THE CHAMPLAIN THRUST SYSTEM IN NORTHWESTERN VERMONT - STRUCTURE AND LITHOLOGY OF THE TACONIC FORELAND SEQUENCE IN THE HIGHGATE CENTER QUADRANGLE

Abstract of
a thesis presented to the Faculty
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Department of Geological Sciences

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ABSTRACT

The Champlain Thrust System, responsible for the final emplacement of the Taconic Allochthon, and which represents the surface trace of the main detachment under the Taconic suite, can be traced throughout the study area in northwestern Vermont, northward into southern Québec. In the Highgate and St. Albans area this imbricated assemblage consists of 3 main carbonate thrust slices (Highgate Springs-, Philipsburg-, Rosenberg slice), each of which contain rocks deposited during some part of early Cambrian to medial Ordovician time. The Highgate Springs sequence comprises a mildly deformed early to medial Ordovician carbonate-shale succession. Late Cambrian to early Ordovician carbonates of the Philipsburg sequence, some quartz-arenaceous to argillaceous, form a northeast-plunging syncline structure that appears as an intercalated wedge within the thrust system. The Rosenberg sequence consists of early Cambrian to medial Ordovician shallow water siliciclastics and carbonates of the continental shelf and shelf edge that has been believed thoughout the last decades to represent a largely unfaulted regional synclinal structure ("St. Albans Synclinorium"). In particular, the eastern contact of the Rosenberg carbonates with the overlying Morses Line Formation has been interpreted as a stratigraphically intact, rapid facies change from shallow water siliciclastics and carbonates into deeper water sediments of the continental slope and/or rise. In contrast, my detailed lithostructural study of the Highgate and St. Albans region favors a substantially faulted nature of this contact. The significant contrast of the dip angle of the average cleavage foliation (~15°) across the Highgate-Morses Line Formation contact suggests a substantial structural discontinuity between the Rosenberg and the Morses Line sequence. This structural break is interpreted as a continuous detachment fault (*"St. Albans Detachment"*) that caused a counter-clockwise rotation of the average cleavage fabric within the Morses Line Formation and juxtaposes mildly deformed Cambro-Ordovician siliciclastics, carbonates and siltstones/shales of the Rosenberg sequence with intensely strained slates of the medial Ordovician Allochthon. The extension of the St. Albans Detachment can be extrapolated between Burlington, Vermont, and Drummondville, Québec, where it causes a strongly varying thickness of the shelf carbonate

strata [Dunham Dolomite, Monkton Quartzite, Winooski Dolomite, Danby Formation] that are exposed adjacent to intensely strained slates, and, in Canada, juxtaposes mélange with Taconic sequences [Stanbridge Nappe, Granby Nappe]. In addition, a closely spaced suite of northeast-southwest trending normal/tear faults cross-cuts the main structural north-south trend in the study area and can be extrapolated across the international border.

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This thesis is a little dream that came true. I'm indebted to everyone who contributed to this project and my time in Albany.

Most grateful, I appreciate the patience of Bill Kidd, the driving force of this project, who not only introduced me, outdoors, to the indeed intricate Taconic geology in Vermont and New York State but also provided, indoors, more than once, some decent breakfast at Friendly's. Cheers ! An admirable thank you to Win Means for pointing out the guidelines and the importance of basic and clear thinking in all stages of my work. The extended use of Greg Harper's enviable hardware equipment enabled me to produce what I produced and is greatly acknowledged. John Delano's scientific enthusiasm provided an challenging ideal for practicing research in all scales. Special thanks to Barry Doolan and his family for their warm hospitality and the really good time in their cabin in the state of the Green Mountains. Diane, an extraordinary "thank you" to you for the general moral support and for suffering my impatient moments (I'll be a good boy !).

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INTRODUCTION

"... No Cambrian area in all North America is more interesting stratigraphically, or shows more complicated structure, than that of northwestern Vermont ..."

Charles Schuchert 1937

1. INTRODUCTION

This study investigates the geology of the Highgate Center quadrangle and of the northwestern portion of the St. Albans quadrangle in northwestern Vermont (***Figure 1.1.***), situated between longitudes 73°00'00" and 73°07'30" west. The international border to Québec/Canada together with the shoreline of Lake Champlain form its borders towards the north and northwest. The western boundary of the field area follows Route 7 into St. Albans. Reasonable access is provided by the north-south Interstate Highway 89 (I-89) together with Route 207 as well as the east-west Route 78, crossing the study area (***Figure 1.2.***). Exit 20 of I-89 marks an approximate southern limit of the core field area. The study area is a strip about 18 km long and 9 km wide, located about 50 km north of Burlington, Vermont and about 100 km south of Montreal.

Topographically, the area is characterized by extended NNE-SSW trending, gently rounded hills, ranging in height between 465 feet (142 m) WNW of Cutler Pond in the northern part of the Highgate quadrangle to 95 feet (29 m) along the shoreline of Lake Champlain (***Figure 1.3.***). The hills, which contain most of the outcrop, are generally covered by woods, whereas pastures largely occupy the lowlands. This landscape is predominantly due to continental glacier activity during the last ice age as striated surfaces on several outcrops and numerous erratic boulders indicate. Areas lower than 250 feet (76 m) are generally underlain by sands and clays, deposited in a periglacial Lake Champlain which followed the retreat of the glaciers.

Systematic mapping was done between May and October 1993 and between May and June 1994, based on enlargements of 7.5-minute topographic maps of the Highgate Center and the St. Albans quadrangle of the U.S.

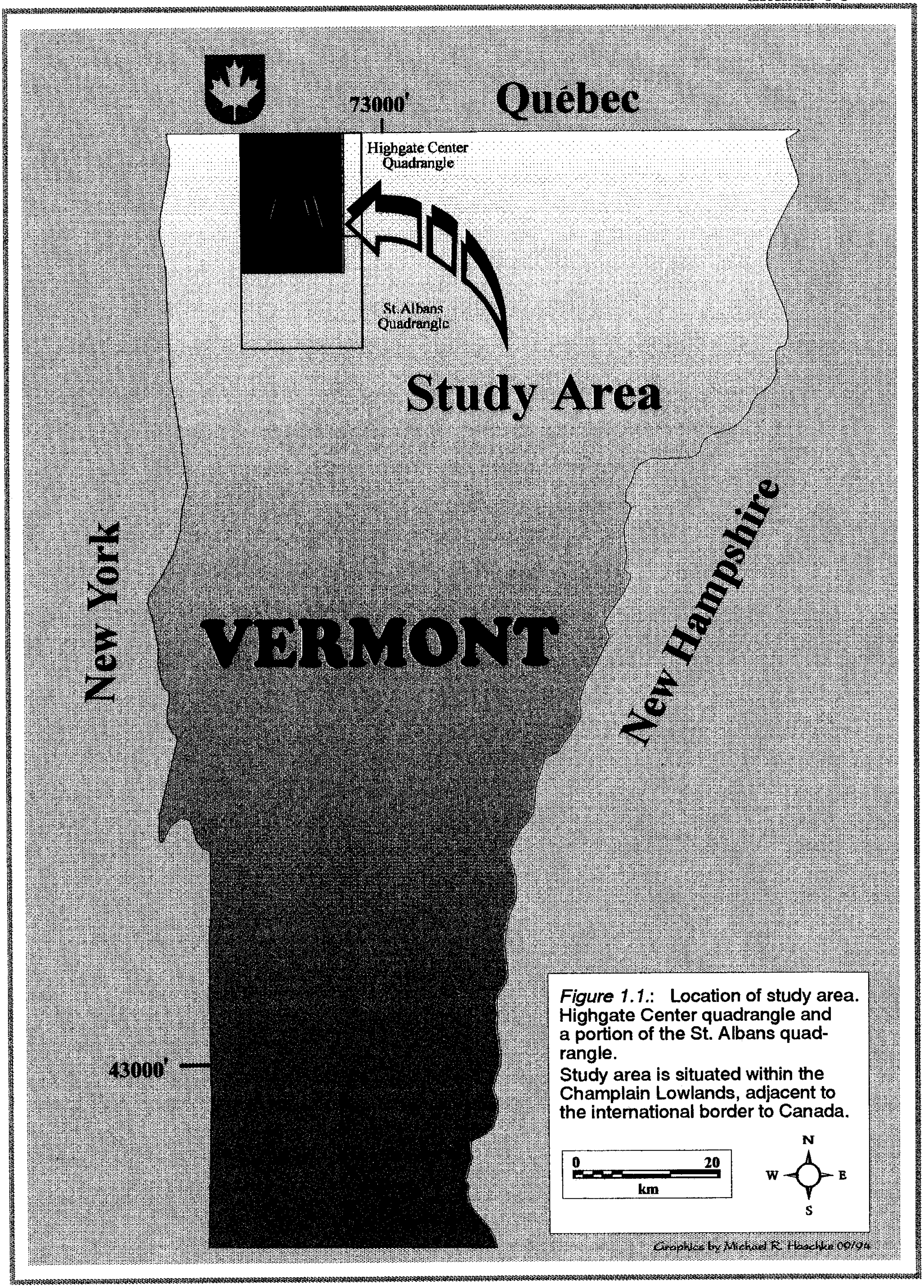


Figure 1.1.: Location of study area. Highgate Center quadrangle and a portion of the St. Albans quadrangle.
Study area is situated within the Champlain Lowlands, adjacent to the international border to Canada.

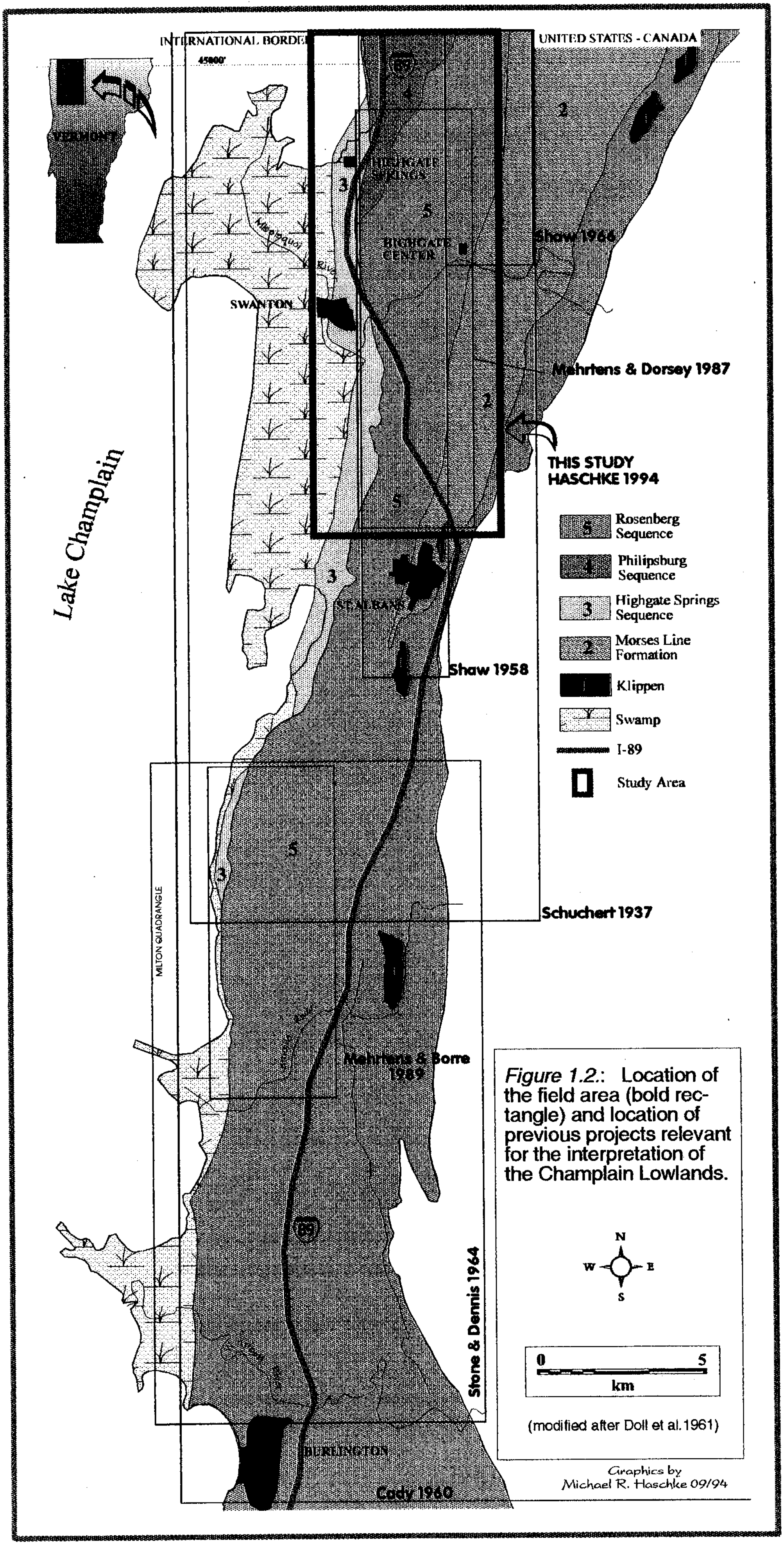


Figure 1.2.: Location of the field area (bold rectangle) and location of previous projects relevant for the interpretation of the Champlain Lowlands.

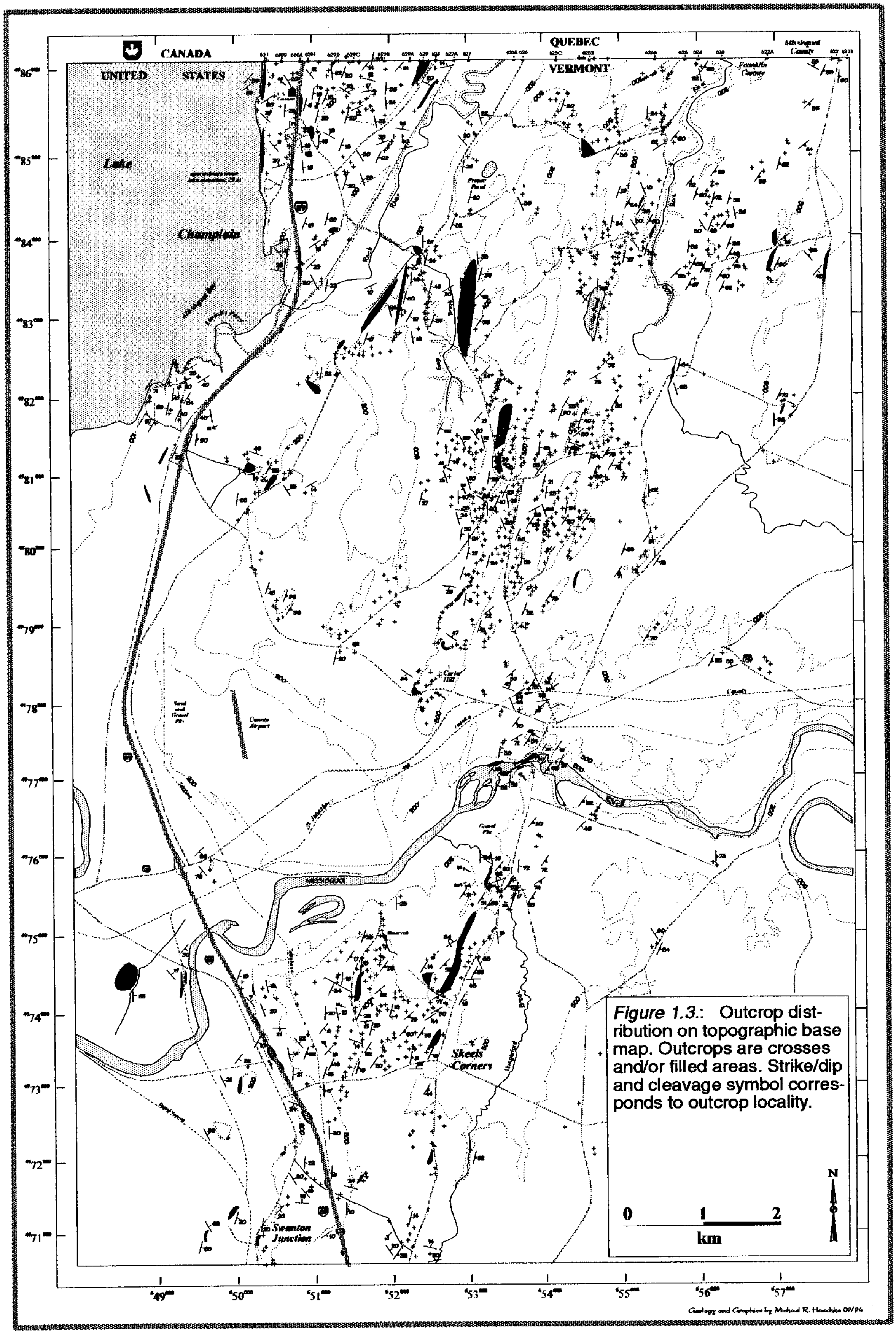


Figure 1.3.: Outcrop distribution on topographic base map. Outcrops are crosses and/or filled areas. Strike/dip and cleavage symbol corresponds to outcrop locality.

Geological Survey. All geologic data is derived from 460 detailed documented outcrops and about 300 additional registered exposures (***Figure 1.3.***). Glacial and postglacial sedimentation is responsible for the lack of outcrop below 250 feet (76 m) and therefore restricts the exposure to about 1% of the entire area. The best exposures are available where the Missisquoi River cuts across strike at Highgate Falls.

1.1. GEOLOGICAL FRAMEWORK

Early Cambrian to medial Ordovician sedimentary strata of the continental shelf, slope and rise are exposed on a northeast-southwest trending strip between Alabama and western Newfoundland. They were deposited in the classic example of a passive continental margin sequence in early Paleozoic times [Bird & Dewey 1970, Read 1989] when a sedimentary sequence, characterized on the rise by mudrocks containing shelf-derived turbidites, debris-flows and contourites [Rowley et al. 1979] and on the shelf by shallow water siliciclastics and carbonates, accumulated along the subsiding continental margin of North America and the adjacent ocean floor of the *"Proto-Atlantic"* (Iapetus Ocean). Eastward directed subduction caused the closure of this at least 500-900 km wide ocean [Bradley 1989] in the early Ordovician and was followed by the collision of the North-American continental passive margin with a collage of volcanic island arcs and/or microcontinents [Chappel 1973, Hiscott 1978, Rowley 1983] during medial Ordovician times. This major diachronous orogenic event, named the Taconic Orogeny, involved large-scale compressional deformation and caused the development of a major thrust and fold belt. Large nappes of thrusted continental shelf slices and allochthonous deeper marine sedimentary strata, and locally ophiolites, were obducted westward and imbricated atop of shelf facies. The tectonic contact between autochthonous and allochthonous rocks, the *Champlain Thrust System*, is characterized by a complex thrust-imbricated and splintered belt of siliciclastic and carbonate shelf slices of varying lithology.

1.2. REGIONAL SETTING

A portion of the continental passive margin sequence is exposed in northwestern Vermont. Early Cambrian to medial Ordovician siliciclastic and carbonate sediments of the shelf, slope and rise form a north-south trending thrust-imbricated sequence of the Taconic foreland in the Champlain Lowlands. All observed thrust slices dip gently towards the east to southeast and exhibit lithologies ranging from non quartz-arenitic shelf carbonates and massive quartz-arenitic, partly recrystallized, calcareous dolostones in the western and central part to shaly limestones and fissile slates towards the eastern section of the study area.

The *Highgate Springs Thrust* represents the most western thrust fault occurrence within the Champlain Thrust System of this area and separates mildly deformed medial Ordovician black shales of the structural foreland from the overthrusted late early Ordovician quartz-arenitic to argillaceous limestones of the Highgate Springs thrust slice (***Figure 1.2.***). Towards the east, from Swanton northward into Québec, the Champlain Thrust System separates the Highgate Springs slice from an overthrusted late Cambrian to early Ordovician northeast-plunging syncline (***Figure 1.2.***), the dolostone and limestone-dominated *Philipsburg sequence* [Cady 1960]. A third branch of the Champlain Thrust System adjoins the Philipsburg thrust slice from Highgate Springs southward and separates the synclinal Philipsburg structure from a overthrusted siliciclastic shelf carbonate sequence ranging from early Cambrian dolostones and shales to early Ordovician limestone strata; this sequence, which is quartzose, dominated by quartz-arenitic dolostones, is termed the *Rosenberg sequence* [Cady 1945]. South of Highgate Springs this sequence lies in immediate thrust fault contact with the Highgate Springs slice. The Rosenberg thrust slice represents the core of the field area (***Figure 1.2.***) and is tectonically juxtaposed with severely deformed and imbricated calcareous to non-calcareous slates, which represent the beginning of the allochthonous strata of the continental rise.

All structural elements follow an approximate north-south trend and can be traced across the international border into Québec as well as southward towards the Burlington area.

1.3. PURPOSE OF STUDY

A detailed structural and lithological investigation of the Highgate and St. Albans area was intended to test previous tectonic models and to propose a structurally reasonable alternative.

All previous studies on the overthrust Rosenberg sequence and its eastern contact to the *Stanbridge Nappe* [Schuchert 1937; Shaw 1958, Mehrtens & Dorsey 1987] show an anomalously thin (1.5-2 km) shelf edge-slope sedimentary sequence, juxtaposed with equivalent deep-water sediments of the early Paleozoic continental rise. This requires anomalously low subsidence rates at the shelf edge throughout the approximately 80 m.y. that this margin existed.

However, subsidence rates of the continental passive margin of early Paleozoic North America in Maryland and Virginia are typical of mature passive margins (3-6 cm/1000 yr.; [Bova & Read 1987]). Maximum subsidence and thickest sections (5 km or more) of shelf sediments are normally expected to occur at the outer edge of a passive margin [Read 1989].

The thermal subsidence history of the northwestern Vermont portion of the northern Appalachians is not fully understood. Reduced subsidence rates of the Vermont portion of the shelf in medial and late Cambrian times, caused by long-term uplift activity of the Adirondacks [Read 1989] and progressive cooling of the passive margin [Mehrtens 1985], however, are unlikely to account for the strongly reduced thickness of the shelf edge-slope sequence.

No previous interpretation provides a satisfactory explanation for the extremely reduced shelf-slope sequence and the structural features observed in this area.

Some further problems and questions investigated are:

- Is the contact between the Rosenberg sequence and the Morses Line Formation/Stanbridge Nappe stratigraphically intact or is it faulted ?
- Can the Highgate Falls Thrust, exposed at the Missisquoi River gorge at Highgate Falls be traced towards the north and south of this locality ?
- What are the structural relations within the Rosenberg sequence ?
- What is the nature of the contact between the Gorge and the adjacent Highgate Formation ? Is it stratigraphically intact [Mehrtens & Dorsey 1987] or is it faulted [Pingree 1982] ?
- What is the deformational character of the Morses Line Slate ?
- Is the Champlain Thrust a single detachment horizon ?

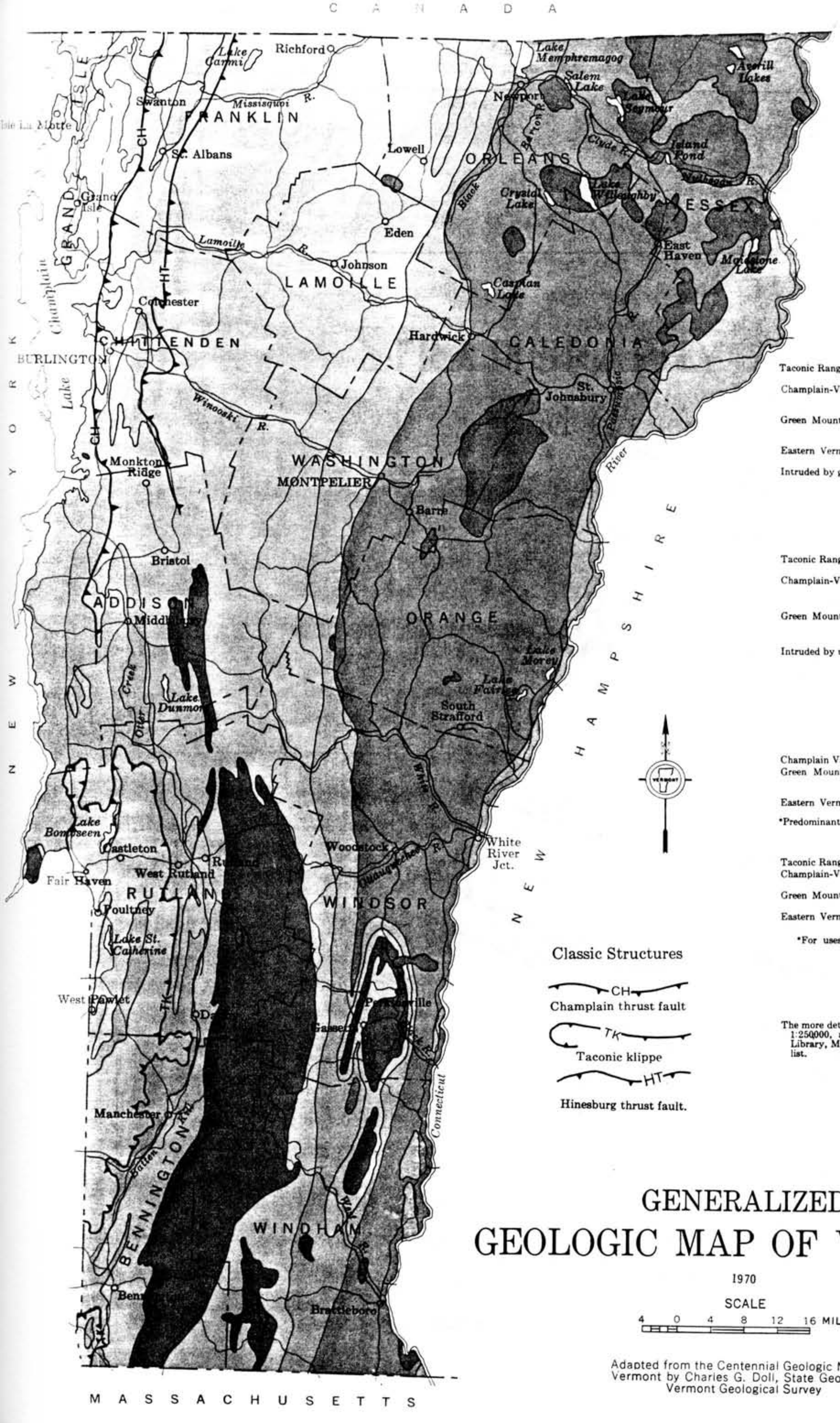
The results of the investigation of these problems support a modified structural interpretation for the Rosenberg sequence, and I propose a new model for the nature of the contact between autochthonous and allochthonous sediments.

1.4. PREVIOUS WORK

The basic outlines of the geology of Vermont are visible on the *Centennial Geologic Map of Vermont* [Doll et al., 1961, 1:250,000] which is the product of a "basic mapping program" of the Vermont Geological Survey between 1947-61, carried out by several workers (***Figure 1.4.***).

The first structural work in northwestern Vermont, but predominantly a paleontological investigation, was conducted by a Canadian worker, William Logan [1863]. Keith's [1932] synthesis of geological information from the 19th century in this area underwent stratigraphic revision by Schuchert [1937, 1:200,000, ***Tab.1.1.***]. Shaw's reconnaissance mapping project [1958, 1:62,500] of the St.Albans area (including Highgate Center) summarized the stratigraphic studies through the late 1950's and represented the basis for the northwestern region of Vermont's Centennial Geologic Map (***Figure 1.5.***). Further work was focused on paleontological aspects and mainly concentrated on isolated exposures in this area [Shaw 1966; Shaw & Clark 1968].

Recent results of the structural and stratigraphic investigation in northwestern Vermont (***Figure 2.4.***) have been published in *Vermont Geology* [Mehrtens 1985], the *Québec-Vermont Appalachian Workshop* [Mehrtens 1987, Doolan 1989] and, most detailed, in the *Special Bulletins of the Vermont Geological Survey* [Mehrtens & Dorsey 1987, 1:24,000 & 12,000]. This most recent review-mapping is incomplete at the scale presented and lacks necessary detailed structural data. Its structural interpretation is unsatisfying and cannot account for the complex structural relations in northwestern Vermont. Also, usage of the many local formation names, both in Québec and Vermont, has hindered understanding of regional patterns.



EXPLANATION

Igneous Rocks*

Granite, syenite, basalt, dunite, peridotite, *serpentinite*.

Silurian-Devonian

Slate, *phyllite*, *limestone*, quartzite, conglomerate, greenstone, schist, *amphibolite*. Intruded by granite, and syenite.

Ordovician

Taconic Range—*Slate*, *graywacke*, quartzite, *limestone*, conglomerate, *marble*.

Champlain-Vermont valleys—*Shale*, dolomite, *limestone*, quartzite, phyllite, slate, sandstone, conglomerate, *marble*.

Green Mountains—*Phyllite*, *schist*, quartzite, *greenstone*, slate, graywacke, gneiss, conglomerate, amphibolite.

Eastern Vermont—Phyllite, quartzite, greenstone, *schist*, gneiss, slate, *amphibolite*.

Intruded by granite, syenite, basalt, *ultrabasic rocks*.

Cambrian

Taconic Range—*Slate*, *graywacke*, quartzite, limestone, phyllite, sandstone, *marble*, dolomite.

Champlain-Vermont valleys—*Quartzite*, *dolomite*, slate, phyllite, sandstone, *shale*, limestone, conglomerate, *marble*.

Green Mountains—*Schist*, *phyllite*, quartzite, graywacke, conglomerate, *greenstone*, dolomite, limestone, *gneiss*, *amphibolite*.

Intruded by ultrabasic rocks, basalt.

Precambrian

Champlain Valley (small area)—*Gneiss*, quartzite, granulite.

Green Mountains—*Schist*, *gneiss*, metagraywacke, quartzite, calcite and dolomite marbles, amphibolite.

Eastern Vermont—*Gneiss*, *schist*, quartzite, calcite, and dolomite marble, amphibolite.

*Predominant and important rocks in *italics*.

Earth Materials*

Taconic Range—Slate, marble.

Champlain-Vermont valleys Limestone, marble, clay, kaolin, roadstone.

Green Mountains—Talc, asbestos, verd antique marble, roadstone.

Eastern Vermont—Granite, talc, roadstone, copper (now inactive).

*For uses consult *The Mineral Industry of Vermont*, U. S. Bureau of Mines, Preprint, obtainable from Superintendent of Documents, U. S. Government Printing Office, Washington, D .C. 20402

Also, use information available from the mineral industries,

The more detailed Centennial Geologic Map of Vermont, scale 1:250,000, available from State Librarian, Vermont State Library, Montpelier, Vermont 05602. Write for publication list.

Classic Structures

CH — Champlain thrust fault

TK — Taconic klippe

HT — Hinesburg thrust fault.

Figure 1.4.

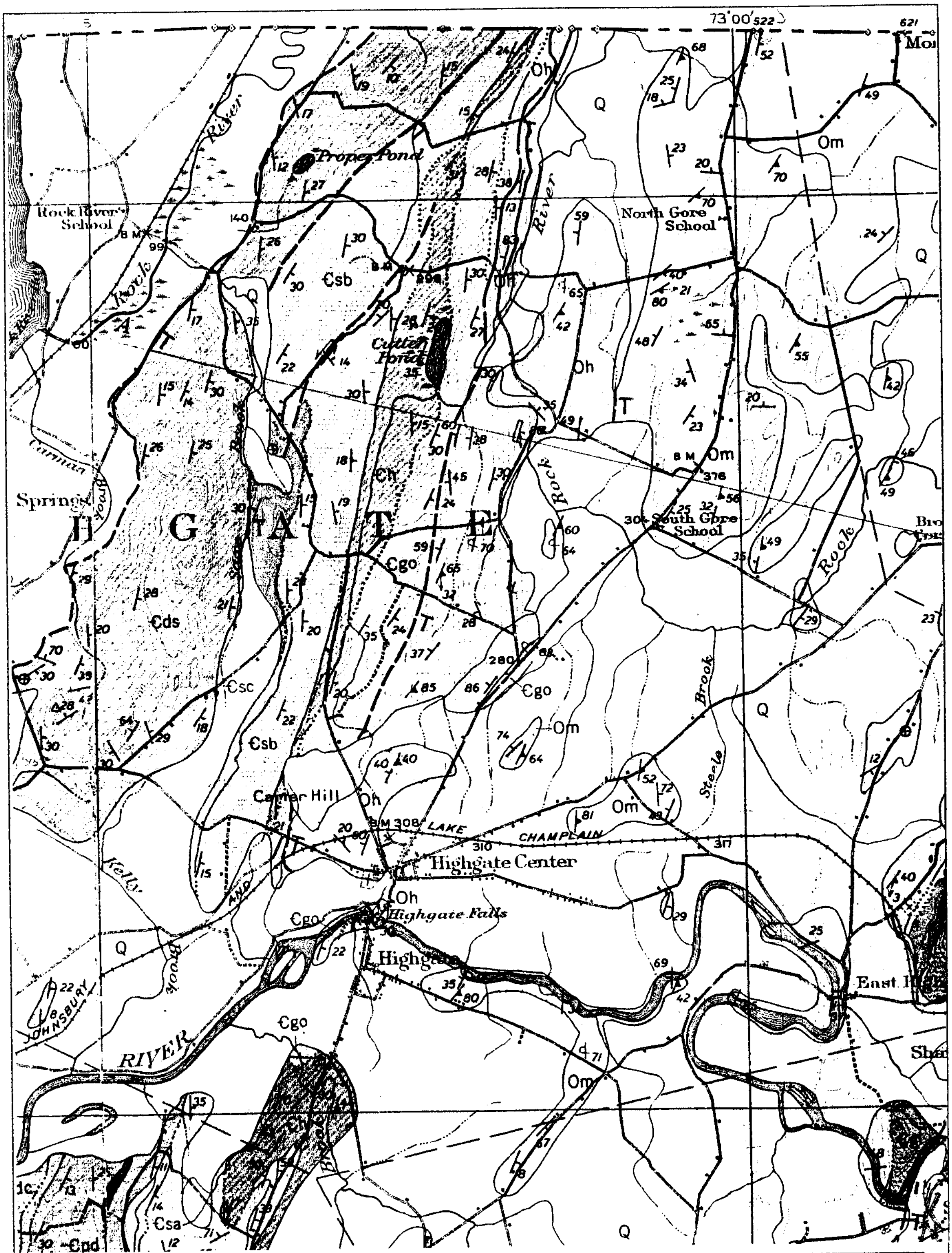
GENERALIZED GEOLOGIC MAP OF VERMONT

1970

SCALE

4 0 4 8 12 16 MILES

Adapted from the Centennial Geologic Map of Vermont by Charles G. Doll, State Geologist. Vermont Geological Survey



Legend Occ = Corliss conglomerate member Om = Morses Line Slate Oh = Highgate Formation
Cgo = Gorge Formation Ch = Hungerford Slate Csb = Saxe Brook Dolomite
Csc = Skeels Corners Slate Crl = Rockledge conglomerate Cm = Mill River conlomerate
Csa = St. Albans Slate Crb = Rugg Brook Dolomite Cp = Parker Slate
Cd = Dunham Dolomite (Cds - sandy facies, Cdc - carbonate facies) Faults —— T = Thrust

Figure 1.5.: A portion of Shaw's [1958] geologic map of the St. Albans area, Vermont.

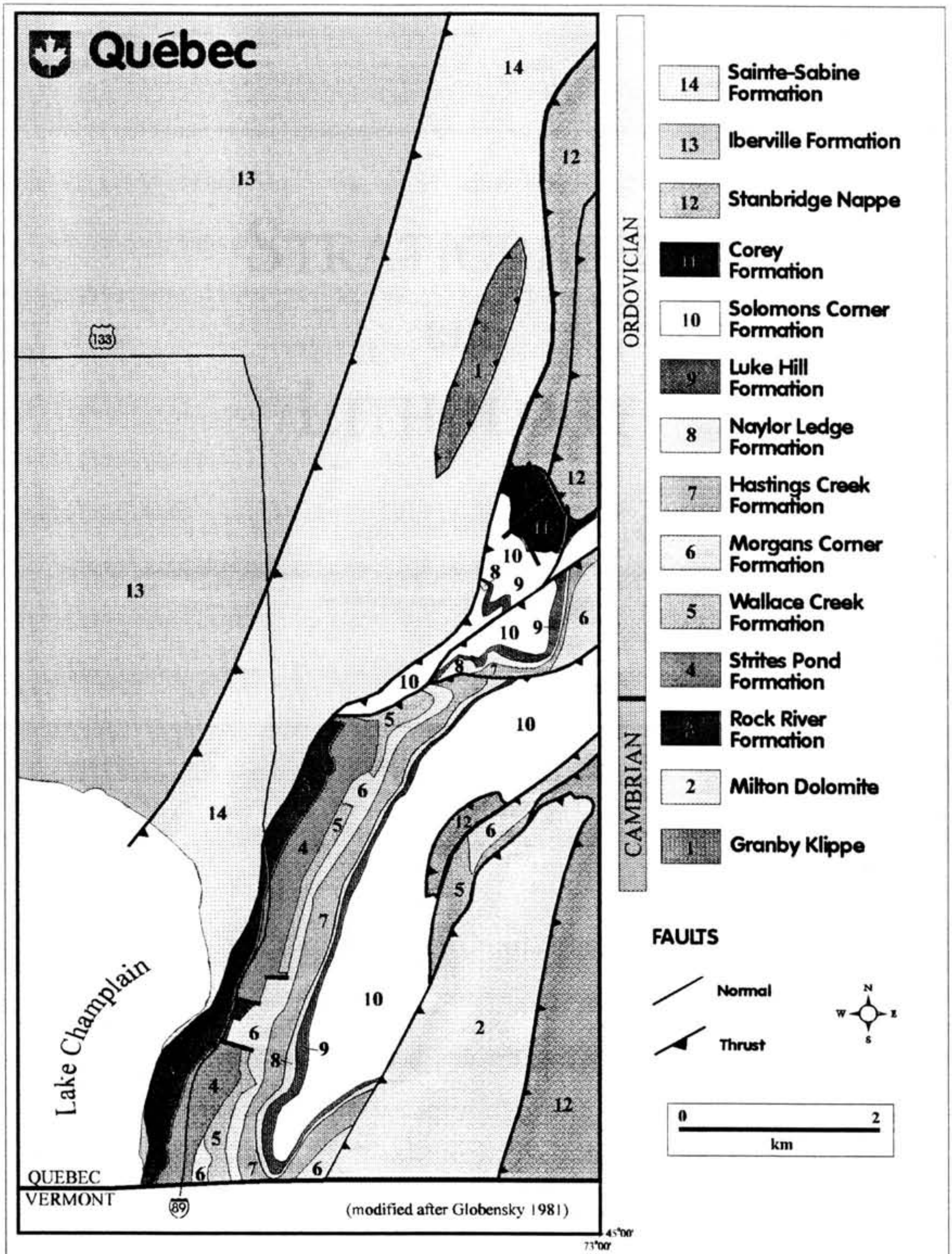
0 1 km

Recent structural and stratigraphic results from Québec (***Figure 1.6.***) are available from Québec's *Ministere de L'Energie et des Ressources.* Avramtchev's [1989, 1:250,000] "Geologic Map" of southern Québec together with Globensky's [1981, 1:63,360] and Charbonneau's [1980, 1:50,000] detailed reports about the geology adjacent to northwestern Vermont represent the most recent Canadian information.

The different stratigraphic rock units in the Champlain lowlands are traditionally separated by trilobite biostratigraphy [Shaw 1958] and their lithologic character into shallow water carbonates ("western shelf sequence") and deeper water sediments of the continental slope or rise ("eastern basinal sequence"; [Mehrtens & Dorsey 1987]; ***Figure 2.4.***).

Schuchert	Highgate Center Quad., St.Albans Quad., Georgia Plains Quad., Milton Quad.	1937	1:200,000
Shaw	Highgatecenter Quad., St.Albans Quad.	1958	1:62,500
Cady	Northern Vermont & Southern Québec	1960	1:487,000
Doll et al.	Centennial Geologic Map of Vermont	1961	1:250,000
Stone & Dennis	Milton Quad.	1964	1:62,500
Charbonneau	Region de Sutton, *Québec*	1980	1:50,000
Globensky	Regions de Lacolle, Saint-Jean(s), *Québec*	1981	1:63,360
Mehrtens & Dorsey	Highgatecenter Quad., St.Albans Quad.	1987	1:12,000 & 1:24,000
Avramtchev	Basses-terres du Saint-Laurent et Estrie-Beauce, *Québec*	1989	1:250,000
Mehrtens & Borre	Colchester Quad., Georgia Plains Quad.	1989	1:12,000 & 1:24,000
Haschke	**THIS STUDY**: Highgate Center Quad., St.Albans Quad.	**1994**	**1:22,000**

TABLE 1.1.: Previously published reports and maps of the Champlain Valley in northwestern Vermont and adjacent Québec *(see **Figure 1.2.**)*



Geologic Map of Southern Quebec
Region des Basses-Terres du Saint Laurent

Figure 1.6.

Graphics by Michael R. Hoschke 12/94

STRATIGRAPHY
and
LITHOLOGY

"...The problem of correlation of rocks north and south of the international boundary
is chiefly one of terminology..."

Wallace M. Cady 1960

2.1. INTRODUCTION

The thrust-imbricated belt of the Champlain Thrust System displays a range of continental passive margin lithologies and comprises shallow water deposits of the shelf region as well as deeper water sediments of the continental slope and/or rise. The thrust faulted nature of the study area causes a restricted exposure of the successive change across the passive margin during the Cambro-Ordovician in northwestern Vermont as a result of its underthrusting the Taconic Allochthon.

The Centennial Geologic Map of Vermont [Doll et al. 1961, ***Figure 1.4.***], and subsequent investigation [Mehrtens & Dorsey 1987], summarizes the general structural outline provided by previous workers, but lacks information on essential structural details, and on features which are necessary to understand the structural arrangement and assembly of this area.

Each thrust slice within the Champlain Thrust System consists of various lithostructural domains of the shallow water carbonate succession and comprises distinctive suites of lithologies. Units within these individual domains, in particular within the Rosenberg sequence and the adjacent deeper water sediments of the Morses Line Formation were traditionally separated by trilobite diversity and their lithologic character into "western shelf" deposits and "eastern basinal" facies [Mehrtens & Dorsey 1987]. Previous sedimentologic models [Mehrtens 1985] of this area suggest a rapid transition and eastward deepening from shallow water calcarenites of the shelf edge region into slaty limestones of the continental slope and rise.

Lithology-descriptive mapping of a representative suite of outcrops, corresponding to each unit, provided the basis for this study. Lithologic criteria of dolostone, limestone and shale/slate units were utilized in order to try to distinguish the apparent abundance of local formations that were established by previous workers in northwestern Vermont.

Most important, my investigation of the applied lithologic criteria revealed a considerable overlap of several units. In particular the separation of dolostone and shale/slate units into apparently distinguishable and characteristic *formations* could not be accomplished using the criteria employed by previous workers. Moreover, Canadian workers in southern Québec proposed a lithostructural stratigraphy of the region adjacent to the study area that differs significantly from the one established in northwestern Vermont. In particular the definition of the *Morses Line Slate* as being a broadly undeformed and unfaulted unit [Mehrtens & Dorsey 1987] strongly contrasts with the northern continuation of this sequence into Québec, where it is described as allochthonous *Stanbridge Nappe*. The interpretation of a tectonic contact between the Rosenberg sequence and the Morses Line Formation/Stanbridge Nappe, however, modifies the stratigraphy dramatically. It is therefore of particular interest to clarify the question:

Is the contact between the Rosenberg sequence, that is the Highgate Formation in this area, and the Morses Line Formation/Stanbridge Nappe a transitional, stratigraphically intact or a faulted one ?

If this contact is faulted, how would this relation modify the "rapid lateral facies change"- hypothesis of Mehrtens and Dorsey [1987] ?

2.2. HIGHGATE SPRINGS SEQUENCE — LOWER-MIDDLE ORDOVICIAN

The Highgate Springs slice (***Figure 2.1.***) extends for about 134 km length and 1.3 km width between north of St. Dominique in Québec (Bagot County) and southwest of St. Albans, Vermont [Kay 1958]. This folded and partly recrystallized (marbleized) carbonate succession, first termed *Highgate Springs Series* (consisting of Trenton, Black River and probably Chazy formations) by Schuchert [1937], corresponds to the lower unit [Division E] of the *Québec Group* of Logan [1863] and with the *Highgate Springs-St. Dominique slice* of Cady [1960]. Both terms are based on regional observations and cannot account for its tectonic significance in Vermont and Québec. This study suggests ***Highgate Springs slice*** as a ***tectonically descriptive*** term to represent the facies of the Highgate Springs sequence, and therefore follows Kay's [1958] lithostructural outline.

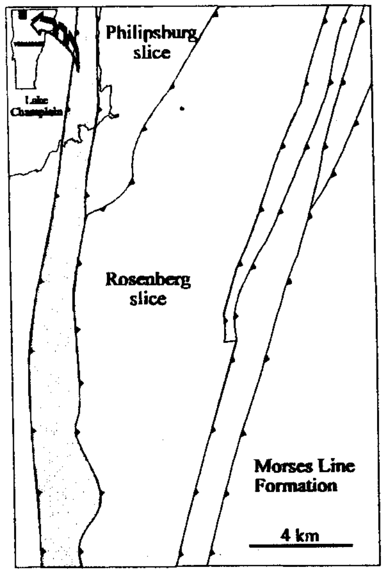


Figure 2.1.: Highgate Springs slice, northwestern Vermont.

The carbonate-shale succession rests on the Highgate Springs thrust fault, representing the most western branch of the Champlain Thrust System in the study area, which separates it from the relatively gently deformed rocks of the structural foreland along Lake Champlain. The internal structure of the Highgate Springs sequence is complicated by multiple thrust faulting and folding within the carbonate and shale units. Therefore, the structural condition of this sequence causes strongly varying thicknesses of all stratigraphic units (***Tab 2.1.***).

The majority of the rocks are exposed in and around the village of Highgate Springs along a number of linear ridges separated by meadows, interpreted as cross-faulted limbs of asymmetrical folds with east-dipping axial planes [Kay 1958]. The ridges show chiefly massive dolomitic limestone, partly interstratified with dolomite bands, whereas the meadows are underlain by thin-bedded argillaceous limestone and shaly argillite. Most detailed studies of this sequence are provided by Kay [1958] and Cady [1960].

Stratigraphic Unit	Logan 1863	Kay 1958
Hortonville Formation	-	**6**
Glens Falls Limestone	**61-122**	**30**
Isle La Motte Limestone	**9**	**21-69**
Youngman Formation	*>18*	**54-**>*91*
Carman Sandstone	**15-30**	**12-37**
Beldens Formation	-	**76-***213*

[all thicknesses in meters] **measured** *estimated*

TABLE 2.1.: Correlation of unit thicknesses of the Highgate Springs Sequence.

Rocks like those at Highgate Springs are recognized in scattered exposures southward through Swanton and St. Albans Bay. Most are isolated small areas exposing one single formation. *Fonda Quarry Hill*, northwest of St. Albans, is considered the best exposure for the nature of the structure of the Highgate Springs sequence and in displaying the contact with the overthrusted Lower Cambrian rocks of the Rosenberg slice [Kay 1958].

2.2.1. BELDENS FORMATION — LOWER ORDOVICIAN

The *Beldens limestone*, named from a locality north of Middlebury [Cady 1945], generally forms conspicuous as well as persistent exposures within the Highgate Springs sequence and is typically a grey-buff weathering, buff-white-grey micritic and well bedded (0.5-2 m), calcitic and quartz-arenitic limestone with interstratified buff-weathering dolomite bands ([Kay 1958]; outcrop # 397, 398, 403, 411, 412, 463). The established gastropod- and brachiopod-stratigraphy of Cady [1945] and Kay [1958] reveals the carbonates of the Beldens Formation as being the oldest stratigraphic unit within the Highgate Springs slice. Some localities show fractured limestone containing calcite veins (1.5 cm, outcrop # 424, 416) without preferred orientation. Rare sections of pure limestone were quarried in *Fonda Quarry,* northwest of St. Albans.

2.2.2. CARMAN SANDSTONE — MIDDLE ORDOVICIAN

The Carman quartzite has its type section at the head of Shipyard Bay in the Highgate Springs area [Kay 1945] where it exposes thin- to medium-bedded quartz-arenites. It also appears with intercalated minor argillaceous and calcarenite layers towards the bottom and the top of the unit [Kay 1958]. Best exposures of the Carman quartzite in the study area are available at Fonda quarry.

2.2.3. YOUNGMAN FORMATION — MIDDLE ORDOVICIAN

At the type locality along the shoreline of Lake Champlain at Highgate Springs (outcrop # 462), the Youngman limestone exposes argillaceous limestone and argillite [Kay 1945, 1958]. Many scattered, smaller exposures of the formation also comprise cleaved calcareous argillite.

Considerable confusion still exists about some single members of the Youngman Formation. The *St. Dominique limestone* in Québec was described as a sandstone member as well as a limestone unit in the same section [Clark 1955]. Moreover, Kay [1945] suggested that the St. Dominique limestone and sandstone can be designated to the *Carman Formation* as well as to the Youngman Formation. The term *Youngman Formation* [Kay 1958] is therefore accepted in this study as comprising both the sandstone and the limestone subunit of the St. Dominique in the Highgate Springs area.

2.2.4. ISLE LA MOTTE LIMESTONE — MIDDLE ORDOVICIAN

The *Isle la Motte limestone* was named by Emmons [1842] from the island town of Grand Isle County, northwestern Vermont. Its typical lithology at the village of Highgate Springs comprises thick-ledged, medium to thin-bedded, and black chert-bearing, argillaceous limestone, locally interstratified with calcarenites [Kay 1958]. Best exposures are located at St. Dominique in Québec within a lime quarry, and northwest of Fonda quarry, northwest of St. Albans.

However, some uncertainty still exists about the criteria of differentiating it from the argillaceous limestones of the Youngman Formation.

2.2.5. GLENS FALLS LIMESTONE — MIDDLE ORDOVICIAN

The Glens Falls Limestone was initially named by Ruedemann [1925] but first defined by Kay [1937] and described by Cady [1945] as a uniformly thin-bedded and black shaly limestone with minor interstratified calcarenite beds in the Highgate Springs area. Its exposure was described as being rather constant throughout the Highgate Springs slice as well as the area east of Lake Champlain [Kay 1958]. According to previous workers [Keith 1932; Cady 1945; Kay 1958], the Glens Falls Limestone is generally overlain by the *Hortonville Shale*, that is described as a rarely exposed laminated black argillaceous shale along the shoreline of Lake Champlain in Highgate Springs village. This stratigraphic order disagrees with the established stratigraphic succession of the Highgate Springs sequence on the State Map of Vermont [Doll et al. 1961], where homogeneously thin-bedded limestones and limestone-bearing black shales of the *Cumberland Head Formation* overly the Glens Falls Formation. According to Doll and others [1961], however, exposures of the Cumberland Head Formation are restricted to the island of Grand Isle in Lake Champlain. My own examination of the Lake Champlain shoreline in Highgate Springs village revealed dark grey thin-bedded shaly limestones and calcareous shales, that indicates a strong lithologic similarity with the Glens Falls Formation. This study, therefore, rejects the term "Hortonville Shale" from the Highgate Springs sequence and favors a designation of the shaly limestones to the Glens Falls Formation.

The Glens Falls Formation can be traced southward into west-central Vermont, West Haven area, where Steinhardt [1983] mentioned it as a dark grey thin-bedded limestone with interbedded shale partings.

2.3. PHILIPSBURG SEQUENCE LATE CAMBRIAN-EARLY ORDOVICIAN

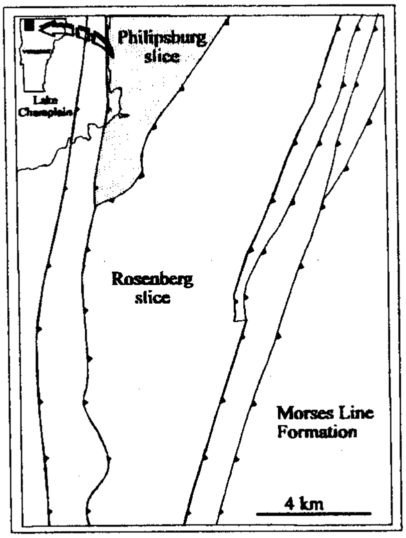


Figure 2.2.: Philipsburg slice, northwestern Vermont.

The Philipsburg slice (***Figure 2.2.***) extends from east of Highgate Springs to south of Bedford in Québec. In Vermont, most of it forms a gently east-dipping homoclinal sequence, about 3.2 km wide [Cady 1960], that can be traced across the international border [Globensky 1981]. This succession comprises arenitic limestones and dolostones, ranging in age from late Cambrian to early Ordovician. Billings' [1861] original "two-unit" classification (***Tab.2.2.***) of this sequence was regrouped by Logan [1863] into **divisions A, B, and C.** In northwestern Vermont, McGerrigle's [1931] revision and subdivision of these units into 10 formations appear in the subsequent literature [Clark 1934] but lacks an adequate definition until Clark and McGerrigle [1944] established a reasonable stratigraphic nomenclature (***Tab.2.2.***). Further previous work in the Vermont-portion of the Philipsburg sequence has been conducted by Keith [1932] and Cady [1945]. The term *Philipsburg slice* for this Cambro-Ordovician succession is an adequate tectonically descriptive term used in this study and replaces former local names.

PERIOD	Epoch	Units of Billings (1861)	Divisions of Logan (1863-64)	Formations of Clark & McGerrigle (1944)	Formations of Globensky (1981)
Ordovician	Middle	Blue laminated and nodular limestone	D_2D_3	Stanbridge Slates	Stanbridge Formation
			D_1	Mystic Conglomerate	
	Early		C_2	Basswood Creek Slates	Corey Formation
			C_1	Corey Limestone	
			B_5	Saint-Armand Limestone	Solomons Corner Formation
			B_4	Solomons Corner Limestone	
			B_3	Luke Hill Limestone	Luke Hill Formation
			Upper B_2	Naylor Ledge Limestone	Naylor Ledge Formation
			Most of B_2 + B_1	Hastings Creek Limestone	Hastings Creek Formation
		Mg-Limestone	A_3	Morgans Corner Limestone	Morgans Corner Formation
				Wallace Creek Limestone	Wallace Creek Formation
Cambrian	Late		A_2	Strites Pond Limestone	Strites Pond Formation
			A_1	Rock River Dolomite	Rock River Formation Member C, B, A (Member A equivalent to Rock River Dolomite)

Table 2.2.: Correlated stratigraphic nomenclature of the Philipsburg sequence and the Stanbridge Nappe in southern Quebec (modified after Globensky 1981).

2.3.1. ROCK RIVER FORMATION UPPER CAMBRIAN

The Rock River borders the Philipsburg sequence along a thrust fault contact adjacent to the Rosenberg slice and provided the local term [Clark & McGerrigle 1932] for the type-locality (1 km southeast of the village of Philipsburg in Québec) of an Upper Cambrian limestone-dolomite within the northeast plunging syncline structure of the Philipsburg slice.

This formation comprises predominantly arenaceous dolomite and, locally, calcarenites. Globensky's [1981] subdivision of the Rock River Formation into individual lithologic members (***Tab.2.2.***; A, B, C) was not applicable to the Vermont portion of this formation. Instead of a dolomitic arenite with interstratified siltstones at the base (member A), a micritic calcarenite in the middle (member B) and a massive arenaceous dolomite on top (member C, [Globensky 1981]), this formation exposes rather thick-bedded pigeon-grey to tan-colored dolomitic limestones with interstratified pale quartz-arenaceous dolomite beds, partly as intraformational conglomerates (outcrop # 342) and green-grey argillaceous calcarenite laminations (outcrop # 176, 342, 347, 396, 389, 391, 392).

Scattered (secondary) chert occurrences are not restricted to the base of the formation [Globensky 1981], but rather are present in small abundances throughout this unit of the Philipsburg sequence. Several minor normal fault occurrences within the Rock River Formation (outcrop # 343) together with its location adjacent to and within a severely deformed thrust belt complicate the determination of stratigraphic thicknesses and relations of this unit in the area.

2.3.2. STRITES POND FORMATION LOWER ORDOVICIAN

Strites Pond is located 2 km southeast of the village Philipsburg in Québec where its western and southern ledge represent the type locality for this partly quartz-arenaceous micrite-dominated formation. Several localities have been quarried by the *"Missisquoi Stone and Marble Company"* in Québec and are exposed as old quarries [Globensky 1981].

Well-bedded pigeon-grey to whitish grey micrite and marble intercalated with quartz-arenite laminations dominate the composition of this formation. Most outcropping limestones show characteristic raised reticulate lines of mat-white dolomite on the weathered surface (outcrop # 314, 364, 365), which separates it from other arenaceous limestones within the Philipsburg sequence. Without this significant feature, it could not be separated from the pigeon-grey micritic portions of the Rock River Formation. The dolomite reticulations are typically associated with spectacular algal laminations and stromatolite structures on weathered surfaces (outcrop # 338).

The Strites Pond Formation has been correlated to the *Shelburne Formation* in west-central Vermont [Globensky 1981] as well as to the *Beauharnois Formation* within the Beekmantown Group in the "*Basses Terres of Saint Laurent/Québec*" [Clark & Eakins 1968].

2.3.3. WALLACE CREEK FORMATION — LOWER ORDOVICIAN

The Wallace Creek Formation, named after the north-south creek near Strites Pond in Québec, is characterized by the increased abundance of interstratified calcareous black shale bands within the pigeon-grey limestones. It is commonly known as *black marble* (*"marbre noir"* in Québec) due to its black-grey mottled appearance [Globensky 1981].

The majority of the formation displays well-bedded argillaceous micrites; nevertheless minor intraformational conglomerates consisting of angular pigeon-grey limestone clasts have been observed.

2.3.4. MORGAN CORNERS FORMATION — LOWER ORDOVICIAN

The Morgans Corners type locality, characterized by the hill about 500 m east of the village of Philipsburg in Québec [Globensky 1981], exposes massive dark-grey microcrystalline dolomite. As for all other formations of the Philipsburg sequence, a type locality was missing until the work of Clark and McGerrigle [1946]. Minor conglomeratic zones are present throughout the Morgan Corners Formation.

The abundance of tiny crevasses on weathered surfaces of the dolomite has been proposed as a distinguishing feature of this formation [Globensky 1981], although this criterion is only applicable in Vermont to restricted exposures along the international border (***Plate 1***, marker 629E-629D). This study applies the occurrence of rather *black-grey massive dolomitic* facies as a reasonable criterion of subdividing it from the underlying Wallace Creek Formation.

2.3.5. HASTINGS CREEK FORMATION — LOWER ORDOVICIAN

The type locality of the Hastings Creek Formation (500 m north of the route between Philipsburg and Saint-Armand-Station within the Hastings Creek valley) varies between mostly white to black thick-bedded argillaceous limestones and algal-bearing dolomitic lithology [Globensky 1981].

Earlier workers [McGerrigle 1931, Clark & McGerrigle 1932, Clark 1934] mentioned the Hastings Creek Formation and several other units of the Philipsburg sequence without defining either their type localities or their lithologic characteristics. All Hastings Creek carbonate exposures are restricted to a small area along the international border (***Plate 1***, marker 629B-628) and represent the youngest unit within the Vermont portion of the Philipsburg syncline.

2.3.6. SUMMARY AND DISCUSSION — PHILIPSBURG SEQUENCE

The wedge-shaped Philipsburg thrust slice comprises early Ordovician argillaceous and arenaceous limestones and dolomites, forming a gently northeast-plunging syncline. This structure was created and additionally complicated by the overthrusting of the Rosenberg thrust slice.

The overlying younger limestone formations (Naylor Ledge-, Luke Hill-, Solomons Corner- and Corey limestone Formations) were not observed within the U.S.-portion of the Philipsburg sequence (***Tab.2.3.***), but are well documented in adjacent Québec [Charbonneau 1980, Globensky 1981]. This reflects the gently north-northeast-plunging nature of the syncline, across the border, which caused the uppermost formations to "pinch out" before entering the Vermont area. Most exposures are restricted to the gently east-dipping, western limb of the syncline whereas the almost vertical dipping eastern limb of the syncline cannot be entirely documented in the Vermont area due to the lack of outcrops within the Rock River swamp area.

FORMATION	McGerrigle 1931	Clark & McGerrigle 1944	Gilmore 1971	Globensky 1981	**Haschke 1994** -This study-
Corey Limestone	91-114	84	-	80-90	-
Solomons Corner Fm.	?	152	222	155-185	-
Luke Hill Fm.	49	49	-	60	-
Naylor Ledge Fm.	9	9	-	-	-
Hastings Creek Fm.	75	96	-	78	137
Morgan Corner Fm.	46	46	71	-	138
Wallace Creek Fm.	-	61-76	35-44	40-55	62
Strites Pond Fm.	-	-	32-72	128	103
Rock River Fm.	-	153	128	143-157	169 ?

TABLE 2.3.: Stratigraphic thicknesses of the Philipsburg sequence.
[Values in this study calculated from cross-section; thicknesses in meters]

2.4. ROSENBERG SEQUENCE LOWER CAMBRIAN-LOWER ORDOVICIAN

The Rosenberg sequence (***Figure 2.3.***), entirely Cambro-Ordovician in age, was named from the village of Rosenburg (erroneously Rosenberg) in Québec by Clark [1934], who briefly described the Canadian section of this siliciclastic carbonate and shale thrust slice. It is bordered to the west by the *Champlain Thrust* fault branch of the Champlain Thrust System and to the east by the contact with the tectonically juxtaposed allochthonous Morses Line Formation. The Rosenberg succession in its entirety comprises a quartz-arenitic dolostone assemblage together with its intraformational conglomerate facies (*"carbonate-quartzite assemblage"*, [Cady 1960]), and is interstratified with shale and siltstone units of Cambrian age. The top of the Rosenberg sequence consists of early Ordovician argillaceous slaty limestones. Considerable effort has been made throughout the last decades in classifying multiple dolomite, arenite and "slate" subunits [Schuchert 1937, Shaw 1958, Stone & Dennis 1964, Mehrtens 1985, Mehrtens & Dorsey 1987, Mehrtens & Borre 1989] that has caused tremendous confusion about the stratigraphic nomenclature (***Figure 2.4.***) within this thrust slice. All subunits are well equipped with regional stratigraphic names but lack applicable and satisfying lithological field criteria to justify the recently proposed [Mehrtens & Dorsey 1987] subdivision of this sequence.

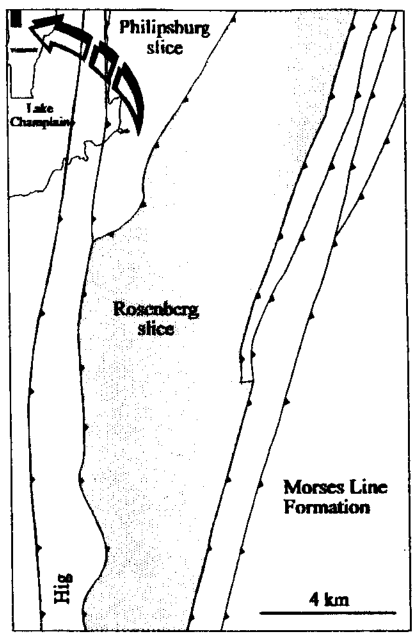


Figure 2.3.: Rosenberg slice, northwestern Vermont.

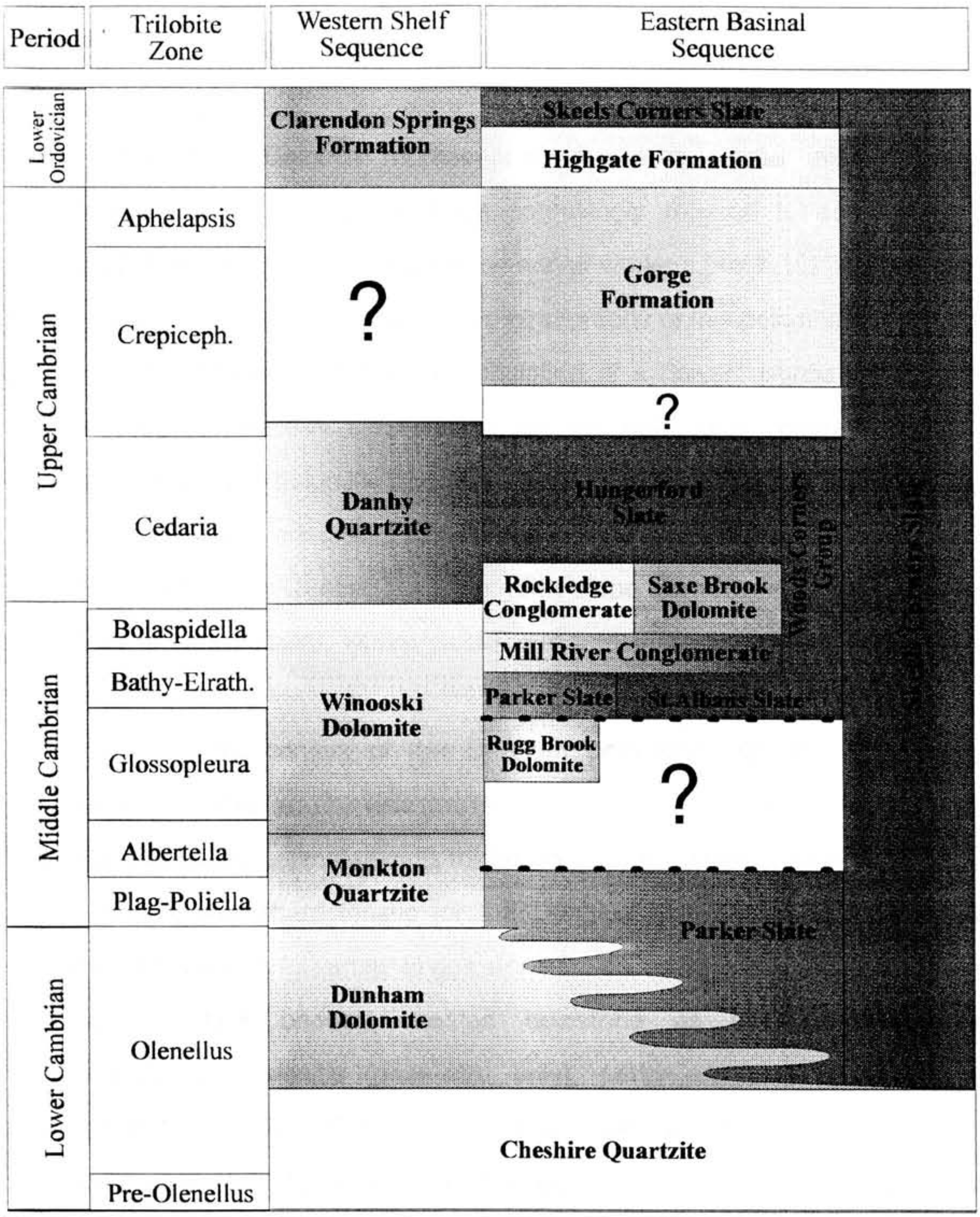


Figure 2.4.: Correlation chart for stratigraphic units of the Rosenberg sequence of northwestern and west-central Vermont.
Ages of the western shelf sequences are approximated and based on intertonguing relationships the eastern basinal shales.

Dolomite
Limestone
Quartzite/ Sandstone
Shales
Unconformity

DATA SOURCES:
Shaw 1958
Doll et al. 1961
Palmer 1971
Dorsey et al. 1983
Landing 1983
Mehrtens & Gregory 1984
Mehrtens 1985

(modified after Mehrtens & Dorsey 1987)

Graphics by Michael R. Haschke 11/94

2.4.1. DUNHAM DOLOMITE — LOWER CAMBRIAN

The Dunham Dolomite represents the lowest dolostone unit within the Rosenberg succession and has been confusingly referred to as *"Swanton Mosaic Marble"* and *"Winooski Marble"* by earlier workers [Keith 1923, Schuchert 1937]. Cady [1945] established a revised stratigraphy of this dolomite-dominated portion of the Rosenberg sequence consisting of a "lower" *Dunham Dolomite* and an "upper" *Winooski Dolomite*, which are commonly separated in the Burlington area by a dolomitic quartzite unit of strongly varying thickness, the *Monkton Quartzite*. Clark [1936] initially defined the type locality of the *Dunham Dolomite* at Oak Hill in Québec (14.5 km north of the international border, 4.5 km southeast of Cowansville), where calcareous and dolomitic marble is exposed [Charbonneau 1980].

The tectonic contact of the Dunham dolostone with the Philipsburg sequence, or (further south) with the Highgate Springs sequence, respectively, prevents reasonable estimates of its thickness in northwestern Vermont.

Lithologically characteristic for the Dunham Dolomite in the map area (***Figure 2.5.***) is a colorful variety of dolostone facies, ranging in lithology between light-grey-tan thick bedded, mottled dolostone as well as dolostone conglomerate and arenitic (grain-size only), partly recrystallized dolostone (outcrop # 308, 310, 312, 421, 423) to dolomitic quartz-arenite (outcrop # 159, 161, 171, 172, 395), and purple-white thin wavy laminated, argillaceous, fine crystalline dolomite marble (outcrop # 160, 309, 311). Quartz-arenaceous portions within the Dunham lithology may be be observed but are restricted to patches within outcrop scale. However, the major portion of the Dunham Dolomite generally lacks quartz-arenite grains, in contrast to the overlying Dolomite-Shale sequence.

Figure 2.5.: **Orange-buff fine-crystalline dolomite of the Dunham Dolomite at outcrop # 311, 1.5 km southeast of Swanton along I-89.**

A north-south trending belt of, partly cleaved, shale and siltstone strata separates the Dunham Dolomite from the overlying *Dolomite-Shale Sequence*, but terminates towards the international border.

Several outcrops within this distinctive lithology display intensively fractured dolomite that is cross-cut by quartz-veins (0.2-1.5 cm thickness).

2.4.2. DOLOMITE-SHALE SEQUENCE LOWER-MIDDLE CAMBRIAN

Considerable confusion exists about the Cambro-Ordovician stratigraphic nomenclature of this sedimentary unit that immediately overlies the Dunham Dolomite. In contrast to the previous studies of the Highgate and St. Albans area [Shaw 1958, Mehrtens & Borre 1987, Mehrtens & Butler 1989], which separated several lithologically identical quartz-arenaceous dolostones (*Saxe Brook-*, *Rugg Brook* Dolomite, *Danby* Formation of the "western shelf sequence") and slates (*Parker-*, *Skeels Corners-*, *Hungerford-*, *Russell-*, *St.Albans* Slate of the "eastern basinal sequence") within the Rosenberg succession, I apply the term *"Dolomite-Shale sequence"* to a variety of folded argillites, shales and siltstones as well as argillaceous dolostones and dolomitic arenite lithology (outcrop # 430, 431; ***Figure 2.6. & 2.7.***). The <u>dominance</u> of quartz-arenaceous dolostones distinguishes this strata from the underlying Dunham Dolomite.

A substantial lithological break crosscuts the homoclinal Rosenberg sequence lithologically as well as stratigraphically, so that a separate consideration of a northern and a southern gently east-dipping subsequence is required. The lithology of the ***northern Dolomite-Shale sequence*** resembles portions of the Dunham Dolomite but contains significantly more quartz-sand and is intercalated with lenticular dolomitic shale units (outcrop # 401).

Figure 2.6.: Dolomite-Shale sequence at outcrop # 430, 2.25 km southwest of Skeels Corners. Non-calcareous shales conformably overlying thin-bedded quartz-arenaceous dolostone layers.

0 30
cm

Figure 2.7.: Southern Dolomite-Shale sequence at outcrop # 431, 2.2 km southwest of Skeels Corners. Non-calcareous, folded and weakly cleaved shales with interstratified thin dolostone laminations.

0 10
cm

Towards the west, a minor thrust fault [Shaw 1958] juxtaposes siltstones and partly fissile black to grey limonite-stained micaceous dolomitic shales (outcrop # 40, 44, 46, 49, 50, 59, 61, 68, 71) with quartz-arenaceous dolostone lithology. This dolomitic shale interval can be traced northward almost all the way to the international border. Included within this dolomitic shale unit are discontinuously lenticular quartz-arenaceous dolostone and dolomitic limestone conglomerate bodies (outcrop # 55, 56, 60) of regionally limited extent (3-5 m thickness, 10-400 m length).

The ***southern Dolomite-Shale sequence*** ranges in thickness between 850 m to 1100 m and is characterized by its clear shale-dominance. Previous authors [Howell 1939, Shaw 1958, Cady 1960, Dennis & Stone 1961, Mehrtens & Dorsey 1987] referred to this fossiliferous sequence as *Parker-, Skeels Corners-, Hungerford-, Russell-* or *St.Albans Slate* ***(Figure 2.4.)***.

Howell [1939], who initially introduced the term ***Skeels Corners Slate***, established its type locality "half a mile northwest of Skeels Corners and 6 miles north of St.Albans", where non-calcareous black shales are exposed. According to Howell [1939] and Shaw [1958], the shale sequence is typically intercalated with dolostones, limestones and quartz-arenite bodies. This wide variation in the lithology of the Skeels Corners Slate causes a broad overlap of lithologies of the previously separated shale units [Hungerford-, Parker-, Russell-, St.Albans Slate]. All shales "characteristically" show interstratified "white to brown limonite-stained quartz sandstone laminations, less than 1mm thick, in a endless succession throughout the slates" ([Shaw 1958]; ***Figure 2.7.***).

Schuchert [1937] proposed the name ***Hungerford Slate*** for a type locality at "Hungerford Brook, 1.25 miles south of Highgate Center, where the brook crosses the Highgate road", where Upper Cambrian "black slates with interstratified alternating *beds of fine-grained white sandstone are exposed"* [Shaw 1958]. Both descriptions, the Skeels Corners Slate and the Hungerford Slate, define a very similar lithologic image and lack reasonable lithological criteria for clearly distinguishing these units.

Ruedemann's [1947] attempt of clarifying the stratigraphy, using the variations of the graptolite content in the shales, could not resolve this problematic lithostratigraphy, but resulted in extending the stratigraphic nomenclature by introducing the term ***Russell Slate***. The Russell Slate itself, however, was not accepted by subsequent workers [Shaw 1958].

The ***St.Albans Slate*** was introduced by Howell [1929] as a Middle Cambrian, black to grey-black cleaved micaceous slate. Its type locality *"at the western edge of St.Albans, north of the road from St.Albans to St.Albans Bay"* [Howell 1929], strongly resembles the *"grey to black micaceous dolomitic and arenaceous Parker Slate at its type locality, 2.25 miles northwest of Georgia Center in the Milton quadrangle"* [Shaw 1958]. According to Schuchert [1937], *"the St.Albans Slate is, if unfossiliferous, hard to distinguish from the Parker Slate"*. The ***Parker Slate***, however, was designated by previous authors to the *Colchester Formation* [Keith 1923] and to the *Georgia Slate* [Walcott 1891].

All other previous workers [Schuchert 1937, Shaw 1958, Mehrtens & Dorsey 1987] accepted this tremendously confusing stratigraphic nomenclature (***Figure 2.5.***) without questioning the criteria for distinguishing those units. Earlier lithostratigraphic reports already mentioned the *"difficulty of separating the Skeels Corners Slate from the Parker- and the Hungerford Slate"* [Shaw 1958]. This study could not apply earlier proposed criteria for distinguishing the shales and therefore summarizes this variety of lithologies as a *Dolomite-Shale sequence* (***Figure 2.8.***).

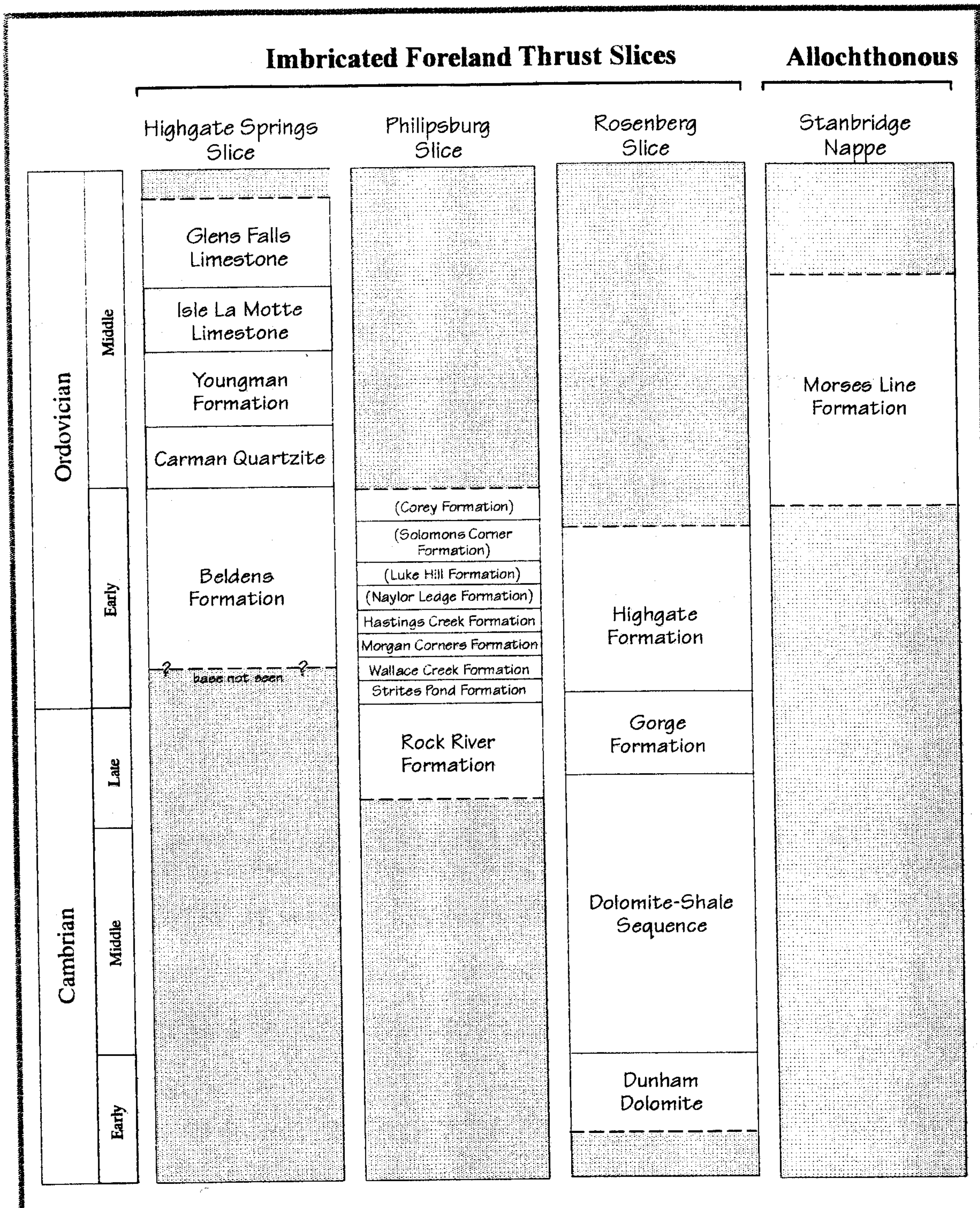


Figure 2.8.: Stratigraphy of the Cambro-Ordovician thrust slices in the Taconic foreland sequence in the St.Albans and Highgate area, northwestern Vermont (this study). Additional units of the Philipsburg slice (in parenthesis) that occur in Quebec only.

Within the southern Dolomite-Shale assemblage, north-south trending, discontinuous lenticular quartz-arenaceous dolostone and quartz-arenite units of considerable extent (5-20 m thickness, 10 m - 4 km length) are episodically intercalated within the folded and generally weakly cleaved dark-grey to black-brown, non-calcareous shale/slate strata (outcrop # 240, 248, 264, 270, 287, 289, 294, 418). In contrast to the northern sequence, the dolostone and quartz-arenite lithologies are only represented by occasionally intercalated lenticular bodies (outcrop # 260, 261, 263, 266, 268, 283, 286, 295, 296, 297), where exposed.

The largely undeformed and non-micaceous shale beds underlying the quartz-arenaceous dolostone breccia of the *Gorge Formation* (***Figure 2.9.***) are not found within the southern sequence of the Rosenberg slice. It is possible that a substantial structural break exists between both subsequences and the two stratigraphic sequences are tectonically juxtaposed. However, the occurrence of white-grey to purple Dunham dolomite lithology below the southern sequence (outcrop # 310, 311, 312, 421) clearly resembles the dolomite exposures observed at the base of the northern sequence (outcrop # 160, 224, 225). This identity together with the proposed continuity of the Gorge and Highgate units above, on the other hand, suggests similar stratigraphic levels and sources of both sequences.

2.4.3. GORGE FORMATION — UPPER CAMBRIAN - LOWER ORDOVICIAN

The Gorge Formation comprises lithologically characteristic massive beds (0.3-1.5 m thickness) of dark-grey, quartz-arenaceous dolomite breccia (outcrop # 23, 78, 100, 137; ***Figure 2.10.***), initially designated as *"Missisquoi Formation"* ([Raymond 1925], ***Figure 2.11.***). The unit commonly forms continuous, largely north-south trending ridges between the international border and St. Albans and

0 50
cm

Figure 2.9.: Contact between southern Dolomite-Shale sequence and Gorge Formation at outcrop # 243. Note dolomitic limestone fragments floating in quartz-arenaceous dolostone groundmass. Breccia is conformably overlying weakly cleaved non-calcareous black shales.

0 5
cm

Figure 2.10.: Type locality of Gorge Formation at outcrop # 137. Sub-angular carbonate mud-derived limestone fragments floating matrix-supported in a quartz-arenaceous dolomitic groundmass. Note incipient imbrication of rip-off fragments.

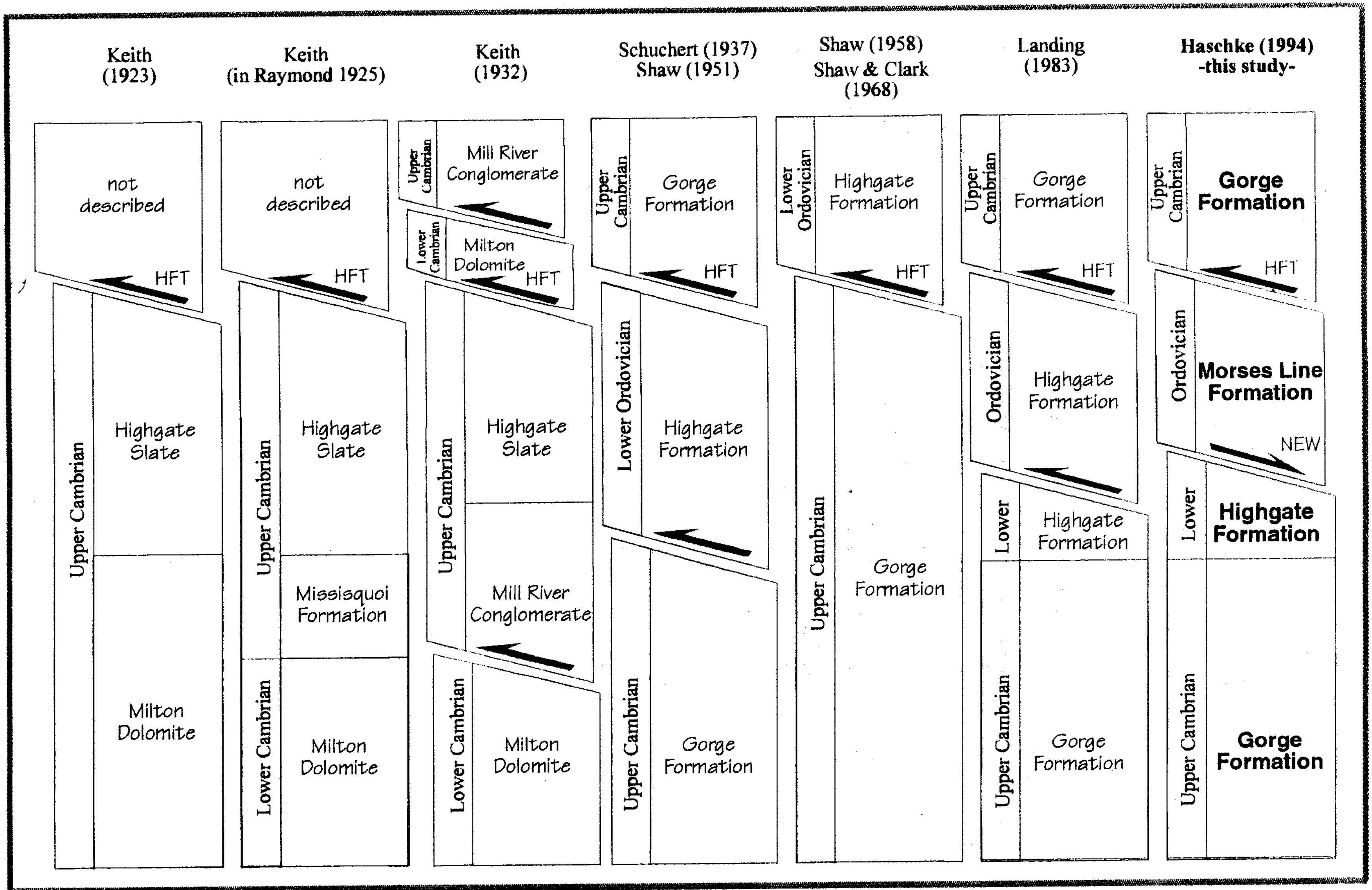
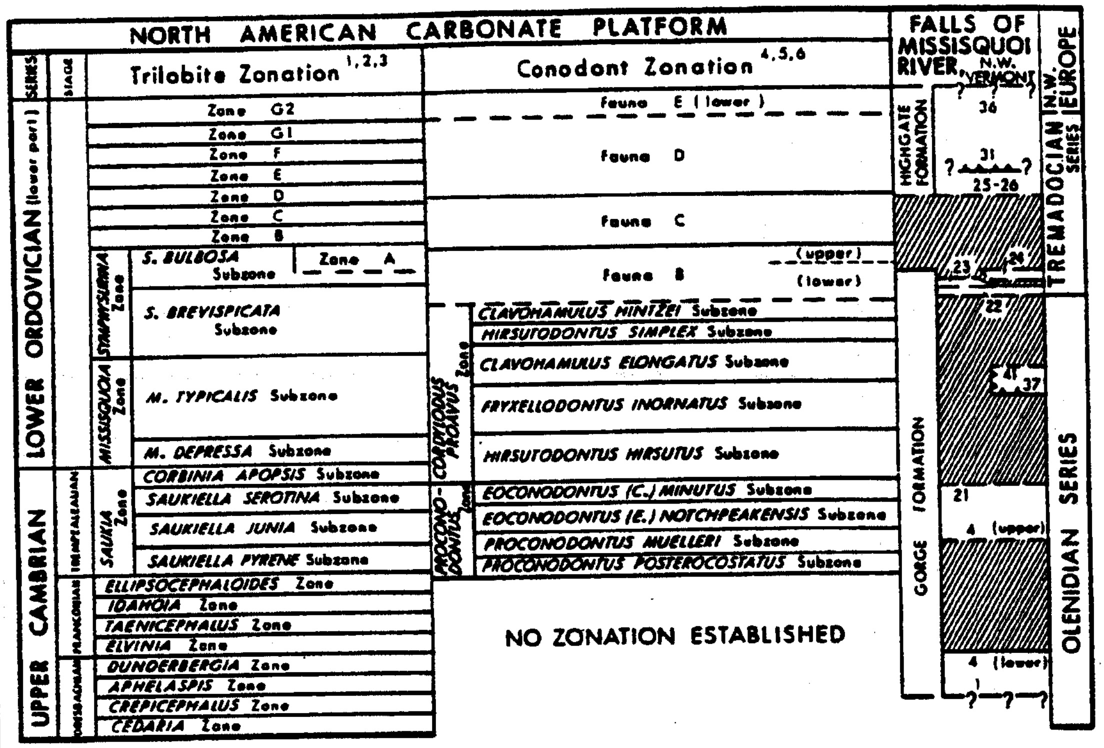


Figure 2.11.: Synthesis of stratigraphic nomenclature from previous workers of the Cambro-Ordovician section in the Missisquoi River gorge at Highgate Falls (modified after Landing 1983) HFT - Highgate Falls Thrust, NEW - New Fault.

conformably overlies non-micaceous and non-calcareous shales and siltstones of the Dolomite-Shale sequence. Keith's [in Raymond, 1925] initial attempt of establishing a type locality for the *Missisquoi Formation* (***Figure 2.11.***) as a independent Cambro-Ordovician unit represented the beginning of a nomenclatural problem of distinguishing the Gorge Formation from the Highgate Formation. The term "Missisquoi Formation" was rejected by Schuchert [1937] who introduced this unit as the *Gorge Formation* (***Figure 2.11.***). Its initial type locality was designated to the section immediately below and above the Highgate Falls thrust fault contact in the gorge of the Missisquoi River at Highgate Falls [Schuchert 1937]. Shaw [1958] revised this type locality and assigned "all rocks below the Highgate Falls thrust" to the Gorge Formation (***Figure 2.11.***). This redefinition is contrary to Keith [1932], who recognized that the upper part of the section below the thrust lithologically represents the *Highgate Formation*. Landing's [1983] recent synopsis of previous stratigraphic outlines (***Figure 2.11.***) reveals the contradiction between lithologic and paleontologic data at the type section of the Gorge Formation. Although the Gorge Formation-Highgate Formation contact is currently believed to be approximately equivalent with the Cambrian-Ordovician boundary, there is no paleontologic record of this time interval in the type section at Highgate Falls [Landing 1983]. This faunal gap caused several lithology-independent interpretations [Keith 1923, Schuchert 1937, Shaw 1958] of the contact between those units. Most recently, the "occurrence of the uppermost Cambrian *Missisquoia typicalis* fauna in bedded limestones" lead Landing ([1983], unit 22; ***Tab 2.4.***, ***Table A-1***) to suggest that the earliest Ordovician occurs within the uppermost section of the Gorge Formation at Highgate Falls.



(modified after Landing 1983)

Table 2.4.: Biostratigraphic correlation of the Gorge and Highgate Formation at Highgate Falls with the North American carbonate platform and the Acado-Baltic Faunal Province.

DATA SOURCES:

Landing & Skevington 1982
Wilson 1958
Stitt 1971, 1977
Ross 1951
Hintze 1953
Landing 1981, 1983
Taylor, Landing & Gillett 1981
Miller 1980
Ethington & Clark 1971

My detailed lithostructural investigation of this locality (outcrop # 399) revealed that the significant change from quartz-arenaceous dolostones of the Gorge Formation to lime-mudstone dominated beds of the Highgate Formation coincides with Landing's [1983] boundary of "unit 20-21 in section A&B", that is the base of the Tremadocian, and not, as proposed by Landing [1983], between "unit 23-25 in section E&F". The locality where I identified the Gorge-Highgate Formation contact (outcrop # 399, ***Figure 2.14.***), that is the obvious lithologic change from quartz-sand containing dolostones to lime-mudstone dominated units, appears to be below anything measured or shown by Landing [1983] in his "section E".

This study prefers to distinguish the two formations by the general quartz-arenaceous dolostone dominance and the absence of limestone beds in the Gorge Formation, whereas the Highgate Formation is dominated by quartz-sand free lime-mudstones.

Typical exposures of the Gorge Formation (***Figures 2.10.***) show a breccia that contains blue to dark-grey subangular, dolomitized, muddy limestone laths (2-5 cm thickness, 3-50 cm length), which are embedded in a brown to dark-grey quartz-arenaceous dolostone matrix (outcrop # 12, 13, 16, 22, 23, **78**, 84, 100, 105, **109**, 137, 151, 153, 399).

Some outcrops lack the abundance of angular clasts but can be separated from other dolostone facies by their significant brown to dark-grey quartz-arenaceous dolomite groundmass. Most of the angular clasts are interpreted to represent reworked rip-up fragments of carbonate mud-derived limestone beds, similar to beds seen in the stratigraphically younger Highgate Formation. Irregular shaped, scattered black to dark-grey chert nodules (5-50 cm Ø, ***Figure 2.12.***) as well as white to grey quartz knots (associated with quartz-

Figure 2.12.: Gorge Formation at outcrop # 151. Blue-grey angular chert nodules within massive quartz-arenaceous dolostone breccia of the Gorge Formation.

0 10
cm

Figure 2.13.: Gorge Formation at outcrop # 151. White irregular shaped quartz vugs and knots within massive quartz-arenaceous dolomite breccia of the Gorge Formation.

0 10
cm

veins, 0.1-3 cm thickness, ***Figure 2.13.***) and cavity fillings (2-100 cm Ø) commonly occur throughout the whole breccia unit (outcrop # 16, 24, **78**, 98, 105, **109**, 151). The occurrence of interstratified, non-calcareous and undeformed shales underlying the dolostone breccia of the Gorge Formation is explicitly restricted to the northern Dolomite-Shale sequence (outcrop # 14, 23, 78, 100) and these can be traced almost all the way to the U.S.-Canadian border (***Plate*** *1*). The dolomitic shales and siltstones and the shales interstratified with the Gorge Formation do not, however, extend quite as far as the international border with Québec, so that only the gently east-dipping quartz-arenaceous dolostone lithology continues across it.

The lithology of the Gorge Formation in northwestern Vermont strongly resembles the lithologic description of the *Milton Dolomite* and the *Clarendon Springs Formation* in west-central Vermont [Schuchert 1937; Dennis & Stone 1964; Mehrtens & Dorsey 1987]. Shaw [1953] proposed an identical fossil species content (*Hungaia magnifica* fauna) in both lithologies. Keith [1932], "confused by the abundance of so many similar dolomites", applied the term *Milton Dolomite* to the entire dolostone succession of the Rosenberg Sequence north of the Canadian border. The type locality of the Milton Dolomite, however, is located about 3.5 miles southwest of Milton village [Schuchert 1937].

2.4.4. HIGHGATE FORMATION — LOWER ORDOVICIAN

The Ordovician strata in the St. Albans area in northwestern Vermont are composed of two major units; a folded and cleaved early Ordovician limestone sequence (*Highgate Formation,* [Keith 1923]) and an allochthonous, complexly deformed medial Ordovician slate sequence (*Morses Line Slate*, [Shaw 1958]).

Substantial thrust faulting and extensive glacial cover complicate the interpretation of the stratigraphic and structural relations between both units. In previous publications the names *"Georgia Slate"* [Keith 1923, 1932], *"Corliss Breccia"* [Schuchert 1937], *"Grandgé Slate"* and *"Corliss Breccia"* [Schuchert 1937] and *"Upper Gorge Formation"* [Shaw 1951, 1953] were assigned for the Highgate Formation. Schuchert [1937] initially suggested *"the upper sequence in the gorge of the Missisquoi River at Highgate Falls"* (outcrop # 113), as the type locality and proposed the *Ceratopyge limestone* (Tremadocian Series, ***Tab 2.4.***) as its biostratigraphic equivalent in northwestern Europe (Scandinavia).

In this study the name Highgate Formation defines blue to dark-grey, mostly thin-bedded (2.5-15 cm) and cleaved argillaceous shaly/slaty limestones (***Figures 2.14.***; outcrop # 93, 101, **103**, **104**, 111, **113**, 148, 150, 152, 399, 432) and breccia facies (***Figure 2.15.***, outcrop # 129, 130, 135, 139, 435), including tan to dark-grey, partly dolomitized, limestone horizons where it is in contact with the underlying Gorge Formation (***Figure 2.14.***; outcrop # 399). Breccia beds within this formation (0.5-1.5 m thickness) commonly consist of chaotically oriented to weakly imbricated blue to dark-grey angular fragments of planar laminated, carbonate mud-derived limestone, embedded in an unsorted, partly arenitic to dolomitic limestone matrix (***Figures 2.15. & 2.16.***). The clasts in these breccias resemble the clasts in the dark-grey dolomitized breccias of the Gorge Formation. As defined here, the Highgate Formation is dominated by muddy limestones and typically lacks a significant quartz-sand content, which separates it from the lithology of the Gorge Formation.

At least one fault occurrence together with a lenticular fault breccia (1-4 cm thickness, 5-15 cm length) are present at the type locality of the Highgate Formation in the Missisquoi River gorge at Highgate Falls (outcrop # 140). The newly observed fault juxtaposes brown-grey calcareous moderately cleaved

0 1

m

Figure 2.14.: Stratigraphic contact between quartz-arenaceous dolostone breccia of the Gorge Formation and shaly limestone beds of the Highgate Formation at outcrop # 399 at Highgate Falls.

Figure 2.15.: Highgate Formation at the Missisquoi River gorge at Highgate Falls at outcrop # 139. Mud-derived limestone clasts floating in argillaceous limestone groundmass.

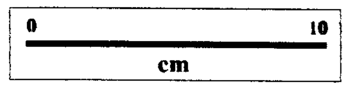


Figure 2.16.: Highgate Formation in the Missisquoi River gorge at Highgate Falls at outcrop # 137. Cross sectional view through a conglomeratic mud-limestone channel within dolomitic limestone beds.

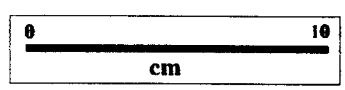


Figure 2.17.: Morses Line Formation at outcrop # 140, Missisquoi River gorge at Highgate Falls. Intensely cleaved non-calcareous slates with cross-faulted tan dolomitic boudins. Note closely spaced set of white calcite veins. Identified as Highgate Formation by all previous workers [Shaw 1958, Landing 1983, Mehrtens & Dorsey 1987].

0 50
cm

Figure 2.18.: "Fault breccia" of Landing [1983, unit 30]. Note depositional contact of cleaved limestone breccia with underlying massive micrite bed of the Highgate Formation.

0 50
cm

slaty limestones with black non-calcareous intensely cleaved slates, locally intercalated with tan dolomitic micrite boudins (***Figure 2.17.***). The intensely cleaved black slate strata in the Missisquoi River gorge strongly resembles the lithostructural character of the Morses Line Formation, and this study refers to the upper slate portion, exposed between the shaly limestones of the Highgate Formation and the Highgate Falls Thrust, as part of the allochthonous Morses Line Formation [Landing 1983, *"unit 31"*].

An additional internal tectonic contact within the Highgate Formation was suggested by Landing [1983] in his "section E", who introduced a cleaved limestone breccia (*unit 30*, ***Figures 2.18.***, outcrop # 139) at Highgate Falls as the fault breccia of an *"unnamed thrust"* between his *"unit 30"* and *"unit 31"*. My own investigation of these units at Highgate Falls, however, revealed one of several sedimentary breccia deposits (unit 30) of the Highgate Formation that displays a depositional contact between two shaly limestone units, but clearly neither a fault breccia nor a fault contact. However, this study identified a fault within Landing's [1983] unit 31. The fault locality at the north bank of the Missisquoi River gorge can be located:

(a) by the obvious color change from medium grey to black,

(b) by the lithologic change from shaly limestones (below) to non-calcareous slates (above).

I estimate that Landing's [1983] sample 9 m up into his "unit 31" is from below this contact.

Other localities of the Highgate Formation, north of Highgate Falls, expose argillaceous shaly/slaty limestones (outcrop # 2). This apparent difference in strain between the non-calcareous uncleaved siltstones and shales that are interstratified with the Gorge Formation, and the slaty limestones of the Highgate

Formation suggests a tectonic contact between the Gorge and the Highgate Formation. Moreover, this would confirm Pingree's [1982] hypothesis about the thrust faulted nature of this contact.

The interpretation of a faulted contact between those sequences, however, disagrees with the actual observation at the Missisquoi River gorge at Highgate Falls, where the Highgate Formation is observed in continuous outcrop to conformably overly the beds of the Gorge Formation (outcrop # 399; ***Figure 2.14.***). At this locality, the first thin-bedded, moderately southeast-dipping (~30°) cleaved limestone beds of the Highgate Formation are in depositional, stratigraphically intact contact with the underlying rippled surface of the uppermost quartz-arenaceous dolostone layer of the Gorge Formation.

2.4.5. SUMMARY AND DISCUSSION — ROSENBERG SEQUENCE

A general understanding of the complex stratigraphic succession of the homoclinal Rosenberg sequence has been hampered by the variety of local terminology. A currently accepted synopsis of previous stratigraphic models ([Mehrtens & Dorsey 1987]; ***Figure 2.4.***) neglected the occurrence of substantial faulting within this thrust slice and therefore cannot account for the stratigraphic and structural relations. Clarification of the stratigraphic relationships used in Québec and their correlation to those in Vermont is necessary to fully understand the equivalence of lithologies on either side of the international border.

Earlier established lithologic criteria for subdividing the quartz-arenaceous dolostone assemblage of the Rosenberg sequence could not be applied in this study. Neither the relative quartz-arenite grain abundance within the dolostones

[Shaw 1958, Mehrtens & Dorsey 1987] nor the degree of dolomitization within different slates provided sufficient information to distinguish the units.

The dolostone lithology of the Rosenberg slice ranges, within outcrop-scale, from a non-arenitic, recrystallized coarse-grained dolomite to a massive quartz-arenite. This variation of lithology on such a small scale does not agree with the previously assigned lithologies of the dolostone units within this sequence. Thus, a wide range of dolomitic-arenitic lithologies have been designated to several different stratigraphic levels; there is in my view a broad overlapping of lithologic characteristics.

Similar efforts of subdividing the slate and shale units within this sequence caused the lack of recognition of a bisected Rosenberg sequence. The dolostone dominance of the northern Dolomite-Shale sequence as well as the shale-dominated lithology of the southern Dolomite-Shale sequence has been explained by local southward-deepening of the shelf platform in Cambrian times; it is, however, debatable if facies change alone can account for this abrupt change in lithology between shallow-marine carbonates and the cleaved deeper-water slates immediately across the Missisquoi River.

None of the shale units within the Rosenberg sequence could be traced to or across the international border (***Plate 1***). The Rosenberg sequence crosses the U.S.-Canadian border as a chert-bearing quartz-arenaceous dolostone (~1270 m thickness), carrying the shaly limestone-dominated Highgate Formation on top. This observation largely agrees with previous Canadian literature [Cady 1960, Charbonneau 1980, Globensky 1981], that lacks any report of shale or slate strata within the northern continuation of the Rosenberg thrust slice. The Canadian work, however, lacks any description about a shaly

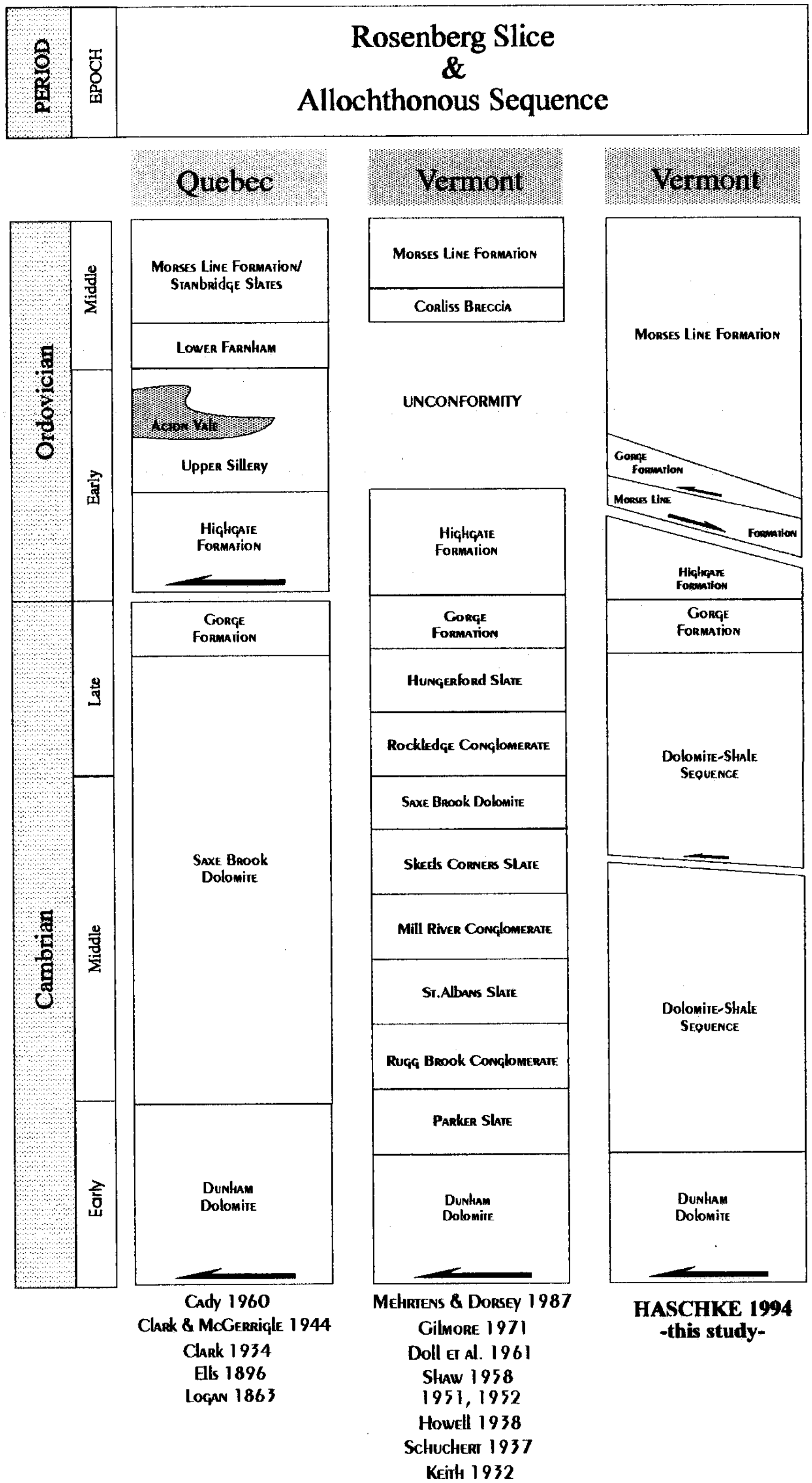


Figure 2.19.: Tectonostratigraphy of the Rosenberg slice and the allochthonous sequence in southern Quebec and northwestern Vermont.

limestone equivalent of the Highgate Formation that crosses the international border forming the uppermost exposed unit of the Rosenberg sequence.

Cady's [1960] initial stratigraphic correlation of northern Vermont and adjacent parts of southern Québec correlated the *graywacke-shale assemblage* of the *Québec Group*, the *Sillery* and the *Farnham* unit, with the rocks of the Rosenberg sequence (***Figure 2.19.***). The correlation of Cady [1960] involves an abrupt facies change from shallow water carbonates and quartzites of the shelf edge to deeper water slates of the continental slope and rise, marked by a anomalously thin zone of lateral transition. This model is in striking contrast to more recent Canadian workers [Charbonneau 1980, Globensky 1981], who suggest an isolated Cambro-Ordovician arenaceous dolostone thrust slice that pinches out northeastward, after crossing the international border [Globensky 1981]. Both investigations from Québec designate the entire dolostone assemblage of the Rosenberg sequence to the *Milton Dolomite*, representing the lowest stratigraphic unit within the Philipsburg slice and therefore the base of the Rock River Formation. However, neither Charbonneau [1980] nor Globensky [1981] recognized the lithostructural diversity of dolostones within the Rosenberg sequence and, moreover, lack reasonable evidence for designating them to the ancient term *"Milton Dolomite"*. The name "Milton Dolomite", however, has been already rejected by Dennis and Stone [1964] and replaced by the *Clarendon Springs Dolomite*.

2.5. ALLOCHTHONOUS SEQUENCE MIDDLE ORDOVICIAN

The thrust-imbricated proposed allochthonous slate belt in the St.Albans and Highgate area (***Figure 2.20.***) overlies the Highgate Formation, and therefore the Rosenberg sequence to the west, and follows a north-northeast to south-southwest trend throughout the study area (***Plate 1***). Most of the allochthonous lithology is represented by the *Morses Line Formation*, which is in Québec included within the *Stanbridge Group* described by Charbonneau [1980] and Globensky [1981].

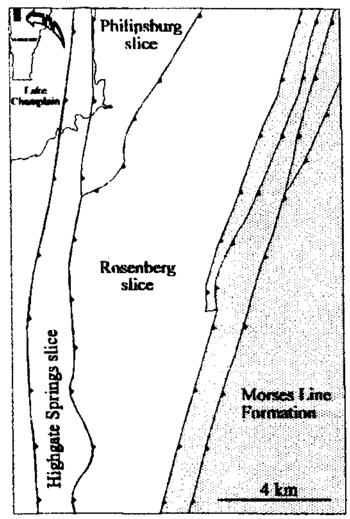


Figure 2.20.: Morses Line Formation, northwestern Vermont.

However, the allochthonous nappe character of this formation in Vermont has not been accepted by all workers [Doll et al. 1961, Mehrtens & Dorsey 1987]. It is one purpose of this study to demonstrate an allochthonous character of the Morses Line Formation, that resembles the lithofacies of the well-established allochthonous Stanbridge Nappe in Québec, and the Taconic Allochthon in southwestern Vermont.

2.5.1. MORSES LINE FORMATION MIDDLE ORDOVICIAN

The name for this formation is taken from a locality southeast of the village of Morses Line along the international border with Québec, where non-calcareous strongly cleaved black slates are exposed [Shaw 1958]. Earlier workers in Vermont referred to these strata as *Georgia Slate* [Keith 1923], *Grandgé Slate* and *Highgate Formation* [Schuchert 1937] and as *Morses Line-Stanbridge Slate* [Cady 1960] in Québec. The *Morses Line Slate* was dated by Shaw [1958] as early Ordovician in age, using the identification of fossils within the clasts of intraformational limestone breccias [Mehrtens & Dorsey 1987]. Cady [1960], in contrast, suggested a medial Ordovician age for the *Morses Line-Stanbridge Slates*. Most recent work [Mehrtens & Dorsey 1987], however, refers to Shaw [1958] again where the Morses Line Slate is early Ordovician (Canadian) in age.

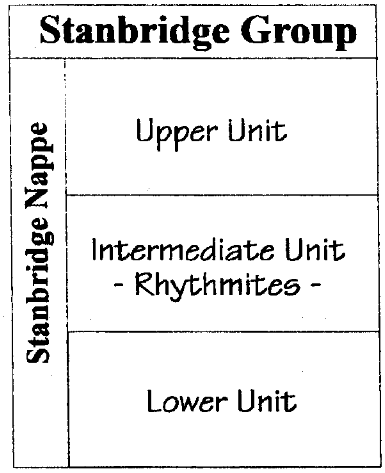


Figure 2.21.: Stratigraphic units of the early-medial Ordovician Stanbridge Group in Quebec. (after Charbonneau 1980)

Shaw [1958] first documented the lithologic similarity of the Morses Line slates with the allochthonous Middle Ordovician *Stanbridge Slate* of Canadian geologists. Charbonneau [1980] initially defined the *Stanbridge Group* (***Figure 2.21.***) in Québec as a northeast-southwest trending band of 9-12 km length in the Stanbridge area, northeast of the Missisquoi Bay, and correlated this sequence (***Tab.2.2.***) in Canada lithologically with the division C_2 and D of Logan [1863], with the *Basswood Creek Slates*, the *Mystic Conglomerate* and the *Stanbridge Slates* of Clark and McGerrigle [1944] and with the *Stanbridge Complex* of Clark and Eakins [1968]. According to Charbonneau [1980] and Globensky [1981], the Stanbridge Group can be lithologically subdivided into 3 lithostratigraphic units (***Figure 2.21.***):

• LOWER UNIT	Thin-bedded (0.1-1cm) dolomitic shales, occasionally interstratified with limestone horizons (10 cm thickness), and discontinuously intercalated limestone conglomerates and breccias;
• INTERMEDIATE UNIT	Alternating siltstone-shale-mudstone rhythmites, partially dolomitized;
• UPPER UNIT	Pyrite stained shales and slates with interstratified siltstone laminations; partly calcareous slates and discontinuous intercalated limestone conglomerates.

In Vermont, the *Stanbridge Group*, as defined by Charbonneau [1980] and Globensky [1981], stratigraphically and lithologically incorporates the *Georgia-* and the *Highgate Slates* of McGerrigle [1931] and Clark and McGerrigle [1932], and comprises the *Highgate Formation* and the *Morses Line Formation* as defined by Keith [1923] and Shaw [1951, 1958].

This study refers to the **Morses Line Formation** as allochthonous, thrust-imbricated, calcareous (outcrop # 141, 204, 324, 328, 329, 433) to non-calcareous (outcrop # 140, 142, 200, 208, 216, 218, 323, 331, 332, 433) slate and slaty limestone strata (***Figure 2.22. & 2.23.***) with episodically intercalated, lenticular argillaceous limestone breccia bodies (outcrop # 202, 203, 212, 327, 330, 432). The term *Corliss Conglomerate* was introduced by Schuchert [1937] to represent several isolated outcropping patches of limestone conglomerate and breccia that occur scattered throughout the entire area of the Morses Line Formation. Limestone conglomerates within the Morses Line Formation (outcrop # 205, 207, 324, 330) typically contain blue-grey angular to subangular argillaceous limestone clasts

Figure 2.22.: Morses Line Formation at outcrop # 141, 2.5 km south of Morses Line at the international border. Note intensely cleaved and tightly folded slaty limestones and slates.

0 5 cm

Figure 2.23.: Morses Line Formation at outcrop # 141, 2.5 km south of Morses Line at the international border. Folded and cleaved slaty limestones, cut by quartz veins.

0 5 cm

(3-30 cm length, 3 cm thickness), floating in a partly quartz-arenaceous and dolomitic, brown-grey to blue-grey carbonate mud-derived limestone matrix. Angular limestone boulders of up to 80 cm diameter were observed at several localities (outcrop # 205, 207, 243). Several breccias are cross-cut by discontinuous quartz-veins (0.5-5 cm thickness). In general, the limestone breccias within the Morses Line Formation strongly resemble the limestone breccia beds of the Highgate Formation observed at Highgate Falls. The slate lithology above the Highgate Formation at this locality, however, is inferred to belong to the Morses Line Formation.

2.5.2. SUMMARY AND DISCUSSION — ALLOCHTHONOUS SEQUENCE

The slaty limestones of the Highgate Formation in the study area represent the lithologic analogue of the *Lower Unit* of the Stanbridge Group in Québec [Charbonneau 1980]. In particular, it can be demonstrated that the banded argillaceous limestone strata of the lowest portion of the Highgate Formation, exposed at the Missisquoi River gorge at Highgate Falls in northwestern Vermont, grades upward into more fissile non-calcareous slates with intraformational limestone conglomerates and breccias.

This lithologic transition within the Highgate Formation causes it to partly overlap with the lithology of the Morses Line Formation and complicates the definition of stratigraphic and structural relations between both formations. In spite of the separate lithologic definition of the slaty limestone of the Highgate Formation and the mostly intensively cleaved slates of the Morses Line Formation, their lithological contact cannot be exactly located. This problem is aggravated by the general lack of outcrop at critical locations. According to Charbonneau [1980], there is no definite lateral lithostratigraphic equivalent to the *Intermediate* and the *Upper Unit* of the Stanbridge Group in Vermont. This study, in contrast, proposes the upper portion of the Highgate Formation (outcrop # 104), exposed 500 m northwest of Highgate Center, as the lithologic analogue of the Intermediate Unit of the Stanbridge Group, as defined by Charbonneau [1980]. Exposed is a succession of blue-grey thin-bedded, folded and cleaved calcareous argillites with interstratified siltstone-mudstone laminations (1-3 mm). This lithology (outcrop # 104) strongly resembles Charbonneau's [1980] description of the *rhythmites*, 3 km north-northwest of Saint-Ignace-de-Stanbridge in Québec, where alternating siltstone-argillite-mudstone cycles are exposed, that characterize the Intermediate Unit of the Stanbridge Group. In addition, I believe that the Morses Line Formation in Vermont (outcrop # 142, 208, 209, 212, 214, 215, 216, 218, 332, 333) represents the lithostratigraphic analogue to the Upper Unit of the Stanbridge Group in Québec.

According to Charbonneau [1980], a typical exposure of the Upper Unit is exposed just north of the international border at Pigeon Hill, 4 km southwest of St.Armand Center in Québec, and shows intensively cleaved calcareous slates with intercalated discontinuous limestone conglomerates and breccias. Therefore, no significant lithologic difference between the Upper Unit of the Stanbridge Group in Québec and the Morses Line Slate in Vermont can be demonstrated. Since there is no reasonable argument why the lithology of the Stanbridge Group should terminate along the international border, this study considers the Highgate Formation together with the Morses Line Formation of northwestern Vermont as the lithostratigraphic equivalent to the well-established Stanbridge Group in Québec .

2.6. SYNOPSIS STRATIGRAPHY

In summary, this study concludes, and agrees with Cady [1960], that the problem of correlation of rocks north and south of the international border is chiefly one of terminology. Previous attempts of clarification of the lithostratigraphic relations in northwestern Vermont added more more confusion to the questionable and un-reproducible work on the stratigraphic nomenclature.

However, most stratigraphic relationships are rather easily explained by taking the structural relationships into account. Many of the units resemble each other, or their lithologic character is partly overprinted by recrystallization (e.g. Dunham Dolomite) and/or tectonic processes, so that it is often difficult to assign a particular outcrop properly to a specific unit. Only the major well-established stratigraphic units are utilized in this study in order not to add more to the already existing confusion about the stratigraphic relations for the rocks in the tectonic foreland of the medial Ordovician Allochthon in northwestern Vermont.

I choose to provide a precise lithologic description for each outcrop, to leave room for changes in lithologic correlation in the light of later evidence. My detailed field work was predominantly focused on resolving critical stratigraphic contacts and on the fault relationships in the field area and the involved rock units. Many of the difficulties in correlating the stratigraphic units are the result of the disruption due to tectonic processes (e.g., fault slivers), which probably contributed to the confusion considerably more than previously thought. Although my detailed field work substantially improves the understanding of the lithostratigraphy in northwestern Vermont, some critical stratigraphic relations still remain unclear.

However, the major differences to previous detailed maps in the field area are:

- A characteristic fine-crystalline, non-arenaceous early Cambrian dolomite section of the Dunham Dolomite can be documented in the northern as well as in the southern portion at the base of the Rosenberg slice.
- The medial to late Cambrian quartz-arenaceous dolostone and shale-siltstone succession of the Rosenberg Sequence cannot be subdivided into the previously described units (Parker Slate, Rugg Brook Dolomite, Mill River Conglomerate, Skeels Corners Slate, Rockledge Conglomerate, Saxe Brook Dolomite, Hungerford Slate, ***Figure 2.4.***) using outcrop-defined lithologic criteria. Lithologic descriptions of both the quartz-arenaceous dolostones, and the shales and slates, mostly overlap with each other.
- The entire Rosenberg slice is bisected and cross-cut by a northeast-southwest trending suite of faults; one of those, near Highgate Falls, appears to tectonically juxtapose a quartz-arenaceous dolostone-dominated northern succession of the Dolomite-Shale sequence with a shale-dominated southern sequence. This is in strong contrast to previous interpretations, which show a largely unfaulted section of siliciclastic dolomite and slate strata.
- The Morses Line Formation, together with slaty portions of the Highgate Formation, represent, in the St. Albans area, the lateral lithostratigraphic equivalents to the well-established Stanbridge Group in Québec. It can be documented that the lower and the upper portion of the Highgate Formation correspond lithologically to the lower and intermediate unit of the Stanbridge Group, respectively. Moreover, the Morses Line Formation represents the lithostratigraphic analogue to the upper unit of the Stanbridge Group.
- The lithostructural character of the slate strata in the Missisquoi River gorge at Highgate Falls, that are exposed between the shaly limestones of the

Highgate Formation and the Highgate Falls Thrust, strongly resemble the Morses Line Formation east of Highgate Falls.

- A carbonate thrust sliver relation can be demonstrated at the dam at Highgate Falls where the resistant quartz-arenaceous carbonate sliver of the Gorge Formation causes the vertical step in the topography and forms a locally repeated sequence of the Gorge Formation. At this locality, the Gorge Formation is interpreted to be an intercalated thrust sliver within the allochthonous Morses Line Formation.

STRUCTURE

"... *Not until 1861 did anyone see a fault - it turned out to be a big overthrust ...*"

Charles Schuchert 1937

3.1. INTRODUCTION

The main focus of this study comprised the investigation of the previously neglected structural features in the St. Albans and Highgate area in northwestern Vermont. In particular, the structural geometry of the Rosenberg sequence within the imbricated thrust slice assemblage of the *Champlain Thrust System* in the study area (*Highgate Springs slice, Philipsburg slice, Rosenberg slice*), and the nature of the contact with the overlying allochthonous Morses Line Formation needed to be clarified.

Most of the uncertainty, whether the contact of the Rosenberg sequence and the Morses Line Formation is a transitional or a faulted one, arose when Shaw [1958] proposed thrust fault-bounded contacts of the Highgate Formation, both with the Gorge, and the Morses Line Formation, respectively. Subsequent work of Canadian geologists [St.-Julien & Hubert 1975, Charbonneau 1980, Globensky 1981, St.-Julien et al. 1983] provided additional support for this interpretation by suggesting a north-south trending thrust contact between the Rosenberg slice and the *Stanbridge Nappe*, the Canadian *Morses Line* equivalent, in Québec. This interpretation, however, disagrees with earlier workers [Schuchert 1937, Cady 1945], who assumed an intact, transitional stratigraphic contact between those sequences. In spite of the well-established, faulted, structural relation between the Rosenberg sequence and the allochthonous Stanbridge Nappe in Québec, most recent work in the immediately adjacent Highgate and St. Albans area rejects a tectonic contact and restates an unfaulted nature of the contact between the Rosenberg sequence, in particular the Highgate Formation, and the Morses Line Formation [Mehrtens & Dorsey 1987].

In this context, previous detailed mapping along the western margin of the Taconic Allochthon [Rowley & Kidd 1981, Bradley & Kidd 1991], about 200 km south of the study area, resulted in a modified interpretation for the structural history of the Taconic Orogeny where the geology is relatively well understood. However, their developed structural interpretation has not been applied to or demonstrated for the Champlain Lowlands in the St. Albans and Highgate area. Using the structural model for the nature of the contact between the foreland thrust belt and the Taconic Allochthon, together with the results of Canadian workers, this study attempts to present a new structural interpretation for the study area that is consistent with equivalent interpretations north and south.

Additionally, I attempted to resolve the high-angle fault relations that cross-cut the stratigraphy of the Rosenberg sequence. High angle faulting and its effect on the stratigraphic succession has been neglected by most previous workers, although some features of the relief can only be reasonably explained by faults.

Faulting is the most prominent deformational feature of the area and can be demonstrated in the field. The amount of displacement along high angle and low angle faults ranges between several meters and hundreds of kilometers, respectively. Traditionally defined thrust slice boundaries typically coincide with topographic breaks.

Structurally, the northern Champlain lowlands are best described as a fragment of an overthrust system (*Champlain Thrust System*) that comprises several large, gently east-dipping thrust-imbricated slices of the Taconic foreland. The Champlain Thrust System can be traced from the Catskill Plateau in east-central New York [Bosworth & Vollmer 1984, Stanley 1987, Plesch 1994], throughout west-central Vermont [Stone & Dennis 1964], northward to and across the international border into Québec. In northwestern Vermont, this thrust fault-controlled zone consists of 3 major thrust slices; the *Highgate Springs-*, the *Philipsburg-* and the *Rosenberg slice*.

Every thrust slice rests on its corresponding low angle thrust fault (e.g., 0-10°, Rosenberg Thrust [Rowley 1982]); in the study area known as the *Highgate Springs-*, the *Philipsburg-* and the *Rosenberg-* or *Champlain Thrust* [Kay 1958, Shaw 1958]. These largely north-south trending thrust horizons dominate and define slice boundaries and correspond to topographic breaks in the study area. Such faults are easily recognized where they place quartz-arenaceous dolostone lithology over argillaceous limestones and slates. The great difference in resistance to weathering produces ledges up to 30 meters high along the Highgate Springs-Philipsburg contact (outcrop # 320, 321, 340, 341), the Highgate Springs-Rosenberg contact (outcrop # 245, 310, 423), and the Philipsburg-Rosenberg contact (outcrop # 161, 179), respectively. Most prominently developed and described is the thrust contact where early Cambrian dolomite of the Rosenberg sequence is placed over medial Ordovician limestone of the Highgate Springs sequence, termed the *Rosenberg-* or *Champlain Thrust fault* [Shaw 1958, Mehrtens 1985, Mehrtens & Dorsey 1987]. This thrust extends throughout the area roughly paralleling the east shore of Lake Champlain.

3.2. HIGHGATE SPRINGS SEQUENCE

The carbonate-shale succession of the Highgate Springs slice forms a peculiarly narrow strip, dominated by the *Beldens Limestone*, that extends north-south for about 134 km but is only up to 1.3 km wide [Kay 1958]. A complete structural resolution of this sequence, in particular south of Highgate Springs is hampered by the general lack of outcrop along the shoreline of Lake Champlain.

3.2.1. FAULTING

The *Highgate Springs Thrust* carries the shelf carbonates of the Highgate Springs sequence and represents the westernmost thrust fault observed within the study area. Its sharp, shallow east-dipping, undulated fault contact can be observed at the *St. Griswold & Co.Inc.* Quarry (***Figure 3.1.***), just south of the village of Swanton, where dolomitic (marbleized) limestones (mapped by Kay [1958] as *Beldens Limestone*) are thrust over heavily cleaved black slates at a low angle (<10°, outcrop # 403). At this locality, thrust faulting caused the interstratified dolomite layers to be drawn out into lens-sectioned boudins along the thrust trace. A confusing attempt at correlating the Highgate Springs Thrust with the *Highgate Falls Thrust* [Mehrtens & Dorsey 1987] lacks essential structural evidence and could not be confirmed in this study.

The Highgate Springs Thrust can be extrapolated northward through the entire study area before it is lost underneath the Missisquoi Bay of Lake Champlain, about 4 km south of the international border. It reappears in Québec, north of Lake Champlain and cuts the east limb of the southern end of the Chambly-Fortierville synclinal basin [Clark & McGerrigle 1944, Avramtchev 1989].

Figure 3.1.: Highgate Springs Thrust fault contact at outcrop # 403.

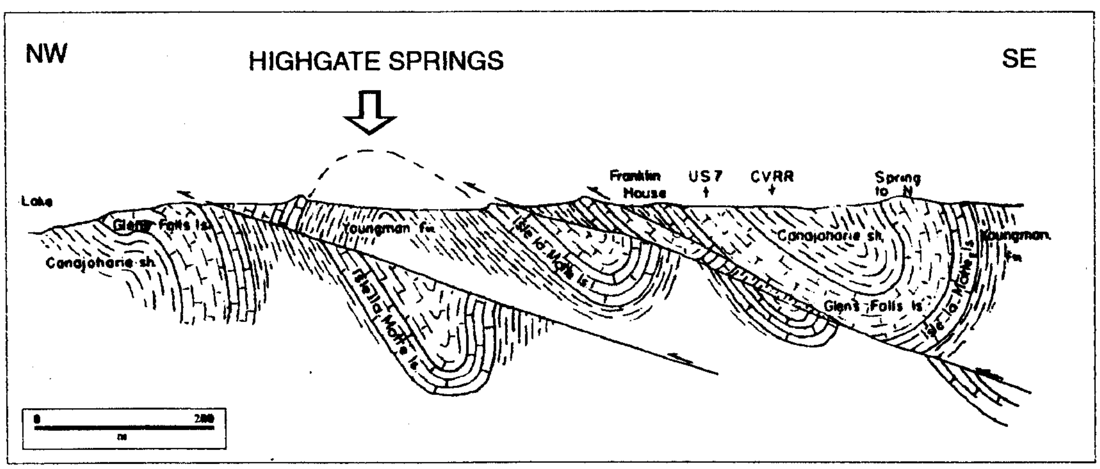


Figure 3.2.: Schematic structural NW-SE interpretation of the Highgate Springs sequence at the village of Highgate Springs (modified after Kay 1958).

3.2.2. FOLDING AND CLEAVAGE

This study largely follows the structural interpretation of the Highgate Springs sequence by Kay [1958]. According to Kay's work [1958], the arenite-limestone-shale assemblage of the Highgate Springs slice reveals intricately imbricated, largely east-dipping fold fragments of regionally limited extent (250-300 m; ***Figure 3.2.***). The majority of outcrops within this sequence is concentrated at the center of Highgate Springs village, where a number of U-shaped ledges, separated by meadows, trace the limbs of asymmetrical folds with east-dipping axial planes. The principal outcrop area delineates a south-southwest plunging anticline, broken by low-angle faults that produce interruptions and repetitions, and by some offsetting high-angle faults that cross the dominant structural trends [Kay 1958].

Fold deformation features can be readily observed throughout the Highgate Springs sequence. The several formations have reacted differently; for example, the *Carman* has been rather severely folded in spite of its quartzose lithologic nature; the *Youngman*, where shaly and argillaceous, has a moderately developed axial-plane spaced cleavage such that the foliation resembles a series of parallel bands and therefore can be easily confused with bedding. The *Glens Falls*, on the other hand, displays an incipient axial planar fracture cleavage, causing the generation of rhomb-shaped cleavage fragments [Kay 1958].

In spite of the tectonically disrupted nature of this carbonate sequence, the Highgate Springs slice forms a persistent narrow strip of shallow water shelf deposits between the Georgia Plains region and Highgate Springs village in northwestern Vermont.

3.3. PHILIPSBURG SEQUENCE

The dolostone-limestone assemblage of the Philipsburg slice appears as an intercalated wedge within the Champlain Thrust System between Highgate Springs village, Vermont, and Bedford, Québec, separating the Highgate Springs succession from the Rosenberg sequence. North of Rosenberg, however, the Philipsburg slice is juxtaposed with the *Sainte-Sabine Formation* to the west and the allochthonous *Stanbridge Nappe* to the east [Cady 1945, Globensky 1981, Avramtchev 1989].

3.3.1. FAULTING

The Philipsburg Thrust appears about 1.6 km southeast of Highgate Springs and carries the carbonate suite of the Philipsburg succession northward, towards and across the international border. Several sharp thrust contact exposures can be documented along the shoreline of Lake Champlain, immediately before entering Québec (***Figure 3.3.***, outcrop # 320, 321, 341). At those localities the fault contact is characterized by steeper dipping (62°), fault-dragged thin-bedded limestones that are overthrusted along a shallow east-dipping (29°) fault surface by moderately dipping (39°) calcareous dolomite beds, causing a steep scarp of about 35 m.

After crossing the Vermont-Québec state line the Philipsburg Thrust continues as Logan's Line [Logan 1863] that places the Philipsburg succession atop of flysch-type lithologies of the Sainte-Sabine Formation [Globensky 1981, Avramtchev 1989]. North of Morgan Corners (about 7 km north of the international border), however, the trace of this thrust cuts upward through the stratigraphic section of the Philipsburg succession and complicates the structural

Figure 3.3. Philipsburg Thrust fault contact at outcrop # 320, Missisquoi Bay, Lake Champlain, just south of the international border.

0 40
cm

Figure 3.4.: Gently southeast dipping limb of the Philipsburg syncline at outcrop # 347, about 1 km southeast of the village of Highgate Springs.

0 5
m

and stratigraphic relations. About 10 km north of the international border the Philipsburg Thrust/Logan's Line juxtaposes the Sainte-Sabine Formation with allochthonous units of the Stanbridge Nappe and, further north, the Granby Nappe [Globensky 1981, Avramtchev 1989].

3.3.2. FOLDING AND CLEAVAGE

The fold deformation features within the Philipsburg sequence in Vermont are restricted to a dismembered homoclinal, southeast-dipping, well-bedded and uncleaved carbonate sequence (***Figure 3.4.***). In its extension into Québec, however, bedding orientation and map units reveal a large, gently northeast-plunging asymmetric syncline [Globensky 1981] with an estimated steeply southeast-dipping axial surface. Moreover, the Solomons Corners Formation in Canada is documented as an incipiently cleaved unit within the Philipsburg carbonate succession. Most outcrops of the Philipsburg slice in Québec are restricted to the western limb of this asymmetric syncline structure [Globensky 1981].

3.4. ROSENBERG SEQUENCE

The Rosenberg sequence represents a large homoclinal Cambro-Ordovician siliciclastic and carbonate thrust slice that extends as a more-or-less coherent sequence between Burlington, Vermont, and Rosenburg, Québec. Its northern end in Canada, about 6 km north of the international border, is tectonically juxtaposed with the allochthonous Stanbridge Nappe [Globensky 1981]. In Vermont, however, this contact is lithostructurally equivalent with the contact to the Morses Line Formation, that has been described as being stratigraphically intact [Shaw 1958, Mehrtens & Dorsey 1987].

3.4.1. FAULTING

Within the Champlain Thrust system in northwestern Vermont, the *Champlain-* or, further north, the *Rosenberg Thrust* (*"St. Lawrence Thrust"* [Schuchert 1937]) carries the Rosenberg sequence; a large homoclinal Cambro-Ordovician siliciclastic carbonate slice, that is immediately underlying the Morses Line Formation/Stanbridge Nappe. This most prominent thrust fault occurrence has been described as the *"master fault in northwestern and west-central Vermont"* [Shaw 1958] since it was first recognized by Logan [1863].

The *Champlain Thrust* can be traced from the vicinity of Middlebury, Vermont, northward through the study area but appears to split into two separate thrust branches about 1.6 km southeast of Highgate Springs. At this locality, the underlying Philipsburg thrust slice is cut out by the northeast trending *Rosenberg Thrust* [Kay 1958], which crosses the international border, but extends only about 6 km into southern Québec, according to Globensky [1981], where it forms the western boundary of the large homoclinal Rosenberg thrust sliver ([Stanley 1987]; ***Figure 1.6.***).

Most of the attention the Champlain and the Rosenberg Thrust received from previous workers was due to the abrupt changes seen across this thrust fault within the early Paleozoic shelf sequence in terms of contrasting thickness, time-stratigraphic interval, and facies [Rowley 1982]. According to Rodgers [1970], the siliciclastics and carbonates of the Rosenberg sequence represent a more distal facies of the shelf edge to slope than the adjacent shelf carbonates of the Highgate Springs- and Philipsburg sequence.

The estimated amount of displacement on the Champlain Thrust and on the Rosenberg Thrust, respectively, varies from worker to worker and from area to area but ranges from 7.2-6.2 km at the latitude of Middlebury to about 32 km near Milton [Cady 1945], but is likely everywhere 80-94 km according to Rowley [1982]. Rowley's [1982] approach of determining the amount of displacement along this fault, however, utilized the simple geometric relationships of a seaward thickening wedge for the cross-sectional shape of a passive margin (Atlantic-type) shelf sequence. Using this method yields an estimate of the amount of displacement about an order of magnitude larger than that of previous workers. In spite of the large uncertainties within the variables employed in Rowley's [1982] estimate, its application yields a significantly larger amount of displacement along the Champlain Thrust than can possibly be determined from surface geology.

The apparent increasing throw towards the north on this largely north-south trending fault led earlier workers to suggest that the upper plate, carrying the Rosenberg succession, essentially pivoted about a vertical axis located some 20 km south of Middlebury [Cady 1945, Coney et al. 1972]. The continuation of this major thrust to the south [Steinhardt 1983], and the tracing of the Champlain Thrust System over a distance of about 320 km (200 miles), as far as the Helderberg unconformity near Albany, New York, where part of it is overlain unconformably by Devonian rocks [Vollmer 1981, Bosworth & Vollmer 1981, Stanley 1987, Plesch 1994], however, confirms the continuous north-northeast to south-southwest trend observed farther north, and therefore I reject the proposed pivoting motion of the upper plate of the Champlain Thrust. Thrusting along the Champlain Thrust occurred during the latest part of the Taconic orogeny of medial to late Ordovician age. Some subsequent movement, however, during the Paleozoic Acadian orogeny can not be ruled out [Stanley 1987].

Additional north-south trending thrust occurrences within the Rosenberg slice have been reported by previous authors [Schuchert 1937, Shaw 1958, Landing 1983]. A *"rather minor"* thrust fault, the ***Saxe Thrust***, has been introduced by Shaw [1958]. It appears as a major topographic break in the dolostone strata of the Rosenberg sequence (outcrop # 108, 162, 170, 183) that can be traced from the international border southwestward, following the Saxe Brook valley. Shaw's [1958] main evidence for this thrust is based on the varying proximity of shales towards the Dunham Dolomite at different locations. According to Shaw [1958], the identical shale strata reappear further south at a greater distance from the Dunham Dolomite. Moreover, Shaw [1958] suggested that the reversal of dip within the slates provides accompanying evidence for the Saxe Thrust.

This study, however, traces the Saxe Thrust contact along the Saxe Brook valley, where the thrust coincides with a prominent westward facing scarp of about 15 m height. Furthermore, the slickenside striated slate strata (outcrop # 59, 61) immediately above the inferred thrust contact (***Figure 3.5.***) suggest thrust faulting along or underneath the slates. Estimates of the total amount of displacement along the Saxe Thrust are not available, but the similarity in lithology within the dolostones east and west of the thrust suggests a minor fault slip. The trace of the Saxe Thrust is lost before it crosses the Missisquoi River valley in the sandy outwash plains.

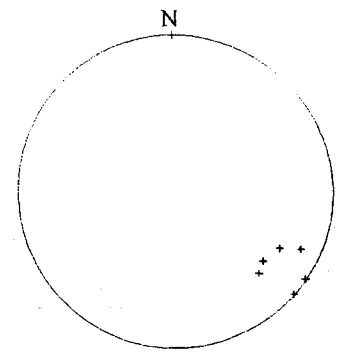
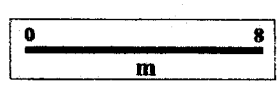


Figure 3.5.: **Lower hemisphere stereographic projection of slickenside striations at outcrop # 59.**

Another thrust within the Rosenberg sequence, the ***Highgate Falls Thrust***, has been recognized by previous authors because of its spectacular fault contact exposure at the Missisquoi River gorge at Highgate Falls (***Figure 3.6.***, outcrop # 137), about 7.85 km south of the international border. Moreover, this locality has been described in detail for its unusual exposure of a complete and continuous outcropping trilobite-containing section from the Upper Cambrian into the Lower Ordovician [Landing 1983]. Shaw [1951, 1953, 1958] first referred to and documented the Highgate Falls Thrust in detail and extrapolated the fault contact north beyond the international border and south of Highgate Falls.

The Highgate Falls Thrust carries an isolated, siliciclastic carbonate sliver of the Gorge Formation, forming the lip of the falls (as exposed, a minimum of about 15 m thickness). On the north bank of the Missisquoi River gorge, the Highgate Falls Thrust places quartz-sand containing dolostones directly over black slates that were interpreted to represent a portion of the Highgate Formation. However, due to the strong lithologic resemblance of these slates with the slate strata of the Morses Line Formation, this study prefers to interpret the sequence, that is immediately underlying the Highgate Falls thrust contact, as a portion of the Morses Line Formation (outcrop # 112). This section corresponds to Landing's [1983] "section E" (*unit 32-36*) that is shown as a strata

Figure 3.6: Highgate Falls thrust contact at outcrop # 137, Missisquoi River gorge at Highgate Falls. View towards the west wall of the Falls.

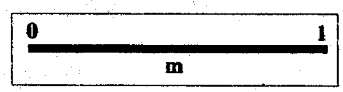


of limestones with some dolostones. I think that this section must be from the southwestern bank opposite and not from the northern bank of the Missisquoi River gorge where Landing [1983] shows his "section E" locality. Moreover, I think that these limestones and dolostones of Landing's [1983] unit 32-36 are isolated in the Highgate Falls thrust system and are not, as shown by Landing [1983] in "section E", stratigraphically above the black slates (unit 31). In any case, they do not occur on the north bank in direct continuity with the slates.

The thrust surface is only exposed in the Missisquoi River gorge at Highgate Falls for about 25m and forms an undulated, gently east-dipping (<10°) fault surface. Thrust faulting at Highgate Falls produced spectacular structural features immediately beneath the fault contact, that were revealed during the construction work on the Highgate Falls dam system in summer and fall 1993. Some views of the structural relations shown in this thesis are based on a unique opportunity to observe the nature of deformation in the vicinity of the thrust contact that are not accessible anymore. Several complex chevron-fold structures were observed beneath this fault zone (***Figure 3.7. & 3.8.***). Further downstream, the outcropping section of the Highgate Formation dips gently to moderately (~30°) towards the southeast (outcrop # 399) where it is exposed in intact stratigraphic contact above the main section of the Gorge Formation.

The thrust fault relations between the Gorge and the Morses Line Formation exposed at Highgate Falls are restricted to the Missisquoi River gorge locality. North and south of Highgate Falls the Gorge Formation contact is generally inferred to be an intact stratigraphic one with the Highgate Formation. However, the presence of this isolated, locally restricted Gorge Formation thrust sliver, carried by the Highgate Falls Thrust, additionally complicates the structural relations in the study area.

Figure 3.7.: Chevron-folded dolomitic limestone layers, immediately below the Highgate Falls thrust contact at the Missisquoi River gorge at Highgate Falls.



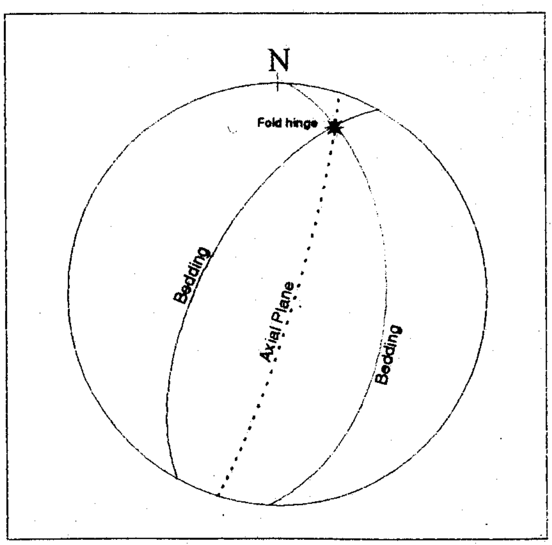


Figure 3.8.: Equal area projection of the chevron fold structure shown above; footwall of the Highgate Falls Thrust at the Missisquoi River gorge.

Shaw's [1958] solution of this structural problem (***Figure 1.5.***) involved the redefinition of the type-lithology of the well-established Gorge Formation and caused substantial confusion in the stratigraphic nomenclature of the Gorge and Highgate Formation. Furthermore, Mehrtens and Dorsey's [1987] attempt of correlating the Highgate Falls with the Highgate Springs Thrust lacks essential structural evidence. The northeast-southwest trending continuation of the Highgate Falls Thrust, however, has been inferred to continue all the way up to and across the international border as an internal thrust imbrication within the Morses Line Formation/Stanbridge Nappe that is marked by occasional carbonate slivers attached to the fault surface (outcrop # 205, 206, 207).

Within the Highgate Formation, this study identified at least one more fault occurrence at the Missisquoi River gorge at Highgate Falls. The exposed northwestern section downstream of Highgate Falls contains a fault that juxtaposes shaly limestones of the Highgate Formation with intensely cleaved black slates that strongly resemble those of the Morses Line Formation (***Figure 3.9.***). The sharp, moderately (39°) southeast-dipping, fault surface is locally characterized by a lenticular granular fault breccia of 3-5 cm thickness (***Figure 3.10.***) and displays a small slickenstriated surface (~9 cm^2; 38/122) at a locality about 3 m above the fault breccia. Steps on the calcite covered fault surface of about 1-3 mm thickness indicate a moderately dipping normal motion of the upper plate towards the southeast (38/122, ***Figure 3.11.***). This newly identified fault is located stratigraphically within Landing's [1983]

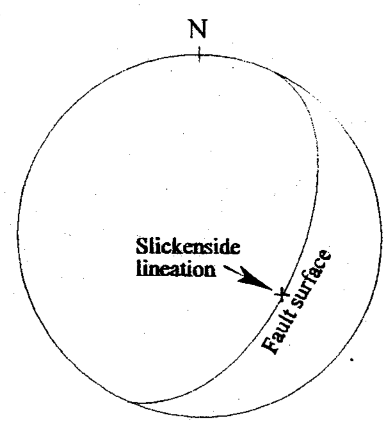


Figure 3.11.: **Equal area projection of the new fault identified at outcrop # 140 in the Missisquoi River gorge at Highgate Falls.**

Figure 3.9: New fault contact identified in the Missisquoi River gorge. Shaly limestones of the Highgate Formation are juxtaposed with intensely cleaved slates of the Morses Line Formation.

0	2
m	

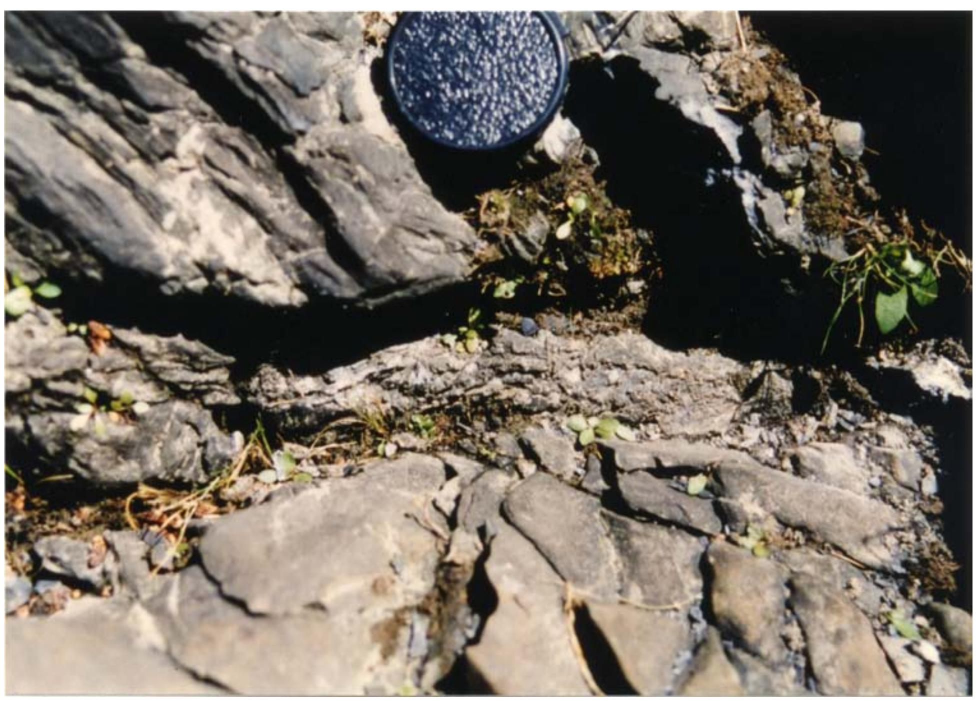


Figure 3.10.: Coarse-grained fault-breccia at the new fault contact of *Figure* above.

0	5
cm	

undivided slate unit (*"unit 31"*) of the "Highgate Formation" in "section E". Landing's [1983] proposed *"unnamed thrust"*-contact between *"unit 30"* and *"unit 31"* (***Figure 2.18.***, outcrop # 139) in this section, however, could not be confirmed as a tectonic contact by this study but was identified as a stratigraphically intact contact, revealing a sedimentary breccia instead of the "fault breccia" proposed by Landing [1983].

3.4.2. FOLDING AND CLEAVAGE

The structures in the siliciclastics and carbonates of the Rosenberg sequence, particularly the dolomites, strongly contrast, within outcrop scale, with structural features seen in intercalated shale and slate horizons due to their significantly different response to strain. Sedimentary layering within the siliciclastics and carbonates represents the dominant primary structure, whereas, within the southern portion of the Rosenberg succession in particular, a tectonically imposed weak to moderately developed slaty (axial planar) cleavage together with mild folding controls the planar structures in the shales and slates, where developed. If the cleavage parallels the thin layering within the shales it is often not possible to clearly determine the observed feature as a shaly parting or a genuine slaty cleavage.

The intensities of fold deformation within the shaly lithologies of the southern Dolomite-Shale sequence together with the slaty limestones of the Highgate Formation of the Rosenberg slice strongly contrast with the degree of fold deformation observed in the allochthonous Morses Line Formation/Stanbridge Nappe. The intensity of folding within the Rosenberg sequence ranges between incipiently deformed siliciclastic carbonates, and chevron-folded dolomitic limestone layers in the vicinity of tectonic contacts

Figure 3.12.: Folding of thin-bedded shaly dolostone of the Dolomite-Shale sequence along the Interstate Highway I-89, at outcrop # 301.

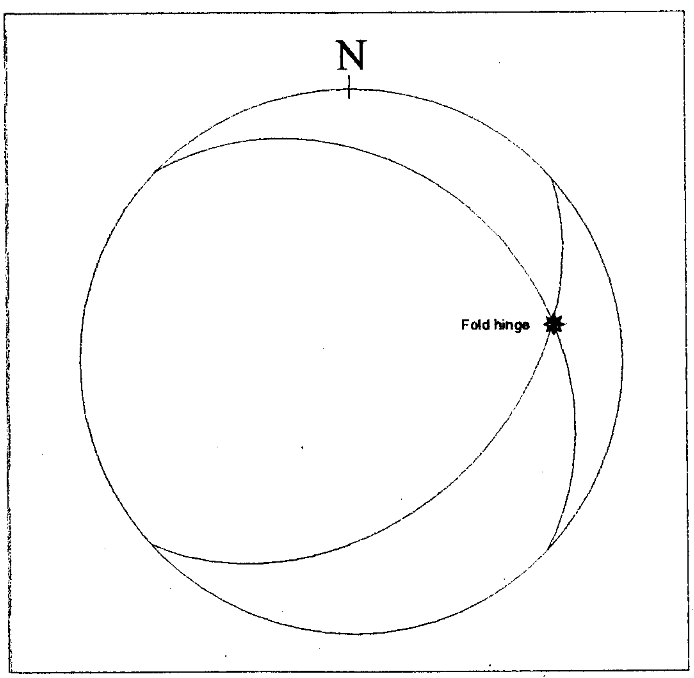


Figure 3.13.: Equal area projection of the fold structure observed at outcrop # 301 (Figure above).

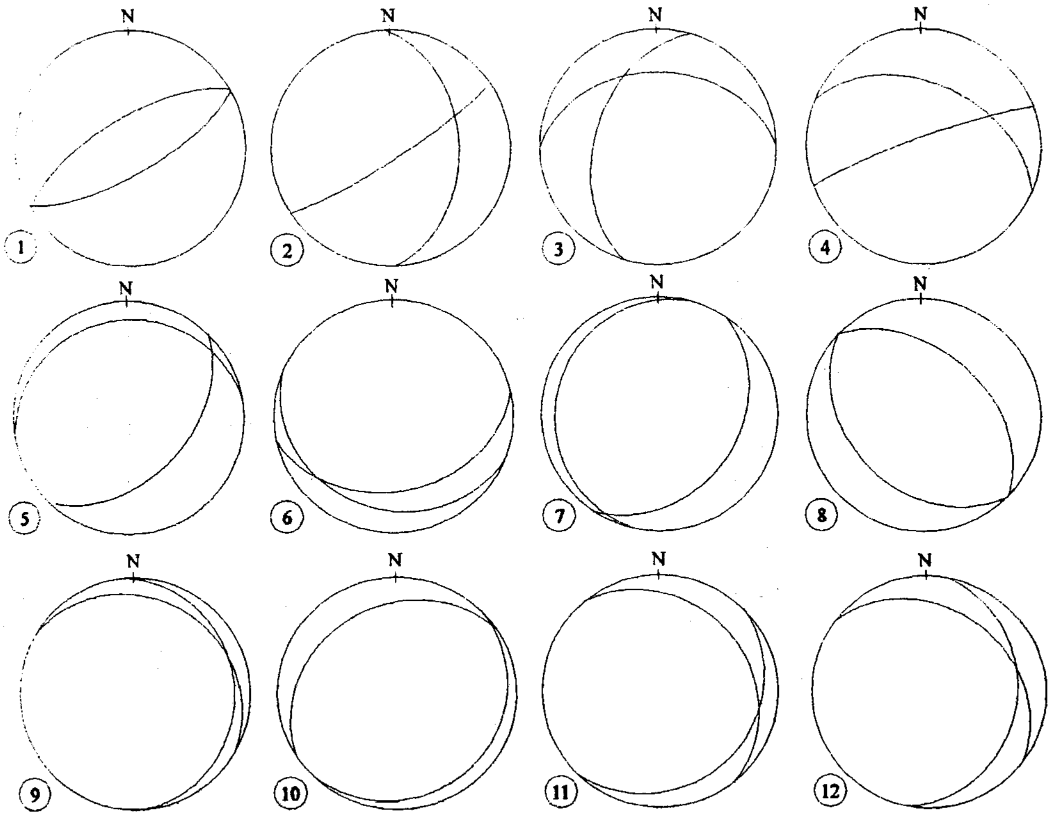


Figure 3.14.: Equal area projections of fold structures, measured within the Dolomite-Shale Sequence in the Highgate Falls and St.Albans area:

	outcrop #	*Fold hinge orientation*
(1)	228	0/241
(2)	238	39/064
(3)	239	34/336
(4)	241	36/066
(5)	242	8/052
(6)	259a	19/230
(7)	259b	4/209
(8)	259c	0/314
(9)	262	14/067
(10)	278	2/053
(11)	288	16/101
(12)	293	23/075

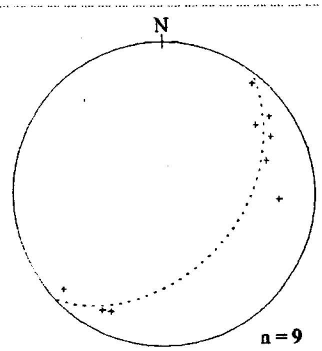
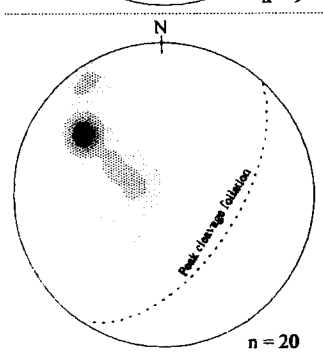


Figure 3.15.: Equal area projection of the measured bedding-cleavage intersection lineations within the Dolomite-Shale Sequence. Approximated common plane (dotted) containing all B/C intersection lineations:

Common plane: 041/42 SE



21 - 25 %

16 - 20 %

11 - 15 %

6 - 10 %

1 - 5 % (per 1 % area)

Figure 3.16.: Equal area projection of the concentration of poles to all measured cleavage foliations and their peak orientation within the Dolomite-Shale Sequence.
Approximated common plane (dotted) of all measured cleavage foliations:

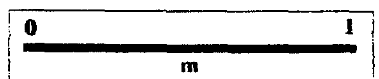
Peak position: 37/308
Peak foliation: 038/53 SE

This common plane corresponds to the measured peak of the concentration of poles to cleavage foliations and its approximated common plane (***Figure 3.16.***), and therefore suggests a thrust-induced compressional deformation, causing the development of a moderately southeast-dipping spaced cleavage. Its weakly developed fabric surfaces, best exposed within the shaly portion of the southern Dolomite-Shale sequence northwest of Skeels Corners (outcrop # 218, 228, 238, 242), are difficult to distinguish from sedimentary layering but can be identified, within the southern portion of the dolomite-shale strata, by its rather wide-spaced (1.5-5 cm) cross-cutting planar surfaces, whereas bedding appears as thin (1-3 mm) limonite-stained laminations. Moreover, where the cleavage fabric is definitely developed, it is oblique to bedding.

3.4.2.2. HIGHGATE FORMATION

Folding within the Highgate Formation exceeds the dimensions of that in the Dolomite-Shale unit (***Figure 3.17. & 3.18.***) and corresponds to fold wavelengths of about 50-100 meters. An initial analysis of structural elements within the Stanbridge Nappe in Québec, as the northern lithostructural equivalent to the Highgate and Morses Line Formation, was provided by Charbonneau [1980], who classified the Stanbridge Nappe as a distinct *"structural domain"*. Previous comparable detailed structural analyses of the Highgate and the Morses Line Formation in Vermont are non-existent. This study represents the initial attempt of characterizing fold deformation features within these units and correlating them to their northern and southern structural equivalent, using analytical structural methods.

Figure 3.17.: Fold structure in slaty limestones of the Highgate Formation at outcrop # 2.



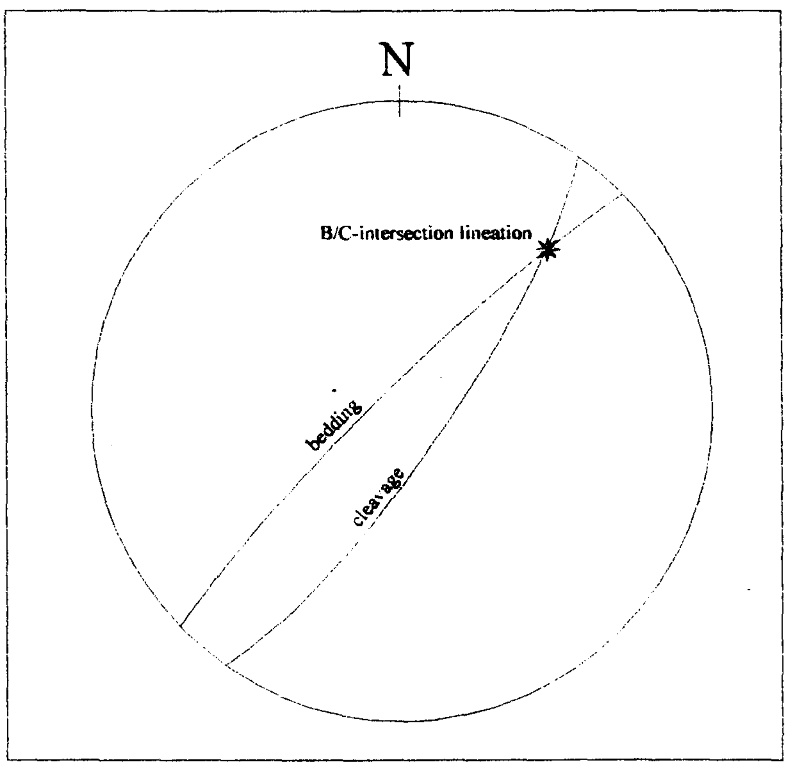
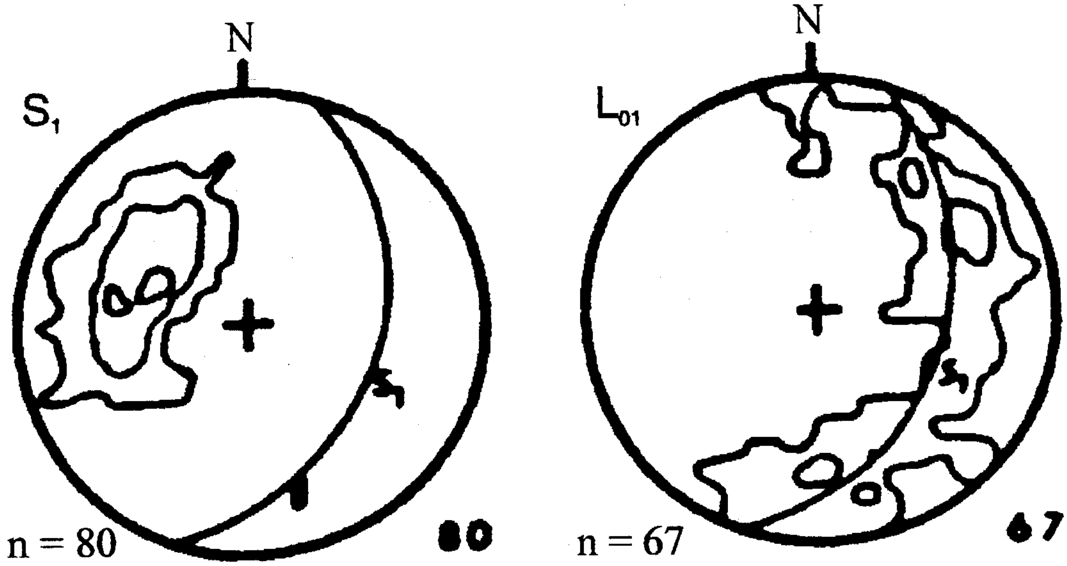


Figure 3.18.: Equal area projection of bedding-cleavage relation in the Highgate Formation, at outcrop # 1 (Figure above).

The Highgate Formation is best exposed at Highgate Falls (outcrop # 138, 139, 140, 399). At this locality the limestones exhibit a shaly to slaty, probably solution-type, cleavage. Direct evidence for a solution-type cleavage, however, is not available. This fabric continues upstream towards the lip of the falls. The new fault identified within the Highgate Formation, however, juxtaposes shaly limestones with more intensively cleaved slates.

In Québec, Charbonneau's [1980] report suggests three distinctive phases of fold deformation within the structural domain that correlate lithostructurally to the Highgate and the Morses Line Formations in Vermont. However, only one generation of fold deformation features can be recognized throughout the Stanbridge Nappe, and that corresponds to the attitude of folding and the orientation of cleavages observed in the Highgate and the Morses Line Formation in the Highgate and St. Albans area. Charbonneau's [1980] fold geometry-analysis (***Figure 3.19.***) indicates a concentration of bedding/cleavage (=B/C) intersection lineations (L_{01}) on a moderately (~50°) southeast-dipping common plane. The preferred orientation of cleavage foliations (S_1) exhibits a concentration of poles within the northwestern quadrant (***Figure 3.19.***), suggesting a moderately southeast-dipping common plane of the cleavage foliations measured by Charbonneau [1980]. The structural analysis of fold deformation features within the Highgate Formation of this study (***Figure 3.20.***) reveals orientations for both the common plane of B/C intersection lineations (***Figure 3.21.***) and the concentration pattern of poles to cleavage foliations (***Figure 3.22.***) that resemble Charbonneau's [1980] results. The similarity of the pattern of fold attitudes and cleavage orientations obtained from this study and Charbonneau's [1980] work therefore supports the lithostructural equivalence of the Stanbridge Nappe in Québec, and the Highgate and Morses Line Formations in Vermont, respectively.



- Contours of 1, 5, and 10 %, per 1 % area (Schmidt Method).

Figure 3.19.: Equal area projection of the concentration of poles to cleavage foliations (S_1) and the distribution of bedding-cleavage intersection lineations (L_{01}), measured within the structural domain of the Stanbridge Nappe in Quebec (from Charbonneau 1980).

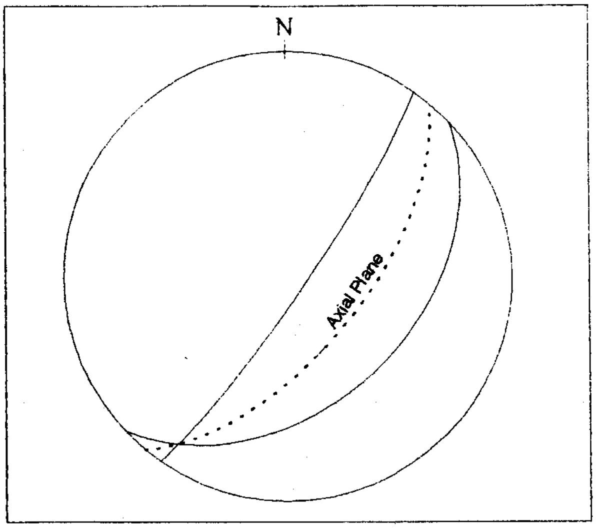


Figure 3.20.: Equal area projection of a fold structure observed within the Highgate Formation at outcrop # 113:

Fold hinge lineation: 12/214
Axial plane foliation: 040/62 SE

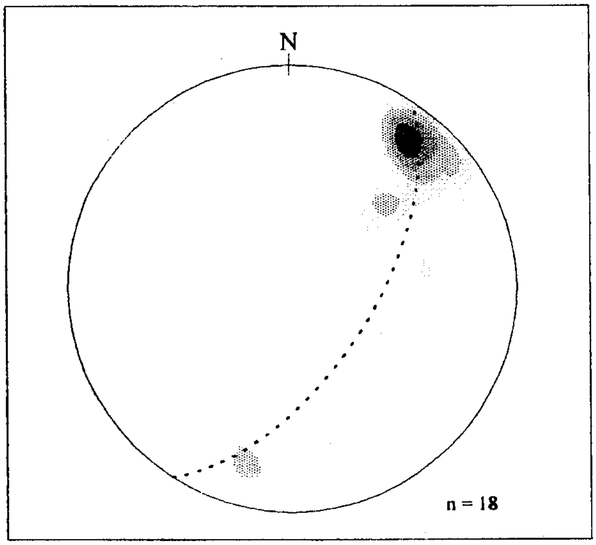


Figure 3.21.: Equal area projection of the concentration of bedding-cleavage intersection lineations within the Highgate Formation.
Approximated common plane (dotted) of all measured B/C intersection lineations:
Peak position: 18/037
Common plane: 033/61 SE

13 - 15 %

10 - 12 %

7 - 9 %

4 - 6 %

1 - 3 %

(per 1 % area)

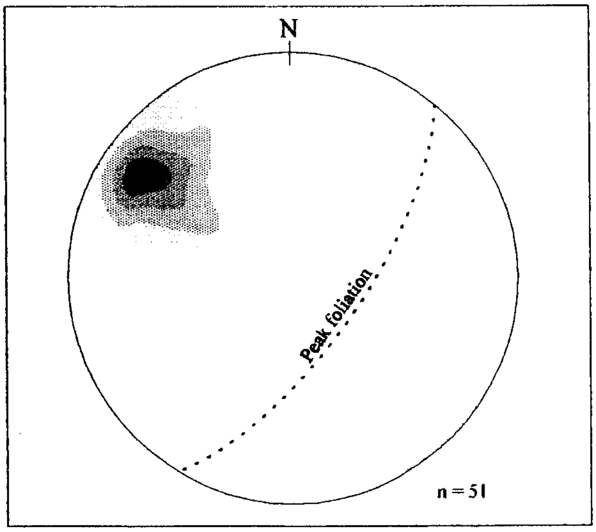


Figure 3.22.: Equal area projection of the concentration of poles to all measured cleavage foliations and their peak orientation within the Highgate Formation:

Peak position: 21/304
Peak foliation: 034/69 SE

20 - 24 %

15 - 19 %

10 - 14 %

5 - 9 %

1 - 4 %

(per 1 % area)

3.5. SUMMARY ROSENBERG SEQUENCE

Structurally, the Rosenberg slice itself has been described by previous workers as forming part of a broad, shallow northeastward-plunging syncline, the *St. Albans Synclinorium* [Clark 1934, Cady 1945, Shaw 1958]. This characterization, however, includes all the Morses Line area as far east as the Hinesburg Thrust. The most recent study of the St.Albans and Highgate area [Mehrtens & Dorsey 1987] follows this earlier model and restricts the degree of deformation and faulting to a *"minor amount"* and proposes *"rapid lateral facies changes"* between the *"western shelf"* and the *"eastern basinal sequence"* within the broadly-defined Rosenberg slice.

The results of my detailed lithostructural mapping project, however, revealed a dramatically different and far more complicated structural geometry and nature of the contact of the Rosenberg sequence and the overlying Morses Line Formation than previously proposed (***Plate 3***). In my view, the Rosenberg thrust slice, exposed in the St. Albans and Highgate area, and its northern continuation in Québec, is a gently east-dipping homoclinal slice, and represents a large, more-or-less coherent, sample of the near-shelf-edge Cambrian siliciclastics and carbonates. The slaty succession east of the Highgate Formation is, I propose, largely part of a separate, allochthonous thrust slice, the Stanbridge Nappe, and should therefore not be included within the Rosenberg slice.

The idea can be presented that the entire Champlain thrust sheet in western Vermont can be viewed as an anomalously large carbonate sliver, attached to the Stanbridge Nappe in the same way as smaller ones are attached to the base of the Taconic Allochthon [Rowley & Kidd 1981]. A carbonate slice of such dimensions, however, widely exceeds the size of the slivers found at the base of the Taconic Allochthon, and which nowhere else includes units lower than Chazy limestones [pers. comm. Kidd 1994]. This interpretation of the Champlain thrust sheet as an anomalously large carbonate sliver, does not help, however, to explain the apparent absence of mélange at the western edge of the Taconic Allochthon rocks (Morses Line Formation/Stanbridge Nappe) in northwestern Vermont, nor the apparent absence of an early Ordovician shelf/shelf edge section above the late Cambrian to earliest Ordovician limestones of the Highgate Formation.

3.6. ALLOCHTHONOUS SEQUENCE

It is interpreted that internal thrust imbrication in the allochthonous sequence in the study area created a north-northeast to south-southwest trending thrust stack within the area mapped as the Morses Line Formation. Proposed slice boundaries typically coincide with topographic breaks and are locally marked by highly fractured slivers of shelf to shelf edge derived siliciclastic carbonates (outcrop # 205, 206, 207), interpreted as originally attached to the moving Taconic thrust front, or intercalated lenticular limestone breccia bodies (outcrop # 141, 206, 327, 330).

A distinctive, moderately (44-60°) southeast-dipping slaty cleavage is the dominant deformation feature within the slates of the Morses Line Formation (***Figure 3.23. & 3.24. & 3.25.***, outcrop # 197, 198, 199, 324, 325, 433). This foliation, however, differs from the less intense but steeper dipping (70-80°) axial planar cleavage within the Highgate Formation (***Figure 3.22.***, outcrop # 5, 6, 8) by exhibiting a reduced dip (~15°) of the cleavage plane. Nevertheless, the consistent north-northeast to south-southwest trend of both cleavages (strike 027-034°) suggests that both foliations are due to the same deformational event.

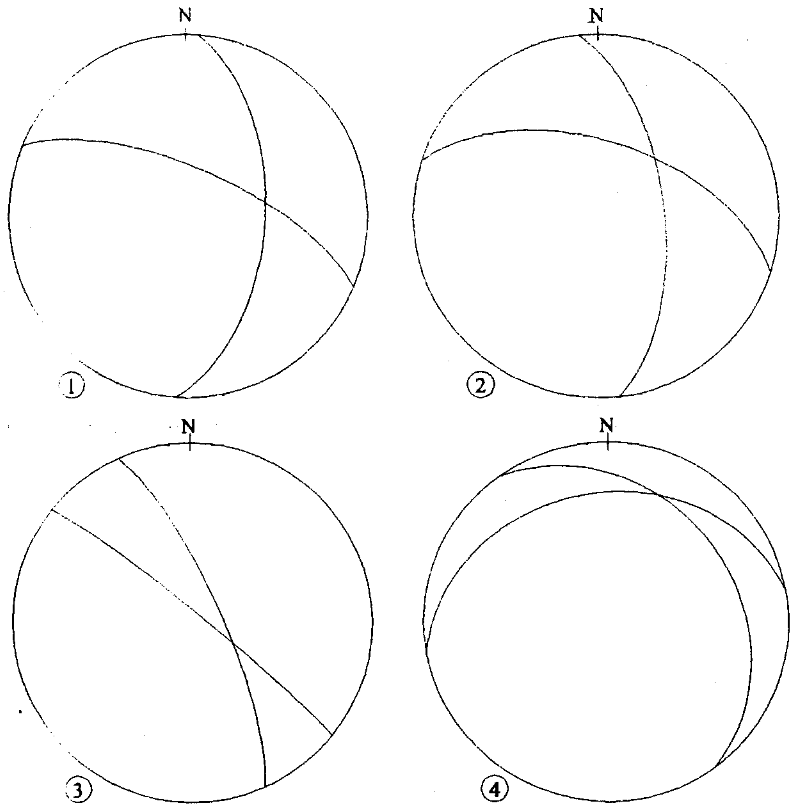


Figure 3.23.: **Equal area projection of fold structures, measured within the Morses Line Formation in the St. Albans and Highgate area.**

	outcrop #	*Fold hinge orientation*
(1)	141a	53/080
(2)	141b	53/044
(3)	211	69/116
(4)	218	25/022

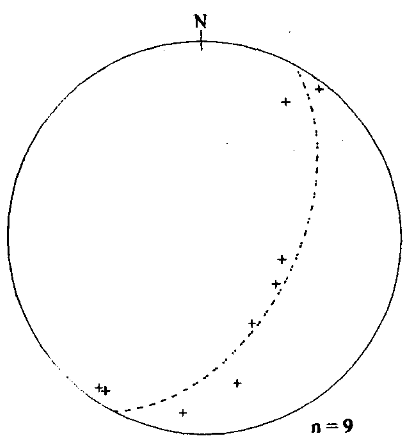


Figure 3.24.: **Equal area projection of the distibution of bedding-cleavage intersection lineations within the Morses Line Formation.**

Approximated common plane (dotted) of all measured B/C intersection lineations:
027/51 SE

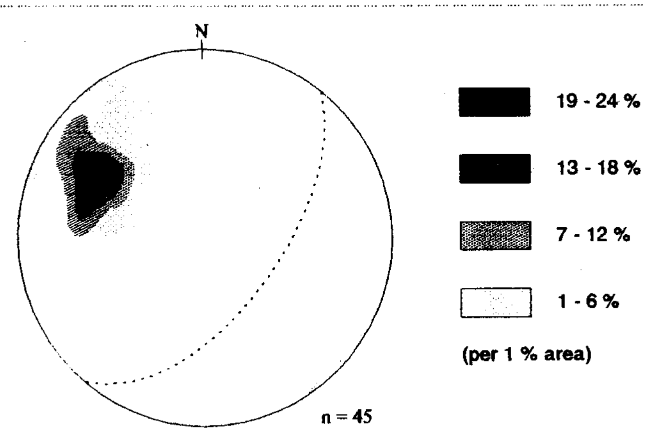


Figure 3.25.: **Equal area projection of the concentration of poles to all measured cleavage foliations and their peak orientation within the Morses Line Formation.**

Approximated common plane (dotted) of all measured cleavage foliations:

Peak position: 36/301
Corresponding foliation: 031/54 SE

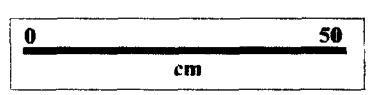
However, this small but important variation in dip between the average cleavage surfaces is proposed to indicate a structural discontinuity between the Highgate and the Morses Line Formation.

The allochthonous character of the Morses Line Formation has not been accepted by all workers in this area. St.Julien and Hubert [1975] proposed the allochthonous Stanbridge Nappe and its correlation with the Upper Highgate and the Morses Line Formation, a proposal endorsed by the work of Charbonneau [1980] and Globensky [1981]. However, Mehrtens and Dorsey [1987] propose that the *"shales"* of the Morses Line Formation represent a *"largely undeformed"* and *"stratigraphically intact"* continental slope-rise deposit.

This study follows the hypothesis that the Morses Line slate package, that includes the slaty and intensely cleaved portion above the Highgate Formation at the Missisquoi River gorge (***Figure 3.26. & 3.27.***, outcrop # 140), is the lithostructural equivalent and structural continuation of the Stanbridge Nappe into northwestern Vermont, and therefore mainly follows the structural outline of Charbonneau [1980] and Globensky [1981]. The structural evidence at Highgate Falls suggests that the separated (upper) slice of Gorge Formation is one of several, lithostructurally distinct out-of-sequence thrust slivers, attached to the westward moving thrust wedge of the Taconic Allochthon. Carbonate slivers are present in many places at the margin of the Taconic Allochthon [Rowley & Kidd 1981]. The occurrence of slivers of shelf carbonate and related rocks at the sole of several allochthonous slices is most easily interpretable as the product of late-orogenic large-scale imbrication of a single initial thrust surface, with carbonates attached to the sole thrust from the shelf.

Figure 3.26.: Folding of lenticular boudins within slates of the Morses Line Formation at outcrop # 140, Missisquoi River gorge at Highgate Falls. Dolomitized boudins represent remainders of bedding, subsequently folded and cleavage penetrated.

Figure 3.27.: Intensely cleaved slates of the Morses Line Formation at outcrop # 140, Missisquoi River gorge at Highgate Falls. Note extensional northeast-southwest trending calcite vein set.



3.6.1. MORSES LINE FORMATION

Fold deformation features within the Morses Line Formation generally comprise discontinuous isoclinally folded layers, characterized by small fold wavelengths of tens of centimeters. Its attitude of folding differs from that of the Highgate Formation by exhibiting less steeply northeast-dipping axial planes accompanied by mostly moderately to steeply (53-69°) northeast-plunging fold hinges (***Figure 3.23. & 3.24.***). According to Vollmer [1981], the steepening of fold hinge lines from an overall horizontal attitude to a steeply plunging one, together with their rotation from a northeastern (southwestern) orientation into a more east-southeastern trend reflects a substantial increase in strain. Following Vollmer's [1981] work, the rotation of fold hinge lines within highly strained, isoclinal folded lithology, as in the Taconic slates of the Morse Line Formation/Stanbridge Nappe, can be accomplished by assuming a heterogeneous distribution of strain, that is the varying intensity of strain within the northwestward moving Taconic thrust front. As a consequence, the development of a shear component within this high strain assemblage causes the rotation of isoclinal fold hinge lines, which can approach the direction of maximum finite extension [Vollmer 1981]. This structural feature can be interpreted in the slate assemblage of the Morse Line Formation in the study area (***Figure 3.23.***).

The common plane to the measured B/C intersection lineations in the Morses Line sequence corresponds to the north-northeast to south-southwest oriented common plane of the B/C intersection lineation data of the Highgate Formation, but dips significantly (~10°) steeper than in the Highgate Formation (***Figure 3.24.***). In addition, the concentration of poles to the measured cleavage foliations within the Morses Line Formation corresponds to a common plane, that is less steeply southeast-dipping (~15°) than the Highgate Formation (***Figure 3.25.***). Furthermore, the cleavage foliation surfaces within the Morses Line Formation are generally closer spaced (1-3 mm), and the cleavage is as a result more prominent than the cleavage induced surfaces of the Highgate Formation.

This small but important difference of cleavage and fold hinge orientation between the Highgate and the Morses Line Formation is interpreted to indicate a structural discontinuity between these units and may support the consideration of the Morses Line Formation/Stanbridge unit as a separate structural domain in northwestern Vermont.

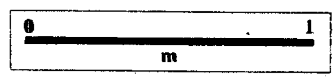
3.7. HIGH ANGLE FAULTING

High angle faulting, mainly northeast-southwest trending, has been mapped along the Champlain Thrust fault and in many parts of the foreland basin sequence [Welby 1961; Doll et al. 1961]. In my study area, however, it has been completely neglected by earlier workers. Previous as well as recent reports about the Highgate and St. Albans area [Schuchert 1937; Cady 1945; Shaw 1958; Mehrtens & Dorsey 1987] lack detailed descriptions about cross-strike high angle faulting, particularly within the Rosenberg sequence, although the topography clearly shows its existence.

High-angle faults within the *Rosenberg sequence* are most easily recognized by the lateral displacement of north-south trending topographic ridge segments (***Plate 1***). Lateral, northeast-southwest oriented offset of those ridge segments may range between tens and hundreds of meters. Rather more difficult to identify, due to the poor exposure in general, are high-angle fault occurrences within the siliciclastic dolostone section and the dolostone-shale sequence of the Rosenberg slice. Some fault plane exposures can be identified by the juxtaposition of well-bedded dolostone to rather slaty lithologies (***Figure 3.28.***). Measurements, where available, indicate steeply northwest-dipping (outcrop # 305) normal faulting (***Figure 3.29.***), involving some, mostly left-lateral strike-slip. The rare exposures of actual fault surfaces also complicates the solution of the timing of high angle faulting in the Champlain Thrust System. Seismic imaging, however, shows that some of the faults are older than the Champlain thrust fault whereas others are younger, probably Mesozoic in age [Stanley 1980]. Synorogenic normal faulting due to flexural extension [Bradley & Kidd 1991] could be responsible for the juxtaposition of shallow water siliciclastics and carbonates with shaly lithologies. However, most of the synorogenic normal faults that accompanied the subsidence of the Taconic foredeep strike parallel with the thrust front [Bradley & Kidd 1991], that is largely north-south in this area. The trend of normal faulting observed in the study area mostly strikes at a high angle to the thrust front of the Morses Line Formation/Stanbridge Nappe. Therefore, normal faulting observed in the Highgate area cannot be correlated with the normal faults that are due to flexural extension and documented from the Taconic foredeep in the Mohawk Valley in New York [Bradley & Kidd 1991]. High-angle normal faulting in the field area is better explained by representing post and/or latest-orogenic extensional normal and/or tear faults. These fault features are best observed where the Gorge-Highgate Formation contact is displaced across strike.

High-angle fault exposures in the *Philipsburg sequence* (***Figure 3.30. & 3.31.***; outcrop # 316, 343) are restricted to subvertical scarps along I-89 and quarry outcrops and exhibit mostly thick-bedded, partially fault-dragged carbonate layers along moderately (50°) northwest and also steep (80°) northeast-dipping surfaces.

Figure 3.28.: High angle normal faulting within the Dolomite-Shale sequence, Rosenberg slice, displaying slickenside lineations, at outcrop # 305.
Fault juxtaposes massive dolostone with vertical oriented dolomitic shale layers.



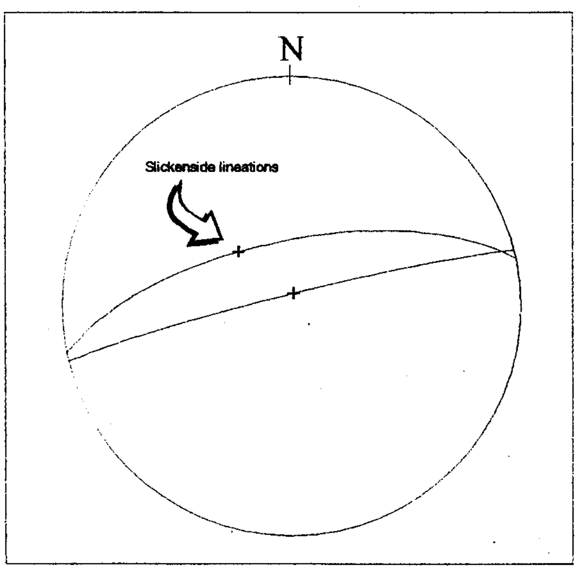


Figure 3.29.:Equal area projection of the orientation of measured fault surfaces and slickenside lineations of high angle faulting within the Dolomite-Shale sequence,Rosenberg slice, exposed at outcrop # 305.

Figure 3.30.: High angle faulting within the Strites Pond Formation, Philipsburg slice, exhibiting slickenside lineations at outcrop # 316.

0 1
m

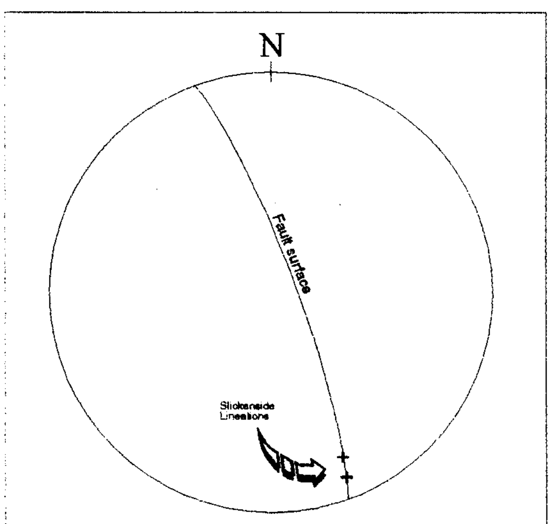


Figure 3.31.: Equal area projection of fault surfaces and slickenside lineations of high angle faulting observed within the Strites Pond Formation, Philipsburg sequence, outcrop # 316.

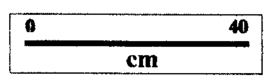
3.8. VEINS

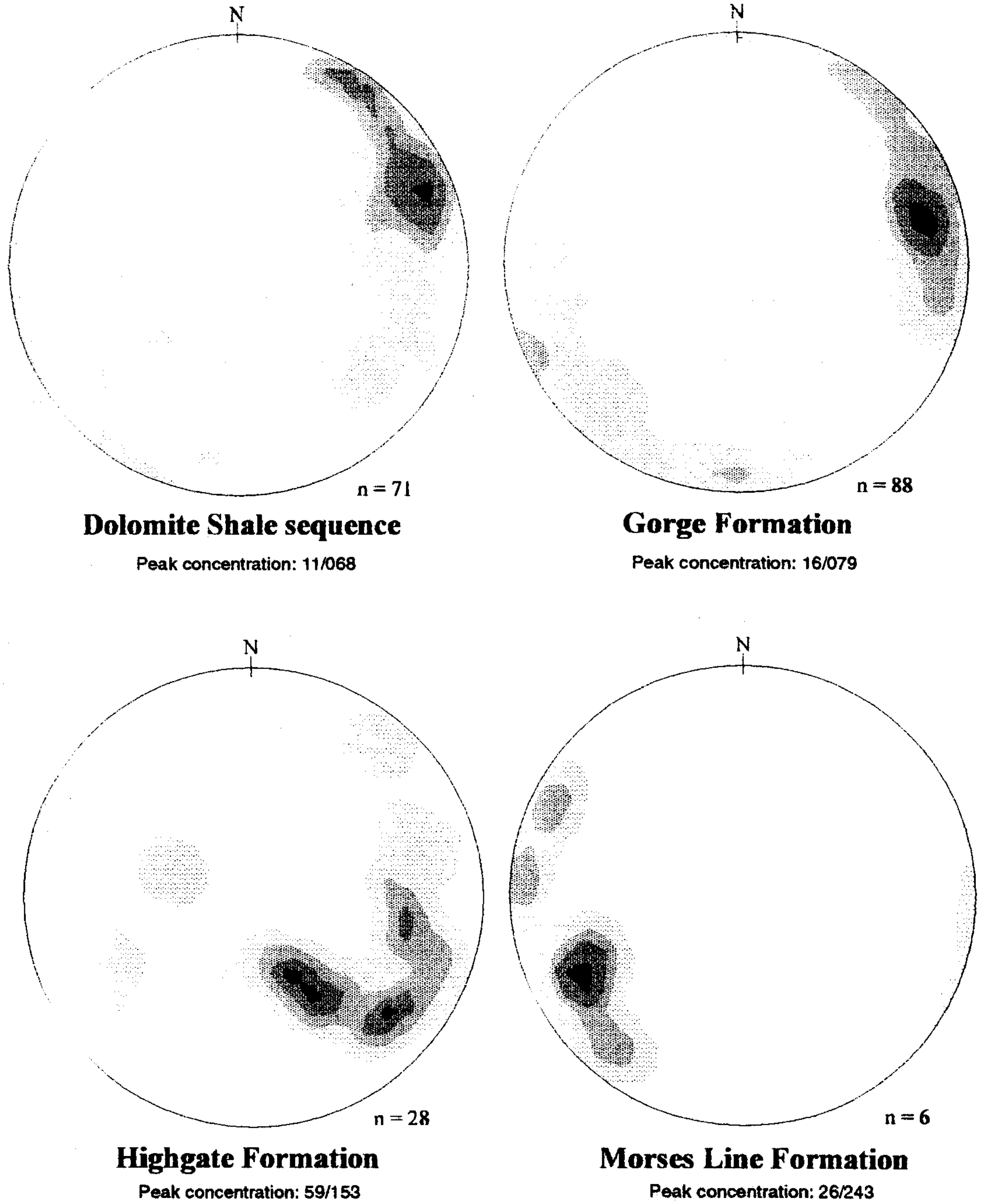
Veins and cavity fillings are characteristic for the dolostone facies throughout the Rosenberg sequence in the study area. They generally occur as a framework of closely spaced (3 cm - 1.5 m), quartz-filled (1 mm - 3.5 cm thickness) cavity fillings of continuous, partly en-echelon oriented fractures within the dolomitic rocks (***Figure 3.32.***). In particular the dolostones of the Gorge Formation and the Dolomite-Shale sequence and, to a lesser degree, the Highgate Formation exhibit a characteristic vein assemblage, partly associated with irregular shaped quartz-filled cavities and knots. Veins within the Highgate and the Morses Line Formation (***Figure 3.27.***) are thinner (1 mm - 1 cm), closer spaced (0.5 - 5 cm) and consist predominantly of calcite.

Previous authors utilized the abundance of quartz-veins and knots within the Gorge Formation [Doll et al. 1961] and, in west-central Vermont, the Clarendon Springs Dolomite [Dennis & Stone 1964] as an additional characteristic feature for identifying those units in the field. Earlier work, however, lacks detailed measurements and structural analyses of these vein sets. This study demonstrates a distinguishing pattern of vein development within the Gorge Formation and the Dolomite-Shale sequence that probably reflects post and/or late-orogenic phases of extensional deformation within the Champlain Thrust System.

Most veins within these units are represented by steeply southwest-dipping planar surfaces (***Figure 3.33.***), partly associated with minor offsets (0.2-1.5 cm), whereas veining within the Highgate Formation is concentrated along moderately to steeply northwest-dipping planes. In contrast, the veins within the allochthonous Morses Line Formation occur as mainly steep northeast dipping surfaces. Moreover, veining observed within the allochthonous sequence is more

Figure 3.32: Quartz-vein set within the Dolomite-Shale sequence. Feather veining oblique to a second widely spaced vein set at outcrop # 132.





16 - 20 %

11 - 15 %

6 - 10 %

1 - 5 %
(per 1 % area)

Figure 3.33.: Equal area projections of the concentration of poles to all measured vein surfaces throughout the Dolomite -Shale, Gorge, Highgate and the Morses Line Formation.

abundant and shows multiple phases of fracturing, such that older veins are folded and crosscut by younger veins.

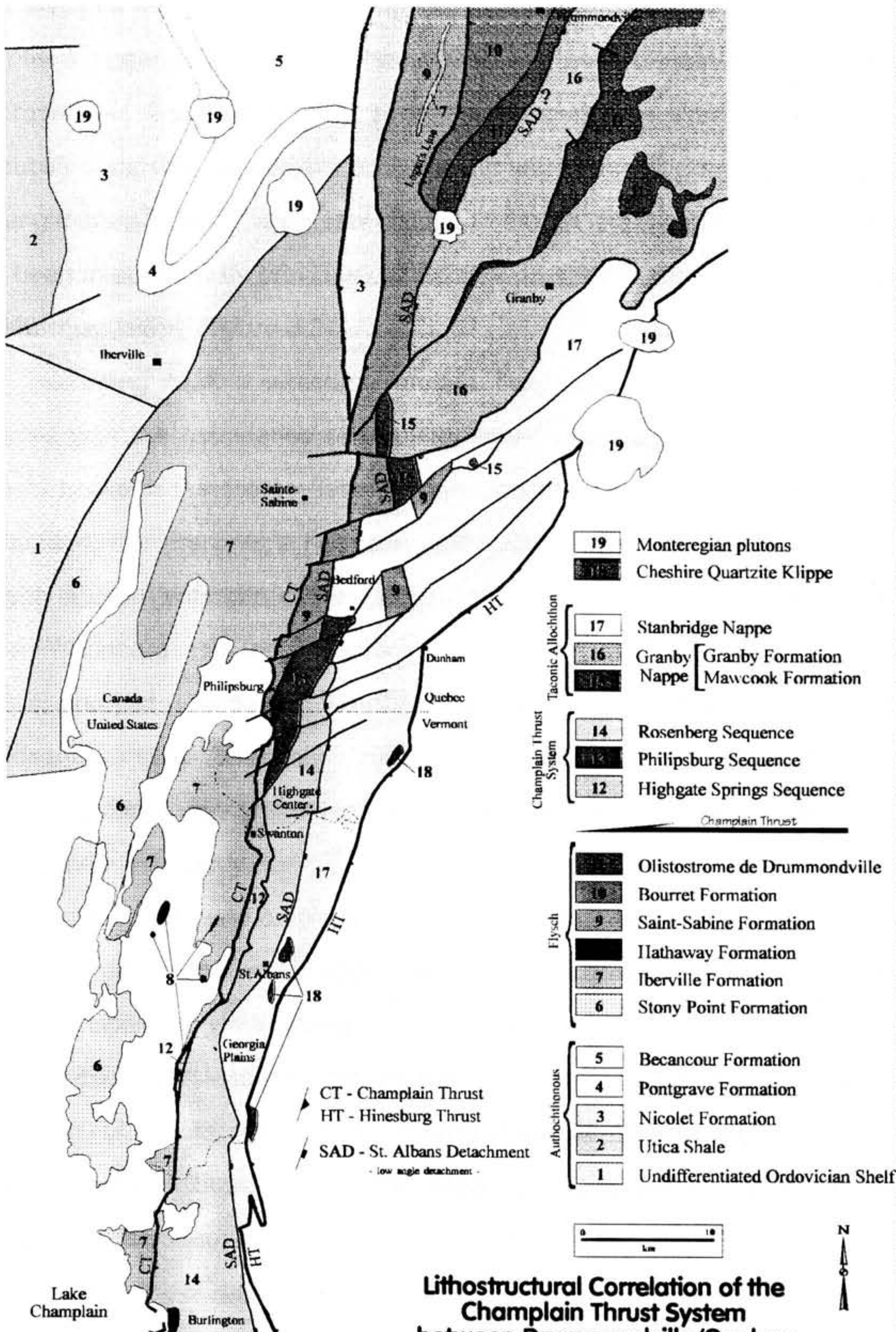
All veins cross-cut cleavage surfaces, where developed, at a high angle and therefore suggest phases of largely north-south trending extensional deformation, causing the development of a suite mostly east-west trending normal faults and a distinctive vein pattern; in particular within the brittle-behaving quartzose dolostones of the Rosenberg sequence. The lack of cross-cutting relations with deformation features younger than those due to the Taconic orogenic event, however, prevents the determination of more accurate timing of fracturing.

3.9. DISCUSSION STRUCTURE

Recent seismic reflection studies north and south of my study area [Ando et al. 1983] show that the Champlain Thrust dips eastward as far as beneath the Green Mountains. The suspicious, abrupt northern end of the Champlain Thrust in Québec raises the question of whether the Champlain thrust fault in northwestern Vermont really represents the same major thrust horizon as the one in the Burlington region where the thrust relations are well established [Stanley 1987].

Alternatively, the Rosenberg Thrust could represent an older thrust horizon relative to the Philipsburg Thrust, so that the Rosenberg thrust fault is cut by the younger Philipsburg Thrust. Following this interpretation, the Champlain Thrust, that is the thrust fault that tectonically juxtaposes siliciclastics and/or shelf carbonate strata with autochthonous shale lithologies of the Taconic foreland basin, is represented by the continuation of the Philipsburg and Highgate Springs Thrusts further north, where they cross the international border to Québec as Logan's Line (["Faille Logan", Avramtchev 1989], ***Figure 3.34.***). If this correlation is correct then the Champlain thrust zone extends not only from Rosenburg, Canada, to east-central New York, as suggested by Stanley [1987], but continues in Québec as roughly paralleling Logan's Line where it juxtaposes medial Ordovician shale and flysch-type lithologies of the Sainte-Sabine Formation with a suite of allochthonous medial Ordovician mélange [Avramtchev 1989], and rocks belonging to the Taconic Allochthon (Granby Nappe).

This alternative tectonic model involves more intricate fault relations such that the Champlain thrust horizon, as it is exposed in the Burlington region [Stanley 1987], is cut by the relatively younger Philipsburg Thrust in the Highgate



Lithostructural Correlation of the Champlain Thrust System between Drummondville/Quebec and Burlington/Vermont

(modified after Doll et al. 1961 & Avramtchev 1989)

Compilation & Graphics by Michael R Haschke 11/94

Figure 3.34.

Springs region that continues throughout southern Québec as a substantial thrust fault [Logan's Line, Avramtchev 1989], and probably represents a major late-orogenic fault within the thrust zone. Furthermore, this structural alternative introduces a closely spaced suite of substantial northeast-southwest trending high angle normal and/or tear faults oblique to the general north-south trend, that have been interpreted by previous Canadian workers as thrust slice boundaries ([Avramtchev 1989], ***Figure 3.34.***).

According to this structural model, the trend of high angle faulting coincides with the boundaries of lithostructural "windows" and other uncertain contacts between the Sainte-Sabine Formation and the carbonate thrust slice assemblage, the Stanbridge Nappe and the Granby Nappe, respectively, that suddenly appear just north of the international border, along the structural trend of the Philipsburg and the Rosenberg slice, within the Stanbridge Nappe in Québec, and just northeast of the abrupt disappearance of the Philipsburg and the Rosenberg slice [Avramtchev 1989]. These proposed faults are of the same family as those forming the high angle fault pattern in the Highgate area documented in this study.

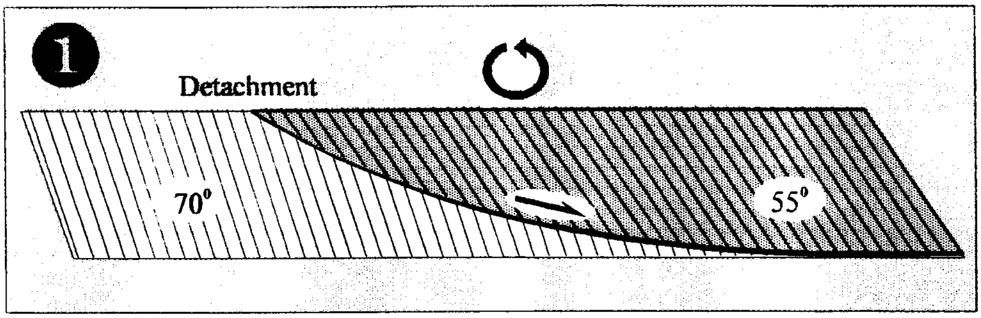
The fault relations employed in the Highgate area and southern Québec alternatively explain the abrupt end of the siliciclastic and carbonate successions (Highgate Springs-, Philipsburg-, Rosenberg-) in Canada. However, this structural geometry leads to the problem of the contrast in thickness and abundance of medial Ordovician mélange-type lithologies adjacent to the Taconic Allochthon between Fort Ann, New York, and the latitude of Granby, Québec, and its sudden increase in abundance north and south of those places. The complete lack of mélange, on the other hand, adjacent to the Morses Line Formation and most of the Stanbridge Nappe may provide the key for understanding the structural relations in northwestern Vermont.

This study proposes a substantial structural break between the siliciclastics and carbonates of the Rosenberg sequence and the allochthonous slates of the Morses Line Formation/Stanbridge Nappe (***Plate 3***). The map pattern (***Figure 3.34.***), together with the distinctive less steep dipping average cleavage foliation within the Morses Line Formation relative to the Highgate Formation, suggests a detachment surface between the two units (***Figure 3.35/1***). The fault geometry requires a listric nature of the fault contact that is able to account for the rotation of the cleavage foliation within the Morses Line Formation. Following this model, the detachment surface probably coincides at depth with a reactivated low angle Taconic fault surface; possibly the Champlain thrust horizon (***Plate 3***). The straight outcrop pattern next to the inferred fault trace, however, prevents to elaborate a more exact geometry of the detachment at depth.

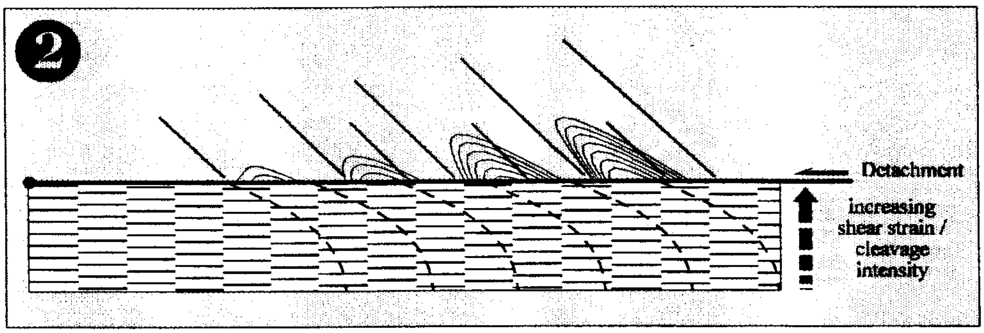
However, the contrast in cleavage steepness only allows alternative explanations (***Figure 3.35/2-3***) such that a planar finite strain within a cleaved strata subsequently undergoes shear distortion. This deformation mechanism could account for a moderately cleaved, steeply dipping sequence that is juxtaposed to a intensively cleaved, less steeply dipping sequence, but does still not provide a satisfying explanation for the absence of mélange in the study area.

The detachment surface between the Rosenberg succession and the Morses Line Formation/Stanbridge Nappe could have developed in response to one of the latest phases of deformation within the Taconic orogen, or post-Taconic, possibly as a result of Mesozoic extension. This detachment would have caused a mild counter-clockwise (looking north) rotation (~15°), of the correct sense, of cleavage foliation within the Morses Line Formation/Stanbridge Nappe that corresponds to the reduced cleavage dip, relative to the Highgate

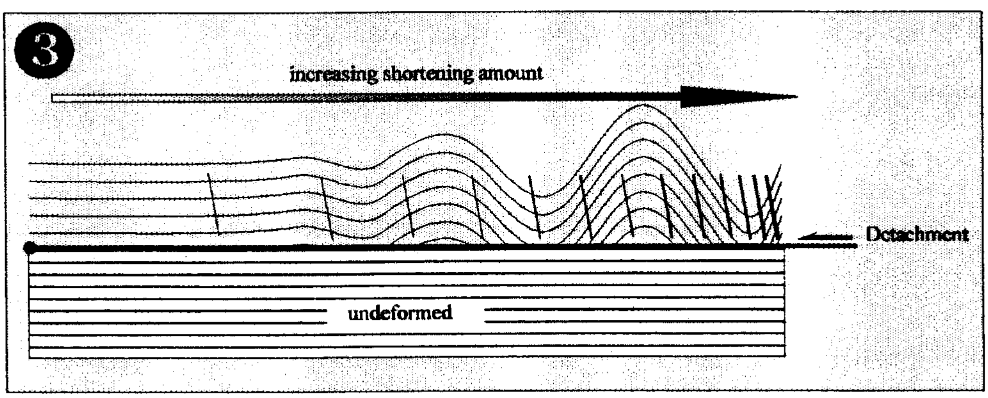
Figure 3.35.: SCHEMATIC MODEL FOR CLEAVAGE DEVELOPMENT



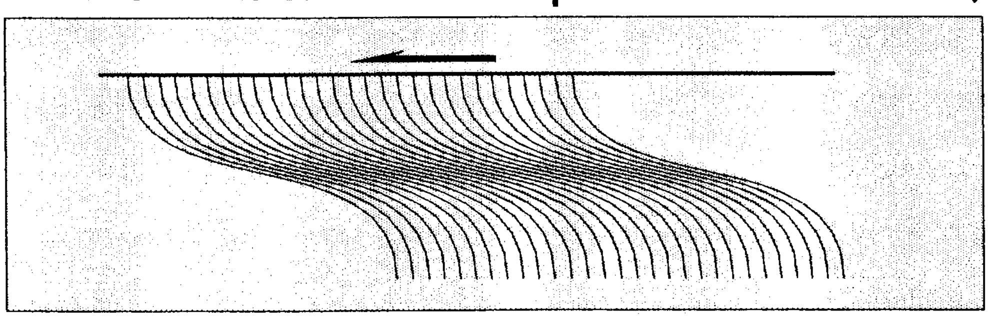
Counter-clockwise rotation of cleavage foliation due to detachment fault.



Increasing shear strain towards the detachment.



▲Planar finite strain - subsequent shear distortion▼



Formation (***Figure 3.35/1***). In addition, this fault, or a single splay of it could be exposed as the new fault contact observed within the Highgate Formation (between the shaly limestone and the slaty portion of the Morses Line Formation) in the Missisquoi River gorge at Highgate Falls (***Figure 3.36.***). At this locality, late and/or post-Taconic imbrication of the Gorge-Highgate Formation thrust sliver predates the St. Albans Detachment, such that thrusting occurred after the detachment.

The timing of the detachment scenario cannot be accurately determined. However, cross-cutting relations of the St. Albans detachment with the high-angle cross-faults indicate that cross-faulting occurred later than the development of the detachment surface. The timing of faulting could reach as far as into the Mesozoic and be related to the Newark/Connecticut graben system [Stanley 1980].

The newly identified detachment surface is interpreted to follow the distinctive boundary between shallow water siliciclastics and carbonates (Dunham Dolomite, Monkton Quartzite, Winooski Dolomite, Danby Quartzite, Clarendon Springs Dolomite) and intensely cleaved slates of the continental slope and/or rise (Skeels Corners Slate, Morses Line Slate), and provides an alternative explanation of the sudden change from shallow water sediments into deeper water deposits of the continental slope and/or rise between Burlington and the northern end of the Rosenberg slice. Recent mapping projects south of my field area, near Orwell, Vermont [Fisher 1984, pers. comm. Kidd 1994], revealed similar map patterns and structural relations that support the structural model proposed in this study.

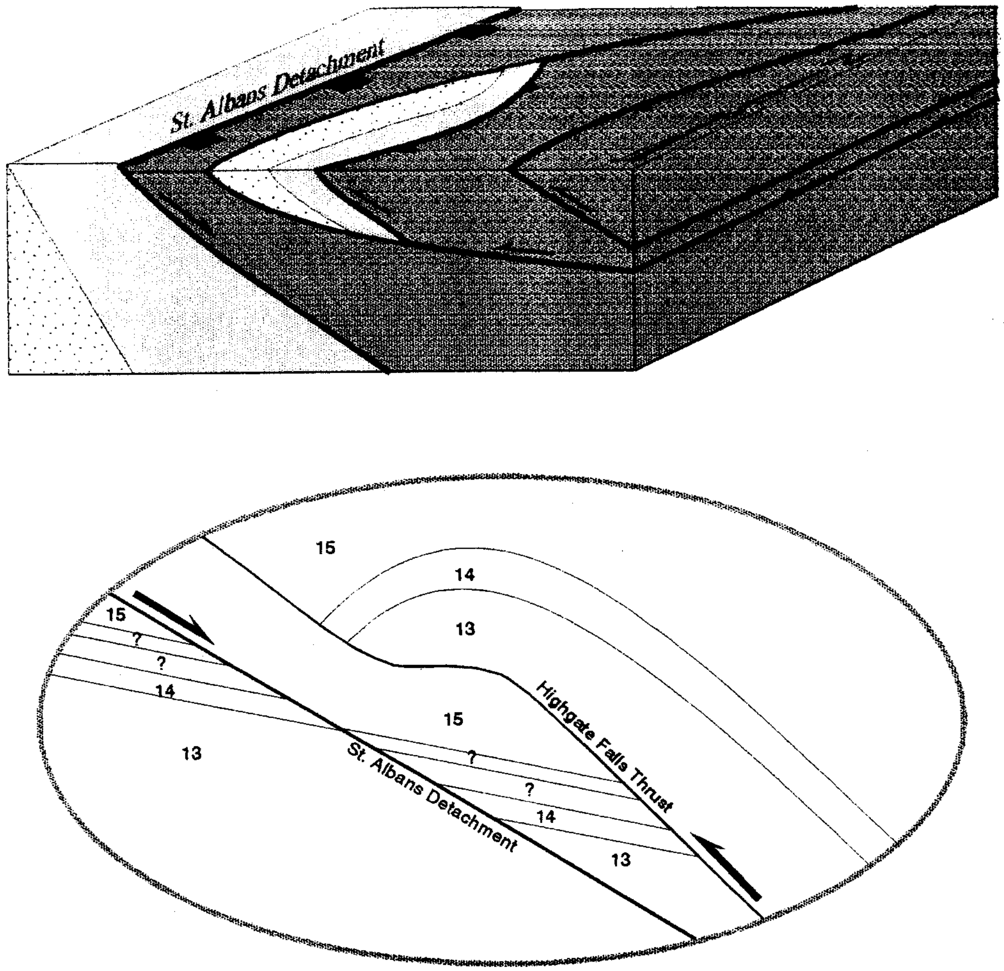


Figure 3.36.: Structural geometry at Highgate Falls. Gorge Formation thrust sliver attached to a thrust branch of the internally imbricated Morses Line Formation.
The new fault/low angle detachment identified in the Missisquoi River gorge juxtaposes slates of the Morses Line Formation with shaly limestones of the Highgate Formation.
Post-detachment imbrication of the Gorge-Highgate sequence along the Highgate Falls thrust.

15 Morses Line Formation

? medial Ordovician melange ?

14 Highgate Formation

13 Gorge Formation

low angle detachment
- St. Albans Detachment -

thrust

(not drawn to scale)

3.10. SYNOPSIS — STRUCTURE

In review of the structural features observed and analyzed in the Highgate and St. Albans region, the model of a largely undeformed, coherent and stratigraphically intact sequence, as suggested by previous workers throughout this century, does not correspond to the complex structural relations in this area. In spite of the relatively well-established thrust relations between the Highgate Springs-, the Philipsburg- and the Rosenberg slices, the structural relations within and adjacent to the Rosenberg sequence (Morses Line Formation/Stanbridge Nappe) has been rather poorly documented by earlier workers. The results of this detailed structural study, together with present Canadian contributions, suggest a new tectonic model and dramatically modify the interpretation of the structural geometry of this region:

- The Champlain Thrust System in the study area is characterized by an about 3-5 km wide belt of shallow east-dipping, thrust imbricated and locally dismembered siliciclastic carbonate thrust sheets (Highgate Springs-, Philipsburg- and Rosenberg slice; ***Figure 3.34.***) that embody the foreland of the Taconic Allochthon. This thrust slice assemblage overthrusts mildly deformed shales and flysch-type lithologies to the west. A lithostructurally distinctive suite of heavily deformed slates that strongly resemble the Taconic Allochthon, both lithologically and structurally, borders and overlaps the Champlain Thrust System to the east. This study extrapolates the largely north-south trending Champlain Thrust System from Drummondville, Québec, throughout the study area, over the Milton and Middlebury region, to the Burlington region, and down to the Helderberg unconformity in east-central New York [Stanley 1987, Plesch 1994].

- The *Highgate Springs slice* is best characterized as a narrow, longitudinally extended, continuously north-south trending sequence of dismembered fold fragments, cross-cut by numerous northeast-southwest trending high-angle faults, that is carried by the Highgate Springs Thrust. The mild deformational nature of this thrust slice is suggested by the only incipiently developed fracture cleavage.
- The wedge-shaped *Philipsburg slice* in northwestern Vermont represents a thrust sliver between the Highgate Springs and the Rosenberg slice, carried by the Philipsburg Thrust. The carbonate assemblage of the Philipsburg sequence is folded into a large, gently northeast-plunging syncline and displays largely uncleaved strata.
- The *Rosenberg slice* forms a gently southeast-dipping, locally internal imbricated (Saxe Thrust) homoclinal sequence. The Champlain Thrust, that carries the Rosenberg succession, continues north of the Burlington, Vermont, area but cuts through the carbonate slice thrust zone at the latitude of Highgate Springs village, so that the major thrust displacement is accomplished by the Highgate Springs Thrust in this region. A closely spaced suite of northeast-striking high angle normal faults cross-cut the carbonate slice assemblage and probably contributes to its abrupt termination in southern Québec.
- The Rosenberg slice is tectonically juxtaposed to the Morses Line Formation/Stanbridge Nappe. This study interpretes this fault contact as a detachment surface (*"St. Albans Detachment"*) that juxtaposes the Rosenberg sequence with deposits of the medial Ordovician Allochthon, and, in addition, accounts for the explicit absence of mélange along the eastern edge of the Rosenberg slice in the study area (***Figure 3.34.***). The trace of this detachment is inferred to extend at least between Burlington, Vermont,

where is forms the contact between shallow water siliciclastics and carbonates and intensely strained deeper water slates (*Hungerford-, Skeels Corners, St. Albans Slate*), and Drummondville, Québec, following the contact that juxtaposes mildly deformed shales and flysch-type lithologies (*Sainte-Sabine Formation, Bourret Formation*) with a suite of strongly deformed allochthonous slates (*Stanbridge Nappe, Granby Nappe*). In particular, the new fault discovered in the Missisquoi River gorge at Highgate Falls, that separates shaly limestones of the Highgate Formation from slates of the Morses Line Formation, probably represents the main detachment surface or a single splay of it. Furthermore, this detachment is believed to be responsible for the counter-clockwise rotation (viewed north) of the average dip of the cleavage fabric within the Morses Line Formation documented.

- The Morses Line Formation, together with the slaty portion of the Highgate Formation, represent the lithostructural equivalent of the Canadian Stanbridge Nappe. The interpreted thrust-imbricated nature of the Morses Line Formation involves the Highgate Falls Thrust as an imbrication that locally carries carbonate thrust slivers (of the Gorge Formation), as the one exposed at Highgate Falls, attached to the thrust surface.

CONCLUSIONS

"... The Taconic system, that great irritant of unacknowledged ignorance, ..."

Charles Schuchert 1937

4.1. CONCLUSIONS

- In spite of the poor outcrop in the Champlain Lowlands of northwestern Vermont, this study documents an abundance of critical - and partly unique - localities that were either not available or not discovered by previous workers, but are of crucial importance for structural constraints;
- Lithostratigraphic confusion within the Cambro-Ordovician stratigraphic nomenclature of western Vermont, caused by overlapping type-lithology definitions, in particular within the Rosenberg sequence, prevented and obscured the understanding of structural relations in the Highgate and St. Albans region;
- The Taconic foreland sequence in the study area comprises at least three large distinctive Cambro-Ordovician carbonate thrust slices (Highgate Springs-, Philipsburg-, Rosenberg slice) that characterize the *Champlain Thrust System*. The Champlain Thrust System juxtaposes mildly deformed shales and flysch-type deposits of the Taconic foreland basin to the west with intensely strained allochthonous slates of the Morses Line Formation/Stanbridge Nappe to the east;
- The significant structural break between the Rosenberg sequence (Highgate Formation) and the Morses Line Formation is interpreted as a detachment fault (*"St. Albans Detachment"*). This detachment is responsible for the absence or rarity of mélange adjacent to the Taconic thrust slice(s) western margin between near Fort Ann, New York, and Bedford, Québec, and juxtaposes the siliciclastic and carbonate-dominated suite of thrust slices of the Champlain thrust zone with the medial Ordovician Allochthon in this region. The detachment probably developped along one or more splays of

faults and was subsequently offset by a suite of northeast-striking high-angle faults.

- A distinctive set of northeast-southwest trending high-angle normal and/or tear faults cross-cuts the Rosenberg sequence in the study area. The available map pattern in southern Québec suggests a continuation of this family of faults northward up to Drummondville, Québec;
- The present stacking order and slice division of the of the Champlain Thrust System and the adjacent Taconic Allochthon presented in this study are, in part, a product of large-scale imbrication during the last stages of the collision. Thin shelf carbonate slivers, found at the base of most of the Taconic slices, are the sign-posts of this event.

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OUTCROP #	FORMATION	AGE	STAGE	FOSSIL SPECIES	AUTHOR
146, 147	Highgate Formation	Lower Ordovician		*Archaeorthis hippolyte*	Keith 1932
				Archaeorthis electra	Schuchert 1937
				Finkelnburgia sp.	
191				Syntrophina sp.	Schuchert 1937
				Archaeorthis sp.	
				Clarkella n.sp.	
				Cholopilus vermontanus	
				Petigurus cybele	
				Gignopeltis rava	
				Hystricurus mammatus	
				Pilekia eryx	
				Leostegium obtectum	
				Lloydia saffordi	
102, 103, 104				Lingulella acuminata	Raymond 1937
				Leostegium puteatum	Keith 1932
399				Symphysurina bulbosa	Landing 1983
	Gorge Formation			Symphysurina brevispicata	
				Missisquoia typicalis	
				Missisquoia depressa	
		Upper Cambrian	Trempealeauan	*Corbinia apopsis*	Landing 1983
				Saukiella serotina	
				Saukiella junia	
				Saukiella pyrene	
			Franconian	Ellipsocephaloides	
				Idahoia	
				Taenicephalus	
				Elvinia	
			Dresbachian	Dunderbergia	
237				*Bovicornellum vermontese*	Howell 1934
				Agnostus	
				Lingulella	
				Hemirhodon schucherti	
				Hemirhodon viator	
205 [pebbles]				Ucebia lata	Raymond 1937
				Stenopilus pronus	
				Illaenurus breviceps	
				Hemirhodon schucherti	
				Bynumia leptogaster	
				Marywillia triangularis	
				Coelopachys strix	
				Agostus innocens	
				Protillaenus marginatus	
				Kaninia ? platus	
				Greylockia minuta	
				Hemodyctia imitatrix	
				Acrohybus argutus	

Table A-1: Synthesis of Cambro-Ordovician biostratigraphic correlation of previous workers -> [continued on following page].

152	**Gorge Formation**	**Upper Cambrian**		Cholopilus magnus	Schuchert 1937
				Diplapatokephalus multispinosus	
				Diplozyga striata	
				Dirachopia sp.	
				Entomaspis clarki	
				Isochilina bilabiata	
				Isochilina eximia	
				Isochilina puteata	
				Isochilina tenuifilum	
				Lecanospira sp.	
				Leiostegium elongatum	
				Leiostegium obtectum	
				Leiostegium puteatum	
				Leiostrototropis phlegeri	
				Leistrototropis laevis	
				Lloydia brevis	
				Metapliomerops latidorsatus	
				Metoptogyrus grandgei	
				Perischodory grandgei	
				Plumulites mobergi	
				Protopliomerops laeviuscula	
				Shumardia granulosa	
				Strototropis elevata	
				Strototropis laeviuscula	
				Triarthroides cyclas	
				Walburgella angelini	
				Walburgella plana	
45	**Dolomite-Shale Sequence**	**Middle Cambrian**		Olenoides marcoui	Schuchert 1937
				Olenoides quadriceps	
				Bonnia	
				Ptychoparella	
				Coleoides (?)	
				Ptychoparella adamsi (?)	
228 [matrix]				Sthenotheca rugosa	Schuchert 1937
				Scenella sp.	
				Olenellus	
				Paracrepicephalus	Raymond 1937
246, 303				Mesonacis vermontana	Edson 1906
				Dactyloitides asteroides	
				Ptychoparella adamsi	
				Acrocephalites ? vulcanus	
				Microdiscus	
				Olenoides marcoui	
				Olenellus thompsoni	
				Hyolithes edsoni	
160, 311, 312	**Dunham Dolomite**	**Lower Cambrian**		Salterella pulchella	Edson 1908
				Ptychoparia adamsi	Schuchert 1937
				Bonniella desiderata	Resser 1937

Table A-1: Synthesis of Cambro-Ordovician biostratigraphic correlation of previous workers.