BEDROCK GEOLOGIC MAP OF THE SAXTONS RIVER 7.5'×15' QUADRANGLE, WINDHAM AND WINDSOR COUNTIES, VERMONT

By Nicholas M. Ratcliffe and Thomas R. Armstrong

INTRODUCTION

Bedrock of the Saxtons River 7.5'x15' minute quadrangle ranges in age from Middle Proterozoic to Cretaceous. All of the rocks, except for the late granitic and mafic dikes, are weakly to very strongly deformed, high-grade metamorphic rocks, such as schists and gneisses. The major geologic feature is the domal exposure of Middle Proterozoic gneiss and granite that makes up the core of the Chester and Athens domes in the eastern one third of the map. The narrowed connection, a structural saddle between the domes, forms the juncture between the generally south-plunging end of the Chester dome to the north and the north-plunging Athens dome to the south. From west of the domes to the border of the quadrangle is a broad belt of structurally very complex cover rocks that range in age from Late Proterozoic to perhaps uppermost Ordovician. A conspicuous structure in this belt is the Spring Hill syncline which has rocks of the Cram Hill Formation in its core. To the west, Middle Proterozoic gneisses appear in the core of the Butternut Hill fold. Near the northwestern corner of the map, Middle Proterozoic basement rocks that correlate with the core rocks of the Chester and Athens domes reappear to form the eastern margin of the Green Mountain massif.

East of the Chester and Athens domes, a moderately steeply east-dipping sequence of Late Proterozoic to Early Devonian cover rocks is exposed. This eastern cover sequence generally correlates with similar cover rocks in the western part of the quadrangle. Subtle and perhaps very significant differences in rock types and stratigraphic relations of these two sections may exist. Although the general map pattern and stratigraphic relations of the cover rocks appear simple, their internal structure and relation to underlying basement rocks is quite complex. Quite possibly there is much more extensive faulting and intricate folding in the cover sequence rocks than is shown on the map.

Previous and generally accepted interpretations of the geology (Doll and others, 1961) portrayed large refolded Middle Devonian nappes (Acadian) that involved basement and cover rocks (see cross section *A-A'* on figure 1). The southern terminus of the Middle Proterozoic rocks in the core of the Butternut Hill fold, for example, was interpreted as a north-plunging inverted sequence of rocks in a recumbent west-facing nappe that roots in the core of the Chester dome to the north. The Spring Hill syncline was also mapped as a north-plunging structure at both its northern and southern closures and was interpreted as the upwardly folded and inverted part of the recumbent syncline that was complimentary to the anticlinal nappe. Likewise, a major belt of metavolcanic rocks in the Devonian metasedimentary rocks east of the domes (Standing Pond Volcanic Member of the Waits River Formation) was interpreted as a major, originally east verging, synclinal nappe. We do not agree with these interpretations, and regard the structures as typical anticlinal and synclinal folds.

PREVIOUS STUDIES AND PRESENT EVALUATION

The Saxtons River 7.5'x15' quadrangle covers the northern half of the former Saxtons River 15-minute quadrangle that was studied by Rosenfeld (1954). His unpublished map and cross sections at a scale of 1:62,500 are a valuable resource and important background for his subsequent publications (Rosenfeld, 1968; Rosenfeld and others, 1988; Thompson and others, 1993). The basic distribution of our map units agrees closely with Rosenfeld's (1954) map except for our greater detail, especially in the core of the Chester and Athens domes and in the cover sequence rocks east of the domes. Because of the present larger scale, a more detailed distribution of rock units is possible particularly in the cover rocks west of Grafton, in the Spring Hill area, and within volcanic rocks of the Waits River Formation.

Major differences, however, exist in our interpretation of the principle structures, chiefly in the area of the Butternut Hill-Spring Hill areas where very critical regional geologic structures are present. According to Rosenfeld (1954), the Butternut Hill fold and adjacent Sugarloaf-Star Hill fold (fig. 1) are late-stage folds related to gravitational upwelling of the Chester and Athens domes. Rosenfeld (1954) interpreted these folds as north-plunging synformal and antiformal folds, respectively, that contain inverted rocks previously overturned in the Proctorsville synclinal nappe. In Rosenfeld's (1954) interpretation, the axial trace of the Proctorsville syncline is folded by the Butternut Hill fold, is folded to the north by the Sugarloaf-Star Hill fold, and returns to the south as the axial trace of the Spring Hill fold. Both the northern and the southern terminations of the Spring Hill fold were interpreted to plunge to the north and the rocks in the core pass from right side up to inverted along the axial trace of the synform. Rosenfeld's (1954) ideas are presented in figure 1.

Our interpretations, presented here in figure 2 and on the form-line map on sheet 2, differ substantially from Rosenfeld (1954). We interpret the Butternut Hill fold as a south- or

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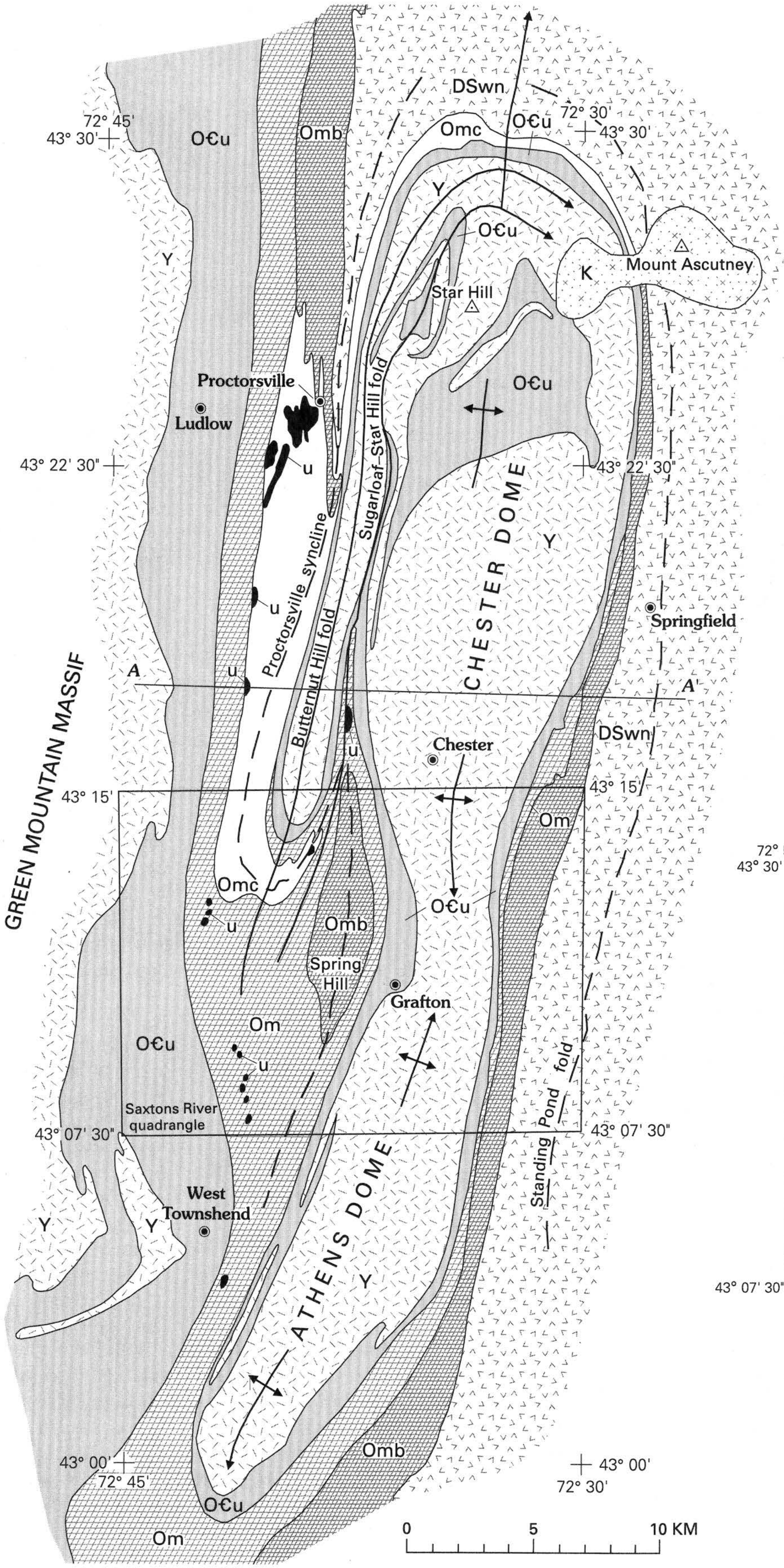

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Sugarloaf–Star Hill fold
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CHESTER DOME
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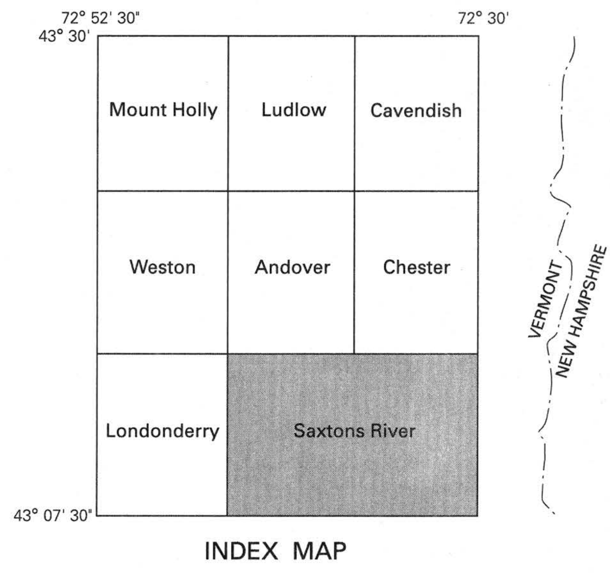
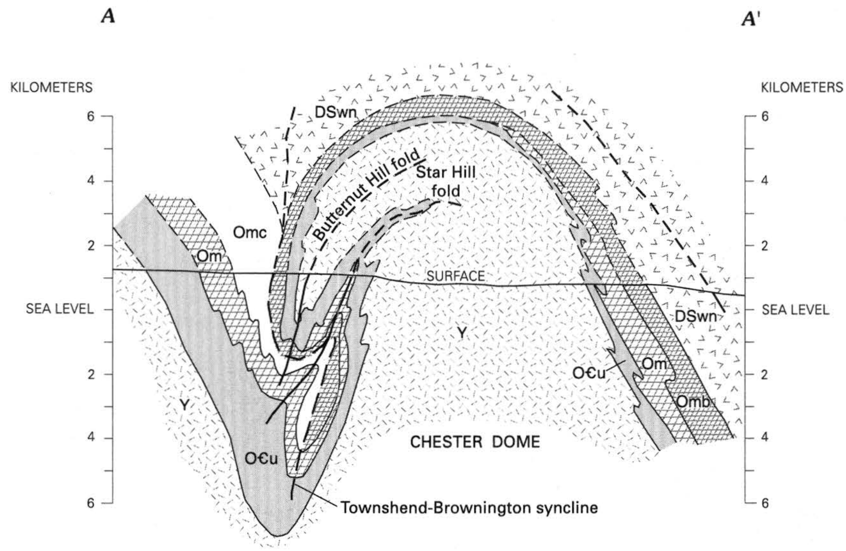


72° 52' 30"
72° 30'
43° 30'
Mount Holly
Ludlow
Cavendish
Weston
Andover
Chester
Londonderry
Saxtons River
43° 07' 30"
VERMONT
NEW HAMPSHIRE

INDEX MAP



EXPLANATION

K — **Alkaline intrusion (Cretaceous)**

D$wn — **Waits River and Northfield Formations, undivided (Devonian and Silurian)**

Omc, Omb — **Rusty schist member** (Omc) **and Barnard Volcanic Member** (Omb) **of Missisquoi Formation (Ordovician)**

Om — **Moretown Member of Missisquoi Formation (Ordovician)**

O€u — **Hoosac, Pinney Hollow, Ottauquechee, and Stowe Formations, undivided (Ordovician to Cambrian)**—Includes Cavendish Formation in Chester dome

Unconformity

Y — **Gneiss and granite, undifferentiated (Middle Proterozoic)**

u — **Ultramafic rocks (age uncertain)**—Included in Omc and Om

Contact

Axial trace of Acadian folds (after Rosenfeld, 1954; and Doll and others, 1961)

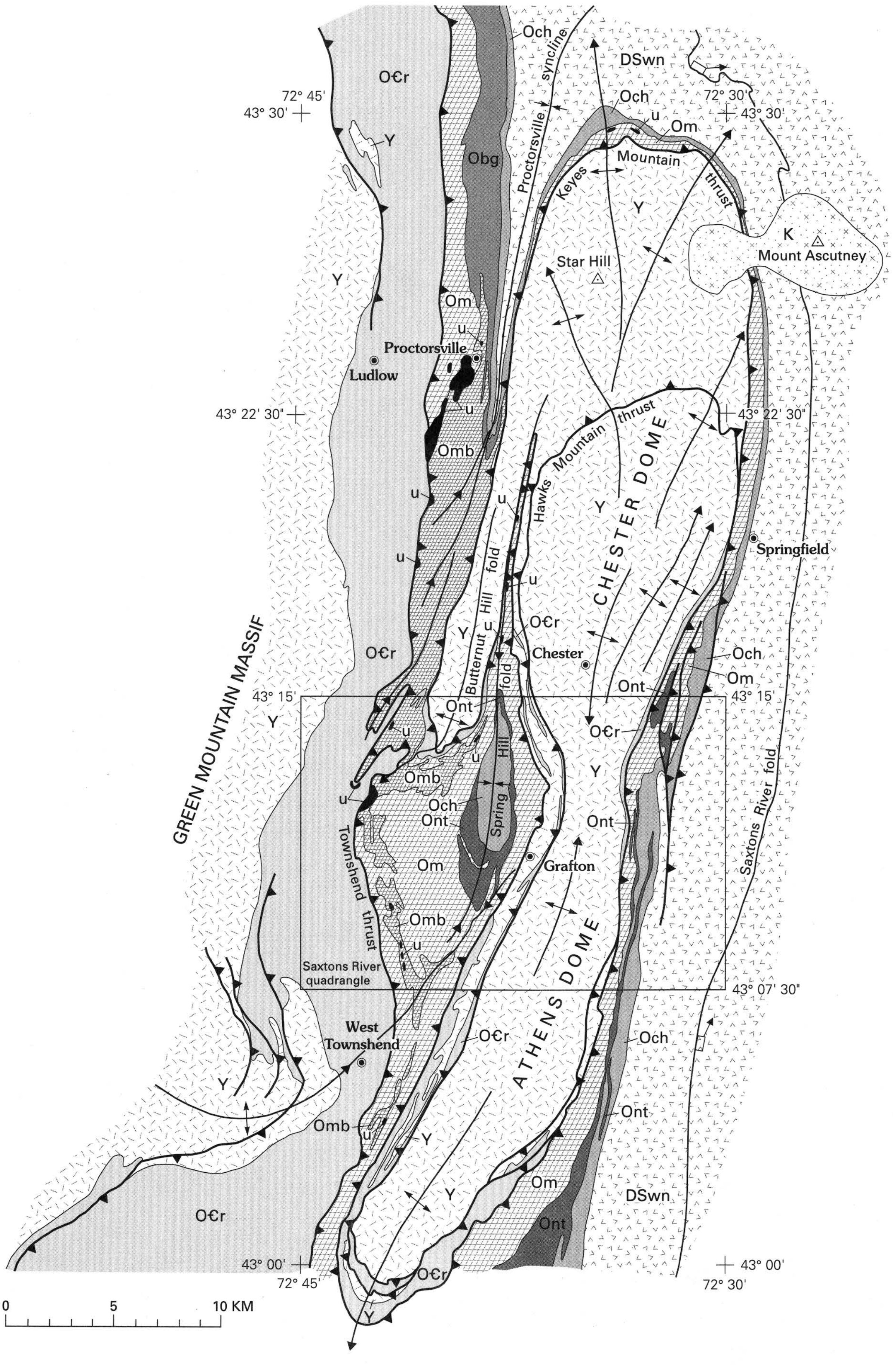
Nappe-stage fold

Dome-stage folds—Showing direction of plunge

Antiform

Synform

Figure 1.—Simplified geologic map of the Chester and Athens domes after Doll and others (1961) and Thompson and others (1990). Shows location and plunge direction of major Acadian folds after Rosenfeld (1954) and Doll and others (1961). *A-A'* is a simplified cross section showing interpretation of geology according to Rosenfeld (1954) and Doll and others (1961, their section *E-E'*).


O€r
72° 45'
43° 30'
Y
Obg
Och
Proctorsville syncline
DSwn
Och
72° 30'
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u
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Keyes Mountain thrust
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K
Mount Ascutney
Star Hill
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Om
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Proctorsville
Ludlow
43° 22' 30"
u
Omb
Hawks Mountain thrust
43° 22' 30"
CHESTER DOME
u
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Springfield
Butternut Hill fold
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GREEN MOUNTAIN MASSIF
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Spring Hill fold
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Townshend thrust
Om
Grafton
Saxtons River fold
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ATHENS DOME
Saxtons River quadrangle
43° 07' 30"
West Townshend
O€r
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43° 00'
72° 45'
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EXPLANATION

K — **Alkaline intrusive rocks (Cretaceous)**

DSwn — **Waits River and Northfield Formations, undivided (Devonian and Silurian)**—Includes volcanogenic members of Waits River Formation (Standing Pond Volcanics of Doll and others, 1961)

Och, Obg, Ont — **Cram Hill Formation (Och), igneous rocks of North River Igneous Suite (Ont), and Barnard Gneiss (Obg) of Richardson (1929) (Ordovician)**

Om, Omb — **Moretown Formation (Ordovician)**—Includes rocks previously mapped as Whetstone Hill member of Doll and others (1961) of Missisquoi Formation. Also includes rusty schist member (Omb)

O€r — **Hoosac, Pinney Hollow, Ottauquechee, and Stowe Formations, and Rowe Schist, undivided (Ordovician to Cambrian)**

Y — **Core gneisses and schists of Green Mountain massif and Chester and Athens domes (Middle Proterozoic)**—Includes units of Mount Holly Complex, Cardinal Brook Intrusive Suite, and Cavendish Formation

u — **Ultramafic rocks (age uncertain)**—Found within units O€u, Om, and Omb

Contact

Thrust fault—Teeth on upper plate

Acadian folds

- Antiform—Showing direction of plunge
- Synform—Showing direction of plunge
- Axial trace of fold—Showing plunge only
- Inclined fold—Barb shows dip direction of axial surface. Arrow shows approximate plunge direction

Figure 2.—Simplified geologic map of the Chester and Athens domes using current terminology and showing plunges of Acadian folds.

southwest-plunging antiformal structure and we find no evidence for inversion along the northern closure of the Spring Hill syncline. Instead, we find abundant evidence that the Spring Hill fold is a late, doubly plunging synform of foliation and bedding. The axial trace of this fold coincides with the southern extension of the Star Hill fold from the Cavendish quadrangle (Ratcliffe, 1995). In addition, we do not correlate rocks of the Spring Hill area with those south of the Butternut Hill fold closure, as was done by Rosenfeld (1954).

Our concept of the connection of significant regional axial traces of folds is shown in figure 2 and the form-line map (sheet 2). A critical point of disagreement between Rosenfeld's (1954) and our work is over the excessively high amplitude of the Chester and Athens domes shown by Rosenfeld. In his interpretation, the crest of the dome is restored to a position 8 km above the present surface (Rosenfeld, 1954, plate 3). Our studies of the folded foliations in the core of the dome do not permit the construction of such high-amplitude structures.

A recently published map of the Chester and Athens domes and surrounding areas by Thompson and others (1993) shows considerable modification of the location of some contacts in the Saxtons River quadrangle from those drawn previously by Rosenfeld (1954) and by Doll and others (1961). Those principal modifications include greater definition in the cover rocks northwest of Grafton and the reassignment of a belt of dark biotite schist in the Round Mountain area from the Hawley or Cram Hill Formation of their previous reports to the Ottauquechee Formation. We do not agree with this assignment and believe that previous correlation of these rocks with the Cram Hill or Moretown Formation as shown on Doll and others (1961) is correct. In addition, Thompson and others (1993) have mapped a loosely defined unit of slabby gneisses of Proterozoic age near the margin of the dome. They show these rocks as forming a belt approximately 0.5 km wide along the southwestern border of the dome at the southern boundary of the Saxtons River quadrangle. In the southern part of the Chester dome, at the northern border of Saxtons River quadrangle, and extending to a point approximately 4 km south, they show a southward-closing belt of Tyson Dolomite (interpreted by them to be Cambrian and (or) Late Proterozoic) which forms the base of their slabby gneiss unit. We map these slabby gneisses as mylonitic varieties of the Mount Holly Complex and the Bull Hill Gneiss and as local occurrences of the Hoosac Formation. Elsewhere, their slabby gneiss-Mount Holly contact appears to coincide with what we map as the Bull Hill-Mount Holly contact (for example, along the eastern margin of the Chester and Athens domes).

MIDDLE PROTEROZOIC ROCKS

MOUNT HOLLY COMPLEX

Gneiss, schist, and metaintrusive rocks of Middle Proterozoic age constitute the Mount Holly Complex which is exposed in the core of the Chester and Athens domes, in the Butternut Hill fold, and in the Green Mountain massif. The units of the Mount Holly Complex in the northwestern corner of the map are poorly exposed but consist of trondhjemitic gneisses (**Yt**), well-layered biotite-quartz-plagioclase gneiss (**Ybg**), and felsic gneiss (**Yfg**). Outside the area to the west, the trondhjemitic gneisses (**Yt**) are continuous with similar rocks that range in age between 1.35 and 1.31 Ga, based on U-Pb zircon data determined by John N. Aleinikoff (*in* Ratcliffe and others, 1991). Throughout the Green Mountain massif, trondhjemitic gneisses in that age range appear to underlie other well-layered gneisses interpreted as metasedimentary and metavolcanic in origin. All of the above-mentioned gneisses were intruded by biotite granite at about 1.25 Ga; subsequently, all Mount Holly Complex rocks were metamorphosed, deformed, and intruded by still younger granites and pegmatite which, in part, postdate the Grenville orogeny (Ratcliffe and others, 1991).

Rocks of the Mount Holly Complex occur in the core of the Chester and Athens domes and are intercalated by faults with cover rocks along the western margin of the dome. These gneisses are typical of the well-layered gneisses and younger granitic gneiss of the Mount Holly Complex in the Green Mountain massif. None of the trondhjemitic gneisses, however, are exposed in this quadrangle, but they are widely present to the north in the Chester and Cavendish quadrangles (Ratcliffe, 1995). In those quadrangles, the trondhjemite gneisses do intrude units of the Mount Holly Complex, such as **Ybg**, **Yrs**, **Yrg**, and **Ya** as mapped here. This trondhjemite gneiss, formally named the Felchville Gneiss in the Cavendish quadrangle (Ratcliffe, in press) has recently been dated by $^{207}Pb/^{206}Pb$ single high-resolution ion-microprobe (SHRIMP) ages for zircon at 1.4 Ga (John N. Aleinikoff, U.S. Geological Survey, written commun., 1995; Ratcliffe and others, 1996). The intrusive relations suggest that parts of the Mount Holly Complex may be 1.4 Ga and older.

Mount Holly Complex in the Chester and Athens domes

The dominant unit in the core of the domes consists of a heterogeneous assemblage of gray, well-layered biotite-quartz-plagioclase gneisses (**Ybg**) that contain beds of quartzite, marble, calc-silicate gneiss and more aluminous schists or gneisses. These enclosed rocks are clearly metasedimentary in origin and are distinguished separately where they can be shown at map scale. The origin of the biotite-quartz-plagioclase gneiss (**Ybg**) unit is uncertain; it may have originated as a heterogeneous assemblage of dacitic metavolcanic rocks or graywackes rich in plagioclase and lacking potassium feldspar. Hornblende-bearing biotite-quartz-plagioclase gneiss and hornblende or hornblende-garnet amphibolites (**Ya**) occur in layers.

Rusty-weathering and more biotitic and muscovitic gneisses (**Yrg**) and schists (**Yrs**) are common. These units are regionally associated with amphibolite (**Ya**), quartzite (**Yq**), and pods or lenses of calc-silicate rock (**Ycs**) or marble (**Ym**). One belt of intercalated schist (**Yrs**), gneiss (**Yrg**), and associated marble and calc-silicate rock (**Ycs**) enters the quadrangle from the north along the western base of Fox Chair Mountain. This belt of rocks is continuous to the north with a belt west of the Bull Hill Gneiss that extends northward to the Black River at the eastern edge of the Chester quadrangle. This belt of schist, gneiss, marble, and calc-silicate gneiss was identified as the base of the cover sequence by Thompson (1950) and subsequently by Thompson and others (1993) as the dolostone member of the Tyson Formation and associated slabby gneisses. They show this as a continuous belt of rocks that bends westward at a point 4 km south of the Saxtons River quadrangle boundary, in the center of the Chester dome, and then extends northwestward for about 2 km. Rocks inside and to the north of this belt were assigned to the Mount Holly Complex and rocks to the south were shown as their slabby gneiss unit, including the Tyson Formation and Bull Hill Gneiss. They regard the Bull Hill Gneiss as a rhyolitic ash-flow deposit and part of the cover sequence (Thompson and others, 1993). The rusty gneiss belt (Yrg) and associated rocks that we map (Yrs and Ycs) do not have the configuration they depict. These units appear to be an integral part of the Mount Holly Complex. Relict gneissosity in the Mount Holly trends north-northeast throughout the southern part of the Chester dome and the map pattern of units in the Mount Holly indicates that the rusty gneiss unit (Yrg) is not coextensive with cover rocks east of the Bull Hill Gneiss.

Mount Holly Complex in the Butternut Hill Fold

A large area of biotite-microcline-plagioclase gneiss and leucocratic gneiss of uncertain correlation extends into the area from the north in the core of the Butternut Hill fold. Exposures are poor and, in the absence of detailed data and subdivisions of the Mount Holly Complex in the Butternut Hill fold in the Andover quadrangle, we cannot at present determine the connection with mapped units in the Cavendish quadrangle to the north. Therefore, this area is shown as Mount Holly Complex, undivided (Ymhu). Some of the more granitic rocks exposed here may belong to the Bull Hill Gneiss, but at present that is conjectural.

Intrusive rocks of the Mount Holly Complex

In the Green Mountain massif, the small area of trondhjemitic gneiss (Yt) in the northwestern corner of the map is regarded as part of a widespread intrusive suite of rocks dated between 1.35 to 1.3 Ga, previously interpreted as older than the bulk of the Mount Holly Complex (Ratcliffe and others, 1991). Similar rocks in the core of the Chester dome in the Chester quadrangle, however, appear to intrude units of the Mount Holly Complex (Ratcliffe, 1995) and indeed all of the paragneisses of the Mount Holly Complex may be older than 1.35 Ga. Several belts of biotite granite gneiss are present in this area and may be intrusive into the other units of the Mount Holly Complex. Alternately, these granitic gneisses (mapped as Ygg) may be felsic volcanic rocks that were transformed in place into migmatitic rock during the Middle Proterozoic metamorphism.

THE BULL HILL GNEISS OF THE CARDINAL BROOK INTRUSIVE SUITE

The Bull Hill Gneiss of Richardson (1929) is named for exposures of coarse-grained microcline augen gneiss on Bull Hill in the Saxtons River quadrangle, 3.5 km northwest of Cambridgeport. On Bull Hill, the rock is a coarse-grained microcline-phenocrystic granite in which the individual microcline crystals may be ovoidal to rectangular and have thin plagioclase rims. This structure commonly is destroyed because of intense mylonitization (see Ratcliffe, 1991, p. 19–25). In the Saxtons River quadrangle, the Bull Hill Gneiss is discordant to units of the Mount Holly Complex. Cliff exposures of an intrusive breccia containing rotated blocks of gneissic Mount Holly can be seen on the slopes above the Saxtons River, 0.8 km west of the crest of Bull Hill. Here, the medium-grained, aplitic, igneous matrix contains small, 1- to 2-cm-long phenocrysts of microcline. Both the gneissic blocks, which may be as much as 10 cm in length, and the igneous matrix are crosscut by a foliation younger than the gneissosity contained in the rotated xenoliths. In the vicinity of Bear Hill, on the west side of the dome, and along the west-facing slopes to the south, very coarse grained, rapikivi-textured, ovoidal microcline forms grains as much as 3 cm in diameter in the Bull Hill Gneiss. These textures are very similar to those present in the much less deformed Somerset Reservoir Granite and Stamford Granite (Ratcliffe, 1991, figs. 3 and 4).

The Bull Hill Gneiss, exposed along the western border of the domes from the northern part of the map southward, is all highly mylonitic. Microcline augen define an orientation and shape fabric with the long axes of the augen plunging down the dip of the foliation to the west, northwest, or southeast, depending upon the dip direction of the foliation.

The Bull Hill Gneiss is confined to the Chester and Athens domes (Ratcliffe, 1991), where it occurs both at the margin and in the core principally in the area south of the Chester quadrangle. Although previously mapped as a continuous unit around the Chester dome (Doll and others, 1961), it actually extends for only a short distance north into the Chester quadrangle. Along the eastern margin it extends as far north as Springfield (Ratcliffe, unpub. data). Along the western margin it extends as far north as Gassetts (perhaps as fault slivers). Previously reported occurrences (Doll and others, 1961) in the Chester and Cavendish quadrangles are not Bull Hill Gneiss but coarse-grained plagioclase-augen gneisses that belong to the Felchville Gneiss (Ratcliffe, 1995).

The Bull Hill Gneiss intrudes the Mount Holly Complex throughout the Athens dome (Ratcliffe, 1991, fig. 6). Two samples of Bull Hill Gneiss from the Athens dome and from the southern part of the Chester dome yielded U-Pb zircon upper-intercept ages of 945 ± 7 and 955 ± 5 Ma, respectively (Karabinos and Aleinikoff, 1990). The latter sample comes from roadcuts in this quadrangle on the south side of Vermont Route 103, 300 m east of the intersection with Sylvan Road. The former sample comes from a point 2 km south of the southern border of the quadrangle from a belt of Bull Hill Gneiss exposed on the western side of Branch Brook and above a belt of the Hoosac Formation (above the Ober Hill fault) that extends into this quadrangle from the south. This belt of Bull Hill Gneiss yielded an age similar to the type Bull Hill, but this belt cannot be traced to the type locality because of faults that separate it from the main part of the Bull Hill in the core of the dome.

The approximately 955 to 945 Ma age for the Bull Hill Gneiss agrees closely with other ages determined for other units of the Cardinal Brook Intrusive Suite (Karabinos and Aleinikoff, 1990). These other units clearly intrude gneissic units of the Mount Holly Complex. Elsewhere, these rocks lack the strong gneissosity present in the Mount Holly Complex. For this reason, the Bull Hill Gneiss is regarded as post-Grenvillian (Ratcliffe and others, 1988, p. 5–9; Ratcliffe, 1991). This observation is important for structural analysis because the foliation present in the Bull Hill Gneiss must be younger than Grenvillian and must, therefore, be Taconian, Acadian, or younger. The comparison between the structure developed in the Mount Holly Complex and that in the Bull Hill Gneiss, therefore, can be a guide to recognizing post-Grenvillian structures in the core of the Chester and Athens dome.

PEGMATITES

Minor pegmatite bodies (Yp), mostly too small to map, are present throughout the Mount Holly Complex but none are found within the Bull Hill Gneiss. The age of these pegmatites is uncertain but they appear to crosscut structures and gneissosity in the Mount Holly Complex. Many of these pegmatites pass into migmatitic gneiss and fine-grained aplitic segregations that cannot be distinguished from integral parts of the Mount Holly units; therefore, we choose to interpret the pegmatites as part of the Mount Holly Complex. Alternatively, we cannot rule out the possibility that all or part of the pegmatites mapped are actually coeval with the Bull Hill Gneiss and largely younger than the Mount Holly Complex.

COVER ROCKS OF THE CHESTER AND ATHENS DOMES AND GREEN MOUNTAIN MASSIF

Rocks which overlie the Mount Holly Complex and Bull Hill Gneiss range in age from Late Proterozoic to Lower Devonian. Stratigraphic correlations and age assignments within these units are quite uncertain. Previous interpretations of Doll and others (1961) and Thompson and others (1993) have shown continuous units and a regular stratigraphic succession. In areas to the south of the Saxtons River quadrangle, detailed mapping by Ratcliffe (1993a, 1997) and Ratcliffe and Armstrong (1999) has shown that the standard Vermont units, Pinney Hollow, Ottauquechee, and Stowe Formations as used by Doll and others (1961), cannot be mapped. We have found that the sequence is heterogeneous and that individual units do not repeat or occur in stratigraphic succession. We have chosen to refer to some of the rocks above the Hoosac Formation and below the Moretown Formation as the Rowe Schist because they are traceable southward into the type locality of the Rowe Schist in Massachusetts (Stanley and Hatch, 1988). There, the Rowe Schist is also an internally complex and heterogeneous unit which Stanley and Hatch (1988) interpreted as a tectonic mélange.

North of the Saxtons River quadrangle and on the west side of the domes, the Pinney Hollow, Ottauquechee and Stowe Formations can be traced continuously along the strike belt to the type localities of the Pinney Hollow and Ottauquechee Formations in the Plymouth quadrangle. At their type localities, the Pinney Hollow and the Ottauquechee, although quite identifiable, are bounded top and bottom by thrust faults and thus their positions as units within a well-defined stratigraphic succession can be questioned (Walsh and Ratcliffe, 1994).

The Stowe and Pinney Hollow Formations both contain chlorite-muscovite±magnetite-quartz schists and amphibolites that are indistinguishable. Where the carbonaceous schists of the Ottauquechee are absent, thin, or occur at several different positions, the distinction between Pinney Hollow and Stowe cannot be made with certainty. In this quadrangle, we have chosen to use the names Pinney Hollow, Ottauquechee, and Stowe only for those rocks that are demonstrably coextensive with those units in the Plymouth area. The name Rowe is applied to rocks between the Stowe and the Pinney Hollow Formation east of the Green Mountains but west of the base of the Moretown Formation, where those distinctions cannot be made. Specifically, we apply the name Rowe Schist to rocks that lie above the Hoosac Formation and below the Moretown in the cover sequence of the Chester and Athens domes. This usage is supported by the lack of clear and continuous units assignable to the Pinney Hollow, which appears to be entirely absent, and by the very sporadic appearance of rocks assignable to the Ottauquechee Formation.

HOOSAC FORMATION

The Hoosac Formation rests unconformably on basement gneisses of the Mount Holly Complex of the Green Mountain massif in the northwestern corner of the map. Conglomerate and an angular unconformity are present at several places along the contact. Rosenfeld (1954, p. 22) made note of the unconformity in well-exposed roadcuts along Vermont Route 11 about 1 km west of the quadrangle border. Elsewhere in the quadrangle, deformation is very intense and clear-cut angularity is difficult to discern.

The four members of the Hoosac Formation in this quadrangle closely follow subdivisions and usages from the Jamaica-Townshend and Mount Snow-Readsboro areas to the south (Ratcliffe, 1993, 1997). To the north, albitic rocks of the Hoosac interfinger with beds of the Tyson Formation which appears to replace much of the Hoosac (Ratcliffe, 1992, 1994). Rocks of the Hoosac Formation resting on the basement rocks of the Chester and Athens domes and in the Butternut Hill fold are, in general, more schistose and less markedly albite-studded than the rocks resting on the Green Mountain massif.

Three discontinuous and partially fault-bounded belts of Hoosac Formation have been mapped along the western border of the Chester and Athens domes. The first and lowermost belt (above the Kenny Pond thrust) is a continuation of a sequence of rocks exposed along the valley of South Branch in the adjacent Townshend quadrangle. There are no exposures of this lowermost belt in this quadrangle from the southern border of this map to the town of Grafton. However, isolated small outcrops of very schistose and feldspathic, slightly rusty schist can be found underlying the steep cliffs of Bull Hill Gneiss west of Vermont Route 35, north of Grafton. Pasture exposures of Hoosac Formation are found on the west side of Vermont Route 35, 0.7 km north of Grafton. The western contact of Hoosac with rocks assigned to biotite-quartz-plagioclase gneiss (**Ybg**) of the Mount Holly Complex lies at the western edge of this pasture. Rocks west of this point are gray, highly foliated muscovite- and biotite-rich feldspathic schists, gneisses, and amphibolites that are very difficult to distinguish from rocks of the Hoosac Formation. The assignment here of some or all of these gneisses to the Mount Holly Complex is quite uncertain. Likewise, schists, which locally contain carbonates, exposed along Vermont Route 35 to the north, although resembling rocks of the Hoosac Formation, may in fact be mylonitic varieties of paragneiss units in the Mount Holly Complex.

The second belt of Hoosac Formation, west of a belt of granitic gneiss, enters the quadrangle from the south where excellent exposures are present in and around Acton. These rocks unconformably overlie the Mount Holly Complex. The third belt occurs north of Grafton and west of Vermont Route 35, where two smaller areas of Hoosac Formation have been mapped, separated by quite distinctive biotitic and granitic gneisses of the Mount Holly Complex. This interleaving could be the result of folding, but intense mylonitization and belts of anastomosing foliation suggest faulting is more likely the cause.

Rocks of the Hoosac Formation are well exposed on the west side of the Butternut Hill fold for a distance of 2.5 km south of the northern border of the quadrangle. The basal beds contain abundant biotite and plagioclase and numerous amphibolite layers interbedded with rusty-weathering, feldspathic, and quite muscovitic schists. The contact with the underlying Mount Holly Complex can be located to within about 10 m. In the interval between the last recognizable gneisses of the Mount Holly and the first beds of well-layered rocks of the Hoosac Formation, there are scattered outcrops of very well foliated rocks not clearly assignable to one specific formation. In this area, dips of bedding and foliation and the plunges of fold hinges indicate that the Hoosac Formation overlies the Mount Holly Complex along a southwest-dipping contact.

Rocks assigned to the Hoosac Formation east of the Chester and Athens domes are quite enigmatic. A belt of rocks typical of the Hoosac Formation that contains quartzites, amphibolites, minor conglomerate, and schist enters the quadrangle from the south and is interbedded with distinctive plagioclase-spotted schists and granofels that is very typical of the Hoosac Formation. These rocks can be traced 3.5 km north from Cambridgeport. North of that point, to the northern border of the quadrangle, the exposures of rocks assignable to the Hoosac Formation are limited to a very narrow belt from 10 to 50 m wide that is in fault contact with the Bull Hill Gneiss to its west. Excellent exposures showing the transition from the Bull Hill Gneiss, through mylonitic slabby gneisses derived from the Bull Hill Gneiss, into nondescript, mylonitic gray gneisses or granofels of uncertain correlation are present in the bed of the Williams River, 0.7 km southeast of the railroad crossing. In these outcrops and in exposures on the slopes to the north, enclaves of Bull Hill Gneiss, in various stages of mylonitization, can be found for a distance of approximately 50 m east of the main outcrops of Bull Hill Gneiss. A similar transition from massive Bull Hill Gneiss into gray, medium-grained to mylonitic Bull Hill Gneiss is exposed in the bed of Saxtons River, 1.5 km northwest of Cambridgeport. In these two areas, part or all of the zone shown as Hoosac Formation on the map contains highly mylonitic rocks that may be in part derived from the Bull Hill Gneiss or other gneisses and in part derived from rocks of the Hoosac Formation or other cover-sequence rocks. These rocks are correlated with the Hoosac Formation but are labeled ЄZh? on the map.

Regardless of the correlation of these rocks with the Hoosac Formation, this belt of rocks is unusually thin and contains few of the variety of rock types normally present in the Hoosac elsewhere. Therefore, the extreme thinness cannot be explained solely by tectonic thinning (of a complete section) but must be explained by original stratigraphic thinning of the section or by faulting (the explanation which we propose).

PINNEY HOLLOW FORMATION

Lustrous schists of the Pinney Hollow Formation (ЄZph) crop out in a broad area extending from Glebe Mountain in the Londonderry quadrangle to the west almost to the village of Windham, a distance of approximately 4 km. Throughout this belt, the rocks are quite uniformly green, chloritic-muscovitic quartz±magnetite schists that are sparingly garnetiferous and rarely albitic, although some feldspathic rocks do occur. Amphibolites are absent, except for a narrow belt near the base of the Pinney Hollow on Glebe Mountain in the Londonderry quadrangle. Near the southwestern corner of the map, the Pinney Hollow thins to less than 1 km in width as it does at the northern boundary of the map. The base of the Pinney Hollow above the Hoosac Formation is interpreted as a fault. Internal units in the Pinney Hollow were not delineated because of the monotonous nature of the rocks. Rare quartzite beds as much as 2 m thick could not be mapped as continuous units.

Regionally, the base of the Pinney Hollow rests on different rocks. North of Ludlow in the Plymouth quadrangle, it is in contact with rocks of the Plymouth Formation, which pinches out southward, as does the underlying Tyson Formation (Ratcliffe, 1992; Walsh and Ratcliffe, 1994; Ratcliffe, 1997). From Ludlow southward, the contact at the base of the Pinney Hollow is marked by a concentration of closely spaced mylonitic S_2 fabric that plicates and shreds an older S_1 schistosity. The intensity of this structure at or near this contact suggests that the Pinney Hollow contact with underlying rocks is a thrust fault. South of this quadrangle (Ratcliffe, 1997), rocks coextensive with the Pinney Hollow become involved in a series of very complex structural repeats and thrust faults (the South Wardsboro thrust, for example) that make detailed connections and correlations impossible. These rocks, which lithically resemble Pinney Hollow-like rocks of the Pinney Hollow Formation, have been mapped either as the Rowe Schist or Rowe(?) Schist (Ratcliffe and Armstrong, 1999; Ratcliffe and others, 1992) in the areas south of the Saxtons River quadrangle.

The upper contact of the Pinney Hollow is also interpreted as a thrust fault, here named the Windham thrust. In the southern part of the map, a series of highly sheared outcrops associated with carbonaceous, feldspathic schists marks the contact. East of this fault, a heterogeneous group of gray, feldspathic schist, amphibolite, and large-garnet schist occurs. To the north, a series of thick amphibolites occur that are interlayered with green, chloritic schists indistinguishable from the Pinney Hollow. The heterogeneous rocks immediately above the Pinney Hollow are here assigned to the Rowe Schist.

OTTAUQUECHEE FORMATION

Near the northern border of the map, and for a distance of about 8 km south, a continuous, narrow belt of dark-gray to black, carbonaceous schist and minor quartzite occurs. Intermittent outcrops that appear to be part of this same belt continue south of Windham. This belt of rock is coextensive with rocks of the type Ottauquechee Formation in the Plymouth quadrangle to the north. There, the formation crops out in a 2.5-km-wide belt that consists predominantly of carbonaceous schists similar to the Ottauquechee of this map area. Rocks typical of the Ottauquechee, therefore, thin from a 2.5-km-wide belt in the type area to a several-hundred-meter-wide belt in this quadrangle (Walsh and Ratcliffe, 1994). From this point southward to the Massachusetts State line, rocks typical of the Ottauquechee are largely absent, although rare lenses of carbonaceous schist occur throughout the Rowe. Some of the black schists and quartzites most closely resembling the Ottauquechee Formation at its type locality are found at or near the contact between the Rowe Schist and the Moretown Formation in the Jamaica and Townshend quadrangles (Ratcliffe, 1997; unpub. data, 1995). The name Ottauquechee is not applied south of this quadrangle because it disappears as a traceable unit and similar rocks are interleaved either structurally or stratigraphically throughout the Rowe Schist, rather than in a single stratigraphic position (Ratcliffe, 1997). Isolated occurrences of Ottauquechee-like rocks within the Rowe

Schist, but near the proper position, are identified as Ottauquechee(?) Formation (€o?). From South Windham northward, the belt of Ottauquechee appears to be fault bounded. At one locality approximately 1 km north of Lawrence Four Corners, an isolated exposure of quartzite appears as a horse between the bounding faults.

ROWE SCHIST AND STOWE FORMATION

As mapped here, the Rowe Schist occurs as a narrow, heterogeneous band of rock, including large-garnet schist, amphibolite, feldspathic schist, and granofels that are all associated with fine-grained chloritic-muscovitic schist. We use the name Rowe Schist generally to refer to this heterogeneous group of rocks that lie above the Pinney Hollow (or Hoosac) Formation, but that underlie the Moretown Formation. In the western part of the map, the Rowe Schist is separated from the Pinney Hollow Formation by the Windham thrust. A nearly continuous belt of chlorite-muscovite±biotite schist lies east of the Rowe Schist here but below the Moretown Formation in the western part of the map. This belt of rocks is coextensive with rocks mapped as the Stowe Formation to the north. In the Plymouth quadrangle, the Stowe Formation overlies the Ottauquechee along a prominent and well-documented fault (Walsh and Ratcliffe, 1994). To the south, the rocks of the Stowe Formation, as mapped here, were shown as correlative with the upper part of the Rowe Schist (Ratcliffe, 1997).

The cover rocks on the Chester and Athens domes immediately above the Hoosac Formation are herein assigned to the Rowe Schist. A fairly persistent sequence of amphibolite, plagioclase-rich granofels, large-garnet schist and green, phyllonitic, feldspathic schist are present along the western margin of the Chester and Athens domes north and south of Grafton. Along the eastern margin of the domes, very limited occurrences of schist assignable to the Rowe Schist are found between the Moretown Formation and the basement gneisses of the dome.

Both east and west of the domes, previous workers (Doll and others, 1961; Thompson and others, 1993) have mapped parallel belts of the Hoosac, Pinney Hollow, Ottauquechee, and Stowe Formations. We do not agree with their mapping and find that rocks assignable to the Pinney Hollow and Ottauquechee are absent and that the Rowe Schist is a very thin belt. The thinness of the Rowe Schist (or equivalent rocks of previous workers) has in the past been attributed to tectonic thinning of a coherent section. Comparison of the map distribution and apparent thicknesses of units within the Rowe Schist west of the domes, as compared to similar units further west, suggest that rocks on the flanks of the domes are not markedly attenuated compared to similar rock units exposed elsewhere.

MORETOWN FORMATION

The Moretown Formation as mapped here consists of three major units and many minor ones. We interpret the base of the Moretown as a regionally important thrust fault (Townshend thrust), based on the observation that the internal stratigraphy of the Moretown is discordant to this contact, and that in the western part of the map the contact is marked by numerous large to small bodies of ultramafic rock. The best exposures of Moretown occur in the area west of Spring Hill and south of the Butternut Hill fold. Pinstriped granofels (Oml), the most distinctive unit of the Moretown, appears to be transitional into garnet schist and granofels (Omgs). The distinction between these two is based largely on the greater abundance of chlorite, muscovite, and garnet in the schist and granofels and the decreased abundance of typical pinstriped granofels. Biotite schist (Ombs) forms a prominent unit that is spatially associated with small to large masses of serpentinite and talc (OZu). The dark-gray, sooty-weathering, and locally sulfidic schist contains beds of quartzite, rusty sulfidic amphibolites, magnetite-rich rocks, and coticule. A nearly continuous belt of biotite schist (Ombs) enters the quadrangle from the north and extends southward to the area of the Townshend Reservoir in the Townshend quadrangle. These rocks are interbedded with pinstriped granofels (Oml) and with garnet schist and granofels (Omgs). In fact, the unit appears to laterally replace the greenish schists of Omgs.

A belt of biotite schist (Ombs) extends northeast from Round Mountain 2.5 km to a small talc quarry. For 10 km northeast of the quadrangle, biotitic schist and associated talc rocks occur as discontinuous masses within the Moretown Formation. A narrow belt of similar rocks occurs interlayered with garnet schist and granofels (Omgs) along Saxtons River, 2.5 km northwest of Grafton. All of these occurrences of the biotite schist unit are within rocks typical of the Moretown Formation and do not appear to be in the stratigraphic position of the Ottauquechee Formation. Rosenfeld (1954) originally mapped this belt as the Hawley Formation, whereas Doll and others (1961) mapped them as the Cram Hill Formation. Recently, Thompson and others (1993) assigned these rocks to the Ottauquechee. We do not accept that correlation for the reasons stated above and regard the biotite-schist (Ombs) as an integral part of the Moretown Formation.

ULTRAMAFIC ROCKS

There are eight localities where ultramafic rocks, either serpentinite or talc and talc-carbonate, occur within the quadrangle. All of these are found within or in close association with rocks of the Moretown Formation. Many of these were quarried for talc. Small prospect pits and larger quarries are abundant; all are currently inactive except for the large Hamm quarry on the southwest side of Mack Road, 3 km southeast of North Windham. The principal quarries are east of Windham, east of South Windham, and on the ridge northeast of Round Mountain. These ultramafic rocks form part of the major ultramafic belt that begins near Proctorsville in the Ludlow quadrangle and extends southward through the Andover quadrangle to the central part of the Townshend quadrangle (fig. 1).

All of the ultramafic bodies occur within the same structural setting and are enclosed in the rocks of the Moretown Formation, or they occur near the border of the Moretown within adjacent rocks. At one locality near Hitchcock Road and Mack Road, the talc quarries appear to be within greenish phyllites of the Stowe Formation.

The ultramafic rocks typically have strongly sheared contacts with adjacent rocks and are bordered by selvages of talc-chlorite-carbonate rock. Because of this deformation, the original contact relations are not preserved. At many localities, at one or both borders, the country rocks are dark, sulfidic, siliceous schists (Ombs) that contain thin (1 to 2 cm), well-layered beds of light-gray, medium-grained biotite-plagioclase-quartz granofels and quartzite. These thinly bedded quartzofeldspathic rocks resemble finer grained varieties of the more feldspathic pinstriped granofels (Oml). Also occurring as thin beds are rusty-weathering layers or laminae of hornblende-chlorite-plagioclase amphibolite or ankeritic amphibolite in layers 1 to 2 m thick. This association of rocks is very common at or near the borders of the ultramafic rocks, even in the smallest occurrences.

The large exposures of ultramafic rocks are commonly interpreted as fault slivers of peridotitic or harzburgitic mantle rocks derived as tectonic pieces of larger ophiolite sheets (Stanley and Ratcliffe, 1985). Alternatively, the ultramafic rocks may be submarine slide blocks and associated turbiditic debris flows shed from submarine fault scarps or from advancing ophiolite thrust sheets. The association of siliciclastic sands, sulfidic pelites, and ankeritic greenstones with bedded talc deposits suggests a mixed source of ultramafic depleted mantle as well as intermediate to mafic volcanic rocks.

CRAM HILL FORMATION

On this map, we have applied the name Cram Hill Formation to a complex collection of intercalated and laterally discontinuous rocks that overlie the Moretown Formation and that are intruded by the North River Igneous Suite (Armstrong, 1994; Ratcliffe and Armstrong, 1999). The felsic and mixed felsic and mafic varieties of igneous rocks have previously been referred to as the Barnard Gneiss of Richardson (1929) or the Barnard Volcanic Member of the Missisquoi Formation of Doll and others (1961). The remaining metasedimentary rocks have been referred to as the Cram Hill Member of the Missisquoi Formation (Doll and others, 1961). Thompson and others (1993) have applied the name Barnard Gneiss to what they regard as largely volcanic rocks and have used Hawley Formation (Emerson, 1917; Stanley and Hatch, 1988) to refer to a collection of metasedimentary rocks and greenstones above their Barnard. Stanley and Hatch (1988) applied the name Hawley to rocks in Massachusetts which are predominantly felsic and mafic intrusive and extrusive rocks and minor amounts of black, splintery, sulfidic schist (known as the schist member) that is mapped in the upper third of the formation. Armstrong (1994) has shown that the Barnard Volcanic Member of the Missisquoi Formation (of Doll and others, 1961) in southernmost Vermont contains metaigneous and metasedimentary rocks that are coextensive with the entire Hawley belt in Massachusetts. Both belts consist almost entirely of amphibolite, metatrondhjemite, metatonalite, and minor intermediate to felsic metavolcanic rocks that are interlayered with sulfidic black schist. Thus, the Hawley Formation of Massachusetts is largely equivalent to rocks mapped by others as the Barnard Volcanic Member of the Missisquoi Formation (Doll and others, 1961) and the names Hawley and Barnard are equivalent; the use of both terms, the Hawley Formation and the Barnard Gneiss, for rocks in Vermont is inappropriate.

In order to resolve this problem, we use the name Cram Hill Formation to refer to the metavolcanic and metasedimentary rocks in Vermont. We equate the Cram Hill with the metavolcanic and metasedimentary parts of the Hawley Formation in Massachusetts, but we exclude the intrusive rocks, which in Vermont are referred to as components of the North River Igneous Suite (Armstrong, 1994).

The name Cram Hill (Currier and Jahns, 1941) originally was applied to greenish-gray to black phyllite that contains minor beds of siliceous volcanic breccias, soda rhyolite, and basalt or basaltic andesite. A basal quartzite mapped as the Harlow Bridge Quartzite Member of the Missisquoi Formation provided a useful marker to distinguish greenish phyllites of the Moretown (below) from identical rocks of the Cram Hill (above) (Currier and Jahns, 1941, p. 1487–1512). The Harlow Bridge Quartzite Member at its type locality is a pinstriped biotite-chlorite-muscovite-plagioclase-quartz quartzite that is identical to the pinstriped granofels member (Oml) of the Moretown Formation of southern Vermont and western Massachusetts. Our use of the Cram Hill, therefore, to refer only to largely dark, carbonaceous schist and included metavolcanic rocks is more restrictive than that of Currier and Jahns (1941). Our usage establishes a one-to-one correlation with identical rocks of the Hawley Formation but leaves open the question of detailed correlation with the type Cram Hill.

Two somewhat similar sequences of Cram Hill Formation, as we map it, occur in the Saxtons River quadrangle. In the correlation of map units and in the description of map units, these belts are treated separately. One column shows relations for Spring Hill; the other is for the more extensively exposed rocks near Rockingham and Cambridgeport.

On Spring Hill, the basal biotite granofels and schist member (Ochfg) rests on rocks of the Moretown Formation, which is in contact with intrusive rocks (Ontg, Ontr) of the North River Igneous Suite. At the north end of the Spring Hill syncline, well-layered felsic and mafic gneisses (Ochv) and amphibolites (Ocha) underlie Ochfg. These rocks, although somewhat resembling rocks of the underlying North River Igneous Suite, are well bedded and contain thin layers of metasedimentary rocks, including quartzite and hornblende feldspathic granofels. Both the Ochv and Ochfg members are interpreted to disconformably overlie the Moretown Formation and rocks of the North River Igneous Suite. U-Pb zircon ages from volcanic and intrusive rocks of the North River Igneous Suite in the Moretown and Cram Hill Formations east of the Chester and Athens domes range from 486±3 Ma to 462±6 Ma (Ratcliffe, Walsh, and Aleinikoff, 1997) suggesting Ordovician ages for these units (see discussion below).

NORTHFIELD FORMATION

Small, garnet-bearing (3 to 6 mm), carbonaceous phyllite (DSn) within the eastern part of the Saxtons River quadrangle lies structurally above rocks within the Cram Hill Formation and intrusive rocks of the North River Igneous Suite. Bedding is defined by homogeneous gray to black carbonaceous phyllite and thin (millimeter to centimeter scale), sandy, sulfidic layers. Well-preserved, graded beds consistently show that beds young to the east. Near the contact with the Northfield Formation, several small, 1- to 2-m-thick, discontinuous lenses of white, vitreous quartzite and quartz-pebble conglomerate are present within the Northfield and are identical to those found within the Cram Hill Formation. All of these discontinuous quartzites and quartz-pebble conglomerates are interpreted as submarine channel deposits. Near the contact with the Cram Hill, the Northfield contains rusty-weathering, feldspathic grit (DSng) which may reflect recycled sediments deposited upon an eroded surface marking a disconformity.

The contact between the Northfield and underlying rocks is sharp. The grit unit (DSng) is in contact with Cram Hill granofels (Ochfg), volcaniclastic rocks (Ochvc), or amphibolite (Ochg) and contains beds of coticule (Ochc). Many of the units of the Cram Hill appear to be truncated along the Northfield contact. However, the gross stratigraphy of the Cram Hill suggests that both the Northfield and the Cram Hill Formations were deposited in a shallow marine environment. The continuous nature of the Northfield grit, its gradational contact relationship with the Northfield carbonaceous schist, and the gradational nature of the contact between the Northfield and Waits River Formation further suggest that the Northfield's its lower contact is a disconformity.

WAITS RIVER FORMATION

The Waits River Formation was originally named the Waits River Limestone in the East Barre quadrangle (Richardson, 1906). Doll (1943) revised the Waits River Formation to the Waits River Group, which included (from youngest to oldest) the Westmore, Barton River, and Ayers Cliff Formations. Both Richardson and Doll believed the Waits River to be Ordovician, based upon equivocally identified fossils and tenuous correlation of Waits River rocks with fossiliferous rocks in southern Quebec (Doll, 1943). The age of the Waits River was changed to Silurian following identification of Silurian crinoids within the Irasburg conglomerate, an informal intraformational unit within the Waits River (Doll, 1951). The Waits River was reduced to formation status by Doll and others (1961) on the Centennial Geologic Map, and included the pelite, limestone, and interbedded volcanic rocks mapped by Doll (1944) as the Standing Pond Volcanic Member.

The Waits River Formation, as used in this report and described in areas outside of the East Barre quadrangle, consists primarily of black to gray, carbonaceous, sulfide-bearing schist, slate, or phyllite with subordinate beds of brown-weathering, punky carbonate or calcareous quartzite. The type section at East Barre contains disproportionately abundant metalimestone and subordinate carbonaceous pelite and, therefore, appears to be atypical of the majority of the Waits River mapped throughout the Connecticut River Valley from northern to southern Vermont. The Waits River includes several metasedimentary and metavolcanic units and at least one metaintrusive unit. The predominant rock type (DSw) is dark-gray to black, carbonate-bearing, carbonaceous garnet schist interlayered with 1-cm- to 30-m-thick beds of discontinuous gray to bluish-gray, siliceous marble. The distinctive brown-weathering crust on these marbles led previous workers to call it "the punky brown." The marble beds are commonly gradational with calcite-bearing quartzite horizons (DSwq) that have a distinctive tan- to brown-weathering, centimeter-scale sedimentary lamination (bedding). Some quartzite beds are white and contain abundant millimeter-scale, rounded detrital grains in a clast-supported framework. Typically, these beds have sharp, planar contacts with the surrounding carbonaceous schist.

The marble beds appear to be distributed rather uniformly throughout the Waits River Formation, but increase in abundance and thickness eastward from the Northfield Formation. The contact with the Northfield is noticeably gradational. Many small (<1-m-thick) beds of the marble are present within the Northfield, but are never as abundant or as large as those within the Waits River. Although the carbonaceous schist of the Northfield is very similar to that within the Waits River, several criteria have been developed in order to delineate a consistent contact between the two formations. First, the Waits River tends to have a more silvery sheen on schistosity surfaces than the Northfield. Second, garnets within the Waits River are typically larger, less abundant, and more widely distributed than those in the Northfield. Third, marble beds in the Northfield are typically less than 1 m thick whereas marble beds in the Waits River are typically 1 to 10 m thick with some approaching 30 m in thickness.

Several metavolcanic and metavolcaniclastic units are present within the Waits River Formation, along with one possible intrusive metaigneous rock type. These units were undivided on previous maps, but were mapped together as the Cobble Mountain Member of the Waits River Formation by Rosenfeld (1954); they were subsequently renamed the Standing Pond Volcanics (Doll and others, 1961) and correlated as a continuous mappable unit with mafic volcanic rocks mapped further north at Standing Pond (Doll, 1944). In the Saxtons River quadrangle, these units include (from top to bottom) felsic volcaniclastic rocks (DSwf), large garnet-hornblende schist (or garbenscheifer) (DSwg), rusty felsic volcanic rocks (DSwc), mafic volcaniclastic rocks (DSwv), megacrystic felsic volcanic rocks (DSwvf), and fine-grained amphibolite rocks (DSwa). The felsic volcaniclastic rocks (DSwf) are actually separated from the rest of the volcanic units by a 200- to 400-m-thick belt of Waits River carbonaceous schist and appear to represent younger volcaniclastic deposition. The garbenscheifer unit (DSwg) occurs at both the upper and lower boundaries of the main volcanic to volcaniclastic sequence and is demonstrably interlayered with both carbonaceous schist and volcanic to volcaniclastic rocks (including DSwvf, DSwv, and DSwc). The garbenscheifer unit may represent the admixture of volcanic and pelitic sediments during both the onset and cessation of local volcanic activity. The fine-grained mafic amphibolite unit (DSwa) and the mafic volcaniclastic rocks (DSwv) appear to be involved in a large facies change on the west flank of the synformal structure that these units define (the "canoe structure" or Standing Pond fold of Thompson and others, 1993; fig. 1). The amphibolite unit grades southward into the volcaniclastic unit and the amphibolite unit is entirely absent near the southern closure of the Saxtons River fold. Along the entire eastern side of the fold, the DSwv mafic volcaniclastic unit is associated with only minor amounts of rusty felsic volcanic rocks (DSwc). The discontinuous, gradational, and interbedded nature of all of the volcaniclastic units suggests that they were deposited as either water-lain volcanic sediments or as volcaniclastic detritus eroded from at least several volcanic (or intrusive) sources.

Coarse-grained, homogeneous hornblende-plagioclase mafic gneiss (DSmg) that is of dioritic composition lies entirely within the amphibolite unit (DSwa). Despite the sharp contacts that parallel compositional layering within the amphibolite, it may be of intrusive origin. The age of the Waits River Formation is Silurian and Devonian based on a Silurian (423±4 Ma) U-Pb zircon age from a volcanic layer on the east limb of the Saxtons River fold (Aleinikoff and Karabinos, 1990; Armstrong and others, 1997).

POST-PROTEROZOIC INTRUSIVE ROCKS

INTRUSIVE ROCKS OF THE NORTH RIVER IGNEOUS SUITE

Many large to small bodies of medium- to coarse-grained biotite-plagioclase-quartz and biotite-absent muscovite-plagioclase-quartz gneiss occur within rocks of the Moretown and Cram Hill Formations on the east side of the Chester and Athens domes and within similar rocks in the Spring Hill syncline. Although contact relationships at most exposures within these rock types are not clear, several key exposures show that the gneisses crosscut either bedding or pre-existing layering within the Moretown or Cram Hill units. Based on these relationships, the felsic gneisses are interpreted, at least in part, to be intrusive metaigneous rocks of tonalitic and trondhjemitic composition. Although all of the mapped felsic gneiss bodies are interpreted as intrusive rocks, it is entirely possible that some, or parts of some, may actually be extrusive in origin. The largest of these bodies (informally called tonalite at Ruger Hill) is located along the enveloping southern part of the Spring Hill syncline and best exposed on a small hill, here referred to as Ruger Hill (a local name), approximately 1.5 km southwest of Spring Hill. Here, numerous coarse-grained tonalitic sills and dikes occur within fine-grained, ankerite-bearing amphibolite and dull-gray, well-bedded felsic volcanic rocks or granofels. The tonalite sills and dikes are in turn intruded by several 1- to 2-m-thick, plagioclase-phenocryst-bearing mafic dikes that are identical to the Whitneyville dikes mapped to the south by Armstrong (1994) within the Townshend, Newfane, and Brattleboro quadrangles of southern Vermont. In addition, many of the ankeritic amphibolites appear to intrude a series of older hornblende-bearing felsic rocks of tonalitic composition. These older rocks may be either intrusive or extrusive, but are easily distinguished from the younger tonalites, which contain biotite but no hornblende. Except for the young mafic dikes, all of these rocks are intruded by a large, white, coarse-grained, 0.5- to 1.5-km-thick, felsic intrusive rock of trondhjemitic composition, that may be equivalent to the West Halifax Trondhjemite of the North River Igneous Suite, mapped by Armstrong (1994) in the Brattleboro quadrangle and by Ratcliffe and Armstrong (1999) in the Jacksonville quadrangle. Although its true thickness appears to be much less than the apparent thickness as shown on the map (due to shallow, north-plunging, late folds), the trondhjemite and tonalite thin dramatically along the western side of the Spring Hill syncline. Numerous dikes and sills of trondhjemite too small to map are found on both sides of Spring Hill within the Moretown Formation, although several small bodies also occur within the Cram Hill Formation on Spring Hill.

On the eastern side of the Chester and Athens domes, several 1- to 2-km-long and 100- to 400-m-wide sills of tonalite and trondhjemite have been mapped within the Moretown. Contacts between the intrusive and host rocks are typically sharp and frequently contain elongate, strained screens of Moretown granofels or garnet schist. Several small bodies of trondhjemite have been mapped within Cram Hill units Ochg and Ochc within the McMaster Mountain area.

Whether or not the felsic intrusive rocks of the North River Igneous Suite postdate Taconian S_2 foliation fabric in this area is unclear. In most areas, contacts between the intrusive rocks and Moretown host rocks are concordant with fabric interpreted to be S_2 foliation, and form line analysis of S_2 foliation does not seem to indicate any structural disharmony between the intrusive rocks and the Moretown host rocks. However, trondhjemite along the east limb of the Spring Hill syncline, due west of the William Putnam State Forest, crosscuts schistosity in the Moretown Formation. To the north in the Plymouth area, however, similar biotite-bearing tonalities and related trondhjemites have been described as postdating S_2 foliation (Walsh and Ratcliffe, 1994). To the south, in the type locality for the North River Igneous Suite, Armstrong (1994) described similar intrusive relationships with felsic rocks crosscutting Moretown-like lithologies without Taconian fabric. If all of these intrusive rocks are related and are coeval, an explanation must be presented that accounts for the different crosscutting relationships and variable development of Taconian fabric in these areas. At present, we do not have enough information to properly address this problem.

Age of the North River Igneous Suite and related rocks

Three U-Pb zircon ages that range from 486±3 to 462±6 Ma have been determined for two intrusive rocks within the Moretown and Cram Hill Formations and for one metavolcanic rock in the Cram Hill (Ratcliffe, Walsh, and Aleinikoff, 1997). These Early to Middle Ordovician ages establish the minimum age of the Moretown Formation and the actual age the Cram Hill Formation as mapped on the eastern side of the Chester and Athens domes. The three dated rocks occur within this quadrangle. A biotite-hornblende metatonalite (Ont) 1.6 km northwest of Bartonsville yielded a conventional upper-intercept age of 486±3 Ma. The trondhjemite gneiss unit (Ont) collected at South Newfane in the Newfane quadrangle yielded an ion-microprobe $^{206}Pb/^{238}U$ age of 462±6 Ma. Trondhjemite from the dated locality extends into this quadrangle above the Atcherson Hollow fault, where it intrudes rocks of the Moretown and Cram Hill Formations. A third sample collected from the top of well-layered, felsic volcanic rocks in the Cram Hill Formation (Ochv), 2 km north of the quadrangle border in the Springfield quadrangle, yielded an ion-microprobe $^{206}Pb/^{238}U$ age of 484±4 Ma. This volcanic unit extends from the sample locality into the Saxtons River quadrangle.

A fourth zircon sample from metatrondhjemite of the Barnard Gneiss (of Richardson, 1924) near Proctorsville yields a $^{206}Pb/^{238}U$ ion-microprobe age of 496±8 Ma. The metatrondhjemite at that locality (Ratcliffe, Walsh, and Aleinikoff, 1997) intrudes the Moretown Formation, which also includes ultramafic blocks of the Proctorsville ultramafic belt.

The U-Pb zircon ages above suggest that intrusion of the tonalitic and trondhjemitic rocks of the North River Igneous Suite, the Barnard Gneiss (of Richardson, 1924) and extrusion of volcanics of the Cram Hill Formation indicate active tectonism, probably fore-arc accretion and intrusion of arc-related igneous rocks in the Early to Middle Ordovician. The mafic and felsic volcanic and intrusive rocks found in the Moretown and Cram Hill Formation have geochemical characteristics of destructive plate margins. For a discussion of the chemistry of the Hawley Formation and intrusive rocks of Massachusetts, see Kim and

Jacobi (1996); see Ratcliffe and others (1997) for similar rocks in Vermont.

DEVONIAN GRANITE AND PEGMATITE

Sills and dikes of biotite-muscovite granite 0.5 to 2 m thick and larger irregular masses of pegmatite intrude rocks in the core and western flank of the Chester and Athens domes. An area of abundant, apparently structurally controlled dikes occurs in a 2-km-wide belt that extends from just east of Spring Hill, north of Grafton, to Fox Chair Mountain at the northern border of the map. The smaller granite dikes are shown only as symbols on the map. Isolated concentration of dikes, too small to map, are also found along the western margin of the domes near the southern border of the map, east of Grafton, and on Bull Hill on the eastern flank of the Athens dome.

Dikes are commonly straight-walled, weakly foliated to nonfoliated, and strongly discordant to the intense folding and schistosity in the country rocks. Where foliated, the foliation trends north-northeast and dips moderately to the northwest, for example, northwest of Grafton. In this area, the northeast-trending, northwest-dipping dikes tend to follow the axial surfaces of the late northeast-trending F_4 folds, but also are deformed into folds having wavelength to amplitude ratios of approximately 5 to 1. On the other hand, folds truncated by the dikes are highly isoclinal and have a strong penetrative axial surface foliation not present in the dikes.

Dikes similar to those found here are widely distributed in the core of the Chester dome to the north, where they tend to be even less deformed than they are here. Approximately 62 dikes have been mapped in the Chester quadrangle (Ratcliffe, in press). They occur in three sets, each trending either northeast, northwest, or nearly east-west. Dips mostly ranging from 45° to 60° are commonly opposed in single outcrops. Vertical dikes are of lesser abundance. Overall, the dikes are nondeformed except near the eastern margin of the Chester quadrangle near the eastern flank of the Chester dome, and there, shortening of dikes in the northeast-trending late fold set is pronounced; folding there postdated dike intrusion.

To the south in the Townshend quadrangle, a series of prominent, subvertical biotite-muscovite granite dikes as much as 4 km long occurs in cover rocks and in core gneisses along the western margin of the Athens dome. There also, the dikes are late to posttectonic and are nonfoliated, or very weakly foliated.

Based on the areal distribution of the granite dikes and the degree of deformation, it appears that post-intrusive deformation is most intense in zones of the late northeast-trending cross folds, regardless of location of the dikes within the dome. This suggests in part, a temporal relationship between intrusion of the dikes and the late northeast-trending cross folds. The absolute age of these granitic intrusions is unclear. However, a U-Pb zircon upper intercept age of 374±4 Ma (John N. Aleinikoff, U.S. Geological Survey, written commun., 1992; Ratcliffe, Armstrong, and Aleinikoff, 1997) for syntectonic Black Mountain Granite in the core of the Guilford dome (Hepburn and others, 1984) suggests that they are Middle to Late Devonian.

CRETACEOUS DIKES

One felsic Cretaceous dike (Kfd) was found in the northern part of the map associated with a marked north-south-trending valley and a concentrated zone of brittle fracturing. The dike, approximately 2 m thick, is exposed south of Horseshoe Farm Road, just north of the Windsor County line. Similar felsic dikes are present in and around the plutons at Mount Ascutney in the Cavendish and Mount Ascutney quadrangles. Regionally, mafic Cretaceous dikes are far more abundant than felsic dikes. The dikes were intruded after all metamorphism had occurred; they are uniformly nonfoliated, strongly crosscut country rock, and commonly exhibit effects of rapid cooling. Spherulitic felsic dikes are relatively common in and around Mount Ascutney (Balk and Krieger, 1936). The spherulitic structure suggests devitrification of originally glassy, quickly cooled rocks. Based on Rb/Sr whole-rock isochron, K-Ar, and $^{40}Ar/^{39}Ar$ biotite ages (122±1.5 Ma), the Mount Ascutney intrusive rocks are Cretaceous (Foland and others, 1985; Hubacher and Foland, 1991).

STRUCTURAL GEOLOGY

INTRODUCTION

The Saxtons River quadrangle contains major geologic structures (see figure 1 and previous discussion) that have figured prominently in interpretations of the models for the tectonic evolution of this part of the New England Appalachians. Previous interpretations, while all recognizing the Chester and Athens domes as antiformal, have depicted the areas to the west of the domes as having been recumbently folded by early Acadian deformation (nappe-stage folds of Thompson, 1950; Rosenfeld, 1968). This inversion involved core rocks of the Butternut Hill fold as well as rocks of the cover sequence west and south of the Butternut Hill fold and rocks of the Spring Hill area. The early nappe fold, in Rosenfeld's (1954, 1968) view, is the Proctorsville syncline (fig. 1) with the upper or inverted limb containing Middle Proterozoic rocks in the core of the late Butternut Hill fold. The geology of the Saxtons River quadrangle is critical for evaluating these ideas.

The deformed bedrock ranges in age from Middle Proterozoic to Devonian. The pre-Silurian rocks have been affected by both Taconian (Ordovician) and Acadian (Devonian) deformation and therefore are highly complex polymetamorphic rocks. The Silurian and younger rocks are highly deformed and perhaps also polymetamorphic, as a result of multiple phases of Acadian metamorphism and deformation. Acadian staurolite-kyanite-grade metamorphism may have exceeded Taconian metamorphic conditions. As a result, little probably remains of the actual minerals and of the mineral chemistry of the Taconian or Middle Proterozoic events. Nonetheless, Taconian schistosities and minor structures, as well as major structures such as thrust faults, are found in the pre-Silurian rocks. Evidence for the Taconian schistosities and structures comes from areal tracing of these features along-strike from areas to the south in

Massachusetts where metamorphic structure of Taconian age has been documented (Sutter and others, 1985).

Pre-Silurian rocks west of the Chester dome, and to some extent east of the domes, preserve coarse- to fine-grained schistosities that are probably Taconian. These features are mapped as foliations S_1 and S_2. The latter is spatially associated with thrust faults that disrupt stratigraphic continuity. A strong downdip rodding, or plunge, of minor reclined folds of schistosity characterize these structures. Throughout southwestern Vermont, structures expressed by fold hinges and elongated quartz rods plunge northwest or southeast downdip of the S_2 surfaces. This lineation is regional and is absent from Silurian and younger rocks.

The major contacts between map units in the cover sequence rocks east of the Green Mountain massif and west of Spring Hill contain zones of intensely developed mylonitic rocks exhibiting S_2 foliation and associated lineations, suggesting that these contacts are synmetamorphic faults. The major contacts in the mantling cover rocks of the Chester and Athens domes likewise contain an intensely developed mylonitic foliation and strong downdip rodding. The repetition of units along with the intense strain suggests that these zones are also faults and that they may be the same age as the Taconian faults in the cover rocks to the west.

STRUCTURE IN THE COVER ROCKS OF THE GREEN MOUNTAINS WEST OF THE MORETOWN FORMATION

The Hoosac and Pinney Hollow Formations contain two schistosities. S_1 is parallel to bedding and is expressed by a fine-grained, phyllitic structure in the more muscovitic rocks. The entire belt of Pinney Hollow also contains a very strong S_2 foliation that is superposed on the older S_1 in almost all outcrops. Near the major thrust faults, such as the Windham and South Wardsboro thrust faults that cut the Pinney Hollow and the subjacent Rowe Schist, the S_2 foliation is intense.

Highly plicated and folded schistosity (S_1 overprinted by S_2) forms highly attenuated folds, the hinge lines of which plunge down the dip of the S_2 foliation. Such folds are referred to as reclined folds. In addition, well-developed quartz rods and chlorite and muscovite spears define this lineation. Observations here and elsewhere throughout southern Vermont, east of the Green Mountain massif, indicate that this lineation and the parallel hinge lines are characteristic of zones near faults.

The lineation is parallel to hinge lines of major folds that appear to be sheath folds (they have a somewhat tubular geometry) elongated in the direction of the thrust faulting. The lineations plunge to the southeast nearly downdip in northeast-striking and southeast-dipping foliation. To the east in the Moretown Formation, this prominent lineation of S_2 foliation fabric is present but difficult to detect because of the very strong later (F_3, Acadian) crenulation cleavage and folding in this area. The Woodburn Road fault zone contains fractured and semibrittle microfaults that suggest this fault is later, perhaps Acadian. Northwest- and northeast-trending crenulate folds, are both termed F_4 folds because the age relations are not entirely clear, especially in the area west of the Spring Hill syncline.

STRUCTURE OF THE COVER ROCKS OF THE CHESTER AND ATHENS DOMES

The belts of rocks immediately overlying the western flank of the Chester and Athens domes are all very strongly foliated and locally well rodded, with a strong, downward dipping, west-northwest plunge of hinge lines and elongated microcline augen. This foliation and linear structure is comparable to the S_2 foliation and lineation found to the west. The Ober Hill, Kenny Pond, Townshend, and other unnamed thrust faults flanking the domes are all associated with highly rodded schists and gneisses.

Mylonitization is intense throughout this belt, especially in rocks on both sides of South Brook and to the north of Grafton along Vermont Route 35. Proterozoic rocks such as the normally very coarse grained Bull Hill Gneiss and schists of the Hoosac Formation are medium-grained blastomylonite in which the grain size of the constituent minerals is greatly reduced and the rocks have a very strong metamorphic and mylonitic structure. The correspondence between the intensity of deformation and the characteristic rodding suggests that this mylonite zone and the faults may be second-generation (F_2) structures. Microscopic and mesoscopic textures associated with the prominent downdip lineation in the fault zones north of Grafton indicate a down-to-the-northwest sense of shear. Locally, the augen are ellipsoidal in the plane of the foliation, which indicates an elongation direction parallel to the lineation. However, recrystallization of potassium feldspar in many exposures has overgrown the original augen. This overgrowth occurred during Acadian metamorphism and clearly is younger than the F_2 foliation. We interpret the Townshend, Ober Hill, Kenny Pond, and associated faults as Taconian thrust faults that are responsible for imbrication of basement and cover rocks.

STRUCTURES RELATED TO THE PROCTORSVILLE SYNCLINE

At the southern closure of the Waits River Formation in Proctorsville Gulf in the Andover quadrangle to the north (fig. 2), gentle northerly plunges are common, indicating that the Proctorsville syncline plunges north. At the closure, there are two north-trending cleavages: an older bedding-plane foliation and a later set of conjugate crenulation cleavages. When traced southward into pre-Silurian rocks, the expression of the Proctorsville syncline is rapidly lost (see form-line map on sheet 2). A highly plicated schistosity is present in rocks of the Cram Hill Formation, in the Barnard Gneiss, and in the Moretown Formation to the south. Analysis of the form lines of folded schistosity in the pre-Silurian rocks of the Andover quadrangle (Ratcliffe, 1996) indicates that the only expression of the Proctorsville syncline in these older rocks is a well-developed crenulation cleavage that transects an older well-developed S_2 schistosity (see form-line map). This crenulation cleavage can be traced southward through the Andover quadrangle into the Saxtons River quadrangle where it trends southwest parallel to similar structures in the Butternut Hill fold (that is, these trend lines cross diagonally across the belt of Moretown). In the critical area south of the Butternut Hill fold, the crenulation cleavage that is coplanar with the Proctorsville syncline does not bend eastward around the Butternut Hill fold. It is associated with a set of southwest-trending F_3 folds. These F_3 folds are in turn crosscut by the Acadian (F_4 and F_5) crenulate folds.

There is no axial-surface foliation or cleavage that can be traced from the Proctorsville syncline around the southern closure of the Butternut Hill fold. These data indicate that the structure responsible for the Proctorsville syncline (which is undisputedly Acadian) cannot connect on the ground with the Spring Hill syncline (see form-line map on sheet 2). The regional structural geology, therefore, does not permit connection of the Proctorsville syncline with the Spring Hill syncline. The Proctorsville syncline must be regarded as a late fold (or alternately, a rather insignificant early F_3 Acadian fold).

Reexamination of the southern closure of the Butternut Hill fold (Ratcliffe, 1993, and this study) indicates that it plunges south rather than north and that cover rocks along the southern border appear to be right-side-up rather than inverted. The Butternut Hill fold is a late-forming antiform produced by a crenulation cleavage event (exactly like the Proctorsville syncline) that folded a steeply dipping foliation or schistosity in the pre-Silurian rocks. Plunges on multiple crosscutting sets of these late crenulation folds (F_4 and F_5, Acadian) are highly irregular and plunge very steeply because of the steep dips of the foliation prior to formation of the crenulation cleavages.

Likewise, structures which form the axial surface of the Spring Hill syncline are north- to northeast-trending crenulation cleavages that plunge both north and south. The Spring Hill syncline is a doubly plunging foliation synform probably of the same generation as the Proctorsville syncline and the Butternut Hill fold. The regional structural data are all compatible and there are no observations that indicate either an inverted succession along the northern closure of the Proctorsville syncline or northerly plunges as previously shown by Rosenfeld (1954).

STRUCTURE OF THE CHESTER AND ATHENS DOMES

One of the most striking features of the rocks in the core of the Chester and Athens domes is the lack of structural conformity of units to the form of the dome itself. This remarkable relationship indicates that the structure of the Mount Holly Complex and the Bull Hill Gneiss do not conform to the shape of the dome. For example, in the southern closure of the Chester dome, units of the Mount Holly trend northeastward across the axial trace of the dome and show little or no deflection in the map pattern. The same relationship is evident in the Athens dome. In general, the internal units in the core of the domes strike and dip at an oblique angle to the form of the domes. The units and the gneissosity generally trend northeasterly and dip moderately steeply. Deformation of the core rocks of the domes was not great enough to bring internal units into conformity with the cover rocks.

The attitude of S_2 foliation in the core rocks is at a high angle to gneissosity and the pattern of the folded foliation illustrates well two domal structures with a structural saddle between the two domes. The shape of the domes as determined by the folded internal foliation is that of an asymmetric antiform overturned eastward (cross section *A-A'*). The projection of the foliation into the lines of section for cross sections *A-A'* and *B-B'* defines a dome with a low amplitude and rather open, nonattenuated form. Map relationships like these are now known to occur throughout the Chester and Athens domes from areas to the north (Ratcliffe, 1995b; Ratcliffe and others, 1997) and to the south (Ratcliffe and others, 1992; Ratcliffe and Armstrong, 1999).

STRUCTURE OF THE SPRING HILL SYNCLINE

Rocks of the Cram Hill Formation form a well-defined synform in the central part of the map. Schistosity in all rocks of the synform is highly folded and the structure is a synform of folded schistosity that has been affected by several sets of crenulation cleavage and cross folds. A moderately persistent zone of black, carbonaceous phyllite and associated quartzite and quartz-pebble conglomerate rim the structure. At the northern end of the syncline, minor folds and elongation directions of quartz pebbles plunge gently to the south-southwest, whereas to the south, plunges of these features are north-northwest.

A prominent belt of plagioclase-spotted amphibolite (Ocha), perhaps intrusive in part, occurs near the core of the syncline. At the top of the amphibolite is a prominent but discontinuous zone of pinkish, magnetite-rich coticule and quartzite (Ochmc), and dark, feldspathic schist (Ochfs). This coticule and associated metasedimentary rocks occur in the central part of the syncline.

The syncline itself is broadly asymmetric and is overturned eastward with a steeply dipping western flank. The main synclinal fold is interpreted as an Acadian F_3 fold. Minor F_3 folds near the center of the structure plunge gently southwest, are horizontal, or plunge gently north. The southern closure of the fold plunges generally north, but the bearing and plunge of minor folds are determined by at least four sets of axial surfaces. The oldest set of folds (F_3) trends broadly northeast-southwest, has northwest-dipping axial surfaces, and is broadly folded by subsequent northwest- (F_4), northeast- (F_4), and north-trending (F_5) folds.

Although numerous folds crosscut the larger synform, the plunges overall define a doubly plunging structure (see cross sections *A-A'* and *B-B'*). Detailed mapping has not revealed evidence for repetition of the units, nor is there any evidence for northward overturning of the strata in the northern part of the syncline. In short, our data do not support the structural interpretation of Rosenfeld (1954) and of Doll and others (1961) that shows the Spring Hill structure as a northerly plunging, upward-closing antiformal syncline (fig. 1).

STRUCTURE OF THE BUTTERNUT HILL FOLD

The axial surfaces of both the Spring Hill syncline and the Butternut Hill fold appear to be related to multiple Acadian refolding of older folds. These late folds refold coarse axial-planar schistosity in the Moretown Formation south of the closure of the Butternut Hill fold. This older schistosity is interpreted as Taconian and the same general age as the composite S_1 and S_2 cleavages in the belt of Hoosac Formation through Stowe Formation to the west. Plunges of folds on the folded schistosity are erratic, particularly in the area immediately south of the closure in biotite schist (Ombs) unit of the Moretown Formation where the general trend of schistosity is east-west and steeply dipping. Analysis of the plunge of minor folds and of the crenulate cleavages indicates that as many as four separate sets of steeply plunging Acadian folds are present and that the Butternut Hill fold does not plunge northward. Instead, plunges of the folded S_2 schistosity are predominantly subvertical and plunge steeply north or northwest.

STRUCTURE OF THE ROCKS EAST OF THE CHESTER AND ATHENS DOMES

The eastern flank of the dome consists of a nearly homoclinal, eastward-dipping section of rocks from Middle Proterozoic basement up through the Devonian and Silurian Waits River Formation. The pre-Silurian section seems relatively uncomplicated by folding but is very strongly mylonitic at or near the margin of the domes. Numerous thrust faults such as the Kenny Pond, Townshend, and South Newfane thrust faults are mapped along this section and account for the irregular along-strike distribution of units. Rocks of the Moretown Formation appear very close to the basement rocks and in certain places are essentially at the contact with Bull Hill Gneiss.

Rocks of the Moretown Formation, Cram Hill Formation, and North River Igneous Suite, as well as of the overlying Northfield and Waits River Formations, all contain a very strong Acadian schistosity and several younger Acadian crenulation cleavages. Acadian thrust faults, such as the McMaster fault, duplicate the section and place Moretown and Cram Hill units onto the Northfield and Waits River Formations. The South Newfane thrust fault has been traced northward into this area from the Townshend area (Armstrong, 1994) where it has been interpreted as a significant Acadian thrust fault. In this quadrangle, it appears to die out northward, and strain appears to be transferred to the more easterly Atcherson Hollow and McMaster thrust faults.

Clearly, intense Acadian deformation and thrust faulting affected rocks on the eastern side of the domes. A more difficult and, at present, unresolved matter is the extent of Taconian deformation in the pre-Silurian rocks. We favor the interpretation that all of these rocks, with the possible exception of some intrusive rocks in the North River Igneous Suite and perhaps some of the Cram Hill, were affected by Taconian deformation and have a schistosity that is older than the first Acadian foliation in the Silurian and Devonian rocks. We conclude this because at the south end of the Athens dome, the Moretown Formation and older rocks are coextensive with rocks to the west that contain the older Taconian structures (Ratcliffe and Armstrong, 1999; Armstrong, 1994). We conclude that the intensity of Acadian structures east of the domes has obscured the older structures and suggest that the schistosities in pre-Silurian rocks east of the domes have a composite Taconian and Acadian schistosity. This is the result of the nearly homoclinal dip and tightly appressed folds along the eastern flank of the dome. The major deformational front that separates the less deformed Taconian schistosity from the composite Acadian and Taconian schistosity is parallel to the strongly developed northeast-trending F_4 folds in the core of the domes.

This transition is illustrated by the folding of the foliation in the Bull Hill Gneiss as it increases from gentle warping in the core of the dome to intense plication in the eastern one-third of the dome. This zone of abundant Acadian shortening is shown in cross sections *A-A′* and *B-B′* by prominent folds of the older Townshend, Kenny Pond, and other Taconian thrust faults. Locally, Acadian thrust faults are present in this zone. The short but well-defined Bull Hill fault forms the western boundary of an intense zone of upright to westerly overturned F_4 folds in the Bull Hill Gneiss and associated cover rocks west of the Townshend thrust. The axial traces of Acadian folds and the approximate position of this zone of intense Acadian shortening are shown on the map. As noted previously, Devonian granite dikes are very strongly folded in this area, whereas in the area to the west, these dikes are highly discordant and only weakly folded.

STRUCTURE IN THE SILURIAN AND DEVONIAN ROCKS EAST OF THE DOMES

Silurian and Devonian rocks east of the Chester and Athens domes contain a strong second-generation Acadian foliation (Acadian S_2) which significantly transposes an early bed-parallel foliation or compositional layering (Acadian S_1). Relict S_1 features are typically found in Acadian F_2 fold hinges within competent quartzite and calcareous (punky) beds within both the Waits River and Northfield Formations, as well as in the granofels and volcaniclastic rocks of the Waits River. No Acadian F_1 folds are recognized. Both S_1 and S_2 features are locally transected by at least three different generations of Acadian crenulation cleavage that trend northeast, northwest, and north. Folds associated with the crenulation cleavages typically produce outcrop-scale plications and digitations to S_1 and S_2 foliations but only rarely (and weakly) affect the map geometry of lithologic contacts. Ductile faults within the Silurian and Devonian section are related either to intense Acadian S_2 foliation development or to subsequent intense localization of younger crenulation cleavage development. Both of these fault sets formed during Acadian regional metamorphism.

The most outstanding structural feature within the Silurian and Devonian section is the Saxtons River fold, a composite S_1 and S_2 synformal structure that is locally transected by the younger Acadian crenulation cleavages. The axial trace of this F_3 structure has a stratigraphic symmetry defined by the volcaniclastic, volcanic, and intrusive units of the Waits River Formation east of the Athens dome. Most mesoscopic folds are associated with the Acadian S_2 foliation development and are defined by folded S_1 foliation that is parallel to bedding. Immediately south of the village of Saxtons River, F_2 axial surfaces are broadly coplanar with Acadian S_1 foliation, and F_2 hinge lines plunge southward at gentle to moderate angles. Rosenfeld (1968) interpreted these folds as nappes related to the earliest Acadian recumbent deformation. He further interpreted their steep axial orientations to subsequent rotation during later dome-stage deformation. The southward plunge of the folds here suggested to him that the southern closure of this Standing Pond fold (the actual closure being in the Townshend quadrangle, 0.3 km to the south of this locality) was antiformal. In addition, Rosenfeld (1968) mapped the core of the antiformal structure as Gile Mountain Formation, stratigraphically above the surrounding Waits River Formation, and noted up-to-the west minor folds. Thus, the antiform was interpreted to be a nappe-stage antiformal syncline.

Our analysis, however, shows that the folds associated with Rosenfeld's (1968) nappe stage are second-generation Acadian folds and fold an older foliation (our Acadian S_1). In addition, Acadian F_2 hinges vary along the strike of the major structure's axial trace, plunging south in some areas and north in others. Because second-generation or later folds are formed on previously folded strata, the plunges and rotation sense of minor folds need not be reliable indicators of original facing directions. The variation in plunge appears to be related to the position of F_2 folds on older F_1 fold limbs. Near the southern closure of the Saxtons River fold, F_1 folds plunge moderately to the north. Bedding in this area dips northwest, north, and northeast around the closure. The actual F_1 closure plunges north and is synclinal, not antiformal. F_2 folds appear to be related to subsequent tightening of previously developed S_1 and F_1 structures, possibly in a strain continuum. Additional variation in F_2 (and F_1) hingeline trends and plunges locally was caused locally by development of additional Acadian crenulation cleavage folds. The northeast- and northwest-trending folds (both termed F_4 folds) appear to be broadly coeval and may be a conjugate pair. They are typically crosscut by even younger north-trending (F_5) crenulate folds. The synformal closure of the Saxtons River fold has a very important bearing on the regional structural interpretation, namely that the Waits River belt east of the syncline tops to the west away from the rocks of the New Hampshire sequence.

METAMORPHISM

PREVIOUS WORK

Rosenfeld (1954) studied the pelitic assemblages in the Saxtons River quadrangle, and described a westward-increasing metamorphic gradient from the chlorite zone along the eastern quadrangle boundary, through the biotite and garnet zones in the central part, to kyanite-staurolite grade rocks within the Spring Hill area in the western half of the quadrangle. Metamorphic grade then decreased to the garnet zone along the immediate eastern flank of the Green Mountain massif in the westernmost part of the quadrangle. Rosenfeld (1954, 1968) believed that this distribution of metamorphic grades was symmetrical about the basement-cored Athens dome and related to upwelling of high-grade rocks during dome development.

In another study just east of the Saxtons River quadrangle, Kruger (1946) showed two metamorphic gradients within the Bellows Falls 15-minute quadrangle. In this area, chlorite- and biotite-grade rocks along the Connecticut River and within the Waits River and Orfordville Formations were separated from higher grade rocks to the east by the Northey Hill fault, which is interpreted as the contact between the Vermont and New Hampshire stratigraphic sequences. Rocks immediately east of the fault and within the New Hampshire sequence were mapped at staurolite and sillimanite grades. Rocks west of the chlorite- and biotite-grade rocks (and within the Vermont sequence), including rocks within the Wellington Hill anticline (Kruger, 1946), were mapped at garnet and staurolite grade. This metamorphic-grade symmetry (about the chlorite- and biotite-grade rocks) was interpreted to be tectonic on the eastern flank and gradational on the western flank, increasing toward the Athens and Chester dome structures.

Boxwell and Laird (1987) reported on structural and petrologic studies of amphibole-bearing rocks at greenschist and epidote-amphibolite facies within the Saxtons River, Springfield, and Townshend quadrangles. Boxwell and Laird included the Silurian and Devonian Standing Pond Volcanics (volcanic rocks of the Waits River Formation, in this study) lying between the underlying Waits River and overlying Gile Mountain Formations on the eastern flank of the Chester and Athens domes and within the Connecticut Valley metamorphic low. Their model included both a prograde and retrograde origin for the low, with initial prograde distribution of isograds controlled by peak-temperature dome-stage deformation, following nappe formation (Rosenfeld, 1968), with subsequent retrograde modification of isograd distribution following dome-stage deformation.

PRESENT STUDY

Armstrong (1992) presented detailed thermobarometric analyses for rocks immediately south of the Saxtons River quadrangle and on both sides of the Athens dome. This work showed that both pressure and temperature conditions increase west to east across the dome (475°C at 6.5 kbars to 625°C at 10 kbars) and that related isotherms and isobars were not substantially deflected by dome development; the results suggest that coeval dome evolution did not produce structures of sufficient amplitude to deflect isotherms and isobars. This interpretation is also consistent with the structure of the domes, which are now known to have low amplitudes (Ratcliffe and Armstrong, 1999; Ratcliffe and others, 1992). Petrographic studies conducted during this study are consistent with a west-to-east increase in pressure-temperature conditions across the western part of the Saxtons River quadrangle. Garnet-chlorite and garnet-chlorite-biotite-plagioclase assemblages are present in rocks of the Rowe Schist belt and Moretown Formation west of Spring Hill. Garnets from these samples contain planar inclusion trails within the cores that are progressively kinked, folded, and deformed toward the actual garnet edges, indicating porphyroblast growth and attainment of peak-temperature conditions during dome development; the planar, undeformed inclusion trails in garnet cores indicate the onset of garnet growth prior to crenulation cleavage development in the domes (Ratcliffe and Armstrong, 1995).

Similar fabric relationships are present within garnet-biotite-staurolite assemblages on the south, east, and west sides of the Spring Hill syncline. The presence of staurolite (and kyanite) in these rocks suggests that pressure and temperature greater in these areas than farther west. The distribution of these assemblages within rock units of the Moretown Formation that are present in both garnet- and staurolite-kyanite-grade areas indicates that isograd distribution and extrapolated isothermal and isobaric surfaces are not controlled entirely by the center of the Chester and Athens domes. The eastward increase in metamorphic grade is consistent with the model of eastward-increasing tectonic loading presented by Armstrong (1992).

The eastern side of the Chester and Athens domes shows several different relationships. First, all rocks east of the garnet schist member (Ochgs) unit of the Cram Hill Formation are at garnet grade; garnets in the unit contain deformed inclusion trails within cores that can be traced into matrix folds and deformational fabrics (Ratcliffe and Armstrong, 1995). Second, deformation on the east side includes development of Acadian foliation that transposes older Taconian S_2 foliation and, at least in part, postdates Acadian garnet growth. Thus, late deformation appears to have been greater in intensity on the east side of the dome than on the west side, and is younger (with respect to porphyroblast development) than Acadian cleavage development on the west side. These relationships strongly suggest that fabric development and related metamorphic mineral growth was more intense on the east side of the domes than on the west side. This conclusion is also supported by the highly deformed Acadian granite dikes east of the domes and the weaker deformation in these dikes along the western side of the domes. The deformation possibly was diachronous and the lower grade metamorphic conditions on the east side occurred when rocks already had been uplifted as a whole to shallower crustal levels and lower pressure-temperature conditions.

SUMMARY AND CONCLUSIONS

Our interpretation of the structure of the Saxtons River quadrangle, shown in cross sections *A-A′* and *B-B′*, is that the Chester and Athens domes, as well as the Butternut Hill fold, are antiformal, right-side-up structures. The cover rocks are complexly faulted and folded and contain a regional penetrative Taconian schistosity, our Taconian S_2 foliation. Form-line maps of this foliation (fig. 3), which is developed in the core of the domes and in the cover rocks, define highly folded patterns. Areas of anomalous plunges exist where the folded schistosity is subvertical and is folded along multiple late Acadian folds that are responsible for the formation of the domes.

Where crenulation cleavages cross the older Taconian S_2 schistosity on the flanks of the domes, highly irregular folds (F_3, F_4, F_5) having limited continuity are abundant. These Acadian folds (fig. 3) are mutually interfering and produce either a basin and dome structure where the reference schistosity is subhorizontal, or vertically or very steeply plunging folds where the older schistosity is steeply dipping. Because of the irregular fold patterns, plunges of F_3, F_4, and F_5 folds developed in the pre-Silurian rocks are not consistent regionally.

Acadian schistosity in the Silurian and Devonian rocks (fig. 3) is also highly folded by Acadian cross folds of several generations. Two sets of F_4 folds are recognized and the age relationships are not clear. In general, the northwest-trending folds, such as those developed north of the domes, tend to be overturned to the southwest. These northwest-trending F_4 folds are also present in the core of the domes and in the pre-Silurian rocks west of the domes. A very prominent set of northeast-trending F_4 Acadian folds affects the Athens dome in a broad belt extending from near Jamaica to Springfield and is responsible for the formation of the structural saddle between the Chester and Athens domes. Still younger upright, north-south-trending F_5 folds add to the complexity.

Our conclusion is that younger and more tightly compressed Acadian (F_4) folds are present on the eastern flank of the domes and in the Silurian and Devonian rocks east of the domes than to the west. This increased late Acadian strain east of the domes was synmetamorphic based on garnet-inclusion structures. Garnets in the western areas tend to overgrow microcrenulations of schistosity and appear to be static, whereas to the east, evidence for continued flattening of matrix schistosity occurred during the growth of the peak metamorphic assemblage (Ratcliffe and Armstrong, 1995).

On a regional basis, we suggest that intense Acadian deformation continued longer in areas east of the domes. On the basis of our studies of the Saxtons River quadrangle and of our regional studies (fig. 3), we conclude that the domes resulted from multiple episodes of Acadian crustal shortening rather than from diapiric uprise. We find no compelling evidence for Acadian nappes and specifically no evidence for recumbent folds of basement and cover.

In general, the sequence of events we propose is as follows: (1) Imbricate thrust faulting of basement gneisses and cover rocks occurred during the Taconian orogeny. In this synmetamorphic event, an easterly inclined thrust stack formed. (2) Following the Taconic orogeny, isostatic rebound occurred, returning the previously inclined thrust stack to a more gently east dipping attitude. (3) Silurian and Devonian rocks were deposited on the Taconian thrust stack with a low angular discordance. (4) Rifting and subsidence occurred during deposition from the Late Silurian through the Middle Devonian. (5) Acadian crustal shortening began in the Middle Devonian with the formation of S_1 and S_2 Acadian folds. Preferential shortening of the Silurian and Devonian rocks took place because of the indurated nature of the pre-Silurian rocks, perhaps producing a zone of accumulated detachment strain in the younger rocks. (6) Reactivation of the Taconian thrust faults in the Taconian thrust stack initiated fault-bend folds in the thrust stack. These folds propagated upward and westward during continued shortening and crustal loading from the east to start dome uplift. (7) Locking of the crustal thrust faults in the west prompted the start of crustal delamination and formation of progressively younger compressional phases from the domes eastward.

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