THE CHARACTERIZATION OF DUCTILE DEFORMATION IN THE UPPER AND LOWER PLATES OF THE HINESBURG THURST FAULT THROUGH DETAILED GEOMETRIC ANALYSIS OF SELECTED OUTCROPS

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Abstract

Two well-known, east-dipping thrust faults that formed during the Ordovician Taconic Orogeny divide the bedrock geology of Vermont’s Champlain Valley into lithotectonic slices. On the west, the Champlain Thrust placed Middle Cambrian–Middle Ordovician sedimentary rocks over Late Cambrian–Middle Ordovician sedimentary rocks. Farther east, the Hinesburg Thrust placed Late Proterozoic–Early Cambrian low-grade metamorphic rocks over rocks of the upper plate of the Champlain Thrust. Subsequent deformation during the Devonian Acadian Orogeny resulted in the folding of both faults and other Taconic structures. This study documents deformation at the meso- and microscopic scales in the upper and lower plates of the Hinesburg Thrust and compares and contrasts deformation in both the upper and lower plates and documents along and across-strike variations.

Outcrops from the Champlain Valley belt and Green Mountains belt were grouped into domains that coincide primarily with lithology and spatial/structural relation to the Hinesburg Thrust. Three generations of folds are observed in the study area and correlated to regional structures that are attributed to orogenic events: tight to isoclinals F1 folds that developed pre- and syn-thrust propagation (Taconic), open to closed asymmetric F3 folds that deform the Hinesburg Thrust (Acadian), and open, asymmetric F4 folds that deform F3 (Acadian). Each fold has an associated axial planar cleavage (S1, S3 and S4, respectively). A west-dipping (S2) cleavage is documented in other studies though no foliations that can be correlated to S2 are documented in this study. F2 folds are also not found in the current study area of the Hinesburg Thrust. In other studies, the second generation of folding deforms S2 and all older fabrics and is, thus, referred to as F3. The F3 and F4 folds have associated crenulation lineations (L3 and L4, respectively) that were used to infer the angular relationship between the two fold sets and, in some cases, infer the presence of F4. The angular relationship between the F3 and F4 fold sets in the study area is consistently orthogonal (85–100°) both along strike of the Hinesburg Thrust and across it.

Ductile structures in the upper plate of the Hinesburg Thrust record all three folding events. The Hinesburg Thrust footwall anticline exhibits F1 and F3 structures; the youngest F4 folds can only be inferred from changes in the plunge of F3 fold axes or L3 from one outcrop to another. Ductile fabrics were only observed locally in the carbonate and siliciclastic sequence in the lower plate of the Hinesburg Thrust; however, some fracture sets within these units show a remarkable similarity in orientation to that of cleavages associated with F3 and F4. Acadian structures are observed in the upper and lower plates of the Hinesburg Thrust.
Dedication

I would like to dedicate this thesis to my loving husband, Nathan, who put his own interests on hold so that I may attend this graduate program. I owe much of my success to his understanding and support—financially, mentally and emotionally. Anam amháin.
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I would first and foremost like to thank my advisor, Dr. Laura Webb, for her guidance, patience and support throughout this process. Laura’s door was always open whenever I ran into trouble or had questions about my research or writing. She allowed this to be my own work, and yet was there to steer me back on course when necessary.

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Chapter 1: Literature Review

Introduction

Vermont bedrock records a complex tectonic history involving multiple orogenic events. Geologic mapping has been incredibly useful as a base for further studies, however spatially, it can be too coarse to establish a chronological order of structures and correlate fabrics across an area. Mapping may be complemented by more detailed work. Detailed analysis of selected outcrops integrated with existing geologic maps allows one to look closely at the deformation style and history as recorded at a single location and understand how it changes over an area.

Previous studies have established the deformation style and history of the Hinesburg Thrust through analysis of fault zone structures and large-scale trends (Tauvers 1982; Gillespie, 1975; Agnew, 1977; Dorsey et al., 1983; Strehle and Stanley, 1986; Kim et al., 2007; Kim et al., 2011). This study aims to build on previous work and understand deformation variation beyond the fault zone in both the upper and lower plates of the Hinesburg Thrust, as well as along-strike variations. The motivating questions are: 1) what is the geometric variation of folds across and along strike of the Hinesburg Thrust? And, 2) are inferred-Acadian structures, based on correlations with known Acadian structures, recorded in both the upper and lower plates of the fault?

Field data were collected from forty six outcrops (stations) within Chittenden County, Vt. during the 2010 and 2011 field seasons. The area lies between the latitudes of 44°38'38"N and 44°16'14"N, and longitudes 73°13'19"W and 73°01'11"W and
includes parts of the towns of Burlington, Colchester, Milton, Westford, Essex, Williston, Shelburne, Charlotte, Richmond, St. George and Hinesburg (Figure 1). The coordinates for each field station are listed in Appendix A along with the names of collected samples, where applicable. Outcrops were selected because of their proximity to previously mapped structures, to represent various lithologies involved in deformation, and to represent structures observed both above and below the Hinesburg Thrust. For each field station, later grouped into domains, outcrop and microstructural analysis of structural fabrics associated with folding and faulting was conducted. These analyses allow the author to synthesize the data into schematic diagrams representing structural domains in the area that illustrate the geometry of folds and associated fabrics. Results are also correlated with fabrics documented in previous studies to build on the existing data.

**Tectonic History**

Over one billion years ago, all landmasses on Earth came together to form the supercontinent of Rodinia (van Staal et al., 1998). A majority of land was located in the southern hemisphere, including Laurentia, the core of present-day North America. Rodinia began to breakup approximately 600–550 Ma (Figure 2a) and Laurentia rifted away and moving toward the equator (van Staal et al., 1998; Hibbard et al., 2007). The southern edge of Laurentia (currently the eastern edge) became a passive continental margin as the Iapetus Ocean opened. The geometry of the continental margin was controlled by rifting and transform segments that resulted in abrupt changes in topography along the edge of the margin. This geometry can be seen by promontories
and reentrants inherited from the margin during the breakup (Figure 3). Late
Proterozoic–Early Cambrian rift clastic and transition rocks were deposited on the rifted
margin of Laurentia and in rift basins (e.g. Allen et al., 2009; Stanley and Ratcliffe, 1985;
Doolan, 1996). Cambrian–Ordovician carbonate and clastic sedimentary rocks were
deposited on top of rift sequence material and mark the development of a passive
continental margin much like the present day eastern edge of North America (e.g.
Mehrtens and Borre, 1989).

Prior to the formation in the Permian of the supercontinent, Pangaea, a number of
landmasses accreted to the present day eastern edge of Laurentia (present-day
coordinates). These rocks deformed during the Paleozoic during a series of major
collisions that have been grouped into three events: the Taconic Orogeny, the Acadian
Orogeny and the Alleghanian Orogeny. The product of these collisions is the
Appalachian Orogen: a northeast trending mountain belt of Late Precambrian to
Paleozoic rocks that runs along the eastern edge of North America from Newfoundland to
Alabama. The orogen is divided into the northern Appalachians and southern
Appalachian with the New York promontory serving as the boundary between the two
(Hibbard et al., 2007). The tectonic history is briefly discussed in order of orogenies;
special attention is paid to the northern Appalachians.

The west–northwest extent of the belt is composed of the Laurentian continental
margin. From the Early Cambrian to the Early to Middle Ordovician, the east-facing
passive margin experienced continuous deposition of stable shelf sequence deposits. This
sequence was later deformed in response to collision with the island arc terranes (Figure 2b; Rowley and Kidd, 1981; Bradley, 1983; Stanley and Ratcliffe, 1985; Karabinos et al., 1998; van Staal et al., 1998; Schoonmaker and Kidd, 2009; van Staal et al., 2009). The Shelburne Falls arc formed above an east-dipping subduction zone at 485–470 Ma and collided with the present-day eastern edge of Laurentia at 475–470 Ma (Karabinos et al., 1998); this is thought to be the Taconic Orogeny in New England. Supra-subduction zone ophiolites were obducted onto the Laurentian margin synchronously with this collision (van Staal et al., 1998). In northern Vermont, the ultramafics and associated garnet amphibolites at Belvidere Mountain represent the vestige of an ophiolite (e.g. Doolan et al., 1982; Gale, 1982; 1986; Laird et al., 1984; 1993). According to Karabinos et al. (1998), this collision was followed by a reversal in subduction polarity and the Bronson Hill Arc was built above this new subduction zone. According to van Staal et al. (2009) the Bronson Hill Arc and correlatives were accreted at ~450 Ma. In contrast, Stanley and Ratcliffe (1985) and Ratcliffe et al. (1998) proposed that the Taconic Orogeny was the result of the collision of a composite Bronson Hill Arc, which included the elements of the Shelburne Fall Arc, with the Laurentian Margin.

Following accretion of the Taconic arc(s), a new basin developed to the present-day east; sedimentation into this basin began in the Late Silurian and continued into the Devonian (Bradley, 1983). Closure of the basin during the Devonian led to the Acadian Orogeny. A large landmass, Avalonia, collided with the present-day eastern edge of composite Laurentia beginning in the north (southern Newfoundland) and progressing
southward into Massachusetts (Figure 2c; van Staal et al., 2009). Avalonia is made up of fault-bounded arc- and backarc-related volcanic-sedimentary terranes (van Staal et al., 2009) that rifted away from Gondwana (van Staal et al., 1998). Avalonia subducted to the west or northwest beneath composite Laurentia, marking the onset of collision.

Siluro–Devonian volcanism shut off at the time of the Acadian Orogeny (Bradley, 1983). The most recent estimate for the timing of the orogeny is 421–400 Ma (van Staal et al., 2009). This range is based on the timing of arc and backarc magmatism, tectonic slivers of subduction-related metamorphism and tectonic loading of Avalonia. Other age ranges have been suggested. Bradley et al. (2000) puts the start of the orogeny at 390 Ma and Ratcliffe et al. (2001) puts the end at 365 Ma or later. In Vermont, the intensity of the Acadian orogeny diminishes to the north (Stanley and Ratcliffe, 1985).

It is more likely that deformation was more or less continuous throughout the Paleozoic rather than the occurrence of three distinct tectonic events (Hibbard et al., 2007). Nevertheless, structural fabrics and deformation styles are traditionally associated with specific pulses of deformation in order to facilitate correlation between rocks throughout the belt and establish the relative chronology of deformation.

As a result of the continuous tectonic activity that formed the Appalachian Orogen, Vermont is made up of a series of north–northeast trending bedrock belts (Figure 4; Doll et al., 1961; Stanley and Ratcliffe, 1985; Kim et al., 2011 and others). The farthest west is the Champlain Valley belt, sometimes known as the Champlain Lowlands (Doolan, 1996); which is defined by the carbonate and siliciclastic rocks deposited on the
Laurentian shelf and upper slope (Kim et al., 2011). To the east of the Champlain Valley is the Green Mountains belt; which is defined by rift- and transitional rift-related metasedimentary and meta-igneous rocks. The current study area includes a portion of each belt.

The geology of the Champlain Valley has been studied extensively for the past 100 or so years (e.g. Cady 1945; Shaw 1958; Welby, 1961; Dorsey et al., 1983; Stanley and Ratcliffe, 1985; Mehrten and Dorsey, 1987; Mehrten et al., 1987; Mehrten and Borre, 1989 Doll et al., 1961; Stone and Dennis 1964; Zen 1961, 1972; Rodgers, 1969). Initial investigations in northwest Vermont were mainly large-scale, reconnaissance work (Agnew, 1977). Progressively, previously mapped areas became more detailed; the data remained unchanged, but the interpretations have evolved over time.

**Summary of Lithologies and Depositional and Tectonic Settings**

The stratigraphy in the area records continuous deposition along the Laurentian margin from Late Precambrian to Early Ordovician (Figure 5; Dorsey et al, 1983). Lithologies in the area can be grouped into three depositional settings: rift clastic and transition sequences, stable shelf sequences and eastern basinal sequence, and foreland basin flysch deposits.

The major lithologies in the rift–transition sequence are coarse-grained alluvial fan clastics of the Pinnacle Formation (Touavers, 1982) overlain by finer-grained siliciclastic sediments that make up the Fairfield Pond Formation. The Pinnacle Formation is a coarse-grained, poorly sorted metawacke (low grade metamorphosed
arenites) with scattered slate and siltstone clasts (Cherichetti et al., 1998). It overlies normally faulted Precambrian crystalline basement (Dorsey et al., 1983). The Fairfield Pond Formation is a siliceous phyllite with quartz-arenite interbeds (Cherichetti et al., 1998).

The Pinnacle Formation and Fairfield Pond Formation are described in detail by Agnew (1977), Carter (1979), Tauvers (1982), and Cherichetti et al. (1998). Synrift siliciclastic alluvial-fan deposits of the Fairfield Pond Formation (Cherichetti et al., 1998) in the New England rift zone began prior to the latest Neoproterozoic–Early Cambrian postrift siliciclastic and carbonate deposits overlay them (Allen et al., 2010). This unit marks the transition from Pinnacle rift clastics to the transgressive sequence of the Cheshire Quartzite (Dorsey et al., 1983).

The lithologies of the stable shelf sequence consist of alternating massive layers of siliciclastic and carbonate sediments (Stanley and Ratcliffe, 1985; Mehrtens et al., 1987, Mehrtens and Dorsey, 1987; Mehrtens and Borre, 1989 The sequence, from oldest to youngest, consists of the Cheshire Quartzite, Dunham Dolostone, Monkton Quartzite, Winooski Dolostone, Danby Formation and Clarendon Springs Formation. The Cheshire Quartzite is at the base of the stable shelf sequence or drift stage (Cherichetti et al., 1998). It is divided into two members of variable thickness: the lower member is an argillaceous arkose; the upper member is a quartz arenite. A detailed study of the eight Cheshire Quartzite lithofacies was completed by Myrow (1983). The Dunham Dolostone is a pink to buff dolomite approximately 400 meters thick. The four lithofacies are
described in detail by Gregory (1982) and Gregory and Mehrtens (1983). The Monkton Quartzite is a 300 meter thick buff to red quartz arenite and dolomite (Mehrtens and Borre, 1989). Lithofacies are summarized by Rahmainian (1981). The Winooski Dolostone is a white to grey dolostone approximately 300 meters thick. The Danby Formation is 35–80 meter thick and consists of a siliciclastic basal unit and upper carbonate unit (Mehrtens et al., 1987). Lithofacies are summarized by Butler (1986). The Clarendon Springs Formation is a recrystallized dolostone (Mehrtens and Dorsey, 1987).

The sequence, as a whole, represents beach, near-shore, and off-shore sandstones and siltstones and shallow marine carbonates (Stanley and Ratcliffe, 1985) and is recognized as the margin of the Cambro–Ordovician platform in eastern North America (Rodgers, 1968) with the Cheshire Quartzite marking the transition from rift basin sediments to stable platform sediments. Continental margin sediments conformably overlay rift clastic rocks in the east (Dorsey et al., 1983) and unconformably overlay Grenville age basement rock in the west (Rowley and Kidd, 1981). The contact between rift, rift–transition and platform rocks is gradational suggesting continuous deposition from the Late Proterozoic to the Late Cambrian (Mehrtens and Borre, 1989). The Cambro–Ordovician platform is bordered to the east by deep marine basin deposits (i.e. Skeels Corner Formation). Deposition of platform material from the west lasted from Lower Cambrian to Lower Ordovician time (Dorsey et al., 1983; Mehrtens and Borre, 1989).
The Skeels Corner Formation is part of the eastern basinal sequences. Other units
make up the sequence but are not exposed in the field area and will not be discussed (see
Mehrtens and Dorsey, 1987 and references therein; Mehrtens and Borre, 1989 and
references therein). The Skeels Corner Formation is a black, calcareous shale with
parallel laminae of limonitic quartz silt and dolomite (Mehrtens and Dorsey, 1987). It
overlies the Dunham Dolostone and Clarendon Spring Formation at different locations
and has an age range from Lower Middle Cambrian to Lower Ordovician (Dorsey et al.,
1983; Mehrtens and Dorsey, 1987).

The Stony Point Formation and Iberville Shale Formation are in the lower plate of
the Champlain Thrust. These units are foreland basin flysch deposits and are comprised
of interbedded black shales and argillaceous limestone (Mehrtens and Borre, 1989) and
studies were conducted by Hawley (1957) and Teetsel (1984).

Stratigraphic facies changes from north to south record a shelf–basin transition
and directly reflects large-scale promontories and reentrants along the continental margin
(Allen et al., 2009; Rodgers, 1968; Mehrtens, 1985; observed by Shaw, 1958). A change
in trend of Precambrian Grenville basement from north to northwest caused a period of
Upper Cambrian shelf instability. More recent studies show along strike variations in
thickness in the synrift and postrift stratigraphy which reflects upper-plate, lower-plate
and transform segments of the margin (Mehrtens, 1985; Mehrtens and Dorsey, 1987;
Cherichetti et al., 1998; Allen et al., 2010).
Sedimentation and stratigraphic studies (e.g. Shaw, 1958; Rowley and Kidd, 1981; Dorsey et al., 1983; Mehrtens and Dorsey, 1987; Mehrtens and Borre, 1989) suggest that the strata reflect a change in depositional environment from fluvial deposits typical of a passive continental margin to progressively finer deep marine muds, particularly flysch deposits. The internal contacts of this succession become younger to the west, indicating a westward progradation of flysch over the carbonate platform sequence (Rowley and Kidd, 1981). These rocks (i.e. Iberville Shale and Stony Point Shale) are only found in the lower plate of the Champlain Thrust in this study area.

**Regional Structures**

Major structural features of western Vermont include folds and faults and generally trend or strike north-northeast (Figure 1; Gillespie, 1975; Stanley 1980; Dorsey et al., 1983; Kim and Thompson, 2001; Kim et al., 2007; Gale et al., 2009) with the exception of an east–west fracture set that is also prominent (Gale et al., 2009). These structures consist of: the north–south trending Hinesburg Synclinorium; the Champlain Thrust; the Hinesburg Thrust; west-verging recumbent folds; north–south trending, asymmetric, upright folds; and east–west trending, open folds.

The Hinesburg Synclinorium (Figure 1) is a foreland fold trending north-northeast through much of the Champlain Valley (Cady, 1969). To the south, this fold plunges southward and is referred to as the Middlebury Synclinorium; to the north, it plunges northward it is referred to as the St. Albans Synclinorium (Gillespie, 1975). It is bounded by the Champlain Thrust to the west and the Muddy Brook and Hinesburg Thrusts to the
east (Shaw, 1958). The Cutting Dolostone is the highest stratigraphic unit of the synclinorium with in the field area (Ratcliffe et al., 2011). Strata defining the regional structures rest uncomfortably on Precambrian basement rock.

Early Taconic F₁ folds are upright and open to recumbent isoclinal (Dorsey et al., 1983; Stanley and Ratcliffe, 1985). Within the Hinesburg Thrust zone, F₁ is observed as a large scale recumbent nappe (Dorsey et al., 1983) and plunges east–southeast (Gillespie, 1975). Traces of F₁ folds to the west change plunge to northeast and the hinge rotates toward the northwest as it approaches the Champlain Thrust fault (Stanley, 1987). The axial surfaces of related F₁ folds dip to the east (Cady, 1969). A regional spaced–closely spaced, penetrative S₁ schistosity (Cady, 1969; Gillespie, 1975; Tauvers, 1982; Dorsey et al, 1983; Stanley and Ratcliffe, 1985; Strehle and Stanley, 1986) is subparallel to the axial surface and limbs of these folds where isoclinal. In the Champlain Thrust zone, a pressure solution cleavage is parallel to F₁ axial planes (Stanley, 1987; Strehle and Stanley, 1986). The cleavage rotated counterclockwise to the west during out-of-sequence westward propagation (Schoonmaker and Kidd, 2009). Gillespie (1975) and Dorsey et al. (1983) attributed these structures to the early stage of Taconic deformation preceding westward thrust displacement. These structures are observed to the south in Lincoln, Vermont (Tauvers, 1982; Strehle, 1986) and to the north in Milton, Vermont (Carter, 1979).
The older, pre-thrust, folds are refolded by syn-thrust recumbent folds. An axial planar schistosity parallels thrust surfaces (Stanley and Ratcliffe, 1985). These folds are observed to the southeast of the current study area.

Folding and displacement within the Hinesburg Thrust zone were once attributed to gravity slides of Middle Ordovician age throughout Quebec and Vermont (Zen, 1972). The gravity slide hypothesis has since changed and earlier structures in the region are now attributed to shearing out of the lower limb of an overturned fold (Tauvers, 1982; Gillespie, 1975; Dorsey et al., 1983; Strehle and Stanley, 1986). Westward displacement of carbonate shelf rocks was attributed to the Oak Hill Thrust extending from the Vermont–Quebec border southward to the Winooski River where it is covered by another, separate thrust surface (Clark, 1934; Cady, 1945). Both faults were considered tear faults in which limbs of the westward verging Hinesburg Synclinorium detached and thrusted westward. Booth (1950) later split the Oak Hill Thrust into the Brigham Hill Thrust east of Georgia Mountain and Arrowhead Thrust to the west. Booth (1950) mapped the Brigham Hill Thrust east of the Brigham Hill area and below the Hinesburg Thrust and the Arrowhead Thrust continuing northward with the Fairfield Pond Thrust. Today the Brigham Hill Thrust is mapped west of Brigham Hill. Shaw (1958) mapped a series of east dipping thrusts in the west and puts the Fairfield Pond Thrust lying east of Clark’s (1934) Oak Hill Thrust. From east to west, thrusts expose successively younger units. Fairfield Pond and Oak Hill Thrusts displace the Lower Cambrian to Middle Cambrian section over Upper Cambrian and Middle Ordovician units.
Stone and Dennis (1964) reinterpreted Shaw’s (1958) finding as the differences in stratigraphy to be a facies change due to interfingering of carbonate and pelitic terrains and eliminating the Brigham Hill Thrust from their mapping. Instead they divide the Gilman Formation into the Fairfield Pond Formation and Cheshire Formation. They said the easternmost patches of the Dunham Dolostone may be considered the core of a south-plunging syncline (Dead Creek Syncline). They map the Cheshire Quartzite continuously between Brigham Hill and Bald Hill and suggest a sliver of Cheshire was thrust over Dunham just north of Brigham Hill — defining the easternmost thrust of the region. This is the Hinesburg Thrust. Stone and Dennis (1964) suggest the Hinesburg Thrust, Champlain Thrust and Muddy Brook Thrust all developed as a stress release mechanism on the overturned limb of the Hinesburg Synclinorium and Georgia Mountain anticline. Dorsey et al. (1983) included the Arrowhead Mountain Thrust in with this group.

The two east-dipping thrust faults, the Champlain Thrust and the Hinesburg Thrust, formed during the Ordovician Taconic Orogeny and divide the bedrock geology of Vermont’s Champlain Valley into lithotectonic slices. On the west, the Champlain Thrust placed Middle Cambrian–Middle Ordovician shelf sequence rocks over highly deformed Middle Ordovician shales with a total displacement of 60–80 km (Stanley, 1987). It extends from Rosenberg, Canada—where it is referred to as the Rosenberg Thrust (Séjourné and Malo, 2007)—to Cornwall, Vermont (Stanley and Sarkisian,
and is located in the lowest dolostone of the carbonate siliciclastic platform sequence throughout its extent (Stanley, 1987).

Farther east, the Hinesburg Thrust placed Late Proterozoic– Early Cambrian chlorite–sericite grade metamorphosed rift clastic rocks over rocks above the Champlain Thrust (Strehle and Stanley, 1985). The Hinesburg Thrust represents the sheared out forelimb of F₁ fold (Dorsey et al., 1983). It extends from Fairfield (Ratcliffe et al., 2011) to Lincoln, Vermont and dies out as a shear zone along the overturned western limb of the Lincoln anticline (Tauvers, 1982; DiPietro, 1983). Zen (1972) concluded that the sedimentary record in western Vermont record late Middle Ordovician age for movement along the Hinesburg Thrust. The presence of a block of Hinesburg upper plate material in the uppermost black shale of the Hathaway Formation of Hawley (1957) supports this finding.

Subsequent deformation during the Devonian Acadian Orogeny resulted in the folding of both faults and other Taconic structures. Previous work (Gillespie, 1975; Dorsey et al., 1983; Kim et al., 2007; Derman et al., 2008; Earle et al., 2010) recognized a set of north–south trending, open–tight, upright folds (F₃). Some of these authors (Kim et al., 2007; Derman et al., 2008; Earle et al, 2010) also recognized an east–west-trending, open fold set (F₄) that, with F₃ folds, creates a dome and basin fold interference pattern (Type 1 of Ramsay, 1962). Derman et al. (2008) observed a fold axis (F₃) plunging to the northeast and southwest, and another fold axis (F₄) plunging to the east and west.
A dome–and–basin interference pattern can be the result of one or multiple deformation phases or events (Ramsey, 1962; Ghosh et al., 1995). For example, Allen et al. (2001) found northwest—southeast trending fold and northeast—southwest trending folds that developed during sinistral transpression and dextral transpression, respectively, along the Karatau Fault System in southern Kazakhstan. The fold hinge and axial planes of these folds are orthogonal to each other, resulting in the classic dome and basin pattern. A refolded fold does not necessarily indicate two distinct deformation events. It is possible for simultaneous development of folds in more than one direction due to differential, regional or local, flattening (Ramsey, 1962 and references therein). This was recognized by Ghosh et al. (1995) via experiments on folding by constriction. They found that originally horizontal, competent layers formed gentle domes and basin that become progressively elongated and irregular with progressive shortening. The data collected as part of this study will contribute to the larger regional database required to resolve the two models.

**Microstructural analysis**

*Foliation Morphology*

In the field, it is important to identify planar and linear fabrics in the rock and measure their orientation for the purpose of quantifying the geometry of structures and correlating fabrics between outcrops based on deformation style. These fabrics are foliations and lineations and are described in terms of their morphological characteristics and what defines them. Microstructural analysis adds to the information collected in the
field by aiding identification of what defines these fabrics on a microscale and revealing evidence regarding crosscutting relationships, etc., that may not be obvious in outcrop.

Foliations are divided into primary foliation, such as sedimentary bedding, and secondary foliation, the products of deformation, and refer to any planar feature that occurs throughout a rock (Passchier and Trouw, 2005). Primary foliation is directly related to the original formation of the rock. Many of the sedimentary rocks in the study area have been deformed so primary foliation is destroyed. The earliest foliation observed is often attributed to the earliest phase of deformation.

Secondary foliation develops after lithification in response to deformation and metamorphism (Passchier and Trouw, 2005 and references therein). The classification of secondary foliation is based on morphology (Powell, 1979; Borradaile et al., 1982) and does not necessarily pinpoint the mechanism involved in its formation, thus the two are discussed separately. Note the terms cleavage and schistosity are used in describing secondary foliations and are defined in Passchier and Trouw (2005) by a preferred orientation of inequant fabric elements in fine-grained and medium-coarse grained rocks, respectively.

The first criterion of describing secondary foliations is whether or not microlithons (layers with a small degree of preferred orientation compared to cleavage domains) are present. Continuous foliations do not have microlithons; there is an even distribution of typically platy mineral grains with a preferred orientation. The foliation can be further described by the grain size (whether or not one can see individual grains
without magnification) of the mineral defining it: a fine-grained continuous foliation is a *continuous or slaty cleavage*; a coarse grained continuous foliation is a *continuous schistosity*. Foliations with microlithons present are referred to as spaced foliation and can also be further described according to grain size: a fine-grained spaced foliation is a *spaced cleavage*; a coarse-grained spaced foliation is a *spaced schistosity*. The initial clay content in a rock controls the cleavage domain spacing (Van der Pluijm and Marshak, 2004). High clay content results in closely spaced domains; however, spacing can change with progressive deformation. If crenulations (microfolds) are observed within the microlithons, the foliation is a *crenation cleavage*. The morphological variations in a spaced foliation require an additional set of criteria that describe the relationship between cleavage domains and microlithons.

**Foliation Development**

As mentioned earlier, secondary foliation develops in response to deformation or metamorphism in the rock. Many processes have been identified (Passchier and Trouw, 2005 and references therein); mechanical rotation of grains, solution transfer, microfolding are briefly discusses here; recrystallization is discussed in the deformation mechanism section.

Mechanical rotation of tabular of elongate grains causes a set of randomly oriented grains to line up parallel to each other and perpendicular to the shortening direction during homogenous ductile deformation. Solution transfer and pressure solution involves material along grain boundaries moving from a high to low strain zone.
This process leaves behind insoluble material, which can be identified by planes of opaque or micaceous material. Existing foliations can fold to form crenulation cleavage. Initially the foliation is defined by the limbs of these folds lining up; solution transfer and crystal-plastic deformation are responsible for foliation development as deformation continues. Microfolding is a dominant fabric in certain domains of this study and will be discussed in detail below.

**Crenulation Cleavage Formation**

Foliation that is shortened in a direction parallel to or at a low angle to the XY plane (strain ellipse) will crinkle and produce symmetric or asymmetric microfolds. This shortening may also result in the development of a new foliation: crenulation cleavage (Van der Pluijm and Marshak, 2004). The term axial-planar crenulation cleavage is used if it can be demonstrated that the crenulation cleavage is parallel or subparallel to the axial plane of the microfolds.

All crenulation cleavages are formed by buckling behavior of anisotropic material and/or pressure solution (Cosgrove, 1976; Gray and Durney, 1979; Worley et al., 1997). At low grade metamorphic conditions, pressure solution transfer is the dominant mechanism (Gray and Durney, 1979; Worley et al., 1997). It is necessary that a rock have an older foliation that can be deformed for crenulations to form. This can be mica-rich bedding or pre-existing continuous cleavage. Many of the rocks in the study area are dominated by quartz, fine-grained mica and/or clay. As a result, pressure solution and grain rotation is an important component to crenulation formation (Mamtani et al., 1999).
Quartz migrates from the limbs of microfolds to the hinge causing the limbs to become enriched in platy minerals with a preferred orientation. As deformation progresses, the older foliation can be obliterated and the rock will appear to have a single foliation that simulates compositional layering on a microscale. (Van der Pluijm and Marshak, 2004)

The later stages of crenulation deformation operate under similar deformation mechanisms as those during mylonitization (Mamtani et al., 1999). In fact, microstructures resembling the fabric found in mylonites can be observed in rocks with asymmetric crenulations. The similarities can be seen in rocks where intra-crystalline crystal–plastic deformation is the dominant mechanism; which is the case in the later stages of development (Mamtani et al., 1999). As mentioned before, alternating domains of quartz and mica result from solution transfer. As deformation progresses, shearing can occur along the surfaces defining the crenulation cleavage (Williams and Schoneveld, 1981); which produces sigmoidal curving of the preexisting (and now folded) foliation. This fabric resembles S–C fabrics found in mylonites and is due to intracrystalline crystal–plastic deformation (Mamtani et al., 1999). This information allows one to describe the relative levels of strain from one sample to another.
Figure 1. Bedrock belts (modified from Doll et al., 1961) and geologic map of study area (modified from Ratcliffe et al., 2011). Note the Hinesburg Thrust separates the Champlain Valley belt from the Green Mountains belt.
Figure 2 Paleogeographic maps summarizing the tectonic evolution of Laurentia from the Proterozoic to Silurian. a) Rodinia breaks apart during the Late Proterozoic. Laurentia drifts northward. b) An island arc collides with Laurentia during the Ordovician Taconic Orogeny. c) Avalonia collides with composite Laurentia beginning in the Early Silurian during the Acadian Orogeny. (Blakey, 2003; Paleographic maps are from Ron Blakey, Colorado Plateau Geosystems, Inc. http://cpgeosystems.com/paleomaps.html).
Figure 3 A block diagram of the eastern Laurentian rifted continental margin (Vermont outlined in bold). The diagram shows low-angle detachment faults that control the along-strike variation in the thickness of rift-clastic deposits. (Figure modified from Allen et al. (2009).)
Figure 4 State map showing the bedrock belts of Vermont. Box approximately outlines the current study area (after Kim et al., 2011; modified from Doll et al., 1961).
Figure 5 Stratigraphic column for units in the study area. Figure is modified from Gale et al. (2009), and utilizes the GSA (1999) Geologic Time Scale and unconformities reported by Landing (2007) and Landing et al. (2003). See Figure 1 for legend (*denotes unit not included in Figure 1).
Chapter 2: The Characterization of Ductile Deformation in the Upper and Lower Plates of the Hinesburg Thrust through Detailed Geometric Analysis of Selected Outcrops

Abstract

Two well-known, east-dipping thrust faults that formed during the Ordovician Taconic Orogeny divide the bedrock geology of Vermont’s Champlain Valley into lithotectonic slices. On the west, the Champlain Thrust placed Middle Cambrian–Middle Ordovician sedimentary rocks over Late Cambrian–Middle Ordovician sedimentary rocks. Farther east, the Hinesburg Thrust placed Late Proterozoic–Early Cambrian low-grade metamorphic rocks over rocks of the upper plate of the Champlain Thrust. Subsequent deformation during the Devonian Acadian Orogeny resulted in the folding of both faults and other Taconic structures. This study documents deformation at the meso- and microscopic scales in the upper and lower plates of the Hinesburg Thrust and compares and contrasts deformation in both the upper and lower plates and documents along and across-strike variations.

Outcrops from the Champlain Valley belt and Green Mountains belt were grouped into domains that coincide primarily with lithology and spatial/structural relation to the Hinesburg Thrust. Three generations of folds are observed in the study area and correlated to regional structures that are attributed to orogenic events: tight to isoclinals F₁ folds that developed pre- and syn-thrust propagation (Taconic), open to closed asymmetric F₃ folds that deform the Hinesburg Thrust (Acadian), and open, asymmetric F₄ folds that deform F₃ (Acadian). Each fold has an associated axial planar cleavage (S₁, S₃ and S₄, respectively). A west-dipping (S₂) cleavage is documented in other studies though no foliations that can be correlated to S₂ are documented in this study. F₂ folds are also not found in the current study area of the Hinesburg Thrust. In other studies, the second generation of folding deforms S₂ and all older fabrics and is, thus, referred to as F₃. The F₃ and F₄ folds have associated crenulation lineations (L₃ and L₄, respectively) that were used to infer the angular relationship between the two fold sets and, in some cases, infer the presence of F₄. The angular relationship between the F₃ and F₄ fold sets in the study area is consistently orthogonal (85–100°) both along strike of the Hinesburg Thrust and across it.

Ductile structures in the upper plate of the Hinesburg Thrust record all three folding events. The Hinesburg Thrust footwall anticline exhibits F₁ and F₃ structures; the youngest F₄ folds can only be inferred from changes in the plunge of F₁ fold axes or L₃ from one outcrop to another. Ductile fabrics were only observed locally in the carbonate and siliciclastic sequence in the lower plate of the Hinesburg Thrust; however, some fracture sets within these units show a remarkable similarity in orientation to that of cleavages associated with F₃ and F₄. Acadian structures are observed in the upper and lower plates of the Hinesburg Thrust.
Introduction

Vermont bedrock records a complex tectonic history involving multiple orogenic events. The Champlain and Hinesburg Thrusts are well-known structures that formed during the Early–Late Ordovician Taconic Orogeny (e.g. Stanley and Ratcliffe, 1985) and were reactivated and deformed, respectively, during the Devonian Acadian Orogeny (e.g. Stanley and Sarkisian, 1972; Dorsey et al., 1983). While the large-scale structures are well-mapped regionally, detailed studies of deformation at the meso- and microscope scales are still necessary to define structural geometries and relative ages, make regional correlations, understand structural variability between lithotectonic slices, and to further resolve Taconic deformation from Acadian overprints.

Previous studies have established the deformation style and history of the Hinesburg Thrust through analysis of fault zone structures and large scale structural trends (Tauvers 1982; Gillespie, 1975; Agnew, 1977; Dorsey et al., 1983; Strehle and Stanley, 1986; Kim et al., 2007; Kim et al., 2011). This study aims to build on previous work and compare and contrast deformation in both the upper the lower plate of the Hinesburg Thrust, as well as along-strike variations. The motivating questions are: 1) what is the geometric variation of structural fabrics, and associated foliation and lineations, across and along strike of the Hinesburg Thrust? And, 2) are inferred-Acadian structures, based on correlations with known Acadian structures, recorded in both the upper and lower plate of the fault?

This paper does not provide a detailed analysis of the structures observed in the Champlain Thrust fault zone; however it is referred to because of its large scale implications for the tectonic history of the region and reactivation during multiple
deformation events. Some documentation on the structures observed in the Arrowhead Mountain fault zone are provided because of the spatial relationship between it and the Hinesburg Thrust and to serve as a comparison between the two.

Geologic/Tectonic Background

Bedrock Belts

The study area is located in northwest Vermont and includes portions of two north–northeast trending lithotectonic belts: the Green Mountains and the Champlain Valley (Figure 6). The Green Mountain belt is composed of Late Proterozoic–Early Cambrian rift- and transitional rift-related rocks of the Pinnacle and Fairfield Pond Formation unconformably overlying normal-faulted Proterozoic Grenville basement (ca. 1.4–1.0 Ga; Aleinikoff et al., 2011) (Agnew, 1977; Carter, 1979; Tauvers, 1982; Dorsey et al., 1983; Thompson and Thompson, 2003; Cherichetti et al., 1998; Allen et al., 2010). Based on U-Pb zircon ages on metavolcanic units, the age of rifting in Vermont and southern Quebec ranges from 571–554 Ma (Kumarapeli et al., 1989; Walsh and Aleinikoff, 1999). Immature synrift siliciclastic alluvial-fan deposition began in the Neoproterozoic and transitioned to postrift siliciclastic and carbonate deposits (Allen et al., 2010). These rocks were deformed and metamorphosed during the Taconic and Acadian Orogenies (Rowley and Kidd, 1981; Dorsey et al., 1983; Stanley and Ratcliffe, 1985; Karabinos et al., 1998; Thompson and Thompson, 2003).

The Champlain Valley is subdivided into the upper plate of the Champlain Thrust to the east and the parautochthon to the west. The upper plate of the Champlain Thrust is composed of Lower Cambrian–Middle Ordovician carbonate and siliciclastic rocks
deposited on the eastern Laurentian margin. These rocks represent a stable shelf environment and grade into deep marine basin deposits further east (i.e. Skeels Corner Formation) (e.g. Dorsey et al., 1983; Stanley and Ratcliffe, 1985; Mehrtens and Borre, 1989). Carbonate and siliciclastic sediments conformably overlie rift clastic rocks in the east (Dorsey et al., 1983) and uncomformably overlay Grenville age basement rock in the west (Rowley and Kidd, 1981). The Champlain Valley belt was deformed and weakly metamorphosed during the Taconic and Acadian Orogenies (Kim et al., 2011). The parautochthon is composed of Late Ordovician black shales with thin carbonate interlayers. These rocks are interpreted as flysch deposits (Stanley and Ratcliffe, 1985; Rowley, 1982) and were deformed by the Champlain Thrust (Kim et al., 2011).

**Major Regional Structures**

The major structural features of western Vermont are folds and faults that generally trend or strike north–south (Gillespie, 1975). These structures consist of the north–south trending Hinesburg Synclinorium; the Champlain Thrust; the Hinesburg Thrust; west-verging recumbent folds; north–south trending, asymmetric, upright folds; and east–west trending, open folds.

The Hinesburg Synclinorium is separated from the Middlebury and St. Albans synclinoria by cross-anticlines near Monkton and Milton, Vermont, respectively. It is presumed to be a large-scale Taconic structure (e.g. Dorsey et al., 1983) that deforms the carbonate–siliciclastic sequence and deep marine basin deposits. The east limb is mostly covered by the Hinesburg Thrust, but what is exposed is overturned. The west limb is truncated by the Champlain Thrust (Cady, 1945). The youngest stratigraphic unit
exposed in the synclinorium is the Lower Ordovician Bascom Formation (Ratcliffe et al., 2011). The Hinesburg, Middlebury and St. Albans synclinoria are the result of en echelon folding that preceded thrusting (Welby, 1961).

The two east-dipping thrust faults, the Champlain Thrust, on the west side of the field area, and the Hinesburg Thrust to the east, are major faults that formed during the Ordovician Taconic Orogeny and divide the bedrock geology of Vermont’s Champlain Valley into lithotectonic slices with the deepest structural level to the west and the shallowest level to the east. The Hinesburg Thrust represents a nappe-and-thrust system that placed Late Proterozoic–Early Cambrian chlorite–sericite grade metamorphosed rift clastic rocks over Cambrian–Ordovician imbricated carbonate–siliciclastic platform rocks (Dorsey et al., 1983; Strehle and Stanley, 1985). The Hinesburg Thrust fault formed from the sheared out limb of a recumbent fold (Gillespie, 1975; Dorsey, 1983). It extends from Fairfield (Ratcliffe et al., 2011) to Lincoln, Vermont and dies out as a shear zone along the overturned western limb of the Lincoln anticline (Taufers, 1982; DiPietro, 1983). The backlimb of this large-scale overturned antiform is sheared out from Essex, Vermont to Lincoln; the forelimb is sheared out to the north of Essex (Dorsey et al., 1983). The Hinesburg Thrust exhibits predominantly ductile deformation as opposed to the brittle fault zone fabrics documented at the Arrowhead Mountain and Champlain Thrusts (Dorsey et al., 1983). The fault contains abundant mylonites and rocks in the upper plate have undergone chlorite/sericite to biotite grade metamorphism (Taufers, 1982; Dorsey, 1983; Strehle and Stanley, 1986) indicating it developed at deeper levels in the crust than the Arrowhead Mountain or Champlain thrusts. The surface trace of the Hinesburg Thrust is the boundary between the Champlain Valley belt and the Green
Mountain belt and thus marks a metamorphic contrast between the two belts. Stanley and Wright (1997) estimated 6.4 km of displacement along the Hinesburg Thrust; Dorsey et al. (1983) estimated 8 to 10 km of displacement.

In the west, the Champlain Thrust is a classic foreland thrust that placed Middle Cambrian–Middle Ordovician shelf sequence rocks over highly deformed Middle Ordovician shales with a total displacement of 60–80 km (Stanley, 1987). It extends from Rosenberg, Canada to Cornwall, Vermont (Stanley and Sarkisian, 1972; Stanley, 1987; Séjourné and Malo, 2007). Ratcliffe et al. (2011), however, have extended the Champlain Thrust well into New York State.

The Arrowhead Thrust represents a fold-and-thrust system and is defined by the sheared out lower limb of the overturned Arrowhead Mountain Anticline (Dorsey et al., 1983). Brittle deformation dominates the fabrics in the fault zone (Dorsey et al., 1983). The surface trace of the Arrowhead Mountain Thrust is restricted to the western flank of Arrowhead Mountain in Milton, Vermont.

Early Taconic $F_1$ folds in the study area are upright and open to recumbent and isoclinal (Dorsey et al., 1983; Stanley and Ratcliffe, 1985). Within the Hinesburg Thrust zone, $F_1$ is observed as a large scale recumbent nappe (Gillespie, 1975; Dorsey et al., 1983) or tight, asymmetric, isoclinal folds. Traces of $F_1$ folds change plunge to northeast and hinges rotate toward the northwest as the Champlain Thrust fault is approached from the east (Stanley, 1987). The axial surfaces of related $F_1$ folds dip to the east (Cady, 1969). A regional spaced–closely spaced, penetrative $S_1$ schistosity (Cady, 1969; Gillespie, 1975; Tauvers, 1982; Dorsey et al, 1983; Stanley and Ratcliffe, 1985; Strehle and Stanley, 1986) is subparallel to the axial surface. Within the Champlain and
Hinesburg fault zones, S_1 can be a cleavage that shows offset parallel to the cleavage plane (Stanley and Sarkisian, 1972; Strehle and Stanley, 1986; Stanley, 1987), a pressure solution cleavage (Carter, 1979; Tauvers, 1982; Dorsey et al., 1983; Strehle and Stanley, 1986; Stanley, 1987) or a mylonitic foliation that parallels F_1 axial planes (Strehle and Stanley, 1985; Kim et al., 2011). Gillespie (1975) and Dorsey et al. (1983) attributed these structures to the early pulse of Taconic deformation preceding westward thrust displacement. The cleavage later rotated counterclockwise during out-of-sequence westward propagation of the thrust system (Schoonmaker and Kidd, 2009).

The older, pre-thrust folds are refolded by syn-thrust recumbent F_2 folds during the second pulse of Taconic deformation (Stanley and Sarkisian, 1972; Gillespie, 1975; Dorsey et al., 1983). They are observed within the Champlain Thrust fault zone, but not in the current study area. An associated axial planar schistosity (S_2) is parallel to the thrust surfaces (Stanley and Ratcliffe, 1985; Strehle and Stanley, 1986).

The structures that deform all older fabrics in the field area are attributed to Acadian deformation based on the across-strike correlation with known Devonian structures in the Green Mountain Anticlinorium (e.g. Kim et al., 2011). The anticlinorium was formed and tightened during the Acadian Orogeny to produce open, upright (F_3) folds. These folds have an associated (S_3) spaced cleavage parallel to axial planes (Thompson and Thompson, 2003). The F_3 folds deform S_2, making them post-Taconic structures. Furthermore, the chlorite has partially replaced garnet is attributed to an Acadian overprint (Thompson and Thompson, 2003 and references therein). The Champlain and Hinesburg Thrusts and other Taconic structures in the current study area are folded. The most recent studies have attributed the two youngest generations of folds
(F₃ and F₄) to post-Taconic deformation (D₃ and younger) and infer the deformation to be Acadian (Kim et al., 2007; Derman et al. 2008; Gale et al., 2009; Earle et al., 2010). They are a set of north–south trending, open–tight, upright F₃ folds (Gillespie, 1975; Dorsey et al., 1983; Strehle and Stanley, 1986; Kim, 2007; Derman, 2008; Gale et al., 2009; Earle, 2010; Kim et al., 2011) and a set of east–west trending, open F₄ folds (Dorsey et al, 1983; Kim, 2007; Derman et al., 2008; Gale et al., 2009; Earle et al., 2010, Kim et al., 2011). Both sets have an associated crenulation or disjunctive axial-planar cleavage (e.g. Kim et al., 2007).

Based on an integrated analysis of groundwater well logs (Becker et al., 2008) and geologic mapping of folds, Kim et al. (2007) and Derman et al. (2008) found that north–south trending folds (F₃) were responsible for highly variable depths to the Hinesburg Thrust in Williston, Vermont. They also found variations in the direction of plunge of these folds and attributed it to a superposed east–west trending fold set (F₄). The two sets of orthogonal folds create a Type 1 dome and basin interference pattern (Ramsey, 1962; Grasemann et al., 2004) that is documented in the upper plate of the Hinesburg Thrust in Williston, Vermont (Kim et al., 2007) and the lower plate of Champlain Thrust in Charlotte, Vermont (Earle et al., 2009). The term interference pattern implies that polydeformed terranes control outcrop patterns (Ramsey, 1962). This paper will use the term refold structure (Grasemann, et al., 2004) instead to reflect the fact that the focus is on the geometric relationship between multiple generations of folds and their associated foliations and lineations, and outcrop patterns are not considered.

Methods
Field Data Collection

Outcrops were selected because of their proximity to previously mapped structures (e.g. inferred-Acadian deformation and refold structures) to represent various lithologies involved in deformation and to represent structures observed both above and below the Hinesburg Thrust. Outcrops are grouped into domains (Figure 7) based primarily on spatial/structural relationship to the Hinesburg Thrust, which coincides with lithologic contrasts. Within each domain, outcrops are discussed in order of importance to illustrate key structural relationships.

Field methods included identifying lithologies, measuring fabrics and establishing relative ages, documenting kinematic indicators and collecting oriented samples for petrographic analysis. The latest Vermont state bedrock map (Ratcliffe et al., 2011) is used as a basemap for this study. Lower hemisphere equal area projections of all fabrics are constructed for individual outcrops.

Thin Sections

Thin section work focused on microstructural analysis, specifically crosscutting relationships and characterizing microscopic fabrics, to constrain the sequence of deformation and document associated metamorphic mineral assemblages. Thin sections from fifteen oriented samples from the lower and upper plates of the Hinesburg Thrust were prepared. Billets were cut from samples in pairs: one billet parallel to lineation and perpendicular to foliation; a second cut perpendicular to lineation and perpendicular to foliation. Samples cut perpendicular to foliation and parallel to the mineral stretching lineation are suitable for determining shear sense (Passchier & Trouw, 2005). Samples
that do not exhibit these features were cut twice as well: one billet perpendicular to the dominant foliation and parallel to the strike; a second cut perpendicular to the foliation and parallel to dip direction. Mineral assemblages, microstructures and deformation mechanisms are not discussed where thin sections were not prepared from an outcrop.

Nomenclature

Nomenclature used to identify fabrics in the field is based on the relative age relationships at individual outcrops. Fabrics are identified according to their association with other structures. For example, $F_{n+1}$ is the earliest fold observed at an outcrop where it is observed to fold $S_n$. If the oldest fold is observed to fold $S_{n+1}$, then $F_{n+2}$ is used. Similarly, a crenulation cleavage and its associated crenulation lineation are both identified as the same generation (i.e. $S_{n+3}$ and $L_{n+3}$) even if the lineation is the only lineation observed. Fabrics within each domain are summarized before any correlations are made. Fabrics are first correlated within domains (Table 1–4) and then correlated from one domain to the next (Error! Reference source not found.) to establish a sequence and description of structures documented in this study. Correlations to regional structures identified in previous studies are made once the results of this study have been established (Table 6).

Data and Observations

Domain A

Domain A is defined as the overturned forelimb on the upper plate of Hinesburg Thrust. Early Taconic structures are not observed at all field sites within this domain, whereas the youngest set of structures are. All outcrops discussed in this domain are
located close to the Hinesburg Thrust fault, the surface trace of which is to the west (Figure 7). Table 1 summarizes the fabric elements observed in Domain A.

**10WLO9**

The mapped surface trace of the Hinesburg Thrust curves around this outcrop from the north, around to the west, and continues due south. The outcrop is composed of fine-grained, light gray, chloritic phyllite of the Fairfield Pond Formation.

Compositional layering ($S_n$) consists of 2–4 mm thick layers of alternating metapsammites (quartz, muscovite, sericite and small amounts of chlorite) and phyllites (clay and sericite) with wiggly cleavage domains (Figure 9). Layering strikes north–northwest and dips 25–45º east. It is folded by open to tight, gently inclined to recumbent, west-verging, disharmonic $F_{n+1}$ folds (Figure 8). The fold hinge doubly plunges to the north and south. An $S_{n+1}$ axial planar cleavage is not observed.

Asymmetric crenulation cleavage trends north–south and deforms all earlier fabrics. The associated evenly spaced crenulation cleavage ($S_{n+2}$) shows a gradual transition between predominantly parallel cleavage domains and microlithons (Figure 9); it strikes approximately north and dips variably to the east and is selectively developed in phyllitic material. An $L_{n+2}$ crenulation lineation, expressed on the $S_n$ surface, gently plunges to the north and south (Figure 8). The asymmetry of microfolds is recognized by the relative sericite enrichment in the western limb.

Weakly developed crenulations rotate mica grains to produce symmetric microfolds that are also selectively developed in phyllitic layers. A spaced $S_{n+3}$ crenulation cleavage is defined by the north flanks of the symmetric microfolds, makes
up 10% of the volume and shows a gradual transition between it and microlithons (Figure 9). It strikes west–northwest and dips steeply to the north. Conjugate cleavage domains are also present, but not common. The associated closely spaced crenulation lineation ($L_{n+3}$) is well developed, also expressed on the $S_n$ surface, and plunges to the east (Figure 8b).

A $S_{n+2}$ crenulation cleavage the strikes north–south is observed at the outcrop scale, but east–west striking $S_{n+3}$ crenulation cleavage is not. An $L_{n+2}$ crenulation lineation plunges to the north and south as a result of east–west trending $F_{n+3}$ folds. They are strongly expressed on the folded surface of phyllitic compositional layers (Figure 8). East–west trending microfolds ($F_{n+3}$) are also strongly expressed on the folded surface of $S_n$. The orthogonal relationship between the two sets of upright folds ($F_{n+2}$ and $F_{n+3}$) produces a Type 1 refold structure (Ramsey, 1962; Grasemann et al., 2004). This pattern is subtle, but apparent in the irregular surface of phyllitic compositional layers.

**11ES01**

The outcrop is composed of fine-grained, light gray, chloritic phyllite of the Fairfield Pond Formation. Compositional $S_n$ layering is defined by alternating metapsammite (recrystallized quartz, muscovite, sericite, and small amount of chlorite) and phyllite (clays and sericite) layers with anastomosing cleavage domains (Figure 11). Bulging recrystallization of quartz is present in layers defining $S_n$. It strikes north and dips moderately to the east and is folded by tight, overturned, moderately-inclined, disharmonic $F_{n+1}$ folds (Figure 10). The axial surface strikes north–northwest and dips west (Figure 11). A zonal, spaced $S_{n+1}$ cleavage is defined by wiggly cleavage domains
of biotite and parallels the axial plane of $F_{n+1}$ (Figure 11). This foliation is preferentially developed in phyllitic layers and has an anastomosing relationship with compositional layering. The transition from cleavage domains to microlithons, which is the layering, is discrete.

All earlier fabrics are deformed by north–south trending, open, steeply inclined, asymmetric $F_{n+2}$ crenulations. An associated spaced, axial planar $S_{n+2}$ crenulation cleavage has parallel cleavage domains defined by insolubles between limbs of microfolds and strikes north–northeast and dips steeply east (Figure 11). An associated $L_{n+2}$ crenulation lineation is expressed on the surface of $S_{n+1}$ and plunges very gently to the northwest and southeast.

A spaced $S_{n+3}$ crenulation cleavage and associated $L_{n+3}$ crenulation lineation crosscut the fabrics associated with $F_{n+1}$. $S_{n+3}$ strikes southeast and dips steeply southwest; $L_{n+3}$, also expressed on $S_{n+1}$ surface, plunges variably to the east–southeast (Figure 11).

The approximately orthogonal relationship between the two sets of upright folds, $F_{n+2}$ and $F_{n+3}$, produces a Type 1 interference pattern. This pattern is subtle, but apparent in the irregular surface of phyllitic compositional layers. (Figure 12)

Biotite in the phyllitic cleavage domains of $S_{n+1}$ indicates low grade metamorphic conditions were reached during the formation of $S_{n+1}$. This is supported by bulging recrystallization observed in quartz grains in compositional layering, which indicates low grade (~300–400°C) conditions (Passchier and Trouw, 2005 and references therein). If $S_{n+1}$ is truly an axial planar cleavage to $F_{n+1}$, then the conditions were reached during the development of the $F_{n+1}$ folds. Microfolds are observed on the limbs of the overturned
$F_{n+1}$ fold are $F_{n+2}$ crenulations and an associated north–south striking $S_{n+2}$ crenulation cleavage makes an acute angle with $S_{n+1}$.

**11CLPD**

Five outcrops along the west and north side of Colchester Pond are grouped together. These outcrops are on the upper plate of the Hinesburg Thrust and the upper plate of the north-striking St. George normal fault. These outcrops are composed of beige massive quartzite and light brown to gray, argillaceous quartzite of the Cheshire Formation.

Thin compositional layering ($S_n$) is defined by alternating layers of fine-grained quartz, feldspar and sericite with fine-grained micas, clay and minor amount of quartz. Quartz-rich layers show preferred orientation of quartz grains and sericite parallel to layering; quartz exhibits undulatory extinction and bulging recrystallization (Figure 13). In hand sample, the rock appears to have a slaty cleavage. The poles to $S_n$ define a girdle of $F_{n+1}$, in thin section $F_{n+1}$ can been seen as recumbent isoclinals folds, gently plunging southwest (Figure 13).

$S_n$ is also deformed by steeply inclined, west-verging $F_{n+2}$ folds that plunge gently to the north and south. Folds are open in quartz-rich layers and tighter in argillaceous layers. A closely spaced $S_{n+2}$ crenulation cleavage is defined by insolubles and the preferred orientation of micas; it strikes north–northeast and dips gently to steeply east–southeast. An associated $L_{n+2}$ crenulation lineation is expressed in the surface of $S_n$; it plunges very gently to the south (Figure 15). Cleavage domains are parallel to each other and there is a gradual transition between cleavage domains and microlithons.
All features are deformed by east-west trending, open to closed, steeply inclined \( F_{n+3} \) folds. The axial planes strike east–west and dip steeply to the north; fold axes plunge moderately to the west (Figure 14). In thin section a weakly developed \( S_{n+3} \) foliation is defined by insolubles and dips to the north (Figure 13). Pinch-and-swell structures are observed in quartz-rich layers (Figure 13).

Bulging recrystallization in quartz grains indicates low-grade conditions (~300–400°C); which is consistent with inferred conditions elsewhere in Domain A. (Stipp et al., 2002; Passchier and Trouw, 2005 and references therein).

10WL11

The outcrop is up to 1 km east of the Hinesburg Thrust fault line, but the thrust is truncated by the north–south striking St. George normal fault. The outcrop is composed of fine-grained, light gray, chloritic phyllite of the Fairfield Pond Formation.

Thin compositional layering \( (S_n) \) is composed of slate and metapsammite. These layers are folded by northeast-southwest trending, moderately inclined, west-verging asymmetric disharmonic folds \( (F_{n+1}) \) (Figure 16). Parallel to \( S_n \) is a spaced slaty cleavage making \( S_n \) a composite foliation. A measured fold axis plunges steeply to the northeast; its axial plane strikes north–northeast and dips moderately southeast. The more competent metapsammite layers of \( S_n \) exhibit microfaults, with apparent top-to-the-west motion, instead of folding.

Spaced crenulation cleavage \( (S_{n+2}) \) has parallel cleavage domains and strikes north and dips steeply to the east. An associated crenulation lineation \( (L_{n+2}) \) plunges to the
south–southwest. A second set of crenulation lineations L_{n+1} plunge to the east and west (Figure 16).

The east–west trending L_{n+2} crenulation lineation plunges to the east and west as a result of north–south trending F_{n+2} folds (Figure 177). This indicates the formation of L_{n+1} before F_{n+2} folds. This is older in relative age than what is observed to the north along strike of the Hinesburg Thrust (10WL09, 11ES01) where east–west folds are the youngest feature. Thus the nomenclature differs from other outcrops within Domain A. Additionally, the angular relationship between the two sets of lineations is consistent with that observed to the north at 10WL09, the axis orientation is rotated clockwise approximately 10°.

Competent layers can behave brittly as seen in microfaults in the plane of S_n (Figure 177). Less competent layers appear to behave ductily as seen in asymmetric boudinage. The apparent sinistral sense of shear is top-to-the-west–southwest based on sigmoidal porphyroclast within S_n.

**IIWF01**

The rocks are close to the hinge of the Georgia Mountain anticline (Dorsey et al., 1983 and earlier) that trends northeast–southwest. The outcrop is composed of gray, metagreywacke of the Pinnacle Formation.

A bedding-parallel S_{n+1} spaced schistosity is vertical and strikes north-south (Figure 18). Thin section analyses of samples from this location reveal anastomosing cleavage domains (200 μm wide) defined by preferred orientation of micas (muscovite, biotite and small amounts of chlorite) and grain shapes in domains of partially
recrystallized quartz, feldspar, and micas (Figure 19). Detrital subhedral quartz and feldspar grains (100–200μm wide) are in a preferred orientation in the microlithon in a fine-grained equigranular matrix of micas. Shearing is also observed within microlithons (Figure 19).

A later disjunctive (S_{n+2}) cleavage, subparallel to S_{n+1}, strikes 010° and dips steeply to the east (Figure 18). It is defined by the preferred orientation of sericite. There is a conjugate spatial relationship between the two cleavage domains and the L_{n+2} intersection lineation plunges moderately to the south. Folding is not observed at this outcrop.

**11ES03**

The outcrop is composed of gray, metagreywacke of the Pinnacle Formation. Two oblique foliations dominate at this outcrop. This first foliation is a spaced schistosity (S_{n}) striking about 190° and dipping steeply to the west. The second is a disjunctive cleavage (S_{n+1}) which strikes due south and dips steeply to the west (Figure 20). Folding is not observed at this outcrop.

**Summary**

A schematic block diagram (Figure 21) illustrates how domes and basins are observed in the Fairfield Pond Formation but not in the Pinnacle Formation. The two outcrops of the Pinnacle Formation (11WF01, 11ES03) do not exhibit the youngest sets of folds and associated crenulation cleavages observed in the Fairfield Pond Formation (10WL09, 11ES02, 11CLPD, 10WL11). The oldest generation of folds is also tighter in Fairfield Pond than the Pinnacle Formation. A summary of fabrics are in Table 1.
Table 1 Correlation of fabric elements in Domain A. Note the two outcrops of the Pinnacle Formation (11WF01, 11ES03) do not exhibit the youngest sets of folds and associated crenulation cleavages observed in the Fairfield Pond Formation (10WL09, 11ES02, 11CLPD, 10WL11).

<table>
<thead>
<tr>
<th></th>
<th>10WL09</th>
<th>11ES02</th>
<th>11CLPD</th>
<th>10WL11</th>
<th>11WF01</th>
<th>11ES03</th>
</tr>
</thead>
<tbody>
<tr>
<td>Compositional layering (S₀)/ Bedding (S₀)</td>
<td>Sₙ</td>
<td>Sₙ</td>
<td>Sₙ</td>
<td>Sₙ</td>
<td>S₀</td>
<td></td>
</tr>
<tr>
<td>Schistosity</td>
<td>Sₙ₊₁</td>
<td></td>
<td>Sₙ₊₁</td>
<td>Sₙ₊₁</td>
<td>Sₙ</td>
<td></td>
</tr>
<tr>
<td>Disjunctive cleavage</td>
<td></td>
<td></td>
<td>Sₙ₊₂</td>
<td></td>
<td>Sₙ₊₁</td>
<td></td>
</tr>
<tr>
<td>Spaced crenulation cleavage N–S</td>
<td>Sₙ₊₂</td>
<td>Sₙ₊₂</td>
<td>Sₙ₊₂</td>
<td>Sₙ₊₂</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spaced crenulation cleavage E–W</td>
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<td>Sₙ₊₃</td>
<td>Sₙ₊₃</td>
<td></td>
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</tr>
</tbody>
</table>

**Domain B**

Domain B includes outcrops in the fine-grained argillaceous rocks of the Cheshire Quartzite lower argillaceous member and the Skeels Corner Formation that partially comprise the lower plate of the Hinesburg Thrust (the Dunham Dolostone, Danby Formation and Clarendon Springs Formation are also in the lower plate of the Hinesburg Thrust, but no outcrops in these units were analyzed as part of this study) (Figure 7). Early Taconic structures are observed at all sites within this domain, however, only one set of later fold structures are observed. All outcrops discussed in this domain are located close to the Hinesburg Thrust fault, the surface trace of which is to the east.

**10WL02**

The lithology is composed of black slates with thin dolostone interbeds of the Skeels Corner Formation. The earliest fabric is a slaty to phyllitic continuous cleavage (Sₙ) with anastomosing cleavage domains. In thin section, Sₙ can be seen to be subparallel with compositional layering (S₀, identified only in thin section) of alternating clays and micas (Figure 24). Layers of quartz and calcite are parallel to Sₙ and much
coarser (1-2 mm in width) than the surrounding material. An associated lineation (Lₙ) is an intersection lineation between S₀ and Sₙ; it plunges gently to moderately west–northwest and east–southeast (Figure 23).

All early fabrics are folded by Fₙ₊₁ folds that plunge gently to the north–south and are steeply inclined to the southeast. Poles to Sₙ define a girdle to Fₙ₊₁; the π-axis of which plunges north–northeast (Figure 233). An axial planar, spaced crenulation cleavage (Sₙ₊₁) dips steeply to the east–southeast. Crenulation cleavage domains are thin and the transition to microlithons is discrete. Conjugate crenulations are observed, but rare. An associated Lₙ₊₁ crenulation lineation plunges gently to the north–northeast. Shear bands are defined by white micas and dips less steeply to the east–southeast than Sₙ₊₁ (Figure 25). Shear bands record apparent top-to-the-northwest motion. The presence of shear bands indicates a period of higher strain relative to other fabrics.

A thin section was prepared perpendicular to the axial plane of Fₙ₊₁ (Figure 25b). Fₙ₊₁ is folded again by tight Fₙ₊₂ crenulations. No associated crenulation cleavage is observed. Apparent offset is also observed along Sₙ₊₁ cleavage planes. This may be due to slip along Sₙ₊₁ cleavage planes or the result of dissolution. All fabrics are deformed by open folds (Fₙ₊₃), as seen only in thin section, which appear to trend east–west. The relative chronology of fabrics without Fₙ₊₃ folds is highlighted in Figure 26.

The lithology is argillaceous quartzite of the Cheshire Quartzite’s lower member at this outcrop. Original bedding (S₀) locally occurs in isoclinal folds that are intrafolial to the earliest generation of cleavage (Sₙ). S₀ is sheared out and locally becomes parallel
to Sn, a pervasive slaty cleavage strikes north–northeast and dips steeply to the east. It is defined by the preferred orientation of very fine-grained elongated quartz and sericite (Figure 27). Quartz grains exhibit undulatory extinction. A spaced Sn+1 spaced crenulation strikes north–northeast and dips moderately to the east. The youngest foliation (Sn+2) is defined by disassociated black insoluble layers; it strikes north and dips moderately to the west (Figure 27). Some insoluble layers appear to be folded. The stretching lineation (Ln) is defined by stretched quartz pebbles and plunges steeply to the south–southeast (see Figure 28 for schematic diagram).

Deformed sand pebbles exhibit extension fractures filled with calcite. Antitaxial fiber veins are composed of calcite and the sense of curvature of the veins indicate top-to-the-northwest motion (Figure 27c).

Undulatory extinction in quartz grains and the absent of dynamic recrystallization indicates low grade temperatures of deformation (<300°C) (Passchier and Trouw, 2005 and references therein). Minor folding of Sn+2 indicates a second generation of folding.

The lithology is composed of black slates with thin dolostone interbeds of the Skeels Corner Formation. Compositional layering Sn composed of phyllite and dolostone is folded by tight to isoclinals Fn+1 folds (Figure 29). A continuous phyllitic Sn+1 cleavage is parallel to Sn and may be axial planar to Fn+1. Sn and Sn+1 strike north and dips moderately to the east.

All early fabrics are folded again by moderately inclined, asymmetric crenulations (Fn+2) (Figure 29). An associated closely-spaced crenulation cleavage (Sn+2) exhibits a
gradual transition between parallel cleavage domains and microlithons. It strikes north and dips moderately to the east. The intersection lineation ($L_{n+2a}$) between $S_n$ and $S_{n+2}$ plunges gently to the south–southeast (Figure 30). An associated $L_{n+2b}$ lineation plunges gently to the south–southeast; it is the intersection between $S_{n+1}$ and $S_{n+2}$ or a crenulation lineation associated with $F_{n+2}$. The youngest fabric ($S_{n+3}$) is a foliation that dips west and is only observed in thin section. Apparent offset parallel to the plane of $S_{n+3}$ displaces $S_n$ and may be due to slip along $S_{n+3}$ cleavage planes or the result of dissolution (Figure 30).

See Figure 31 for a schematic diagram of all fabrics of 11ML02_03.

Table 2 Correlation of fabric elements observed in Domain B. Note that stretched pebble lineations are only observed at 10WL01; crenulation cleavage is seen in the Skeels Corner Formation (10WL02, 11ML02_03), but not in the Cheshire Quartzite (10WL01).

<table>
<thead>
<tr>
<th>Fabric Element</th>
<th>10WL02</th>
<th>10WL01</th>
<th>11ML02_03</th>
</tr>
</thead>
<tbody>
<tr>
<td>Compositional layering ($S_n$)/ Bedding ($S_0$)</td>
<td>$S_0$</td>
<td>$S_0$</td>
<td>$S_n$</td>
</tr>
<tr>
<td>Schistosity/slaty cleavage</td>
<td>$S_n$</td>
<td>$S_n$</td>
<td>$S_{n+1}$</td>
</tr>
<tr>
<td>Spaced crenulation cleavage</td>
<td>$S_{n+1}$</td>
<td>$S_{n+1}$</td>
<td>$S_{n+2}$</td>
</tr>
<tr>
<td>Tight, isoclinal folds</td>
<td>$F_{n+1}$ (closed folds)</td>
<td>$F_n$</td>
<td>$F_{n+1}$</td>
</tr>
<tr>
<td>N–S Crenulations</td>
<td>$F_{n+2}$</td>
<td></td>
<td>$F_{n+2}$</td>
</tr>
<tr>
<td>Stretched pebble lineation</td>
<td>$L_n$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Intersection lineation between two oldest foliations</td>
<td>$L_n$</td>
<td></td>
<td>$L_{n+2a}$</td>
</tr>
<tr>
<td>Crenulation lineation</td>
<td>$L_{n+1}$</td>
<td></td>
<td>$L_{n+2b}$</td>
</tr>
</tbody>
</table>

Summary

A schematic block diagram (Figure 32) of fabric elements in Domain B shows how the oldest generation of folds deforms compositional layering/bedding. From north to south, the second generation of folds plunges down to the northeast then up to the northwest. The stretched pebble lineation only observed at 10WL01; Crenulation cleavage seen in Skeels Corner Formation (10WL02, 11ML02_03), but not Cheshire Quartzite (10WL01). A summary of fabrics documented in Domain B is in Table 2.
**Domain C**

The outcrops in Domain C represent the block between the Hinesburg Thrust to the east and the Arrowhead Thrust to the west (Figure 7). Structures associated with Taconic deformation are observed, however later deformation is not. The Arrowhead Mountain Thrust zone exhibits predominantly brittle deformation in contrast to the predominantly ductile deformation experienced in the Hinesburg Thrust zone.

**11ML10**

This field station is made up of outcrops found throughout a small island on the Lamoille River in Milton, Vermont. The lower plate of the thrust is composed of massive dolostone of the Dunham Dolostone; the upper plate is composed of grey to beige massive quartzite of the Cheshire Quartzite’s upper member. The Arrowhead Mountain Thrust dies out at this locality (Dorsey, et al., 1983; Ratcliffe et al., 2011) where it is cross cut by Cretaceous dikes. These outcrops represent the Arrowhead Mountain Thrust fault zone.

The earliest foliation is a bedding parallel schistosity ($S_0/S_n$). Bedding is defined by alternating layers of fine-grained dolostone (dolomite and fine-grained quartz) and fine-grained dolomitic mudstone (clay, dolomite and fine-grained quartz). The schistosity is defined by preferred orientation of dolomite and lenticular mineral aggregates. Poles to $S_0/S_n$ define a girdle to $F_{n+1}$ (Figure 33). These are tight, isoclinal folds are inclined gently to moderately to the northeast with a hinge, defined by the poles to the girdle that plunges gently to the northwest.
A slaty cleavage ($S_{n+1}$) is selectively developed in argillaceous layers and defined by domains of clay (Figure 33). It strikes north and dips steeply to the east. This foliation has shallower dips in the upper plate (lower member of the Cheshire Quartzite) of the thrust than in the lower plate (Dunham Dolostone); both of which are at this locality. It intersects with $S_0/S_n$ to form an intersection lineation ($L_{n+1}$) that plunges gently to the south (Figure 333).

$S_0/S_n$ represents the limbs and possibly axial planar cleavage, respectively, to an overturned anticline inclined to the east. $S_{n+1}$ has shallower dips in the upper plate of the Arrowhead Mountain Thrust suggesting counterclockwise rotation associated with fault movement (Figure 34).

**11ML04, 11ML05 and 11ML06**

These three outcrops are in very close proximity to each other and will be treated as a single field station. The lithology is grey, phyllitic quartzite of the Cheshire Quartzite’s lower argillaceous member. $S_0/S_n$ composite foliation is composed of transposed layers of quartzite interbedded with mudstone (3–5 mm and 1–2 mm wide, respectively) and a spaced schistosity with anastomosing cleavage domains (1–1.5 cm wide) (Figure 355). $S_0/S_n$ strikes northeast and dips to the southeast becoming shallower from east to west, and mud layers anastomose around quartz-rich clasts. $L_n$ mineral lineations are defined by elongated fragments of quartzite and plunge to the northwest and east–southeast (Figure 35).

The east-dipping schistosity $S_n$ anastomoses around quartz-rich clasts. Bedding-parallel schistosity and the mineral lineations are the only evidence of deformation
among these outcrops. The mineral lineations lie within the plane of $S_0/S_n$ suggesting the lineation and second foliation formed during the same deformation event. Later deformation fabrics are not present.

**11ML09**

The lithology is composed of grey to beige massive quartzite of the Cheshire Quartzite’s upper member. Bedding or compositional layering ($S_0$) is folded by open folds ($F_n$) (Figure 366), which plunge gently to the south–southeast. Spaced cleavage ($S_n$) is parallel to the axial plane of $F_n$. $S_n$ dips less steeply than bedding, which indicates the folds are overturned.

One generation of folding is observed at the outcrop scale. It is part of the larger, overturned Arrowhead Mountain Anticline; which is associated with fold-and-thrust system that produced the Arrowhead Thrust (Dorsey et al., 1983).

**Summary**

Bedding and schistosity are observed through the domain, but the spaced cleavage in documented only within the fault zone (see the schematic block diagram (Figure 37). The first generation of folds is isoclinal and its limbs are often parallel to the axial planar schistosity. Folds in the upper plate of the Arrowhead Mountain Thrust tighten as they approach the thrust. The slaty cleavage is shallower in the upper plate within the fault zone. A summary of fabrics documented in Domain C is in Table 3.
Table 3 Correlation of fabric elements observed in Domain C. Bedding and an a schistosity are observed through the domain, but the spaced cleavage is seen in the fault zone.

<table>
<thead>
<tr>
<th></th>
<th>11ML10</th>
<th>11ML456</th>
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<tr>
<td>Schistosity</td>
<td>$S_n$</td>
<td>$S_n$</td>
<td>$S_n$</td>
</tr>
<tr>
<td>Spaced cleavage</td>
<td>$S_{n+1}$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Domain D**

Domain D is defined as the folded carbonate–siliciclastic sequence between the Champlain Thrust and Hinesburg Thrust (Figure 7). All units are deformed by the Hinesburg Synclinorium, however structures at the outcrop scale are predominantly original bedding and two sets of fracture patterns. The lithologies have experienced little or no metamorphism. Special note will be made when outcrops are close to other thrust faults (i.e. Muddy Brook Thrust, Hinesburg Thrust).

**11CL01**

The lithology is composed of grey to beige beds of quartzite interbedded with sandy dolostone of the Danby Formation. The outcrop is in a fault block bounded by northeast–southwest trending normal faults. Poles to bedding define a girdle to $F_n$ (Figure 38). In the field $F_n$ can be observed as an open, gently inclined folds deforming bedding though a measurable surface was not available. Microfaults (Figure 38) show apparent sinistral offset in bedding and subvertical fractures with an apparent top-to-the-north–northeast sense of motion. This may be associated to normal faulting during the Mesozoic and is not likely related to Taconic or Acadian deformation. The fracture patterns suggest two dominant sets striking north–northwest and east–west (Figure 38).

**11CL10, 11BU01, 11BU02, 10CH03 and 11SB01**
These four outcrops are described together because they exhibit similar structures and are located on the western limb (11CL10, 11BU01, 11BU02, 11SB01) and the hinge zone (11CH03) of the Hinesburg Synclinorium. The lithologies are composed of buff, well-bedded dolostone of the Winooski Dolostone at 11CL10, 11BU01, and a portion of 11SB01, reddish–brown bedding sandstone with interbeds of dolostone of the Monkton Quartzite at 11BU02 and a portion of 11SB01, and grey to beige beds of quartzite interbedded with sandy dolostone of the Danby Formation at 10CH03. Original bedding ($S_0$) is the dominant structure in these rocks and is almost horizontal (Figure 39).Brittle deformation in the form of orthogonal vertical fracture patterns (one north–south striking set; one east–west striking set; Figure 39) are the only later structures.

*11FE01*

The exposure at this locality is a well-known teaching outcrop known as “the Oven” (Kim et al., 2011). The lithology is composed of reddish–brown beds of sandstone interbedded with grey-weathered dolostone. Original bedding ($S_0$) is preserved and composed of quartz, feldspar, micas and clays alternating between very fine-grained and fine-grained. Bedding is folded by open, upright $F_n$ folds (Figure 40). Flexural slip between bedding planes is observed at the outcrop scale (Figure 40). A disjunctive cleavage ($S_n$) is parallel to the axial plane of $F_n$. This is one of the few examples of folding at the outcrop scale in Domain D. Undulatory extinction in quartz indicate very low grade ($<300°C$) conditions (Stipp et al., 2002; Passchier and Trouw, 2005 and references therein).

*10SB02*
This outcrop is close to the St. George fault. The lithology is composed of blue–grey dolomitic limestone of the Bascom Formation. It has one subvertical fracture set that strikes north–south and a pair of conjugate fractures (Figure 41). The angle between the conjugate pair faces approximately east–northeast and west–southwest.

**10HB01**

The lithology is composed grey, massive dolostone of the Clarendon Springs Formation. The outcrop exhibits two sets of fractures that are observed in all other outcrops in this domain. The angular relationship between the sets are approximately orthogonal, however, both sets are rotated clockwise in relation to that seen throughout the domain (Figure 42). This may possibly be attributed to downthrown block rotation along the St. George fault.

Table 4 Correlation of fabric elements in Domain D. Fractures are noted as present or not, but not correlated as generation of foliation.

<table>
<thead>
<tr>
<th></th>
<th>11CL01</th>
<th>11CL10</th>
<th>11BU01</th>
<th>11BU02</th>
<th>10CH03</th>
<th>11FE01</th>
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<tbody>
<tr>
<td>Bedding</td>
<td>$S_0$</td>
<td>$S_0$</td>
<td>$S_0$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Disjunctive cleavage</td>
<td></td>
<td></td>
<td>$S_n$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fracture set N–S</td>
<td>Yes</td>
<td>Yes</td>
<td></td>
<td></td>
<td></td>
<td>Yes</td>
<td>Yes</td>
<td></td>
</tr>
<tr>
<td>Fracture set E–W</td>
<td>Yes</td>
<td>Yes</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Yes</td>
</tr>
<tr>
<td>Conjugate fractures</td>
<td></td>
<td></td>
<td></td>
<td>Yes</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

**Summary**

Bedding, the Hinesburg Synclinorium and two fracture sets are the dominant structures, however, a second generation of folding and its associated axial planar cleavage is observed at 11FE01(see the schematic block diagram (Figure 43)). A summary of fabrics documented in Domain D is in Table 4. Fractures are noted as present or not, but not correlated as generation of foliation.
Discussion

The preceding sections documented the geometry, style, and relative timing of structures recorded in both the upper and lower plates of the Hinesburg Thrust and initial correlations of fabrics within domains were presented. The following discussion aims to 1) correlate fabrics between structural domains and with published studies from the region; and 2) describe the geometric variations, if any, of structures across and along strike of the thrust.

Correlations

Table 5 summarizes the next intermediate set of correlations, those made between Domains A–D. An intermediate set of terminology is also presented in Table 5. The earliest generation of foliation, for example, is identified as “S_i”. The structures associated with a generation of folding are given the same subscript as the fold (i.e. F_{ii}, S_{ii}). The sequence of fabrics documented in this study is presented in Table 5.

Compositional layering and depositional bedding are separately identified as the earliest foliation at all outcrops of this study and, thus, collectively S_i (Table 5). This foliation is correlated with the oldest foliation (S_0) documented by Dorsey et al. (1983) and Strehle and Stanley (1986) (Table 6).

At least one generation of folding is observed in all domains, but not all outcrops. S_i is folded by tight to isoclinal F_{ii} folds; in Domain D, F_{ii} is the Hinesburg Synclinorium. F_{ii} correlates with F_1 identified in all studies referenced in Table 6. In the upper plate and footwall anticline of the Hinesburg Thrust, F_1 are tight, isoclinal folds that become mylonites at the fault (Dorsey et al., 1983; Strehle and Stanley, 1986). A regional
cleavage ($S_{ii}$) is parallel to the axial plane of $F_{ii}$ and forms a low angle to $S_1$ in areas of higher strain where $F_{ii}$ is isoclinal (Table 5). It is characterized as an east-dipping, slaty to phyllitic cleavage in fine-grained argillaceous rocks and schistosity in schist and quartzite. Cleavage domains are often wiggly or anastomosing. Foliation development in coarse-grained material results in anastomosing cleavage domains due to mica grains warping around quartz and feldspar grains (e.g. Figure 19a, Figure 35). In terms of relative age and style, it correlates to $S_1$ of Dorsey et al. (1983), Strehle and Stanley (1986) and Kim et al. (2007) (Error! Reference source not found.). $F_1$ folds deforming and/or compositional layering. This is identified as Taconic deformation (e.g. Dorsey et al., 1983; Stanley and Ratcliffe, 1985; Strehle and Stanley, 1986) and is documented throughout the study area.

Table 5 Correlation of fabrics between domains, excluding fractures.

<table>
<thead>
<tr>
<th></th>
<th>Domain A</th>
<th>Domain B</th>
<th>Domain C</th>
<th>Domain D</th>
</tr>
</thead>
<tbody>
<tr>
<td>$S_1$</td>
<td>Compositional layering</td>
<td>Compositional layering/bedding</td>
<td>Bedding</td>
<td>Bedding</td>
</tr>
<tr>
<td>$S_{ii}$</td>
<td>Schistosity or phyllitic cleavage</td>
<td>Slaty to phyllitic cleavage</td>
<td>Bedding-parallel, spaced schistosity</td>
<td>Disjunctive cleavage</td>
</tr>
<tr>
<td>$S_{iii}$</td>
<td>Crenulation or disjunctive cleavage</td>
<td>Spaced crenulation cleavage</td>
<td>Spaced cleavage</td>
<td></td>
</tr>
<tr>
<td>$S_{iv}$</td>
<td>Crenulation or disjunctive cleavage</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$L_1$</td>
<td>Stretched quartz pebbles</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$L_{ii}$</td>
<td>Intersection lineation between two older foliations</td>
<td>Intersection lineation between composite foliation and spaced cleavage</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$L_{iii}$</td>
<td>N–S crenulation lineation</td>
<td>N–S crenulation lineation</td>
<td></td>
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</tr>
</tbody>
</table>
Gillespie (1975), Agnew (1977) and Dorsey et al. (1983) identify an overturned, east-verging fold and an associated, axial planar cleavage. This fold set deforms bedding ($S_0$) and an early regional schistosity ($S_1$) and is therefore correlated as $F_2$; the axial planar cleavage being $S_2$. Strehle and Stanley (1986) observed asymmetric shear bands indicating west–over–east displacement within the Hinesburg Thrust fault zone. The shear band crosscuts compositional layering ($S_0$) and a pervasive schistosity ($S_1$) and is defined as $S_{1.5}$ (Table 6). However, such folds and shear bands are not observed in the current study area, therefore, no correlation can be made to structures documented in this study.

Taconic structures are deformed by north–south trending upright, open to closed $F_{iii}$ folds (Table 5). A spaced cleavage ($S_{iii}$) is parallel to the axial plane of $F_{iii}$. In domains A and B, a spaced crenulation lineation ($L_{iii}$) is also associated with $F_{iii}$ (Table 5). In all cases, $S_{iii}$ is expressed as either a spaced crenulation cleavage or a disjunctive cleavage. $F_{iii}$ and its axial planar cleavage ($S_{iii}$) correlates to $F_2$ and $S_2$, respectively, in Dorsey et al. (1983) and Strehle and Stanley (1986) and $F_3$ and $S_3$, respectively, in Kim et
al. (2007) (Table 6). Since $S_{iii}$ correlates to the third foliation identified in the literature, it is logical to refer to it as $S_3$ as opposed to $S_2$. To remain consistent in terminology, $F_{iii}$ is correlative to $F_3$ even though it is the second generation of folding observed (Table 6). The associated lineation ($L_{iii}$) is, therefore, correlative to $L_3$ and has been identified by Kim et al. (2011).

Thompson and Thompson (2003) identified $F_3$ folds in the Green Mountain Anticlinorium that deformed Taconic structures and tightened the anticlinorium. The fold set has an associated ($S_3$) spaced cleavage that is parallel to its axial plane (Thompson and Thompson, 2003). The retrograde mineral assemblage (fine-grained muscovite and chlorite) that is associated with this foliation (Walsh, 1992) yielded a 363 Ma $^{40}$Ar/$^{39}$Ar age (Lanphere and Albee, 1974). Laird et al. (1984) also reported a $^{40}$Ar/$^{39}$Ar metamorphism age of 376–386 Ma from muscovite and biotite in pelitic schists west of the Green Mountain Anticlinorium. Assuming that $S_3$ is coeval with $F_3$, this fold set and associated structures are Devonian and presumed to have developed during the Acadian Orogeny.

Outcrops in the upper plate and lower plates of the Hinesburg Thrust exhibit an $S_3$ disjunctive cleavage (e.g. 11WF01, $S_{n+2}$; 11ML345, $S_n$), even where $F_3$ is not observed. For example, within Domain A, outcrops of the Fairfield Pond Formation in the upper plate of the Hinesburg Thrust readily exhibit folding (e.g. 11WL09, $F_{n+1}$), whereas, outcrops of the Pinnacle Formation, also in the upper plate, do not (e.g. 11WF01). $S_3$ is axial planar to $F_3$ folds where folding is observed, however, its presence is not dependent
Table 6 Correlation of structures observed in study area with regional structures identified in previous studies and the generation of folding with which they are associated. ¹Kim et al., 2007. ²Dorsey et al., 1983. ³Strehle and Stanley, 1986. ⁴This study.

<table>
<thead>
<tr>
<th></th>
<th>Hinesburg Thrust upper plate ¹</th>
<th>Hinesburg Thrust lower plate ¹</th>
<th>Milton Quadrangle ²</th>
<th>Hinesburg Thrust ³</th>
<th>Hinesburg Thrust upper plate ⁴</th>
<th>Hinesburg Thrust lower plate ⁴</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acadian</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S₄</td>
<td>E–W trending crenulation or fracture cleavage</td>
<td>not reported</td>
<td>not reported</td>
<td>not reported</td>
<td>E–W trending crenulation cleavage</td>
<td>not reported</td>
</tr>
<tr>
<td>S₃</td>
<td>N–S trending crenulation or fracture cleavage</td>
<td>wispy, black, insoluble-bearing cleavage</td>
<td>S₂ N–S trending slip or widely-spaced fracture cleavage</td>
<td>S₂ poorly-developed axial surface schistosity to &quot;younger&quot; folds</td>
<td>N–S trending spaced crenulation or disjunctive cleavage</td>
<td>N–S trending spaced crenulation or disjunctive cleavage</td>
</tr>
<tr>
<td>S₂</td>
<td>new foliation that bounds asymmetric shear bands</td>
<td>spaced cleavage at low angle to S₁</td>
<td>Sₘ pressure solution cleavage</td>
<td>S₁₅ asymmetric shear bands</td>
<td>not reported</td>
<td>not reported</td>
</tr>
<tr>
<td>Taconic</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>S₁</td>
<td>spaced cleavage parallel/ sub-parallel to bedding, mylonites at the Hinesburg Thrust</td>
<td>cleavage containing stretched &quot;pebble&quot; lineation</td>
<td>S₁₅ schistosity axial planar isoclinal and open folds; Sₘ mylonites at the Hinesburg Thrust</td>
<td>S₁ schistosity axial planar isoclinal and open folds; mylonites</td>
<td>continuous slaty to phyllitic cleavage</td>
<td>continuous slaty to phyllitic cleavage</td>
</tr>
<tr>
<td>S₀</td>
<td>transposed bedding in metamorphic rocks</td>
<td>transposed carbonate layers or &quot;pebbles&quot;</td>
<td>compositional layering of phyllites, siltstone, and quartzite</td>
<td>compositional layering severely transposed by S₁</td>
<td>compositional layering of metapsammites and phyllites</td>
<td>compositional layering of metasiltstone and phyllites</td>
</tr>
</tbody>
</table>

on the presence of mesoscopic-scale folding. S₃ is a crenulation cleavage where a preexisting foliation is reworked by microfolding. The schistosity (S₃) and disjunctive cleavage (S₃) observed in the Pinnacle Formation (e.g. 11WF01, Sₙ₊₁ & Sₙ₊₂) are correlated, according to their relative ages, with the S₃ and S₃ in the Fairfield Pond Formation (e.g. 11ES01, Sₙ₊₁ & Sₙ₊₂) (Table 5). It is clear these foliations are associated with generations of folds in the phyllite of the Fairfield Pond Formation. The quartzite of
the Pinnacle Formation, however, does not exhibit folding at the outcrop- or micro-scale in the study area. The presence of folding may be dependent on the competency of lithology.

The third generation of folding (F_{iv}) deforms F_{3}, its associated structures and all older fabrics. F_{iv} fold sets are east–west trending upright, open folds with a spaced crenulation cleavage (S_{iv}) that is parallel to the axial plane of F_{iv} (Table 5). A spaced crenulation lineation (L_{iv}) is also associated with F_{iv} (Table 5). F_{iv} and S_{iv} are correlative to F_{4} and S_{4}, respectively, of Kim et al. (2007; 2011) (Table 6). The associated lineation (L_{iv}) correlates to L_{4} of Kim et al. (2011). All east–west striking cleavages observed in the current study area are correlated to S_{4} of Kim et al. (2007; 2011). Though there is a danger in basing the correlation of S_{4} on orientation alone, there is only one generation of foliation that strikes east–west and, thus, it remains an important criterion.

Morphological characteristics are also considered. All east–west striking cleavages are parallel to the axial planes of east–west trending, open folds and are identified as a spaced crenulation cleavage or disjunctive cleavage. They do not appear in all other previous work; rather, an intersection lineation between S_{3} and earlier foliations is discussed (Gillespie, 1975; Agnew, 1977).

F_{3} is overprinted by F_{4} so F_{4} is taken to be Acadian or later. It is possible that F_{4} is, in fact, part of the same generation of deformation as F_{3} (i.e. the result of constriction), but the cross-cutting relationship observed in this study points to F_{3} being older than F_{4}.

**Geometric Variations**
The first two generations of folding (F\textsubscript{1} and F\textsubscript{3}) deform all lithotectonic slices in the study area. F\textsubscript{1} folds are consistent in style and precede thrust displacement, as has been stated in previous studies (Gillespie, 1975; Dorsey et al., 1983; Strehle and Stanley, 1986; Kim et al., 2011). F\textsubscript{3} folds are expressed as crenulations (e.g. 11ML02_3, 10WL02) or open folds (e.g. 11FE01). Along the strike of the Hinesburg Thrust, F\textsubscript{3} folds change plunge direction between approximately north and south. F\textsubscript{4} folds deform rocks specifically in the upper plate of the Hinesburg Thrust and are recognized at the km-scale by refolded (F\textsubscript{3}) folds that plunge to the north and south.

The L\textsubscript{4} crenulation lineations are particularly informative because they are commonly observed in outcrop in association with presence of F\textsubscript{4}. This is particularly true in fine-grained rocks; the crenulation lineations are almost always expressed on the surface of a phyllitic layer. Moreover, the angle between L\textsubscript{3} and L\textsubscript{4} is a means for describing the geometry of refold structures. Figure 44 illustrates these variations and their spatial relationship to each other. L\textsubscript{3} and L\textsubscript{4} are consistently present in the Fairfield Pond Formation in the upper plate of the Hinesburg Thrust. Along strike of the thrust, L\textsubscript{3} alternatively plunges between north and south. An inflection point occurs at 10WL09 where L\textsubscript{3} plunges to the north and south. Other inflection points are inferred somewhere between two outcrops where a change in plunge direction of L\textsubscript{3} is observed. From south to north across the field area, four inflections points are inferred and one is observed (11WL09).

The same outcrops in the upper plate of the Hinesburg Thrust exhibit L\textsubscript{4} crenulation lineations as well. All but two outcrops show L\textsubscript{4} to plunge to the east (10WL11, L\textsubscript{4} plunges to the east and west; 11CLPD, F\textsubscript{4} crenulation fold hinge plunges

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west). This suggests that \( F_4 \) developed after the other generations of folds and, hence, refolded \( F_3 \) folds; which resulted in highly variable plunge directions of \( L_3 \). The plunge of \( L_4 \) is fairly consistent indicating it is an associated fabric of the youngest generation of folding.

The angular relationship between \( L_3 \) and \( L_4 \) varies slightly in the upper plate of the Hinesburg Thrust along strike. In the southern portion of the upper plate, within the study area, the two lineations are orthogonal to each other (outcrops 10WL05, 10WL11, 10WL09 in Fig. 30). In the northern portion, the two lineations are 90–100° apart.

\( F_3 \) folds in the lower plate of the Hinesburg Thrust are recorded as crenulations in fine-grained argillaceous rocks and tight folds in quartzites and dolostones. If \( F_4 \) is inferred by the variable plunge direction of \( F_3 \) then it is possible that the variable plunge of \( F_3 \) recorded in the lower plate of the Hinesburg Thrust may also be the result of east–west trending folds. In the lower plate, \( F_3 \) folds are most readily observed in the Skeels Corner Formation. The plunge of crenulation lineations change along strike of the Hinesburg Thrust between 10WL02 and 11ML02_3, from which \( F_4 \) is inferred. This is deduced from two outcrops and does not account for potential variation between them that is covered by the Hinesburg Thrust.

\( S_3 \) and \( S_4 \) are axial planar to \( F_3 \) and \( F_4 \), respectively. These foliations developed as a disjunctive cleavage in the carbonate siliciclastic sequence. One example of \( F_3 \) folding is seen in the carbonate siliciclastic sequence in Domain D. 11FE01 exhibits tight, asymmetric \( F_3 \) folds (Kim et al., 2011) that deform bedding. There is also an associated \( S_3 \) axial planar disjunctive cleavage. This sequence of events is very different from that documented in the finer-grained rocks to the east. It also contrasts with the structures
elsewhere in Domain D, which are predominantly horizontal bedding and two sets of fractures patterns. Since the north–south striking fractures are confirmed to be axial planar to F₃ at this outcrop, further correlation can be made between S₃ disjunctive cleavage associated with F₃ and the north–south striking fracture set. Figure 45 shows the dominant orientation of fracture sets and (S₃) disjunctive cleavage in domain D. There is a clear regional pattern consisting of approximately orthogonal fracture sets. Comparing rose diagrams of disjunctive cleavage patterns and crenulation lineations across strike of the Hinesburg Thrust (Figure 46), there is very close agreement in orientation and angular relationship. The two outcrops (11CL01, 11HB01) that deviate from this pattern are on the downthrown block of the St. George Fault.

**Conclusions**

Selected outcrops are grouped into structural domains based primarily on spatial/structural relationship to the Hinesburg Thrust, which coincides with lithology contrasts. Documenting the ductile structures in the upper and lower plates of the Hinesburg Thrust indicates a contrast in recorded deformation between lithotectonic slices from multiple orogenic-related deformation events. Three generations of folds are observed in the study area and correlated to regional structures that are attributed to orogenic events. Tight to isoclinal F₁ folds that are pre- and syn-Taconic thrust propagation deform compositional layering and bedding. A regional schistosity (S₁) is parallel to the axial plane of F₁. A west-dipping (S₂) cleavage is documented in other studies but in the current study. Other studies do not associate S₂ with a generation of folding. The subsequent generation of folds deforms all older fabrics, including S₂, and...
is thus referred to as \( F_3 \). \( F_1 \) and \( S_1 \) are folded by open to closed asymmetric \( F_3 \) folds that also deform the Hinesburg Thrust. A spaced cleavage (\( S_3 \)) is parallel to the axial plane of \( F_3 \) and locally observed crenulation lineations (\( L_3 \)) are parallel to the fold hinge of \( F_3 \). This fold set is correlated to a regional fold system that is Devonian in age and thus attributed to the Acadian Orogeny. All fabrics are deformed by open, asymmetric \( F_4 \) folds; which also has an associated axial planar, spaced cleavage (\( S_4 \)) and locally observed crenulation lineation (\( L_3 \)). These folds overprint \( F_3 \) and are taken to be Acadian or later.

The upper plate domain of the Hinesburg Thrust records all three generations of folds although fine-grain argillaceous rocks more readily record \( F_3 \) and \( F_4 \). The later folds are often expressed as crenulations. The fault block between the Hinesburg and Muddy Brook Thrusts, the Hinesburg Thrust footwall anticline, records the pre-thrust \( F_1 \) folding event and only one Acadian event, \( F_3 \). However, variation in the direction of \( F_3 \) plunge may indicate the presence of large scale \( F_4 \) folds.

The carbonate siliciclastic sequence in the lower plate of the Hinesburg Thrust records little, if any, ductile deformation. Nevertheless, fracture sets, the dominant structures in these rocks besides bedding, show a remarkable similarity in orientation to the orientation of cleavages associated with \( F_3 \) and \( F_4 \).

\( F_3 \) folds and associated structures trend approximately north–south; \( F_4 \) folds and associated structures trend approximately east–west. The plunge of \( F_3 \) fold axes and crenulation lineations fluctuates between northwest to northeast and southwest to southeast, but the plunge of \( F_4 \) linear structures is almost always east. The angular relationship between the two fold sets is consistently orthogonal (85–100°) both along
strike of the Hinesburg Thrust and across it creating Type 1 refold structures (Grasemann et al., 2004).

This study is a geometric analysis of structures observed in the upper and lower plates of the Hinesburg Thrust. Further analyses are needed to document kinematics and quantify strain in order to develop a model for crustal deformation in this area.

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Figure 6 Bedrock belts (modified from Doll et al., 1961) and geologic map of study area (modified from Ratcliffe et al., 2011). Note the Hinesburg Thrust separates the Champlain Valley belt from the Green Mountains belt.
Figure 7  Location of Domains A, B, C and D Domains are highlighted on the geologic map from Figure 5 (Ratcliffe et al., 2011). Outcrops are grouped into domains based primarily on structural relationship with the Hinesburg Thrust, which coincides with lithology contrasts. Please see map legend in Figure 6 for explanation of units.
Figure 8 Annotated field photos of 10WL09. a) Recumbent $F_{n+1}$ fold deforms $S_n$ compositional layering in the Fairfield Pond Formation. It is folded again by north-south trending, open, steeply-inclined, asymmetric crenulations. An associated crenulation cleavage $S_{n+2}$ strikes north and dips to the east. Photo of vertical outcrop looking south. b) $S_{n+2}$ is observed as “steps” in the outcrop; east–west trending crenulations observable by an associated closely spaced $L_{n+3}$ crenulation lineation. Photo of $S_n$ surface looking to the ground, north is to the right.
Figure 9 Stereonet and photomicrographs of sample 10WL09. a) Lower hemisphere equal area projections of outcrop fabric elements. There is an orthogonal relationship between the two sets of crenulation cleavage, $S_{n+2}$ and $S_{n+3}$. Both sets of crenulations are observed in thin section: north–south crenulations (b) and east–west crenulations (c). Crenulation cleavages are defined by the realignment of micas. b) Thin section is cut parallel to $L_{n+2}$ and perpendicular to $S_n$. c) Thin section is cut parallel to $L_{n+3}$ and perpendicular to $S_n$. 
Figure 10 Annotated field photos of 11ES01. a) Closed to tight, moderately inclined, west-verging $F_{n+1}$ folds deform compositional layering ($S_n$) and quartz veins. Phyllitic axial-planar cleavage ($S_{n+1}$) strikes approximately north and moderately dips to the east. Photo is looking north-northeast. Scalebar is 0.5 m. b) Early foliations are folded by north-south trending, asymmetric crenulations ($F_{n+2}$). The associated crenulation $S_{n+2}$ cleavage strikes northeast and steeply dips southeast. Photo is looking northeast. Scalebar is 0.5 m.
Figure 11 Stereonet and microphotographs of sample 11ES01. a) Lower hemisphere equal area projections of all fabric elements at the outcrop b) Photomicrograph of $S_n$ and subparallel, anastomosing $S_{n+1}$. Thin section is cut parallel to $L_{n+2}$ and perpendicular to $S_{n+1}$. c) North–south trending crenulations have an associated crenulation cleavage $S_{n+2}$ dipping east. Thin section is cut perpendicular to $L_{n+2}$ and perpendicular to $S_{n+1}$. 
Figure 12 Block diagram summarizing the fabrics observed at 11ES01. Diagram highlights two obliquely oriented sets of crenulation lineations.
Figure 13 Microphotographs of sample 11CL03. a) N–S trending crenulations have an associated crenulation cleavage ($S_{n+2}$) dipping east. Thin section is cut parallel to $L_{n+2}$ and perpendicular to $S_n$. b) $S_{n+3}$ dips to the north and is selectively developed in clay-rich layers. c) Pinch-and-swell structures are observed in quartz-rich layers. Thin section is cut perpendicular to $L_{n+2}$ and perpendicular to $S_n$. 
Figure 14 Annotated outcrop photos of 11CLPD. a) $F_{n+2}$ set are open, steeply inclined folds tilting to the west refolds earliest foliation ($S_n$). Photo looking north. Scalebar is 0.1m. b) Youngest generation of folding ($F_{n+3}$) deforms all older fabrics.
Figure 15 Field sketch and stereonets for 11CLPD. a) Field sketch showing that \( L_{n+2} \) and \( L_{n+3} \) are orthogonal to each other and can be observed on the surface of \( S_n \). Two sets of crenulations make a Type 1 refold structure. \( S_{n+3} \) is not seen from this view. b) Compositional layering is folding; poles to compositional layering define a girdle with a \( \pi \)-axis. c) Lower hemisphere equal area projections of all outcrop fabrics except \( L_{n+3} \), which was observed but not measured. \( S_{n+2} \) is axial planar to \( F_{n+2} \) and have a highly variable dip.
Figure 16 Outcrop photo and stereonet for 11WL11. a) $S_n$ compositional layering is folded by moderately inclined, west-verging, asymmetric $F_{n+1}$ crenulations; which have an associated spaced crenulation $S_{n+1}$ cleavage. An associated $L_{n+1}$ crenulation lineation plunges to the south. Microfaults (bold line) observed in competent metapsammite layers show apparent top to the west-southwest motion. Photo is looking south. b) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop. The youngest structures are $L_{n+2}$ crenulation lineations that plunge to the east and west.
Figure 17  Block diagram of 10WL11. The two sets of folds are expressed on an $S_n$ surface much like that seen in other outcrops in this domain. They are orthogonal to each other producing a Type 1 refold structure. Quartz-rich layers exhibit microfaults with top-to-the-west sense of offset.
Figure 18  Annotated outcrop photo and stereonet for 11WF01. Nearly vertical $S_{n+1}$ schistosity forms a south-plunging $L_{n+2}$ intersection lineation with steeply-dipping disjunctive $S_{n+2}$ cleavage. Photo is looking north. b) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to foliations are plotted).
Figure 19 Photomicrographs of sample 11WF01.  a) Bedding-parallel S_{n+1} schistosity is near vertical and defined by the preferred orientation of detrital quartz and feldspar. Thin section is cut perpendicular to both S_{n+1} and L_{n+2}.  b) Sigmoidal quartz clasts show an apparent sinistral sense of shear. Thin section is cut approximately normal to S_{n+1} and parallel to L_{n+2}.
Figure 20  Stereonet of field data from 11ES03. Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to S surfaces are plotted). Quartz veins are seen to be folded in the field.
Figure 21 Schematic block diagram of fabric elements in Domain A. The diagram specifically shows domes and basins in the Fairfield Pond Formation but not in the Pinnacle Formation. The oldest generation of folds are tighter in Fairfield Pond than the Pinnacle Formation.
Figure 22  Vertical outcrop photos of 11WL02, looking south. a) Slaty cleavage is folded by $F_{n+1}$ and $F_{n+2}$. b) $S_{n+2}$ crenulation cleavage truncate $S_n$. 
Figure 23  Field sketch and stereonets of 10WL02.  a) Field sketch showing shear band cutting all earlier fabrics.  b) Poles to $S_n$ define a girdle to $F_{n+1}$ with a $\pi$-axis.  Field measurements show the fold to plunge to the north–northeast and south–southwest.  c) Lower hemisphere equal area projection of macroscopic fabric elements.  The spaced $S_{n+1}$ cleavage is parallel to the axial plane of $F_{n+1}$.  $L_n$ plots on the same great circle as poles to $S_n$.  $L_{n+1}$ crenulation lineation plunges to the north–northeast.
Figure 24 Photomicrographs of sample 10WL02a. a) $S_n$ is defined by a combination of fabric elements: preferred orientation of platy minerals and compositional layering. Thin section is cut parallel to $L_n$ and perpendicular to $S_n$. b) $S_n$ has parallel cleavage domains in the YZ plane except in coarse layers (in this example it is a layer of carbonate material). Thin section is cut perpendicular to $L_n$ and perpendicular to $S_n$. Blue color is epoxy.
Figure 25 Photomicrographs of samples 10WL02a and 10WL02b, respectively. a) Conjugate crenulation cleavage cuts early compositional layering and phyllitic cleavage. Thin section is cut perpendicular to both $L_n$ and $S_n$. b) A thin section oriented east–west and cut perpendicular to the axis of $F_{n+1}$. Apparent offset may be due to slip along $S_{n+1}$ or the result of dissolution. $S_{n+1}$ is folded by subsequent folding. See text for more discussion. Blue color is epoxy.
Figure 26 Block diagram illustrating fabrics documented at outcrop 10WL02.
Figure 27 Photomicrographs of sample 10WL01.  a) Two foliations observed in a thin section cut perpendicular to the foliation $S_{n+2}$ and parallel to the respective down-dip direction. Foliation is defined by fine-grained elongated quartz. b) Antitaxial veins formed in a deformed quartz pebble in a matrix of fine-grain argillaceous quartzite. It is identified by an inverse symmetry about a weak median line. Thin section was selectively cut to highlight a quartz pebble in cross section view. c) Lower hemisphere equal area projections of macroscopic fabrics.
Figure 28  Schematic diagram showing stretched pebbles and three generations of cleavage documented at outcrop 10WL01.
Figure 29 Outcrop photo and field sketches for 11ML03. a) Isoclinal (Fn+1) folds deform compositional layering. Phyllitic cleavage (S_{n+1}) is parallel to Fn axial planes. S_{n} is folded again by (F_{n+2}) asymmetric crenulations. Space (S_{n+2}) crenulation cleavage is oblique to S_{n+1}. b) Quartz vein deformed by isoclinal F_{n} folds. Field sketch of isoclinal F_{n} folds deforming S_{n} compositional layering. S_{n+1} phyllitic cleavage is parallel to the axial plane of F_{n+1}. 
Figure 30 Photomicrograph of sample 11ML02 and stereonet for 11ML02_3. a) A west-dipping foliation, not observed in the field, can be seen in thin section. Thin section is cut perpendicular to L_{n+2} and perpendicular to S_n. Apparent offset may be due to slip along the west-dipping cleavage plane or the result of dissolution. b) Lower hemisphere equal area projection of macroscopic fabric elements. S_{n+1} and S_{n+2} are subparallel, but S_{n+1} has more variability of strike. The intersection lineation between S_n and S_{n+1} is parallel to the crenulation lineation associated with F_{n+2}. Blue color is epoxy.
Figure 31  Schematic diagram of fabrics documented at outcrops 11ML02 and 11ML03.
Figure 32. Schematic block diagram of fabric elements in Domain B. The oldest generation of folds deforms compositional layering/bedding. From north to south, the second generation of folds plunges down to the northeast then up to the northwest.
Figure 33 Photomicrograph and stereonets for 11ML10. a) Poles to \( S_0/S_n \) define a girdle to \( F_{n+1} \) whose hinge plunges gently to the northwest. b) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to \( S \) surfaces are plotted). \( S_{n+1} \) is strikes north and dips to the east–southeast and formed an intersection lineation (\( L_{n+1} \)) with \( S_0/S_n \) that plunges south. c) Photomicrograph of compositional layering (\( S_0 \)) cut by slaty cleavage (\( S_{n+1} \)) selectively developed in argillaceous layers and dies out in quartz-rich layers. Thin section is cut vertically and parallel to the strike (northeast) of \( S_0 \).
Figure 34 Cartoon illustrating the fabrics documented at 11ML10. $S_{n+1}$ has shallower dips in the upper plate of the Arrowhead Mountain Thrust suggesting counterclockwise rotation associated with fault movement.
Figure 35  Annotated field photo of 11ML05 and stereonet for 11ML456. a) Bedding parallel schistosity exhibits anastomosing cleavage domains possibly due to competency contrast between fine-grained mud layers and coarser quartz-rich layers. b) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to S surfaces are plotted). $S_0/S_n$ strikes northeast and dips to the southeast becoming shallower from east to west. Mineral lineations ($L_n$) plot in the plane of $S_0/S_n$. 
Figure 36  Field sketch and stereonet for 11ML09.  a) Field sketch showing bedding folded by tight folds.  Spaced cleavage is axial planar.  Cleavage dips less steeply than bedding indicating the fold is overturned.  b) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to S surfaces are plotted).  $F_1$ fold hinge plunges gently to the south–southeast.
Figure 37  Schematic block diagram of fabric elements in Domain C. First generation folds are isoclinal; limbs are often parallel axial planar schistosity. Folds in the upper plate of the Arrowhead Mountain Thrust tighten as they approach the thrust. A slaty cleavage is shallower in the upper plate within the fault zone, however, it is not observed further east (away from the fault).
Figure 38 Annotated field photos stereonets and a rose diagram for 11CL01. a) Field photo showing bedding ($S_0$) and lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to $S$ surfaces are plotted). Original bedding is folded by open, gently inclined $F_1$ fold. No associated cleavage is observed. Microfaults show sinistral offset in bedding (b) and fractures (c). d) Rose diagram showing the dominant orientation of fractures. e) Poles to microfault surfaces.
Figure 39  Outcrop photo of 11CL10 and stereonets and rose diagram of 11CL10, 11BU01, 11BU02, 10CH03.  a) Lower hemisphere equal area projection of poles to bedding ($S_0$) show them to be almost horizontal b) Example photo looking down at the pavement of 11CL10 shows two sets of vertical fractures and flat-lying bedding. c) Contoured poles to fractures and a rose diagram both show north–south and east–west fracture sets.
Figure 40  Field sketch and stereonets for 11FE01 and photomicrograph for sample 11FE01. 
a) Photomicrograph shows fine-grained bedding b) Flexural slip between bedding planes is observed at the outcrop scale. c) Poles to $S_0$ define a girdle to $F_n$, whose hinge plunges gently south. d) Lower hemisphere equal area projection of macroscopic fabric elements at the outcrop (poles to $S$ surfaces are plotted). $S_n$ is parallel to the axial plane of $F_n$. 
Figure 41 Rose diagram and stereonet for 10SB02. Both the rose diagram (a, strikes are plotted) and lower hemisphere equal area projection (b, poles to fractures are plotted) show a north-south strike set of fractures and a pair of conjugate fractures.

Figure 42 Rose diagram and stereonet for 11HB01. Both the rose diagram (a, strikes are plotted) and lower hemisphere equal area projections (b, poles to fractures are plotted) show north–northeast and northwest striking subvertical fracture sets.
Figure 43  Schematic block diagram of fabric elements in Domain D. Bedding, large-scale syncline and two fracture sets are the dominant structures, however, a second generation of folding and its associated axial planar cleavage is observed at 11FE01.
Figure 44  Rose diagrams showing the plunge direction of crenulation lineations recorded in Domain A and B. $L_3$ (black circles) typically plunge approximately north and/or south; $L_4$ (gray circles) typically plunges approximately east and/or west. See text for more discussion.
Figure 45  Rose diagrams showing the dominant orientation of subvertical fractures and (S₃) disjunctive cleavage in Domain D. Patterns generally suggest two sets of fractures striking approximately north–south and east–west. See text for more discussion.
Figure 46 Rose diagrams of L$_3$ and L$_4$ crenulation lineations from Domain A and B and fractures from Domain D. There is a clear regional pattern consisting of approximately orthogonal fracture sets. Comparing the dominant orientation of the fracture patterns to that of the crenulation lineations across strike of the Hinesburg Thrust there is very close agreement in orientation and angular relationship.


Gillespie, R., 1975, Structure and Stratigraphy along the Hinesburg Thrust, Hinesburg, Vermont [MS: University of Vermont, 63 p.


Kim, J., Gale, M., Thompson, P. J., and Derman, K., 2007, Bedrock geologic map of the Town of Williston, Vermont, scale 1:24,000.


Appendix A Field station coordinates and associated samples

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(*) thin sections not prepared from collected sample