ACADIAN EXTENSION AROUND THE CHESTER DOME, VERMONT

by

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INTRODUCTION

Low-angle normal faults are important both in regions undergoing crustal extension and in mountain belts dominated by crustal shortening. Extension and shortening may be coeval in mountain belts, as has been proposed for the high Himalayas (Hodges et al., 1992) and Tibetan plateau (Molnar and Tapponnier, 1978), or extension may postdate shortening and thickening of continental crust as has been suggested for the Basin and Range province (Coney, 1979; Sonder et al., 1987). Extension was also an important process in the evolution of ancient mountain chains (e.g. the Alps, Selverstone et al., 1984; Selverstone, 1985).

Although extension appears to be a fundamental process during orogenesis, major Paleozoic normal faults in the Appalachians have not been widely recognized. A notable exception is the Lake Char-Honey Hill fault in southeastern New England, which has been interpreted as an Alleghenian low-angle normal fault (Goldstein,1989; Getty and Gromet, 1992). Before Goldstein's (1989) breakthrough work showed that the sense of displacement was normal on this fault, it was universally regarded as a thrust. Astonishingly little evidence exists for normal faulting related to the Acadian orogeny in the New England Appalachians. Castonguay et al. (2001) presented thermochronological evidence for Silurian to Early Devonian exhumation and extension along the St. Joseph fault in Quebec, and Spear (1992) suggested that the Monroe Line along the New Hampshire-Vermont border might be a normal fault based on contrasting styles of metamorphism of rocks across the contact.

Karabinos (1999, 2000) proposed that faults previously interpreted as Taconic thrusts around the Chester dome in southeastern Vermont (Thompson et al., 1990; Ratcliffe et al., 1997) are really Acadian normal faults. This field trip presents evidence for an extensional shear zone around the Chester dome and explores the tectonic implications of Early Devonian crustal extension. If normal faulting was important in western New England, an entirely new aspect of the Acadian orogeny will have to be taken into account in tectonic reconstructions, interpreting geologic maps, and models of ore genesis within the belt.

GEOLOGIC SETTING

Lithotectonic Units

The Berkshire and Green Mountain massifs and the Chester dome are cored by Middle Proterozoic Grenvillian basement (see Karabinos and Aleinikoff (1990), Ratcliffe et al. (1997), and Karabinos et al. (1999) for some geochronological constraints). West of the massifs, the Taconic klippen are composed of Late Proterozoic to Middle Ordovician slate and phyllite originally deposited as shale and siltstone on the continental slope and rise of the passive Laurentian margin (Fig. 1). The klippen structurally overlie a coeval sequence of clastic and carbonate rocks, which formed on the continental shelf of Laurentia. East of the massifs, the Tyson, Hoosac, and Pinney Hollow Formations are equivalent to the basal units found in the Taconic klippen. The Rowe Formation in Massachusetts and the Ottauquechee and Stowe Formations in Vermont form the remnants of an accretionary wedge of oceanic crust and sediments and the Moretown Formation contains forearc basin deposits (Fig. 1; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985). The Shelburne Falls arc (Fig. 1) is composed of the Barnard Volcanic Member of the Missisquoi Formation in Vermont and the Hawley Formation in Massachusetts (Karabinos et al., 1998). The Collinsville Formation and the Hallockville Pond Gneiss in Massachusetts are also part of the Shelburne Falls arc, but they form isolated bodies. The Connecticut Valley trough (Fig. 1) contains metasedimentary and metavolcanic rocks of Silurian to Early Devonian age.

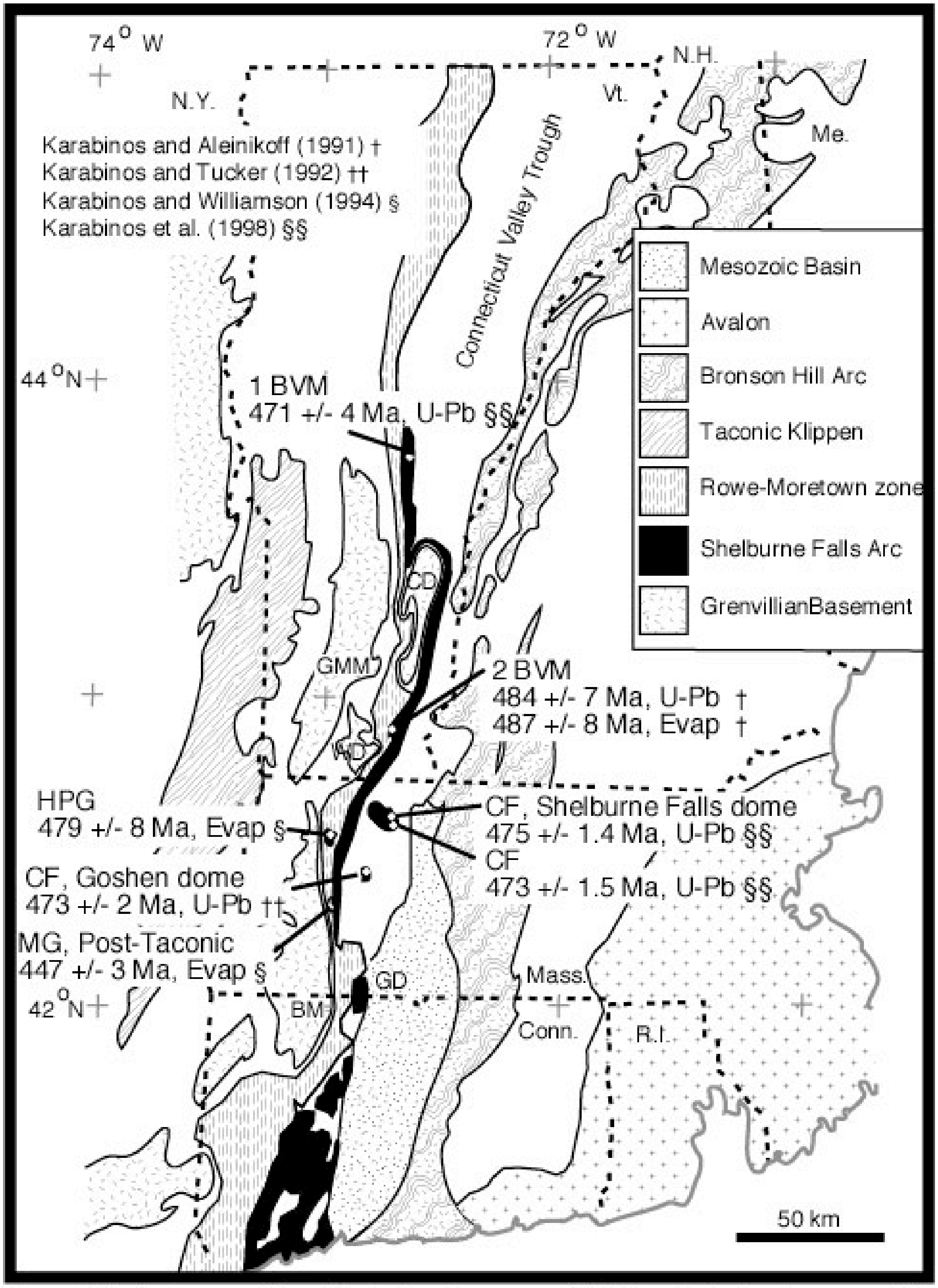

Figure 1. Tectonic map of New England and summary of U-Pb and single-grain evaporation (evap) zircon ages from the Shelburne Falls arc in western New England. Shelburne Falls arc rocks include the Barnard Volcanic Member (BVM), Collinsville Formation (CF), and Hallockville Pond Gneiss (HPG). The post-Taconic Middlefield Granite (MG) indicates that Taconic thrusting ended before ca. 447 +/- 3 Ma. GMM- Green Mountain massif. BM- Berkshire massif. WD- Wilimington dome. GD- Granville dome.CD- Chester dome.

Taconic and Acadian Orogenies

During the Ordovician Taconic orogeny (470 to 455 Ma), Laurentia collided with an island arc that formed above an east-dipping subduction zone. The characteristic deformation pattern was westward-directed thrusting of rocks of the continental margin, accretionary wedge, forearc basin, and arc complex (Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985). Until recently the Bronson Hill arc in western New Hampshire and central Massachusetts (Fig. 1) was commonly identified as the arc that collided with Laurentia. However, Tucker and Robinson (1990) pointed out that the 454 to 442 Ma age range of volcanic and plutonic rocks in the Bronson Hill arc is younger than some metamorphic cooling ages from rocks caught in the Taconic collision zone (e.g. Laird et al., 1984). Karabinos et al. (1998) argued that the older Shelburne Falls arc (485 to 470 Ma) in eastern Vermont and western Massachusetts (Fig. 1) collided with Laurentia during the Taconic orogeny, and that the Bronson Hill arc formed above a west-dipping subduction zone <u>after</u> a reversal in subduction polarity. Karabinos et al. (1998) further suggested that this new west-dipping subduction zone accommodated plate convergence, thus bringing the Taconic orogeny to an end and setting the stage for the Acadian orogeny. According to this model, the Laurentian margin was active during the Silurian and the Connecticut Valley trough formed as an extensional back-arc basin above a west-dipping subduction zone (Karabinos et al., 1998; Karabinos, 1998).

The Acadian orogeny began in the Late Silurian and continued into the Middle Devonian; it resulted from the protracted collision of Laurentia and Composite Avalon (Robinson et al., 1998, Bradley et al., 2000, Tucker et al., 2001). Studies using high-precision geochronology have demonstrated that what was formery regarded as a single ‘Acadian’ orogeny is, in fact, a complex series of tectonic events spanning tens of millions of years (Robinson et al., 1998; Bradley et al., 2000, Tucker et al., 2001). Bradley (1983) proposed that the collision occurred above two subduction zones, one dipping beneath each continental margin, but other models invoke a single subduction zone under Avalon (e.g. Robinson et al, 1998, Tucker et al., 2001).

In western New Hampshire and eastern Vermont, including the area of the Chester dome, two important phases of the Acadian orogeny have long been proposed and were portrayed on the Geologic Map of Vermont (Doll et al., 1961). According to this interpretation, an early nappe stage created kilometer-scale recumbent folds, and a later dome phase refolded the early nappes (e.g., Rosenfeld, 1968; Hepburn et al., 1984; Thompson et al., 1993). Ratcliffe et al. (1997) and Hickey and Bell (2001), however, have questioned this long-standing structural interpretation and argued for more complex Acadian folding histories to explain the structures in and around the Chester dome.

In my opinion, the nappe and dome stages of deformation explain much of the structural geometry of the region, although I am not convinced that the nappes involved units structurally below the Silurian and Devonian sequence. However, an important gap in our understanding of the tectonic history of this area is a mechanism for the dramatic thinning of units around the Chester dome and other structures, including the Wilmington dome and the Jamaica anticline. I believe that the evidence points to an intermediate period of extension along a normal-sense shear zone after the nappe stage of deformation but prior to the doming stage.

Timing of Acadian Events

The age of Acadian deformation and metamorphism in southeastern Vermont is reasonably well constrained by a variety of isotopic studies. $^{40}Ar/^{39}Ar$ hornblende cooling ages indicate that Acadian metamorphism was waning by 380 Ma in the vicinity of the Chester dome (Laird et al., 1984; Karabinos and Laird, 1988; Spear and Harrison, 1989). This agrees remarkably well with two studies of the age of garnet growth. Christensen et al. (1989) used the radial variation of $^{87}Sr/^{86}Sr$ to estimate that garnet porphyroblasts from Townsend, Vermont, grew at about 380 Ma over an average interval of approximately ten million years. Vance and Holland (1993) used Sm-Nd and U-Pb measurements and chemical zoning to argue that garnet from Gassetts, Vermont, grew at about 380 Ma during increasing temperature and decompression of 2.5 kbar, corresponding to exhumation of approximately 7 km. Based on U-Pb dating of granitic intrusions and cross-cutting relationships, Ratcliffe et al. (2001) concluded that Acadian deformation in the vicinity of the Chester dome began before 392 Ma and that it persisted with uneven distribution in southeastern Vermont for another thirty million years. To the south, along strike in western Massachusetts, the end of intense Acadian deformation is tightly bracketed by a 376 ± 4 Ma age (U-Pb zircon, TIMS) of a highly deformed tonalitic sill from the Granville dome (Karabinos and Tucker, 1992) and a 373 ± 3 Ma age ($^{207}Pb/^{206}Pb$ zircon, evaporation) for the post-kinematic Williamsburg Granodiorite (Karabinos and Williamson, 1994).

THE CHESTER DOME

Together the Chester (Thompson, 1950) and Athens (Rosenfeld, 1954) domes form an elongated basement-cored structure approximately twelve km wide and thirty km long (Fig. 2). For simplicity, the entire structure is here called the Chester dome. Since the original work by Thompson and Rosenfeld, this classic mantled gneiss dome has been the focus of numerous structural (Rosenfeld, 1968; Nisbet, 1976; Bell and Johnson, 1989; Ratcliffe et al., 1997; Hickey and Bell, 2001) and metamorphic studies (Rosenfeld, 1968; Thompson et al., 1977; Cook and Karabinos, 1988; Christensen et al., 1989; Chamberlain and Conrad, 1993; Vance and Holland, 1993).

Doll et al. (1961) showed a continuous stratigraphic sequence in the mantling units around the Chester dome with two major unconformities: one at the base of Late Proterozoic clastics and the other at the base of the Silurian and Devonian sequence (Fig. 3). Because some of the units are quite thin, Doll et al. (1961) were obliged to combine them on the map as “Pinney Hollow, Ottauquechee, and Stowe Formations, undifferentiated”. More recently, some of the lithological contacts have been interpreted as thrust faults (Thompson et al, 1990; Ratcliffe et al., 1997 and references therein). Mylonitic textures are indeed widespread in some of the mantling units around the dome, supporting the fault interpretation of some of the lithologic contacts. The problem addressed on this trip, however, is that instead of structural repetition and thickening of units, we observe that the Late Proterozoic to Ordovician sequence around the Chester dome is dramatically thinner than it is on the east side of the Green Mountain massif and further north (Doll et al., 1961). This observation is more consistent with normal faulting and extension around the Chester dome than it is with thrusting. I believe that it is worth testing the extension hypothesis because if it is correct, we will gain important insight into the Acadian orogeny. And if we can demonstrate that the hypothesis is wrong and that normal faulting did not occur, then some other explanation for the dramatic attenuation of units around the Chester dome must be sought.

The Chester dome is outlined by the contact between Late Proterozoic albite schist of the Hoosac Formation and Middle Proterozoic rocks of the Bull Hill Gneiss and Mount Holly Complex. This contact dips away from the dome except along the southwestern margin where it is steeply overturned to the east. Structurally higher and approximately parallel to this important boundary is a very strong mylonitic fabric, mostly contained within the cover sequence. The thickness, intensity, and position within the cover sequence of this mylonite zone varies around the dome, but it appears to completely surround the dome. It also coincides with attenuated and/or omitted units of the cover sequence. This mylonite zone predates doming and its orientation helps define the geometry of the dome. Locally, a steep crenulation cleavage overprints the mylonitic fabric and strikes parallel to the long axis of the Chester dome. I interpret this as a dome-stage deformation fabric. Late-stage porphyroblasts commonly overgrow the strong mylonitic fabric, but these textures do not prove that “ peak-temperature metamorphic mineral growth occurred after the majority of dome-related deformation” as claimed by Ratcliffe et al., (1997, p. B6-28) because the mylonitic fabric predates the doming phase of deformation.

Stratigraphic Problems

As noted in the previous section, the Chester dome is outlined by the contact between the Hoosac Formation and structurally lower Middle Proterozoic rocks. In the northern part of the Chester dome, quartzite, marble, albite schist and aluminum-rich schist (commonly referred to as Gassetts schist) are exposed between the

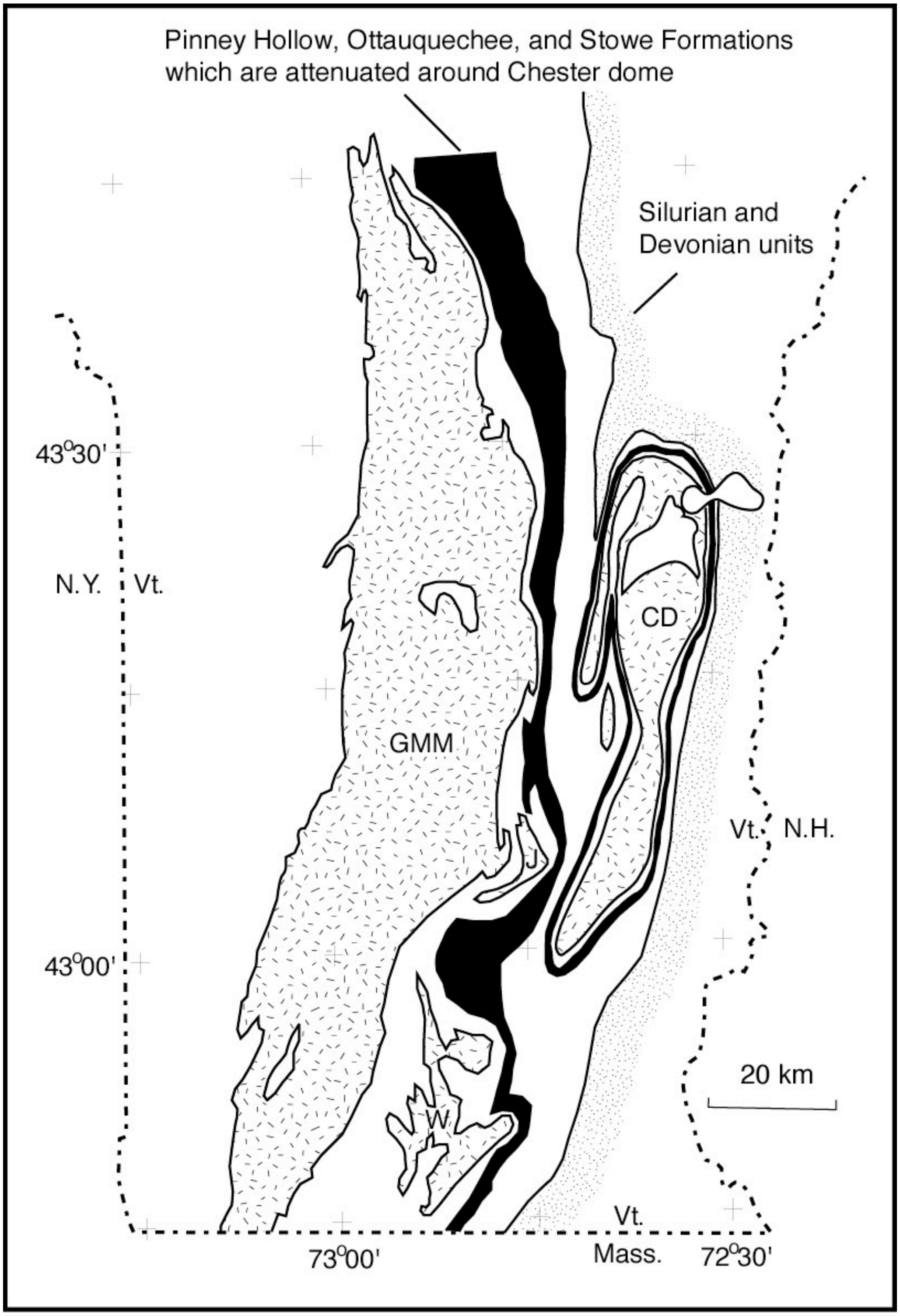

Figure 2A. Simplified map showing the distribution of the Pinney Hollow, Ottauquechee, and Stowe Formations (black), which are extremely thin around the Chester dome (CD). Grenvillian basement- hatchured pattern, Silurian and Devonian units- dotted pattern. G- location of Gassetts road cut, GMM- Green Mountain massif, J- Jamaica anticline, W- Wilmington dome. Based on Doll et al. (1961).

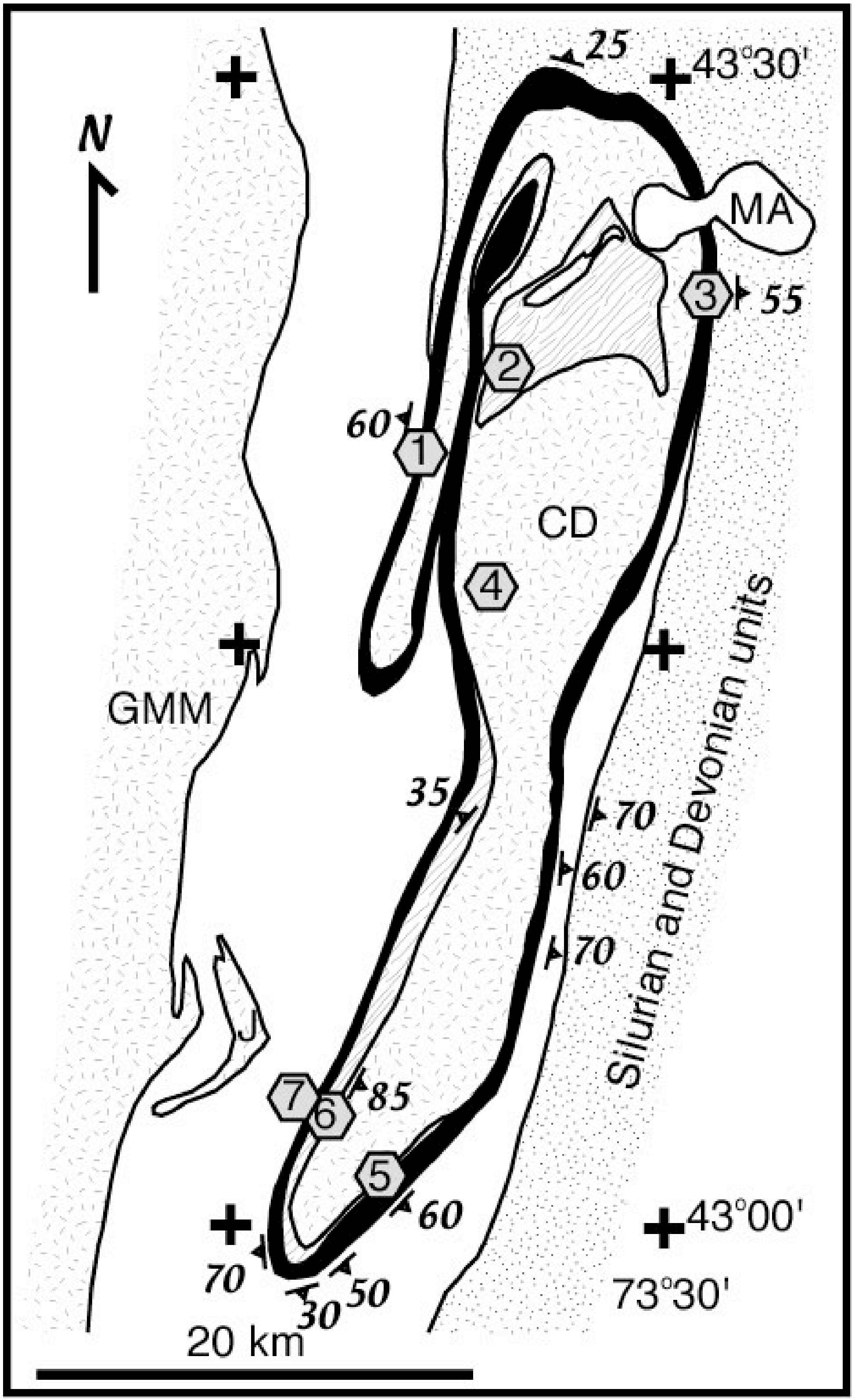

Figure 2B. Map of the Chester dome (CD) showing the distribution of the mylonite zone (black) and the location of field trip stops. Grenvillian basement- hatchured pattern. GMM- Green Mountain massif. J- Jamaica anticline. MA- Mount Ascutney. Strike and dip symbols show the orientation of the mylonitic fabric.

Age (Doll et al., 1961; Karabinos et al., 1998)	Unit Names (Doll et al., 1961)	Tectonic Setting (Karabinos et al., 1998; Karabinos and Thompson, 1997)
Devonian-Silurian	Gile Mountain Fm. Standing Pond Volcanics Waits River Fm. Northfield Fm. Shaw Mountain Fm.	Conn. Valley Trough Back-arc basin deposits
Ordovician	Missisquoi Fm	Fore-arc basin deposits, arc volcanics and plutonic rocks
Ordovician-?Cambrian	Stowe Fm. Ottauquechee Fm.	Accretionary wedge deposits
Cambrian-Late Proterozoic	Pinney Hollow Fm. Hoosac Fm. Tyson Fm.	Slope-rise deposits on Laurentian margin
Middle Proterozoic	Mount Holly Complex	Grenvillian Basement rocks of Laurentia

Figure 3. Diagram of lithologic units in southeastern Vermont together with their ages and interpreted tectonic settings.

Hoosac Formation and the basement rocks or as isolated patches within the basement (Doll et al., 1961). Doll et al. (1961) included these rocks in the Cavendish Formation and interpreted them as part of the cover sequence. Other workers have correlated these and similar rocks with the Tyson and Hoosac Formations (e.g. Thompson et al., 1977; Karabinos, 1984; Thompson et al., 1990; Karabinos and Thompson, 1997). Ratcliffe et al. (1997) reassigned these metasediments to the Mount Holly Complex; a stratigraphic change that requires important structural revisions. This problem is so important to tectonic interpretations of the Chester dome that it requires further discussion.

An ongoing problem in Vermont is distinguishing Grenvillian basement (the Mount Holly Complex of Doll et al., 1961) from pre-Taconic cover rocks. Nowhere is this problem more acute than around the Chester dome where metamorphic grade reached the kyanite zone and intense deformation fabrics overprinted textures useful for indentification of the units. The age of the Cavendish Formation is currently a controversial issue that is central to structural interpretations of the Chester dome. Doll et al. (1961) included the Cavendish Formation in the pre-Taconic cover sequence stratigraphically below the Cambrian Tyson and Hoosac Formations. The metasedimentary units in the Cavendish Formation were later reassigned to the Tyson and Hoosac Formations by Thompson et al. (1977), Karabinos (1984), and Thompson et al. (1990), and the age assignment of these formations was changed to Late Proterozoic to Cambrian. Another difficulty with the Cavendish Formation, as defined by Doll et al. (1961), emerged when Karabinos and Aleinikoff (1990) used U-Pb zircon dating to show that the distinctive augen gneiss member, the Bull Hill Gneiss, forms a major post-Grenvillian plutonic suite that intruded the Mount Holly Complex at ca. 965 to 945 Ma. The Bull Hill Gneiss member must be substantially older than the schist and impure marble units in the Cavendish Formation of Doll et al. (1961), assuming that correlation of these metasediments with the Late Proterozoic to Cambrian Tyson and Hoosac Formations is correct.

Most workers informally abandoned the term “Cavendish Formation” because its metasedimentary members were reassigned to other formations and its metaigneous member was reinterpreted as a post-Grenvillian intrusive suite stratigraphically unrelated to the metasedimentary units (Karabinos and Aleinikoff, 1990). However, in a major departure from earlier interpretations, Ratcliffe et al. (1996, 1997) revived the name Cavendish Formation and assigned its metasedimentary units to the Mount Holly Complex. This new interpretation, if correct, would require a dramatic change in structural reconstructions of the Green Mountain massif and Chester dome (e.g., Ratcliffe, 2000a, b).

To test this new age assignment, Karabinos et al. (1999) separated detrital zircons from a quartzite layer from Cavendish gorge, the type locality of what Richardson (1929) called the Cavendish Schist. This is also the location where Ratcliffe et al. (1996, 1997) obtained a 1.42 ± 0.02 Ga age on a tonalite, the Felchville Gneiss, that they interpreted as cutting the Cavendish Formation. It is important to note, however, that the interpretation of an intrusive contact between the Felchville Gneiss and the Cavendish Formation was based on map-scale patterns rather than direct field observations (Ratcliffe et al., 1997). There are no xenoliths of Cavendish Formation in the tonalite, nor are there apophyses of tonalite in the Cavendish Formation. The contact between the 1.42 Ga tonalite and the Cavendish Formation, where best exposed in Cavendish gorge (Stop 2), is highly sheared, hence a more plausible explanation for the truncation of the metasedimentary units is a fault rather than an intrusive contact.

The zircon grains analysed by Karabinos et al. (1999) varied from nearly euhedral to nearly spherical. Virtually all of the grains had pitted surfaces and showed some rounding of edges and terminations. The grains exhibited oscillatory zoning typical of zircons that crystallized from a magma. This is important because it precludes the possibility that metamorphic overgrowths were inadvertently analysed. Karabinos et al. (1999) analyzed ten zircons grains by the Pb evaporation method and fifteen grains by the sensitive high resolution ion microprobe (SHRIMP). All twenty-five grains yielded ages less than 1.42 Ga. Seven of the grains gave ages consistent with derivation from the Bull Hill Gneiss, which postdates the Grenville orogenic cycle. The ages of eight grains fall in the interval of 1.0 to 1.1 Ga and four in the interval of 1.1 to 1.2 Ga; these are very common igneous ages in the Grenville terrane of Canada and the Adirondacks. Three grains yielded ages of approximately 1.3 Ga, consistent with derivation from trondhjemitic gneisses of the Mount Holly complex (Ratcliffe et al., 1991). There is no reason to suspect that this wide range of single grain ages reflects multiple episodes of metamorphic growth of zircon or complicated Pb-loss events, as suggested by Ratcliffe (2000a). On the basis of these zircon ages, the Cavendish Formation should not be assigned to the Mount Holly Complex. In spite of these age constraints, however, Ratcliffe (2000a, b) did include the Cavendish Formation in the basement sequence. Obviously, the correct age assignment of these rocks is crucial for understanding the structural geometry of the northern part of the Chester dome.

Metamorphic Isograds around the Chester Dome

It is important to understand the timing and complex nature of metamorphism around the Chester dome because if extension did occur, there should be a record of contrasting pressure-temperature paths in footwall and hanging-wall rocks. If such a record exists, it would provide strong evidence for extension, constrain the location of the normal shear zone, and help date the extension.

Rosenfeld (1968) was the first to recognize that truncated inclusion trails in garnet porphyroblasts record two separate periods of garnet growth. He suggested that the first growth episode occurred during Taconic metamorphism and that the second episode occurred during Acadian metamorphism. Karabinos (1984) showed that garnets with textural unconformities also preserve reversals in chemical zoning that developed during a period of partial resorption of the garnet cores before the rims grew. Laird and Lanphere (1981) and Laird et al. (1984) documented polymetamorphic textures, zoning, and isograds in mafic schist from Vermont that are largely consistent with the polymetamorphic history inferred from pelitic rocks. Laird et al. (1984) used $^{40}Ar/^{39}Ar$ cooling ages to show that Taconic metamorphism affected mafic schist in northern Vermont. To date, however, no unequivocal evidence for Taconic metamorphism in the Chester dome has been published, although J. Cheney and F. Spear have unpublished ion-microprobe monazite ages that are consistent with Ordovician metamorphism (J. Cheney, personal communications, 2002).

Cook and Karabinos (1988) used inclusions in garnet porphyroblasts that contain textural unconformities to construct two sets of isograds in southeastern Vermont. The ‘older’ set of isograds is based on inclusions found in garnet cores. The ‘younger’ set is based on inclusions in the garnet rims and the matrix assemblage. The age of the younger metamorphism and garnet rims is clearly Acadian. The garnet cores, however, could have grown during either Taconic or an early stage of Acadian metamorphism. In southeastern Vermont, there is a strong spatial correlation between high grade areas of both the older and younger sets of isograds *and* structures that expose Grenvillian basement (Fig. 4). The strong spatial correlation of the high grade areas of the two sets of isograds suggests to me that both episodes of garnet growth are Acadian, and that the hiatus in garnet growth reflects a major tectonic event, probably extension along a normal-sense shear zone.

Figure 4 shows that the staurolite isograd closely follows the outline of the Chester dome. One reasonable interpretation of this pattern is that the staurolite isograd formed as a horizontal surface and was folded during the doming stage of deformation. Another possibility is that the staurolite isograd merely reflects the distribution of lithologies favorable to staurolite growth rather than a boundary between rocks that experienced different metamorphic conditions. The bulk composition argument could be a reasonable explanation for the early staurolite isograd (Fig. 4A). This isograd is defined by the reaction chloritoid + quartz = garnet + chlorite + staurolite + H_2O. Appropriate compositions are found only in the high-Al layers of the Hoosac (i.e., Gassetts Schist) and Pinney Hollow Formations, and these are located in the lowest part of the mantling sequence and as outliers in the Chester dome (Cook and Karabinos, 1988). It seems unlikely, however, that bulk composition alone controls the late staurolite isograd shown in Fig. 4B. The second staurolite isograd is defined by the reaction garnet + chlorite = staurolite + biotite + H_2O, and appropriate lithologies for this reaction are widely distributed throughout the area of Fig. 4. Although there may be more subtle bulk compositional controls on the temperature and pressure conditions

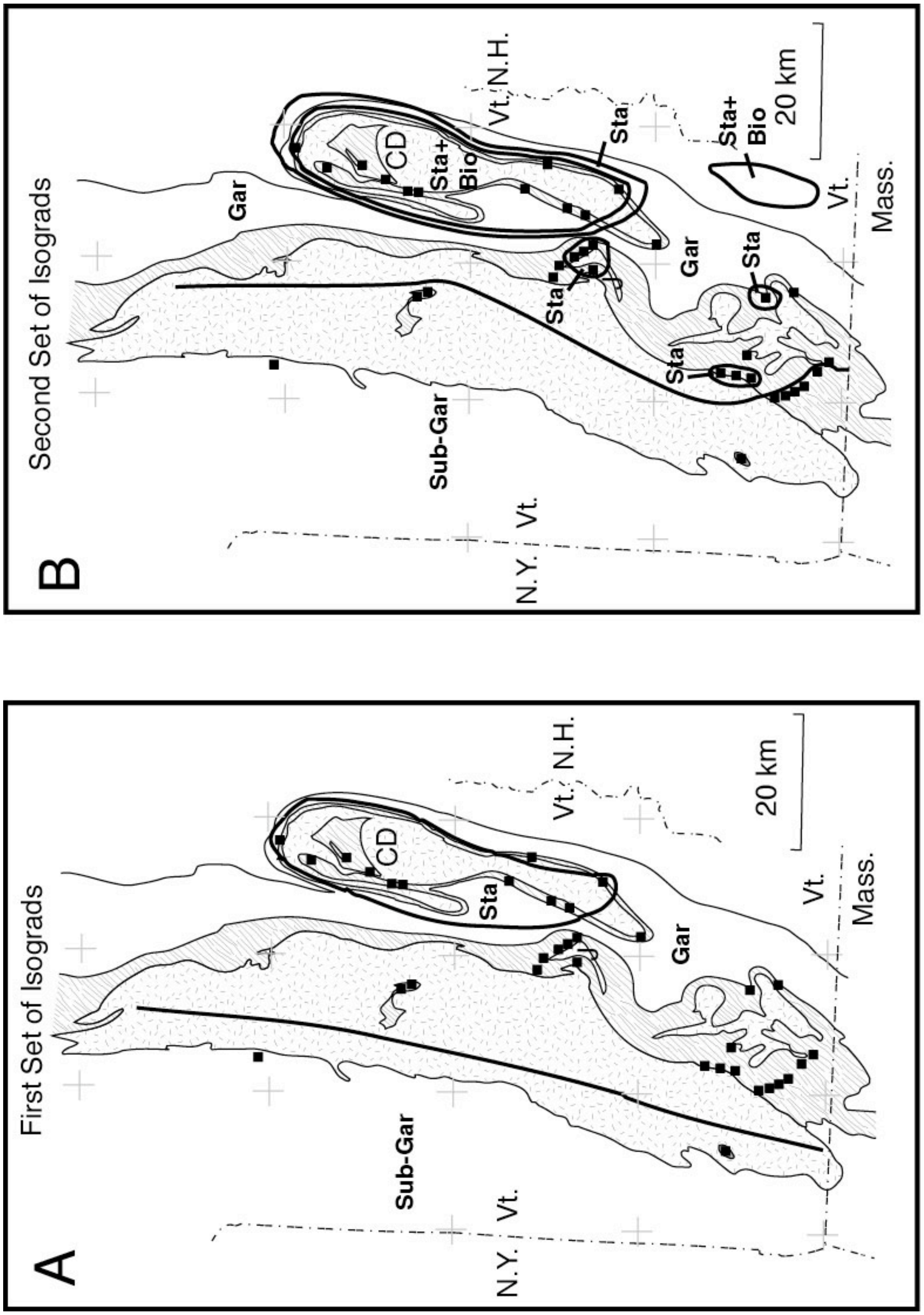

Figure 4. Distribution of two sets of isograds in southeastern Vermont based on Cook and Karabinos (1988). First set of isograds (A) based on inclusions in first stage garnet (gar) cores. Second set of isograds (B) based on matrix assemblages. Sta isograd defined by first appearance of staurolite by the reaction chloritoid = garnet + chlorite + staurolite. Sta + Bio isograd defined by first appearance of biotite with staurolite by the reaction garnet + chlorite = staurolite + biotite. Grenvillian basement-hatchured pattern, Units containing highly aluminous schist in diagonal pattern. CD- Chester dome, J- Jamaica dome. Note the spatial correlation of isograds from the first and second sets and that the higher isograds of the second set are correlated with the basement-cored domes. The Sta+Bio isograd in the southeast corner in B is located over the Guilford dome in Silurian and Devonian units.

at which staurolite is first produced in garnet and chlorite bearing rocks, it is probable that this isograd records real variations in metamorphic conditions.

Garnets with textural unconformities invariably contain rutile in the cores but lack rutile in the rims and chemically altered margins of the cores; rutile is also absent from the matrix (Karabinos, 1984, Cook and Karabinos, 1988). In contrast, ilmenite is concentrated along the altered margins of the cores and widely distributed in the rims and matrix. This indicates that the garnet cores grew under higher pressure conditions than the rims (Bohlen et al., 1983), consistent with the P-T path determined by Vance and Holland (1993), who suggested that garnet from the Gassetts schist grew during 2.5 kbar of decompression. Rocks containing evidence for multiple episodes of garnet growth are relatively common below the mylonite zone around the Chester dome but not above it (Cook and Karabinos, 1988), suggesting that rocks on either side of the mylonite zone experienced different metamorphic histories.

Detailed analysis of samples from the Chester dome is underway to test the hypothesis that rocks on either side of the mylonite zone record different P-T paths. Existing data suggest that this may be the case. Peak metamorphic conditions estimated for rocks in the Chester dome range from approximately 500 to 650 ^{o}C and 8 to 11 kbar (Kohn and Spear, 1990; Vance and Holland, 1993; Kohn and Valley, 1994). To the north, Menard and Spear (1994) showed that rocks in the Strafford dome experienced similar peak pressures of approximately 10 $\pm$ 1 kbar, whereas rocks 7 km to the east near the garnet isograd only reached 5 $\pm$ 1 kbar. Menard and Spear (1994) ascribed this contrast in peak pressure to differential uplift; the close proximity of these rocks could also be due to normal faulting.

"Stratigraphic" Separation Diagram

The most dramatic characteristic of the Chester dome is the zone of extreme ductile thinning or omission of some of the mantling units (Fig. 2). The units shown in black in Figure 2 include rocks mapped as the Pinney Hollow, Ottauquechee, and Stowe Formations (Fig. 3). East of the northern end of the Green Mountain massif the outcrop width of these units is approximately 8 km, and the outcrop width of the same or equivalent units north of the massif increases substantially (Doll et al., 1961). In sharp contrast, the outcrop width of these three units is 150 m around the southern half of the Chester dome. The zone of attenuation is bounded by the Hoosac Formation (below) and Missisquoi Formation (above). Around the northern half of the Chester dome the zone of attenuation includes rocks mapped as the Hoosac and Missisquoi Formations. The outcrop width of all of these units along the northeastern margin of the Green Mountain massif is 19 km, but along the northeastern margin of the Chester dome it is less than 500 m! This dramatic change in outcrop width cannot be accounted for entirely by variations in dip or density of thrust faults. Figure 2 shows that the Pinney Hollow, Ottauquechee, and Stowe Formations are also structurally thinned east of the Jamaica and Wilmington domes, but not as dramatically as around the Chester dome. Taken at face value, the extreme ductile thinning of lithologic units strongly suggests that low-angle normal faults or shear zones extended the crust in this region.

It is important to note that we do not really know the stratigraphic thickness of any of the units in the eastern Vermont sequence because of structural complications. Furthermore, we do not have very good age constraints on these units. A notable exception is a recent 571 $\pm$ 5 Ma U-Pb zircon age on a metafelsite from the Pinney Hollow Formation near the north end of the Green Mountain massif (Walsh and Aleinikoff, 1999). This constrains the Tyson, Hoosac, and Pinney Hollow Formations to be Late Proterozoic, if the stratigraphic continuity of these units is accepted. Another age constraint is that the Moretown Member of the Missisquoi Formation must be older than the 479 $\pm$ 8 Ma Hallockville Pond Gneiss, which cross-cuts the Moretown in western Massachusetts (Karabinos and Williamson, 1994). Because the units formed in disparate tectonic settings and the boundaries between the units are commonly interpreted as thrust faults (e.g. Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985; Ratcliffe et al., 1997), the relative ages of the units are also uncertain. Despite these limitations, it seems reasonable, as a first-order approximation, to compare the thickness and stacking order of the units on the east flank of the Green Mountain massif with the mantling sequence around the Chester dome.

Figure 5 shows schematically which units are affected by ductile thinning and omission in the mylonite zone around the Chester dome. Blank portions of the "column" represent units that are either highly deformed in the mylonite zone or absent. The thinning and omission of units is most extreme along the northern and northeastern margin of the dome. The structural section is most intact along the southern and southwestern margin of the dome. These observations are consistent with recent mapping by Ratcliffe (2000a, 2000b), who showed subdivisions of the Moretown Formation directly on the Mount Holly Complex along the northern and eastern margin of the Chester dome, but interpreted the contact as a thrust fault.

In my opinion, normal faulting or extension along a wide shear zone is a more logical explanation for the observations. Ductile thinning can occur along thrust faults, but it usually affects rocks on the overturned limb of thrust nappes, if at all. There is no evidence for overturned limbs or structural repetition around the Chester dome. Conceivably, there might be some creative explanation for the thinning and omission of units that invokes tectonic erosion during Taconic subduction. Any explanation, however, must also account for the strongly developed mylonitic fabric in the attenuated zone, which is an Acadian feature.

kinematic indicators

working model

ROAD LOG

Mileage

0.0 **MEET AT 8:00 AM IN THE CENTER OF PROCOTRSVILLE** at the intersection of Rt. 131 and Depot St. near the Crows Corner Bakery Café. Drive S on Depot St.

0.5 Intersection of Rt. 103 and Depot St., turn left, S, on Rt. 103.

2.0 Park in pullout on right, W, side of road.

STOP 1. BASEMENT-COVER CONTACT, NW MARGIN OF CHESTER DOME. (45 min.) Gneiss on E side of road, opposite pullout, belongs to the Middle Proterozoic Mount Holly Complex (i.e. basement) and is composed of quartz, plagioclase, biotite, epidote, ± muscovite, ± microcline. It contains abundant highly-deformed pegmatites that distinguish it from plagioclase schist and gneiss of the Late Proterozoic Hoosac Formation. Walk N along road. The first outcrop on the W side of the road belongs to the cover sequence. Before looking at it, head into the woods on the E side of the road to see more of the Mount Holly complex. The contact appears to be buried in the small drainage on the E side of the road. The basement gneisses are highly sheared and the mylonitic fabric dips steeply to the W. Outcrops up the hill contain large-scale C-S fabric and rotated boudins that indicate a tops-to-the-west sense of shear.

The first outcrop on the W side of the road is quartz, muscovite, biotite, chlorite, ± plagioclase, ± garnet schist. Many layers contain graphite. Further N are exposures of quartz-rich gneiss with a distinctive pinstripe texture. I think these rocks belong to the Cram Hill and Moretown Members of the Missisquoi Formation of Doll et al. (1961).

Reverse direction and drive N on Rt. 103.

2.6 Outcrop on E side of road of Barnard Gneiss.

3.5 Turn right, N, on Depot St.

4.0 Turn right, E, on Rt. 131.

6.1 Turn right on Power Plant Rd.

6.3 Park at gate.

STOP 2. CAVENDISH GORGE, AGE OF THE CAVENDISH FORMATIONS. (30 min.) Walk from gate up the dirt road to the power station. One of the tonalite samples dated by Ratcliffe et al. (1997) was collected from the outcrop next to the power station. Ratcliffe et al. (1997) argued that the 1.4 Ga tonalite intruded the metasedimentary units of the Cavendish Formation here in the gorge, the type locality of the Cavendish Schist of Richardson (1929), making them older than the tonalite and part of the Mount Holly Complex. I think that the contact between the tonalite and the marble, calc-silicate, quartzite, and schist is tectonic in origin. I collected a sample of quartzite from the gorge about 50 m upstream from the power station and separated detrital zircons. Single grain Pb-evaporation and SHRIMP ages of 25 grains are all younger than 1.4 Ga (Karabinos et al., 1999), consistent with a tectonic contact between the tonalite and the metasedimentary rocks, but inconsistent with the intrusive contact interpretation. Furthermore, the plagioclase schist here in the gorge does not contain abundant highly-deformed pegmatite like the gneiss we saw in the basement at Stop 1. I correlate the plagioclase-rich schist with the Hoosac Formation and the marble, calc-silicate, and quartzite with the Tyson Formation.

Turn around and drive back to Rt. 131.

6.5 Turn right, E, on Rt. 131.

12.9 Intersection of Rts. 131 and 106. Continue on Rt. 131.

14.2 Turn right on Gulf Rd.

14.5 Intersection of Gulf Rd and Plains Rd. Continue straight.

15.4 Park in pullout on left side of road.

STOP 3. BASEMENT-COVER CONTACT, NE MARGIN OF CHESTER DOME. (45 min.) Outcrop immediately E of pullout is quartz-rich, biotite, chlorite, ± plagioclase schist with graphite in some layers. Further E are more outcrops of this lithology and some layers contain highly fractured garnet crystals partially altered to chlorite along the cracks. These rocks are part of the Silurian to Devonian sequence, probably the Northfield or Waits River Formation. As we will see, the intense zone of shearing along the northeastern margin of the Chester dome extends all the way from Middle Proterozoic basement through the pre-Taconic cover sequence and into the Silurian units.

Walk back down the road to the W. Outcrop beneath roots of fallen tree on S side of road is similar to the first outcrop and contains a fine-grained, unmetamorphosed mafic dike, probably related to Mount Ascutney, a Cretaceous intrusion. Outcrop along road is spotty but it is much better in the woods on both sides of the road. Below the schist of the Northfield Formation are highly deformed felsic and mafic gneisses that contain a mylonitic fabric. It is difficut to assign these rocks to a specific unit with certainty, because they are highly deformed and out of context, but they look more like Barnard Gneiss to me than anything else. This is the only rock type belonging to the Late Proterozoic to Ordovician sequence that I have found along this traverse.

Only 0.3 mile W of the pullout is an outcrop of highly sheared augen gneiss. This rock was mapped as Bull Hill Gneiss by Doll et al. (1961) but not all layers contain microcline, which is typical of Bull Hill Gneiss. In thin section it is clear that the eyes in some layers are composed of plagioclase aggregates. Doll et al. (1961) showed Bull Hill Gneiss directly below Hoosac Formation around much of the Chester dome. It is common to find Bull Hill Gneiss at the basement-cover contact in the southern part of the dome, but other feldspar-rich gneisses belonging to the Mount Holly Complex are more typically found at the contact in the northern part of the dome.

15.6 Intersection of Gulf Rd. and Gravelin Rd. Outcrop on corner contains unmetamorphosed dike. Turn left, N, on Gravelin Rd.

16.0 Turn left, W, on Rt. 131.

19.4 Turn left, S, on Rt. 106.

24.1 Turn right, W, on Rt. 10.

28.4 Turn right, S, on Rt. 103.

28.9 Park in pulloff on right.

STOP 4. GNEISSES IN MOUNT HOLLY COMPLEX WITH ACADIAN DIKES. (20 min.) The tonalitic gneiss seen here is a common lithology in the Mount Holly Complex of the Chester dome. Also, the deformation fabric in this outcrop is typical of that found in the Mount Holly Complex rocks within the core of the dome, away from the basement-cover contact. Late cross-cutting felsic dikes, like those seen here, are found in many outcrops in the dome below the mylonite zone. Ratcliffe et al. (2001) reported a zircon SHRIMP age from a granitic dike near Gassetts of 392 ± 6 Ma.

Continue S on Rt. 103.

32.9 Bear right on Rt. 11 S to connect with Rt. 35 S.

33.1 Go straight across main highway to take Rt. 35 S.

40.1 In village of Grafton turn right.

40.3 Turn left, S, on Townshend Rd.

The narrow valley that the road follows S of Grafton parallels the W flank of the dome near the contact between the Hoosac Formation and the Bull Hill Augen Gneiss.

46.8 Bear right. Back on Rt. 35 S.

50.1 Townshend. Turn left on Rt. 30 S.

51.7 Turn left on Ellen Ware Rd.

51.8 Park in pullover on left.

STOP 5. BASEMENT-COVER CONTACT, SE MARGIN OF CHESTER DOME. (45 min.) Walk back to W to first outcrop; look for signs of blasting dating from construction of the old railway bed parallel to the West River. Rock is quartz, microcline, plagioclase, biotite, epidote gneiss belonging to the Mount Holly Complex. This does not look like typical Bull Hill Gneiss to me, although that is what is shown below the Hoosac Formation by Doll et al. (1961). Walk E along road past pullout. There are some outcrops of plagioclase schist that looks like Hoosac Formation. Continue E along road to where the river is just S of the road. River outcrop of mafic schist with some pelitic schist layers. The pelitic schist contains some large garnet crystals which were incompletely altered to chlorite. The extent of retrogression increases significantly to the E along this traverse. This outcrop et al. further E contain cleavage surfaces with sprays of biotite after amphibole. There are some outcrops along the road to the E of quartz-rich schist with distinctive pinstripe texture. I would assign the mafic and pelitic schist found in the river outcrops to the Moretown Formation. I see no rocks here that I would assign to the Pinney Hollow, Ottauquechee, or Stowe Formations.

Reverse direction and head back to Rt. 30.

51.9 Turn right, N, on Rt. 30.

53.5 Back in the center of Townshend, stay on Rt. 30.

55.1 Covered bridge on left, outcrop of sheared Bull Hill augen gneiss on right. The Bull Hill Gneiss intruded the Mount Holly Complex at approximately 960 Ma (Karabinos and Aleinikoff, 1990), after the Grenville orogeny. Therefore, the deformation fabric in this unit is from the Taconic and Acadian orogenies.

55.2 Park in pullover on left without blocking gate.

STOP 6. TOWNSHEND DAM SPILLWAY, MYLONITE ZONE, SW MARGIN OF CHESTER DOME. (30 min.) Walk around gate and down dirt road past the equipment sheds for Townshend Dam. Just before the road crosses a small and unappealing stream, walk through the reeds to the large outcrop to your right. Here is highly sheared plagioclase schist and gneiss of the Hoosac Formation. Some dark biotite-rich layers are present. Note the extreme isoclinal folds and boudinage in this outcrop.

If it is not too wet, it is possible to walk up the spillway and see rocks on both sides of the cut. Another approach is to walk back to the road, continue across the bridge about 100 m, and head into the pine trees on the right. Look for the end of the chain link fence and then walk around and past it to the N, toward the edge of the spillway using extreme caution- there are some steep cliffs with a lot of vertical relief. From the top of the spillway it is possible to look to the other side and appreciate the extreme deformation of these rocks. There are abundant folds, boudinage, and truncated layers. The rocks on this side of the spillway are made up of interlayered mafic and pelitic schist, with sparys of biotite after amphibole, typical of the Moretown Formation. Between the rocks of the Hoosac Formation just seen and the Moretown Formation, there are some layers of garnet-rich pelitic schist that could be part of the Pinney Hollow Formation. They contain quartz, plagioclase, muscovite, biotite, chlorite, and garnet up to 3 cm in diameter. Some of the garnets have quartz inclusion trails that indicate a counter-clockwise sense of rotation when viewed to the N. This is consistent with other kinematic indicators that show tops to the W. An interesting feature of this section is that the degree of retrogressive alteration of garnet to chlorite is highly variable. Furthermore, some layers contain graphite, usually as inclusions in plagioclase and garnet. I believe that the graphite was precipitated from infiltrating fluids during metamorphism and that the variable retrogression reflects irregular fluid flow, probably during the normal-sense shearing that preceded the doming phase of deformation.

This is the location of garnet dated by Christensen et al. (1989) using $^{87}Sr/^{86}Sr$ variations; they found that the garnet grew at approximately 380 Ma over during a 10 m.y. interval. Thompson et al. (1993) assigned rocks in the spillway to the Pinney Hollow and Ottauquechee Formations. If this assignment is correct, these units are very thin compared to their outcrop widths on the E flank of the Green Mountain massif.

Continue N on Rt. 30.

55.5 Turn left in parking area of Townshend Dam.

STOP 7. TOWNSHEND DAM ROADCUT, MORE OF THE MYLONITE ZONE, SW MARGIN OF CHESTER DOME. (30 min.) Please be very careful looking at this roadcut. It is a dangerous curve and cars drive by very fast. Stay off of the pavement!

The most prominent feature of this long roadcut is the strong mylonitic fabric parallel to the compositional layering. There are abundant folds, boudinage, and rotated garnets. But what impresses me is the extent of disrupted layering throughout the outcrop. The pelitic layers contain quartz, plagioclase, muscovite, biotite, chlorite, and garnet. The mafic layers contain quartz, plagioclase, hornblende, muscovite, biotite, garnet, and epidote.

Rosenfeld et al. (1988) called these rocks Ottauquechee Formation. I would prefer to assign them to the Moretown Formation. The rocks on the W side of the roadcut show the distinctive pinstripe texture of the Moretown Formation. Rocks in the middle and eastern end of the roadcut are admittedly more difficult to assign, but look like Moretown Formation to me. Whatever units these intermediate rocks belong to, there is not much distance between rocks belonging to the Hoosac Formation to the E and rocks that are clearly part of the Moretown Formation to the W. The typical sequence seen on the E flank of the Green Mountain massif, only several kilometers to the W, is highly attenuated.

Outcrops W of here along Rt. 30 are mapped as Mortown Formation and show the distinctive pinstripe texture. They do not have the strong mylonitic fabric seen in this roadcut, however.

END OF TRIP.

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