, total iron is reported as ferrous iron. Table 4-2 also shows two common methods for assigning atoms to various amphibole crystallographic sites, on the basis of 23 oxygens or on a total of 15 cations without Na or K. Both methods point out the relatively high aluminum contents in both the tetrahedrally and octahedrally coordinated sites. As such, these amphiboles are classified as alumino-ferro-tschermakites by the method of Leake (1968).

Volcanic and Mafic Igneous Rocks.

Table 4-3 lists the assemblages found in the metamorphosed volcanic and mafic igneous rocks. The large number of assemblages from the garnet zone is due to the extensive development of this metamorphic grade in the pre-Silurian rocks, all of which are in the garnet zone. Part of Standing Pond band #3 (Figure 2-5) is also in the garnet zone. The assemblages from the chlorite zone are from the Putney Volcanics. Mafic igneous rocks in the biotite zone are only found in the eastern band (band #3, Figure 2-5) of the Standing Pond Volcanics, where it occurs east of the garnet isograd. Assemblages from the staurolite-kyanite zone are from the Standing Pond bands #1 and #2.

Calcareous Rocks.

The calcareous assemblages found in the metamorphosed impure marbles of the Waits River, Standing Pond, Gile Mountain, and Northfield Formations are summarized in Table 4-4. The calcite in the Waits River Formation coexists with ankerite in the lower

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[1]Forty-four electron microprobe analyses of hornblendes, cummingtonites and garnets from the Brattleboro area are presented in Hepburn, 1972A to which the interested reader is referred.

Chlorite zone

Quartz-albite-calcite
Quartz-albite-calcite-epidote
Quartz-albite-chlorite-sericite

Biotite zone

Quartz-plagioclase-biotite-chlorite-hornblende
Quartz-plagioclase-biotite-chlorite-hornblende-calcite
Quartz-plagioclase-biotite-chlorite-hornblende-epidote

Garnet zone

Quartz-plagioclase-hornblende-chlorite-epidote-calcite
Quartz-plagioclase-hornblende-chlorite-epidote-garnet
Quartz-plagioclase-hornblende-chlorite-epidote-biotite
Quartz-plagioclase-hornblende-chlorite-epidote-biotite-K-feldspar
Quartz-plagioclase-hornblende-chlorite-epidote-biotite-calcite
Quartz-plagioclase-hornblende-chlorite-epidote-K-feldspar
Quartz-plagioclase-hornblende-chlorite-epidote-
Quartz-plagioclase-hornblende-chlorite-cummingtonite
Quartz-garnet-chlorite-hornblende-cummingtonite-calcite
Quartz-hornblende-biotite-chlorite
Quartz-hornblende-biotite-chlorite-calcite
Quartz-plagioclase-biotite-epidote-calcite
Quartz-plagioclase-hornblende-chlorite-biotite
Quartz-plagioclase-hornblende-garnet-biotite-epidote
Quartz-plagioclase-hornblende-garnet-biotite-epidote-chlorite
Quartz-plagioclase-hornblende-biotite-chlorite
Quartz-plagioclase-hornblende-epidote

Staurolite-Kyanite zone

Quartz-plagioclase-biotite-hornblende-epidote-calcite
Quartz-plagioclase-hornblende-epidote-calcite
Quartz-plagioclase-chlorite-hornblende-epidote-calcite
Quartz-plagioclase-biotite-garnet-epidote-calcite
Quartz-plagioclase-biotite-hornblende-epidote
Quartz-plagioclase-biotite-chlorite-garnet-hornblende

Table 4-3. Observed mineral assemblages in the volcanic and mafic rocks of the Western Brattleboro area.

metamorphic grades (Rosenfeld, 1954). Commonly, the tremolite forms radiating clusters at the boundary between the schistose and calcareous beds. Figure 4-4 is a photomicrograph of the tremolite in one of these clusters. In the staurolite-kyanite zone of the Waits River Formation large zoisite crystals, up to several inches in length, are also found in the "skarn" between the schistose and calcareous beds or adjacent to quartz lenses.

A. CALCAREOUS ROCKS

Biotite zone

Quartz-plagioclase-muscovite-biotite-chlorite-calcite
Quartz-muscovite-chlorite-calcite

Garnet zone

Quartz-muscovite-chlorite-calcite
Quartz-muscovite-biotite-garnet-calcite

Staurolite-Kyanite zone

Quartz-plagioclase-zoisite-biotite
Quartz-plagioclase-biotite-chlorite-tremolite-calcite-clinozoisite
Quartz-biotite-tremolite-calcite
Quartz-plagioclase-muscovite-garnet-biotite-chlorite-calcite-zoisite
Quartz-plagioclase-muscovite-garnet-biotite-actinolite-chlorite-clinozoisite

B. CONTACT METAMORPHOSED ROCKS

Quartz-plagioclase-garnet-hornblende-chlorite-diopside-calcite-clinozoisite
Quartz-plagioclase-biotite-chlorite-diopside-calcite-clinozoisite
Quartz-plagioclase-biotite-diopside-calcite-epidote

Table 4-4. Observed mineral assemblages in the (A) calcareous rocks, and (B) the contact metamorphosed calcareous rocks of the Western Brattleboro area.

In the western Brattleboro quadrangle, tremolite does not become prominent until the staurolite-kyanite zone, although it may first appear just below this isograd. In the Woodsville quadrangle, Vermont, White (1946) also indicates that actinolite does not first appear until well into the staurolite zone.

Trommsdorff (1966) described an area in the Lepontine and Bergell Swiss Alps where the metamorphism of the calcareous rocks appear to be similar to that in the Brattleboro area. He found that the assemblages tremolite-calcite did not form until the staurolite-kyanite zone; diopside did not occur until well within the amphibolite facies. Tommsdorff attributed the "late" appearance of tremolite and diopside to high pressure associated with Barrovian type metamorphism. Diopside is only found in the Brattleboro area in the contact aureole around the Black Mountain Granite. The diopside is not found in the regionally metamorphosed, impure marbles of the staurolite-kyanite zone. This suggests that the activity of H_2O or of CO_2, or both, were high in these rocks and that the crystallization of staurolite and kyanite took place at lower temperatures than the crystallization of diopside.

Figure 4-4. Photomicrograph of tremolite (Tr) growing between quartz (Q) and carbonate (C). Crossed polars, magnification 40X. Sample is from the Waits River Fm., 0.25 mile S.22°W. from B.M. 566′ Guilford Center, Guilford.

Contact Metamorphism.

Three assemblages from impure marbles in the contact aureole adjacent to the Black Mountain Granite are presented in Table 4-4B. Diopside is abundant in these rocks, indicating higher temperatures and most likely lower H_2O and CO_2 activities than elsewhere in the staurolite-kyanite zone.

Muscovite pseudomorphs apparently after an aluminum silicate are present near the granite contact. However, it has not been possible to tell whether these are pseudomorphic after original sillimanite or kyanite.

Retrograde Metamorphism.

The effects of retrograde metamorphism are generally minor throughout the area. They are most pronounced in the pre-Silurian rocks, where garnet and biotite commonly have been partially replaced by chlorite. Chlorite and biotite in these pre-Silurian rocks are not uncommonly found with a long, narrow, blade-like habit. These may represent the retrograde reaction products of an amphibole.

In the rest of the area, retrograde effects are only locally present. Garnet, staurolite, hornblende, and biotite have been partially altered to chlorite. Kyanite commonly has a thin rim of muscovite. As noted above, aluminum silicates in the area of contact metamorphism have been completely altered to muscovite.

Large Crystals in the Standing Pond Volcanics.

Particularly large crystals of garnet and hornblende occur along the Standing Pond-Waits River contact nearest to the center of the Guilford dome (Standing Pond band #1, Figure 2-5). The large crystals are extensively developed only along this particular contact in the Brattleboro area. Occurrences of large porphyroblasts near the contacts of the Standing Pond have been noted by Thompson (in Billings et al., 1952), Rosenfeld (1954), Lyons (1955), Doll (1944), and Howard (1969) in areas to the north. In these areas, the large crystals are not restricted solely to the contact of the Waits River and Standing Pond, although they appear to be more abundant along this horizon.

It is believed that the large crystals in the Brattleboro area may be related to the particular structural and stratigraphic position of the horizon along which they occur. The rocks of the Standing Pond near this contact must have had an appropriate composition. However, large garnet and zoisite crystals are also found in the Waits River Formation within a few feet of the contact, indicating that the composition of the rocks may not be the only important factor. It is thought that the large size of the crystals near this contact in the Brattleboro area may be due to a concentration of volatiles along this particular horizon. The Waits River Formation would have been particularly rich in volatiles. As the volatiles in the Waits River Formation in the center of the Guilford dome were driven away from the thermal high associated with the dome, the Standing Pond Volcanics, draped over the dome, may have acted as a less permeable blanket, trapping the volatiles along its lower contact. The added volatiles would have increased ion migration and led to the growth of the large crystals.

Conditions of Metamorphism.

The lower greenschist facies metamorphism (chlorite-biotite zones) in the Brattleboro area likely took place at temperatures in the 300°C to 400°C range (cf. Turner, 1980). To investigate the maximum metamorphic conditions in the area (staurolite-kyanite zone of the amphibolite facies), two samples of mica schist were analyzed with the electron microprobe in order to determine paleo-temperatures. Both schists include the assemblage garnet-biotite-staurolite-kyanite. Sample #100G is from the Gile Mountain Formation in the southern Guilford dome (0.93 mile S.81°W. from 831′ road junction, Hinesburg, Guilford) relatively close to the mapped staurolite-kyanite isograd (Plate I). Sample #500 is from the Standing Pond Volcanics (0.56 mile N.69°E. from B.M. 530′, Brattleboro) closer to the center of the northern Guilford dome and is farther inside the staurolite-kyanite isograd. The temperature estimates were made using the method of Ferry and Spear (1978), which utilizes the distribution of Fe and Mg between coexisting garnet and biotite. Although Ferry and Spear caution that additional components such as Ca and Ti could cause deviations in the temperature of natural mineral assemblages from the ones used in their idealized system, this method should be accurate to ±50°C for the samples analyzed. Sample #100G gave a tight spread of temperatures centered around 490°C for several biotite-garnet pairs, although it should be noted that the garnets did contain up to 6.5 wt.% CaO and the biotites up to 1.2 wt.% TiO_2. An average temperature of 561°C was obtained from sample #500. Individual analyses in this sample gave a range of temperatures from 540°C to 580°C. Since sample #500 is from an area where greater uplift of the Guilford dome has occurred when compared to sample #100G, it is felt the temperature differences recorded by these two samples are real and give an approximate estimate of the range of temperatures for the staurolite-kyanite zone in the Guilford dome area.

Kyanite is the only aluminum silicate phase developed during regional metamorphism in the Western Brattleboro quadrangle and serves as a guide to the pressures obtained during the metamorphism. Using the temperatures obtained above and Holdaway's (1971) aluminum silicate phase diagram and triple point (501°C, 3.76 kbars), the pressure in the staurolite-kyanite zone in the Guilford dome area is estimated to have been at least 5 to 5.5 kilobars (equal to a burial depth of some 18 to 20 kilometers). Combining these pressure-temperature estimates, a geothermal gradient for the area is estimated to be approximately 27°C/km to 30°C/km near

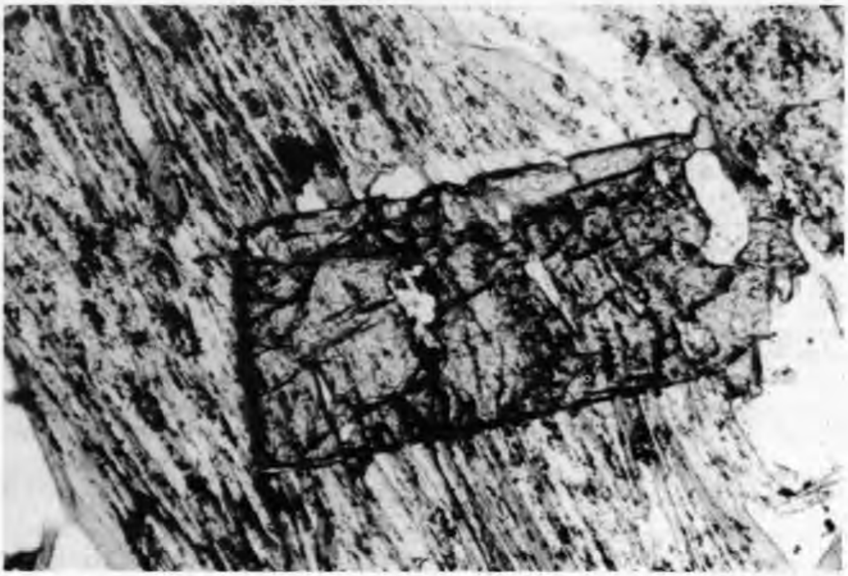

Figure 4-5. Photomicrograph of a staurolite crystal growing across the previously formed schistosity. Sample is from a roadcut in the Northfield Formation on the Green River Rd., 1.5 miles S.72°E. from road junction, Reid Hollow, Halifax. Magnification 40X.

the thermal maximum. This gradient is well within the range established for Barrovian metamorphic terrains elsewhere (Turner, 1980).

Relation of Metamorphism to Deformation.

The mapped isograds (Plate I) indicate that the highest grades of regional metamorphism in the area are centered on the Guilford and Athens domes. In the Guilford dome area, kyanite and staurolite porphyroblasts have grown across the S_1 and S_2 foliations, parallel to which minute inclusion trains of graphite and quartz had been incorporated (Figure 4-5). In no case were these porphyroblasts observed to have a rotated or discordant internal schistosity. Thus in the Guilford dome area, textural evidence indicates the peak metamorphic conditions of the staurolite-kyanite zone were reached subsequent to the F_2 recumbent folding stage (Table 4-5).

The extent of the metamorphism prior to and during the development of the Prospect Hill fold is unclear. The highest grade assemblages (i.e., staurolite-kyanite zone) had not formed. Rosenfeld (1968) indicates that syntectonic garnet growth took place during the folding of the Ascutney sigmoid in the Saxtons River quadrangle to the north. However, garnets with highly rotated spiral

Minor folds	Major folds	Metamorphism
F_5, large, steeply plunging sinistral folds in Brattleboro syncline; kink bands.	----------------------------	----------------------------
F_4, open folds, buckles, or warps in schistosity; may have weak axial planar slip cleavage.	----------------------------	----------------------------
F_3, open folds prominent west and southwest of Guilford dome; moderate plunges; prominent slip cleavage parallel to axial surfaces.	Guilford dome and Brattleboro syncline.	Staurolite-kyanite zone.
F_2, tight to isoclineal folds in schistosity and banding; slip cleavage to schistosity developed parallel to axial surfaces; plunge northeast and southwest.	Major recumbent folding (Prospect Hill recumbent fold).	Garnet zone.
F_1, small isoclinal folds in bedding or banding; a principal schistosity parallel to their axial surfaces.	----------------------------	----------------------------

Table 4-5. Summary of Acadian minor folds, major structures and metamorphism in the Guilford dome area, after Hepburn 1975.

inclusion trains seem to be absent in the central Brattleboro area, suggesting that the metamorphic grade may not have reached the garnet zone when the Prospect Hill recumbent fold was forming.

The exact temporal relation between the peak of metamorphism and the doming stage of deformation is more difficult to establish. The rise of material in the dome undoubtedly carried heat upward. Whether the metamorphic conditions of the staurolite-kyanite zone were first introduced into the rocks during or after the doming, or whether the doming merely carried rocks already at these conditions to their present levels, is unresolved. However, the undeformed nature of the staurolite and kyanite porphyroblasts suggests that the peak of the metamorphism occured either late in or following the doming stage of the deformation. Also, staurolite porphyroblasts have grown across the S_3 slip cleavage.

Garnets such as shown in Figure 4-6 give additional evidence that the metamorphism was late. These garnets from the Guilford dome have slightly rotated cores and clear apparently unrotated exteriors. The rotation axis of the core of the garnet in Figure 4-6 lies in the plane of schistosity; the direction and amount of rotation is consistent with its having been formed during the doming phase of deformation (Rosenfeld, 1968). A detailed zoning profile across this garnet (Figure 4-7) indicates that there is no break or reversal in the zoning. As this profile is from a polished thin section, the central portion may not pass through the exact center of the garnet; and thus the zoning curve is somewhat flattened. The continuous zoning from rolled through unrolled parts in this garnet was observed in several other garnets from the Guilford dome area (Hepburn, 1972A) and is interpreted to mean that the metamorphism outlasted the major deformation in this area.

The exact relations between the peak of metamorphism and the F_4 and F_5 stages of minor folding have not been determined, although it appears likely that the metamorphic peak preceded at least the F_5 folding. Some of the kink bands along the eastern edge of the area may possibly be related to later faulting, as has been suggested by Rumble (1969).

The emplacement of the Black Mountain Granite at the current level of erosion followed most of the deformation in the area, as has been noted. It is believed that the contact aureole developed nearly synchronously with the peak regional metamorphism as a local "hot spot".

Reconstructed schematic cross sections showing the evolution of the structural and metamorphic features of the Guilford dome area

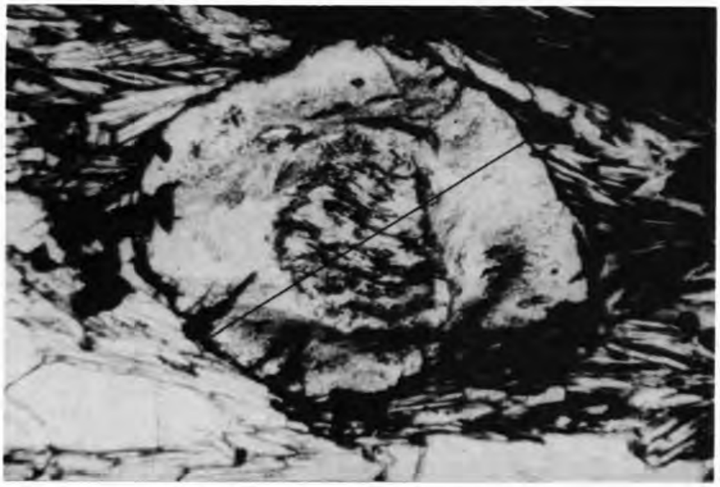

Figure 4-6. Photomicrograph of garnet #257B from the Gile Mountain Formation. The interior of the garnet shows syntectonic growth. The clear unrolled exterior grew after deformation ceased. Line indicates the electron microprobe traverse shown in Fig. 4-7. The garnet is approximately 1 mm in diameter. Location, 0.75 mile N.30°E. from B.M. 719', Green River Village, Guilford. After Hepburn, 1975.

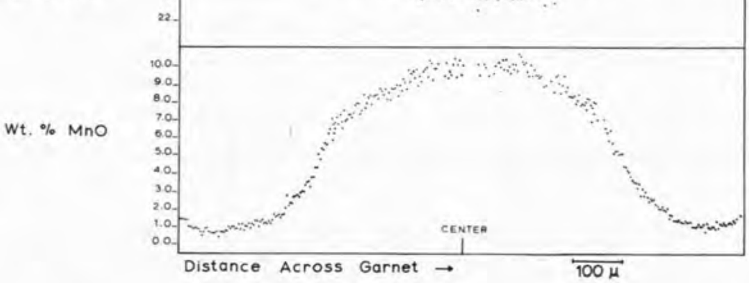

Figure 4-7. Plot of the variation is wt.% FeO and MnO determined by electron microprobe. Traverse is across the zoned garnet in Figure 4-6. The garnet has a rolled central portion surrounded by a clear, unrolled exterior. The analysis is from a thin section which probably did not pass through the exact center of the garnet, thus causing the curves to appear somewhat flattened. After Hepburn, 1975.

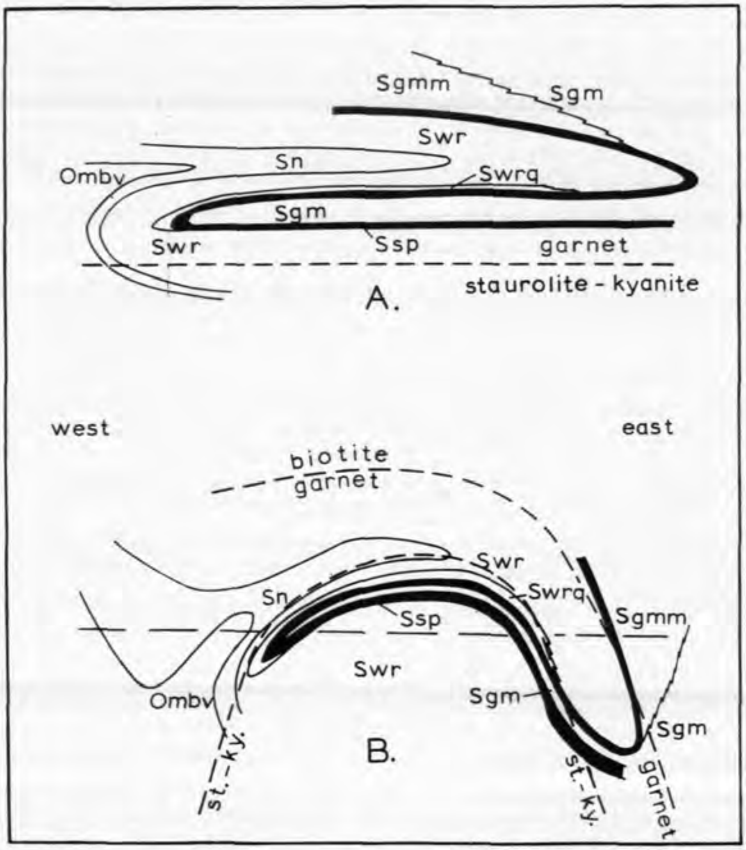

Figure 4-8. Schematic cross sections showing the evolution of the structural features in the two major stages of deformation in the Guilford dome area. The Standing Pond Volcanics is shown in black. (A) Prospect Hill fold at the end of the first major stage of deformation, before the rise of the Guilford dome. The dashed line represents a hypothetical staurolite or kyanite isograd. (B) Prospect Hill fold following the second major stage of deformation, after the rise of the dome. Horizontal line represents the present erosion surface. The dashed lines show the assumed present distribution of isogrades. Modified after Hepburn, 1975. Formations: Sgm = Gile Mt.; Sgmm = Gile Mt., marble mbr.; Ssp = Standing Pond Volc.; Swr = Waits River Fm.; Swrq = Waits River Fm., quartzitic mbr.; Sn= Northfield Fm.; Ombv, Missisquoi Fm., Barnard Volc. Mbr.

are shown in Figure 4-8. The Prospect Hill recumbent fold formed during the first major deformational event (Figure 4-8A), contemporaneously with the F_2 stage of minor folding. The recumbent folding affects a previously developed schistosity that is parallel to the axial surfaces of the F_1 minor folds. The garnet grade of metamorphism was not exceeded during the recumbent folding in the Guilford dome area.

Subsequent to the recumbent folding (Figure 4-8B), the rising of the Guilford dome arched the axial surface of the recumbent fold and probably produced the F_3 minor folds and related prominent slip cleavage. Staurolite-kyanite zone metamorphism is associated with the formation of the dome.

Possible Role of the Black Mountain Granite.

Why did the isograd surfaces in the area of the Guilford dome rise later with respect to the tectonism than those around the Athens and Chester domes to the north? This fact is particularly apparent when comparing the small amounts of rotation of very large garnets in the Standing Pond Volcanics along the west side of the Prospect Hill fold, south of Prospect Hill, to the large degree of rotation for very large garnets at the same structural position on the east side of the Athens dome, west of Saxtons River Village (Saxtons River quadrangle). This is true despite the fact that the garnets in the former location are at a considerably higher grade of metamorphism, and the tectonic history of the rocks at both sites is very similar. At the latter location, the garnets show typical syntectonic snowball fabric indicating rotations of more than 3 radians during recumbent folding, Event I of Rosenfeld (1968), with little or no rotation during Event II of Rosenfeld (1968), doming (see Rosenfeld, 1970, Plate 12). Not only are the rocks at a given structural level in the Guilford dome of higher metamorphic grade than rocks at the same structural level just to the north, but the same levels in the Guilford dome have not been upthrust to nearly the same degree during the doming. Unlike the Chester and Athens domes, which bare their Precambrian cores, the Guilford dome contains only the Waits River Formation and the Black Mountain Granite in its presently exposed core.

It is believed that the presence of the Black Mountain Granite may have been the cause of the contrast in the thermo-tectonic history between the Guilford dome and the vicinity of the Chester and Athens domes to the north. Peraluminous granites such as the Black Mountain are believed to result from partial, quasi-eutectoid melting of pelitic rocks during ultrametamorphism. Trace element studies (Hepburn, 1982) clearly support this type of origin for the Black Mountain Granite. The maximum temperatures of metamorphism in the staurolite-kyanite zone of eastern Vermont are within 100°C to 200°C of the partial melting temperatures of H_2O-saturated pelitic rocks. Using Shields' (1977) estimate that the Black Mountain Granite extends for 5 km below the present surface, and a geothermal gradient of 30°C/km, the base of the Black Mountain Granite is clearly placed in a zone where such melting could have occurred. Thus, if locally at depth under the future site of the Guilford dome, there were more H_2O-rich pelitic rock than in nearby areas, its partial melting would have acted as a temperature buffer. This would delay the rise of the isothermal surfaces while the pelitic rock was melting. Such a delay would persist until sufficient granitic magma had amassed to permit it to intrude overlying strata. The convected heat would then dissipate due to thermal relaxation as the magma solidified, causing the maximum rise in temperature of the intruded rocks to follow the intrusion. The intrusion of the granite also released most of the components that lowered the density in the rocks from which the granite was derived. The increased density of the so-called restite could account for the relatively small upthrust of the Guilford dome, just as a balloon ceases its rise after puncture releases its buoyant gas.

Age of Metamorphism and Deformation.

The principal metamorphism and deformation in the post-Precambrian rocks of the Brattleboro area took place during the Acadian orogeny. Evidence summarized elsewhere (Cady, 1969; Billings, 1956) indicates that the metamorphosed and deformed Littleton Formation contains Early Devonian fossils. The Triassic deposits in Massachusetts are unmetamorphosed and truncate the older metamorphosed rocks. In the Brattleboro area, the Rb-Sr date on coarse muscovites from the Black Mountain Granite (377 m.y. and 383±7 m.y., Naylor, 1971) places an upper limit of early Middle Devonian on the major phases of the orogenesis.

The Precambrian core rocks of the Athens and Chester domes underwent a previous metamorphism in Precambrian time (Faul et al., 1963). No conclusive evidence of a Taconic fabric or metamorphism was found in the pre-Silurian rocks of the Brattleboro area, although this is probably due to the severity of the Acadian overprint.

CHAPTER 5

STRATIGRAPHY OF THE EASTERN SEQUENCE

East Flank of the Connecticut River-Gaspé Synclinorium and West Flank of the Bronson Hill Anticlinorium

General Statement.

The rocks east of the depositional outcrop trace of the Putney Volcanics (Plate I) probably were syndepositional with the rocks to the west, but are sufficiently distinct in detailed succession and lithofacies to warrant separate treatment. The stratigraphic sequence for the western flank of the Bronson Hill anticlinorium, as now understood, is in its major features essentially the same as that first worked out by M. P. Billings (1937) and coworkers in the area around Littleton, New Hampshire.

The part of the Eastern Sequence exposed in the Brattleboro quadrangle consists, in ascending order, of the Ordovician Ammonoosuc Volcanics and Partridge Formation, the Silurian Clough and Fitch Formations and the Devonian Littleton Formation. A stratigraphic column for the Eastern Sequence and its correlations to the Western Sequence are shown in Figure 2-1. Type localities of the Ammonoosuc, Partridge, Clough, Fitch and Littleton Formations are in the Littleton-Moosilauke area of northern New Hampshire (Billings, 1937). Systematic mapping has shown that these formations extend southward to the Massachusetts line and beyond (Moore, 1949; Billings, 1956; Thompson et al., 1968).

Previous workers in the adjacent areas of Massachusetts (Emerson, 1898; Balk, 1956A, B) have used stratigraphic names different from those used in the present study. The Bernardston Formation of Massachusetts has in the past included strata mapped here as the Partridge, Clough and Fitch Formations and part of the Littleton Formation. Also, the term Leyden Formation has been used in Massachusetts to designate rocks that are herein included in part of the Littleton Formation, the Putney Volcanics and the Gile Mountain Formation.

The grade of regional metamorphism increases progressively from west to east across the Eastern Sequence from the chlorite zone to the staurolite zone (Plate I). Rocks of the same bulk composition may thus have different mineral compositions depending on the grade of metamorphism. The Littleton Formation crops out over the full range of metamorphic zones; it grades from slate and phyllite in the west to staurolite schist in the east and reaches the sillimanite zone just east of the quadrangle. The Partridge and Clough Formations crop out in the garnet and staurolite zones. The Ammonoosuc Volcanics are confined to the staurolite zone as, with only minor exceptions, is the Fitch Formation.

ORDOVICIAN

Ammonoosuc Volcanics

General Statement and Areal Distribution. The Ammonoosuc Volcanics crop out in two distinct bands. A southern band in Vernon and Hinsdale townships runs approximately north-south, and a northern band exposed in the area of Streeter Hill crosses the Westmoreland-Chesterfield town line. The Ammonoosuc Volcanics are more resistant than the structurally underlying plutonic rocks of the Oliverian Series. They form prominent ridges along most of the length of the two outcrop bands.

Quartz-feldspar-biotite-gneiss constitutes approximately two-thirds of the Ammonoosuc Volcanics, and amphibolite approximately one-third. The two rock types are interlayered on all scales and in all proportions and probably represent original volcanic stratification. Most commonly, beds of amphibolite averaging 30 to 50 feet thick are interbedded with much thicker layers of quartz-feldspar-biotite gneiss. Amphibolite is most abundant in the lower part of the formation. A pod of amphibolite 1200 feet thick at the base of the formation at the northeast corner of the southern outcrop band is the thickest amphibolite in the area.

Fine-grained and medium-grained quartz-feldspar-biotite gneiss and amphibolite at the base of the formation alternate with coarse-grained quartz diorite gneiss of the underlying plutonics at several places along the southern outcrop band. The layers of coarse-grained gneiss may represent sills of the Oliverian Plutonic Series that were intruded into the lower part of the Ammonoosuc Volcanics. The exact contact between the Oliverian Plutonic Series and the Ammonoosuc Volcanics was not observed within the area.

Rock Description. The quartz-feldspar-biotite gneiss of the Ammonoosuc Volcanics is mostly well foliated. Weathered outcrops are light gray to white; in some localities, pink, light-brown, yellowish-brown and rust-colored outcrop surfaces were noted. Fresh gneiss is uniformly light gray. Medium-grained to fine-grained quartz and plagioclase feldspar are the principal constituents of the gneiss, with quartz in most specimens particularly abundant. No potash feldspar was noted in thin sections, although a systematic study of the petrology of the gneisses was not made. Moore (1949, p. 1622) also reported that the light-colored gneiss in the Ammonoosuc Volcanics of this area contained mostly plagioclase. Biotite is present in the gneiss in amounts ranging from less than 1 percent to 10 percent. Most of the gneiss contains garnet averaging 3 mm in diameter but ranging up to as much as 10 mm in diameter along some layers. Ellipsoidal, milky-white quartz megacrysts are conspicuous in many layers; the average intermediate dimension of these megacrysts is 5 mm. Rare ellipsoidal aggregates of quartz and feldspar are also present. Breccias with fragments of similar composition as their hosts, averaging 5 to 8 cm in diameter, are found at widely separated localities in the quartz-feldspar-biotite gneiss.

Amphibolites in the Ammonoosuc Volcanics are medium- to coarse-grained and include both schistose and non-foliated types. Weathered outcrops are medium gray to dark gray; fresh surfaces are typically greenish gray. Hornblende needles in the schistose amphibolites are several millimeters long. The more massive amphibolites contain layers with hornblende laths as long as 3 cm. Epidote, pyrite, garnet and quartz are present in the amphibolites in minor amounts.

A muscovite-rich rock, with coarse-grained muscovite constituting as much as 90 percent of the rock, is found in discontinuous lenses near the top of the Ammonoosuc Volcanics throughout the area. The rock has a well-developed schistosity and commonly contains quartz and plagioclase in addition to the muscovite. Isolated surfaces of foliation are covered with a rust-colored rind.

Origin. Lahee (1916) first established the volcanic origin for most of the Ammonoosuc Volcanics. Billings (1956, p. 18) has pointed to fragmental textures in the formation at its type locality, where the grade of metamorphism is low, as evidence that most of the volcanics are pyroclastics rather than flows. Fragmental textures are rare in the Brattleboro area but may have been largely destroyed by metamorphism. Small areas of breccia, megacrysts of quartz, and rounded fragments of quartz-feldspar gneiss within the quartz-feldspar-biotite gneisses probably indicate a pyroclastic origin. No fragmental textures were noted in the amphibolites. Some of the thicker amphibolites may be metamorphosed flows. The high content of quartz in some of the gneisses may be due to leaching of original volcanic material. Leo (1981) indicates that the amphibolites and felsic gneisses of the Ammonoosuc Volcanics are part of an original bimodal assemblage of tholeiitic basalts and volcanoclastic quartz keratophyres, similar to those found in modern island arcs or associated with continental margin magmatism. The highly aluminous, muscovite-rich schists are believed to be the product of intense leaching of original volcanic material prior to metamorphism.

Thickness. The thickness of the Ammonoosuc Volcanics in the southern outcrop area is 1600 feet. The formation totals 700 feet in thickness in the Streeter Hill area. The lower part of the formation has been intruded by the Oliverian Plutonic Series so that the original base of the formation is not exposed. Thus these figures represent only partial thicknesses.

Partridge Formation

General Statement and Areal Distribution. The Partridge Formation is a varied sequence of schists, amphibolites, quartz-feldspar-biotite gneisses and granulites. The Partridge crops out in three tracts all in the east-central and northeast portions of the Brattleboro quadrangle. These occur in two distinct structural environments. The northern and smaller of the three is on the southwest slope of Streeter Hill in Chesterfield. The other two tracts are in the Sargent Hill and the Hubbard Brook-Chesterfield area.

Quartz-mica schist makes up approximately 70 percent of the Partridge Formation; amphibolite, 20 percent; quartz-feldspar-biotite gneiss and granulite, 10 percent; conglomerate and calc-silicate granulite, less than 1 percent. The two larger tracts contain the greatest variety of rock types. The area on Streeter Hill consists mostly of schist with small amounts of amphibolite and quartz-feldspar-biotite granulite.

Two subunits of the Partridge Formation have been designated on Plate I. The subunits do not have the status of members and the boundaries are not clearly defined. The schist subunit, **Ops**, consists mainly of mica schist, with subordinate, very thick to very thinly interbedded amphibolite and quartz-feldspar-biotite gneiss. Locally in this subunit, quartzose layers and quartzo-feldspathic layers averaging one inch in thickness are interbedded. The amphibolite and schist subunit, **Opas**, consists of interbeds of amphibolite and schist, with slightly more amphibolite than schist.

The thicker amphibolites and quartz-feldspar-biotite gneisses are the most resistant rocks in the Partridge Formation and form pronounced ledges and hogbacks. The schists crop out on rounded, irregular ledges on steep slopes and in association with more resistant layers. Many of the outcrops of the formation have a distinctive rust-colored rind. Contacts between the Partridge Formation and adjacent formations are sharp in most cases.

Rock Description. The mica schists of the Partridge Formation are gray on fresh surfaces. About 60 percent of the schists are sulfidic, and weather with oxidized surface films ranging from deep reddish brown to ocherous yellow. The oxidized material is mostly goethite. Rocks with a high content of goethite commonly break apart very easily and fresh samples cannot be obtained from the surface. A water well that started in highly rusty-weathering schist at the surface encountered gray schist with abundant fresh sulfide several feet below the surface. The fresh sulfide is a weakly magnetic pyrrhotite. The sulfidic schists are fine-grained and have a dense, poorly-foliated appearance in the garnet zone of metamorphism. In general, the grain size increases with increasing grade of metamorphism. Garnet and staurolite porphyroblasts are relatively rare in the sulfidic schists.

Medium-grained to coarse-grained amphibolites are interbeded throughout the Partridge Formation. They are most abundant in the amphibolite and schist subunit (**Opas**). The amphibolites are dark greenish gray on fresh surfaces and weather to various shades of gray. In the central portions of many of the thicker, coarse-grained amphibolites, the hornblende and plagioclase form an interlocking mosaic of lath-shaped grains with no discernible foliation. A foliation is present in thinner, medium-grained amphibolites and on the margins of the thicker bodies. With rare exceptions, the amphibolites of the Partridge Formation do not exhibit alignment of hornblende crystals to produce a lineation. The proportion of hornblende and plagioclase in the amphibolites varies considerably. Garnet is locally present in the amphibolite in amounts up to 10 percent.

Fine- to medium-grained quartz-feldspar-biotite gneiss and granulite are present as scattered interbeds within the schist (**Ops**) and the amphibolite and schist subunits (**Opas**). Most of the gneiss and granulite is light gray in weathered outcrop, with a few beds of yellow, pink, or brown. Fresh surfaces of the gneiss and granulite are light to dark gray. Many of the fine-grained and medium-grained gneisses closely resemble the light-colored gneisses of the Ammonoosuc Volcanics. They are well foliated and contain thin streaks and beds of amphibolite, megacrysts of quartz up to 2 mm in diameter, isolated areas of breccia, and garnet in restricted layers. Plagioclase is the predominant feldspar.

Other quartzo-feldspathic rocks in the Partridge Formation are fine-grained granulites. Some of these are gradational into granulitic schists but others, up to 20 feet thick, make sharp contacts with the adjacent schists. These granulites may contain megacrysts of quartz, plagioclase, garnet and biotite. A few thin beds of fine-grained, dense, gray calc-silicate granulite are scattered throughout the formation. These calc-silicates may be highly sulfidic.

Origin. The quartz-mica schists of the Partridge Formation were derived from sulfidic and non-sulfidic shales laid down under reducing, most likely marine, conditions. The amphibolites and quartz-feldspar-biotite gneisses indicate that volcanic activity continued, though more sporadically than during deposition of the Ammonoosuc Volcanics. The light-colored gneisses with scattered fragmental textures are metamorphosed pyroclastics. Some of the fine-grained granulites may be the metamorphosed equivalents of quartzo-feldspathic sandstones or of volcanic tuffs.

Thickness. The thickness of the Partridge Formation cannot be measured in the two larger tracts because the base is not exposed. Any estimate of a minimum thickness is difficult because of structural complications. Only the basal few hundred feet are exposed on Streeter Hill and the formation is cut out elsewhere because of an unconformity at the base of the Clough Formation (Billings, 1956).

SILURIAN

Clough Formation

General Statement and Areal Distribution. The Clough Formation is a relatively thin, distinctive unit that is present over much of western New Hampshire. In the Brattleboro area, two-thirds of this unit is composed of quartz-pebble conglomerate and one-third is quartzite with minor amounts of gray mica schist. The Clough Formation, similarly to the Partridge, crops out in two distinct structural settings. The extensive exposures in Vernon and Hinsdale, and those in the immediate vicinity of Streeter Hill, Chesterfield are parts of the mantling sequences of the Vernon and Keene domes, respectively. The remaining exposures, many of them discontinuous lenses, are mainly in the overriding Bernardston nappe. Gray schists and minor amphibolite in the valley of Catsbane Brook near the eastern border of the quadrangle, shown on Plate I as **Scs** (schist), are interpreted as lying in the structurally still higher Skitchewaug nappe.

Where the quartzite contains very thick beds of conglomerate, it forms light-gray to white, resistant ledges and cliffs. Where it is non-conglomeratic or contains only thin beds of conglomerate, it forms light-gray to rust-colored ledges that are still generally more prominent than outcrops of the overlying and underlying strata. The dip slopes of the ledges are parallel to formation contacts and to bedding.

Interbeds of muscovite schist, muscovite-garnet schist and micaceous quartzite are generally rare except, as noted, in the valley of Catsbane Brook.

Rock Description. Beds of conglomerate in the Clough Formation are most evident on the surfaces of the dip slopes of ledges. The pebbles have been flattened and elongated in the plane of the foliation and now have the form of elongated ellipsoids. In some outcrops, outlines of pebbles overlap while in others the pebbles are separated from one another by homogeneous quartzite. The size of the pebbles in the Clough varies widely. Some of the largest pebbles are found at Mine Ledge just south of Wantastiquet Mountain. Typical pebble dimensions here are 20 x 8 x 4 centimeters. Larger-than-average pebbles are also present at the south end of the same outcrop band. More typical pebbles in the rest of the formation have average dimensions of approximately 10 x 4.5 x 1 centimeters. Pebbles may become even more highly elongated in some areas.

The pebbles are mostly coarse-grained vein quartz in the elliptical outcrop band in the central part of the area south of Wantastiquet Mountain, whereas elsewhere up to half of the pebbles are composed of fine-grained quartzite. A few pebbles of granite were noted in an outcrop 1500 feet northwest of the New Boston Cemetery, Chesterfield. The problem of clearly distinguishing pebbles in many outcrops of the formation makes it difficult to estimate accurately the relative amounts of pebbles versus matrix. In outcrops containing large overlapping pebbles, the matrix commonly makes up less than 10 percent of the rock. In other areas, the matrix of medium-grained quartzite appears to constitute as much as 50 percent of the rock. Locally, a matrix of mica schist occurs that can be seen to grade along strike into the quartzitic matrix.

The quartzite commonly consists of a mosaic of interlocking, flattened quartz grains, the average diameter of which is approximately 1 mm. Discontinuous, minute laminae of schist, in some cases sulfidic, are common. These lenticular interbedded schists are rich in muscovite and may contain large garnets.

Origin. The Clough Formation was derived from a relatively thin sequence of conglomerate and quartzose sandstone with minor amounts of shale and, outside the Brattleboro area, limestone and felsic volcanics. The size of the pebbles in the conglomerate indicates that the source rock was relatively near. Their composition suggests that the source underwent severe chemical weathering so that only the most resistant fragments, of vein quartz and quartzite, survived.

Thickness. The thickness of the Clough Formation ranges from 0 to 450 feet. On Plate I, the outcrop width of the unit has been exaggerated in many places for clarity. The formation is thickest in the band through Vernon and Hinsdale, but has been duplicated by folding at Mine Ledge in Chesterfield at the north end of this band. In most of the rest of the outcrop areas, the formation is less than 100 feet thick and highly discontinuous.

Fitch Formation

General Statement and Areal Distribution. The Fitch Formation, another relatively thin and distinctive unit, overlies the Clough in much of western New Hampshire. It is characterized everywhere by a high content of calcium-rich minerals. The formation crops out mainly in two bands in the vicinity of Sargent Hill, Chesterfield (Plate I).

In most outcrops, the Fitch Formation consists of interbedded fine- to medium-grained calc-silicate granulites, granulitic schists and mica schists. Many of the granulites and granulitic schists of the typical Fitch Formation weather to a distinctive purplish-brown color; they are gray with a purplish cast on fresh surfaces. The fine-grained, purplish-brown granulites are a tough, splintery rock that forms pronounced ledges; other rock types within the formation are less resistant. As a whole the unit is more resistant than the Littleton and Partridge Formations and less resistant than the Clough Formation. Bedding is commonly shown in most outcrops by differences in the relative amounts of the calc-silicate minerals and calcite. Laminae and very thin beds commonly contain concentrations of actinolite, diopside and calcite. Thicker beds, with a few exceptions, consist of quartz-biotite granulite, granulitic schist and mica schist. Thin light-gray beds of quartzite are locally present. Relatively rare rocks within the formation are sulfidic, laminated epidote-rich granulite and garnet-rich granulite, which typically form highly contorted, thin beds.

In the western band across Sargent Hill, a section of 300 feet of gray mica schist with interbedded light gray, medium-grained quartzite intervenes between the highest pebble conglomerate and quartzite of the Clough and the lowest calc-silicate granulite of the Fitch. The mica schist in this section is similar to the mica schist in the Littleton Formation. These rocks are included in the Fitch Formation here on the basis of their stratigraphic position, although they are not typical of the formation in other areas. In places, the upper part of the Fitch Formation contains thinly-interbedded mica schist, quartzite and calc-silicate granulite. The quartzites and schists exhibit inverted graded bedding at one locality 0.5 mile north of Catsbane Brook.

Rock Description. The purplish-brown quartz-biotite granulites and granulitic schists of the Fitch Formation have a fair to good foliation that is visible on weathered surfaces but less evident on many fresh surfaces. Quartz and biotite are present in all of the granulites and granulitic schists. Calcite, actinolite, clinozoisite, plagioclase, and garnet are the principal remaining constituents; they are present in a wide range of proportions. In thin section, widely separated laminae and very thin beds of coarse-grained actinolite, diopside and plagioclase are scattered throughout the granulites and granulitic schists. A few laminae and very thin beds contain garnets up to 5 mm in diameter. There are also rare thin beds with a high proportion of calcite. The beds rich in calc-silicate minerals weather light gray and are greenish gray on fresh surfaces.

Rare beds of sulfidic calc-silicate granulite in the Fitch Formation average 6 inches in thickness. The granulite is well laminated and consists of fine-grained quartz, muscovite, and zoisite, with an unusually high content of iron sulfide and a black opaque material mixed with the sulfide that is propably graphite. A garnet-rich granulite with anthophyllite crops out in Chesterfield, 0.75 mile east-northeast of the New Boston Cemetery. Needles of anthophyllite are up to 10 mm long. Many of the garnet granulites have a conspicuous black crust of manganese oxide on weathered surfaces, suggesting that the garnet has a high content of spessartine; fresh surfaces are pink to gray. Clifford (1960, p. 1390) reported manganese-rich garnets from granulites in the Fitch Formation from Walpole, New Hampshire, 10 miles north of the Brattleboro quadrangle.

Origin. The Fitch Formation in Chesterfield and Hinsdale townships was derived from a variegated sequence of shales, calcareous shales, dolomitic shales and sandstones.

Thickness. The formation ranges from 0 to 500 feet in thickness in Chesterfield and Hinsdale townships. In a few places, the thickness shown on Plate I has been exaggerated for clarity.

DEVONIAN

Littleton Formation

General Statement and Areal Distribution. The Littleton Formation, a relatively thick, dominantly pelitic and quartzose unit, is the most widespread formation in western New Hampshire and underlies a large part of the eastern third of the Brattleboro quadrangle. Throughout most of its outcrop, the formation consists largely of interbedded quartzite and metamorphosed argillaceous rocks. A few scattered lenses of conglomerate occur at its base. The pelitic rocks in the Littleton grade from slate to phyllite to schist through the various metamorphic zones. The quartzites range in thickness from a few millimeters to several feet. Outcrops of the formation are dominantly medium gray to dark gray with a few rust-colored patches. There are continuous exposures on parts of Wantastiquet Mountain northeast and east of the city of Brattleboro. Elsewhere the formation crops out as a series of isolated ledges. Exposures are generally better in the areas of higher-grade metamorphism. They are also numerous in areas of low-grade metamorphism along the western portion of the main outcrop band; but in many cases, the limited extent of individual outcrops here hinders a complete analysis of minor structural features.

Quartz veins one inch to one foot thick are scattered throughout the Littleton Formation. Where bedding is distinctly shown by compositional layering, the quartz veins characteristically are parallel to bedding. Many of the quartz veins are intricately folded and thicken and thin markedly over short distances. The host rock has a rust-colored rind a few inches thick on either side of many of the veins. The quartz veins and quartzite beds weather out in relief, giving the rock a furrowed appearance.

A few widely scattered lenses of light gray plagioclase granulite with an average thickness of 6 inches are present in the western portion of the main outcrop band of the formation.

Conglomerate. A thin discontinuous conglomerate (**Dlc**, Plate I) occurs at the base of the Littleton, just east of the interlayered feldspathic granulites and gray phyllites of the Putney Volcanics. The conglomerate has pebbles of vein quartz and slate, "floating" in a phyllite matrix. The pebbles have been somewhat flattened and, in at least one outcrop, elongated. The average long direction of the flattened pebbles is 6 mm. Some slate pebbles have a long dimension up to 20 cm. The present outlines of the pebbles are sub-rounded. The clasts do not show grading. In some outcrops, the pebbles account for 40 to 50 percent of the rock (Figure 5-1), while in others the pebbles amount to only 15 percent. In thin section, small quartz and feldspar clasts appear to be "suspended" in the sericite of the phyllite matrix. At the northern conglomeratic lens in Dummerston, small spessartine-rich garnet porphyroblasts are locally present in the phyllitic matrix of the conglomerate.

The type locality and best exposure of the conglomerate and its contact with the Putney Volcanics is in an abandoned chicken yard just west of Vermont Route 5, 0.45 mile N.34°E. from road junction 428 feet, south of Dutton Pines State Forest, Dummerston.

Rock Description in the Chlorite-Garnet Zones. With the exception of the conglomerate lenses, dark gray to black phyllite makes up most of the Littleton Formation in the lower-grade zones. Inter-

Figure 5-1. Conglomerate (D1c) at the base of the Littleton Fm. Quartzite and slate pebbles in a fine-grained slate matrix. Location is from an abandoned chicken yard, 100 feet west of Vt. Rt. 5, 0.45 mile N.34°E. from road junction 428′, south of Dutton Pines State Forest, Dummerston.

bedded quartzite and quartzose laminae are present in approximately one-half of the outcrops (Figure 5-2). Pencil-thin quartzose laminae are especially abundant along the western margin of the main outcrop band and give the rock a ribboned appearance (Figure 5-3). Farther east, quartzite interbeds range from one-half to 2 inches thick and are spaced as much as 6 inches apart. These quartzites are light gray and fine-grained, but are coarser-grained than the pencil-thin laminae to the west. Bedding is readily shown by the quartzite beds, but the most prominent foliation in many outcrops is a secondary foliation oblique to the bedding and parallel to the axial surfaces of folds. A belt of slate is present in the Littleton Formation along the western margin of the main outcrop area. In this belt, the quartzose laminae are missing; the foliation is essentially uninterrupted by crinkles; and the phyllitic sheen is considerably diminished, producing a regular slate that was quarried for a number of years.

Fine-grained quartz and muscovite are the principal constituents of all the slates, phyllites and quartzites in the low-grade zones.

Figure 5-2. Interbedded phyllite and laminae of quartzite in the Littleton Formation that have been complexly folded. Pencil in center is 4 inches long. Outcrop is at 740′ contour on north end of the north-south ridge, 0.9 miles northwest of the summit of Wantastiquet Mountain, Chesterfield.

Figure 5-3. Interbedded phyllite and quartzite in the Littleton Formation, along Broad Brook, 0.65 mile east of the village of Guilford.

Chlorite is present as medium-grained porphyroblasts and as the alteration product of garnet and biotite, although it is common in only a few specimens. Medium-grained chloritoid was noted in three thin sections but could not be identified in hand specimen. Garnet increases in the Littleton in both amount and grain size from west to east as the grade of metamorphism increases. Biotite is notable mainly for its rarity in rocks of the low-grade zones, even in specimens with large amounts of fresh garnet. Elongate, prismatic aggregates of muscovite and chlorite with square cross sections were noted in one locality 1.65 miles southeast of the village of Guilford. These are believed to be pseudomorphs after andalusite that were formed by contact metamorphism adjacent to a small granodiorite intrusion and that have since undergone retrograde metamorphism. It is considered unlikely that the andalusite formed during the main episode of progressive metamorphism. Very fine-grained, opaque, carbonaceous material disseminated along foliation planes accounts for the dark color of the phyllites and slates of the low-grade zones. Among the other accessory constituents, leucoxene stands out strikingly in many hand specimens as white or light gray, disk-shaped masses up to 4 mm in diameter and 1 mm thick. Accessory iron sulfide is much less common than in the Partridge Formation, although cubes of pyrite are widely scattered throughout the phyllite along the western border of the main outcrop band.

Rock Description in the Staurolite Zone. Schist, quartzose schist and quartzite alternate in the Littleton Formation of the staurolite zone. Both quartzite and schist weather to various shades of gray. Some outcrops have localized areas with a rust-colored rind, but the formation as a whole has a distinctly gray to steel-gray aspect. Fresh schist is dark gray to black, and the quartzites are light gray on fresh surfaces. The schists grade from fine-grained to coarse-grained and consist of quartz, muscovite, and biotite, with porphyroblasts and pseudomorphs after biotite, garnet and staurolite. Finely disseminated graphite accounts for the steel-gray color of the schists. The medium-grained quartzites form thin to thick interbeds within larger bodies of schist. They are thickest and most abundant on Wantastiquet Mountain in Chesterfield township, where they are also much duplicated by folding.

Many of the staurolite porphyroblasts in the staurolite zone of metamorphism have been extensively altered to muscovite and chlorite. The pseudomorphs of staurolite are dull gray to brown and may attain a length of 75 mm, although the average length is about 10 mm. Pseudomorphs of staurolite are especially abundant in the vicinity of quartz veins.

Porphyroblasts of garnet, which average 2 mm in diameter in the staurolite zone, are commonly fresh; but a few are armored with a thin rim of chlorite. Tourmaline, plagioclase, dusty carbonaceous matter and ilmenite, partly altered to leucoxene, are common accessories in the staurolite zone.

Origin. The phyllites, slates and schists of the Littleton Formation were originally black marine shales with interbedded quartzose sandstones. The origin of the mud matrix-supported conglomerate is more difficult to explain. The pebbles are too sparse for the conglomerate to have formed by mud-filling in a preexisting porous network of pebbles. Such matrix-supported conglomerates are generally regarded as debris flows (Harms et al., 1975). They may result from subaqueous slumps and mudflows, and have been frequently observed in association with deepwater turbidy current deposits.

Thickness. Structural complications in the Brattleboro area make the thickness of the Littleton Formation impossible to accurately determine, but 5,000 to 6,000 feet appears to be a reasonable estimate of the minimum thickness.

INTRUSIVE IGNEOUS ROCKS

General Statement. Rocks of two major plutonic series, the Oliverian Plutonic Series and the New Hampshire Plutonic Series, are exposed in the eastern Brattleboro area. A roughly elliptical tract in the southeast part of the area and a small area about the summit of Streeter Hill in Chesterfield are composed of rocks of the Oliverian Plutonic Series. These rocks form parts of the cores of two gneiss domes, the Westmoreland lobe of the Keene dome in the north, and the Vernon dome to the south. Rocks of the New Hampshire Plutonic Series form relatively small bodies of aplite, granodiorite and quartz diorite throughout the area. Moore (1949) has given a complete description of the petrography and areal relationships of these units. A brief summary is presented here.

Oliverian Plutonic Series

Rocks of the Oliverian Plutonic Series in the northern gneiss dome (Keene dome) consist of granodiorite and quartz monzonite gneiss; in the southern gneiss dome (Vernon dome) they consist of quartz diorite gneiss. The gneisses are light gray to grayish-pink, medium-grained, sub-porphyritic rocks with hypidomorphic to granoblastic texture. Much of the quartz shows undulatory extinction and granulation; mortar and sutured textures are common. Flattened and granulated eyes of milky-white quartz are locally present. A specimen from a railroad cut in the central part of the Vernon dome is notable for containing 1 per cent zircon. Foliation in the Oliverian Plutonic Series is best developed around the margins of the gneiss domes. Layers of quartz diorite gneiss as much as 100 feet thick alternate with beds typical of the Ammonoosuc Volcanics at the contact between the two units in the Vernon dome.

New Hampshire Plutonic Series

Numerous small stocks or bodies of granodiorite and less commonly aplite, assigned to the New Hampshire Plutonic Series, are present throughout the eastern Brattleboro area. They are most abundant just west of the band of the Clough Formation in Vernon and east of the Clough in Westmoreland. The bodies range in size from those large enough to be shown on Plate I down to small lenses of dikes 20 to 30 feet wide. Both concordant and discordant contacts have been observed. The granodiorites are light gray to medium gray and generally medium-grained. One third are well foliated and the rest, massive. The well-foliated granodiorites have undergone intense deformation. Both quartz and feldspar have undergone granulation. The plagioclase in these rocks is saussuritized oligoclase with a slight greenish cast. The aplites are white, medium-grained, muscovite-rich rocks with a granitic composition and typical aplitic texture.

AGE AND CORRELATIONS, EASTERN SEQUENCE

The formations of the Eastern Sequence are unfossiliferous in the Brattleboro area, but are coextensive with fossil-bearing rocks elsewhere, notably a few miles to the south near Bernardston, Massachusetts, and to the north in the vicinity of Springfield, Vermont and Claremont, New Hampshire (Boucot et al., 1958; Boucot and Thompson, 1963). These fossil occurrences are in the upper part of the Clough Formation and indicate a late Early Silurian age. Fossils from the Fitch and Littleton Formations farther north (Billings and Cleaves, 1934; Boucot and Arndt, 1960; Harwood and Berry, 1967) indicate that these formations are Silurian and Early Devonian, respectively. Harwood and Berry (1967) also report Middle Ordovician graptolites from black slates, probably correlative with the Partridge Formation. Radiometric dating of the Ammonoosuc Volcanics (Naylor 1969, 1975) is also consistent with an

Ordovician age for this unit.

Naylor (1968) has presented evidence that the Oliverian Plutonic Series is Ordovician in age and generally intrusive into the Ammonoosuc Volcanics. The 377-383 m.y. date on the Black Mountain Granite (Naylor, 1971, see discussion in Chapter 2) is assumed to be approximately the same age as the other small bodies of granodiorite assigned to the New Hampshire Plutonic Series in the quadrangle.

CHAPTER 6
STRUCTURE OF THE EASTERN TERRANE

Introduction

The geologic features in the eastern portion of the Brattleboro area, underlain by rocks of the Eastern Sequence, are more akin to those of western New Hampshire than those of the rest of eastern Vermont. The dominant structural feature of western New Hampshire is the Bronson Hill anticlinorium (Billings, 1937, 1956), characterized by a series of gneiss domes in a more or less *en echelon* array. Most of these gneiss domes lie in a zone about 20 miles wide just east of the Connecticut River in western New Hampshire. The domes of the Bronson Hill anticlinorium extend southward through central Massachusetts and Connecticut to Long Island Sound, east of New Haven.

The general geology of the Bronson Hill anticlinorium in the latitude of the Brattleboro area has recently been described elsewhere (Trask, 1964; Trask and Thompson, 1967; Thompson et al., 1968; Thompson and Rosenfeld, 1979; Robinson et al., in press). Although the gneiss domes are the most conspicuous feature of the Bronson Hill anticlinorium, they represent a relatively late stage in its structural evolution, and were preceded by a series of fold nappes having hinge-to-hinge displacements of as much as fifteen miles. The core gneisses of two of the domes are exposed in the Brattleboro area. These belong to the Vernon dome and to the Westmoreland lobe of the Keene dome (Moore, 1949; Trask, 1964). The staurolite schists on Bear Hill, north of Putney Village, are believed to belong to the mantling cover of yet a third dome, the Wellington Hill anticline of Thompson and Rosenfeld (1979). Rocks assigned to two of the major nappes, the Bernardston and Skitchewaug nappes, are also exposed here. The positions of these features are indicated in Figures 3-1 and 6-1. The Bernardston and Skitchewaug nappes were once thought (Thompson et al., 1968) to be coextensive, but subsequent work (Thompson and Rosenfeld,

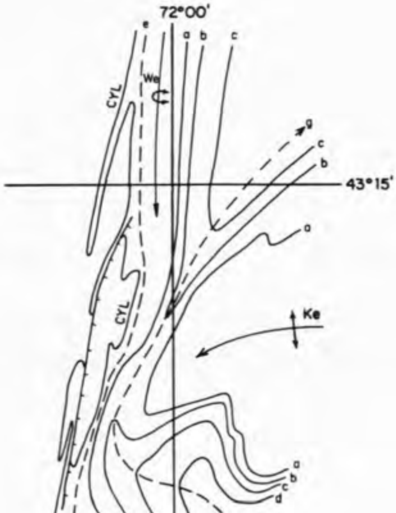

Figure 6-1. Axial elements east of the Chicken Yard line (CYL). Traces of axial surfaces of the nappe-complex are shown by solid lines. a-a is synclinal surface beneath the Bernardston nappe; b-b is the axial surface of the Bernardston nappe; c-c is the synclinal surface between the Bernardston and Skitchewaug nappes; and d-d is the axial surface of the Skitchewaug nappe. Post-nappe axial elements include: e-e, axial surface of the Brattleboro syncline; Ve, Ke, and We, the axes of the Vernon and Keene (Westmoreland lobe) domes, and of the Wellington Hill anticline, respectively; and g-g the axial trace of a major synformal fold that deforms the older axial surfaces. Arrows on g-g show plunges away from an axial culmination west of the Keene dome.

1979) has shown that they are distinct, the Skitchewaug nappe occupying the higher tectonic level. The axial surfaces of both nappes have been strongly deformed by the later rise of the gneiss domes. Subsequent tectonic events include an east-side-north displacement shown by steeply plunging folds at and near the boundary between the rocks of the eastern and western sequences (the "Chicken Yard line"), and post-metamorphic high-angle faulting, believed to be associated with the Mesozoic rift systems of the Connecticut valley.

Vernon Dome

The core gneisses of the Vernon dome and the rocks of its western and northern mantle crop out in the southeast part of the area, athwart the Connecticut River. The gneisses of the core are assigned to the Oliverian Plutonic Series; and the Ammonoosuc, Clough, and Littleton Formations form the surrounding mantle. The bedding in the mantling strata along the west side of the dome dips steeply east. Minor folds and cobble elongations along the west side of the Vernon dome show gentle southward plunges near the south edge of the map area. These pass through the horizontal, and they plunge northward and eventually northeastward as the north end of the dome is approached. Exposures on Mine Ledge and Wantastiquet Mountain at the north end of the dome show intense minor folding. Detailed analysis by Trask (1964) of minor fold axes, and of other linear features such as the elongation of cobbles and pebbles in the conglomerates of the Clough Formation, show an average plunge of about 35^{0} in a N.65^{0}E. direction. Trask's further analysis, making use of extensive observations outside the present map area, shows that the Vernon dome is a tongue-like body protruding upward toward the west-northwest, as indicated schematically in Figure 6-2. The asymmetry of the earlier minor folds is consistent with a flow of the mantling strata away from the crest of the dome during its emplacement. Many of these earlier folds have been refolded about later axes. These later axes plunge moderately to the north at the end of the dome. Interference patterns produced by polyphase folding on different axes can easily be seen on good exposures, as shown in Figure 6-3.

Keene Dome

The core gneisses of the Westmoreland lobe of the Keene dome are exposed extensively in the Keene quadrangle to the east of the present map area, where they can be seen only in a small area about the summit of Streeter Hill in the northern part of Westmore-

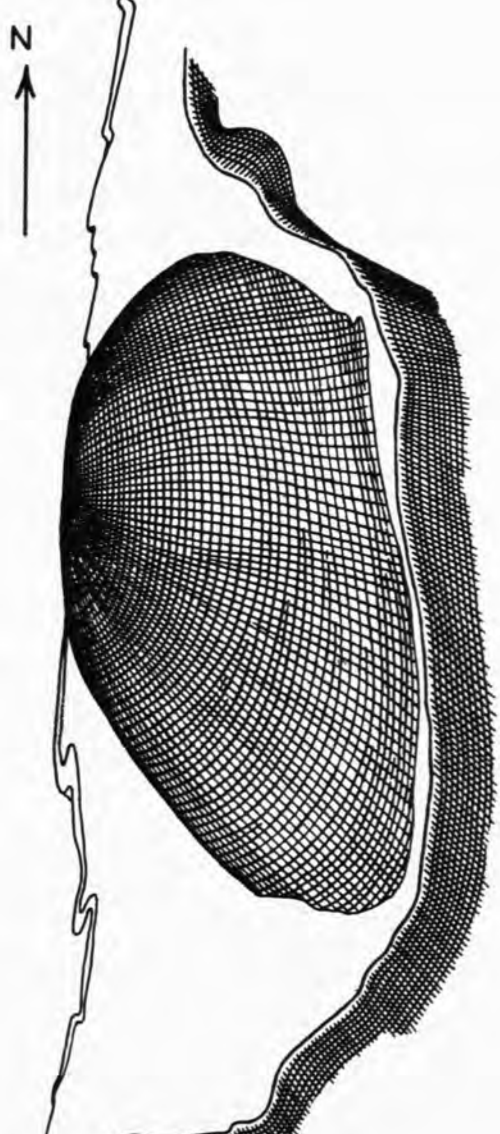

Figure 6-2. Schematic block diagram showing form of the Vernon dome. Horizon shown is the base of the Littleton Formation. View looking down and to the north.

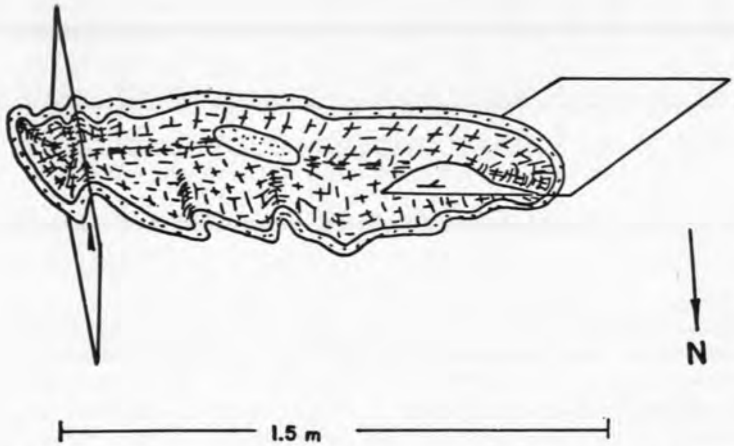

Figure 6-3A. Folded bed of quartzite in Littleton Formation on north side of Wantastiquet Mountain. Plan view with perspective. Axial plane of early fold strikes N.72°W. and dips 60°SW. Early fold has been deformed by late fold with axial plane that strikes N.8°W. and is vertical. Attitude of late folds is relatively uniform in vicinity. Early folds are much deformed. Location, 1000-foot contour due north of summit of Wantastiquet Mountain, Chesterfield township, New Hampshire.

Figure 6-3B. Domes and basins in the Littleton Formation formed by superposed folds. Oblique view looking southwest. Pencil is parallel to axes of early folds that strikes N.75°E. Axes of cross folds strike N.10°W. Knob 1040*w*, 1.1 miles south-southwest of bridge over Connecticut River, Chesterfield township, New Hampshire.

land. The gneisses are mantled successively by the Ammonoosuc, Partridge, Clough, and Littleton Formations. These dip south, southwest, and northwest as one proceeds clockwise around the west end of the lobe. The schists of the Partridge Formation, however, are present only to the south and are absent to the west and northwest, presumably because of an erosional unconformity at the base of the Clough.

The Keene dome is a much larger feature than the Vernon dome, although only the western tip extends into the present area. The west-plunging axis of the Westmoreland lobe corresponds to a marked culmination in the axial surfaces of the tectonically overlying nappes.

Wellington Hill Anticline

Although core gneisses of the Oliverian Plutonic Series are not exposed at the present erosion surface, it is probable that the Wellington Hill anticline of Thompson and Rosenfeld (1979) contains a core of such rocks at depth. In the Bellows Falls quadrangle to the northeast of the present area, erosion has, however, cut down to metavolcanic rocks of the Ammonoosuc Formation. Staurolite grade metamorphism on Bear Hill in Putney suggests that the structure extends southward into the north part of the Brattleboro area. If so, it would be a buried feature comparable in dimensions to the Vernon dome, with which it is in approximate alignment.

Bernardston Nappe

All exposures of the Partridge Formation in the Brattleboro area except those on the south slope of Streeter Hill are believed to lie in the core of the Bernardston nappe. The sulfidic schists of the Partridge are separated in the inverted limb of the nappe from the schists of the structurally underlying Littleton Formation by a discontinuous band of quartzite and conglomerate of the Clough Formation. Minor exposures of calc-silicate granulite of the Fitch Formation were observed between the Clough and Littleton at several localities but were too thin to map separately. The discontinuous nature of this belt of Clough may be in part tectonic in origin. Owing to the structural culmination related to the rise of the Keene dome, the Clough and Fitch on the inverted limb of the nappe have been removed by erosion on the southwest slopes of Bald Hill, north of Governors Brook near the Chesterfield-Westmoreland town line. The Clough and Fitch on the upper, or normal, limb of the Bernardston nappe are less extensively exposed. The trace of these units on the upper limb is in the band passing northwesterly along the west slope of Sargent Hill, then recurving eastward through the village of West Chesterfield.

It is evident from Figure 6-1 that the axial surface of the Bernardston nappe, and the axial surface of the syncline separating the nappe from the relatively autochthonous gneiss domes, have been strongly refolded, both by the rise of the domes and by subsequent events. Minor folds are conspicuous throughout the area of the nappe, as documented by Trask (1964). Some are probably related to the period of nappe formation but cannot now be separated with certainty from those produced by later phases of deformation.

Skitchewaug Nappe

The axial surface of the syncline separating the Bernardston nappe from the structurally higher Skitchewaug nappe lies in the belt of Littleton Formation that passes northwesterly over the crest of Sargent Hill, thence easterly through the southern part of the village of West Chesterfield. Only a small portion of the inverted limb of the Skitchewaug nappe is exposed in the Brattleboro area. This includes the belt of the Fitch Formation that is exposed on the east slope of Sargent Hill and to the north across Catsbane Brook. Graded beds in some of the granulites in this belt of Fitch are consistent (Figure 6-4) with an inverted sequence. Gray schists east of this zone were formerly regarded (Thompson et al., 1968) as part of the Partridge Formation but are now assigned to the Clough, owing to their close association with conglomerates just outside the map area to the east.

The Chicken Yard Line

The boundary between the rocks of the Eastern and Western Sequences of this report has come to be known as the "Chicken Yard line" from the exposures in a former chicken yard on the west side of U.S. Route 5 about 0.6 mile south of Dutton Pines State Forest Park. The line separates conglomerate and slate of the Littleton Formation to the east from greenstone and thinly interbedded slates and sandstones of the Putney Formation to the west. Fine cross-bedding and possible load casts in the sandy laminae of the Putney, and apparent channeling in the conglomeratic beds of the Littleton, all indicate that stratigraphic tops lie to the east.

We interpret this line as an unconformity at the base of the Littleton. This implies that the Shaw Mountain, Northfield, Waits River, Gile Mountain, Putney Volcanics, and the Standing Pond Volcanics, are pre-Littleton, probably ranging in age from Silurian to earliest Devonian. They would then be collectively equivalent to

Figure 6-4. Inverted graded bedding in the upper part of the Fitch Formation, 1.0 mile southeast of village of West Chesterfield, Chesterfield township, New Hampshire. Knife on ledge gives scale. Section view looking southeast. Overturned beds dip gently to southeast.

the Clough and Fitch Formations of the Bronson Hill anticlinorium, and represent a more westerly facies of much greater thickness. Recent mapping by Thompson and Rosenfeld to the northeast in the Bellows Falls-Walpole area has shown that the post-Partridge, pre-Littleton rocks of the Skitchewaug nappe are much thicker and have a higher content of pelitic schist than do the Clough and Fitch as exposed either in the Bernardston nappe or in the mantles of the gneiss domes. Inasmuch as the rocks in the Skitchewaug nappe are tectonically transported from a more easterly source, this would suggest that what is now the Bronson Hill anticlinorium may have been a relatively elevated region during Silurian and earliest Devonian time, with much greater sedimentary accumulation to the east and to the west. This is consistent with the interpretation of Moench and Boudette (1970) of the nature of the southeastern flank of the Bronson Hill anticlinorium (northwestern limb of the Merrimack synclinorium) in northwest Maine.

The above interpretation, however, is not accepted by all workers in adjacent areas. Robinson et al. (in press), for example, regard the Chicken Yard line as the trace of a pre-metamorphic thrust carrying an **older** Littleton Formation westward over the **Younger** Putney and Gile Mountain Formations. This alternative interpretation is based primarily on stratigraphic arguments. Robinson et al. regard their "Erving Formation" of central Massachusetts as correlative with the younger units of our Western Sequence, but also as unconformably overlying the Littleton Formation. We are not convinced, however, that the pre-Erving "Littleton" as mapped in Massachusetts is really equivalent to the Littleton elsewhere, and we propose that it is in fact older. These conflicting views cannot be resolved with the scanty paleontologic and stratigraphic data now at hand. We see no evidence, however, of the large-scale movement required by the "Whately Thrust" hypothesis of Robinson et al. in any of the exposures of the Chicken Yard line that we have examined. This is particularly disconcerting in view of the fact that the metamorphic grade is lower at and near the Chicken Yard line than anywhere else in the Brattleboro area, and of the fact that primary sedimentological features are better preserved in this zone than elsewhere.

The Brattleboro Syncline

Our interpretation of the Chicken Yard line places the axis of the Brattleboro syncline in the Littleton Formation east of the Chicken Yard line and west of the domes and nappes of the Bronson Hill anticlinorium. The axis is thus in an area of steep dips and at least three generations of complex folding. The earliest major folds are about subhorizontal or gently plunging axes that appear to carry the Littleton beneath (by projection) the older Gile Mountain and Putney Formations. These folds have superposed on them a set of steeply plunging sinistral folds indicating an east-side-north displacement. These, in turn, are modified by a still younger set of gently plunging folds (Figure 6-5). Both of the latter two sets are associated locally with kink-banding.

Post-Metamorphic Faulting

The last tectonic event to have affected the rocks of the Brattleboro area is high-angle, post-metamorphic faulting, presumably related to the Mesozoic rift system of the Connecticut Valley. Only two of these faults have sufficient displacement to be shown in the geologic map. One, just west of the Connecticut River in Dummerston and part of Brattleboro and Putney, offsets the

Figure 6-5. Phyllite of the Littleton Formation with two sets of crinkles on one surface of foliation. Road cut on Route 5, 1.1 miles southeast of village of Guilford.

Chicken Yard line. The main movement appears to be down on the east side, but near-horizontal slickensides at one locality indicate a late-stage, strike-slip component.

A relatively small fault is exposed in a gully on the southwest side of Daniels Mountain in Hinsdale. The Ammonoosuc, Clough, and Littleton Formations are offset across the gully with the relative movement, again, down on the east. The fault strikes N. 20°E. and has a stratigraphic throw of about 600 feet.

CHAPTER 7

METAMORPHISM OF THE EASTERN TERRANE

Introduction.

The grade of metamorphism of the Eastern Brattleboro area ranges from low in the west to high in the east. Rocks of low metamorphic grade occupy the core of the Brattleboro syncline (Plate I) and extend northward along the Connectiuct River-Gaspé synclinorium into Canada (Doll et al., 1961). Eastward, the grade of metamorphism increases to the sillmanite zone over the Bronson Hill anticline to the east of the area. The isograds generally parallel the regional north-south strike of the major structural features.

Isograds are drawn on the basis of the first appearance of biotite, garnet, and staurolite in the pelitic rocks. Mafic igneous rocks also show systematic differences in mineralogy according to metamorphic grade but are less useful as an index of metamorphism in this area. Calcareous rocks are restricted to one grade of metamorphism with minor exceptions. The mineralogy of the pelitic rocks, mafic igneous rocks, and calcareous rocks will be discussed in turn below.

Pelitic Rocks.

In the eastern Brattleboro area, lenticular bands of slate and rocks with approximately equal amounts of quartz, muscovite and chlorite occur along the western border of the Littleton Formation and probably belong to the chlorite zone of metamorphism. Most of the phyllites, slates and quartzites of the Littleton Formation in this area consist almost wholly of fine-grained quartz and muscovite. Thus, mapping of the biotite isograd east of the Connecticut River is difficult and is only shown approximately on Plate I. Rocks immediately west of the garnet isograd are in the biotite zone, although biotite is scarce due to composition. The mineral pair chloritoid-chlorite with quartz and muscovite is present in these rocks. Figure 7-1A shows the probable form of the AFM projection for the chlorite and biotite zones in the eastern Brattleboro area and the mineral assemblages observed.

The garnet zone is characterized by the assemblages garnet-chloritoid-chlorite and garnet-biotite-chlorite with quartz and muscovite. Some of the chlorite in the garnet zone is believed to be the product of the retrograde alteration of garnet. Figure 7-1B shows the AFM projection for the garnet zone and the assemblages observed. Note that the coexistence of biotite and chloritoid was not observed.

Much of the chlorite in the staurolite zone is also the product of retrograde alteration. However, a significant amount of staurolite is present as large, clear porphyroblasts without any textural evidence that it has replaced earlier ferromagnesian minerals. Assemblages in the staurolite zone are listed in Table 7-1A and shown on Figure 7-1C. The four phase AFM assemblage, biotite-garnet-staurolite-chlorite, appears to violate the phase rule. As noted previously (Chapt. 4), the two most likely explanations for the four-phase AFM assemblage are: (a) the chlorite is in fact the product of retrograde metamorphism and did not form at the same time as the other three minerals, or (b) the garnet is the extra phase and coexists with the other three because it contains additional compo-

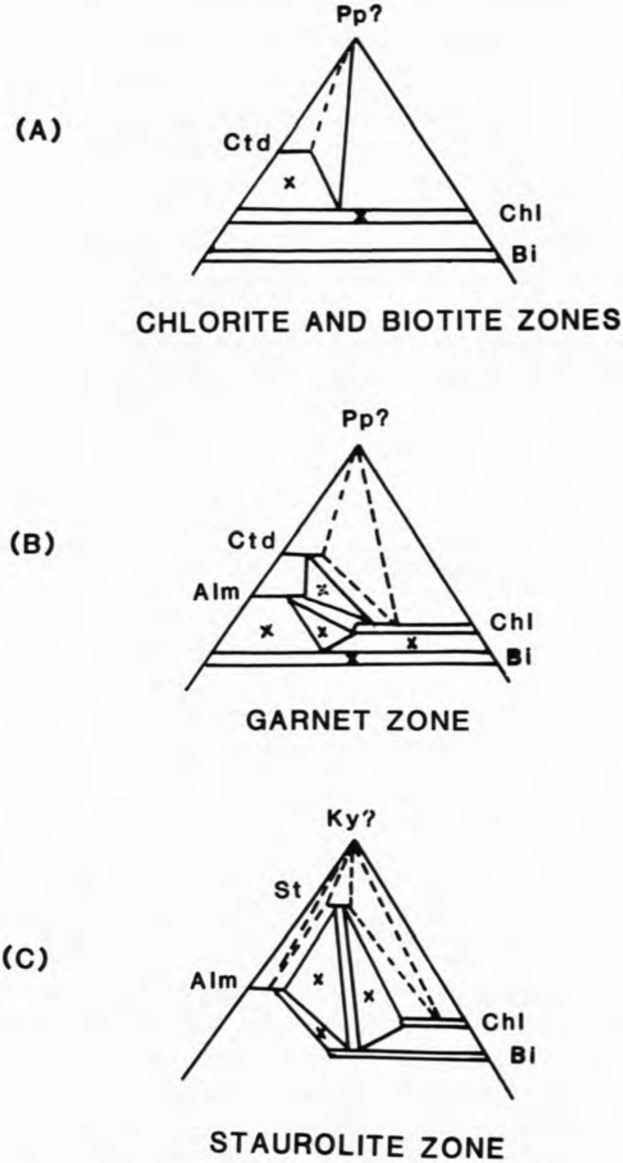

Figure 7-1. Observed mineral assemblages in the pelitic rocks of the Eastern Brattleboro area plotted on AFM projections (Thompson, 1957). X's indicate mineral assemblages present with quartz and muscovite. Inferred geometry of projection for each metamorphic zone shown by dotted lines. Pp = pyrophyllite, Ctd = chloritoid, Chl = chlorite, Bi = biotite, Alm = almandine, St = staurolite, Ky = kyanite.

A. PELITIC ROCKS

Chlorite and Biotite zones	All Assemblages Include Quartz and Muscovite
Chlorite	
Chlorite-andalusite (?)	
Chlorite-chloritoid	
Garnet zone	
Biotite	
Biotite-garnet	
Biotite-chlorite	
Biotite-garnet-chlorite	
Garnet-chlorite-chloritoid	
Staurolite zone	
Biotite-garnet	
Biotite-garnet-chlorite	
Biotite-garnet-staurolite	
Biotite-chlorite-staurolite	
Biotite-garnet-staurolite-chlorite	

B. MAFIC IGNEOUS ROCKS

Staurolite zone
Quartz-plagioclase-hornblende
Quartz-plagioclase-hornblende-biotite
Quartz-plagioclase-hornblende-garnet
Plagioclase-hornblende
Quartz-plagioclase-epidote-hornblende

Table 7-1. Observed mineral assemblages in the (A) Pelitic rocks, and (B) Mafic Igneous rocks, of the Eastern Brattleboro quadrangle.

nents not represented in the AFM diagram. In Figure 7-1C, it is assumed that the chlorite formed in equilibrium with the other phases, but other interpretations are possible. Kyanite is thought to be the stable alumino-silicate in the staurolite zone, but was not actually observed here.

Mafic Igneous Rocks.

Metamorphosed mafic igneous rocks are not present in the eastern Brattleboro area below the upper garnet zone. Here amphibolites in the Partridge Formation commonly contain the assemblage hornblende-plagioclase. Most of the metamorphosed mafic igneous rocks of the area are amphibolites in the Ammonoosuc and Partridge Formations metamorphosed to staurolite zone conditions. Table 7-1B shows the observed assemblages in these rocks, typified by hornblende-plagioclase±garnet±epidote.

CALCAREOUS ROCKS

Staurolite zone

Quartz-biotite-muscovite-epidote-garnet
Quartz-epidote-actinolite
Quartz-biotite-epidote-actinolite
Quartz-biotite-actinolite-calcite
Quartz-muscovite-epidote
Quartz-biotite-calcite-garnet-actinolite-plagioclase
Biotite-plagioclase-actinolite-diopside-epidote
Quartz-plagioclase-diopside-calcite-epidote
Quartz-actinolite-diopside-epidote-calcite
Quartz-microcline-actinolite-diopside-epidote
Quartz-biotite-epidote-garnet-calcite
Quartz-plagioclase-biotite-epidote-muscovite-garnet

Table 7-2. Mineral assemblages observed in metamorphosed calcareous rocks, Eastern Brattleboro area.

Calcareous Rocks.

Table 7-2 summarizes the mineral assemblages found in the metamorphosed calcareous rocks of the Fitch and Partridge Formations. The Fitch Formation in the staurolite zone in Chesterfield consists mostly of various combinations of the minerals actinolite, diopside, epidote, plagioclase, calcite, biotite, garnet and quartz.

Retrograde Metamorphism.

The effects of retrograde metamorphism are widespread in the eastern Brattleboro area. They are particularly abundant where slip cleavage is prominent on the east flank of the Brattleboro syncline. No doubt the planes of the slip cleavage provided pathways for later migrating fluids. Staurolite has been altered to muscovite and chlorite along the western edge of the staurolite zone. Garnet is commonly armored with chlorite in some cases and completely replaced in others. Biotite has been partly altered to chlorite with a patchy distribution.

CHAPTER 8
ECONOMIC GEOLOGY

Slate and granite were once extensively quarried in the area, although currently there are no active bedrock quarries or mines in the quadrangle. Traces of sulfide mineralization have been reported in the area, and there is some potential for soapstone or talc.

The major economic deposits currently being exploited are sand and gravel from the Connecticut River Valley.

The Black Mountain Granite is a light-colored two-mica stone known as "West Dummerston White Granite" (Dale, 1909) and is quite similar to the well known Barre Granite. It was actively quarried as a building stone in the early part of the century. The Presbrey-Leland quarry (on the east side of the West River, east of West Dummerston Village) was the largest quarry in the Black Mountain Granite, although numerous smaller quarries were active for short periods of time. The sheeting in the granite (Figure 3-17) greatly aided the removal of the stone. This granite forms an excellent building stone and has an ultimate compressive strength of 27,810 pounds per sq. inch (Dale, 1909). Although the sheeting becomes thicker with depth, making quarrying more difficult, considerable reserves of this granite still exist. Most quarrying operations stopped in the 1930's when the railroad line along the West River failed, making removal of the granite difficult.

Numerous slate quarries were active in the 19th and 20th centuries in and around Brattleboro and Guilford (Dale, 1914) and supported a moderate-sized industry. Most of the slate was used as roofing slate and was well known in the region as Guilford or Brattleboro "gray slate". The quarried slate was almost exclusively from a band of chlorite zone Littleton Formation just east of the Putney Volcanics. The rock here is quite regular, has few crinkles, and the cleavage generally parallels bedding or crosses it at only a low angle. Considerable reserves of gray slate exist in the area, if demand should ever return.

The small "tear-dropped" shaped ultramafic body in the very northwestern corner of the area (Plate I) likely contains the potential for soapstone or talc. One small pit was found dug into the body, but its potential is largely unknown.

Scattered occurrences of sulfide mineralization, other than pyrite, were noted in two formations, the Standing Pond Volcanics and the Ammonoosuc Volcanics. Pyrrhotite occurs as scattered grains in amphibolites of band #2 (Figure 2-5) of the Standing Pond Volcanics on the southeast side of the Guilford dome, south of Guilford Center. Massive sulfides were mined from the Standing Pond Volcanics to the north in the Strafford quadrangle, Vermont (Elizabeth Mine, see Howard, 1969). However, nothing was observed in the present study to indicate the mineralization in the Brattleboro area is of economic value, although detailed studies of mineralization were not undertaken.

Mineralized areas of the Ammonoosuc Volcanics are known from locations outside the Brattleboro quadrangle. Although scattered sulfide minerals were occassionally noted in the Ammonoosuc Volcanics, no areas of economic potential were found in the present study. Both the Standing Pond and Ammonoosuc Volcanics are potentially worth further investigation for sulfide deposits.

Large ecomonically valuable sand and gravel deposits occur in the valleys of the larger rivers in the area, particularly along the Connecticut River. Interested readers are referred to Stewart and MacClintock (1969) for more information on the Pleistocene geology of the area.

CHAPTER 9

GEOLOGIC HISTORY

Connecticut River-Gaspé Synclinorium
East Flank of Green Mountain Anticlinorium

The inferred generalized geologic history of the western Brattleboro area is summarized as follows:

1. Precambrian: Development of a metamorphic complex exposed in the core of the Athens dome, approximately 1,000 m.y. ago.
2. Cambrian-Middle Ordovician: Deposition of marine sediments and bimodal volcanics (Hoosac through Missisquoi Formations) on the eroded Precambrian basement.
3. Late Ordovician-Early Silurian (Taconic orogeny): Non-deposition, probable thrusting, tectonic emplacement of small ultramafic bodies, metamorphism, uplift and consolidation, followed by erosion and development of two regional unconformities, one above and one below the unnamed schist-amphibolite unit.
4. Silurian: (A) Deposition of the sands and conglomerates of the Russell Mountain Formation over the Late Ordovician unconformity surface, followed by volcanism; (B) Development of a structural basin as a result of the Taconic events, with deposition into it of a thick sequence of mudstones (Northfield Formation), interbedded muds and calcareous sands (Waits River Formation), and interbedded sands and muds (Gile Mountain Formation), mostly under deep water conditions. Volcanic activity, largely mafic, during this time produced the Standing Pond and Putney Volcanics.

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UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

BRATTLEBORO QUADRANGLE
VERMONT—NEW HAMPSHIRE
15 MINUTE SERIES (TOPOGRAPHIC)

Mapped, edited, and published by the Geological Survey

Control by USGS and USC&GS

Topography from aerial photographs by multiplex methods
Aerial photographs taken 1951. Field check 1953 and 1954

SCALE 1:62500

CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL

APPROXIMATE MEAN DECLINATION, 1954

QUADRANGLE LOCATION

ROAD CLASSIFICATION

Heavy-duty — Light-duty
Medium-duty — Unimproved dirt
U. S. Route — State Route

PLATE I

GEOLOGIC MAP

BRATTLEBORO, VT.—N. H.
N4245—W7230/15

1954

LEGEND

STRATIFIED METAMORPHIC ROCKS

"WESTERN SEQUENCE"

EAST FLANK GREEN MOUNTAIN ANTICLINORIUM
WEST FLANK CONNECTICUT RIVER—GASPÉ SYNCLINORIUM

LOWER DEVONIAN

SILURIAN

Spv

PUTNEY VOLCANICS

Spv: Buff to light brownish weathering feldspathic granulite and feldspathic phyllite; light greenish gray chloritic phyllite; greenstone; interbedded gray slate or phyllite.

Sgm
Sgmm

GILE MOUNTAIN FORMATION

Sgm: Interbedded light gray to tan, micaceous and feldspathic quartzite and mica schist with porphyroblasts of biotite, garnet, kyanite and staurolite (western band). Gray phyllite and slate with interbedded, thin, micaceous quartzite and rare, punky-brown weathering impure marble (eastern band).

Sgmm: Marble member, dark gray micaceous phyllite and schist with prominent interbeds of punky-brown weathering impure marble.

Ssp

STANDING POND VOLCANICS

Ssp: Black medium-grained amphibolite and epidote amphibolite; very coarse-grained garnet-hornblende fasciculitic schist; minor impure quartzite, coticule and schist (western bands). Dark gray to gray-green, massive plagioclase-biotite-quartz and plagioclase-biotite-hornblende-quartz-carbonate granulite and gneiss (eastern band).

Swrq
Swr

WAITS RIVER FORMATION

Swr: Dark gray to gray mica schist and calcareous mica schist with abundant interbeds of punky-brown-weathering impure marble; thin interbeds of impure quartzite toward eastern contact. Zone where impure marble averages 1% to 5% in gray garnet-mica schist is indicated on the map.

Swrq: Quartzitic member, gray, feldspathic and micaceous quartzite interlayered with muscovite schist; rare, thin beds of coarse-grained to fasciculitic garnet amphibolite.

Sn

NORTHFIELD FORMATION

Sn: Gray, commonly crinkled, mica schist with conspicuous almandine porphyroblasts; thin impure quartzite interbeds; minor impure marble.

Srm

RUSSELL MOUNTAIN FORMATION

Srm: Light gray to white quartz-pebble conglomerate and quartzite; coarse-grained hornblende fasciculite schist; minor amphibolite, garnet amphibolite and mica schist.

ORDOVICIAN OR SILURIAN

OSsaa
OSsa

UNNAMED SCHIST—AMPHIBOLITE UNIT

OSsa: Rusty-weathering, fine grained black carbonaceous phyllite and schist with thin black quartzite interbeds; medium-grained porphyritic and non-porphyritic black amphibolite; minor gray mica schist with garnet porphyroblasts; distinctive thin coticule layers.

OSsaa: Zone where porphyritic and non-porphyritic amphibolite predominates with lesser amounts of rusty-weathering black phyllite and felsic schist.

ORDOVICIAN

Ombv
Omm

MISSISQUOI FORMATION

Ombv: Barnard Volcanic Member, dark gray to black amphibolite and porphyritic amphibolite with feldspar megacrysts; gray to light gray quartzo-feldspathic gneiss, schist, granulite, and layered gneiss with interlayers of amphibolite; minor black, rusty-weathering mica schist; rare metatuff with elongated quartz grains (X).

Omm: Moretown Member, gray to light gray interbedded, quartzite, quartz-mica schist, silvery mica schist and impure quartzite; garnets are a common porphyroblast in the schist; distinctive laminated "pinstriped" quartzite is locally abundant; 10% to 20% black amphibolite and garnet amphibolite; rare quartz-pebble conglomerate or metatuff (X).

CAMBRIAN

€o

OTTAUQUECHEE FORMATION

€o: Dark gray to silvery graphitic mica schist with abundant pyrrhotite; conspicuous garnet porphyroblasts; thin-laminar micaceous granulite; minor epidote amphibolite; rare coticule.

€ph

PINNEY HOLLOW FORMATION

€ph: Light green quartz-muscovite-chlorite phyllite and schist; garnet and magnetite porphyroblasts common; quartz lenses abundant; black, thinly laminated ligniform epidote amphibolite.

LATE PRECAMBRIAN? or CAMBRIAN?

p€-€h
p€-€hs

HOOSAC FORMATION

p€-€h: Albite-rich mica schist and gneiss; albite forms conspicuous porphyroblasts or small augen; minor biotite-epidotite amphibolite.

p€-€hs: Light gray to silvery muscovite-quartz schist with conspicuous porphyroblasts of garnet, staurolite and kyanite.

p€-€bh

BULL HILL GNEISS

p€-€bh: Augen gneiss, medium-grained, massive quartz-biotite-plagioclase gneiss with conspicuous microcline augen or flaser; quartz-plagioclase-biotite ± muscovite gneiss.

PRECAMBRIAN

p€m

MOUNT HOLLY COMPLEX

p€m: Quartz-biotite-plagioclase gneiss with sporadic muscovite, microcline, garnet and epidote.

"EASTERN SEQUENCE"

WEST FLANK BRONSON HILL ANTICLINORIUM
EAST FLANK CONNECTICUT RIVER—GASPÉ SYNCLINORIUM

Dl
Dlc

LITTLETON FORMATION

Dl: Gray to dark gray slate, phyllite and mica schist with interbedded quartzite; biotite, garnet and staurolite porphyroblasts common at the appropriate metamorphic grades.
Dlc: Polymict conglomerate with a fine-grained slate to phyllitic matrix; vein quartz and slate pebbles abundant to scarce.

Sf

FITCH FORMATION

Sf: Purplish-brown quartz-biotite granulite and granulitic schist with porphyroblasts of garnet and actinolite common; greenish-gray calc-silicate granulite, rare sulfidic calc-silicate granulite and thin beds of quartzite.

Scs
Sc

CLOUGH QUARTZITE

Sc: White to light-gray quartz-pebble and quartzite-pebble conglomerate in a quartzite and quartz mica schist matrix; white quartzite; muscovite schist.

Scs: Gray mica schist.

Opas
Ops

PARTRIDGE FORMATION:

Ops: Rusty-brown weathering sulfidic mica schist; gray quartz-mica schist; biotite, garnet and staurolite porphyroblasts present at appropriate metamorphic grades; lesser amounts of amphibolite and quartz-feldspar-biotite gneiss and granulite.

Opas: Interbedded schist and amphibolite as above with amphibolite predominant over schist.

Oam

AMMONOOSUC VOLCANICS

Oam: Light-gray, well-foliated quartz-feldspar-biotite gneiss with thick interbeds of amphibolite more abundant in the lower part of the formation; minor sulfidic muscovite-quartz schist near top; scattered lenses of breccia.

PLUTONIC ROCKS

MIDDLE DEVONIAN

nhg

NEW HAMPSHIRE PLUTONIC SERIES

nhg: Medium- to coarse-grained, light-gray, to gray, granite, granodiorite and quartz diorite; some bodies weakly to moderately foliated; aplite.

ORDOVICIAN

ol

OLIVERIAN PLUTONIC SERIES

ol: Medium- to coarse-grained, gray to pink, sub-porphyritic, granodiorite, quartz diorite and quartz monzonite gneiss, strongly to weakly foliated.

um

ULTRAMAFIC ROCKS

um: Massive, dark green serpentinite, gray pitted talc-carbonate, local actinolite alteration at contact with country rock.

MAP SYMBOLS

CONTACTS

Good Control

Fair Control

Inferred

D High-angle fault (D on downthrown side, dashes and dots as above)

Thicknesses of particularly thin units have been slightly exaggerated (locally) so that these units may be shown on the map at this scale.

STRUCTURAL SYMBOLS

56 Strike and dip of bedding (upright or direction of top unknown)

34 Strike and dip of overturned bedding

67 Strike and dip of foliation

Strike and dip of vertical foliation

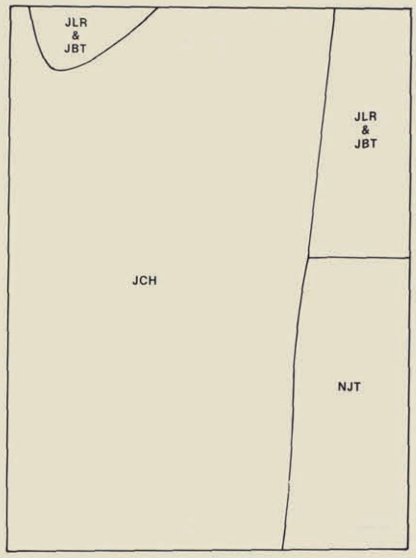
METAMORPHIC ISOGRADS

Bt Isograd, accurately or approximately located, hachures on side of higher metamorphic grade. Bt = Biotite, Gt = Garnet, St = Staurolite, St-Ky = Staurolite-Kyanite, Ch = rocks in the Chlorite Zone.

CONTACT METAMORPHOSED ROCKS

Rocks altered by contact metamorphism in aureole surrounding the Black Mountain Granite, numerous dikes and sills of granite in the country rock as main body is approached.

Larger granitic dikes or sills



KEY TO MAPPING

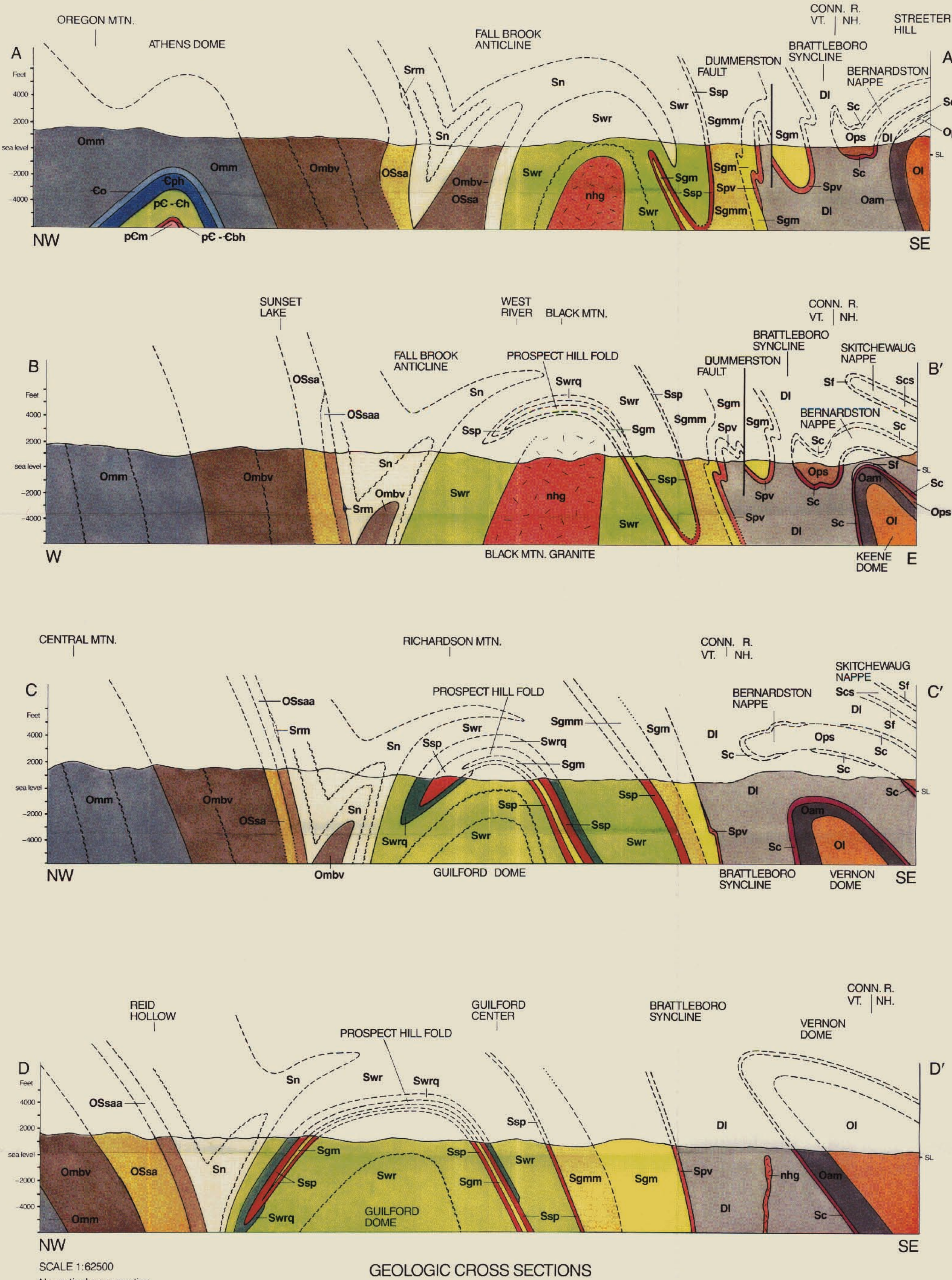
GEOLOGY BY

J. C. HEPBURN, N. J. TRASK, J. L. ROSENFELD, J. B. THOMPSON, JR.

GEOLOGIC MAP OF THE BRATTLEBORO QUADRANGLE VERMONT–NEW HAMPSHIRE

VERMONT GEOLOGICAL SURVEY

Charles A. Ratté, State Geologist
(Bulletin No. 32)



SCALE 1:62500

No vertical exaggeration.
See Plate I for explanation of formation symbols and location of sections

GEOLOGIC CROSS SECTIONS
BRATTLEBORO QUADRANGLE,
VERMONT – NEW HAMPSHIRE

PLATE II