1980

Complex Structure and Stratigraphy of Lower Slices of the Taconic Allochthon Near Middle Granville, New York

David B. Rowley
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A thesis presented to the Faculty
of the State University of New York
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in partial fulfillment of the requirements
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Master of Science

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ABSTRACT

The Precambrian (?) to medial Ordovician rocks of the Taconic Allochthon are characterized by argillaceous and arenaceous sediments with lesser associated carbonates, carbonate breccias, and cherts of predominantly deep-water aspect. These allochthonous rocks tectonically overlie an autochthonous to parautochthonous coeval sequence of dominantly shallow marine clastics and carbonates of the Champlain and Vermont Valley sequences. Facies, thickness, sedimentologic, and paleontologic considerations suggest that these coeval sequences represent a carbonate shelf continental rise pair of the east-facing early Paleozoic Atlantic-type margin of North America. This margin formed by the opening of an ocean in latest Precambrian time. The stratigraphy of the shelf suggests that it experienced a complex transgressive-regressive history which is recorded on the rise by marked changes in type of sediment and mode of sedimentation. This Atlantic-type margin was destroyed in the medial Ordovician by eastward subduction and consequent collision beneath the Ammonoosuc volcanic arc. This resulted in the progressive east to west stacking of the rise sequence and subsequent obduction onto the shelf. Obduction involved an exceedingly complex deformation history of folding and imbrication of the shelf, Allochthon and Grenville basement.

The stratigraphy of the study area varies considerably across strike. Regions of different, though comparable stratigraphy occur in thrust bonded slices. In the west a stratigraphy closely similar to that defined by Jacobi (1977) is observed. All units, including Bomoseen, Truthville, Browns Pond, Mettawee, Hatch Hill-West Castleton, Poultney, Indian River, Mount Merino, and Pawlet are present. A central

region with a similar stratigraphy is recognized, but characterized by less carbonate, thinner and commonly more fine-grained quartzites, which among other aspects suggests that it represents a somewhat more distal (easterly) facies. To the east, the sequence is Bullfrog Hollow Lithozone, Poultney, Indian River (?), Mount Merino (?), and Pawlet. The name Bullfrog Hollow Lithozone is introduced for the basal, apparently thick sequence of purple, green and gray slates and argillites, with associated minor thin quartzites. A thin gray slate with interbedded quartzite and black calcareous quartz wacke lies within the Bullfrog Hollow and is tentatively correlated with the Browns Pond. A new name is used because direct correlations with the Truthville and Mettawee slates of western regions was not possible and other names, such as Bull, St. Catherine, or Mettawee were considered inappropriate because of misuse, poor definition, or the inclusion of untis not observed in this area. Pawlet and Poultney are usually in stratigraphic contact, but locally Indian River and/or Mount Merino are also observed. The Poultney-Pawlet contact appears to be a disconformity. Pawlet and Bullfrog Hollow are locally justaposed, but their contact is everywhere interpreted to be structural.

Structurally, the study area is quite complex. Four phases of tectonic deformation associated with at least three generations of thrust faults are recognized. Earlier, pre-tectonic, syndepositional deformation features (D_0) are also recognized. The earliest tectonic deformation (D_1) is only locally recognizable. It involves macroscopic isoclinal and initially recumbent folds (F_1) and axial surface-parallel thrusts (T_1). F_1 folds and T_1 thrusts are refolded by prominent west-verging, asymmetric, overturned folds (F_2) with an axial surface slaty cleavage (S_2). Thrusting (T_2) parallel or somewhat less steep than F_2 axial surfaces imbricates and dismembers the F_2 folds. These structures pre-date the Giddings Brook Thrust. Mesoscopic refolding of D_2 and earlier structures by F_3 folds which are associated with an axial surface crenulation cleavage (S_3) is observed, but is not macroscopically significant. A third generation of thrusts (T_3) that dip significantly less steeply east than F_2 axial surfaces are prominent in this area and may be temporally associated with F_3 folds, but this cannot be proven. T_3 thrusts may be of the same age as the Giddings Brook Thrust. Rare vertical kink bands (F_4) represent the fourth tectonic deformation and are not considered to be significant to the regional structure.

Acknowledgements

To the many people, including faculty, fellow graduate students, and friends who have contributed to the successful completion of this thesis, I wish to express my most sincere thanks. I wish to specifically thank Dr. W.S.F. Kidd for introducing me to the problems of the Taconics, for helpful discussions during the duration of the project, and for meticulously reading and correcting the text of the thesis. I would also like to thank Drs. J.F. Dewey and W.D. Means for helpful discussions and comments in the field and during the writing of this thesis. Dr. S.E. DeLong provided much needed financial support during the summer in which the field work was done, through a grant from the Petroleum Research Foundation of the American Chemical Society. Dr. W.S.F. Kidd provided financial support for the procuring of thin sections and the Department of Geological Sciences contributed to the purchasing of aerial photographs, and providing teaching assistantships over the years; this assistance is greatly appreciated. I thank Dr. Kevin Burke and Dr. P.J. Fox for financial assistance for travel to San Diego so that some of the ideas developed in this thesis could be presented.

I gratefully extend my thanks and appreciation to Wayne Freisatz and David Gallo for assistance and companionship in the field and at Furlows. I also thank Louise Delano for encouragement and for having provided the necessary stratigraphic framework without which little new would have come out of this thesis. Stimulating and often very useful discussions with J. Casey, J. Grippi, B. Idleman, R. Livaccari, D. Nelson, D. Potter, J. Rodgers, E. Rosencrantz, A.M.C. Şengör, and

E-an Zen provided food for thought and greatly helped clarify my own ideas. I also thank D. Kelly for typing sections of early drafts of this thesis.

I also wish to express my thanks and appreciation to the residents of my field area and Granville, New York for their kindness and helpfulness and for allowing me to carry out fieldwork on their property. I want to specifically thank the Aldous's for allowing me to camp on their property, for encouragement, friendly conversation, and delicious hot meals and cold beer. Ernie and Bedford Hills of H. & H. Slate Co. always took time off from their labors for friendly and useful conversations. Their generosity with their time and knowledge of slate quarrying and the history of quarrying in the area are greatly appreciated.

I owe unlimited gratitude to my parents and family for moral and financial (P.B.A.) support during the duration of my graduate career, not to mention the rest of my life.

Last but by no means least, I wish to thank Diane Paton for general moral support and for typing, and retyping this thesis.

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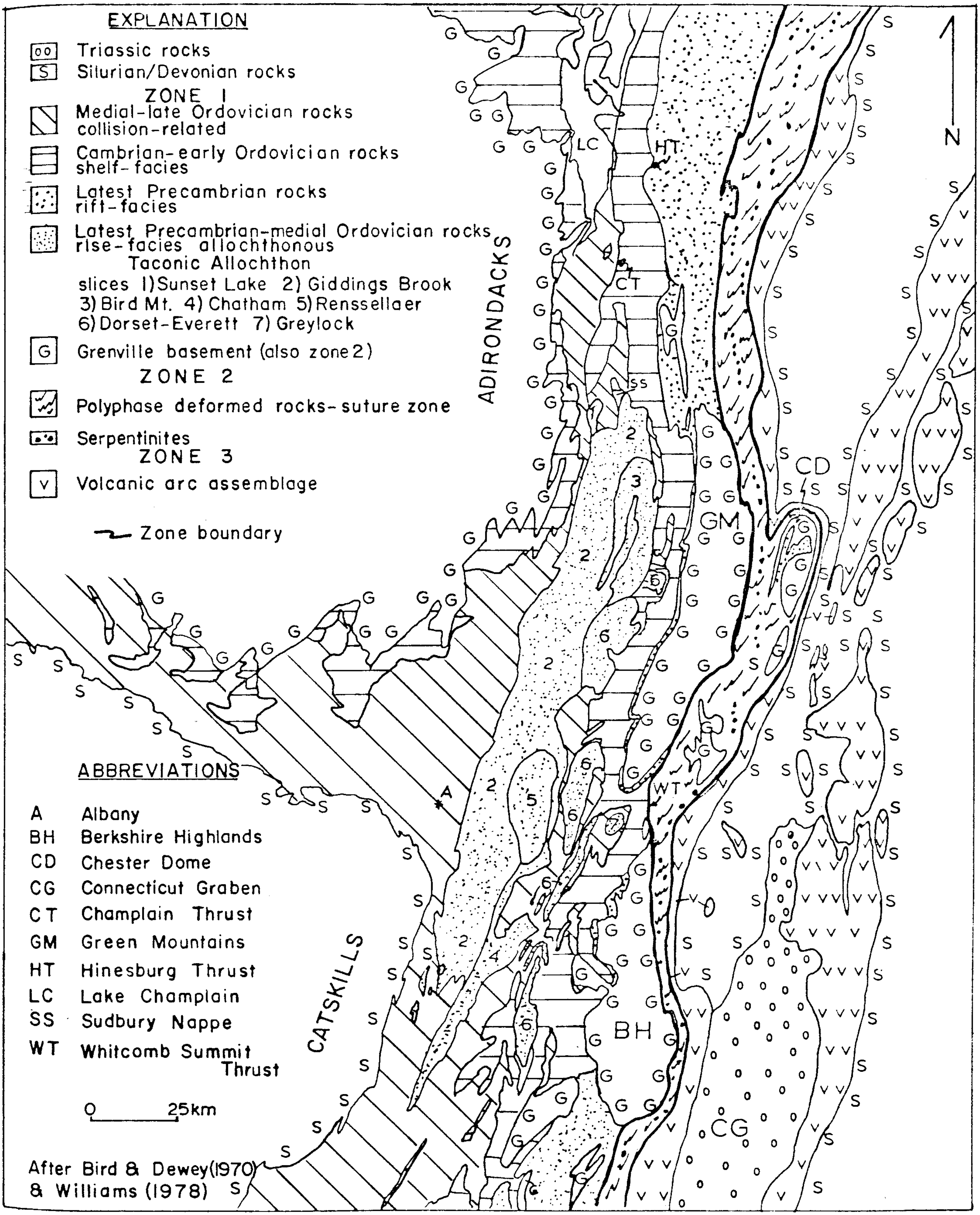
CHAPTER 1
INTRODUCTION

The Taconic Allochthon consists of predominantly deep water, argillaceous and arenaceous, with lesser calcareous and siliceous rocks of Pre-Cambrian (?) to Medial Ordovician (medial Caradocian) age. The Allochthon crops out in an elongate belt approximately 200 kilometers long from Sudbury, Vermont to Poughkeepsie, New York, and extends laterally 20 to 30 kilometers (Figures 1 and 2). The Allochthon roughly parallels the New York, Vermont, Massachusetts, Connecticut state borders. Structurally, the Allochthon consists of a series of imbricate and partially nested thrust slices with complex internal deformation. Six major thrust slices are generally recognized; they are, from structurally lowest (west) to highest (east), the Sunset Lake Slice, Giddings Brook Slice, Bird Mountain-Chatham Slice, Rensselaer Plateau Slice, Dorset Mountain-Everett Slice and Greylock Slice (Zen, 1967). Major slices are defined on the basis of differing, though comparable stratigraphies and are usually marked by topographic boundaries. Stratigraphically the lowest two slices contain the most complete stratigraphic sections (Figure 3). Taconic rocks, now predominantly slates, commonly show evidence of at least two phases of deformation and have undergone low grade (chlorite to biotite) regional metamorphism. In general, deformation and metamorphism increases from west to east within the Allochthon.

The Taconic Allochthon tectonically overlies and is surrounded by an authochthonous to parautochthonous, coeval sequence of dominantly shallow marine clastics and carbonates of the Champlain and Vermont

Figure 1

Geological map of western New England and eastern New York.


EXPLANATION
Triassic rocks
Silurian/Devonian rocks
ZONE 1
Medial-late Ordovician rocks collision-related
Cambrian-early Ordovician rocks shelf-facies
Latest Precambrian rocks rift-facies
Latest Precambrian-medial Ordovician rocks rise-facies allochthonous
Taconic Allochthon
slices 1) Sunset Lake 2) Giddings Brook 3) Bird Mt. 4) Chatham 5) Renssellaer 6) Dorset-Everett 7) Greylock
Grenville basement (also zone 2)
ZONE 2
Polyphase deformed rocks-suture zone
Serpentinites
ZONE 3
Volcanic arc assemblage
Zone boundary
ABBREVIATIONS
A Albany
BH Berkshire Highlands
CD Chester Dome
CG Connecticut Graben
CT Champlain Thrust
GM Green Mountains
HT Hinesburg Thrust
LC Lake Champlain
SS Sudbury Nappe
WT Whitcomb Summit Thrust
0 25km
After Bird & Dewey (1970) & Williams (1978)
ADIRONDACKS
CATSKILLS
N

Valley Sequences (Figure 1) (Shumaker, 1967; Zen's Synclinorium Sequence, 1967). Facies, thickness, sedimentologic and paleontologic considerations suggest that the coeval clastic-carbonate and argillite-clastic sequences represent a carbonate shelf-continental rise pair of the east-facing, early Paleozoic, Atlantic-type North American continental margin (Bird and Dewey, 1970; Rodgers, 1968, 1970). Emplacement of the Taconic sequence onto the coeval shelf occurred during the medial Ordovician (Caradocian) Taconic Orogeny. Plate tectonic corollary modelling suggests that emplacement of the Allochthon probably resulted from attempted subduction of the Atlantic-type continental margin in an east-dipping subduction zone (Chapple, 1973, 1979; Rowley and Delano, 1979; and Rowley, Kidd and Delano, 1979).

Location

The study area lies approximately 100 km NNE of Albany and is located athwart the New York-Vermont state line in northern Washington (New York) and Rutland (Vermont) counties (Figure 2). The field area includes portions of the Granville, Wells, and Poultney 7-1/2 minute quadrangles. The area is roughly triangular, covering about 45 square kilometers. It is bounded on the west by New York Route 22A, Lake St. Catharine on the east, and extends approximately 9 km south from the northern boundary of the Poultney River. The exposure in the area is good, at least for the Taconics, although unfortunately not good enough for detailed mapping in some key areas.

Geomorphologically, the area consists of well glaciated low rolling hills and open valleys with a dominant north-south grain. Glacial deposits are ubiquitous, mostly consisting of unsorted to poorly sorted till and gravel, possibly of kame or kame terrace origin. Glacial

Figure 2a

Location map of study area.

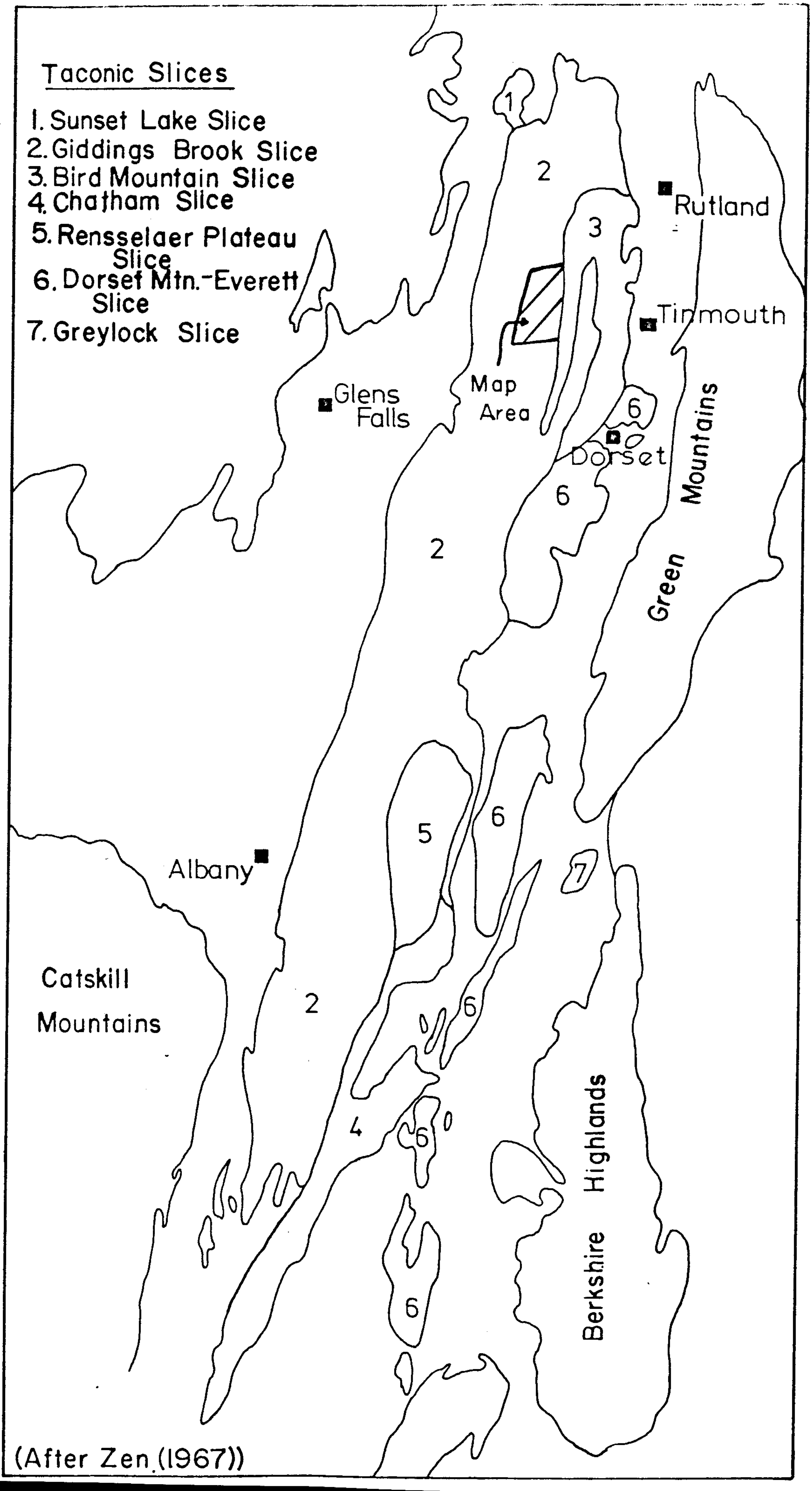
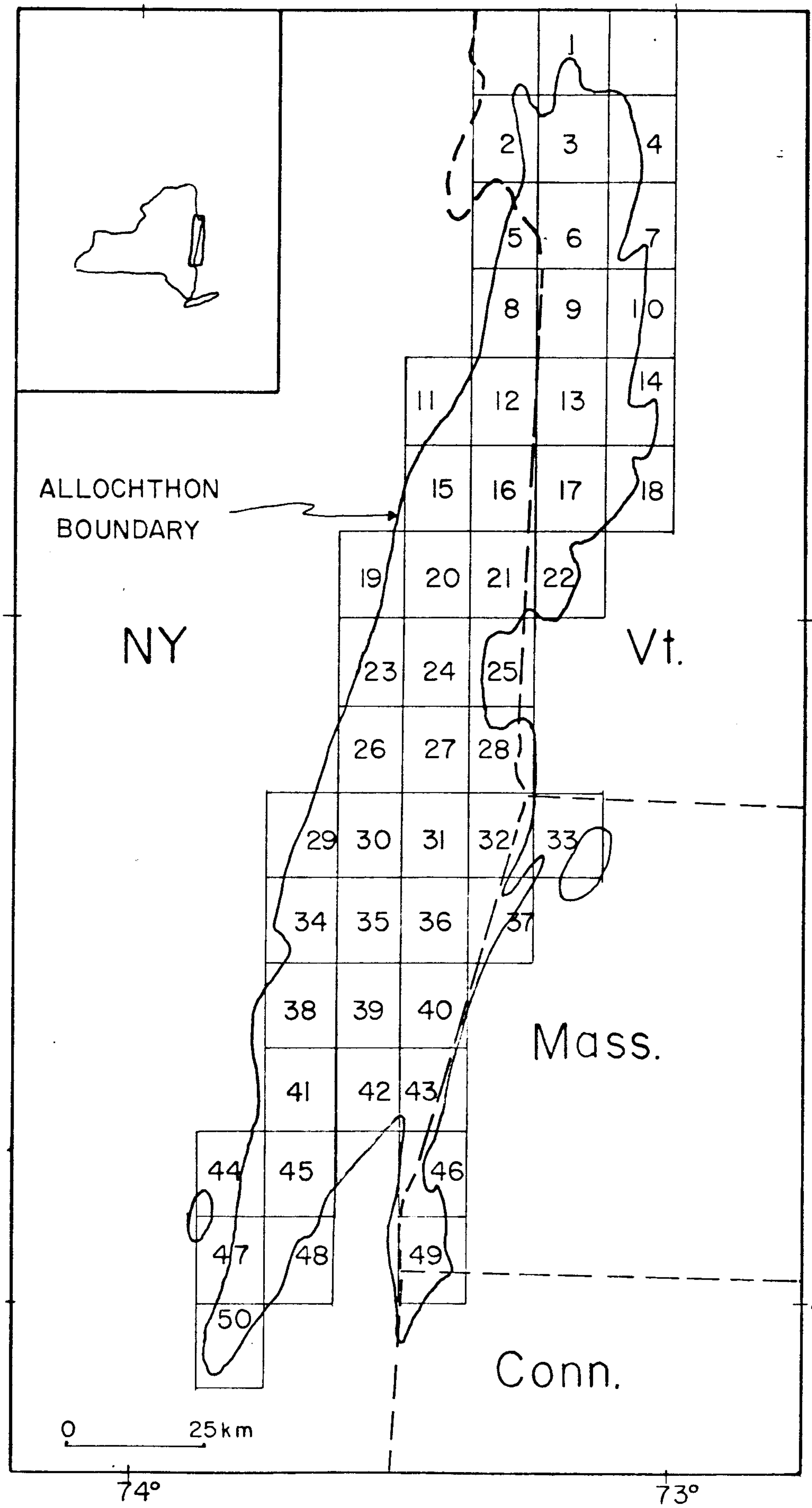

Taconic Slices
1. Sunset Lake Slice
2. Giddings Brook Slice
3. Bird Mountain Slice
4. Chatham Slice
5. Rensselaer Plateau Slice
6. Dorset Mtn.-Everett Slice
7. Greylock Slice
Rutland
Tinmouth
Glens Falls
Map Area
Dorset
Green Mountains
Albany
Catskill Mountains
Berkshire Highlands
(After Zen (1967))

Figure 2b

The geographic setting of the Taconic Allochthon. The relevant states and counties are outlined and labeled, and the quadrangles are numbered. An identifying list follows:

1. Sudbury
2. Benson
3. Bomoseen
4. Proctor
5. Thorn Hill
6. Poultney
7. West Rutland
8. Granville
9. Wells
10. Middletown Springs
11. Hartford
12. West Pawlet
13. Pawlet
14. Dorset
15. Cossayuna
16. Salem
17. Equinox, N.W.
18. Equinox, N.E.
19. Schuylerville
20. Cambridge
21. Shushan
22. Equinox, S.W.
23. Schaghticoke
24. Eagle Bridge
25. Hoosick Falls
26. Tomhannock
27. Grafton
28. North Pownall
29. Troy South
30. Averill Park
31. Taborton
32. Berlin
33. Williamstown
34. East Greenbush
35. Nassau
36. Stephentown Center
37. Hancock
38. Kinderhook
39. East Chatham
40. Caanan
41. Stottville
42. Chatham
43. State Line
44. Hudson South
45. Claverack
46. Egremont
47. Clermont
48. Ancram
49. Bashbish Falls
50. Rock City

(modified from Zen, 1967)



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38
39
40
41
42
43
44
45
46
47
48
49
50
ALLOCHTHON BOUNDARY
NY
Vt.
Mass.
Conn.
43°
42°
74°
73°
0
25 km

Figure 3

Stratigraphies of the different Taconic Allochthon slices. The most complete stratigraphies are recognized in the structurally lowest slices.

		FIRST		STACKING SEQUENCE			LAST (after Rowley et al)
		(Mutual Relations Uncertain)		(Mutual Relations Uncertain)			
		Dorset Mtn. Slice and Greylock Slice	Rensselaer Plateau Slice	Chatham Slice	Bird Mountain Slice	Giddings Brook Slice	Sunset Lake Slice
ORDOVICIAN	MIDDLE					Pawlet	Pawlet ?
					Indian River	Indian River	?
	LOWER			?	Poultney	Poultney	Poultney
CAMBRIAN	UPPER			?	Hatch Hill?	Hatch Hill	?
	MIDDLE			?	?	rocks mapped as W. Castleton	?
	LOWER			West Castleton	West Castleton	West Castleton	West Castleton
		Greylock	Mettawee	Bull	Bull	Bull	Bull
PRECAMBRIAN ?		Bellowspipe	Rensselaer	Rensselaer	Biddie Knob	Biddie Knob	
		upper part of Berkshire					

(after Zen, 1967)

cover and alluvium only constitute a serious impediment to bed rock mapping adjacent to Lake St. Catharine and in the valley to the east of Route 22A.

Detailed outcrop mapping at a scale of 1:12,000 was carried out on photographically enlarged 7-1/2 minute (1:24,000) topographic quadrangle sheets. A set of aerial photographs (1974, scale 1:20,000) were used to improve station locations in Vermont and for more complete location of active and abandoned slate quarries. State, county, local, quarry access and logging roads provided excellent access to all portions of the field area.

Mapping was carried out over a period of nearly 18 weeks from June to mid-November, 1978. Mapping mainly involved east-west traverses during which every outcrop was described and located.

Rationale and Findings

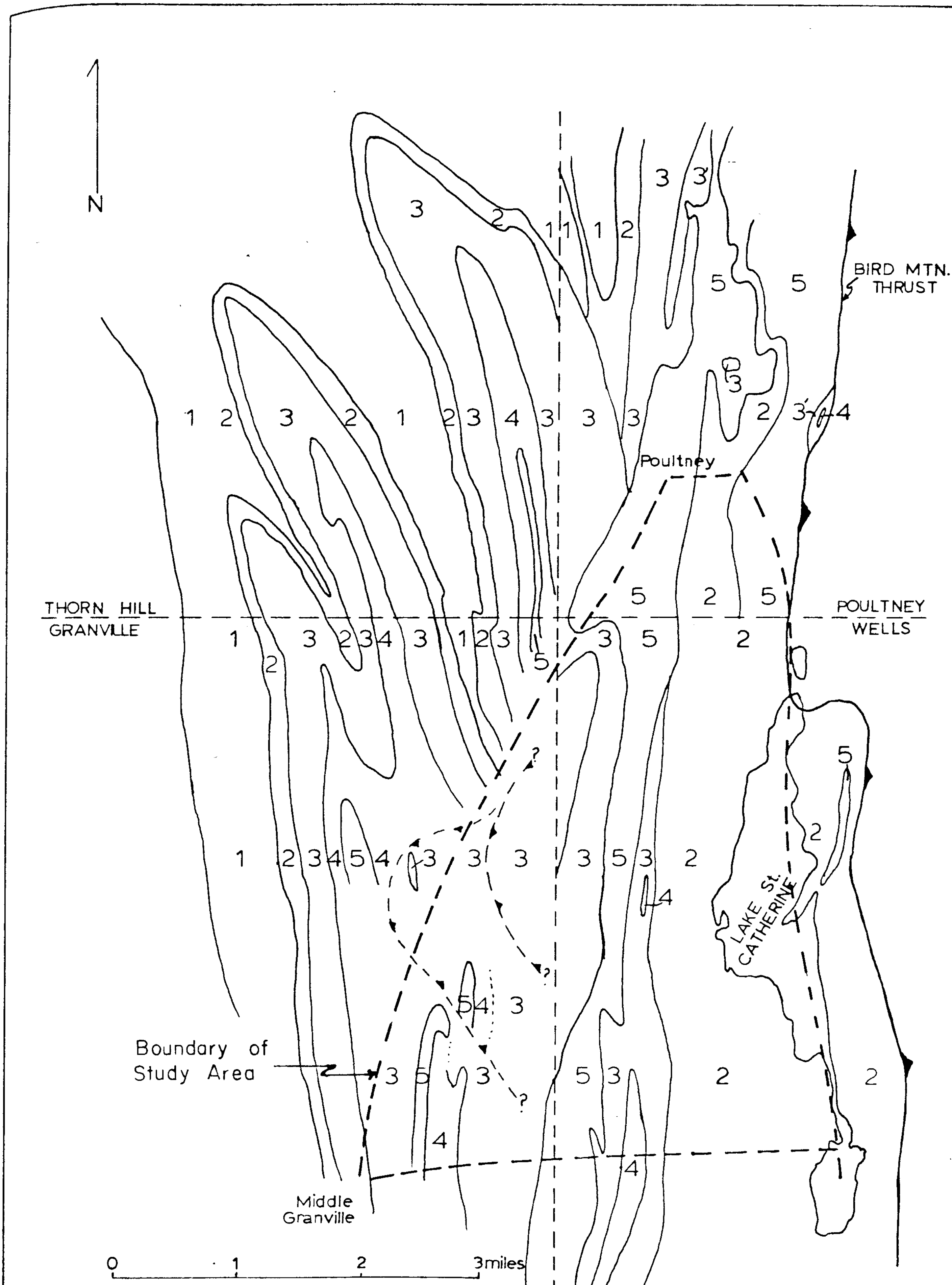
Compilation of geologic maps from eastern Whitehall (Jacobi, 1977; Theokritoff, 1963, Castleton (Zen 1964), and Pawlet (Shumaker, 1967) quadrangles suggest that a marked change in stratigraphy occurs near the New York-Vermont state line in this area (Figure 4). Detailed mapping by Jacobi (1977) demonstrated an apparently conformable lithostratigraphic sequence from basal Bomoseen to Pawlet at the top in the western part of the 'Giddings Brook Slice'. Adjacent to the east many prominent units such as the Bomoseen, Hatch Hill-West Castleton, Indian River and Mount Merino are thin or absent on Zen and Shumaker's maps. Detailed outcrop mapping was undertaken in order to ascertain the nature of this change. Two specific problems were outlined prior to the initiation of mapping. One, to study the nature of the contact at the base

Figure 4

Geological map of study area and vicinity based on pre-existing maps.

Thorn Hill and Granville quadrangles after Jacobi (1977) and Theokritoff (1964), Poultney quadrangle after Zen (1961, 1964) and Wells quadrangle after Shumaker (1967).

Units	Thorn Hill & Granville	Poultney	Wells
5	Pawlet	Pawlet	Pawlet
4	Indian River & Mount Merino	Indian River	Indian River
3	Poultney, Hatch Hill, West Castleton	Poultney West Castleton	Poultney
2	Mettawee Browns Pond	Bull Formation	St. Catherine
1	Truthville Bomoseen	Bomoseen Mbr.	


N
BIRD MTN. THRUST
THORN HILL
GRANVILLE
POULTNEY WELLS
Poultney
LAKE St. CATHERINE
Boundary of Study Area
Middle Granville
0 1 2 3miles

of the Pawlet, specifically its relationship with underlying Taconic units. Two, to define the lateral stratigraphic and structural variations across the eastern part of the 'Giddings Brook slice'.

Zen (1961,1964) and Shumaker (1967) in the course of quadrangle mapping proposed the presence of an angular unconformity at the base of the Pawlet in the Castleton and Pawlet 15 minute quadrangles. Potter (1972) and Wright (1970) among others supported this proposal. Detailed mapping by Jacobi (1977) in the western part of the Giddings Brook slice indicated that the Pawlet Formation was everywhere conformable with the underlying Mount Merino Formation. It was hoped that detailed remapping in areas where the unconformity was originally identified would shed light on this apparent discrepancy. Mapping by the author indicates the existence of an apparent disconformity below the Pawlet, as it is usually in contact with Poultney and less commonly Mount Merino-like black slates.

Mesoscopic and macroscopic relations at the Pawlet-Poultney, and less commonly Pawlet-Mount Merino-like black slate contact, do not suggest any angularity at this contact in this ares. The common absence of Mount Merino and Indian River does apparently necessitate a disconformity.

Lithostratigraphic units have been defined and are in part directly correlative with lithostratigraphic units defined by Jacobi (1977) and others (including Zen, 1961; 1967; Theokritoff, 1963; Shumaker, 1967; Potter, 1972, etc.) in this and surrounding areas. The area is both stratigraphically attenuated and structurally more complex than the area adjacent to the west; mapped by Jacobi (1977). Without the work of Jacobi, particularly with respect to stratigraphy, mapping in the study area would have been considerably less fruitful.

Previous quadrangle style mapping in contiguous segments of this area by Dale (1899) Zen (1961, 1964), Theokritoff (1964) and Shumaker (1967) was primarily stratigraphically and paleontologically oriented and carried out under the assumption of stratigraphic and structural continuity unless otherwise required (Zen, personal communication, 1978). Their mapping has been significantly improved upon. The mapping by the author and Jacobi (1977) indicates that large scale mapping (scale 1:12,000 or better), outcrop by outcrop, is required before we can ever hope to unravel the complex stratigraphy and structure of the Taconics.

Structurally the area is quite complex. Four phases of deformation and three generations of thrust faults are recognized in this area. The earliest deformation is only mappable where downward-facing folds are observed, and involves macroscopic, probably initially isoclinal and recumbent folds (F_1) and possible axial surface-parallel thrusting (T_1). These folds and thrusts are refolded by prominent west-verging asymmetric, overturned folds (F_2) with an axial surface parallel slaty cleavage (S_2). This cleavage and generation of folds are the most prominent fold-related features in this area. Thrust faults (T_2) parallel to F_2 axial surfaces, imbricate and dismember the F_2 folds. Minor coaxial refolding (F_3) of F_2 folds and earlier structural elements, is observed, but this generation of structures gives rise to only mesoscopically observable redistribution of earlier structures. A third generation of thrusts (T_3), which dips less steeply east than F_2 axial surfaces is recognized. This generation may be associated with D_3, but this cannot be demonstrated. The final phase of deformation involves north-west striking vertical kink bands, which are not considered to be significant mesoscopic or macroscopic structural elements.

The major structures are believed to have resulted from the tectonic emplacement of the Allochthon, in the medial Ordovician. These structures, particularly when associated with the structural elements in the western part of the Giddings Brook slice indicate that the low Taconic, Giddings Brook slice was emplaced as a non-soupy, essentially lithified nappe. The structures in the Giddings Brook slice are completely compatible with structures mapped in other fold thrust belts, (Rowley and Delano, 1979).

Succeeding chapters focus on stratigraphy, stratigraphic variations in the Giddings Brook slice, depositional environment, petrography, structure, and depositional and tectonic history of eastern New York and western New England, during the period prior to the end of the Ordovician. This final chapter will attempt to place this history within the context of continental margin evolution and destruction, referred to as the Wilson Cycle by Dewey and Burke (1974).

CHAPTER 2

STRATIGRAPHY

Introduction

The development of stratigraphic thought in the Taconics involves many contributors and almost one hundred and fifty years of investigation. Early workers, such as Dewey and Eaton, Emmons, Walcott, Dana and others (see Merrill, 1924) grouped both carbonates and argillites together and considered them to be a stratigraphically continuous sequence. Emmons in 1842 called this succession the Taconic System. Walcott's paleontologic work (1888) indicated that the carbonates and argillites were at least in part time equivalent sequences of Cambrian and Ordovician age. The work of Dale (1899, 1904) provided the first comprehensive description of the slate belt, and further documented the existence of a Pre-Cambrian (?) to Ordovician sequence within it. Dale (1899, 1904) defined a detailed lithostratigraphic section (Figure 5) which extended from Dale's units A and B (sub-Rensselaer Grit)shales to the Hudson Grit (Pawlet/Austin Glen Greywackes). This lithostratigraphic sequence served as the framework for all later stratigraphic discussions. Recent work (Jacobi, 1977) demonstrated the basic validity of Dale's stratigraphy and indicates the need for only slight modification. The most important modification of Dale's work involves the reinterpretation of Dale's Berkshire Schist (see Zen, 1964b for a discussion of this problem) and the introduction of stratigraphic names in place of the letters and cryptic names used by Dale.

Ruedemann (in Cushing and Ruedemann, 1914) began the process of introducing names and refining Dale's stratigraphy. Ruedemann was the first to suggest and partially document that the slate belt was allochthonous (1909). Dale (1899), aware of the problem of having two different, and yet coeval sequences juxtaposed, had considered the slates to be in situ facies equivalents of the carbonates (see Voight, 1972 for a discussion of Dale's post 1909 work). The suggestion by Ruedemann that the Taconic slates were allochthonous initiated a famous controversy among Taconic geologists, a controversy that was characterized by a distinct polarity between northern and southern Taconic workers (Zen, 1967). The debate between 'klippenists' and 'non-klippenists' raged for nearly half a century. The history and arguments of this debate and the polarization of workers are well summarized by Zen (1967), Metz (1969) and Jacobi (1977). Zen (1961) following and expanding earlier suggestions by Cady (1945) and Rodgers (1951) suggested that the Taconics are allochthonous and were emplaced as soft-sediment submarine gravity slides. Zen's important contribution was in his discussion of the nature of the basal contact that might result from such an emplacement history (the so called 'black slate problem'); in particular, the characteristic lack of a distinct 'thrust contact'. Almost all subsequent workers have accepted Zen's submarine, soft-sediment slide hypothesis. The emplacement history of the allochthon as viewed by the author has been discussed elsewhere (Rowley et al, 1979) and in a later section (see Chapter 6).

The post-Dale history of Taconic stratigraphic thought is characterized by a refinement of the sequence, addition of age constraints and, in particular, the introduction, definition, revision, and renaming of units. This later practice, in part resulting from widely separate map areas, but also due to structural complexities, stratigraphic variations, variable but seldom good exposure, and confusion between litho-, bio-, and chrono-stratigraphic designations, gave rise to a plethora of stratigraphic names, definitions and synonymies. Zen (1964b) compiled most Taconic stratigraphic names into an invaluable reference.

Several stratigraphic sections from the Giddings Brook slice are compiled in Figure 5. It is readily apparent that there is considerable variation in stratigraphic sequence within the Taconics, even within very short distances (less than one to several kilometers). Potter (1972, p. 5) states:

> "In general, stratigraphic units within the Taconic Sequence show great lateral continuity north and south, and maximum change in thickness and lithologic character across strike from east to west. . . .The east-west variations. . . .are accentuated by the telescoping of the facies within the allochthon."

When adjacent areas are compared, such as that described by Jacobi (1977) and in this thesis, a striking lateral (across strike) variation is obvious. This type of lateral variation is extremely important and has been little discussed, especially in any detail.

The nature of the lateral variation, particularly with respect to presence or absence of units, thickness, sedimentologic variations (grain size, bed thickness, carbonate content, etc.) are important for palinspastic reconstruction and will be discussed in some detail later. An important step in Taconic geology and stratigraphy will be the delineation of regions of markedly different stratigraphic sequences, beyond that already done by Zen (1967) to define his slices. Such differences may be helpful in subdividing slices, particularly the Giddings Brook slice, into new slices or subslices.

In the following section, lithostratigraphic units are described,

Figure 5

Compilation of stratigraphies from the 'Giddings Brook slice' (after Jacobi, 1977).

	Dale, 1899	Potter, 1972	Theokritoff, 1964	Zen, 1967	Shumaker, 1967	Jacobi, 1977	Rowley, 1980 Eastern Region
ORDOVICIAN	HUDSON Grit Red and Green Thin Quartzites White Beds CALCIFEROUS	NORMANSKILL Austin Glen Mt. Merino Indian River POULTNEY Owl Kill White Creek	PAWLET INDIAN RIVER POULTNEY C B A	NORMANSKILL Pawlet POULTNEY	PAWLET INDIAN RIVER POULTNEY	PAWLET MT. MERINO INDIAN RIVER POULTNEY Crossroad Dunbar	PAWLET MT. MERINO INDIAN RIVER POULTNEY
CAMBRIAN	FERUGINOUS BLACK SL. & LS. BLACK PATCH GT.	HATCH HILL Eagle Bridge WEST CASTLETON NASSAU Mudd Pond Zion Hill Mettawee'C	HATCH HILL WEST CASTLETON BULL Mettawee	HATCH HILL WEST CASTLETON BULL Mettawee	HATCH HILL WEST CASTLETON ST. CATHERINE Castleton Cg.	HATCH HILL WEST CASTLETON METTAWEE	BULLFROG HOLLOW Lithozone
PRECAMBRIAN (?)	OLIVE GRIT	Bomoseen Mettawee 'B' Rensselaer Mettawee 'A'	Bomoseen	Bomoseen Rensselaer Biddie Knob	Zion Hill	TRUTHVILLE BOMOSEEN	after Jacobi(1977)

All capitals are Formations. Lower Case are members nomenclatural problems discussed, and correlations suggested. Lateral stratigraphic variations, as evident from a traverse across the Giddings Brook slice (integrating Jacobi (1977)and my observations) and possible mechanisms responsible for this variation are described and assessed in Chapter 3.

Stratigraphic units are described in ascending order; however, because of the nature of lateral variations and structural complexity the order is not always clear in this map area. As a result, the author has relied on the stratigraphic sequence determined by Jacobi (1977) and others in less complex regions. Some units are grouped together because they cannot be distinguished in the field. This problem is very reminiscent of stratigraphic discussions of Zen, Shumaker, Potter and others, in contrast to the work of Jacobi (1977). This problem results from the presence of an apparently thick succession of purple, green, and minor gray argillites, lacking commonly observed 'marker' horizons or units at its boundaries, or internally. The work of Jacobi (1977) suggests that the names of Bull Formation, Nassau Formation or other presently used stratigraphic names for the lower succession in the Giddings Brook slice may not be appropriate. I hesitate to use names employed by Jacobi (1977) unless correlation is clearly indicated. As a result, I have introduced a new name, Bullfrog Hollow lithozone (See Hedberg, 1976).

Bomoseen

Along the western edge of the Allochthon the lowest lithostratigraphic unit is a massive, hard, olive gray-green micaceous wacke referred to as the Bomoseen Formation (Jacobi, 1977; Bomoseen Greywacke of Zen, 1961). In this area, lithologically correlative rocks constitute an unfortunately small volume and are poorly exposed. The only outcrops assigned to the Bomoseen lie between the two strings of quarries northwest of South Poultney (Figure 6). Here the Bomoseen is an olive gray-green, silty, mica spangled slate, associated with hard, thin, green, mica spangled wacke, and minor, often highly slump disrupted thin, (1 to 10 cm) commonly lensoid and discontinuous quartzites). The presence of some hard wacke warrants the correlation with the Bomoseen to the west.

There is only a limited thickness (maximum of 20 meters) in this area and the lower contact is faulted.

Dale (1899) called this unit the Olive Grit. Ruedemann (in Cushing and Ruedemann, 1914) renamed this lithology the Bomoseen Grit for exposures near Lake Bomoseen described by Dale. Most recent workers, notably Zen (1961, 1964a & b), Bird (1962), Potter (1972) Theokritoff (1964) have considered the Bomoseen as a member of the Bull Formation or its equivalents (Nassau Formation, St. Catharine Formation). Jacobi (1977) mapped the Bomoseen as a separate formation. She found the Bomoseen to be a prominent lithology, readily recognizable, that everywhere mapped out in the same stratigraphic position. Furthermore, the only other lithology in her area that has anything in common with it is the overlying Truthville slate, both of which are mica spangled. Other

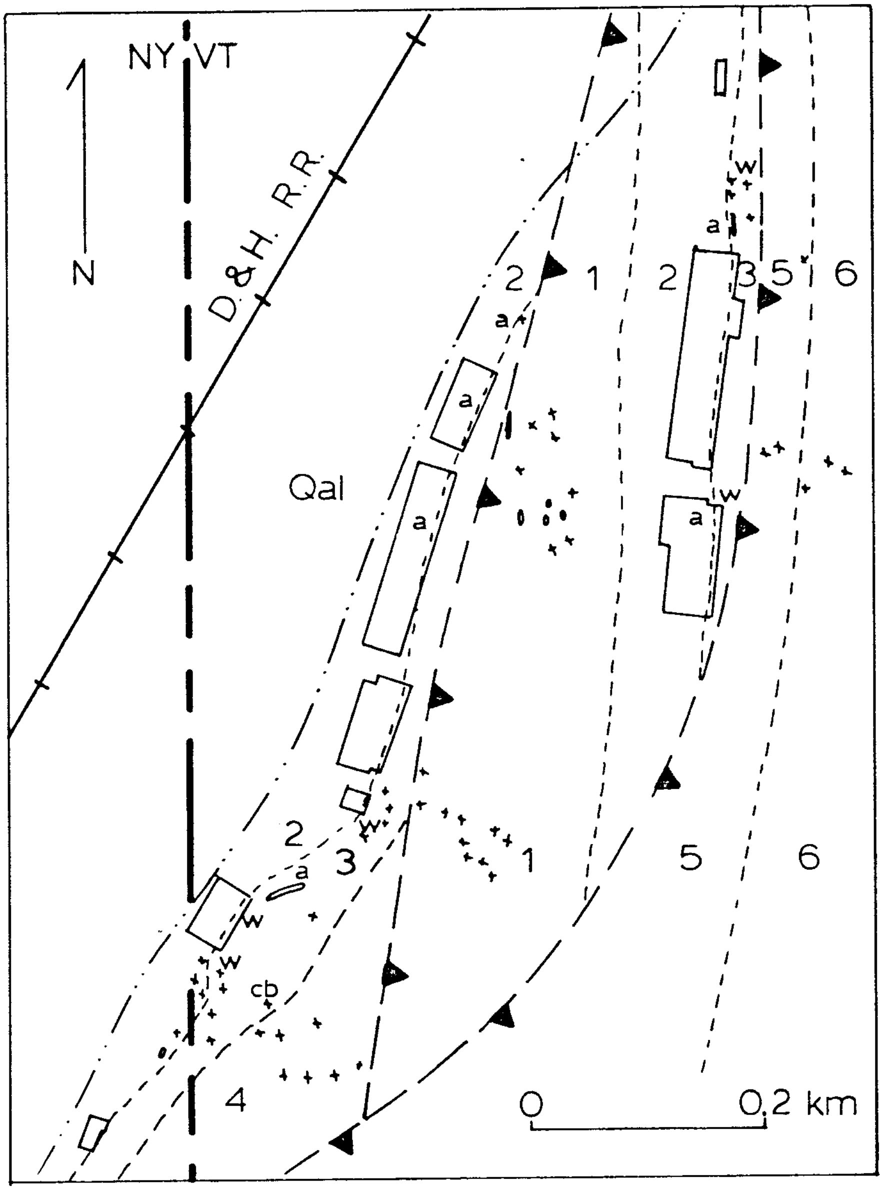


Figure 6

Location map showing outcrops of Bomoseen wackes and slates northwest of South Poultney. a-arenite, cb-carbonate breccia, w-Eddy Hill Grit. Units 1-Bomoseen, 2-Truthville, 3-Browns Pond, 4-Mettawee, 5-West Castleton and Hatch Hill, 6-Poultney.

lithologies included within the Bull or its equivalent units, including the Browns Pond, and Mettawee slate of Jacobi (1977) lack this detrital mica component, and are never found in any other position in her area, suggesting that previous inclusion within a large Bull Formation was unnecessary. This distinction is maintained here.

Bomoseen wackes and associated lithologies everywhere underlie the lowest fossil-bearing strata. As shown by Jacobi (1977) the transition from Bomoseen to the fossil-bearing Browns Pond limestones and slates is gradational and conformable. These rocks are therefore considered to be Pre-Cambrian (?) in age.

Truthville

The name Truthville slate was informally introduced by Jacobi (1977) to refer to olive weathering, gray-green, soft, silty, well-cleaved, mica-spangled slates, lying conformably on Bomoseen wackes. Slates typical of the Truthville primarily crop out to the east of South Poultney; however, the lack of Bomoseen wackes and only poorly developed possible Browns Pond lithologies makes direct correlation impossible. Where Bomoseen and Browns Pond are recognized, slates equivalent to Truthville are mapped as such. However, most of these slates are not identical to 'type Truthville', and are commonly more similar to Mettawee Slate, as used by Jacobi (1977), and Mettawee Slate *sensu stricto* of Rowley et al (1979).

The stratigraphically best-constrained Truthville-equivalent slate outcrops occur along the two short strings of quarries to the northwest of South Poultney (Figure 6). This correlation is based on the following observations:

(1) Browns Pond lithologies occur above and to the east of both strings of quarries. Where the Browns Pond is best developed, e.g., to the east of the western string of quarries, internal stratigraphic relations suggest that the Browns Ponds is right side up and east-facing. The contact between the Browns Pond lithologies and quarried slates is sharp and appears to be sedimentary.

(2) Bomoseen lithologies crop out between the two strings of quarries. A stratigraphic sequence including Bomoseen, Truthville, and Browns Pond is interpreted for the eastern string of quarries.

(3) Both strings of quarries contain mesoscopically identical slates and associated lithologies.

These Truthville-equivalent slates include green, gray-green, and gray, well-cleaved, in places finely silty, slightly calcareous slates that are not remarkably mica spangled. Quartzites are interlayered with these slates close to the Truthville-Browns Pond contact. One or two, lensing, thick (up to 1 m) and fine to medium grained, variably calcareous and pyritiferous, sometimes vitreous, apparently massive quartzite often occurs within 2 to 4 m of the upper contact. Commonly associated with the thicker quartzites are thin (1 to 15 cm) extension-gashed, fine to medium grained, quartzites. These thinner quartzites are also unlaminated and may be calcareous and rotten weathering.

These slates presumably are conformably overlying Bomoseen wackes, but the contact was never observed, so this must remain uncertain. Truthville slates in these exposures have an outcrop width-determined thickness of at least 30 meters. Jacobi (1977) reported thicknesses in the range of 0 to 30 m.

A series of roadcuts and associated outcrops along Fox Road (Figure 7) just west of the New York-Vermont state border are also mapped as Truthville. The correlation is based on gross lithologic similarity and the association with probably Browns Pond lithologies to the east. The slates along Fox Road are different from those just described. Here the slates and associated argillites are olive to tannish weathering, green, gray-green, and gray, slightly to moderately silty, moderately to well cleaved, and they commonly have a brownish stain on cleavage surfaces. Broken surfaces often show a prominent detrital white mica component. Some silty, moderately hard, moderately to poorly cleaved argillites are also present. Locally interlayered with these slates and argillites are 5 to 15 cm thick quartzites. These quartzites are gray to slightly

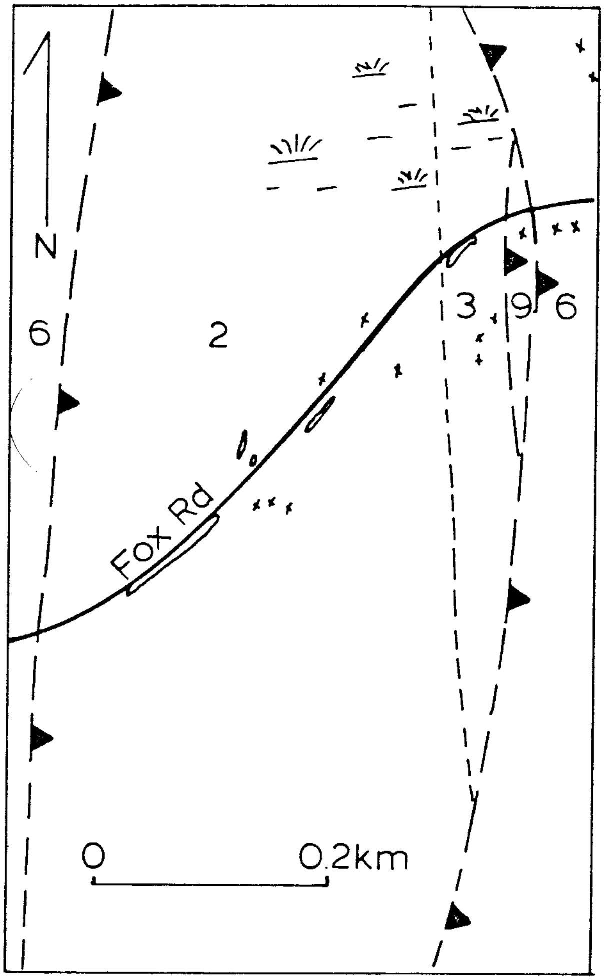


Figure 7

Location map of Truthville outcrops along Fox Road. Units 2-Truthville, 3-Browns Pond, 6-Poultney, 9-Pawlet.

brownish weathering, and on fresh surfaces they are light gray and may be speckled with iron stains. They are commonly slightly calcareous, fine to medium grained, with subangular quartz grains, and appear massive. White micas may be found on the surfaces of some of the quartzites.

Jacobi (1977) introduced the name Truthville Formation. Other workers presumably included these slates with either the Bomoseen member or "Metawee slate facies" of the Bull or equivalent formations. Shumaker (1967) mapped the quarries to the northwest of South Poultney as St. Catharine Formation. The name Truthville slate is applied here only to those slates that are demonstrably equivalent to type Truthville slate. Other slates, even lithologically identical slates are mapped as Bullfrog Hollow lithozone.

Truthville slates directly underlie the lowest fossil-bearing strata in Jacobi's area and are therefore considered to be Pre-Cambrian (?) in age.

Bullfrog Hollow Lithozone

The name Bullfrog Hollow Lithozone (see Hedberg, 1976, for a discussion of lithozones) is introduced here to describe the extensive exposures of purple, green, and gray argillaceous rocks of this area, which cannot be demonstrated to be equivalent to other lithostratigraphic units because of the lack or only marginal development of key horizons, such as the Bomoseen, Browns Pond, and Hatch Hill-West Castleton formations. Some of these argillit es are extensively quarried.

The Bullfrog Hollow lithozone is divided into two lithofacies based on predominant coarseness of the argillites and associated lithologies. Lithofacies-A, probably, in general, lower in the unit, consists of purple, green and minor gray, silty argillite to mudstone that weathers buff to purple or green. The argillites are moderately to coarsely cleaved, less commonly well cleaved, and commonly have fine to silt sized mica on broken surfaces. Locally, interlayered with the argillites, are thin (0.5 to 3.0 cm) silty to fine sandy, greenish to white weathering quartzites. Most of the thin quartzites appear massive, although rarely they are parallel and cross-laminated. Mica may be present within the quartzites, and sometimes is prominent on bedding surfaces. Fine to medium grained, 0.3 to 2.0 m thick, lensing quartzites are sometimes present. These thicker quartzites tend to weather white or gray, are greenish on fresh surfaces and may be slightly calcareous. Rare rust speckles may also be present on fresh surfaces, but most are clean and vitreous. Thin silty quartzites are commonly associated with these thicker quartzites.

Where purple argillites are absent, and thin quartzites are abundant in green or locally gray slates it is virtually impossible to distinguish lithofacies-A from Poultney (see Poultney discussion). In all cases where this problem arose Occam's razor was applied and the interpretation requiring the simplest structural consequences was chosen; although this is not always a good assumption, as demonstrated elsewhere in this area.

Lithofacies-B of the Bullfrog Hollow Lithozone, consists of purple, green, and gray, with minor black, red, and maroon often porcellanous, variably fissile, well cleaved slates. These slates are extensively quarried, including all commercial quarries to the east of the South Poultney Thrust, and some quarries slightly to the west (south-southwest of South Poultney) (Figure 2b). These slates weather buff to brownish, or locally greenish or purple, depending on their carbonate content (Dale, 1899). They are often extremely well cleaved. Bedding is locally observed as thin (0.5 to 2.5 cm) brownish weathering, green micritic to calcisiltitic limestones, best observed in quarries. Local quarry operators sometimes refer to these limestones as "rubber beds" because of their resistance to blasting. Rarely, thin (less than 3.0 cm) lensing, extremely rotten weathering sandstones, with subrounded to rounded grains, were observed associated with finely color laminated green, gray and purple slates. Color laminations may or may not reflect bedding (Dale, 1899). Reduction spots of green within fields of purple slate are sometimes observed, and, locally, good ellipses are seen on cleavage surfaces.

The two lithofacies are interlayered, sometimes on a mesoscopic scale, and do not appear to represent mappable units. Comparisons with other areas, particularly the area mapped by Potter (1972) suggest that lithofacies-A probably is more characteristic of the lower part of the Bullfrog Hollow Lithozone than lithofacies-B. In this sense lithofacies-B probably corresponds with the Mettawee Slate of Jacobi (1977) and Ruedemann (in Cushing and Ruedemann, 1914). Dale (1899) mapped all of the quarries within the Vermont part of this area as Cambrian Roofing Slates, which supports the tentative correlation. However, as noted in a previous section describing the Truthville slate to the northwest of South Poultney this correlation is not certain.

Locally within this formation a thin (3 to 20 m) sequence of arenaceous rocks is found, interlayered with primarily medium gray argillites, and believed to be correlative with the Browns Pond of Jacobi (1977). These lithologies are described below. The exact position of this thin sequence is indeterminable within the monotonous sequence of purple, green and gray slates of the Bullfrog Hollow Lithozone. A similar thin sequence was reported by Wright (1970) to the north near Lake Bomoseen. This sequence does not appear to separate distinguishable slates, or lithofacies of the Bullfrog Hollow Lithozone. Quarried purple and green slates occur both to the east and west of this sequence, although in a general way, the slates to the east of these Browns Pond (?) rocks are siltier and purple is more common. Gray is more common to the west of the Browns Pond (?).

The thickness of the Bullfrog Hollow Lithozone is, like all of the other units, impossible to accurately estimate. Outcrop width calculations, including Browns Pond (?) lithologies, suggest a maximum thickness of 600 to 700 m, however, the significance of this figure is highly questionable. No faults or macroscopic folds are mappable within this monotonous sequence to the east of the second string of quarries to the east of South Poultney, but this is not intended to indicate that they do not exist, only that they cannot be demonstrated at the present mapping scale. The density of meso- and macroscopic folds and thrusts in the Pawlet Formation, for example, might suggest that this thickness could be only 20 to 30% (120-160 m) of that stated above.

A new name is introduced here because previous names, including Cambrian Roofing Slate (Dale, 1899), Mettawee Slate (Ruedemann, in Cushing and Ruedemann, 1914; Jacobi, 1977), Mettawee Slate Facies of the Bull Formation (Zen, 1961, 1964a, 1964b, 1967; Theokritoff, 1964). Mettawee Slate Facies of the Nassau Formation (Bird, 1962; Potter, 1972) or St. Catharine Formation (Doll and others, 1961; Shumaker, 1967) are not precisely or demonstrably correlative, were poorly or improperly defined, or contain marker horizons not observed within this area. As discussed previously, the Bull and equivalent formations are not at present, a clearly closely related association of lithologies.

Dale (1899) described all purple, green, grayish-green and variegated purple and green slates, which are commonly quarried as Cambrian Roofing Slates. Calcareous quartzites and limestone conglomerate or breccia were included within this unit. Ruedemann (in Cushing and Ruedemann, 1914) introduced the name Mettawee Slate to replace Dale's name and specified a type locality on the west side of the Mettawee River, in a series of quarries north of Middle Granville, New York. Most subsequent workers have used the name Mettawee Slate or Mettawee Slate facies to refer to any primarily purple and green with lesser gray slates and argillites (Zen, 1961, 1964a, 1964b; Potter, 1972; Fisher, 1977). Zen (1961, 1964) followed the usage of Swinnerton (1922) and suggested that "Mettawee" slates and argillites form the matrix of the Bull Formation. Included in the Bull Formation as members are Bomoseen Graywacke, Zion Hill Quartzite and Graywacke, Mudd Pond Quartzite, and North Brittain

Conglomerate. Zen (1961) referred to the matrix slate as Mettawee Slate facies of the Bull Formation. Zen did not distinguish between the slates and argillites at the different levels within the Bull Formation. Potter (1972) (Figure 1), mapping in the central Taconics,subdivided the Mettawee slate facies (of the Nassau Formation) into three subfacies, lower, middle, and upper, based primarily on stratigraphic position relative to prominent marker horizons.

According to Potter (1972), the lower Mettawee subfacies (Mettawee-a) lies below and interfingers with the Rensselaer graywacke member. Mettawee-a is interbedded dusky red, fine grained slates, fine grained quartzites, and thin lenses of Rensselaer graywacke; also locally green and purple slate and thin lenses of Bomoseen graywacke. Mettawee-b, the middle Mettawee subfacies, lies between the Rensselaer and Bomoseen graywacke members and constitutes the thickest slate unit of the Mettawee. It consists of soft, green, gray, or purple silty argillite. The argillite varies from thinly cleaved soft slate to crudely cleaved siltstone. Mica spangles are common on broken surfaces. Purple silty slate and silty argillite are abundant, but do not constitute mappable units. Fine grained quartzites in beds 2.5 to 45 cm thick occur sparingly. The upper Mettawee, Mettawee-c, generally lies above the Bomoseen graywacke member and below or interbedded with the West Castleton Formation. Mettawee-c is lithologically similar to Mettawee-b, but is distinguished by finer grain size, less thin quartzites, more common rusty weathering, and the presence of black slates above or locally interbedded with them. These slates form the matrix for the Mudd Pond Quartzite and fossiliferous limestone lenses. Mettawee-c probably corresponds to Dale's Cambrian Roofing Slate.

Jacobi (1977) mapped in the type locality of the Mettawee Slate and found that Mettawee Slate refers specifically to a purple, green with minor gray, extensively quarried slate horizon lying between two black slate units; Browns Pond below and Hatch Hill-West Castleton above. Below the Browns Pond lie the Truthville slate described above.

Her work implies that to be stratigraphically kosher, the name Mettawee Slate should be used only to refer to purple, green and lesser gray slates lying within this stratigraphic position, and that the common usage of Zen, Potter, and others is incorrect. Rowley et al (1979) noted this problem but did not deal with it satisfactorily. They defined two types of Mettawee, based on whether the slates could be demonstrated to lie in the correct stratigraphic position. Those slates that did not were referred to as Mettawee Slate (sensu lato). It is proposed here that the use of the name Mettawee Slate should be entirely restricted to those slates demonstrably lying in the same stratigraphic position as mapped in the type locality. Another name should be used for similar or possibly correlative slates where this position cannot be demonstrated.

One possible name is St. Catherine Formation (spelled St. Catharine, by Shumaker, 1967) informally introduced by Shumaker (1960) and formally introduced by Doll and others (1961). The original definition includes: "purple, gray-green, and variegated slate and phyllite containing minor interbeds of white to green quartzite, locally albitic. Purple and green chloritoid-bearing slate and phyllite". It also contains (1) the Bomoseen Graywacke member and (2) the Zion Hill Quartzite. Shumaker (1960, 1967) also included the so-called Castleton Conglomerate, a name initially used by Zen (1959) but later changed to North Brittain Conglomerate (Zen, 1961) in order to avoid confusion with the West Castleton Formation. A specific type locality was never designated by Doll and others (1961) or

Shumaker (1960, 1967). Shumaker indicates that typical exposures of St. Catherine Formation can be seen around Lake St. Catherine.

The name St. Catherine Formation is not resurrected because:

(1) As originally defined, and from map distribution this formation is identical to the Bull and Biddie Knob Formations of Zen (1961, 1964). It would thus be better to continue the use of the more familiar name than reintroduce an old one.

(2) Doll and others (1961) and Shumaker (1967) map large tracts of the higher Taconic, Bird Mountain and Dorset Mountain slices as St. Catherine Formation even though stratigraphic equivalence cannot be demonstrated.

(3) Neither of the members mentioned by Doll and others are found in this area.

Fisher (1977) suggests that the name Nassau Formation should be applied to all rocks previously mapped as Bull Formation in the northern Taconics, since their stratigraphic equivalence is well established. A critical requirement of a formation designation is that it be characterized by "a substantial degree of overall lithogic homogeneity" (Hedberg, 1976, p. 31). Jacobi (1977) demonstrated that the lithologies included within the Bull Formation, and included by her in the Bomoseen, Truthville, Browns Pond, and Mettawee Slate Formations do not possess such a lithologic homogeneity and therefore were separated. Her work indicates that the concept of Bull and Nassau Formation needs revision. In order to avoid confusion and because many of the lithologies commonly attributed to the Bull Formation and/or Nassau Formation are not present in the study area these names are not applied. Instead the name Bullfrog Hollow lithozone is introduced to refer to purple, green, and lesser gray slates and argillites, with minor micritic limestones, thin silty quartzites, rarer thicker quartzites, and a thin member of medium gray slate and associated arenaceous lithologies. The type locality (Figure 8) is designated as lying on the east side of Bullfrog Hollow, which lies one km to the west of Little Lake at the south end of Lake St. Catherine, and between South Poultney and Wells, Vermont. The type section extends from the quarries of green and gray slates on the west of Bullfrog Hollow Road, just north of the fork in the road, to the top of the 787 foot high hill approximately 800 m east northeast. This section includes all lithologies commonly found within the Bullfrog Hollow lithozone. The term lithozone is used instead of formation because neither the top nor the bottom contacts are known, being everywhere mapped as thrust contacts and because the section is complexly folded and probably faulted, although present data does not allow accurate delineation of these complexities.

No fossils have been collected from within this particular area, so that the age is uncertain. The lack of Hatch Hill-West Castleton lithotypes further complicates the age determination. By correlation with the section exposed to the west (Jacobi, 1977) the most likely age assignment is Pre-Cambrian (?) to early Cambrian.

Mettawee Slates and Browns Pond in their type areas contain *Elliptocephala asaphoides* faunas of medial early Cambrian age. No fossils have been reported from the Truthville or lower part of the Browns Pond. If the siltier part of the Bullfrog Hollow corresponds to the Truthville or part of the Bomoseen this would suggest a Pre-Cambrian (?) age. A correspondance between lithofacies-B and type Mettawee would suggest a medial early Cambrian age. The lack of fossils in this area precludes a precise

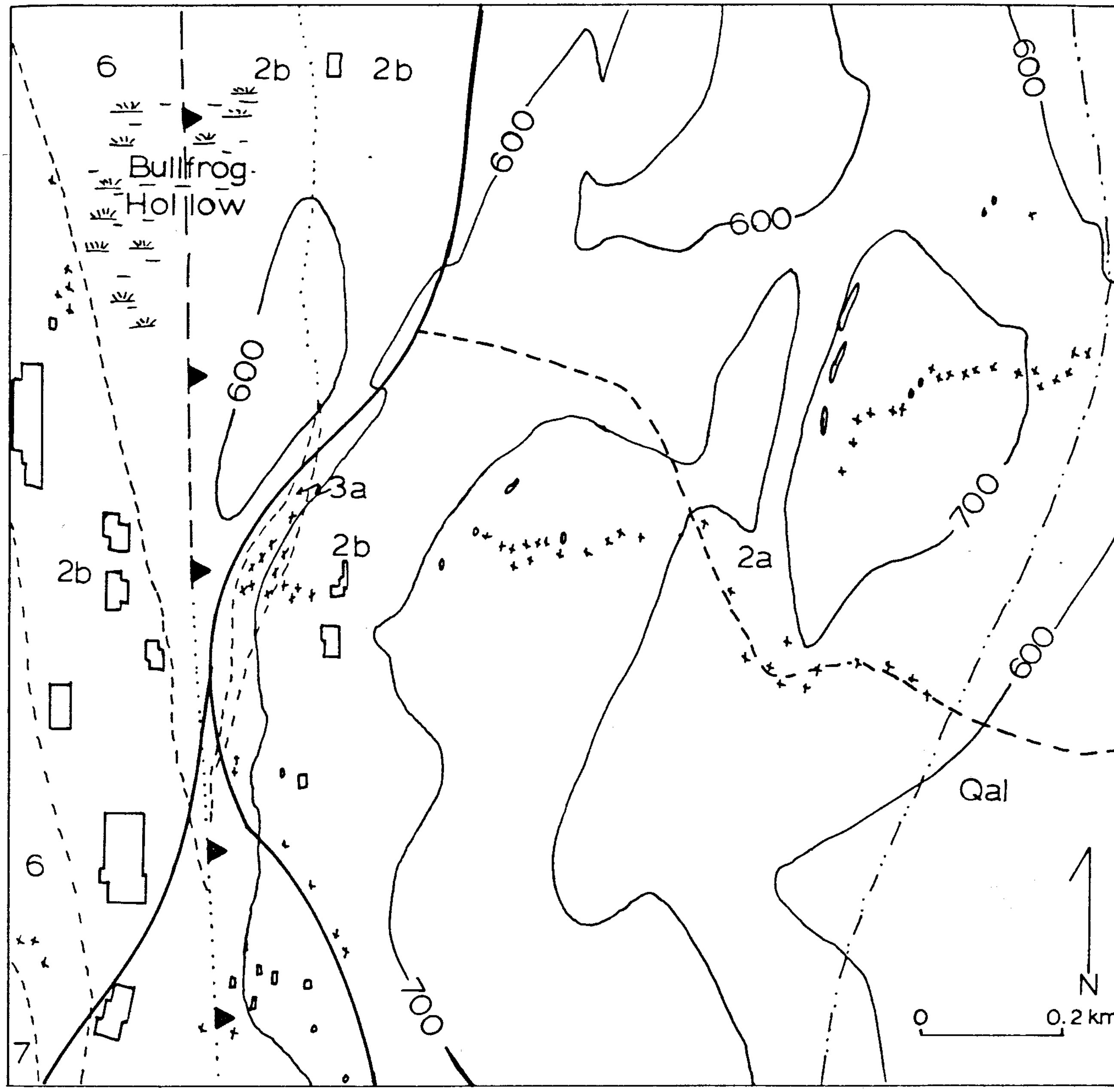


Figure 8

Location map of type locality of Bullfrog Hollow Lithozone along the east side of Bullfrog Hollow. Units 2a-both lithofacies a and b of Bullfrog Hollow, 2b-only lithofacies b, 3a-Browns Pond (?), 6-Poultney, 7-Indian River.

Figure 9

Folded, brown weathering carbonate in purple and green slates.

Figure 10

Dismembered folds of brown weathering sandy carbonate and light weathering arenite in purple, green and gray slates.

Figure 9: Folded, brown weathering carbonate in purple and green slates.

Figure 10: dismembered folds of brown weathering sandy carbonate and light weathering arenite in purple, green and gray slates.

biostratigraphic correlation. Shumaker (1967) reports probable *Elliptocephala asaphoides* age fossils from his Castleton Conglomerate, but his localities were not from the present area.

Browns Pond Formation

The Browns Pond Formation, informally introduced by Jacobi (1977), consists of a heterogeneous assemblage of calcareous and arenaceous rocks within a primarily black slate matrix. The matrix is predominantly black, with lesser dark gray, finely cleaved, and rather fissile slate. Interbedded are limestones, limestone conglomerate and breccia, black calcareous quartz wacke, thin dolo- to calc-arenites, and one or two thick clean quartzites. Descriptions of these lithologies are in Jacobi (1977) and Rowley et al (1979).

Browns Pond lithologies are recognized in two localities in this area to the west of the South Poultney Thrust. Browns Pond (?) rocks have been mapped to the east of the South Poultney Thrust, but the correlation is tentative. The two different occurrences are described separately below.

Browns Pond - West of the South Poultney Thrust

Browns Pond rocks are exposed northwest of the South Poultney along the east sides of the two prominent short strings of Truthville slate quarries (Figure 11), and along Fox Road just west of the New York-Vermont state border (Figure 12). The best development is to the east of the westernmost string of Truthville quarries. Here the Browns Pond consists of black and dark gray slates and interbedded thick quartzite, black quartz wacke, calcarenites to calcisiltites, thin quartzites, and limestone breccia.

The matrix slate is black to dark gray, often brownish to rusty weathering, well cleaved, fissile and sometimes somewhat silty. The other lithologies interlayered with this slate are lensing and inter-

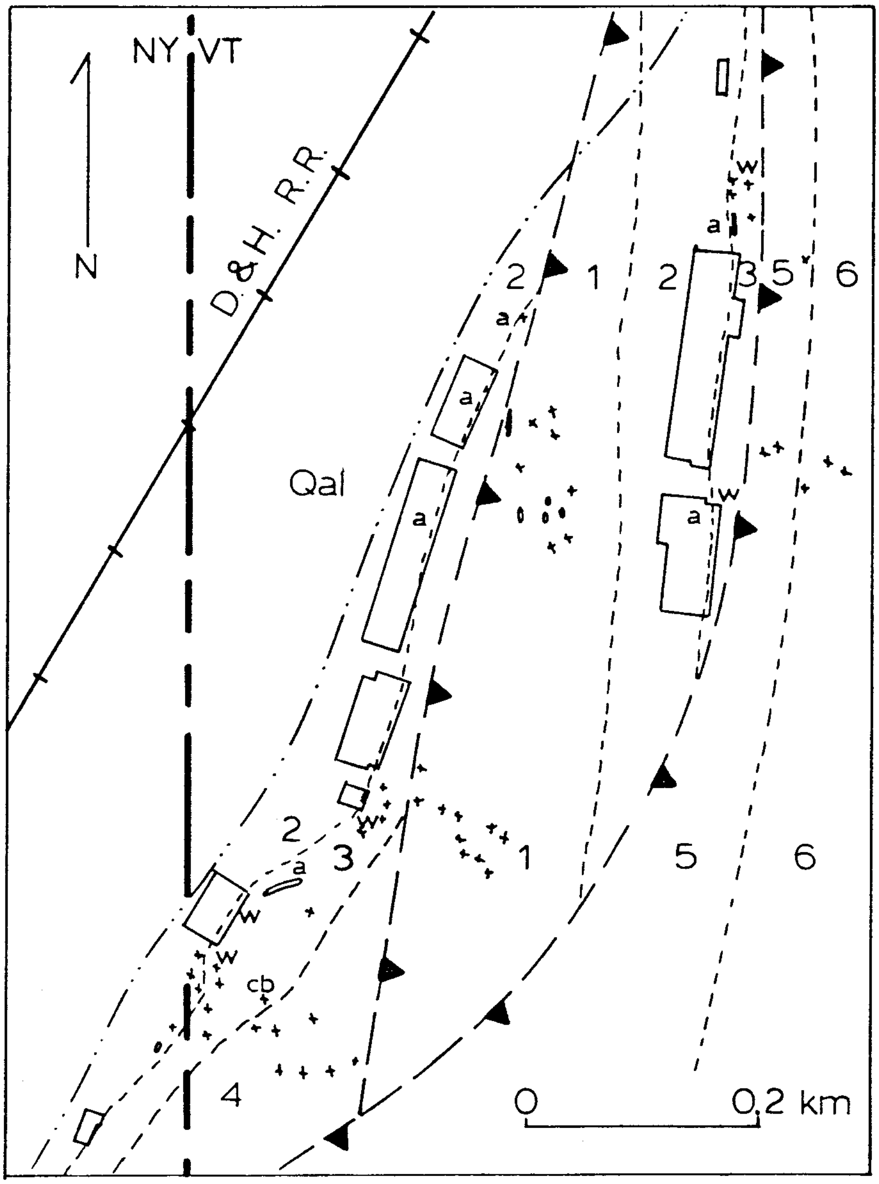


Figure 11

Location map of outcrops of Browns Pond Formation northwest of South Poultney. a-arenite, cb-carbonate breccia, w-wacke ('Eddy Hill Grit'). Units 1-Bomoseen, 2-Truthville, 3-Browns Pond, 4-Mettawee, 5-West Castleton and Hatch Hill, 6-Poultney.

mittently present. These lithologies are described from west (bottom) to east (top) as found in the section east of the western string of quarries. The top of the underlying Truthville Formation consists of green and gray fissile slates that become siltier towards the contact with the Browns Pond. Just below the contact of gray slates with dark gray slates, one or two lensing 1-1.5 m thick gray weathering vitreous quartzites are commonly present. The silty gray-green slate is in sharp contact with dark gray slate, often associated with widely dispersed rounded fine to medium grained quartz. This lithology somewhat resembles the calcareous quartz wacke which occurs from one to five meters above the contact, except that it is finer grained, lacks black slate rip-up clasts, and the quartz grains tend to be more widely dispersed. Usually there is a break in outcrop at this point; where outcrop is present, dark gray to black slate, is sometimes associated with another thick (1-2 m), clean, vitreous medium grained quartzites, and thin silty to fine grained quartzites. To the east brownish weathering, black calcareous quartz wacke crops out. This wacke is characterized by dispersed medium to coarse, rounded quartz grains within a black slaty matrix. The quartz grains appear black due to the darkness of the surrounding matrix. In one extremely good exposure (Figure 13a) a wide variety of lithologies, sometimes occurring as large (up to 1-1.5 m diameter) disoriented blocks were found dispersed within a matrix primarily composed of calcareous quartz wacke. These other lithologies include (1) dark gray calcarenites to calcisiltites thinly interlayered with black to dark gray slate; (2) finely laminated dark gray to black silty, calcareous siltstone, and lighter gray, thin (1-5 mm) calcisiltites; (3) lithologies similar to 2, with round (1 to 4 cm diameter) rotten

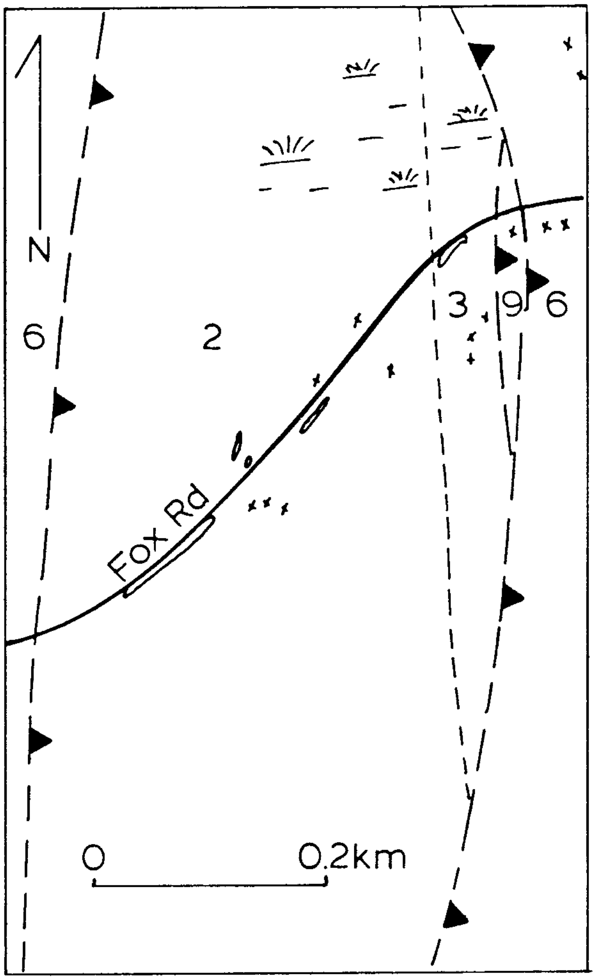
Figure 12

Location map of Browns Pond along Fox Road. Units 2-Truthville, 3-Browns Pond, 6-Poultney, 9-Pawlet.

Figure 13a

Location map of outcrop of slump blocks within the Browns Pond Formation.

sb-slump blocks. Units 1-Bomoseen, 2-Truthville, 3-Browns Pond, 4-Mettawee, 5-West Castleton and Hatch Hill, 6-Poultney, 7-Indian River, 8-Mount Merino.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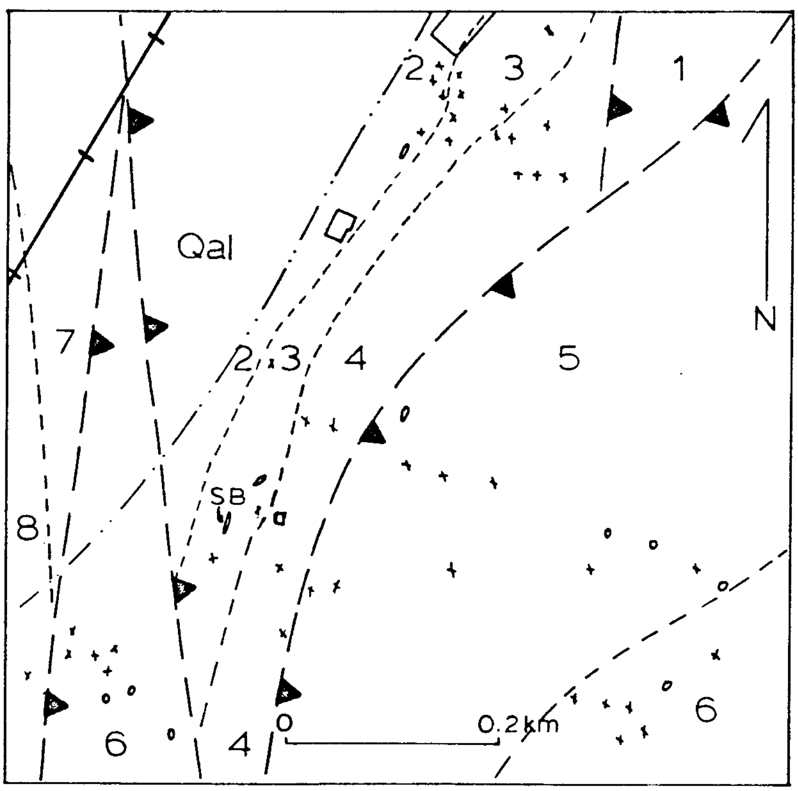
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Figure 13b

Outcrop sketches of blocks within a silty to sandy wacke matrix in the Browns Pond Formation.

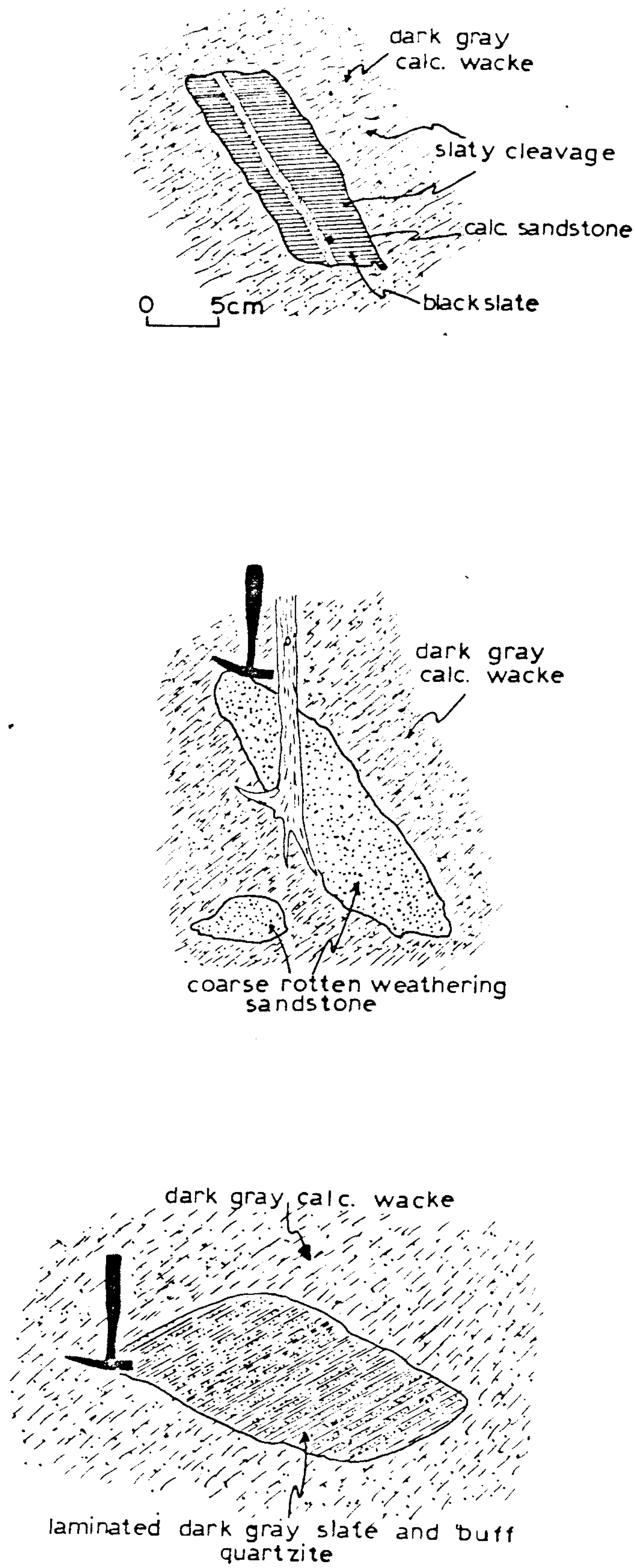

dark gray
calc. wacke
slaty cleavage
calc sandstone
black slate
0
5cm
dark gray
calc. wacke
coarse rotten weathering
sandstone
dark gray calc. wacke
laminated dark gray slate and buff
quartzite

FIGURE 13.

weathering, medium grained sandstone balls; (4) gray calcarenites interlayered with brown weathering, gray to dark gray silty slate. The calcarenites usually have a brown weathering rind a few millimeters thick, but when fresh are hard and fine grained; (5) extensively rotten weathering, friable, medium to coarse grained subrounded quartz sandstone; (6) laminated dark gray to black slate interlayered with thin (1-1.5 cm) buff quartzites. This outcrop is interpreted to be a debris fall or rock fall, and is probably a channel-fill deposit. Black slates, up to 5 meters thick are found above this wacke-matrix horizon. Locally above this is a thin (less than 1-2 m) lensing limestone breccia. This lithology is characterized by dove gray weathering, dark gray calcisiltite clasts (up to 10 cm maximum dimension) closely packed in brownish weathering sandy to slaty matrix. The matrix constitutes less than 10 to 20% of the rock volume. This limestone occurrs at or near the top of the succession, similar to its position farther to the west (Jacobi, 1977; Rowley et al, 1979). In some exposures (1 to 3 cm thick) gray weathering dark gray laminated, calcisiltites are interlayered with black slates; however, the position within the sequence is unknown. Nowhere in this area is the whole range of lithologies associated with the Browns Pond to the west developed.

The Browns Pond Formation in this area appears to have a maximum thickness of 25 to 30 meters, and commonly only 10 to 15 meters. Jacobi (1977) reports a thickness of from 25 to 130 meters, with typical values of 80 meters.

Browns Pond (?): East of the South Poultney Thrust

The Browns Pond (?) of this region crops out in an apparently discontinuous belt to the east of the eastern most prominent string of quarries of Bullfrog Hollow, lithofacies-B. The lithologies described here as Browns Pond (?) are essentially a thin (less than 10 m) member of the Bullfrog Hollow lithozone, but are separated out because of their probable correlation to the thick and prominent Browns Pond Formation to the west. Wright (1970) describes a similar thin member to the north, which he included with the Bull Formation.

The Browns Pond (?) consists of arenites and calcareous quartz wacke interlayered with gray to dark gray and locally black slate. The slate is fine grained, well cleaved, and commonly fissile. It may be brownish weathering, presumably indicating presence of carbonate. The wacke is brown weathering, dark gray to black, and calcareous with medium to coarse, rounded quartz grains dispersed in a slaty to silty wacke matrix. Black slate rip-up clasts are locally prominent. This wacke is often mesoscopically identical to the black quartz wacke mapped to the west (Jacobi, 1977). Associated with this wacke are one or several 20 to 150 cm thick quartzites. These quartzites are variable and include: gray weathering, gray,clean, vitreous, massive quartzite; rotten weathering, medium grained sandstone (when fresh, it is highly calcareous); and argillaceous gray quartzite. The arenaceous layers are demonstrably lensing, and not always present. Some small, white weathering, black phosphate (?) pebbles, and very locally elongate lenses are also observed. Commonly to the east of the Browns Pond (?) is a gray-green silty, moderately well cleaved slate with thin (1-2 cm) white weathering, silty quartzites. These gray-green slates may be finely mica spangled. These are in turn continuous with quarried purple and green slates of the Bullfrog Hollow Lithozone.

The tentative correlation with the Browns Pond Formation is based on the common association of calcareous quartz wacke and clean vitreous quartzite, found here and in the type locality. Zen (1961, 1964a, 1964b) has also noted this common association and referred to it as the Mudd Pond Quartzite facies of the Bull Formation. The lack of a black or even sometimes dark gray slate matrix, and lack of other characteristic lithologies, particularly carbonates, means the correlation can only be tentative and partial at best. The only other possible correlation might be with the Hatch Hill-West Castleton Formation. This is the tentative correlation that Shumaker (1967) made (see Poultney discussion); however, he never showed this on his map. This correlation is not favored because of the lack of black slates, well developed dolomitic quartz arenites and associated limestones, and the apparent stratigraphic position within purple and green slates of the Bullfrog Hollow. Furthermore, lithologically correlative horizons to the calcareous quartz wacke, and clean, vitreous quartzite are not known from either of these units.

Rowley et al (1979) discussed the problems of mapping and lithologic correlation using marker horizons within the Taconics. They assessed the reliability of many commonly used lithologies, including clean, vitreous quartzite, often called the Mudd Pond Quartzite (Zen, 1961, 1964a, 1964b); calcareous quartz wacke referred to in Taconic literature as the Black Patch Grit (Dale, 1899), or Eddy Hill Grit (Ruedemann, in Cushing and Ruedemann, 1914); and the limestone conglomerate or breccia horizon, variously named the North Brittain Conglomerate (Zen, 1961), Castleton Conglomerate (Zen, 1959; Shumaker, 1967), Ashley Hill Limestone (Dale, 1892; Bird 1962) or Stuyvesant Conglomerate (Fisher, 1961, 1977). Almost all, except Potter (1972) mapped the limestone conglomerate-breccia [referred to as a brecciola by Lowman (1961); see Friedmann (1972)] as part of the Bull or Nassau or equivalent formations. Potter (1972) includes it in the West Castleton Formation. Most consider the Mudd Pond Quartzite and associated Eddy Hill/Black Patch Grit lithology as part of the Bull Formation or equivalent units. However, Fisher (1977) in a recent compilation of New York stratigraphy, shows these lithologies as occurring at the base of the West Castleton Formation. The analysis of Rowley et al (1979), based on the detailed mapping of Jacobi (1977) indicates that the calcareous quartz wacke occurs only in the Browns Pond Formation, the limestone breccia occurs in either the Browns Pond Formation (most common) or the Mettawee Slate. Medium grained vitreous quartzites occur through a large range within the stratigraphy and are therefore not indicative of stratigraphic position. The presence of the calcareous quartz wacke in the Browns Pond (?) is an important aspect of the proposed correlation, and supports it.

The limestone breccia and other limestones within the Browns Pond Formation contain the medial early Cambrian Elliptocephala asaphoides faunal assemblage (Theokritoff, 1964). These are the oldest fossils yet reported from the Taconic Allochthon.

Hatch Hill-West Castleton

Hatch Hill and West Castleton lithologies are not differentiated in this area due to very limited and generally poor exposure. The lack of distinguishing arenaceous component further hampers this situation. The bulk of the rocks mapped as Undifferentiated Hatch Hill-West Castleton probably belong to the Hatch Hill Formation. Hatch Hill-West Castleton lithologies are not recognized to the east of the South Poultney Thrust. Where observed, the Hatch Hill-West Castleton is chracterized by an apparently thick sequence of monotonous black with very minor very dark gray slate, locally interbedded with rotten weathering arenites. The slates are soft, fissile, well cleaved, and commonly rusty brown weathering. These slates tend to occur as large areas of abundant float, particularly in fields, and are associated with small outcrops.

The arenites are brownish weathering, medium grained, locally cross-laminated, and are not very abundant. The arenites range in thickness from 2 to 25 cm, with most only 4 to 8 cm. Locally the arenites are veined with both quartz and carbonate, but seldom as extensively as to the west (Jacobi, 1977). They are distinguished from Poultney quartzites by their brown weathering, and generally coarser grain size. The association with fissile black slates is also important. Very rarely, thin (5 to 12 cm) gray weathering, medium gray calcarenite to calcisiltite are observed. These have a well developed spaced 'solution' (?) cleavage. They occur as dense, multiply bedded, layers, close to the Poultney-Hatch Hill contact and may be correlative with the lower Poultney-Dunbar Road Member of Jacobi (1977). They are interlayered with soft, fissile black to dark gray slates and are here included in the Hatch Hill- West Castleton Formation.

Nowhere was the base of the Hatch Hill-West Castleton observed. It is always mapped in thrust contact with surrounding units so that the nature of the lower contact and the thickness of this unit are indeterminable. Outcrop width calculations suggest a poorly constrained minimum thickness of 30 meters. Jacobi (1977) found considerable variation in thickness, and reports a range from 35 to 205 meters, and typically about 150 meters.

To the east of the South Poultney Thrust, this unit is unrecognized. Possible reasons for its absence are discussed in a later section. (See Chapter 3).

The name Hatch Hill Formation was introduced by Theokritoff (1964). Zen (1961) introduced the name West Castleton, discarding the previously used formation name, the Hooker Slate of Swinnerton (1922), because at the type locality the rocks were unfossiliferous (Zen, 1961,p.304). Being a lithostratigraphic unit, the presence or absence of fossils is irrelevant. It is this type of confusion between litho-, bio-, and chronostratigraphic terminology which has contributed significantly to the nomenclatural problems of the Taconics, and no doubt other areas. [In Zen's favor, is the fact that Keith's (1932) original description of the Hooker Slate was "a notably black slate" (Zen, 1964b, p. 43).]

In Jacobi's area the Hatch Hill constitutes the bulk of the lithologies present at this stratigraphic level. Rowley et al (1979) suggest that the West Castleton might be a facies of the Hatch Hill, in which case it would be best to consider it as a member of it and not as a separate formation. This suggestion is supported by the presence of dolomitic arenites typical of the Hatch Hill at most localities where distinguishing limestones of the West Castleton crop out.

Based on fossil collections from the area to the west by Theokritoff (1964), Rowley et al (1979) demonstrated that the faunal assemblage of the West Castleton (lower Hatch Hill?) includes the late early Cambrian *Paedeumias-Bonnia*, *Pagetia connexia*, and *Pagetides elegans* faunas, but not the *Elliptocephala asaphoides* faunas of medial early Cambrian age as commonly reported. Bird and Rasetti (1968) reported medial Cambrian *Bathyuriscus-Elrathina* and *Glossopleura* fauna in correlative lithologies from the southern Taconics. Theokritoff (1964) reports Dendroid graptolites from the upper part of the Hatch Hill, indicating a Tremadocian age. The sequence is apparently conformable, suggesting an age range of late early Cambrian to latest Cambrian or earliest Ordovician. Landing (1977) reports conodonts of medial to late medial Trempealeauan age from the Germantown Formation in the southern Taconics.

Poultney Formation

Poultney Formation crops out extensively in this area. It is characterized by medium to dark gray argillite, with lesser green, gray and black. Fine scale color laminated argillites, often displaying bioturbation features are fairly common. The argillaceous component of the Poultney varies from well cleaved slate to silty, moderately to hard, coarsely to well cleaved argillite or mudstone. The argillite commonly weathers to a chalky white. In places fine white micas are observed.

Commonly interlayered with the argillite are thin, 0.2 to 3.0 cm, though mostly $\leq$ 1.0 cm, white weathering, white to light greenish silty to fine sandy quartzites. These thin quartzites are commonly finely parallel- and cross-laminated; dark minerals outline the laminae. Some of the quartzites are slightly calcareous, but most are not. Extension fracturing perpendicular to bedding is commonly well developed and makes the distinction between cleavage and fracturing in fold hinges extremely difficult in these quartzites. Spacing between quartzites ranges from a few millimeters to greater than 10 cm. Most often a spacing of about 5 cm is observed. Weathered outcrops commonly have a ribbed or pinstriped appearance, due to differential weathering of the slates and quartzites. These quartzites are characteristic, but not diagnostic of the Poultney. Thin quartzites are also observed in the Bullfrog Hollow, the basal part of the Indian River, and in the Hatch Hill of this area.

Thicker, silty, fine to medium grained quartzites are locally interlayered with the thin quartzites described above. These quartzites are mesoscopically identical to the thin quartzites, including fine parallel and cross-laminated character. No grading was observed in this area, although Zen (1961) reports well developed grading farther to the north

Figure 14

Color laminated slate. Color laminae interpreted to represent bedding.

Figure 15

Interbedded thin silty quartzites in gray silty Poultney slates. Structures include open F_3 fold and cross-cutting F_4 kink 'band'.

Figure 14: Color laminated slate. Color laminae interpreted to represent bedding.

Figure 14: Interbedded thin silty quartzites in gray silty Poultney slates. Structures include open F_3 fold and cross-cutting F_4 kink 'band'.

Figure 16

Photograph of hand specimen of mud chip conglomerate from Poultney.

Figure 17

Photomicrograph of mud chip conglomerate. 10X magnification. Field of view is mm across.

Figure 16: Photograph of hand specimen of mud chip conglomerate from Poultney.

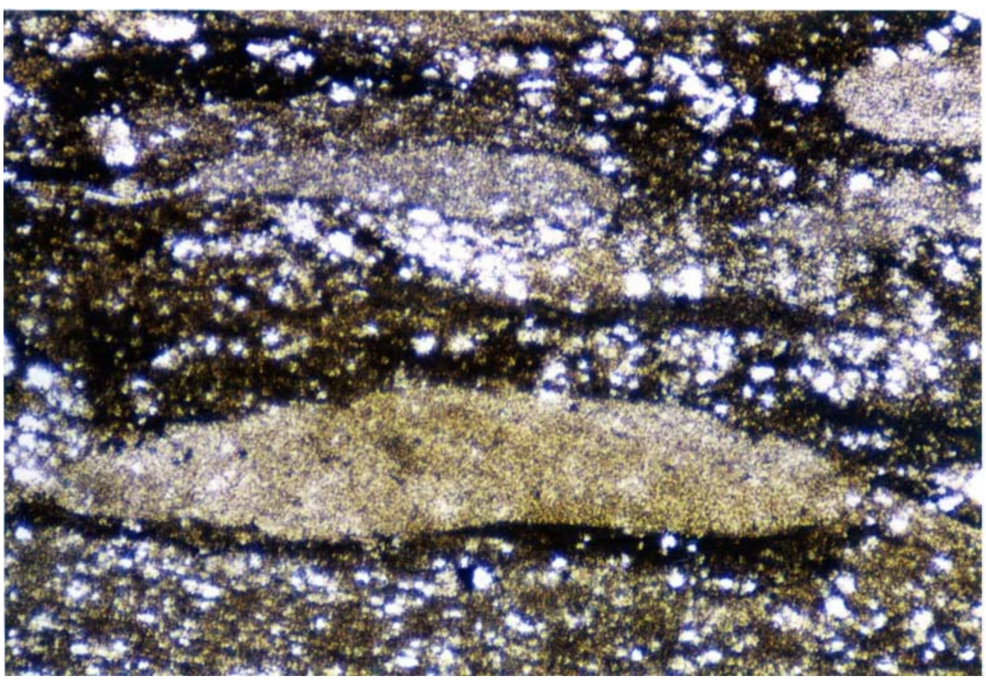


Figure 17: Photomicrograph of mud chip conglomerate. 10X magnification. Field of view is   mm across.

in similar lithologies. Also interlayered with thin quartzites at two localities are one or several 3 to 5 cm thick mud and slate chip conglomerates. The chips vary in color from gray to dark gray, and lie within a brownish gray weathering mud to silty slate matrix. The chips vary from 3 to 8 mm in maximum dimension and are mostly 0.5 to 2 mm thick. The chips are imbricated. These occurrences are interpreted as intraformational conglomerates.

Locally, gray weathering, medium gray, sometimes cross-laminated calcisiltites to calcarenites interlayered with dark gray argillites and slates are found. At one locality, in a beautifully exposed stream-cut to the west of Stoddard Road, dark gray, soft, fissile, brownish weathering slates are interlayered with brownish weathering, medium gray calcareous siltstones to fine calcarenites. These are sometimes finely cross-laminated and usually 0.5 to 1.5 cm thick. At this locality, medium gray fine silty argillites with thin quartzites and bioturbated color laminations typical of the upper Poultney (Jacobi, 1977) are exposed continuously to the east. Structural relations suggest that the dark slates and calcareous siltstone and arenites lie below the typical upper Poultney lithologies. This dark slate unit was not recognized elsewhere to the north so that the extent of it is unknown. Regional relations suggest that this unit lies in the middle of the Poultney as the upper and lower contacts lie significantly to the south and north of this outcrop, respectively. Nowhere else has a dark slate with calcareous siltstone to calcarenites been described from the middle Poultney. Jacobi (1977), Theokritoff (1964), and Potter (1972) describe a similar unit from the base of the Poultney in nearby areas. This may indicate that a more complex structural interpretation of this area is warranted.

From the map pattern it is obvious that the thickness of the Poultney is quite variable. Calculations based on outcrop width, neglecting minor folds and internal thrusting suggest a range between 30 and 150 m. The error in this calculation is likely to be extremely large, and results from the above mentioned factors as well as the fact that bounding sedimentary contacts are often not demonstrable. Jacobi (1977) reported thicknesses of 23 m and 150 to 175 m for her Dunbar and Crossroad members, respectively, with a typical thickness of 185 m (Rowley, et al, 1979). Zen (1964) and Shumaker (1967) estimated maximum thicknesses of approximately 180 and 210 m, respectively. Both noted that thickness variation was considerable, and locally Poultney was not observed.

To the west of the South Poultney Thrust, Poultney appears to conformably overlie Hatch Hill lithologies. The contact was never observed, but is placed at the first appearance of fissile black slate with or without rotten weathering quartz arenites.

To the east of the South Poultney Thrust neither Hatch Hill-West Castleton nor lower Poultney (Dunbar Member) lithologies are recognized. Poultney is everywhere in contact with Bullfrog Hollow slates. The contact has been placed at the first appearance of thin, extension fractured silty quartzites in either medium to dark gray or locally green argillite. In places the basal quartzites are calcareous.

A real difficulty arises in the eastern region because of the occurrence of silty quartzites in green argillites in both the Poultney and Mettawee formations. Since sequence is not always reliable, partly because we are often dealing with only two units, and because of the cryptic nature of most of the thrust faults, the unit assignment of some outcrops is highly uncertain. The outcrops at the north end of Bullfrog Hollow serve as a good example. They are green and greenish gray fine silty argillites with and without thin silty quartzites. They are associated with purple, green and gray quarried slates to the east, and Pawlet, with some possible Mount Merino to the west. Regional relations allow assignment to either Poultney or Bullfrog Hollow (i.e. the position of this thrust is dependent upon the stratigraphic position of outcrops - Figure 18). Dale (1899) mapped all of the quarried slate in this area as Cambrian roofing slate (Mettawee, in present usage), but only recognized a small amount of "Silurian" rocks in the middle of the string of quarries to the east of South Poultney. Shumaker (1967) mapped the outcrops in question and the western string of quarries near South Poultney as Poultney Formation. Shumaker's rationale for this placement was:

> "(1) A thin black slate (West Castleton?) separates the slates in question from purple and green pelite of the St. Catharine Formation.
>
> (2) The upper contact of the black slate (West Castleton?) is marked by a pyrite zone which gives the surface the appearance of an unconformity.
>
> (3) Irregular dark markings of an organic(?) origin are common both to these slates and to known Poultney slates farther to the west. These markings, however, occur to a lesser degree in slates of the St. Catharine Formation.
>
> (4) The lowest 10 to 15 feet of the slate in question contains quartzites very similar to those of the Hatch Hill and Poultney formations.
>
> (5) Very thin seams of brownish weathering, limy sand give the unit a subdued pinstriped appearance.
>
> (6) The upper contact of the slate is apparently gradational with red slates of the overlying Indian River Formation." (p. 28-29).

Points 1 and 4 refer to outcrops that are here mapped as Browns Pond (?) because of the local presence of calcareous quartz wacke with medium to coarse rounded to subrounded quartz grains very similar to rocks referred to as Black Patch or Eddy Hill Grit (Zen, 1964b) and apparently confined to the Browns Pond Formation to the west (Jacobi, 1977; Rowley, et al., 1979). Both east and west of the Browns Pond (?) lithologies are a series of quarries of green, gray and purple slates, and considered to be Bullfrog Hollow.

The irregular dark markings described in point 3 and the limy layers mentioned in 5 are common to both sides of the parallel string of quarries east of South Poultney. Lithologically these strings of quarries are identical and are separated by Poultney and to the south also Indian River slates. The outcrop in Bullfrog Hollow is not good enough, as implied by Shumaker's 6 statement, to demonstrate that the lithologies in question grade into Indian River. Zen (personal communication, 1978) stated that to his knowledge, Poultney slates are not commercially quarried anywhere in the northern Taconics. For these reasons, I have mapped the quarried slates as Bullfrog Hollow and the intervening slates as Poultney. The uncertain outcrops have been included in Poultney but this is tentative at best.

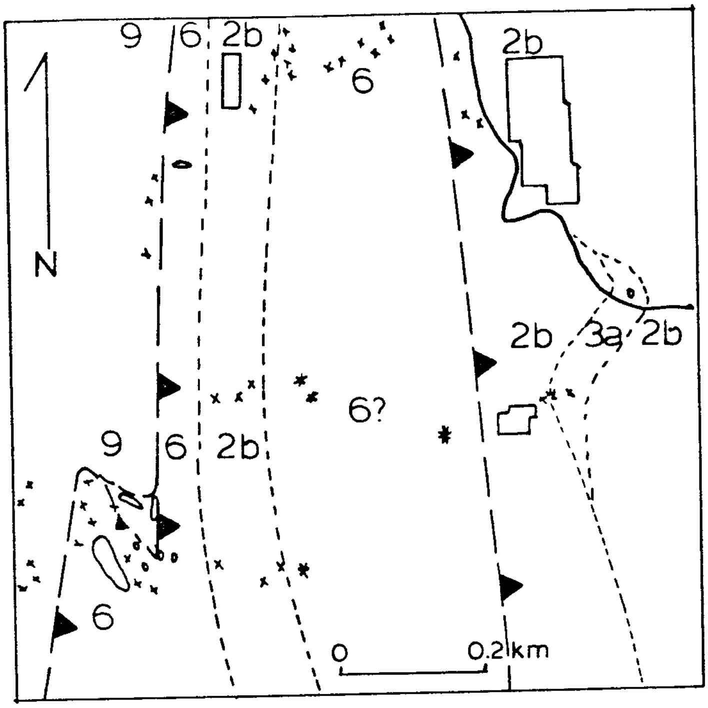
Another problematic example is the presence of several outcrops of gray to medium gray argillites with thin silty quartzites to the east of the H & H Slate Company quarries and a quarry about 300 m to the north (Figure 19). If these are Poultney then another thrust fault is required along the eastern side of this string of quarries. I have included these with the Bullfrog Hollow because both to the north and south Mettawee slate outcrops occur across this zone. However, as shown

Figure 18

Location map of outcrops of uncertain affinities (*) and interpreted to be Poultney in Bullfrog Hollow. Location of thrust fault dependent upon assignment of these outcrops. Units 2b-lithofacies-b of Bullfrog Hollow, 3a-Browns Pond (?), 6-Poultney, 9-Pawlet.

Figure 19

Location of Poultney-like outcrops (*) to the east of quarries in Bullfrog Hollow lithofacies-b. These are assigned to the Bullfrog Hollow lithozone. Units 2b-lithofacies-b of Bullfrog Hollow, 3a-Browns Pond (?), 6-Poultney, D-Jurassic (?) diabase dike.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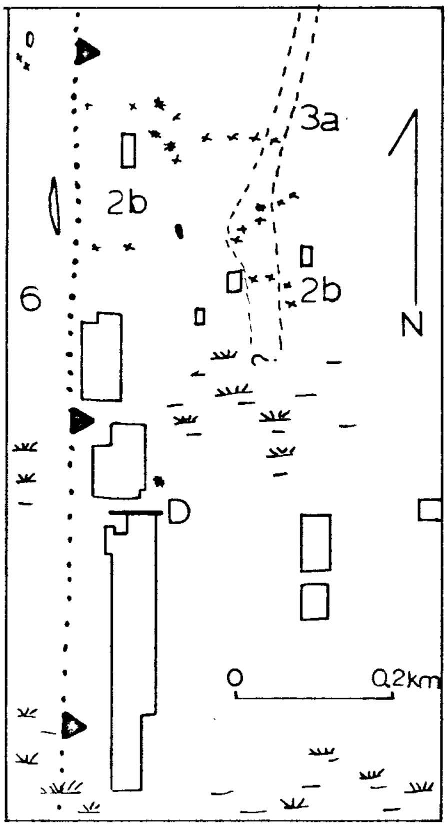
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in other places this hardly rules out the presence of a fault.

Jacobi (1977), following the work of Theokritoff (1964) and Potter (1972) and others, mapped two members of the Poultney. The lower member, Dunbar Member, is known to be lenticular and is characterized by dark gray to black slates interlayered with gray medium to fine, sometimes silty limestones. The upper member, Crossroads Member, consists of variegated argillites with thin silty quartzites and chert near the top. These are lithologically similar to Potter's White Creek and Owl Kill members, respectively. Lithologies similar to both members are present in the area to the west of the South Poultney Thrust, but nowhere is the lower member sufficiently abundant to map it separately. The problem of a dark slate and calcarenite and calcareous siltstone horizon apparently in the middle of the Poultney has already been discussed above. To the east of the South Poultney Thrust almost all of the Poultney is similar to Jacobi's Crossroad Member.

Keith (1932) introduced the name Poultney for outcrops located in the Poultney River just east of Hampton, New York. However, he included lithologies now separately mapped as Hatch Hill in his original description. In the southern Taconics, correlative units are mapped as Stuyvesant Falls Formation (Fisher, 1961) for the upper member and upper Germantown for the lower member (Fisher, 1961, 1977). Graptolites (Berry, 1962) and conodonts (Landing, 1976) of early Ordovician age have been reported from Poultney and correlative units. Poultney lithologies may extend into early medial Ordovician time (Llanvirnian to lowest Llandeilian) however, no fossils of this period have been reported, possibly because of unsuitable host rocks.

It is interesting to note that Zen (1961) in his original description of the West Castleton Formation states that it "ranges from a dark gray, hard, poorly cleaved sandy or cherty slate that weathers white or pale red, to a jet black, fissile, graphitic and pyritic slate that contains many paper thin white sandy laminae and commonly also black cherty nodules, and when weathered displays alum bloom. Locally interbedded in the fine slate are beds of buff to yellow-weathering black dolostone or dolomitic quartzite, a few inches thick, some of which, however, become massive, siliceous, and heavily bedded in the harder black slate. The varieties of black slate do not form mappable units but grade into each other along strike." (p. 304). The dark gray, hard, poorly cleaved lithology described by Zen is very similar to rocks mapped here as Poultney. The inclusion with the Poultney and not the West Castleton is based on (1) interlayering of these lithologies with lithologies containing thin Poultney quartzites, (2) locally a stratigraphic succession from good Poultney lithologies, through these siliceous slates, into Indian River, Mount Merino, and Pawlet can be demonstrated, (3) to the west Jacobi (1977) found only black fissile slates locally interlayered with limestones and/or dolomitic arenites in a position demonstrably equivalent to the West Castleton. Siliceous black or dark gray slates are unknown at this stratigraphic position in her area. This suggests to me that Zen (1961, 1964) has probably mismapped thse lithologies, and that they should be included in the Poultney Formation instead.

Indian River

The Indian River Formation is one of two extensively quarried slates in this area. Rocks mapped as Indian River include red, blue-green (sea green), pale green, and some gray (at the top) porcelaneous, well cleaved slates. Green cherts have also been included where associated with porcelaneous slates. The slates range from soft and fissile to hard and siliceous. All slates have a characteristic porcelaneous or waxy luster. Round grains may be present on cleavage surfaces. In thin section these grains are round to elliptical and composed of polycrystalline quartz. Jacobi (1977) interpreted these aggregates as radiolaria tests. These rounded grains are restricted to the Indian River and Mount Merino Formations. Some thin, extension-fractured finely laminated silty quartzites identical to those found in the underlying Poultney are present in the lower part of the Indian River, particularly in the eastern exposures. Yellowish weathering, pale gray and green flinty layers, interpreted as silicic tuff bands (W.S.F. Kidd, personal communication; Jacobi, 1977; Rowley, et al, 1979), were also noted in both western and eastern regions, however, they are not as extensively developed in this area as compared to the area farther to the west.

Jacobi (1977) interpreted the quarried gray-green hard slate just to the northwest of the intersection of Stoddard Road and Fox Road as part of the Pawlet Formation. These outcrops and similar lithologies to the north are reinterpreted as Indian River because they lie between typical Mount Merino to the east and Poultney gray slate with thin quartzites to the west. This requires that the Indian River undergo a drastic change in color from the east limb where they are typical red and sea green slates to sea green and gray-green slate on the west limb of this syncline. To the north, on the eastern limb, the Indian River appears to thin and become more maroonish colored associated with sea green slates. By the hinge exposed on Butler Road just south of Raceville the Indian River is primarily green with lesser red and maroonish slate. This along strike color variation supports the assignment of the gray-green slates to the Indian River.

The thickness of the Indian River is quite variable. In the west, the thickness ranges from 10 to 30 m. A very thick section of red slates is exposed in the hinge of the anticline east of Stoddard Road. Jacobi (1977) reports a thickness of 25 to 55 m with typical values of 50 m (Rowley, et al, 1979). To the east, the Indian River is considerably thinner and varies from 0 to 25 m in the core of the truncated syncline, southeast of South Poultney. Zen (1964) reports a thickness of approximately 30 m (100 feet) while Shumaker (1967) mapped up to 25 or 30 m of Indian River.

The lower contact of the Indian River is not observed in this area. To the west the contact is placed at the first red, sea green or pale green porcelaneous slate, lacking thin quartzites. To the east, in the hinge of the truncated syncline, the basal Indian River is red slates interbedded with thin silty Poultney-like quartzites. Stratigraphically below this the thin quartzites are interbedded with medium gray and locally green slates of the Poultney Formation. In this area the contact was placed below the first appearance of red slates.

Only red and associated green slates have been mapped as Indian River east of the South Poultney Thrust. A sequence from Poultney though green slates to black slates or Pawlet was never observed making it impossible to definitely assign green to the slates Indian River. The great abundance of green slates in this eastern region, almost all of which have been mapped as Bullfrog Hollow, makes me suspect that some of them may be Indian River, but nowhere could this be demonstrated. For example, nowhere were rounded grains observed in the east that might lead one to suspect Indian River. To be conservative I did not include any of these non-red slates in the Indian River from this area.

Keith (1932) introduced the name Indian River for the extensively quarried red slates of the Granville area. Jacobi (1977) followed Theokritoff (1964) and included sea-green slates and cherts in the Indian River as these are commonly interlayered with red slates in many quarries. This practice is continued here.

The Indian River corresponds to the lowest member of Ruedemann's (1942) Normanskill Formation. Berry (1962) found N. gracilis zone graptolites near North Granville and Hampton which were reportedly from the Indian River, but are now believed to have been collected from the overlying Mount Merino (W.S.F. Kidd, personal communication, 1979).

Berry (1968) suggested that the red coloration of the Indian River slates was primary and due to the erosion and redeposition of a fossil soil. Bird and Dewey (1970) suggested that the terra rosa soil developed on the sub-Trenton karst surface on the carbonate platform the source of the red clays. The time correspondence between the time of deposition of the Indian River and the development of the karst surface on the carbonate platform strongly support their contention. The presence of thin silicic tuff bands is the first evidence of volcanogenic material in the Taconic sequence above the Cambrian (?) mafic flows described in the Rensselaer Greywacke (Potter, 1972) and Everett Formation (Ratcliffe, 1969). The presence of only tuff bands may indicate that the source was still a long way off.

Mount Merino

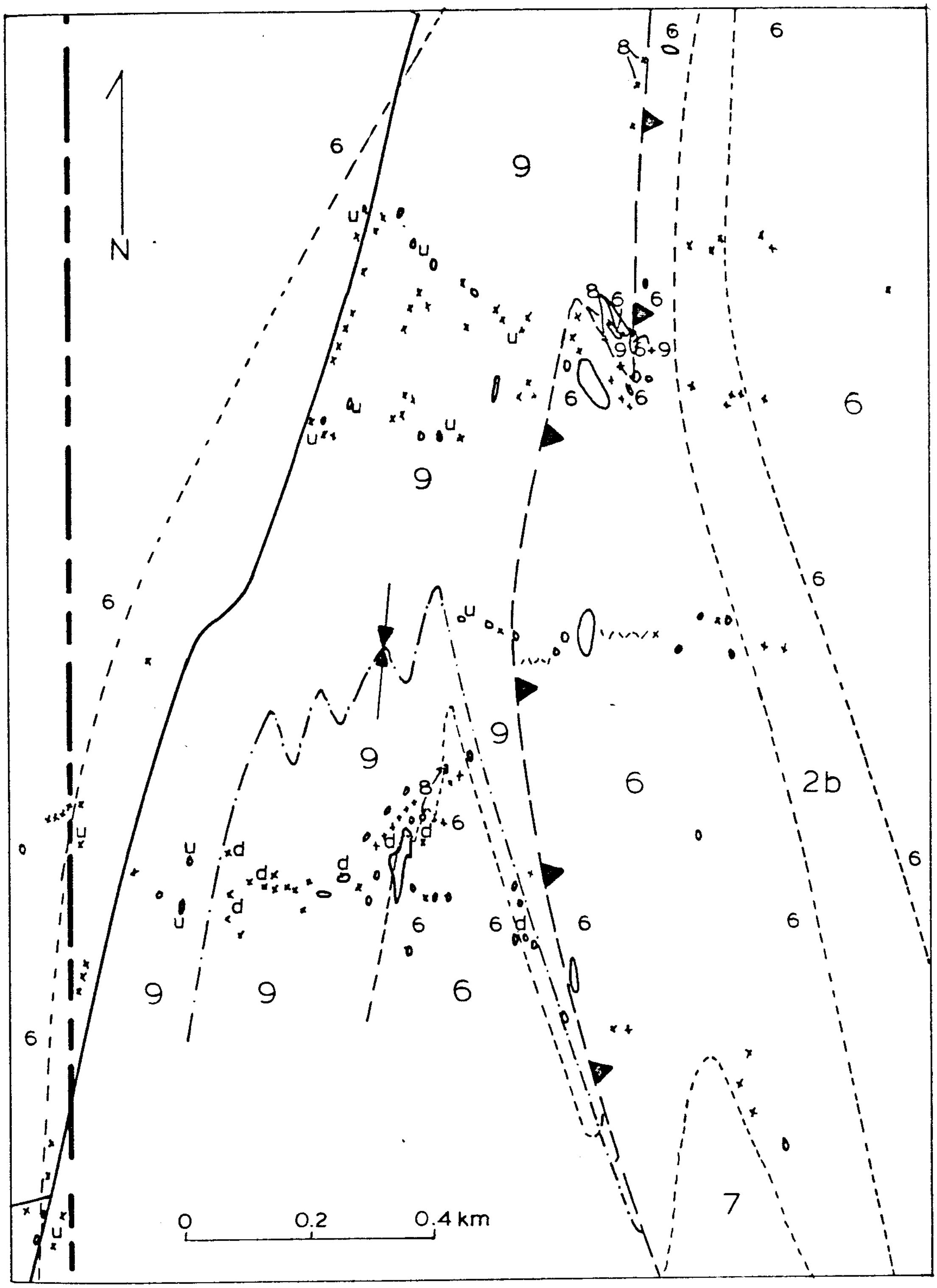
Mount Merino Formation consists of two members; a lower banded chert and slate member overlain by an upper sooty black, graptoliferous slate. The lower member, typical Mount Merino, is characterized by banded black and lesser dark green cherts 3 to 5 cm thick. The cherts are poorly cleaved or uncleaved and may be cut by quartz veining. Interlayered with the cherts are black, often siliceous well cleaved slates. Both cherts and interlayered slates weather to a chalky white. Fine color laminations within the cherts are locally observed. Round grains, probably radiolaria are sometimes present on slaty cleavage or bedding surfaces. Both chert and slate are pyrite-bearing.

The upper member is a distinctive coal black, rusty weathering, soft, slightly silty, graptoliferous slate. Slaty cleavage is well developed. Where bedding, marked by graptolite layers are oblique to slaty cleavage the graptolites act as passive strain markers and are shortened perpendicular to cleavage and appear to be extended parallel to it.

Mount Merino is only well developed and well exposed to the west of the New Boston Thrust. Farther to the east, Mount Merino is at best poorly developed. Very locally, small outcrops of black chert and/or black graptoliferous slate are present near the Pawlet-Poultney contact. Nowhere is a continuous succession, including Mount Merino cherts followed upward by black slates, found in this eastern region. The Mount Merino is not a mappable unit in this region, although outcrops where lithologies typical of the Mount Merino have been differentiated on the map (Plate 1 and Figure 20). The localized occurrence of the Mount Merino is important to interpretations of the lower Pawlet contact.

Figure 20

Location map of outcrops of Mount Merino exposures along Poultney-Pawlet contact northeast of Route 31-Fox Road intersection. u-upward facing structures, d-downward facing structures, dot-dash line with arrows is axial trace of F_1 fold. Units 2b-lithofacies-b of Bullfrog Hollow, 6-Poultney, 7-Indian River, 8-Mount Merino, 9-Pawlet. Circle-location of sketch shown in Figure 22.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9
9
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6
7
0
0.2
0.4 km

Mount Merino is best exposed and developed in the syncline just east of the Stoddard Road Thrust. Both members are well developed and attain a thickness of approximately 50 m. In the area to the west of the New Boston Thrust the thickness of the Mount Merino varies from 15 to 50 m and only one or the other member may be present, at least in outcrop. To the east, at most a few meters may be present.

The lower contact with the Indian River is sharp and marked by both a color and hardness change. The contact is well exposed due to extensive quarrying of the Indian River slates, for example, just east of Stoddard Road or in the large quarry south of Fox Road to the east of its intersection with Stoddard Road.

The contact between the lower and upper members was nowhere observed but is marked by the loss of the siliceous component characteristic of the lower member.

Ruedemann (1942) was the first to use the name Mount Merino for exposures in the southern Taconics. Jacobi (1977) was the first to map the Mount Merino as a separate formation in the Northern Taconics. Previously Theokritoff (1964) had included these lithologies with either the Indian River or with the Poultney. Poultney and Mount Merino can look disconcertingly alike in poor exposures and where the characteristic thin silty quartzites of the Poultney are absent. Jacobi (1977) did not include the coal black graptoliferous slate in the Mount Merino Formation and instead placed these at the base of the overlying Pawlet Formation, following their placement by Shumaker (1967) and Zen (1961, 1964). Rowley, et al,(1979) placed the coal black slates with the black cherts of the Mount Merino because of the greater lithologic and sedimentologic similarity between black cherts and slates than between black slates and greywackes. To the east such formational status is unwarranted for the sporadically occurring black slates and cherts and thus Shumaker's and Zen's placement is understandable.

Lang (1969) suggested, on the basis of detailed petrographic studies of the Mount Merino cherts, that they were of volcanogenic origin. Bird and Dewey (1970) followed her conclusions. The presence of suspected radiolaria on some bedding and cleavage surfaces made Jacobi (1977) suspicious of this conclusion.

Both the cherts and overlying slates contain a well developed graptolite fauna of N. gracilis zone, and possibly multidens zone (?) (Lang, 1969) of Riva (1974).

Pawlet Formation

The Pawlet Formation is the uppermost unit of the Taconic sequence. It consists of an easterly derived (see Petrography, Chapter 4) flysch-like sequence of very distinctive interbedded greywackes and dark argillites. The argillites are medium to dark gray, gray to brownish weathering, and vary from slates to fine silty mudstones. Most of the argillite is soft, fissile, and very well cleaved. Fine mica may be present on slaty cleavage surfaces. Some of the argillite is slightly calcareous. Thickness of argillite beds vary from a few centimeters to a few meters, but discontinuous outcrops makes the maximum argillite bed thickness indeterminable; it may be several meters.

The greywackes are medium to dark gray, brownish to gray weathering, and variably calcareous. Most are lithic wackes. Dark slate is the most abundant lithic fragment and often occurs as rip-up clasts. Greywacke and chert (poly-crystalline quartz) fragments are less common. Grain size varies from coarse sand to silt, but most commonly, medium to fine sand, dispersed within a clay matrix. Locally, scattered small pebbles (less than 3-5 mm) are observed. Wacke bed thickness ranges from 3 to 300 cm; locally evidence of composite beds are observed. Average bed thickness is 30 to 40 cm. In good outcrops and well exposed sections greywackes comprise between 40 and 60% of the formation thickness. Shumaker (1967) suggests that greywackes comprise up to 70%, while Zen (1961, 1964) reports that greywackes and slates are present in equal amounts. Considerable variation in greywacke to slate ratio was found in this area. More greywackes were found, on the average, in the eastern belt running through South Poultney than to the west near Stoddard Road.

The greywackes commonly show evidence of turbidite deposition, including grading, bottom structures, hints of Bouma sequences, poor sorting, and rip-up clasts of slate. Bottom structures, including low amplitude load casts, current lineations, and possible tool markings have been observed. Flute casts or good polarity current related linear structures have not been noted. The non-polarity lineations suggest transport in a north to northwest-south to southeast direction. Internal structures other than grading were seldom observed, presumably reflecting the nature of the outcrop and lack of good roadcut or river exposure. Where observed, internal structures include parallel and cross-laminations, minor convolute bedding, and entrained dark slate clasts. To the west, i.e. west of Stoddard Road, fine sandy to silty greywackes, often moderately to highly calcareous, are more common. These wackes, though interbedded with more typical lithic wackes tend to be characterized by thinner beds (3 to 15 cm, average 10 cm), more prominent parallel and cross-laminations, a common lack of a basal massive to graded section, and sharp contacts to both bottoms and tops of beds are sharp. These are typical DE-type turbidites; suggestive of more distal character (Bouma, 1962).

South of South Poultney (Figure 21), lenses of Poultney-like medium gray, fine silty argillite with thin silty quartzites were found apparently interlayered with Pawlet greywackes and slates. These lenses may be intraclasts or fault-related slivers, but, no evidence to support either of these possibilities was found. Another possibility is that Poultney-type deposition continued into or started again during Pawlet time, although this seems unlikely. Alternatively contour currents may have reworked the tops of some greywacke beds, winnowing and sorting the silt

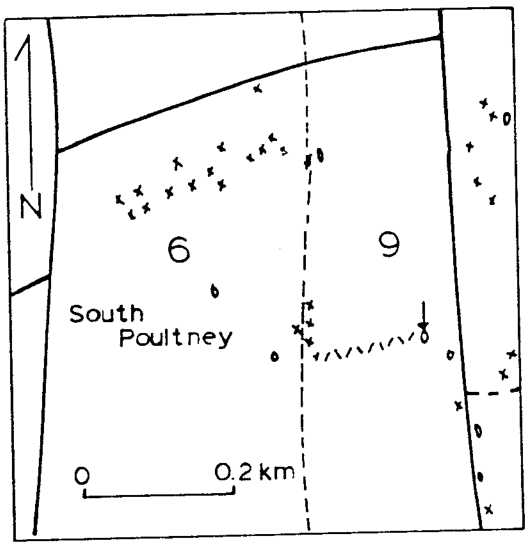


Figure 21

Poultney-like thin quartzites and dark gray slates in Pawlet near South Poultney. Units 6-Poultney, 9-Pawlet. Arrow points to specific outcrop.

component to produce Poultney-like silty quartzites. Present data does not allow a choice between these possibilities.

The original thickness of the Pawlet Formation is indeterminable as the top is nowhere exposed. The present thickness of the Pawlet is extremely difficult to establish in this area due to the presence of numerous small parasitic folds (demonstrated both mesoscopically and by short wavelength changes in facing direction). Thrust faulting and slicing on both meso- and macroscopic scales also contribute to this problem. A maximum thickness, of 100 to 175 m, neglecting small scale folds and tectonic excisions can be obtained from outcrop width calculations. The error in this estimate may be 50% or greater. Other workers in this and surrounding have estimated thicknesses ranging from less than 100 m (Shumaker, 1967) to better than 300 m (Potter, 1972). Jacobi (1977) estimated a thickness of approximately 70 m while Rowley, et al (1979) calculate a thickness in excess of 150 m in the area adjacent to the west.

The complications mentioned above and other factors, such as original thickness variations and variations in strain in different areas make it difficult to judge the quality of the estimates and also means that most estimates are susceptible to large errors.

The nature of the lower contact is quite variable in this area. To the west, near the Mettawee River and in the area mapped by Jacobi (1977), Pawlet greywackes and slates appear to be everywhere conformable on Mount Merino black cherts and black graptoliferous slates. The completeness of the section in these areas and the structural coherence indicate that this is probably the 'normal' position of the Pawlet Formation. Faunal evidence strongly supports this contention (Berry, 1962). In the

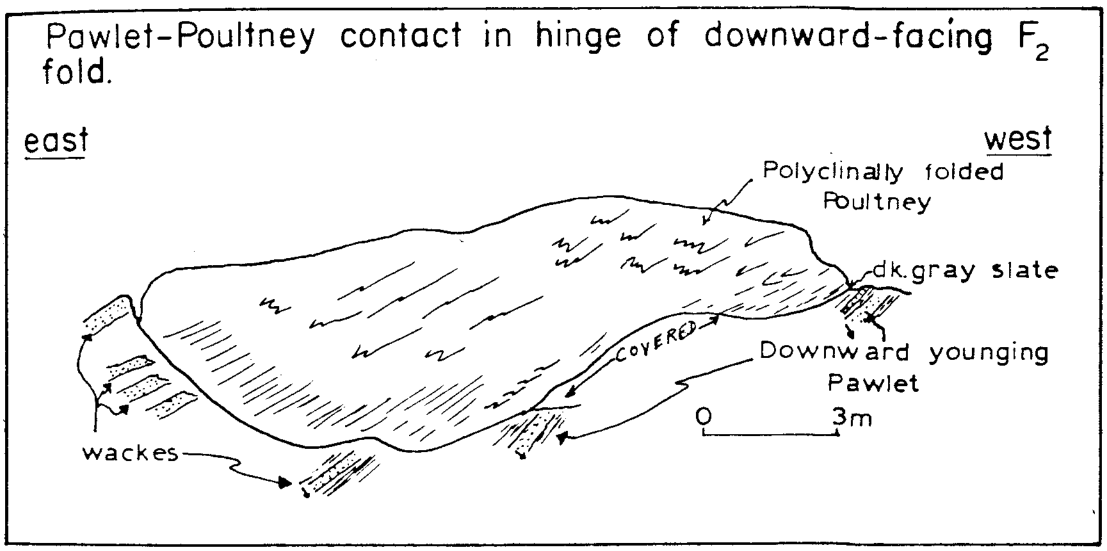


Figure 22

Outcrop sketch of Pawlet-Poultney contact in hinge of downward-facing F_2 fold. Note transposition of Pawlet wackes and Poultney quartzites into axial surface of the F_2 fold. Location of sketch shown by circle on Figure 20.

area to the west of the Stoddard Road Thrust the Pawlet maintains its 'normal' position, except where it is in demonstrable thrust contact with surrounding Poultney as in the small klippe west of the intersection of Stoddard Road and New Boston Road. Between the New Boston Thrust and South Poultney Thrust the Pawlet occurs in two modes: one mode is as very thin tectonic slivers along the base of thrust faults. The second mode is apparently sedimentary, although the contact is nowhere observed, with Pawlet overlying Indian River slates. Although Mount Merino lithologies were not observed between, this does not demonstrate their absence, so that the nature of this contact is uncertain. Farther to the east, Pawlet is primarily in contact with Poultney and less commonly Mettawee. On the hill northeast of the intersection of Fox Road and Vermont Route 31 (Figure 22), where Poultney overlies Pawlet(in the axial region of downward facing F_2 folds), the contact is exposed and marked by a 10 to 30 cm thick layer of well cleaved, gray weathering, medium gray fine silty slate. The slate has a Poultney-like appearance, but without the thin silty quartzites characteristic of the Poultney. This contact appears to be sedimentary. Less than 30 m to the east, a sooty black graptoliferous slate, less than 2 to 3 m thick, identical to the slate at the top of the Mount Merino, intervenes between Pawlet greywackes and Poultney argillites containing thin quartzites. The silty gray slate is not observed here. Black graptoliferous slate is present sporadically in the area proximal to the base of the Pawlet.

In other places, for examples northeast of South Poultney, a silky gray slate intervenes between Pawlet and Poultney Formations. A similar silky slate has been described by Zen (1961). A thrust fault is inferred in this area along or at least near this contact for the following reasons: (1) Pawlet and Poultney occur complexly interdigitated (see Stop 8 description in Rowley, et al, 1979), and (2) the thickness of inverted Pawlet does not approach the apparent thickness of right-side up Pawlet on the western limb of this synform. Other justifications are discussed later. The presence of a thrust along this Pawlet contact also negates Shumaker's (1967) contention that Pawlet rests unconformably on Mettawee (Shumaker's St. Catherine Formation).

Zen (1964, p. 29) states that: "half a mile southeast of the Poultney River along Route 30 (the road between Poultney and Wells), an outcrop 200 feet south of the road shows the Mettawee within two feet of the Pawlet Formation; the Pawlet is to the east and overlies the Mettawee. This is the closest that the unconformity has been located in the present area." At this outcrop and others farther south along this belt, Pawlet greywackes are highly deformed. In outcrops to the south (Figure 23) the greywackes are tectonically interleaved with Poultney dark gray argillites and thin quartzies, and locally have lost all evidence of original layering due to the intensity of deformation. This contact is interpreted as a thrust fault (Number 18, Figure 74)to explain the presence of transposed quartz veins and related structural complexity. This contact is not an unconformity as Zen supposed. (Highly deformed Pawlet greywackes and slates may be observed in a low road cut along the north side of Route 30 where the proposed thrust fault crosses the road.)

Figure 14 of Shumaker (1967, p. 32) is a photograph of a road cut just south of East Poultney which supposedly shows the angular unconformity below the Pawlet. This outcrop was visited, but evidence of an unconformity was not observed. Instead, an apparently continuous though overturned sequence from Pawlet greywackes, through Poultney, including

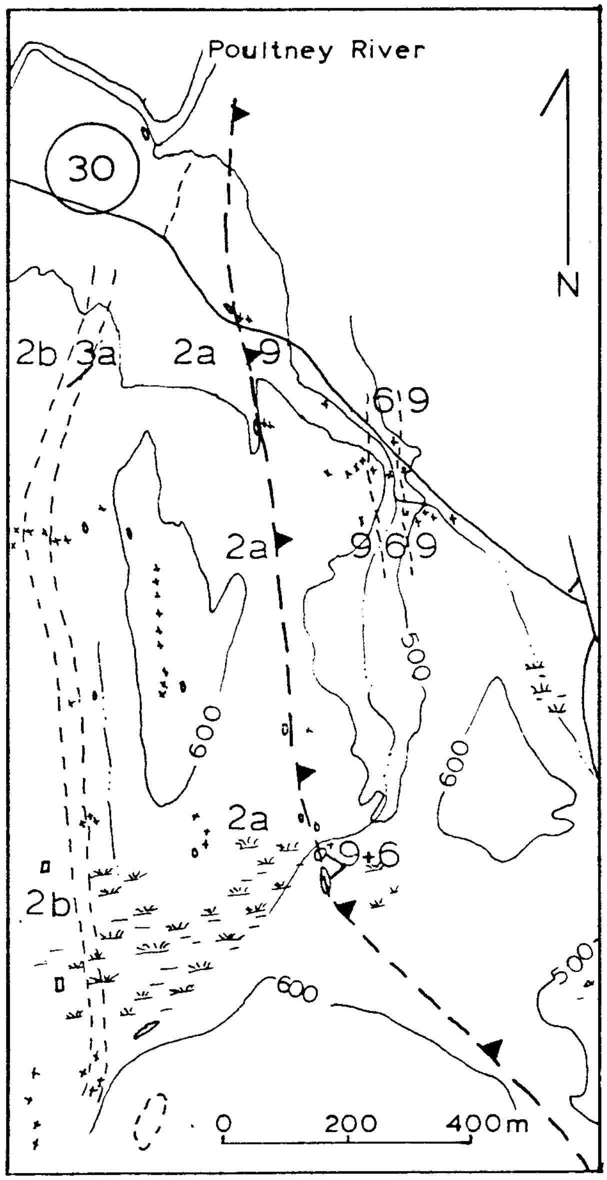


Figure 23

Location map of outcrops of Pawlet in contact with 'Mettawee', interpreted to be unconformable by Zen (1964) and in thrust contact from this study. Outcrops marked 9+6 show beautiful evidence of intimate interleaving of different units and thrust related lithologies shown in Figure 55.

Mount Merino cherts and black slates and red Indian River was observed. The angular relationship noted by Shumaker is between bedding in the Pawlet and slaty cleavage in the stratigraphically lower Mount Merino, which is not evidence of an angular unconformity.

Wright (1970) also suggests that the Pawlet unconformably overlies lower Taconic lithologies. He suggests that the underlying units experienced a pre-Pawlet phase of deformation. His description of the structures which he associates with this phase of deformation, including dismembered layering, injection structures, and his figures suggest that this phase may simply reflect syn-depositional, soft-sediment disruption. At least in this area, and farther to the south (Potter, 1972) the Pawlet is completely involved in the earliest phase of regional deformation. No evidence exists in this or the adjacent area, for an even earlier phase deformation.

No evidence of a pre-Pawlet angular unconformity was found in this area. The absence or sporadic distribution of graptoliferous slates and other lithologies that 'normally' intervene between the Pawlet and Poultney may be due to (1) non-deposition or sporadic deposition of these lithologies; (2) pre-Pawlet erosion; or (3) consistent tectonic slicing out of these units. The presence of Indian River and Mount Merino in the eastern part of the area, even though they are rare, suggests that non-deposition is an unlikely possibility. The presence of what appears to be a sedimentary contact between Pawlet and Poultney (Figures 22 and 27) means that even if significant tectonic slicing out of intervening units does occur, there remains, at least locally, a problem of their absence. Pre-Pawlet erosion, thus appears to be the remaining alternative. A reasonable possibility is that vigorous bottom current activity scoured the eastern part of the area, but left regions now represented by Jacobi's area and the western part of this area untouched. This type of scouring could conceivably leave some areas relatively untouched, and could explain the sparse presence of both Indian River and Mount Merino in the eastern part of this area.

Zen (1961) formally introduced the name Pawlet, after the informal usage of Shumaker (1960). In the southern and central Taconics the name Austin Glen (Ruedemann, 1942) is applied to correlative rocks. Fisher (1977) suggested that the name Pawlet be discarded in favor or Austin Glen. In the present usage, the name Pawlet only applies to entirely allochthonous sediments that possess the same structural history as the underlying Taconic sequence. In contrast, Austin Glen is applied to both allochthonous and autochthonous, and presumably parautochthonous greywackes and slates of the Taconic flysch sequence (for example, Potter, 1972; Rickard and Fisher, 1973). The use of the name Pawlet is continued in this thesis in order to maintain and strengthen the distinction that presently exists between allocthonous and other variably transported types of flysch.

Bird (1969) suggests that the Pawlet and Austin Glen are 'epikinallochtonous'. This implies that they range from flysch which is allochthonous, parallochthonous, parautochthonous, and neoautochthonous. Distinguishing between these different tectono-sedimentary units is important, but likely to be extremely difficult without detailed faunal and stratigraphic control. At present this type of control is not possible.

Graptolites of Riva's (1974) N. gracilis (lower to middle Caradocian) have been collected from the Pawlet and allochthonous Austin Glen. Graptolites as young as C. americanus and O. ruedemanni have been reported from probable parautochthonous to neoautochthonous Austin Glen.

CHAPTER 3

LATERAL STRATIGRAPHIC VARIATIONS

The following section compares and contrasts the stratigraphic sections from three regions spanning the width of the "Giddings Brook Slice". The western, central, and eastern sections extend from the allochthon-autochthon boundary to the New Boston Thrust, New Boston Thrust to the South Poultney Thrust, and South Poultney Thrust to the Bird Mountain Thrust on the east, respectively. Each region is characterized by a first order coherence of the internal stratigraphy, and marked difference with adjoining regions. The extent of the variation noted here within the rest of the "Giddings Brook Slice" is unknown. Figure 28 shows the stratigraphy of each of the regions.

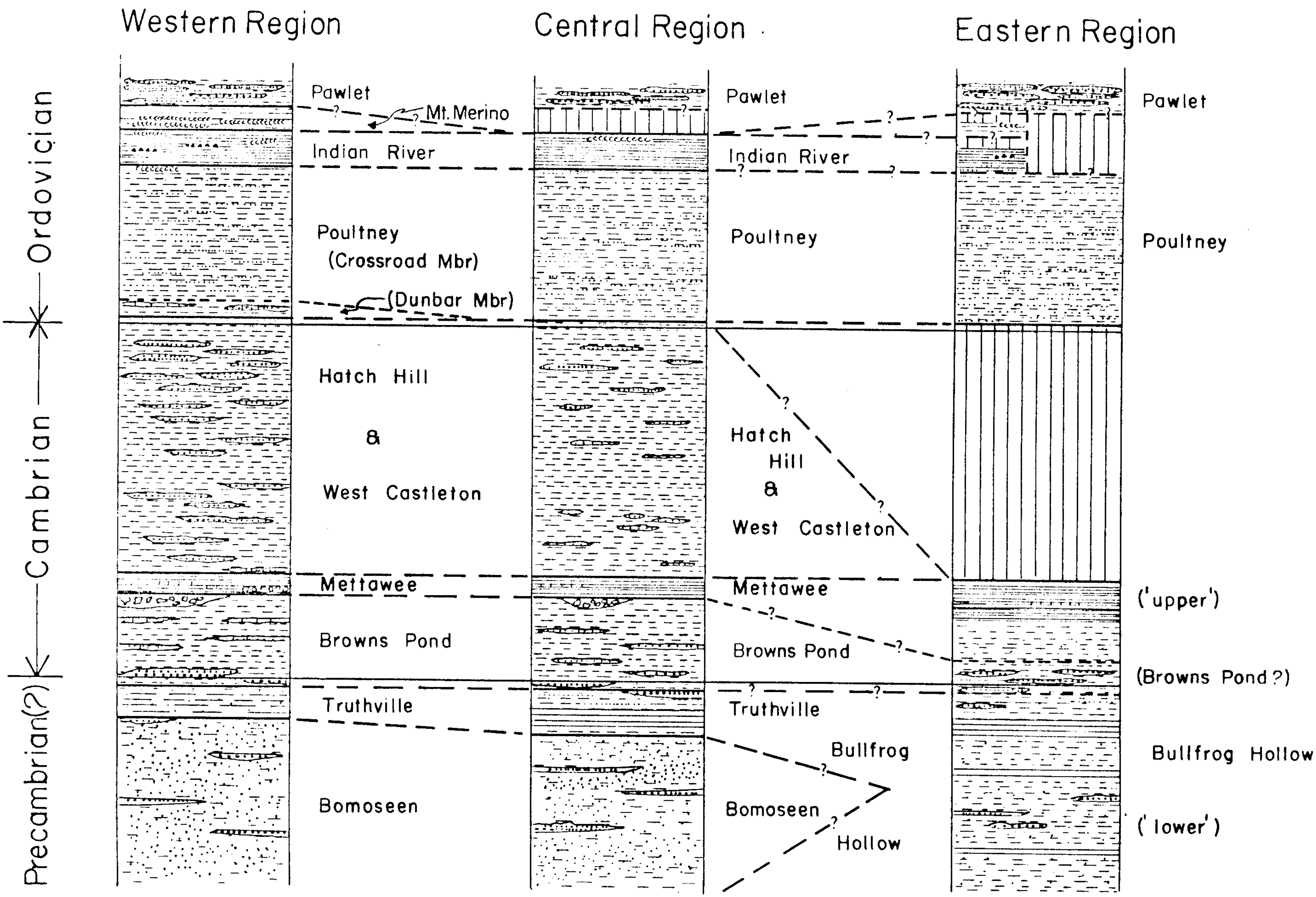
Lateral variations may result from many mechanisms, including facies changes, lateral variations in the activity of contour currents, slope morphology, large and small scale, essentially syn-depositional slumping, or from convergent tectonic phenomena, particularly removal of section due to thrusting, but also including strain variations giving rise to thickness variations. These mechanisms will be discussed more fully in a later section after the nature of the lateral variation for each unit has been described. The discussion proceeds from bottom of the section to top and from west to east.

The Bomoseen formation crops out extensively in the western region, only marginally in the central region and is unrecognized in the eastern region. Some part of the Bullfrog Hollow Formation, presumably part of the 'lower subfacies' may be equivalent to the Bomoseen. In the west, the Bomoseen is predominantly a quartz-rich wacke with lesser arenites and slates. The wacke varies from hard, poorly cleaved, dull and mica-

Figure 24

Proposed stratigraphic correlations and comparisons across the 'Giddings Brook slice'. Region boundaries are marked by thrusts, these are from west to east, Allochton-Autochthon boundary, New Boston Thrust, South Poultney Thrust, and Bird Mountain Thrust. Standard symbols are used to depict lithologies.

STRATIGRAPHIC COMPARISONS & CORRELATIONS



spangled to softer silty wacke. Thin quartz arenites are also present, most are less than 10 cm thick. In the central region similar lithologies are present, but the proportion of hard coarser wacke to softer silty wacke has decreased considerably. All lithologies are mica spangled. Farther east, in the Bullfrog Hollow Formation the lower subfacies is characterized by silty, locally mica-spangled purple and green argillite and slate with thin greenish arenites. Thicker, 0.5 to 1.0 m thick medium to coarse arenites are present in all three regions. Accepting the proposed correlations, a progressive eastward fining sequence, characterized by increasing clay and silt component and also a change from predominantly green to purple and green is recognized. The thicker arenites appear to be roughly the same.

The Truthville Formation is only distinguishable in the western region. Here the formation is characterized by a soft, silty, well cleaved, fissile slate. Rare, thin arenites are locally present. Nowhere in the western region is this slate commercially quarried. In the central region, to the northwest of South Poultney, two strings of quarries appear to be in Truthville-equivalent slates. This equivalence is based solely on the internal stratigraphy of the overlying Browns Pond Formation, which suggests that it faces east and thus the quarried slates underlie it both structurally and stratigraphically. These slates are fine grained, well cleaved, gray-green, green and gray sometimes with wormy color layering. Near the Browns Pond-Truthville contact the slates become slightly silty, weakly mica spangled, and interlayered with 1-100 cm thick quartzites. The outcrops along Fox Road just west of the Vermont-New York border are more typical Truthville soft silty slates. Gray-green soft, silty, mica spangled Truthville-like slates occur in the eastern region. However, they do not appear to form mappable horizons within the Bullfrog Hollow lithozone. The presence of probable Browns Pond lithologies suggests that some slates are stratigraphic equivalents; however, a lithologic distinction between slates above and below is not possible. By implication, Truthville slates become finer grained to the east and become indistinguishable from "Mettawee" slates.

In the type area the Browns Pond Formation is a heterogeneous assemblage of arenaceous and calcareous lithologies lying within a dark, mostly black slate matrix. Non-slate lithologies include limestones, limestone breccias, calcareous quartz wacke, thick clean quartz arenites, and thin dolo-to calc-arenites. (See Jacobi, 1977, and Rowley, et al, 1979, for more complete description). The Browns Pond of the central region is considerably less diverse, containing only calcareous quartz wacke, clean quartz arenite, and limestone breccia, in a dark slate matrix. The thickness of the unit has decreased three to four fold compared to the west. In the eastern belt, Browns Pond (?) consists of only a thin sequence of arenaceous rocks lying within a medium to dark slate matrix. Carbonate is much less abundant to the east. The overall picture suggests a thinning and fining to the east, associated with decreasing carbonate component and lighter slate matrix.

Mettawee slates are porcelaneous, well cleaved, extensively quarried, purple, green and minor gray slates in the type area. Locally, thin micritic limestones and limestone breccia are present. In the central region Mettawee slates are only demonstrable in the south between the two short strings of quarries to the northwest of South Poultney. Here it consists of buff weathering, green, porcelaneous slates. Outcrop is not extensive and other lithologies were not noted. In the eastern region

Mettawee-like slates are mapped as 'upper subfacies' of the Bullfrog Hollow lithozone. If this correlation is at least in part correct then it is characterized by green, purple, and gray, porcelaneous slates with minor thin micritic limestones and rare thin quartzites. Mettawee and upper subfacies of the Bullfrog Hollow lithozone are nearly identical suggesting that there is little lateral variation of this unit.

Hatch Hill-West Castleton are grouped together. In the western region Hatch Hill-West Castleton lithologies constitute a thick sequence of interlayered limestones and dolomitic arenites within black fissile slates. The limestones, and particularly the arenites constitute a significant porportion of this unit. To the east, in the central region, limestones are not observed and dolomitic arenites are thinner and usually finer grained than typical Hatch Hill to the west. The dolomitic arenites make up only a small part of the unit in this central section and the bulk of the unit consists of monotonous black fissile slates. No Hatch Hill-West Castleton is recognized to the east of the South Poultney Thrust. The absence of Hatch Hill-West Castleton lithologies in this area is important and may result from either sedimentary or tectonic causes. Five possibilities are described below, four of which have been discussed elsewhere (Rowley, et al, 1979).

(1) An unconformity below the Poultney due to:

a) Vigorous scouring by bottom current.

b) Non-deposition, presumably due to either lack of sediment or to continuous removal of sediments by bottom currents.

c) Large scale down-slope slumping resulting in local removal of this unit.

(2) A facies change from black slates with interbedded limestones

and dolomitic arenites to the west to purple, green, and gray slates with interbedded micrites to the east.

(3) Systematic removal of these units during D_1 folding and thrusting.

Arguments that bear on the validity of these mechanisms include the following points:

(a) The presence of Hatch Hill-West Castleton lithologies to the east in the Edgarton Half-window (Shumaker and personal observation) suggests that possibilities, 1a, 1b, and 2 require ad hoc modifications in order to explain only local unconformities, sites of non-deposition, or facies changes to purple, green and gray slates.

(b) Thick sequences of presumably chaotic, olistostrome-like deposit s of intermixed black slate, dolomitic arenite, and limestone have not been described farther to the east and thus support for mechanism is lacking.

(c) The tectonic model requires systematic removal of these lithologies along all Bullfrog Hollow-Poultney contacts. This seems improbable at least after folding has been initiated. This suggests that removal has to be pre to syn-F_1 or possibly F_2 but evidence for removal of section at this time is not observed.

The absence of such a prominent horizon is as yet unresolved and may reflect complex interactions of one or more of the above suggestions or other mechanisms.

Generalizations can only be made between the western and central regions. Hatch Hill-West Castleton thins and is associated with thinning and fining of the dolomitic arenite beds to the east.

Poultney Formation is prominent in all three regions. In the west Jacobi (1977) distinguished two members, a lower black slate plus limestone member, called the Dunbar Member, and an upper variegated and color laminated gray, green and minor black, red, and maroon slate called the Crossroad Member. The upper slates are somewhat silty and commonly associated with thin silty quartzites. Locally, a dark slate and calcisiltite to calcarenite members occurs high in the Poultney (Rowley, et al, 1979).

In the central region only the upper lithologies corresponding to the Crossroad Member of Jacobi are recognized. Here the slates are less variegated, consisting of grays, green and lesser black, and harder (more siliceous ?). The slates are somewhat silty and large thicknesses lack the charcteristic thin quartzites. Ribboning of quartzites occurs, but is less prominent than to the west. Still farther east, grays, green and minor black slates typify the Poultney along with thin silty quartzites. Very closely spaced quartzites were not observed in this region. The lower member is not recognized east of the New Boston Thrust.

In general, the spacing of quartzites, at least in terms of 'ribboning' becomes progressively less prominent to the east. The upper chert horizon locally prominent to the west in only rarely observed to the east. Color variability decreases but this may reflect problems with distinguishing between Poultney and Bullfrog Hollow Formation when thin quartzites are not present. The general character and thickness of Poultney thin quartzites is the same across the area.

The Indian River outcrops extensively only in the western region. It consists of predominantly red and sea green slates, with minor cherts and green and pale gray silicic tuff bands. Gray slates may be present near the top of the formation and along the eastern margin of the western region sea green and light green slates and cherts are more abundant and may predominant over red.

In the central region light green and sea green slates and cherts occur without red between Poultney and Pawlet formations. Farther east, red and sea green slates crop out associated with minor maroon slates. Thin quartzites occur near the base and a few thin silicic tuff bands were also observed. The overall pattern suggests increasing proportion of green to red, thinning, and loss of round grains on slaty cleavage surfaces to the east.

West of the New Boston Thrust the Mount Merino Formation is well developed and characterized by black and dark green cherts and inter-layered slates overlain by sooty black graptoliferous slates. No Mount Merino was observed in the central region. In the eastern region Mount Merino-like lithologies including both black cherts and black slates crop out sporadically near the Poultney-Pawlet contact, but nowhere constitute a mappable unit. The absence of Mount Merino lithologies to the east of the New Boston Thrust may reflect non-deposition, sporadic deposition, sub-Pawlet erosion, or tectonic excision along this contact.

The Pawlet Formation crops out extensively in both the western and eastern regions and sparsely in the central region. In the western region the Pawlet is somewhat variable, including medium grained lithic wackes that grade to silty wackes interlayered with dark slate and quite calcareous, better sorted, ungraded wackes interlayered with dark slate. Interbedded slates constitute roughly 40 to 50% of the section. In the central region coarse to fine lithic wackes occur. Pawlet greywackes and slates are best developed in the eastern region where they constitute a thick sequence of coarse to fine lithic wackes and dark slates. The proportion of wackes to slates varies from 40 to 60% depending on the section. To a first approximation maximum bed thickness and proportion of wacke to slate decrease from east to west. Also, and importantly, the Pawlet is in contact with Mount Merino, Indian River, and Poultney in the western, central, and eastern regions, respectively. This is a successive sequence and is the best evidence for a pre-Pawlet unconformity in this area. However, it is important to note again that,

(1) Although Mount Merino was not observed in the central region this may reflect thinning and poor outcrop characteristics of this unit.

(2) Indian River has been mapped in the eastern region. The upper contact of the Indian River was never observed so the stratigraphic sequence cannot be determined with certainty. In particular, it cannot be demonstrated to be an unconformity.

(3) The sporadic occurrence of Mount Merino-like lithologies near the Pawlet-Poultney contact may indicate local erosional remnants or thrust slivers. Since Indian River was never found associated with the Mount Merino lithologies it is difficult to constrain this interpretation further.

These observations indicate that a simple model of progressive down-cutting and erosion in pre-Pawlet time is not viable. However, it does seem likely that the base of the Pawlet does represent a disconformity. The nature of it and the tectono-sedimentary mechanisms responsible for it are as yet unclear.

The following generalizations appear to be warranted.

(1) The average grain size for all units except the Pawlet decrease from west to east.

(2) The percentage of carbonate decreases to the east. Once again the Pawlet may be an exception.

(3) The percentage of dark slate decreases to the east.

(4) The average bed thickness, for non-silty thin quartzites, decrease to the east. The Pawlet shows the opposite trend.

(5) The proportion of silt size and coarser clastic material relative to argillite decreases to the east.

If attention is focussed on the western and central regions, where the correlations are best and the section most complete, all of the above mentioned generalization are valid and completely compatible with increasing distance eastward from the westerly, continental shelf source. The same conclusion is indicated if all three regions are considered, but the absence of lower Poultney and Hatch Hill-West Castleton indicates that a simple model is inadequate.

CHAPTER 4

PETROGRAPHY

Jacobi (1977) studied the sedimentary petrography of Taconic rocks in detail. This petrographic analysis contributed to her understanding of sediment source and environment of deposition but did not provide an independent means of distinguishing different lithostratigraphic units. For this reason petrographic studies have been essentially restricted to Pawlet greywackes and a few samples of Browns Pond wackes. The Pawlet has been singled out because of its tectono-sedimentary importance, particularly with respect to source terrain. The Browns Pond wackes were studied since they provide a basis for comparison of sediment source and also it is important to know whether the two wackes are distinguishable.

The Browns Pond wacke usually occurs as thin beds or zones near the bottom of the Formation. It is commonly associated with a massive, vitreous, very well indurated quartzite, which led Zen (1961) to propose the name Mudd Pond facies for these lithologies. The wacke is characterized by very well rounded coarse sand-size quartz dispersed in a dark gray to black argillite matrix. Other components of the framework include minor plagioclase, lithic fragments, primarily dark slate, and rare zircons, tourmaline, and opaques. The matrix is composed of fine angular quartz, plagioclase, carbonate and clay. The framework grains usually float within the matrix with little contact between adjacent grains. This is markedly different from that observed in most Pawlet wackes and provides one means of distinguishing them. Another means of distinguishing between them is lack of variety of lithic fragments within the Browns Pond wacke when compared with the Pawlet. The composition of the Browns Pond wacke is completely compatible with being drived from the continental source to the west on the continental shelf.

Pawlet greywackes and slates are flysch-like sediments which are considered to have been derived from an easterly source (Zen, 1961). Evidence supporting an easterly derivation include the abundance of argillite as matrix and argillite clasts within the wackes, and temporal association with the 'initiation' of the Taconic orogeny. Coeval sediments to the west on the shelf are limestones and dark shales which are unlikely sources for the Pawlet greywackes. Paleocurrent data from the Pawlet within the Taconics is sparse and difficult to interpret.

Unrestored paleocurrent data collected during this study suggest north-south, axially directed transport of the sediments. Restoration of data to pre-fold orientation was not attempted due to the complexity of the fold history; however, the unrestored data are considered reasonable approximations because they approximately parallel the F_2 fold axes, which is a direction parallel to which large amounts of reorientation of linear structures is not expected, at least within the plane of the bedding. Axial transport is commonly observed in flysch sequences (Reading, 1978 and references therein), but does not help support the notion of an easterly source terrain.

Petrographic study of the Pawlet greywackes was undertaken with the hope of finding independent evidence for an easterly source. Eight thin sections were examined to determine general composition, but modal analyses were not done because the number of thin sections is probably not statistically significant, not all thin sections were made from the same interval of the beds, and previous work by Weber and Middleton

(1961) found significant operator-induced variability in their study of 'Normanskill' greywackes.

Pawlet greywackes are lithic wackes (in the usage of Pettijohn, 1976) and are primarily composed of quartz, plagioclase, potassium feldspar, micas, carbonate, lithic fragments, clays, heavy minerals and opaques. Quartz is the most abundant mineral phase, constituting 15-42% in Weber and Middleton's (1961) samples. It varies considerably in grain size, grain shape, and habit, both within and between thin sections. Grain size of quartz ranges from clay or fine silt to coarse sand. In general the coarser fraction tends to be better rounded, while the finer fraction is commonly quite angular. Quartz occurs in various habits, including single grains, polygranular aggregates, or clasts of fine grained, polycrystalline quartz. Single grains are usually clear, exhibiting undulatory extinction, and in the coarser grains are sometimes very well rounded. Fine needles of rutile (?) or sillimanite (?), and small subhedral to euhedral zircons sometimes occur as inclusions within the quartz. Polygranular aggregates are similar to the single grain occurrences except that they contain more than one grain. Within the aggregates the grains are usually anhedral and interlocking and are sometimes associated with minor white mica or clear carbonate (calcite). In a few cases the grains within the clasts exhibit strong shape preferred orientation not observed in adjacent clasts, suggesting pre-depositional deformation. These polygranular quartz clasts are considered to be derived from either deformed or undeformed vein quartz or in a few cases indurated quartzites. Much of this quartz probably has a protracted, poly-cyclic sedimentary history, and is probably derived from quartzites or other quartz-rich sediments in the Taconics, from which it has been suggested (Zen, 1967) these greywackes were at least in part derived.

Clasts of fine grained polycrystalline quartz are quite common in the Pawlet greywackes. It typically occurs as interlocking networks of quartz commonly associated with small rhombs of clear or brownish carbonate and sometimes a fine grained opaque phase. Most, if not all of this polycrystalline quartz is derived from chert, although some may be devitrified and recrystallized volcanic glass. The most likely source for chert is the erosion of Mount Merino or lower cherts from within the Taconic sequence. Volcanic glass shards or fragments could be derived from the volcanic arc farther to the east, which is also the source of bentonites observed on the shelf to the west (Brun and Chagnon, 1979 and references within).

Plagioclase is a common, but minor phase, constituting only a few percent of the greywackes by volume. It usually occurs as angular to subrounded grains which range from fine silt to medium sand, but are mostly silt or fine sand size. The plagioclase is commonly albite-twinned. It varies from fresh grains to cloudy grains partially altered to sericite and carbonate. Five fresh plagioclase composition determination from three thin sections utilizing the Michel-Levy method and extinction angle curves in Kerr (1959, third edition, p. 258) were made. Extinction angles ranged from 27° to 40°, corresponding to plagioclase compositions An_{48} to An_{70}. The majority of determinations fell in the range An_{55} to An_{60}, or in the middle of the labradorite field. The presence of labradorite suggests a mafic source terrain. The grain size and shape is compatible with a gabbroic or doleritic source, although an amphibolite source cannot be excluded. In general, most of the plagioclase in Adirondack anorthosites is more albitic (average An_{45}) (Buddington, 1969) so that this is an unlikely source for the plagioclase in these greywackes. This supports an easterly source.

Potassium feldspar is also present, but constitutes less than 1% of the rock by volume. It occurs as microcline, perthite, and as untwinned clear to cloudy, light gray grains. The grains are usually rounded and vary from silt to fine sand size. These grains are believed to be multi-cycle phases derived from the erosion of 'Taconic' sediments such as the Rensselaer grit.

White and greenish micas occur as clay to silt size detrital flakes within the greywackes. White, high birefringent mica, presumably muscovite is the most common variety of mica, and occurs as both detrital and secondary flakes. Greenish, pleochroic mica, presumably chlorite, although it could also be iron-rich muscovite is also present. Minor, brownish pleochroic biotite is present, but only as a detrital phase.

In hand specimen many of the greywackes are calcareous. In thin section, carbonate occurs as individual rounded grains, polycrystalline lithic fragments and commonly as part of the matrix. It is not always possible to distinguish between these different modes of occurrence, due to secondary recrystallization. Carbonate is either clear or brownish, and may occur as euhedral to subhedral rhombs. Twinning is quite common. Lithic fragments of carbonate are usually micrite to calcisiltite, which sometimes contain minor quartz.

Lithic fragments constitute a large proportion of these rocks. Shale and slate (?) fragments are most abundant and commonly indistinguishable from the argillite matrix. Wacke, silty quartzite (Poultney ?), limestone, chert, low grade pelitic metamorphic, and volcanic lithic fragments are also observed. Weber and Middleton (1961) report the

Figure 25

Photomicrographs of volcanic fragments within Pawlet wackes. a. Felt-textured rounded fragment (plane polarized light) (100 X magnification 1.46 mm field of view). b. a under crossed nichols. c. Dark, subrounded grain with phenocrysts of plagioclase (100 X magnification 1.46 mm field of view). d. albite-twinning in plagioclase phenocryst from grain shown in c. (250 X magnification; 0.59 mm field of view).

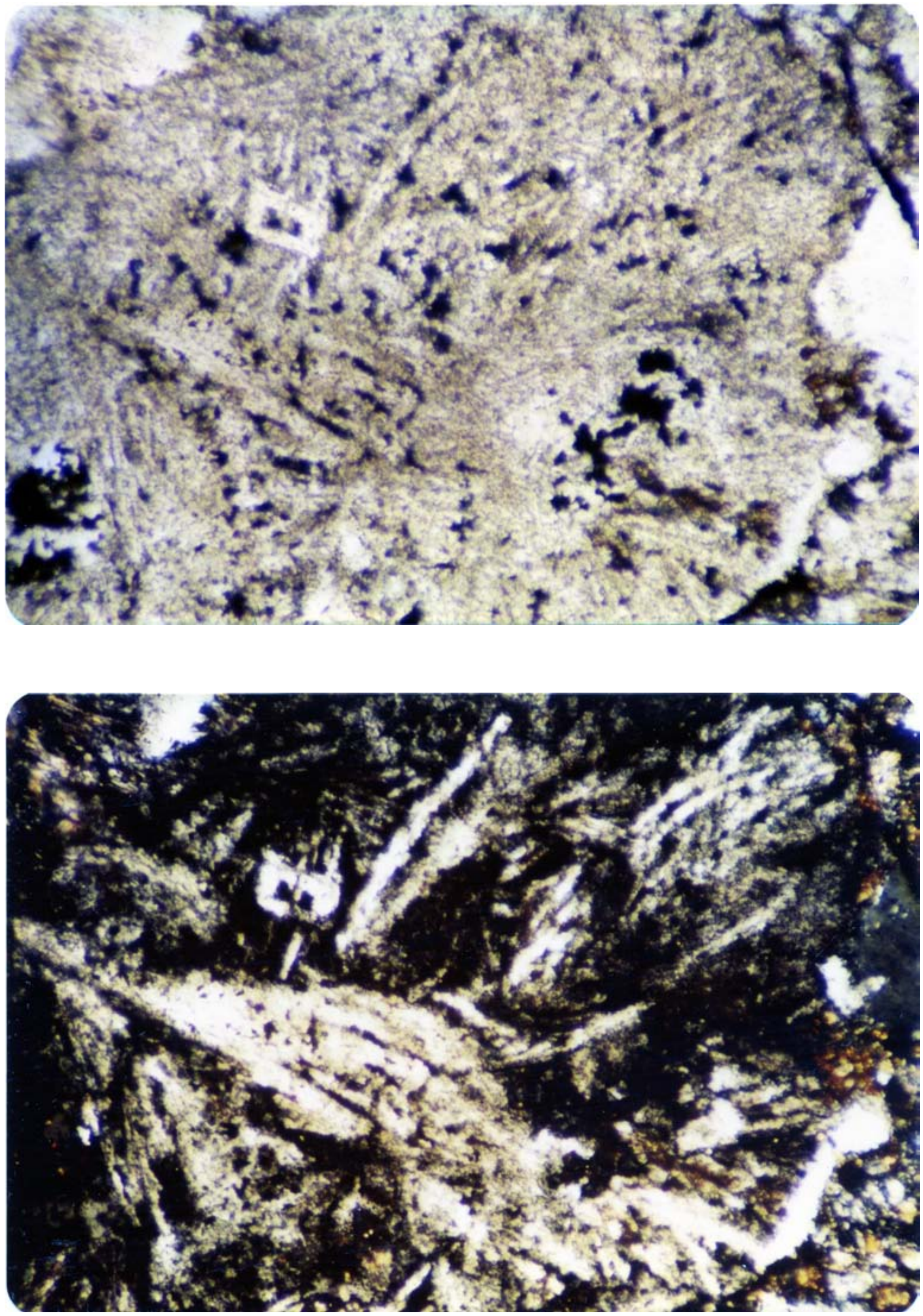
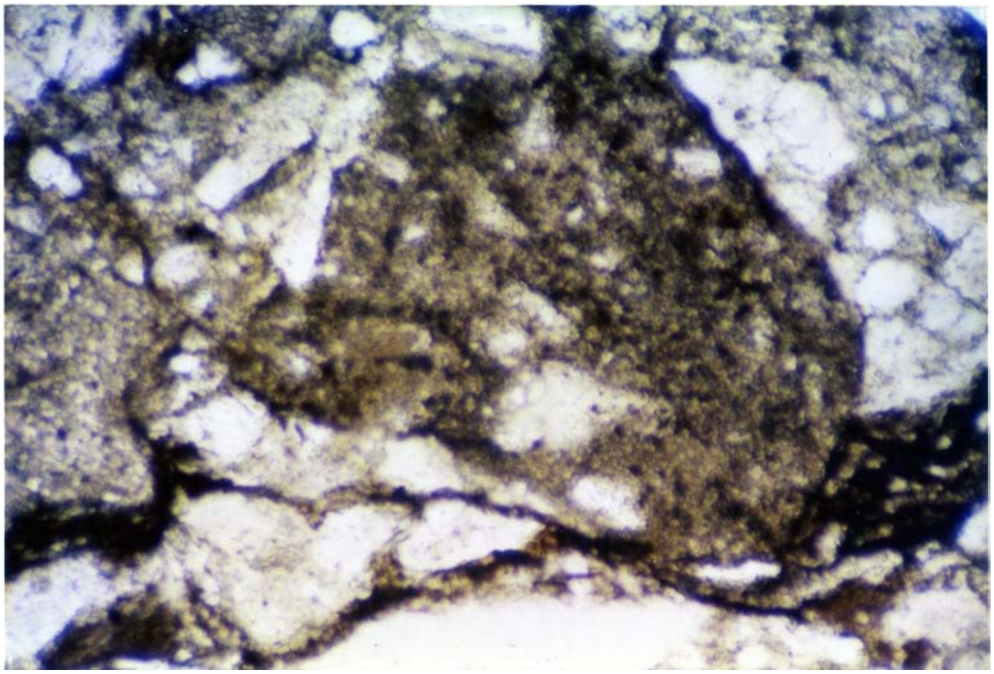


Figure 25: Photomicrographs of volcanic fragments within Pawlet wackes.
a. felt-textured rounded fragment (plane polarized light). 100x magnification, 1.46mm field of view. b. [same view] under crossed nicols.



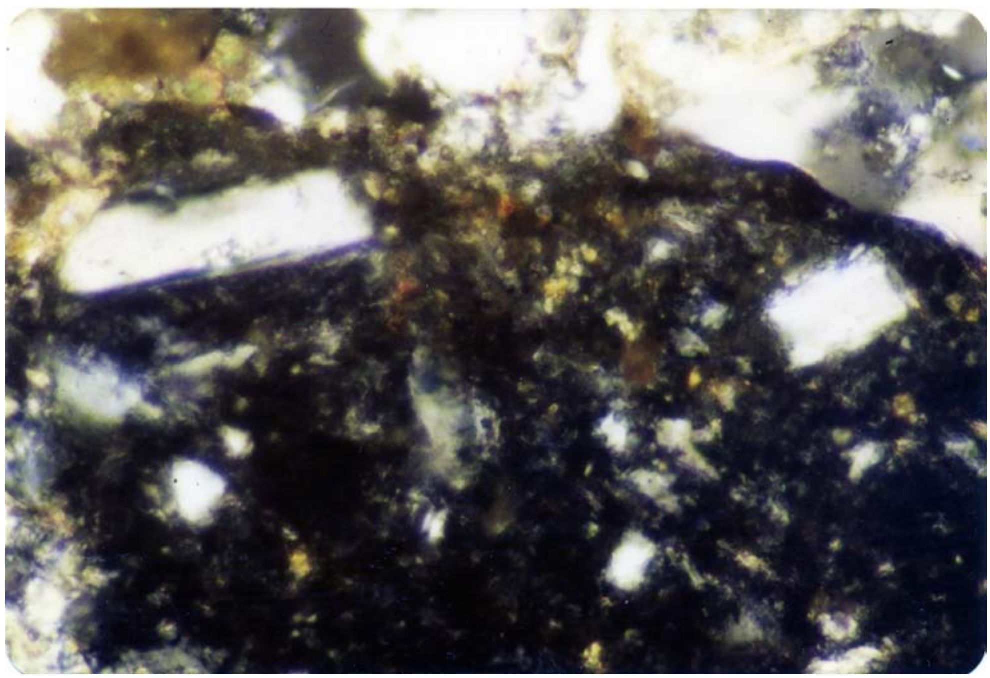


Figure 25 (cont): Photomicrographs of volcanic fragments within Pawlet wackes.
c. Dark, subrounded grain with phenocrysts of plagioclase (100x magnification, 1.46mm field of view). d. albite-twinning in plagioclase phenocryst from grain shown in c. (250x magnification; 0.59mm field of view).

following abundance of lithic types from a bed of 'Normanskill' greywacke: 40% shale and slate, 8% dark micrite, 8% sandstones and siltstones, 25% limestones, 13% siliceous metamorphics and 7% mafic to intermediate volcanic fragments. Their results agree in a general way with my observations, however, I have only been able to identify a few, probably mafic volcanic fragments (Figure 25) and the possible metamorphics are characterized by fine grained quartz and white mica with a phyllitic or fine grained schistose appearance, not "siliceous metamorphics". For the most part the cleavage within the shale and slate clasts is parallel to the 'slaty' cleavage of the matrix. However, in a few cases, a foliation defined by shape-preferred orientation of fine micas and quartz is observed and is crenulated by the regional slaty cleavage (Figure 27). Shumaker (1967) described similar observations. We (myself, W.D. Means, and W.G. Gregg) were unable to determine whether the slaty cleavage-like folation of the clasts is a well developed 'bedding plane fissility' or a slaty cleavage. There was no obvious evidence of recrystallization or metamorphism. It is unfortunate that a distinction could not be made, as the structural histories implied by them are significantly different.

The volcanic fragments are characterized by dark intermeshed, felted textures. The fragments are subrounded to rounded, fine sand size clasts. They are probably derived from mafic to intermediate volcanics, and supports the earlier suggestion of a mafic source to the east.

Zircons, tourmaline, and opaques constitute the heavy mineral suite of these greywackes. Zircon occurs as either rounded to subrounded detrital grains or as inclusions in quartz. Tourmaline occurs only as silt to fine sand size, rounded grains with a greenish to brownish ap-

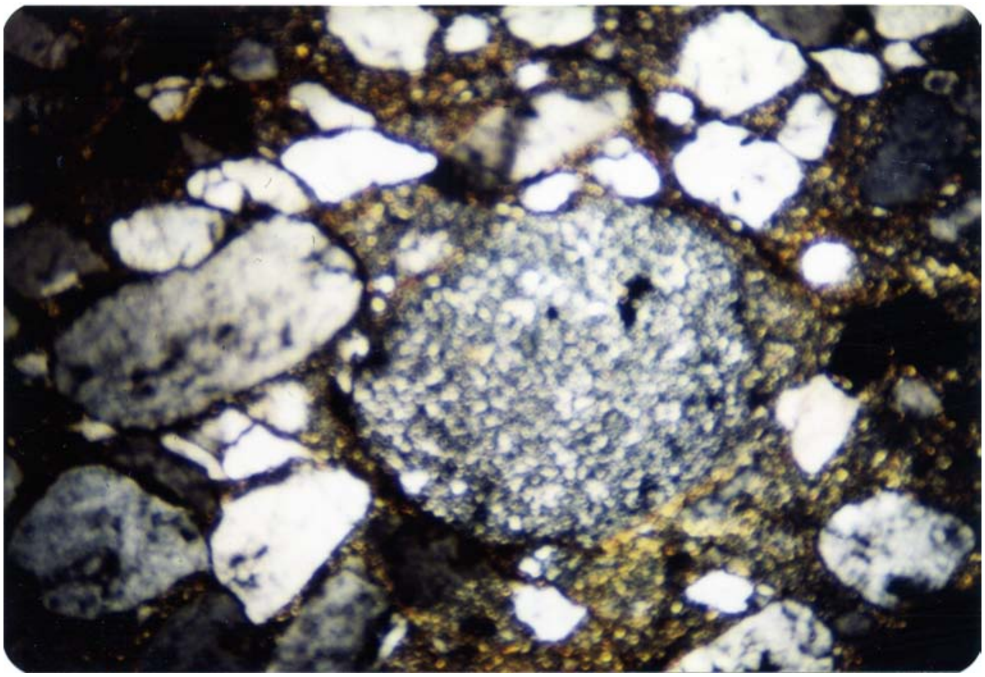


Figure 26: Photomicrograph of polycrystalline quartz grain within Pawlet wacke, surrounded by monocrystalline quartz grains (40x magnification, 3.64mm field of view).

Figure 27

Photomicrographs of argillite clasts in Pawlet wacke. Clasts contain a pre-depositional foliation, possibly a slaty cleavage that is crenulated by the 'regional' S_2 slaty cleavage. a. 40 X magnification, 3.64 mm field of view. b. 100 X magnification, 1.46 mm field of view.

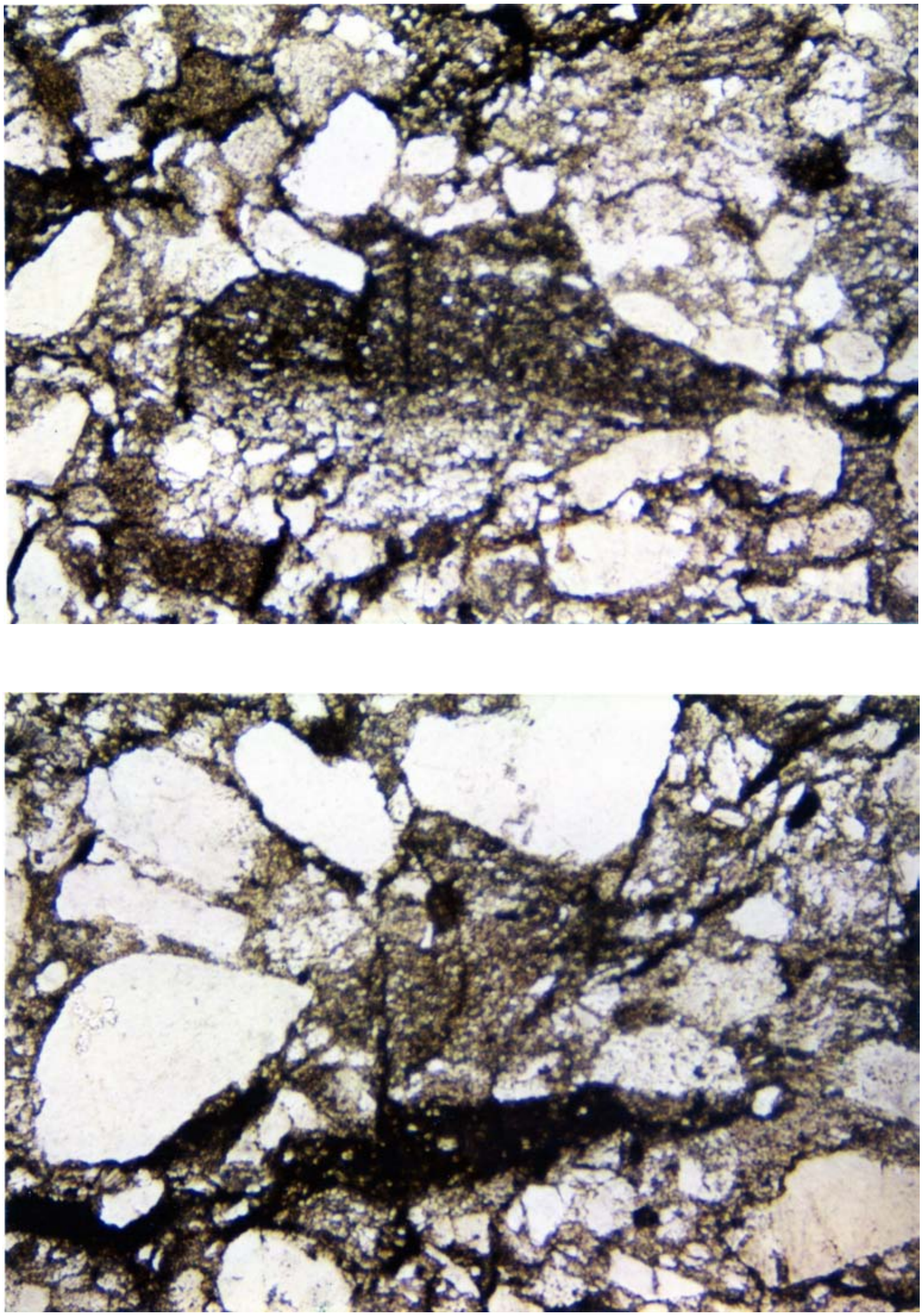


Figure 27: Photomicrographs of argillite clasts in Pawlet wacke. Clasts contain a pre-depositional foliation, possibly a slaty cleavage that is crenulated by the 'regional' S_2 slaty cleavage.
a. 40x magnification, 3.64mm field of view. b. 100x magnification, 1.46mm field of view.

pearance. Pyrite and titanomagnetite (?) with leucoxene alteration are the primary opaque phases. Opaque chrome-spinel or other opaque phase may be present but were indist inguishable from other opaque phases. Translucent, reddish brown chromite commonly reported from tectonically equivalent middle Ordovician flysh sequences, such as the Blow-me-down Brook Formation (Stevens, 1970), Tourelle Formation of the Gaspe peninsula (Hiscott, 1978) and Beaupre Formation of southern Quebec (Belt, et al, 1979) are not observed in sections from this area.

The composition of the Pawlet greywackes strongly supports an easterly derivation. Most of the greywacke materials can be derived from 'Taconic' sources, however, minor phases such as labradoritic plagioclase and mafic volcanics require an additional, but volumetrically less important source terrain. Two possible mafic igneous source terrains can be identified to the east. An ophiolitic source terrain can be postulated on the basis of the presence of serpentinized peridotites, gabbros, and amphibolites in the Connecticut River "synclinorium", which when traced to the north can be demonstrated to belong to a belt of ophiolites, including the Quebec ophiolites and farther north the Bay of Islands and Hare Bay ophiolites. Syn-obduction erosion of these ophiolites, as demonstrated by Casey and Kidd (1979) producing an angular unconformity on the ophiolites and shedding ophiolitic debris in front (Stevens, 1970; Hiscott, 1978) of the moving allochthon. If an ophiolite sheet similar to that observed to the north existed to the east of the Taconics it could have provided the necessary source for the mafic materials observed within the Pawlet greywackes.

An alternative source terrain for these mafic materials could be the Ammonusuc- Ascot-Weedon volcanics, which lie to the east of the

Connecticut River-Gaspe "synclinorium". These volcanics are Ordovician in age and presumably represent an island arc terrain on the east side of the closing ocean (Rowley et al, 1979).

CHAPTER 5

STRUCTURE

Introduction

The structural history of this area of the Taconics is quite complex and involves two phases of macroscopically important deformations and two later phases of deformation associated with only mesoscopic-scale structures. These 'tectonic' deformations are locally superimposed on an earlier, essentially syn-depositional, soft-sediment deformation. The rationale for this deformational scheme and the structures associated with different phases of deformation are the subject of this chapter. The chapter itself is divided into five parts. Part 1 discusses criteria used to distinguish different generations of structures. Part 2 describes the mesoscopic and associated microscopic structural elements of this area and attempts to associate them with different generations of structure. Data on the orientation of various structural elements is discussed in Part 3. Thrust faults are important structures in this area and are discussed separately in Part 4. Part 5 describes the macrostructure of the area and focusses on lateral variations of structure and sequence observed across the width of the 'Giddings Brook Slice'. These structural relationships are integrated in Chapter 6 into a tectonic model for the emplacement of the Taconic Allochthon.

Scale

The structural interpretation of this area, as in most, involves the integration of observations made on various scales. The terms macroscopic, mesoscopic, and microscopic are used extensively in this chapter and refer, respectively, to larger than a single outcrop, a single outcrop to hand specimen, and microscopic scales of observation.

Part 1: Generations of Structures: Distinguishing Characteristics

In regions of complex structures, where outcrop is discontinous and small compared to the scale of macroscopic structures being mapped, distinguishing structures belonging to different generations is often very difficult. In areas such as this, various criteria have been used to distinguish between different generations of structure, including folding of earlier folds, folding and overprinting of earlier axial surface foliations, nature of the axial surface foliations, fold style, or orientation of structural elements. These criteria are often used without apparent awareness of their reliability.

In the following section these criteria are briefly discussed based on the work of others in areas with better exposure and often in rocks with abundant primary younging evidence. This brief discussion provides a basis for assessment of the structural sequence worked out in this area.

Classically, fold style is one of the most commonly used criterion to distinguish folds of different generations in multiply-folded terrains. Fold style refers to the morphology of folds in profile. In many regions, particularly where structures are fairly simple this criterion appears to be useful. However, in at least some areas of complex, polyphase deformation, fold style has been demonstrated, using other criteria to distinguish folds of different generations, to be non-discriminatory (Means, 1963, 1966; Williams, 1970). Both Means and Williams found that, in their areas, folds of any one generation may vary extensively in style, including open to isoclinal, harmonic to disharmonic, or concentric to angular, to name a few of the morphologic characteristics commonly employed. Their findings suggest that style should be used only with caution.

The orientation of mesoscopic and microscopic structural elements might be useful in distinguishing different generations of structures. Some workers have utilized this approach, apparently with some success. However, in his study of the Bermagui area of southeast Australia, Williams (1970, 1971) found that different generations of structure are indistinguishable on the basis of orientation of bedding, axial surface foliations, fold hinge orientations, or any combination of these data. This finding suggests that caution must be exercised when using this criterion.

In the Taconics, as in many other areas the type or nature of the axial surface foliation is a commonly employed criterion to distinguish structure of different generations. Thus folds with slaty cleavage are of one generation, whereas those with some other type of foliation belong to a different generation (Wright, 1970; Zen, 1964, 1972). Hoeppener (1956, 1960) and Williams (1970, 1971, 1972a) among others have demonstrated that axial surface foliations may vary considerably within a single fold generation.

Williams (1971, 1972a) found slaty cleavage, crenulation cleavage, and differentiated layering, axial surface to a single generation of folds, and also that considerable overlap of types of axial surface foliations existed between folds of different generations. Williams also found that foliation of different fold generations may be microstructurally indistinguishable. Without evidence to the contrary, this criterion should also be considered to be not necessarily discriminatory.

From his detailed mapping, Williams (1970, 1971) found that refolding of folds provided the best criterion for distinguishing different generations of folds; but this method also may have some pitfalls. For example, in areas where two or more generations of folds are refolded by a regionally prominent generation of generally tight folds, it may be difficult to demonstrate more than two generations of folds, unless one is fortunate and finds all generations in a single outcrop. Since one is commonly not so fortunate, many workers extend the refolding criterion to include folding of an axial surface foliation. Because not all foliations develop axial surface to folds and not all generations of folds are associated with axial surface foliations, this method may sometimes be unreliable. Williams (1971, 1972a) describes one or possibly two generations of bedding-parallel differentiated layering which are folded about his first and second generation folds. This layering is mesoscopically and microscopically similar to the axial surface foliations except that it is everywhere parallel to bedding. In areas where facing evidence is less abundant, this differentiated layering might be mistaken for an earlier axial surface foliation. A similar situation can exist where diagenetically produced structures, such as bedding plane partings in shales are well developed, and subsequently folded. These bedding-parallel foliations may be mistaken for an earlier axial surface foliation, and may be assumed to have resulted from an early phase of isoclinal folding. These diagenetic foliations may be characterized by microstructural fabrics that are indistinguishable from slaty cleavage (Means, personal communication, 1979), further complicating matters.

This brief summary of criteria commonly employed to distinguish folds of different generations paints a bleak picture for mappers of complexly folded terrains characterized by only small, discontinuous exposures and sparse younging evidence. The structural analysis of such regions should be done cautiously and only after very detailed field observations. On the basis of the field work completed during this study the following four phase structural sequence can be demonstrated using the refolded fold criterion (Williams, 1970, 1971) extended to include folding of earlier, presumed axial surface foliations. D_1 is associated with early, tight to isoclinal and possibly recumbent, west-facing folds (F_1). This generation of folds is demonstrated by stratigraphic and structural (bedding-slaty cleavage (S_2)) relations indicating both upward- and downward-facing F_2 folds in the eastern part of the map area. These folds do not appear to be associated with an axial surface foliation. The most prominent macroscopic and mesoscopic folds in this area and the adjacent area to the west (Jacobi, 1977) are west-verging and usually west-facing, overturned folds with an essentially penetrative axial surface slaty cleavage (S_2). F_2 folds and S_2 cleavage are refolded by upright to west-verging open to barely tight F_3 folds. A spaced crenulation cleavage (S_3) is commonly developed axial surface to these F_3 folds. Finally, the crenulation cleavage and all earlier foliations are folded by upright kink bands, here referred to as F_4. Only D_2 related structures are penetratively developed in this area.

The separation of structures belonging to different phases of deformation shown on Plate 1, and described in this chapter relies on the assumption that once the structural sequence has been determined using the refolded fold criterion that the type of axial surface foliation, and to a lesser extent style and orientation provide a means of distinguishing structural elements of different generations in this area, where refolding can not be demonstrated on a local (mesoscopic or macroscopic) basis. In other words, it is assumed that, for example, slaty cleavage is everywhere the same 'age', unless local relations demonstrate otherwise; which occurs only rarely in this area. Without this assumption it would be almost impossible to draw any conclusions concerning the macroscopic structure of this area.

Part 2: Mesoscopic and Microscopic Structural Elements of the Study Area

The discussion of mesoscopic and microscopic structural elements observed in the study area is divided into six sections, based on association with the four phases of regional deformation and earlier syn-depositional structures. Separate sections are devoted to mesoscopic and microscopic structures associated with thrust faults and veins.

Syn-depositional to penecontemporaneous structures

Bedding and other primary features, such as grading, bottom structures, internal laminations, and bioturbation features are often indisputably recognized, but younging criteria are almost entirely restricted to Pawlet greywackes (Figure 28), and even then are not ubiquitous. Bedding may be delineated by grain size variations or compositional layering, including color laminations (Figure 29). Bedding thickness, continuity, and prominence vary considerably, and is primarily dependent upon position within the lithostratigraphic sequence. Bedding may be confused with secondary color changes (Dale, 1899; Jacobi, 1977),

Figure 28

Graded bedding in a bed of Pawlet greywacke. Bedding is overturned and less steeply dipping than slaty cleavage (S_2) indicating an antiformal syncline to the right (west).

Figure 29

Color laminated slate. Color laminations interpreted to represent bedding.

Figure 28: Graded bedding in a bed of Pawlet greywacke. Bedding is overturned and less steeply dipping than slaty cleavage (S_2) indicating an antiformal syncline to the right (west).

Figure 29: Color laminated slate. Color laminations interpreted to represent bedding.

secondary layering (Dale, 1899; Zen, 1961), or locally transposed layering (Jacobi, 1977).

Secondary color changes are mostly observed in purple and red slates due to reduction of ferric to ferrous iron shown by change to a green color. Secondary color layering is best observed in slate quarries, and may be spatially associated with joints, veins or other fracture surfaces. Reduction spots are also a manifestation of this, and are generally assumed to have been produced prior to deformation (Wood, 1973).

<u>D_0-Deformation:</u> <u>Penecontemporaneous, soft-sediment deformation</u>

Syn-depositional slumping (D_0) is observed in many units, and is particularly well developed in the Bomoseen, Browns Pond, and Poultney Formations (Rowley et al, 1979). D_0 deformation usually involves dismemberment or folding of one or several beds. Slump folds are often demonstrably intrafolial and have a disharmonic character. Slump folds vary from open to tight or isoclinal and are locally associated with transposition. Individual syn-depositional slump structures do not involve more than a few beds, and never influence the macroscopic distribution of litho-units, with the possible exception of the Hatch Hill-West Castleton from the eastern parts of the study area (see Chapter 3). It is important to note that all D_0 structures have been significantly modified by later tectonic deformations. There are no unambiguous criteria for distinguishing D_0 structures from D_1 or some D_2 structures. In this area, the only criterion used was the presence of intrafolial folds in sequences otherwise not mesoscopically folded (Figure 30). This criterion is not always reliable, since outcrops may in some cases be small enough to mask larger folds, particularly of D_1 generation.

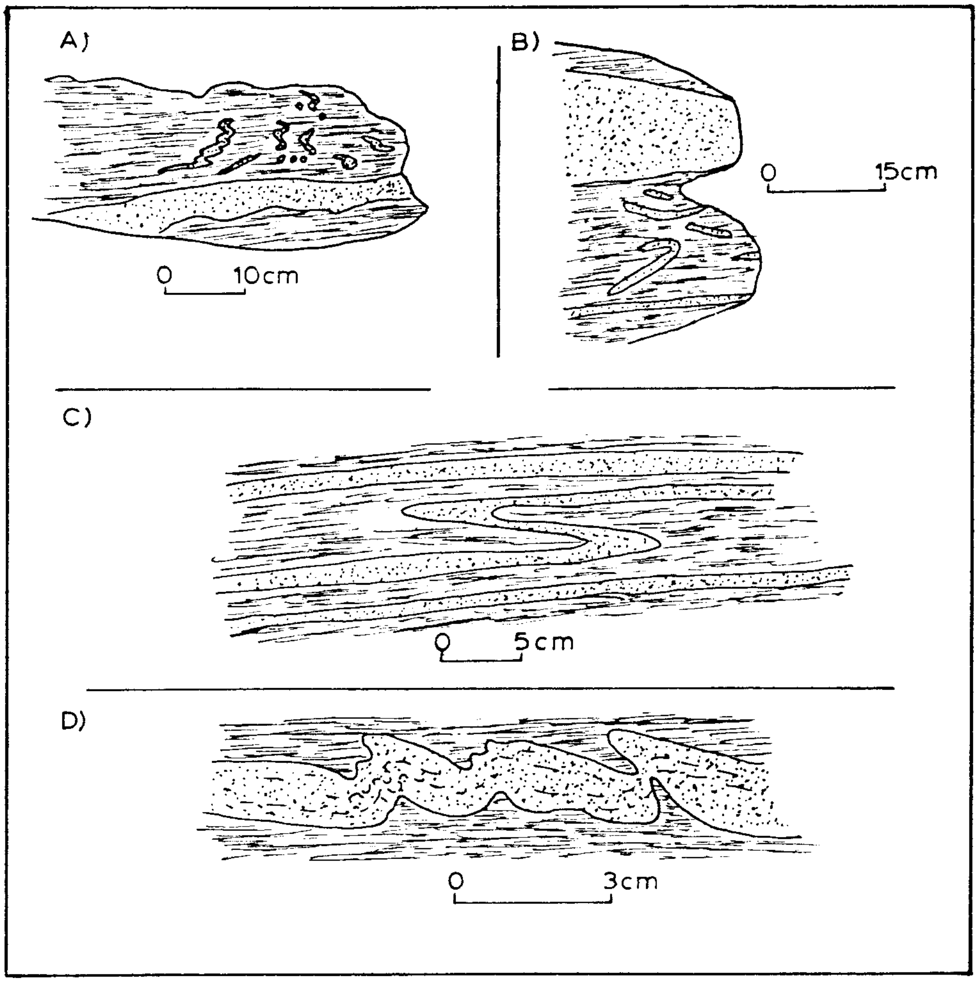


Figure 30

Outcrop sketches of structures interpreted to result from pene-contemporaneous slump folding.

Wright (1970) suggested that detached fold hinges, folds with an axial surface foliation lacking evidence of recrystallization, or folds where both limbs are transected by slaty cleavage are all evidence of soft-sediment deformation. However, at least in this area, slaty cleavage appears to be associated with D_2, and D_1 is not considered to be penecontemporaneous with sedimentation, so that all of the features used by Wright (1970) could be ascribed to D_1 just as well, if not better than D_0.

Tectonic Deformations

Four phases of tectonic deformation are recognized in this area. The earliest two phases are responsible for the map pattern, while the later deformations are only mesoscopically and microscopically recognizable. However, it is the intermediate phases D_2 and D_3 that are responsible for the regionally recognizable tectonic fabrics, slaty cleavage and crenulation cleavage.

D_1 Deformation

Mesoscopic evidence for D_1 deformation consists of a small number of outcrops in which already inverted bedding in the Pawlet has been refolded by F_2 giving rise to downward-facing F_2 fold structures (Figure 28). Mesoscopic folds and related axial surface foliations of the D_1 generation are not demonstrable in outcrop. At a number of localities within the Poultney Formation, a well developed bedding parallel foliation is observed and is folded by F_2 (Figure 31). This foliation might be associated with the F_1 phase of folding, but it is equally well interpreted as a bedding plane parting. Unfortunately, the possible regional significance of the D_1 generation of structures was not recognized until late in the investigation so that little attention was paid to this foliation, both mesoscopically and microscopically. In a single thin section (Figure 32) from the hinge of an F_2 fold in which this foliation may be observed on the limbs, this foliation is not discernible, and only the slaty cleavage is seen. Since most other evidence for this phase of deformation derives from macroscopic relations, further discussion is deferred until part 5 of this chapter.

D_2 Deformation

D_2 structures are characterized by a prominent, usually well developed axial surface slaty cleavage. This slaty cleavage is axial surface to F_2 folds of F_1 and the F_1 axial surface. Folds of slaty cleavage are observed, but these folds are either associated with a spaced crenulation cleavage or lack a cleavage altogether. In the area adjacent to the study area on the west, slaty cleavage is axial surface to folds with similar morphology, amplitude, wavelength, and general orientation (Jacobi, 1977; Rowley, et al, 1979). There, no convincing evidence of an earlier deformation is observed, and structures post-dating the folds and slaty cleavage are rare. Thus on a regional scale, slaty cleavage is demonstrably associated with a major early phase of deformation. These observations support the use of slaty cleavage as a distinguishing characteristic of the D_2 generation of structures in this study area.

F_2 Folds

F_2 folds affect bedding and associated bedding-plane fissility, some veins and, possibly, an earlier axial surface foliation (Figure 31). F_2 folds are most conspicuously developed in interlayered thin silty quartzites and argillites of the Poultney and Bullfrog Hollow litho-

Figure 31

Bedding parallel foliations folded by F_2 folds. Foliation is cross-cut by 'regional' slaty cleavage in the hinge of an F_2 fold.

Figure 32

Thin section from hinge of an F_2 fold with a mesoscopically visible bedding-parallel foliation. This foliation is not observed, and only the slaty cleavage developed axial surface to the fold is observed.

Figure 31: Bedding parallel foliations folded by F_2 folds. Foliation is cross-cut by 'regional' slaty cleavage in the hinge of an F_2 fold.

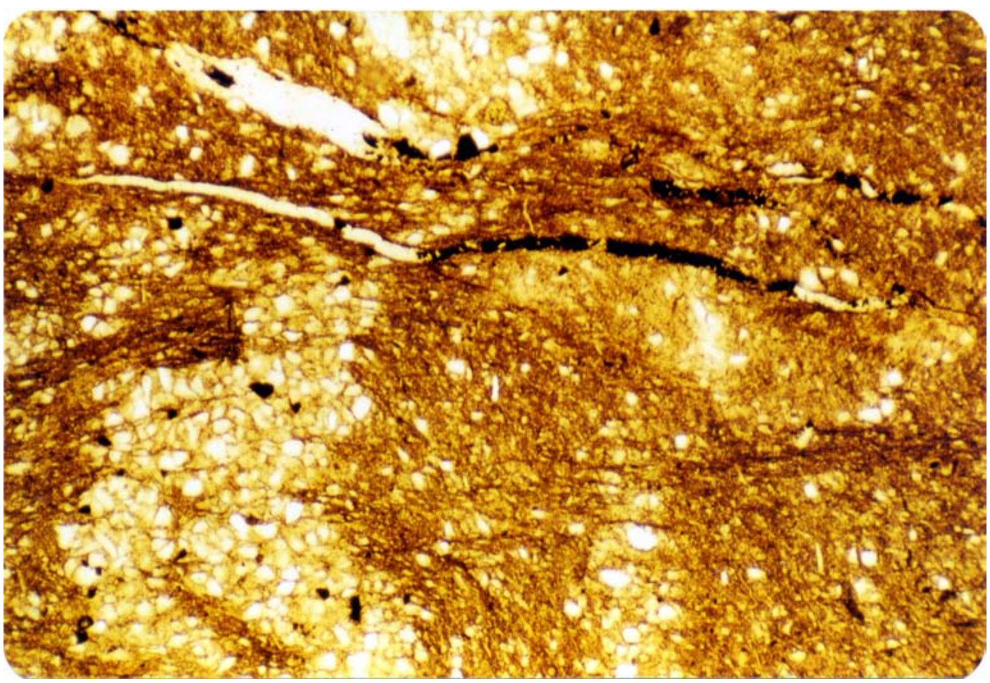


Figure 32: Thin section from hinge of an F_2 fold with a mesoscopically visible bedding-parallel foliation. This foliation is not observed, and only the slaty cleavage developed axial surface to the fold is observed.

facies 'A'. F_2 folds vary considerably in profile morphology, more commonly referred to as style. This variation is dependent in part on their relationship to larger scale folds, layer thickness and thickness variation, and layering interval. In general, F_2 folds are tight, but rarely precisely isoclinal. Interlimb angles are generally between 15° and 40°, but may locally exceed 60° (Figure 34). F_2 folds may be harmonic or disharmonic (Figure 34).

Folds in predominantly thicker-bedded (greater than 5-10 cm) sequences with little interbedded argillite tend to approach cylindrical folds, whereas folds developed predominantly in argillite tend to approximate similar folds (Figure 34). At several localities polyclinal (Turner and Weiss, 1963) folds were observed (Figure 35). Cleavage in the argillite generally forms divergent fans (Figure 36) (Ramsay, 1967), while cleavage in coarser lithologies, particularly wackes, forms convergent fans (Figure 36).

S_2 Cleavage

The most prominent mesoscopic and microscopic fabric element in this area is the almost ubiquitous axial surface S_2 cleavage. The parallelism of this cleavage with the axial surface of F_2 folds in profile is locally well demonstrated (Figure 37). This parallelism is assumed to be maintained elsewhere in the study area. This cleavage varies continuously from an extremely well developed, penetrative, slaty cleavage to a spaced cleavage and, locally, to a fracture cleavage. The degree of development is primarily dependent upon the grain size and proportion of pelitic component in the cleaved rock. Uniformly fine-grained argillites of the Indian River, Mettawee, Truthville, and Bullfrog Hollow lithofacies 'B' are penetratively cleaved (Figure 9),

Figure 33

Photographs of F_2 folds showing variations in style.

Figure 33: Photographs of F_2 folds showing variation in style.

Figure 34

Outcrop sketches of F_2 folds. B. shows divergence of cleavage in the argillites. C. shows truncation and overprinting of earlier foliation by an axial surface parallel foliation. Cleavage in the arenite is convergent, while cross-cutting cleavage in the argillite is divergent. E. cleavage is spaced on a mesoscopic scale in a silty argillite. H through J show details of form surfaces of folds. H. shows disharmonic folds.

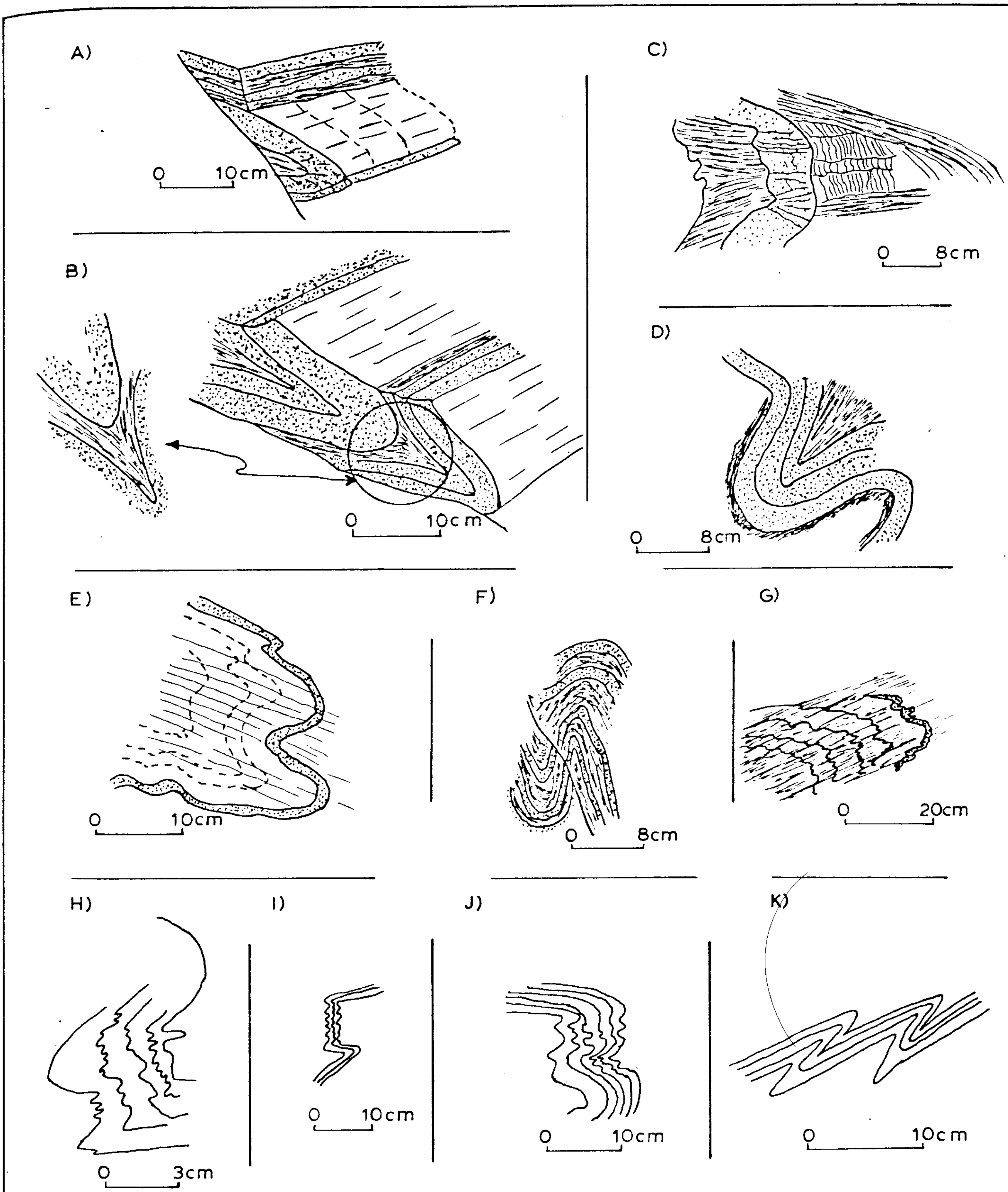

A)
0
10cm
B)
0
10cm
C)
0
8cm
D)
0
8cm
E)
0
10cm
F)
0
8cm
G)
0
20cm
H)
0
3cm
I)
0
10cm
J)
0
10cm
K)
0
10cm

Figure 35

Polyclinally folded thin, silty, Poultney quartzites. a. Photograph (lens cap is 50 mm across). b. Outcrop sketch.

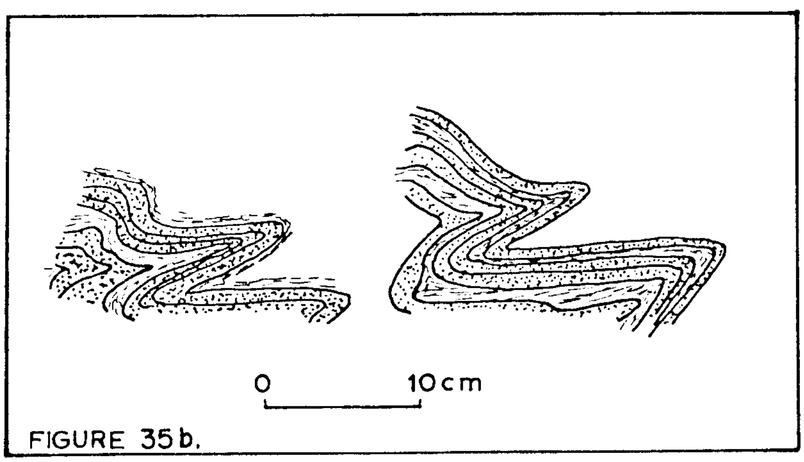


Figure 35: Polyclinally folded thin, silty, Poultney quartzites. a. Photograph (lens cap is 50mm across). b. Outcrop sketch.

Figure 36

Outcrop sketch of upward-facing F_2 fold of Pawlet greywackes. Cleavage in the argillites is divergent, whereas that in the wackes is convergent.

Figure 37

Photograph of tight to isoclinal fold in quarry of Indian River slates from Jacobi's area demonstrating, at least two-dimensionally that cleavage is axial surface to tolds.

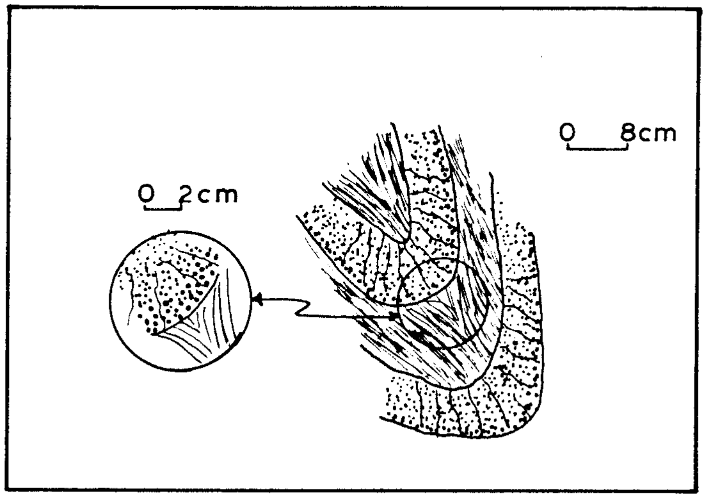


Figure 36: Outcrop sketch of upward-facing F_2 fold of Pawlet greywackes. Cleavage in the argillites is divergent, whereas that in the wackes is convergent.

Figure 37: Photograph (by W.S.F. Kidd) of tight to isoclinal fold in quarry of Indian River slates from Jacobi's area demonstrating, at least two-dimensionally that cleavage is axial surface to folds. [Bruce Idleman for scale]

and are commonly quarried for roofing slate or tile. Coarser argillites and some greywackes are well cleaved, but are not suitable for quarrying. Relatively clean silty to medium-grained quartzites are poorly cleaved or uncleaved.

The S_2 cleavage is generally associated with a prominent plane of parting. Bedding and locally S_3 may also be associated with planes of parting. Where these planes are oblique to one another, their intersection may produce a pencilling, color or other bedding-cleavage intersection lineation (Figure 38), or crenulation lineation. However, the general tightness of F_2 folds means that bedding and the S_2 cleavage are commonly parallel, at least mesoscopically.

In fine argillites the slaty cleavage is defined by a homogeneously developed preferred orientation of fine micas and clays, and less common fine silt-size white or light green micas (See figure 5.8b, p. 224, Hobbs et al, 1976). In siltier argillites the slaty cleavage is commonly defined by compositionally-defined microlithon domains. One is primarily composed of fine micas, mica films, and elongate detrital grains with well developed preferred orientation and commonly a corroded appearance (Figure 39). Dark films are also commonly associated with this type of domain. A second domain is commonly richer in quartz and other clastics, and any mica present is often less well oriented. Quartz in this second type of domain may or may not have a corroded appearance. Mica, quartz and less common carbonate beards are observed and usually help define the slaty cleavage. In most cases the beards parallel the slaty cleavage, but in one section beards are markedly curved, from initial perpendicularity to the host grain (Williams, 1972b) into parallelism with the cleavage (Figure 40). This relationship suggests relative

Figure 38: Well-developed bedding-cleavage lineation.

Figure 39

Outline of quartz grains. (A) Detrital grains; (B) Corroded grains. Redrawn from Williams (1972a, Figure 5, p. 10.).

Figure 40

Fibrous calcite forming a 'pressure shadow' on pyrite. Note curvature of fibers. (100 X magnification; 1.46 mm field of view)

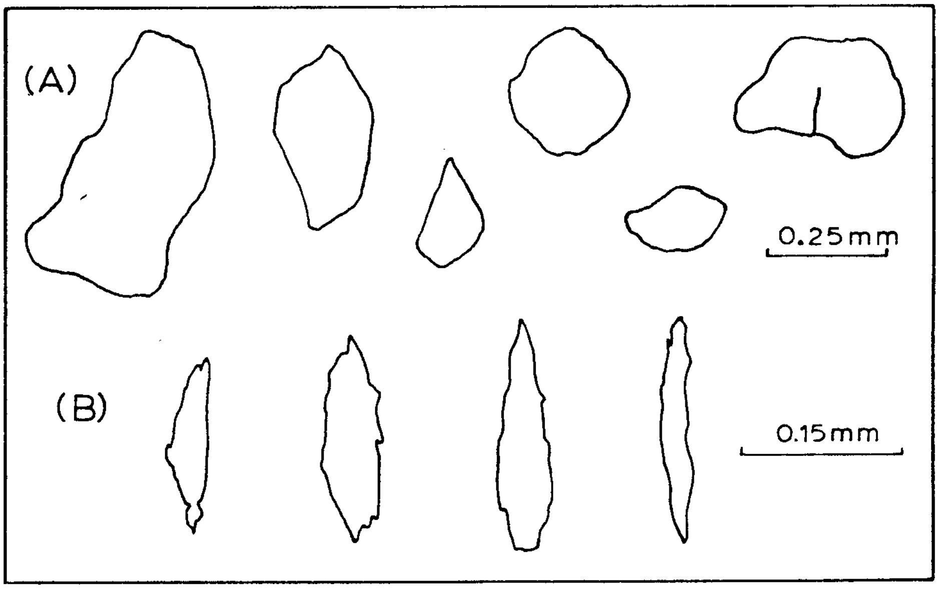


Figure 39: Outline of quartz grains. (A) Detrital grains; (B) Corroded grains. Redrawn from Williams (1972a, Figure 5, p. 10).

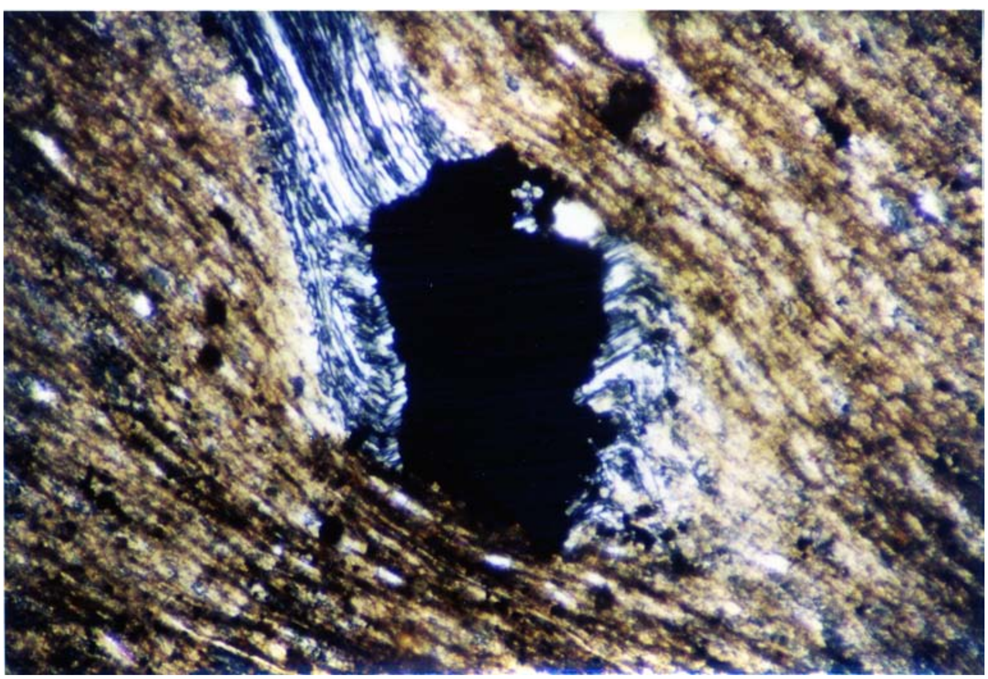


Figure 40: Fibrous calcite forming a 'pressure shadow' on pyrite. Note curvature of fibers. (100x magnification; 1.46mm field of view).

rotation between the host grain and slaty cleavage, during cleavage development.

In coarser sediments such as wackes, the S_2 foliation is typically defined by anastamosing and interweaving of fine mica films, sometimes quite dark, and preferred orientation of clays and clastic grains, which comprise one domain. The other domain may have an undeformed appearance, in which rounded or otherwise detrital-looking grains are preserved (Figure 41). In others the cleavage may be morphologically similar to the slaty cleavage in a silty argillite, but developed on a somewhat coarser scale. The most conspicuous aspect of the foliation in these cases is the shape-preferred orientation of quartz, feldspar, and lithic fragments parallel to the cleavage. Quartz and mica beards are commonly well developed as well.

Ellipsoidal reduction spots (Figure 42) and grain are other mesoscopic structural elements associated with the D_2 slaty cleavage. Grain is a secondary plane of weakness in many slates; best developed in roofing slates in this area. This plane of weakness is developed perpendicular to the slaty cleavage and commonly intersects the cleavage in a down-dip direction. Wright (1970) suggests that the grain results from a secondary preferred orientation of cross-micas, but other explanations have been given in other areas. No attempt was made to study this feature.

Ellipsoidal reduction spots are observed in several quarries of purple slate (Bullfrog Hollow lithofacies 'B') in this area, but were not studied during this project. Wood (1973, 1974), however, utilized reduction spots in similar slates from an area approximately 15 kilometers to the north of the study area, to determine finite or total

Figure 41

Difference in appearance of quartz grains. Some are distinctly detrital, others appear corroded. Two cleavages present, an early cleavage approximately parallel to compositional layering and a later crenulation cleavage oblique (vertical) to the early cleavage. 40 X magnification, 3.64 mm field of view.

Figure 42

Ellipsoidal reduction spot on cleavage surface.

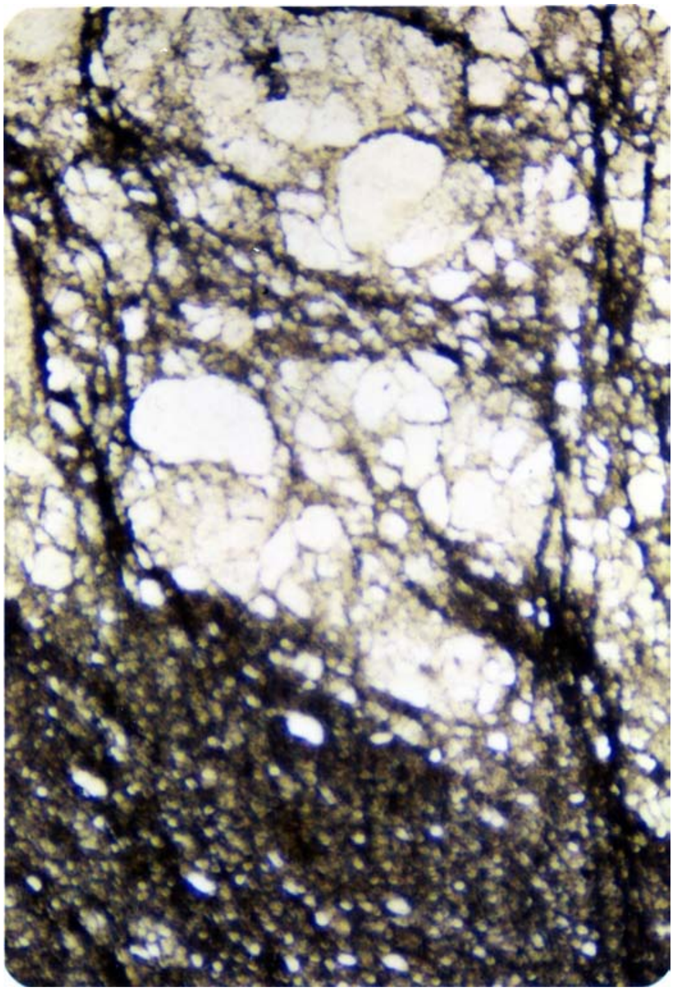


Figure 41: Difference in appearance of quartz grains. Some are distinctly detrital, others appear corroded. Two cleavages are present, an early cleavage approximately parallel to compositional layering and a later crenulation cleavage oblique (vertical) to the early cleavage. 40x magnification, 3.64mm field of view.

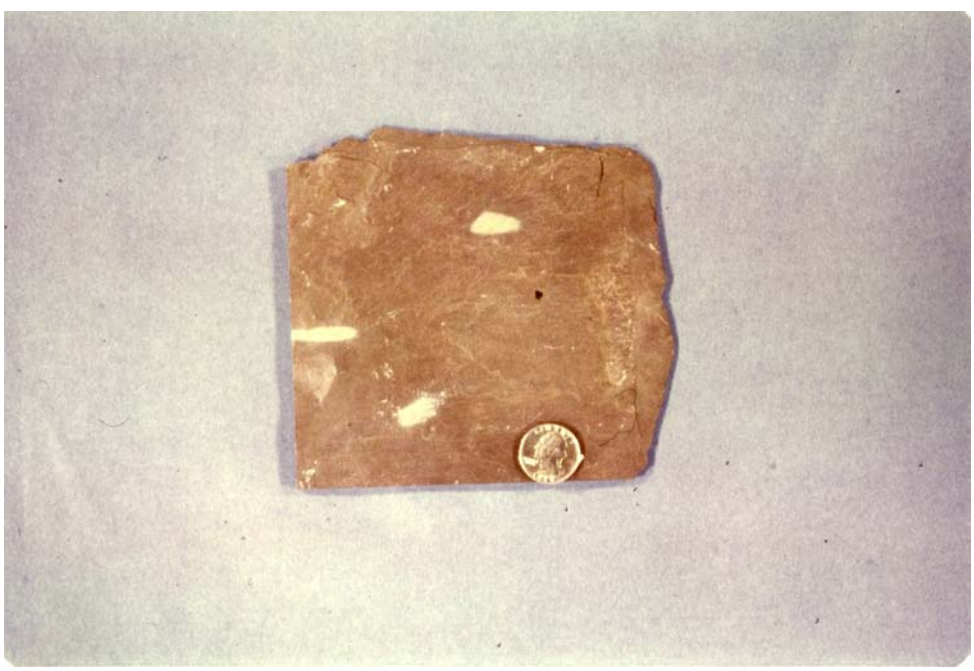


Figure 42: Ellipsoidal reduction spot on cleavage surface.

(Means, 1975, 1976) strain associated with cleavage development. His analysis indicates that the slates have enjoyed approximately 75% shortening perpendicular to slaty cleavage (Z or λ_3 direction) and, simultaneously, approximately 150% and 50% extension approximately perpendicular and parallel to F_2 hinge lines (X or λ_1 and Y or λ_2 directions, respectively) within the slaty cleavage plane. Slaty cleavage was found to lie precisely coincident with the plane of maximum total shortening. [See Hobbs et al, 1976 and Williams (1976) for discussions of parallelism of foliations to principle planes of the total strain ellipsoid.] The cleavage grain was found to lie parallel to the direction of maximum total elongation.

In order to calculate the total strain associated with these reduction spots the initial shape of the reduction spot must be assumed to have been essentially spherical just prior to slaty cleavage development. Wood has made this assumption. He attempts to insure a uniform shape by studying only those reduction spots that have smooth, elliptical cross sections. If the reduction spots, just prior to cleavage formation were markedly non-spherical, but instead ellipsoidal, reflecting an earlier phase of folding and flattening, then the total strains calculated by Wood (1973, 1974) cannot be entirely ascribed to slaty cleavage development. If this were true, as might be indicated from the deformational sequence in this area, the strains calculated by Wood could either be too high or too low depending upon the shape and orientation of the reduction spot ellipsoid prior to slaty cleavage development. Unfortunately, the regional extent of the D_1 deformation is unknown, and therefore the implication of this earlier phase of deformation to Wood's studies is unknown, but it could be substantial.

An important question to be considered in rocks possessing a slaty cleavage is; What was the nature of the rocks at the time of cleavage formation? Many reviews of slaty cleavage formation have been published, including Wood (1974), Williams (1977), Hobbs et al (1976) among others. The general consensus among reviewers is that although dewatering of unconsolidated sediments (Maxwell, 1962; Altermen, 1973) may play a role in slaty cleavage development, it is by no means the most important mechanism. Further discussion of slaty cleavage formation is beyond the scope of this thesis and only a few observations pertaining to this question are described below.

In several places in the study area, at least one generation of veins are observed to be deformed, most commonly folded, with slaty cleavage axial surface to these folds (Figure 43). Win Means also pointed out several locations in Jacobi's area where fibrous veins had developed; the fibers tended to be oriented in a down-dip direction, and the veins were truncated by cleavage surfaces (Figure 44). These observations may indicate that cleavage formation occurred in rocks sufficiently consolidated to be veined, and accord well with the work of Wood (1974), Buetner et al (1977), Williams (1976, 1977), Gregg (1979 and 1979b) and many others. Their work indicates that slaty cleavage developed in partly to wholly lithified rocks undergoing deformation, and not by 'tectonic dewatering' of unconsolidated to semi-consolidated sediments.

D_3 Deformation

F_2 folds, characterized by an axial surface slaty cleavage are locally observed to be refolded (Figure 45) and the axial surface slaty

Figure 43

Isoclinally folded vein quartz with slaty cleavage axial surface to folds.

Figure 44

Deformed fibrous vein quartz. Deformation may be associated with late stages of slaty cleavage development.

Figure 43: Isoclinally folded vein quartz with slaty cleavage axial surface to folds.

Figure 44: Deformed fibrous vein quartz. Deformation may be associated with late stages of slaty cleavage development.

cleavage is overprinted by a spaced, axial surface crenulation cleavage. The later generation of folds is designated as F_3. The axial surface crenulation cleavage is designated S_3. Commonly, earlier folds are not observed to be refolded, but the S_2 slaty cleavage is folded and often overprinted by a crenulation cleavage. These folds are also assigned to F_3.

$\underline{F_3 \text{ Folds}}$

F_3 folds are not commonly observed in this area. They vary from essentially upright to inclined asymmetrical, and locally overturned west-verging folds (Figure 46). F_3 axial surfaces are generally steeply east-dipping and are associated with gently plunging fold axes. Fold style ranges from open to less commonly tight, with either rounded, concentric or angular hinges. Chevron folds were not observed. Interlimb angles range from 50° to almost 180°, but are most commonly 100° to 120°. F_3 folds have low amplitudes (less than 1-2 m, and mostly less than 0.5 m) and short wavelengths (generally less than 1-3 m), when compared to earlier generation of folds in this area.

Where folds are open and the axial surface is defined by a steeply east-dipping crenulation cleavage, F_3 folds are easily distinguishable from F_2 and older mesoscopic folds. Where the axial surface S_3 cleavage mesoscopically approaches a slaty cleavage, the distinction is more difficult. In these cases, microscopic examination of the axial surface foliation usually allows them to be distinguished.

$\underline{S_3 \text{ Foliation}}$

The F_3 axial surface foliation (S_3) varies from a spaced crenulation cleavage to a spaced, 'differentiated' crenulation cleavage and

Figure 45

Photograph of slab section of refolded fold. An F_2 tight to isoclinal fold with slaty cleavage axial surface to fold is openly folded by an F_3 fold. A spaced crenulation cleavage is developed axial surface to the F_3 fold.

Figure 46

Outcrop photograph of an F_3 fold of Poultney thin quartzites and cleaved silty argillites. Lens cap is 50 mm across.

Figure 45: Photograph of slab section of a refolded fold. An F_2 tight to isoclinal fold with slaty cleavage axial surface to fold is openly folded by an F_3 fold. A spaced crenulation cleavage is developed axial surface to the F_3 fold.

Figure 44: Outcrop photograph of an F_3 fold of Poultney thin quartzites and cleaved silty argillites. Lens cap is 50 mm across.

locally becomes a slaty cleavage (Figure 50). Spacing of cleavage surfaces varies from less than 0.1 to 10 mm or more. At the coarser end, the S_3 cleavage becomes indistinguishable from closely-spaced joints unless folds are present.

On a regional scale, the predominance of S_3 varies considerably in this area, but in a general fashion the frequency with which it is observed and its intensity increases to the east. Farther west, in Jacobi's area this crenulation cleavage is even less frequently observed, supporting the contention made here and elsewhere (Zen, 1961; Wright, 1970) that it increases to the east.

In a few places an apparently conjugate relationship between two sets of crenulation cleavages is observed (Figure 48). Conjugate relationship observed for crenulation cleavages are not commonly reported in the Taconics but are also known to occur at Cedar Point Quarry on Lake Bomoseen. However, close examination of the Cedar Point examples indicated that the crenulation cleavages are of different ages (Means, personal communication, 1979).

Secondary layering, often referred to as differentiated layering (Williams 1972a; Hobbs, et al, 1976) is observed at a few localities to form parallel to the axial surface of mesoscopic F_3 folds (Figure 57). More commonly, secondary layering is not associated with mesoscopic folds, but usually parallels the regional S_3 cleavage orientation (Figure 58). Folds of secondary layering are not observed. For these reasons secondary layering is considered as a member of the S_3 foliation, although some occurrences may be associated with D_2.

Secondary layering is most easily recognized where it crosscuts indisputable bedding in the hinge of a mesoscopic fold (Figure 49).

Figure 47

Photomicrograph of silty argillite with two cleavages. Slaty cleavage (S_2) approximately parallels the grain size layering and is crenulated by a near vertical crenulation cleavage. The crenulation cleavage is in part defined by dark opaque rich films and by reorientation of the slaty cleavage. (100 X magnification, field of view 1.46 mm)

Figure 48

Field sketch of slate with two crenulation cleavages with apparent conjugate relationships. Arrows depict sense of rotation associated with the crenulations.

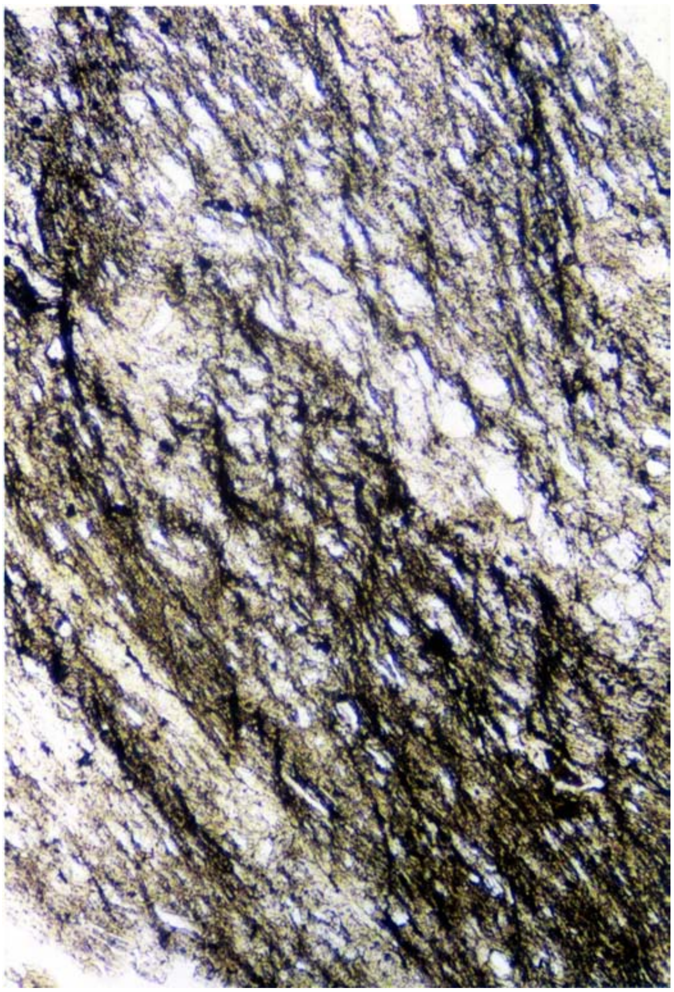


Figure 47: Photomicrograph of silty argillite with two cleavages. Slaty cleavage (S_2) approximately parallels the grain size layering and is crenulated by a near vertical crenulation cleavage. The crenulation cleavage is in part defined by dark opaque rich films and by reorientation of the slaty cleavage. (100x magnification, field of view 1.46 mm)

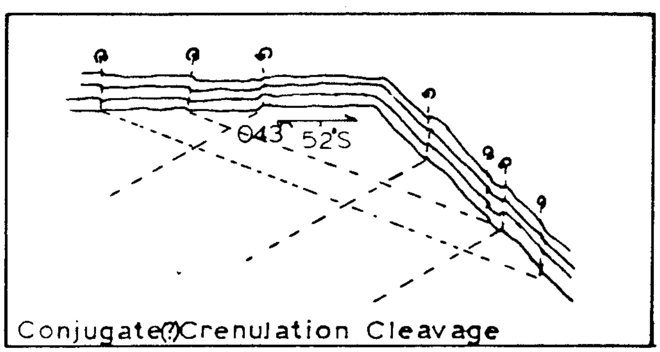


Figure 48: Field sketch of slate with two crenulation cleavages with apparent conjugate relationships. Arrows depict sense of rotation associated with the crenulations.

Often, indisputable bedding is not present, but may be represented by color laminations that cut obliquely across the secondary layers. In these cases the distinction between primary and secondary layering becomes difficult.

In this area, secondary layering is characterized by interlayering of two types of domains with different mesoscopic and microscopic characteristics. The domains are (1) individual mudstone to fine siltstone layers, and (2) usually well-cleaved finer-grained argillite. The argillite domains vary from 1 - 10 mm thick and are mesoscopically similar to surrounding slates. This cleavage may be parallel or oblique to the secondary layering. The coarser-grained layers range from 1 to 15 mm thick and lack bedding features parallel to their surfaces, but often contain grain size or color layering highly oblique to their bounding surfaces. Commonly these layers are discontinuous on an outcrop scale. The earlier S_2 cleavage is often refracted through them. Dale (1899) and Zen (1961) also describe secondary layering and refer to such layers as cleavage bands. Dale (1899) suggests that cleavage bands result from variable development of cleavage. Thus, the mudstone layers were simply less well cleaved. Zen (1961) suggests they result from intense shearing, giving rise to variations in texture.

Talbot and Hobbs (1968) and Williams (1972a), among others, describe secondary layering from other slate regions of the world, and ascribe it to metamorphic differentiation or domainal solution transfer mechanisms. The secondary layering in this area is believed to result from domainal solution transfer of quartz and other relatively soluble minerals, concentrating relatively insoluble micaceous and opaque

Figure 49

Photograph of mesoscopic F_3 fold of thin silty quartzites and cleaved argillite. Thin silty argillite layers are developed where spacing between quartzites is great. These silty argillite layers are axial surface to the F_3 folds and clearly cross-cut bedding demonstrating their secondary origin.

Figure 50

Field sketches of other examples of secondary layering. Abbreviation s.l. is secondary layering.

Figure 49: Photograph of mesoscopic F3 fold of thin silty quartzites and cleaved argillite. Thin silty argillite layers are developed where spacing between quartzites is great. These silty argillite layers are axial surface to the F3 folds and clearly cross-cut bedding demonstrating their secondary origin.

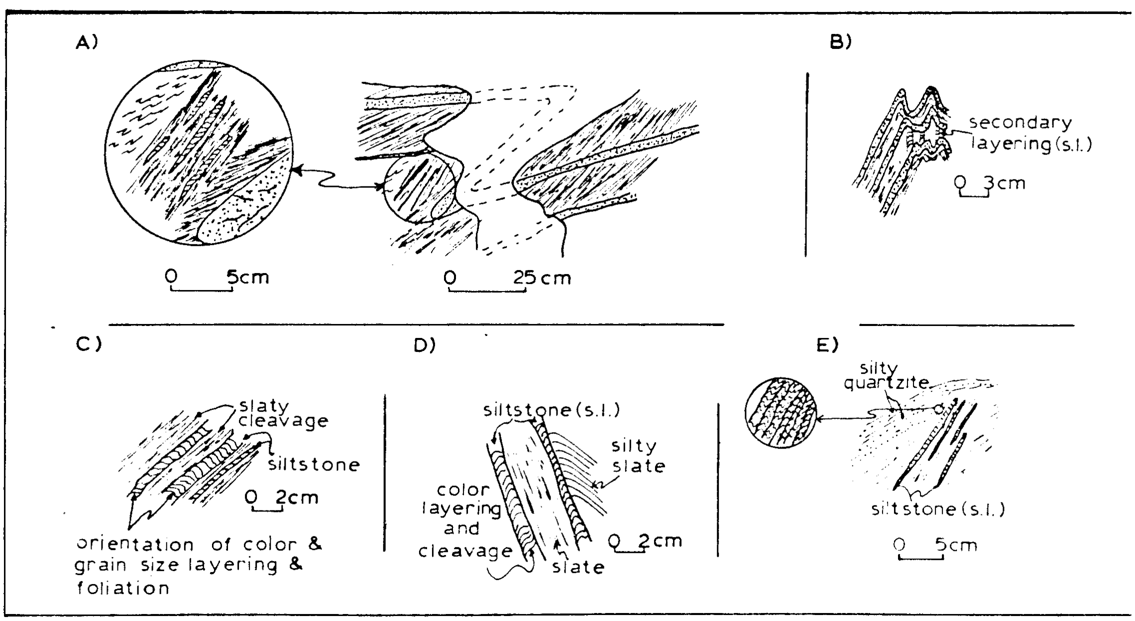


Figure 50: Field sketches of other examples of secondary layering. Abbreviation s.l. is secondary layering.

materials to produce the argillite layers. This appears to be a domainal process which leaves some regions relatively untouched giving rise to the interlayered aspect (Figure 51).

Microstructurally, the S_3 crenulation cleavage ranges continuously from zonal to discrete end member types of Gray (1977a, 1977b, 1979) (Figure 52). Most often the S_3 crenulation cleavage is characterized by straight or anastomosing mica and iron-stained films which 'offset' earlier fabric elements. In some sections, this type of crenulation cleavage changed parallel to the cleavage surface into a simple warping of the earlier fabrics. These changes commonly involved transitional types where some earlier foliation could be traced continuously while others could not.

No new minerals were observed to be associated with the S_3 cleavage though my observations are not detailed enough to state this unequivocably. Neither Zen (1960) nor Potter (1972) observed new mineral phases associated with this cleavage.

D_4 Deformation

This deformation is associated with essentially vertical kink bands which cut across all earlier fabrics (Figure 15). These kink bands are only observed in a few places and strike east to southeast. They vary from 3 to 10 cm wide and are bounded by sharp essentially vertically oriented surfaces. This generation of structures has no macroscopic impact.

Fault Related Rocks

Mesoscopic and Microscopic Structures

Major thrust faults are fairly widespread structures in this area.

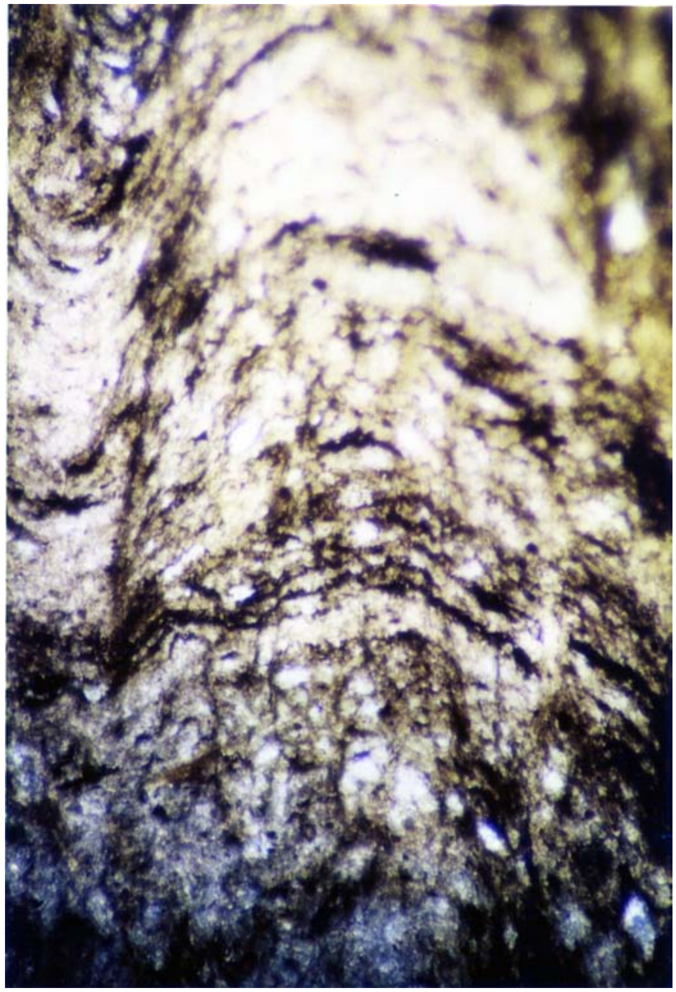
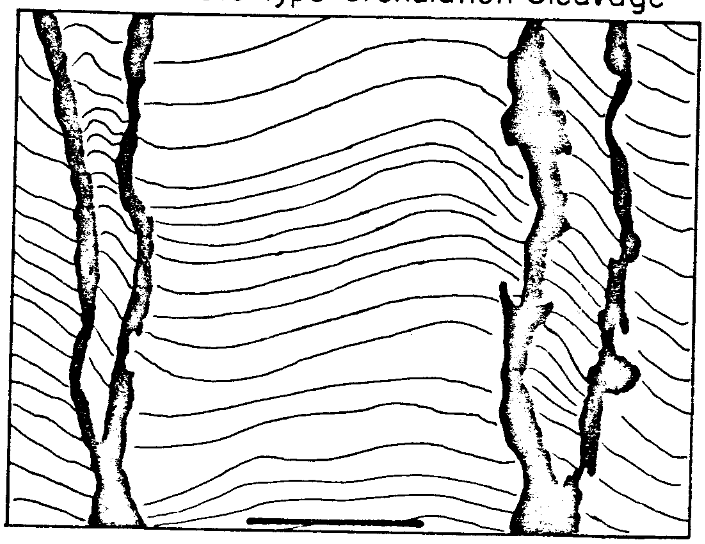


Figure 51: Photomicrograph of a secondary siltstone layer. Bedding and an early cleavage are openly folded and crenulated by a vertically oriented cleavage. The siltstone layer was developed parallel to this crenulation cleavage (40x magnification, 3.64mm field of view).

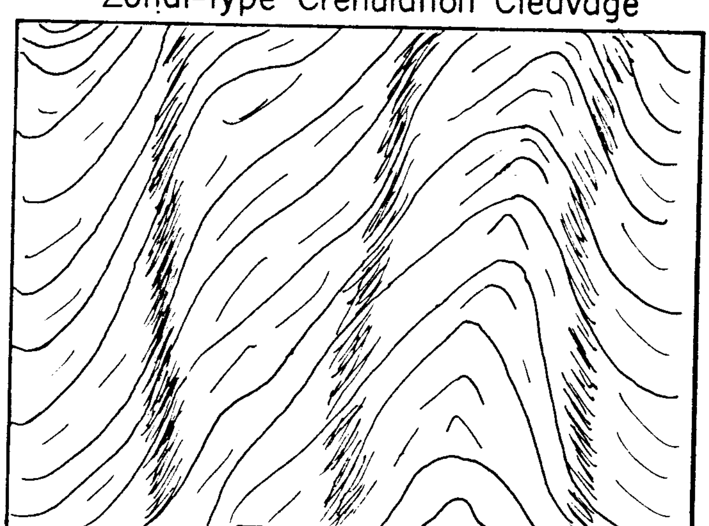
Figure 52

Figure illustrates differences between discrete-type and zonal-type crenulation cleavages as defined by Gray (1977a, 1979). Figure drawn from photographs in Gray (1979, Plates 1A and 1B, pp. 98 and 99). Bar in upper diagram is 250 microns long and in lower 500 microns long.

Discrete-type Crenulation Cleavage



Zonal-type Crenulation Cleavage



Unfortunately, these structures and lithologies related to them are poorly or more commonly not exposed. None the less, in a number of localities, thrust fault-related rocks are observed. These major thrust faults appear to be associated with D_2. This is based on the following observations: (1) they commonly appear to be parallel to the D_2 axial surface foliation; (2) they are not folded by F_2; (3) they are openly folded by F_3; (4) the thrust zones contain a well developed slaty cleavage which is in most respects similar to the regional slaty cleavage. The faults are not associated with deformation structures ascribable to soft-sediment deformation. They are, however, associated with a number of features indicative of 'hard rock' deformation as is described below.

Mesoscopic Structures

Thrust fault zones; i.e. rocks apparently associated with thrusting, vary from 3 cm to more than 4 m thick. There does not appear to be a correlation between thickness and stratigraphic displacement, although this is extremely difficult to demonstrate. The fault zones are characterized by a well cleaved argillaceous matrix within which lithic 'fragments' and vein quartz are commonly observed. The relative proportions of these three components varies, but usually argillite, and less commonly vein quartz predominate.

The argillaceous matrix is usually characterized by a well developed slaty cleavage. In some localities these may have a silky or phyllitic appearance. The cleavage in the argillites tends to parallel the regional slaty cleavage (S_2).

Lithic 'fragments' are commonly observed associated with the thrust zones. Silty quartzites (usually Poultney) and greywackes (Pawlet) are the most abundant lithic-types, although silty argillites and slates are also observed. These fragments occur as lenticular masses with high aspect ratios, and are sometimes elongate in the third dimension, producing a down-dip lineation. These lenticular masses are usually surrounded by argillite and have a preferred orientation parallel to the slaty cleavage. In some cases primary bedding laminations in silty quartzite (Poultney) are observed to be truncated by argillite almost perpendicular to the orientation of the laminations (see Figure 53). The lenticularity and truncation of primary features suggest transposition, boudinage, and/or slicing. Folds or isolated fold hinges are rarely observed (Figure 54) to support the suggestion of transposition, but unequivocable evidence of slicing has not been seen.

Vein quartz is commonly associated with these fault zones. The veins commonly show the complex structure within these zones more precisely than the lithic fragments, presumably due to different and more varied orientations with respect to the deformation and possibly also multiple generations recording different increments of the strain. The veins are commonly tightly to isoclinaly folded and transposed parallel to the slaty cleavage (Figure 55). In some places the veins occur as isolated lozenge-shaped pods surrounded by argillite. From regional observations it appears that lithologies containing highly deformed quartz veins are restricted to thrust fault zones. This may provide a tool for mapping faults where compelling lithostratigraphic evidence is absent.

Microscopic Structures

In thin section, fault related rocks are characterized by more

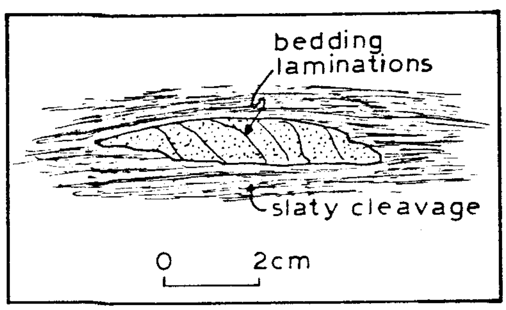


Figure 53

Field sketch of a bedding fragment of silty quartzite in well-cleaved argillite matrix associated with the Stoddard Road Thrust. Note bedding laminations in quartzite are clearly truncated by the cleavage, suggesting folding associated the deformation.

Figure 54

Photographs of deformation associated with thrust faults. Bedding is dismembered and discontinuous, cleavage is variably developed and appears roughly axial surface to the mesoscopic folds. C and D. Photomicrographs of folds associated with faults. D. is detail of hinge of C. Note large chlorite porphyoblast (?) with cleavage oriented perpendicular to slaty cleavage of fold. (C. 10 X magnification field of view. D. 40 X magnification, 3.64 mm field of view).

Figure 54a: Photograph of deformation associated with thrust faults. Bedding is dismembered and discontinuous, cleavage is variably developed and appears roughly axial surface to the mesoscopic folds.

Figure 54b: Photograph of deformation associated with thrust faults. Bedding is dismembered and discontinuous, cleavage is variably developed and appears roughly axial surface to the mesoscopic folds.

Figure 54c: Photomicrograph of deformation associated with faults. (10x magnification, field of view).

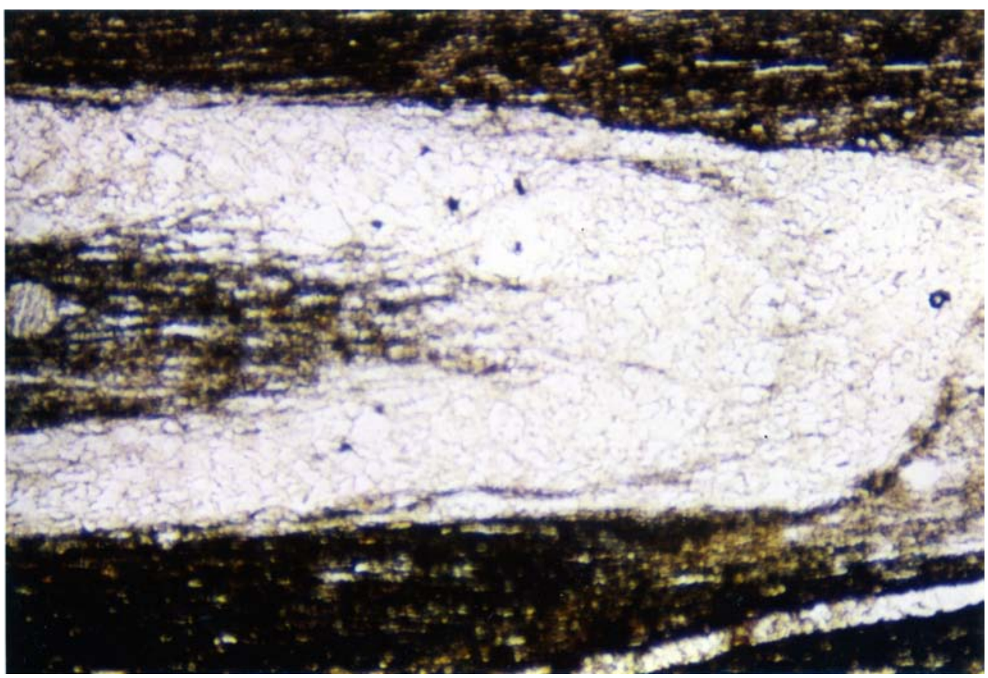


Figure 54d: Photomicrograph of deformation associated with faults. Detail of the hinge of Figure 54c above. Note large chlorite porphyroblast (?) with cleavage oriented perpendicular to slaty cleavage of fold. (40x magnification, 3.64 mm field of view).

Figure 55

a. Photograph of hand specimen of fault rock collected from locality 9+6 on Figure 23. Note folded and highly dismembered quartz veins with cleavage axial surface to tolds. A weak lineation is apparent on the cleavage surface. b & c are outcrop sketches of a few other faults.

Figure 55a: Photograph of hand specimen of fault rock collected from locality 9+6 on figure 23. Note folded to highly dismembered quartz veins with cleavage axial surface to folds. A weak lineation is apparent on the cleavage surface.

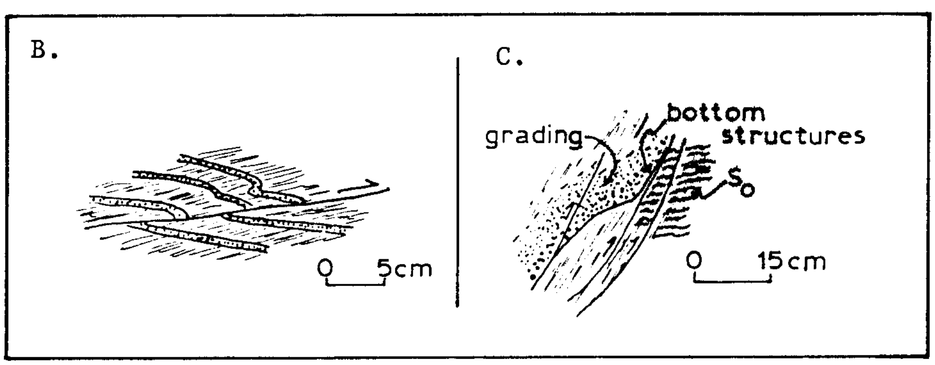


Figure 55b, c.: Outcrop sketches of a few other faults.

complex microstructures than observed in surrounding rocks and commonly appear to be characterized by composite fabrics. The complex fabric relations are best displayed in the argillaceous components of these rocks. The earliest fabric, defined by mica and slightly opaque-rich films, is only recognized in isolated areas where it is oblique to the pervasive slaty cleavage and often tightly folded (Figure 56). This foliation may be an earlier cleavage or a bedding plane parting; the two commonly appear to be indistinguishable (Means, personal communication, 1979). This early foliation is pervasively overprinted by a well developed slaty cleavage, which is the dominant fabric in the argillites. This slaty cleavage is very similar to the regionally developed slaty cleavage (S_2). It is defined by mica and opaque-rich films, quartz and mica beards, preferred orientation of larger white and green micas, and elongate, lozenge-shaped stringers of vein quartz. In some sections, somewhat richer in quartz, the cleavage is defined by a fine, anastamosing and interweaving network of thin mica films. This produces an elongate polygonal character to the grain boundaries. It is very difficult to know in such cases whether this fabric results from a single complex fabric or the intersection of two or more 'generations' of fabrics. In some this question could be resolved, however, more often it could not.

In some thin sections, the slaty cleavage is crenulated. The nature of the crenulation fabric varies widely, but includes end members characterized by (1) an evenly spaced crenulation cleavage with essentially the same cleavage morphology throughout the section (Figure 57) to (2) isolated or heterogeneously developed crenulation fabrics which vary both in morphology and intensity within a single section. The

Figure 56

Photomicrographs of isolated 'early' foliations truncated by the penetrative slaty-like cleavage of the fault rocks (a and b, 40 X magnification, 3.64 mm field of view).

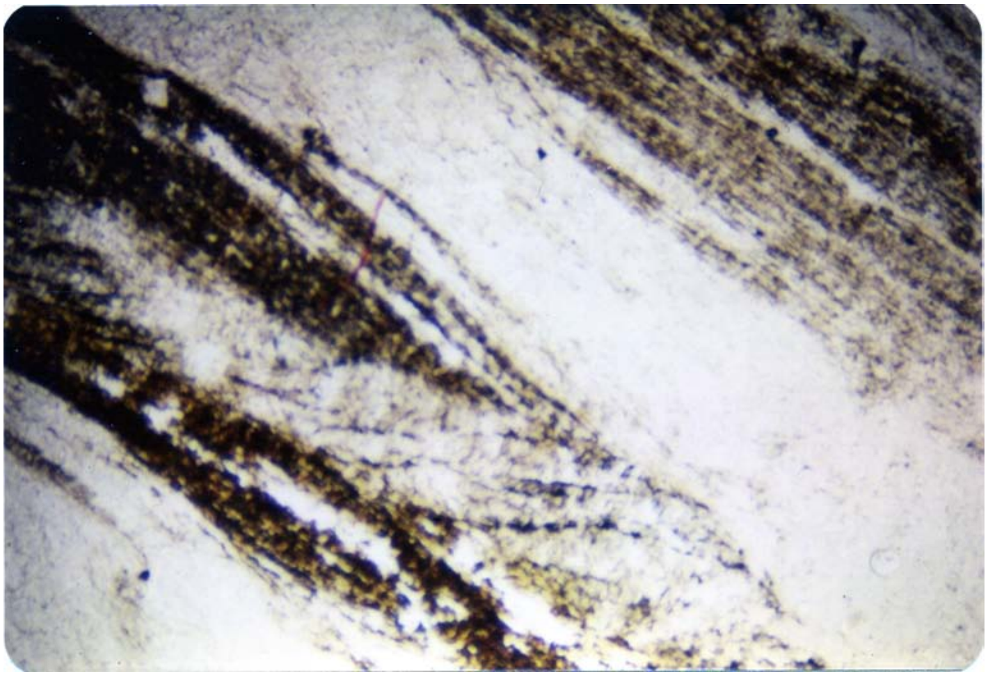


Figure 56a: Photomicrograph of isolated 'early' foliations truncated by the penetrative slaty-like cleavage of the fault rocks (40x magnification, 3.64 mm field of view).

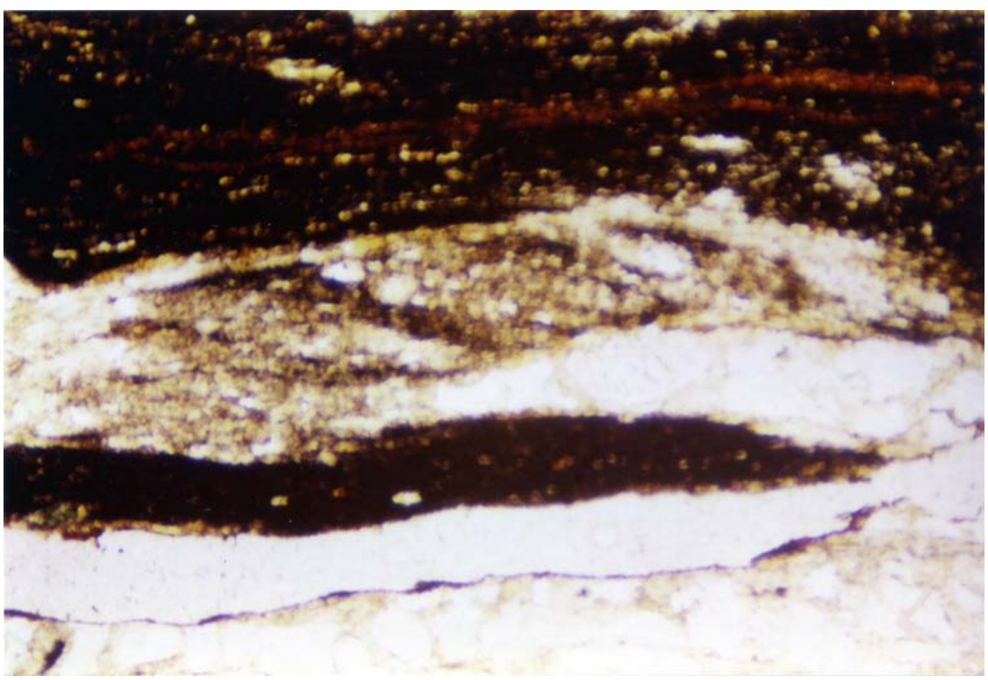


Figure 56b:
Figure 56a: Photomicrograph of isolated 'early' foliations truncated by the penetrative slaty-like cleavage of the fault rocks (40x magnification, 3.64 mm field of view).

first type closely resembles the regional S_3 crenulation cleavage and probably correlates with it. In sections containing the second end member type the resemblance and correlation with the regional S_3 is not apparent. In several sections the slaty cleavage is well developed and uncrenulated, except in a single isolated place where the slaty cleavage is openly to tightly folded (Figure 58). These isolated crenulations never affect more than a few tenths of a square millimeter. In other sections, both discrete and zonal type crenulation cleavages are well developed, but in a heterogeneous manner, when the whole section is considered. In these cases the discrete-type cleavages are commonly defined by dark, opaque-rich films which anastamose across the section. These opaque-rich films are best developed where they are adjacent to quartz vein boundaries or where they are bounded on both sides by vein boundaries. These films closely resemble features ascribed to solution transfer of quartz and other soluble material and simultaneous concentration of mica and opaques.

Vein quartz in these rocks ranges from relatively large (several millimeters thick) discrete veins to thin (less than 1-2 mm) veins which are visibly deformed with the argillites. Vein quartz is identified by the interlocking, polygranular character of the quartz. Grain size varies from less than 0.1 mm to more than 1.0 mm. Grain shapes include polygonal to elongate and locally fibrous ones. Polygonal grains are most common.

Vein quartz occurring in the massive mode commonly appears to be essentially undeformed. Grains lack obvious shape preferred orientation but do have undulatory extinction. In some sections carbonate filled cracks are observed within some of these veins. The cracks tend to be

Figure 57

Photomicrograph of argillaceous fault rock with well developed slaty cleavage and spaced, weak crenulation cleavage. Crenulation cleavage is oriented parallel to the regional crenulation cleavage which suggests a pre-D_3 timing for this fault (Stoddard Road Thrust). (100 X magnification, 1.46 mm field of view).

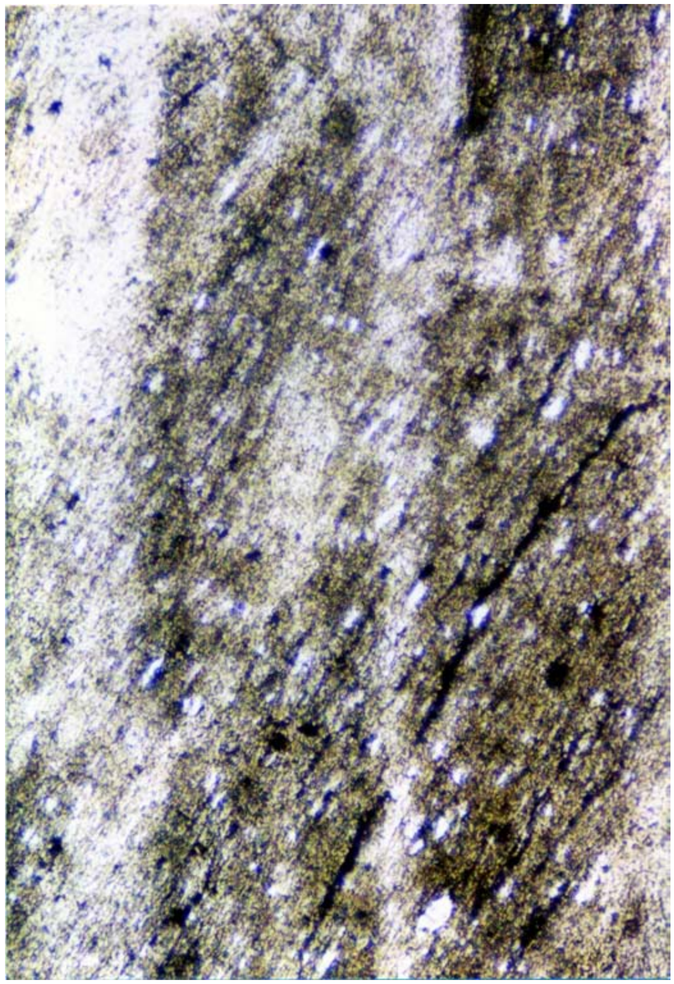


Figure 57: Photomicrograph of argillaceous fault rock with well developed slaty cleavage and spaced, weak crenulation cleavage. Crenulation cleavage is oriented parallel to the regional crenulation cleavage which suggests a pre-D_3 timing for this fault (Stoddard Road Thrust). (100x magnification, 1.46 mm field of view).

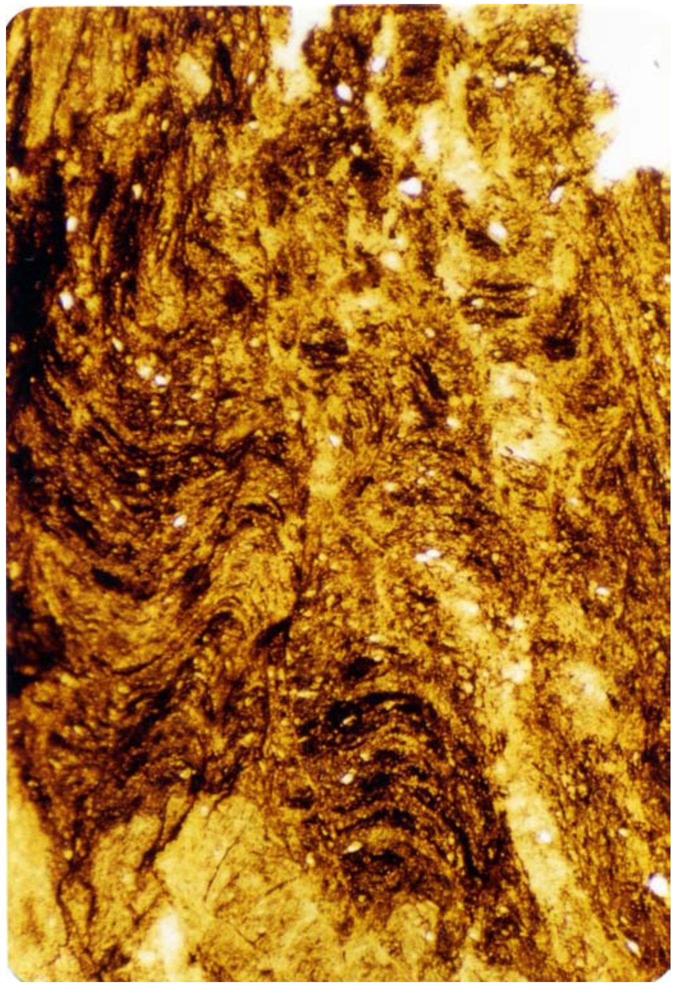


Figure 58: Photomicrograph of isolated folds of cleavage in fault rock. Cleavage in the rest of the thin section approximately parallel to the long dimension of the photomicrograph. The cleavage in the rest of the thin section can be traced into cleavage in these folds; it therefore predates the folds. (4x magnification, 2.1 mm field of view).

oriented perpendicular to the slaty cleavage, possibly indicating a genetic relationship. Vein boundaries are commonly sharp and associated with dark mica films. Some grains along the boundaries have shapes suggestive of corrosion and solution removal of quartz (Figure 59). Quartz and mica beards may be present along some vein boundaries, but only where the boundaries are oblique to the cleavage.

Thin veins of quartz are commonly observed to be associated with argillite. Most of these veins are tightly to isoclinally folded and transposed (Figure 60). Thin elongate, lozenge-shaped stringers of fine grained polygonal quartz are commonly strung out parallel to slaty cleavage and contribute to its definition (Figure 61). Locally, isolated intrafolial fold hooks are observed. In some sections, folds of veins have not been dismembered. These folds vary from tight to isoclinal, with either rounded or angular hinge s. Slaty cleavage is in all cases parallel to the axial surface. Rounded hinges commonly occur in relatively thick (approximately 1 mm) veins in which the enveloping surface of the folded vein is oriented perpendicular to the slaty cleavage. Quartz grains in these veins are commonly elongate, and have sutured boundaries (Figure 62). The shape preferred orientation of these grains defines convergent fans around the hinges of the folds. The cleavage in the adjacent argillites forms divergent fans. Folded veins with angular hinges are more commonly observed, and are characteristic of thin (0.1-0.5 mm) veins. The quartz grains within these folds do not show obvious evidence of internal deformation, although recrystallization may have occurred.

The presence of thin, isolated, elongate, lozenge-shaped vein quartz within argillite can be attributed to (1) transposition and

Figure 59

Photomicrograph of fault rock. Polygonal grains of vein quartz are cross-cut by a thin selvage of dark opaque-rich argillite. The truncation of the quartz grains suggests that solution of quartz has occurred. Note the deformation lamellae in the vein quartz. (40 X magnification, 3.64 mm field of view - section very thick).

Figure 60

Photomicrograph of highly deformed and dismembered vein quartz from fault rock. Note planar preferred orientation of quartz within well cleaved argillite matrix (100 X magnification, 1.46 mm field of view).

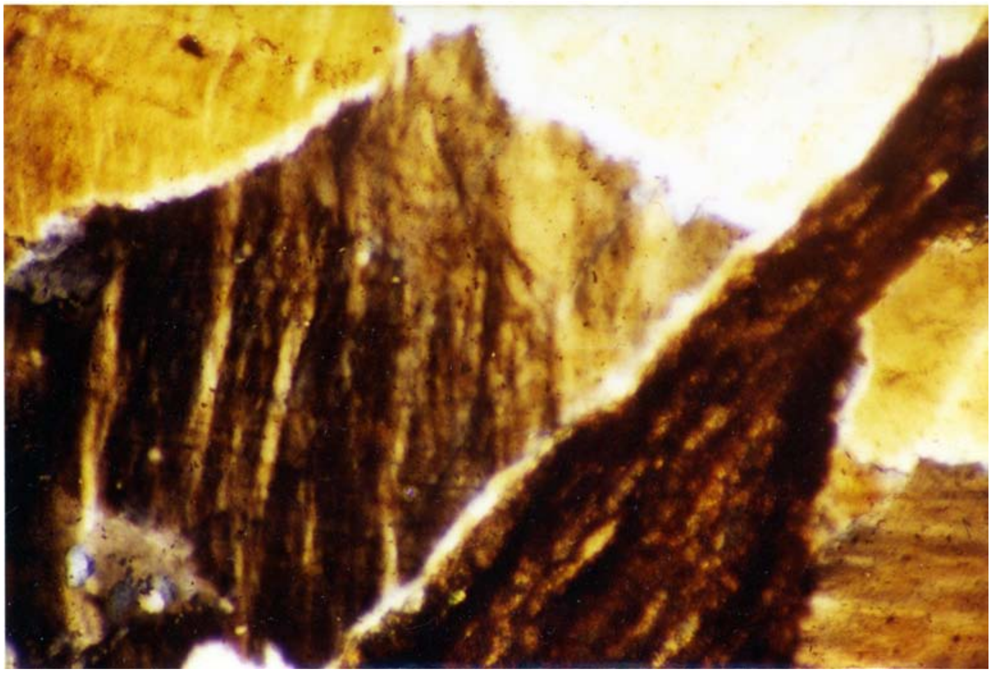


Figure 59: Photomicrograph of fault rock. Polygonal grains of vein quartz are cross-cut by a thin selvage of dark opaque-rich argillite. The truncation of the quartz grains suggests that solution of the quartz has occurred. Note the deformation lamellae in the vein quartz. (40x magnification, 3.64 mm field of view – section is very thick).

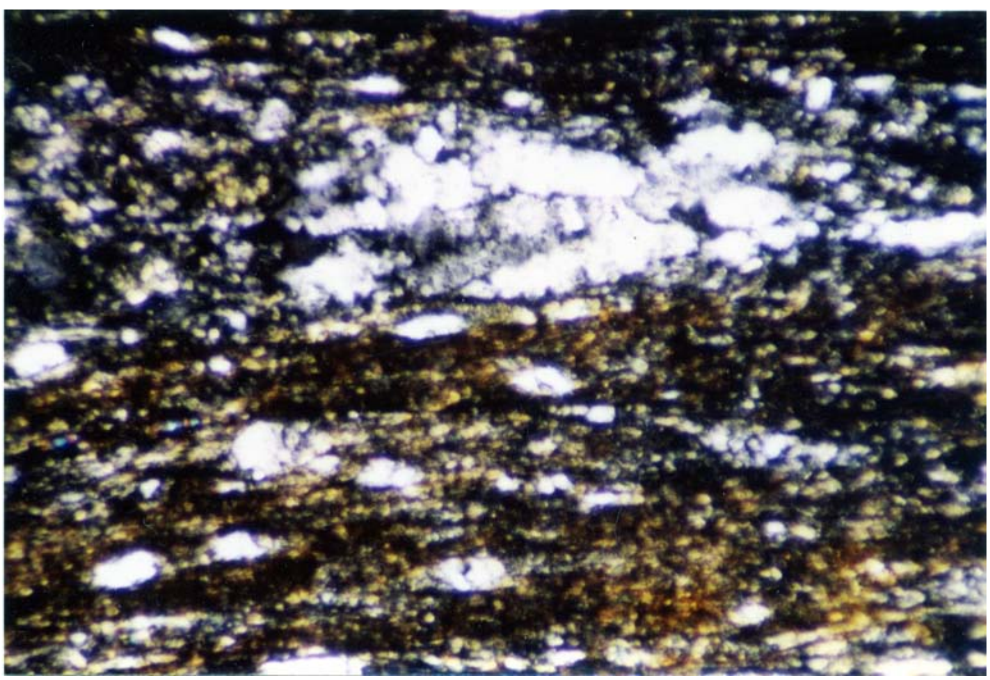


Figure 60: Photomicrograph of highly deformed and dismembered vein quartz from fault rock. Note planar preferred orientation of quartz within well cleaved argillite matrix. (100x magnification, 1.46 mm field of view).

Figure 61

Photomicrograph of isolated and dismembered fold hinge of vein quartz within well cleaved argillite matrix of fault-related rock. (40 X magnification, 3.64 mm field of view)

Figure 62

Photomicrograph of a relatively wide tightly folded and partially dismembered quartz (section very thick) vein. Vein initially oriented approximately perpendicular to present foliation. Note the lack of polygonal grains and sutured appearance of grain boundaries. The elongate shape of the grains define convergent fans within the veins. The well developed foliation within the argillite matrix is axial surface to the folds of the vein and form divergent fans in the hinge areas (40 X magnification, 3.64 mm field of view).

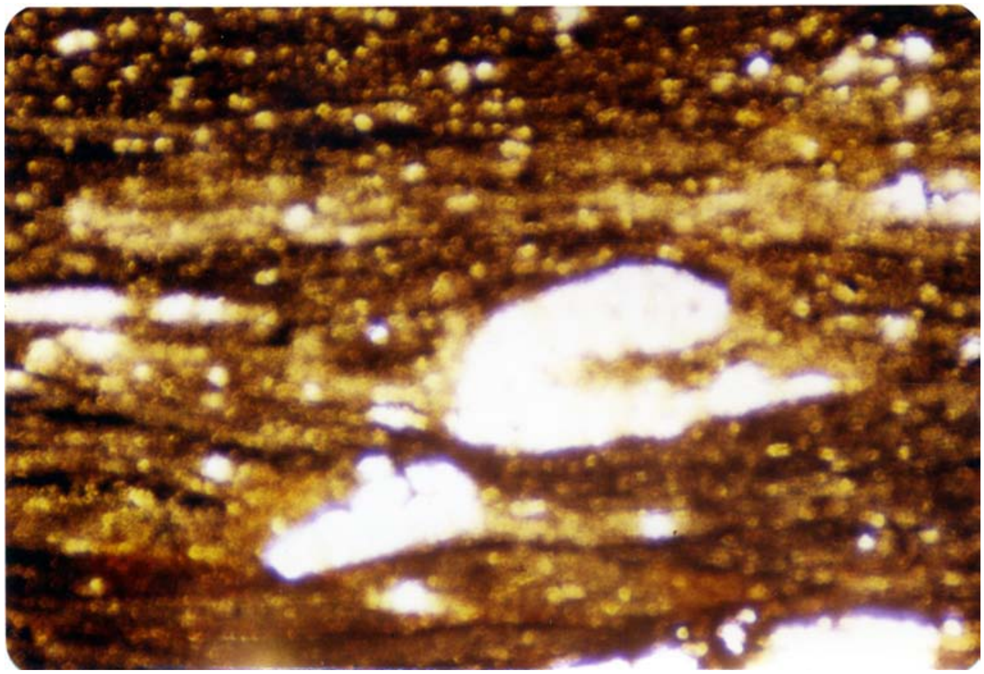


Figure 61: Photomicrograph of isolated and dismembered fold hinge of vein quartz within well cleaved argillite matrix of fault-related rock.. (40x magnification, 3.64 mm field of view).

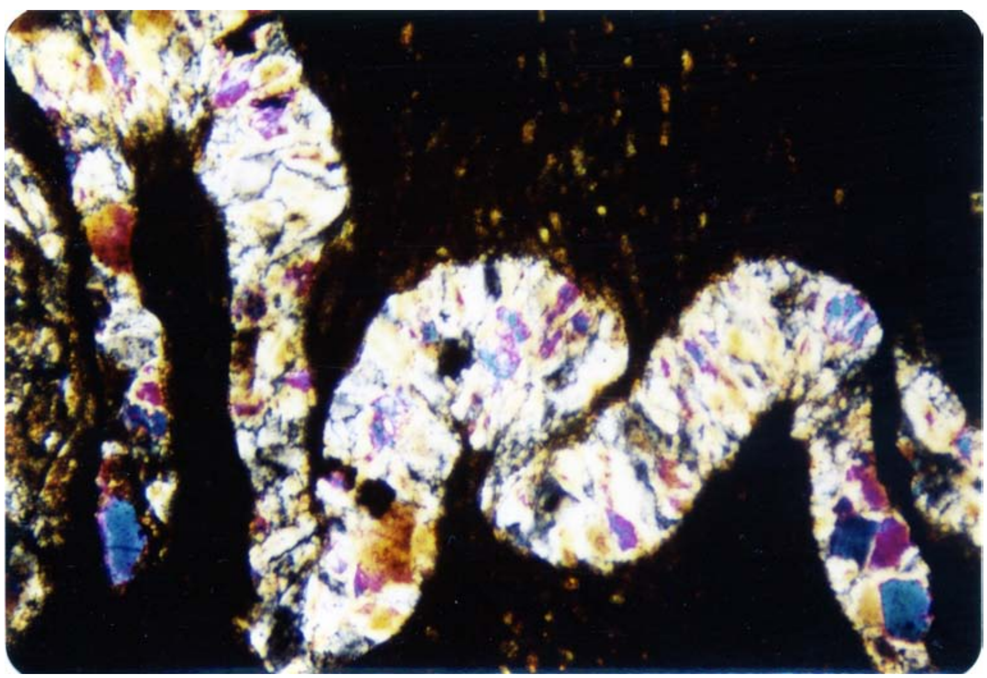


Figure 62: Photomicrograph of a relatively wide tightly folded and partially dismembered quartz (section very thick) vein. Vein initially oriented approximately perpendicular to present foliation. Note the lack of polygonal grains and sutured appearance of grain boundaries. The elongate shape of the grains define convergent fans within the veins. The well developed foliation within the argillite matrix is axial surface to the folds of the vein and form divergent fans in the hinge areas. (40x magnification, 3.64 mm field of view).

slicing parallel to slaty cleavage, (2) heterogeneous and incomplete solution removal of quartz from folded veins, with solutioning occurring preferentially in the hinges, or (3) some combination of (1) and (2).

Tight to isoclinal folds are easily demonstrated, but slicing is not. Good markers, composed of insoluble material, are not observed, and thus offsets in veins or other objects can be equally well attributed to solution removal of material as to slicing. Several observations are compatible with solution transfer processes. These include (1) dark mica and opaque-rich films, presumably representing the insoluble residuum (Williams, 1972a; Gray, 1977b and others). (2) Corroded and truncated quartz and carbonate grains (Figure 63). These grains have distinctly dissimilar appearance to surrounding apparently unsolutioned grains. (3) The intimate relationship of (1) and (2).

No conclusive evidence of slicing has been observed in these rocks, but because of the lack of markers this is not surprising. Solution transfer processes appear to have operated and may be the dominant mechanism responsible for the isolation of the thin vein quartz material, however, the possibility of combined slicing and solution cannot be discounted and like slicing alone, will be very difficult to demonstrate.

Thin fibrous veins are also observed in some sections. These veins tend to be oriented parallel to the crenulation fabric (Figure 64). The fibers are relatively straight and appear to be optically continuous across the veins, suggesting antitaxial growth (Durney and Ramsey, 1973). Individual fibers appear to connect cleavage folia across the vein. These relations suggest extension perpendicular to the crenulation cleavage direction presumably after its development, as the veins themselves appear undeformed.

Figure 63

Photomicrograph of isolated remnant of sedimentary carbonate within a fault-related rock. Foliation in the matrix wraps around the clast. The right side of the clast appears truncated against a dark oxide rich film. This truncation suggests possible solution transfer of the carbonate along this boundary. (100 X magnification, 1.46 mm field of view).

Figure 64

Photomicrograph of thin veins of fibrous quartz in a fault-related rock. Veins are oriented parallel to a crenulation cleavage. Fibers connect cleavage folia across the vein suggesting antiaxial growth (100 X magnification, 1.46 mm field of view).

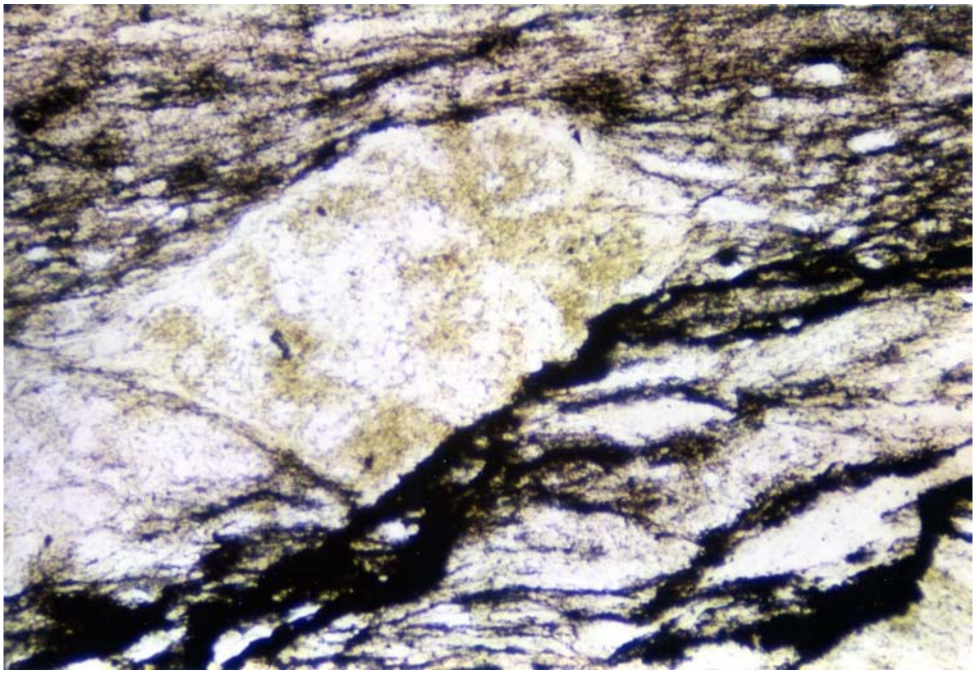


Figure 63: Photomicrograph of isolated remnant of sedimentary carbonate within a fault-related rock. Foliation in the matrix wraps around the clast. The right side of the clast appears truncated against a darker oxide rich film. This truncation suggests possible solution transfer of the carbonate along this boundary. (100x magnification, 1.46 mm field of view).

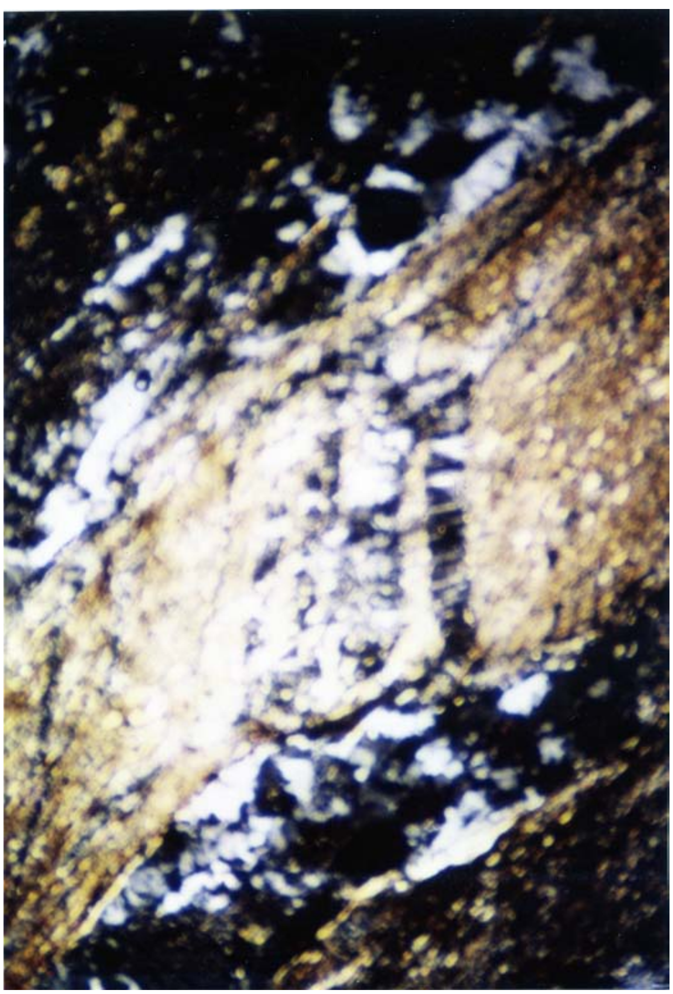


Figure 64: Photomicrograph of thin veins of fibrous quartz in a fault-related rock. Veins are oriented parallel to a crenulation cleavage. Fibers connect cleavage folia across the vein suggesting antitaxial growth (100x magnification, 1.46 mm field of view).

Summary

These fault related rocks appear to be characterized by tightly to isoclinally folded and transposed quartz veins in a well cleaved argillite matrix. The slaty cleavage is morphologically and mineralogically similar to the 'normal' slaty cleavage. In some cases the slaty cleavage transects an earlier fabric of unknown affinities. Slicing parallel to the slaty cleavage could not be demonstrated. Solution removal of quartz and other soluble material is frequently suggested by corroded grain shapes and association with mica and opaque-rich films. The cross cutting crenulation cleavage may in some cases be correlated with the regional S_3 crenulation cleavage, but this correlation is usually less clear. In the former cases, this correlation supports a pre-D_3, syn- to late D_2 timing. In the later cases, the timing is not constrained by the mesoscopic or microscopic observations. These fabrics and microstructural relations may have resulted from a 'single' progressiv e polyphase deformation history associated with either D_2 or D_3, or they could have complex, reactivation histories reflecting D_1, D_2 and/or D_3. Present microstructural observations do not constrain the interpretation. Regional structural relations suggest that three generations of thrust faults are present in the study area, but data is insufficient to constrain timing enough to determine thrusts which have been reactivated.

The presence of quartz veins suggests a relatively brittle behavior of the slates. The relative concentration of quartz veins near faults suggests a possible genetic relationship. Tight to isoclinal folding and transposition of these veins suggest a relatively ductile behavior at least of the veins. The strength of quartz is significantly influenced by water (fluids), resulting in hydrolytic weakening (Hobbs et al, 1972). The presence of features suggestive of solution removal of quartz and other soluble material strongly supports the presence of a fluid phase during deformation. Combined these observations suggest that deformation occurred near the brittle-ductile transition for the materials involved in the presence of fluids. The observations are completely incompatible with the development of these faults in soft-sediments, or as 'late', near surface (few 100m??) structures.

Other Faults

Two other types of thrust faults are recognized in this area. These are (1) slickensided, usually veined surfaces which mostly parallel either bedding or the S_2 axial surface slaty cleavage. (2) Steep, 'late' schuppen, sometimes associated with veining and slickensided surfaces. Both types are believed to be of secondary importance, and associated with less than a few meters offset. Neither type of fault was observed on map scale to recognizably offset stratigraphy.

The slickensided, and veined variety of faults belong to at least two generations. Both generations are characterized by 2-5 cm thick zones composed of argillite and quartz. The earlier generation is believed to have formed synchronously with the major D_2 thrusts described above. This correlation is based on the presence of a single, well developed slaty cleavage within argillites associated with these faults. This slaty cleavage parallels the cleavage in the rest of the outcrop. At one locality, one of these faults was openly folded, coaxially with surrounding bedding. The slickenside lineation always trends east to southeast. These faults may be small splays off more major thrusts with which they are commonly spatially associated.

The second generation of veined, slickensided faults post-date D_2, as demonstrated by the presence of disoriented angular slate clasts within a massive vein quartz matrix. Slickenside lineations on these faults also trend roughly east-west. These 'late' schuppen are characterized by fairly narrow (less than 3-5 cm), steeply east-dipping displacement zones. The nature of these schuppen depends primarily on the lithology of the associated rocks. In hard slates the schuppen tend to be narrow, sharp sided faults, with or without slickenside lineations. In softer slates these faults are usually characterized by dark 'smeared' surfaces. Veins may be associated with these faults, but they are never observed to be deformed. Bedding and cleavage sometimes are slightly bent into these faults suggesting east over west transport. These faults tend to be oriented parallel to the S_3 crenulation cleavage. These faults must post-date D_2 and may be associated with D_3 or may be later structures.

In some siliceous rocks slickensided bedding surfaces are also observed. These surfaces probably record a phase of flexural slip folding. These lineations are usually southeast trending. Becuase of the coaxial nature of the folding in this area, it is difficult to know whether this reflects D_2 or D_3 folding.

Veins

Veins are fairly common features in this area. Veins are usually filled with quartz and/or carbonate, and may contain lithic 'fragments'. They vary in thickness from less than a millimeter to more than a meter. Three generations of veins are distinguishable within this area. The earliest generation is pre- to early D_2. These veins are tightly to isoclinally folded and locally boudinaged. Transposition of these veins has occurred locally where they are associated with thrust faults. The second generation of veins is characterized by fibrous quartz or carbonate. In some places fibrous veins appear to be deformed, during the late stages of slaty cleavage development. In some veins that show this relationship, the fibers have a down-dip orientation. In other occurrences of fibrous veins, the veins appear to be unaffected by slaty cleavage deformation. In these, the orientation of the vein filling cracks do not appear to be structurally controlled and the fibers are oriented perpendicular to the crack boundaries.

The latest generation of veins cross cut the regional cleavages without apparent deformation. These are most commonly composed of quartz. They vary from a few millimeters to more than a meter thick. Angular and disoriented slate and lithic clasts are commonly observed within these veins (Figure 65). Internally the veins lack evidence of deformation, being characterized by interlocking polygonal grains.

Part 3: Orientation of Mesoscopic Structural Elements

The orientation of bedding (S_0), slaty cleavage (S_2), mesoscopic fold axes (probably both F_1 and F_2, and possibly F_0 as well) and bedding-slaty cleavage intersection lineations, and crenulation cleavage are plotted on equal area stereonets which are presented in Figures 66 to 73. Data are presented for three areas, these are the western and eastern subareas of Part 5, and Jacobi's area to the west. It is recognized that misidentification of fabric elements in the field can give rise to error; however, it is hoped that the large number of measurements used for these plots will minimize the effect of such errors.

Figure 65: Hand specimen of late, vein quartz rich fault-related rock containing disoriented angular clasts of slate and other lithic fragments.

Contours are constructed using the method described in Turner and Weiss (1963, p. 60-64).

The following inferences and conclusions can be drawn from these projections, when field observations are also considered.

(1) The concentration of poles to bedding (Figure 66) and slaty cleavage (Figure 70) are at essentially the same point maxima (Figure 73) which reflects the tight to isoclinal overturned folding during D_2.

(2) The high concentration of poles to slaty cleavage (S_2) (Figure 70) and lack of any tailing of contours strongly supports the observation that later deformations have not significantly redistributed S_2 and older structures.

(3) There is a marked difference in orientation of slaty cleavage (S_2 and S_{J-1} *) (Figure 67) between Jacobi's area to the west and the study area. This change is also observed in the macroscopic fold orientation (F_2 and F_{J-1} *) (Figure 74). Comparison of Figures 67 and 68 shows that this change occurs across the Middle Granville thrust. This difference in orientation probably reflects either a difference in strain history during slaty cleavage and fold development or relative rotation of these elements after their formation.

The second possiblity is considered most likely, since the Middle Granville thrust is a third generation thrust (Part 4) that truncates macroscopic F_2 and F_{J-1}, folds and second generation thrusts. The amount of relative rotation is uncertain, but the difference in average slaty cleavage orientation is 34°.

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* F_{J-1}, S_{J-1}, etc., refer to generations of structure observed to the west by Jacobi (1977).

(4) Pre-F_3 mesoscopic fold axes and bedding-slaty cleavage intersection lineations (Figure 71) are primarily either gently south or north-plunging. The weak girdling of these elements is approximately coplanar with slaty cleavage. This weak girdling can be explained by (a) passive rotation of early formed folds (F_0, F_1, and possibly F_2) towards parallelism with the direction of maximum total elongation, which from total strain analyses is essentially in the down dip direction (Wood, 1973, 1974); (b) an early phase of approximately east-west crossfolding; (c) markedly non-coplanar orientation of bedding prior to folding, presumably due to slumping; (d) some combination of these processes.

Passive rotation of pre- and early-D_2 structural elements is highly probable considering the large amounts of elongational strain (~ 150%) (Wood, 1973, 1974) associated with the D_2 deformation.

(5) Thrusting within both the western and eastern subareas did not involve demonstrable reorientation of structural elements (Figures 68 and 69). However, this in itself does not constrain the timing of thrusting.

(6) The crenulation cleavage (S_3) (Figure 72) is distinctly more sttepIy east-dipping than the slaty cleavage (S_2) (Figure 70). Slaty cleavage and crenulation cleavage have almost the same strike. D_3 folding does not appreciably redistribute earlier structural elements (Figure 70), supporting the contention that D_3 does not noticeably influence the map pattern.

Figure 66 through 73

Stereonet projection of structural elements from the study area. Contours were constructed according to the method described by Turner and Weiss (1963, p. 60-64).

Figure 66

Poles to bedding (S_0). 237 points. Contours are 1, 5, 10, 15, and 20% per 1% area. 'Mean' orientation of S_0 derived from point maximum (Point Maximum at 48^{o} to 300^{o}).

Figure 67

Poles to slaty cleavage S_{J-1} in area adjacent to the west mapped by Jacobi (1977). Orientations taken from her map. 160 points. Contours are 1, 10, 20, and 30% per 1% area. (Point maximum at 47^{o} to 262^{o}).

Figure 68

Poles to slaty cleavage (S_2) from the central region of the study area. 77 points. Contours are 1, 5, 10, and 15% per 1% area. (Point maximum at 42^{o} to 293^{o}).

Figure 66. S_0

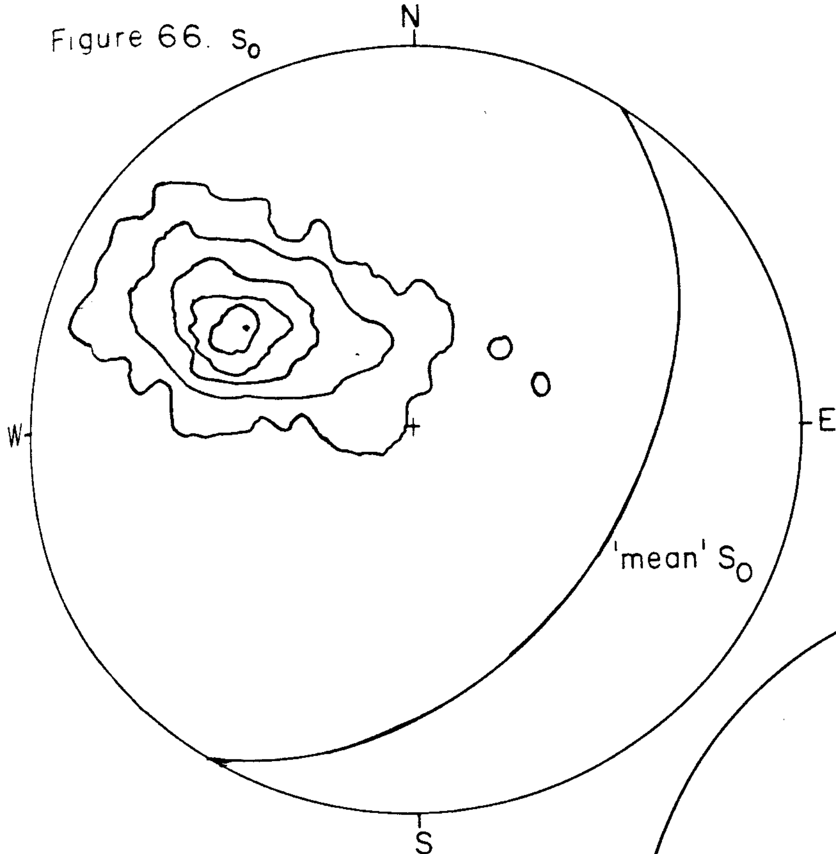


Figure 67. S_{j-1}

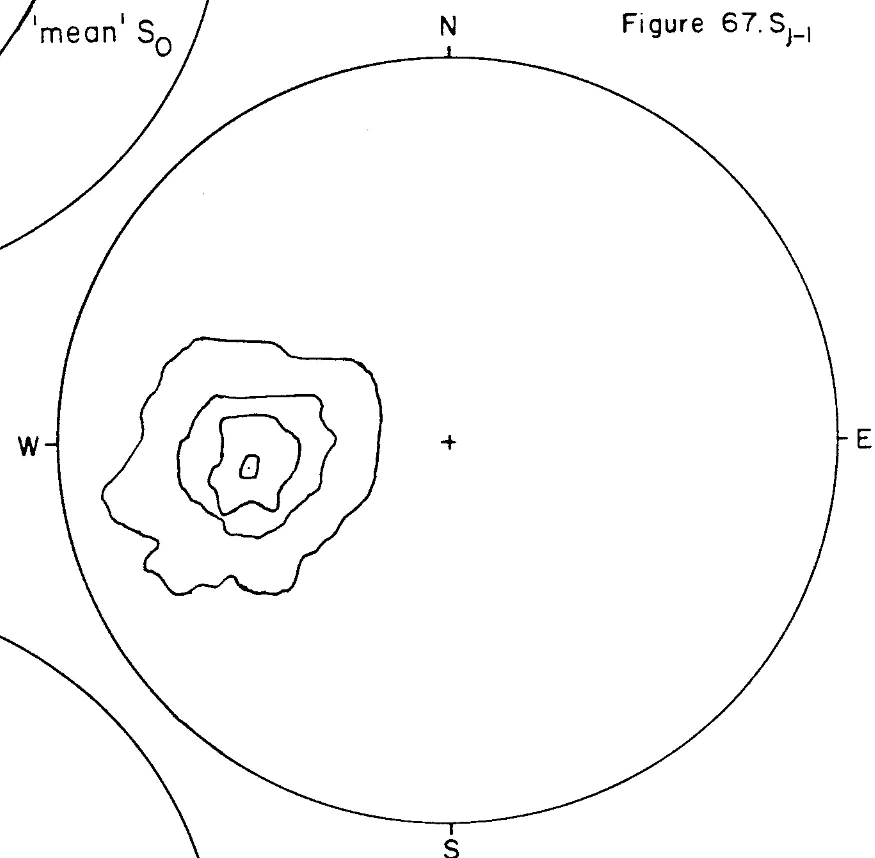


Figure 68. S_2 Central Region

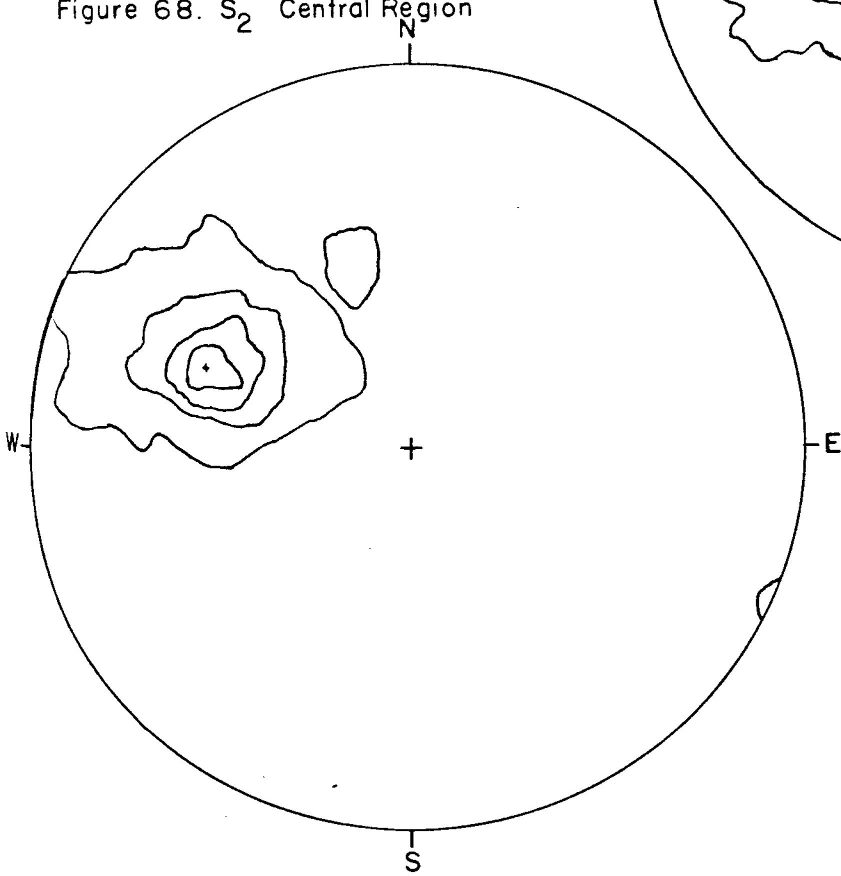


Figure 69

Poles to slaty cleavage (S_2) eastern part of study area. 86 points. Contours are 1, 10, 20, and 30% per 1% area. (Point Maximum at 45^o to 304^o).

Figure 70

Poles to slaty cleavage measurement from across the study area. 614 points. Contours are 1, 10, 20, and 30% per 1% area. (Point Maximum at 42^o to 297^o).

Figure 71

Hinge orientations and bedding-slaty cleavage intersection lineation orientation for the study area. Hinge orientations are primarily for F_2 folds. Some F_1 and possibly F_0 hinges may have been included because of inability to distinguish between them. The orientation of 'mean' slaty cleavage (S_2) is also plotted, based on Figure 70. 205 points. Contours are 1, 5, 10, and 15% per 1% area.

Figure 69. S_2 Eastern Region

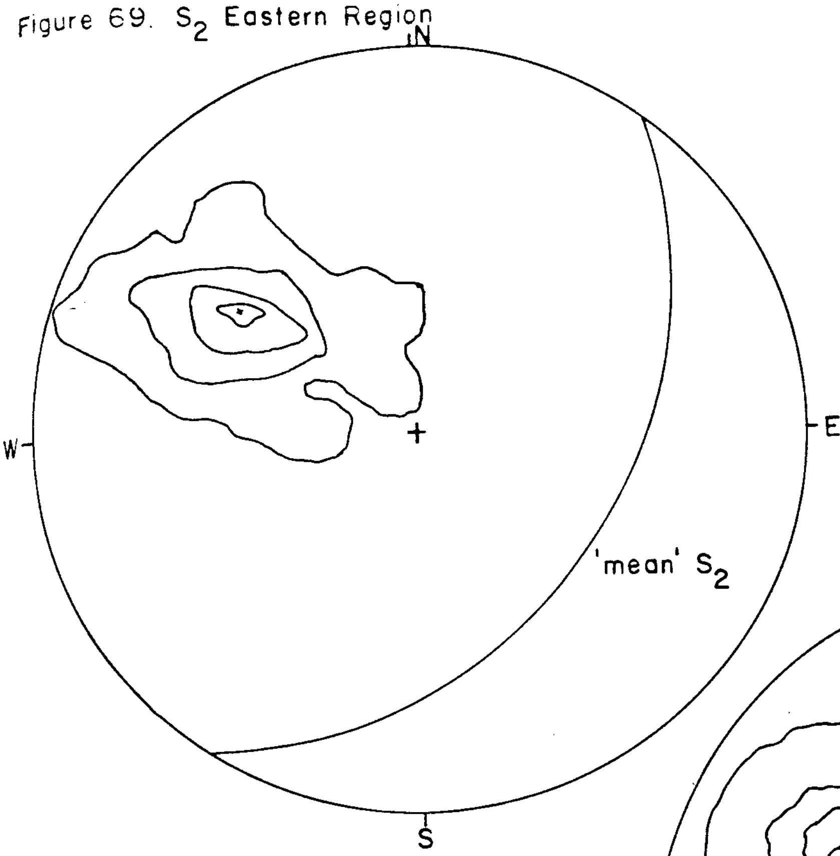


Figure 70. S_2 Whole Area

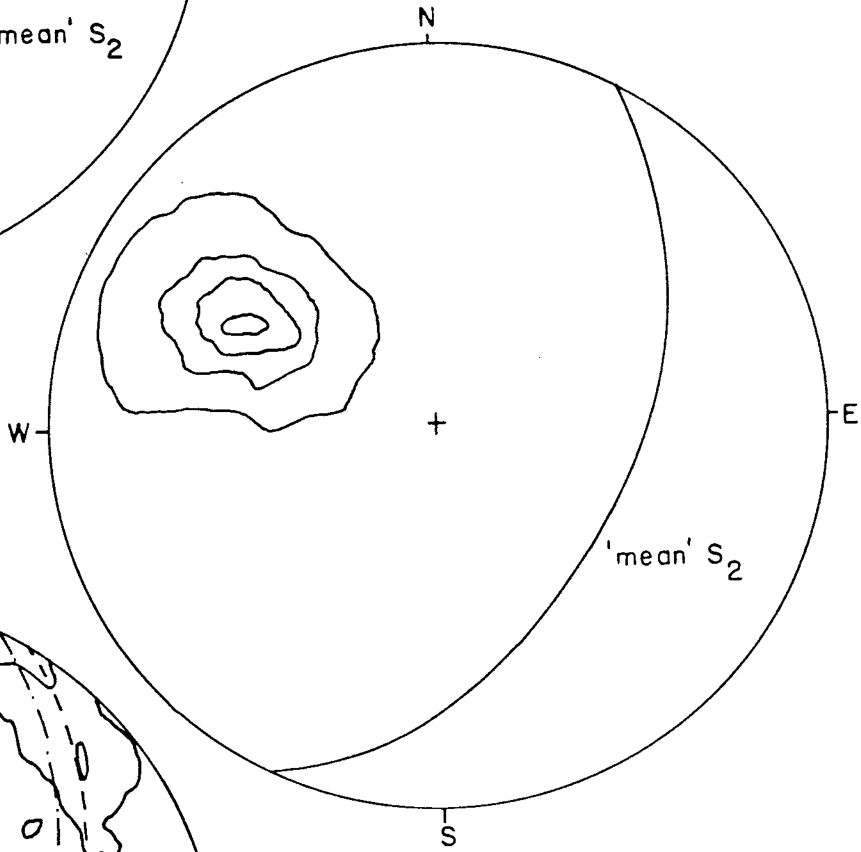


Figure 71. F_2 & L_2^0

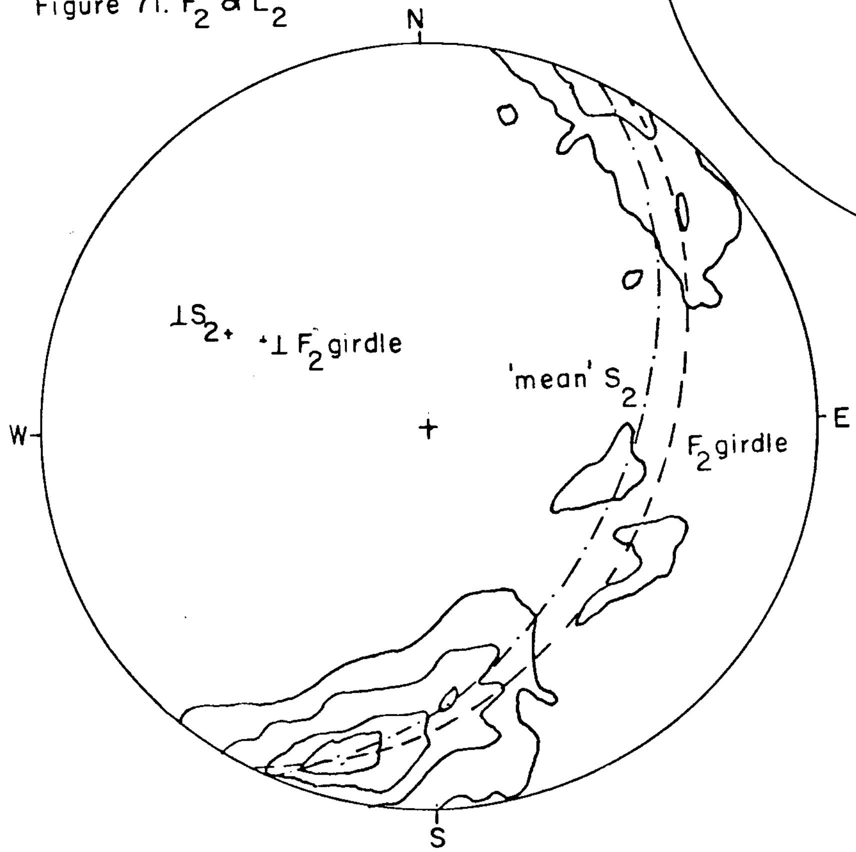


Figure 72

Poles to crenulation and fracture cleavage (S_3) for entire area. 62 points. Contours are 1, 10, and 20% per 1% area. (Point Maximum at approximately 22^o to 287^o).

Figure 73

Synoptic diagram showing point maxima for S_0, S_{J-1}, S_2, S_3, and pole F_2 hinge girdle and 'mean' orientations of S_0, S_2, S_3, and girdle defined by F_2 hinges.

Figure 72. S_3 Whole Area

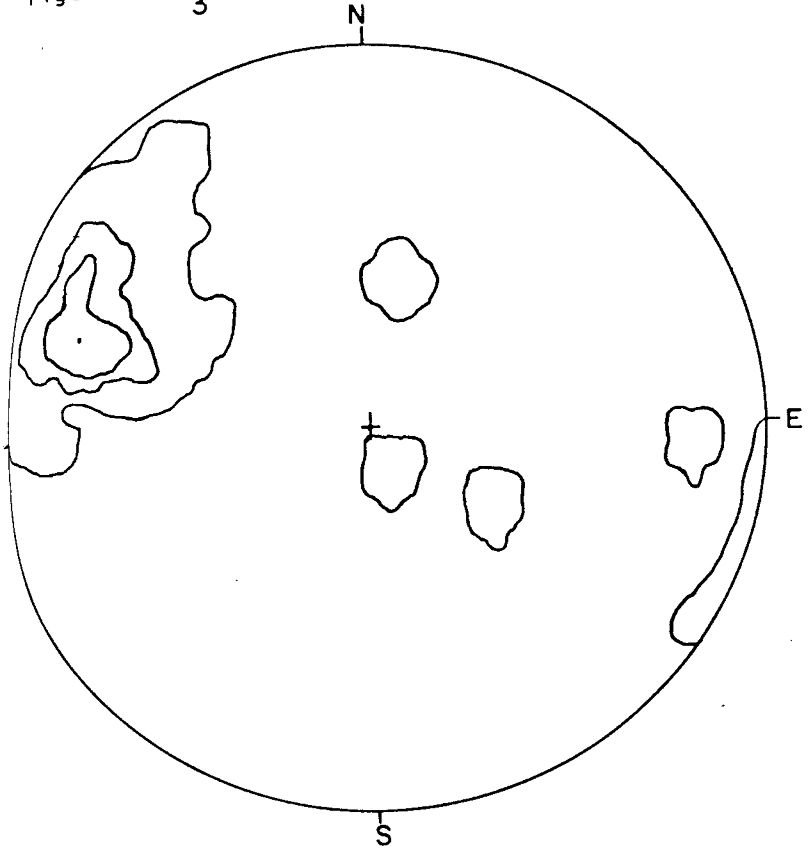
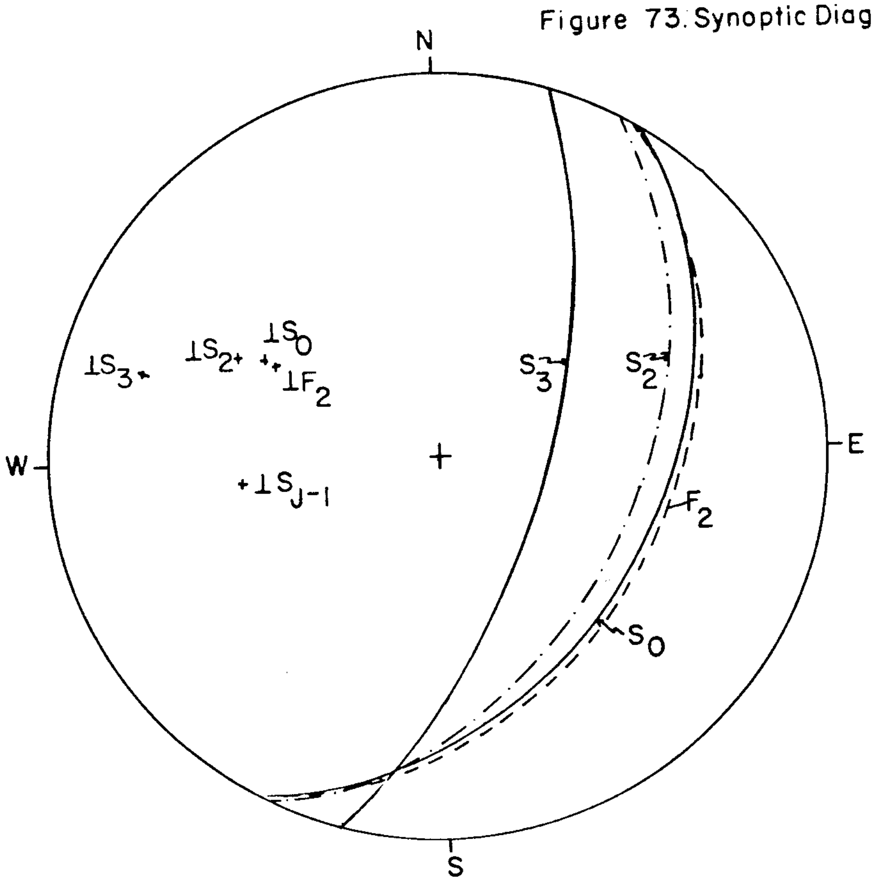


Figure 73. Synoptic Diagram



Part 4: Thrust Faults: Macroscopic

Thrust faults are infrequently described structural elements in the northern Taconics. Previous workers recognized major slice boundary thrusts, but did not map intraslice thrusts in this or in surrounding areas (Zen, 1961, 1964; Theokritoff, 1964; Shumaker, 1967). Jacobi (1977) mapped two intraslice thrusts on the eastern edge of her area (i.e. the western edge of the study area), but her mapping was not detailed enough in this peripheral part of her area to precisely locate them or constrain their timing. The positions of these thrusts have been modified and are here referred to as the Middle Granville and Raceville thrusts.

Three generations of major thrust faults are recognized in this area. The term major here refers to thrusts that influence the map distribution of lithostratigraphic units. They are numbered on figure 74 for convenience in the following discussion.

The placement of thrust faults in this area is based on the following list of criteria.

(1) Pawlet greywackes younging directly into older lithostratigraphic units without an intervening oppositely younging fold limb (1(?), 3(?), 5, 10, 13).**

(2) Truncation of structures, including folds (1, 4, 5, 13, 14, 15), or strike belts (3, 7, 8, 9, 10, 13).

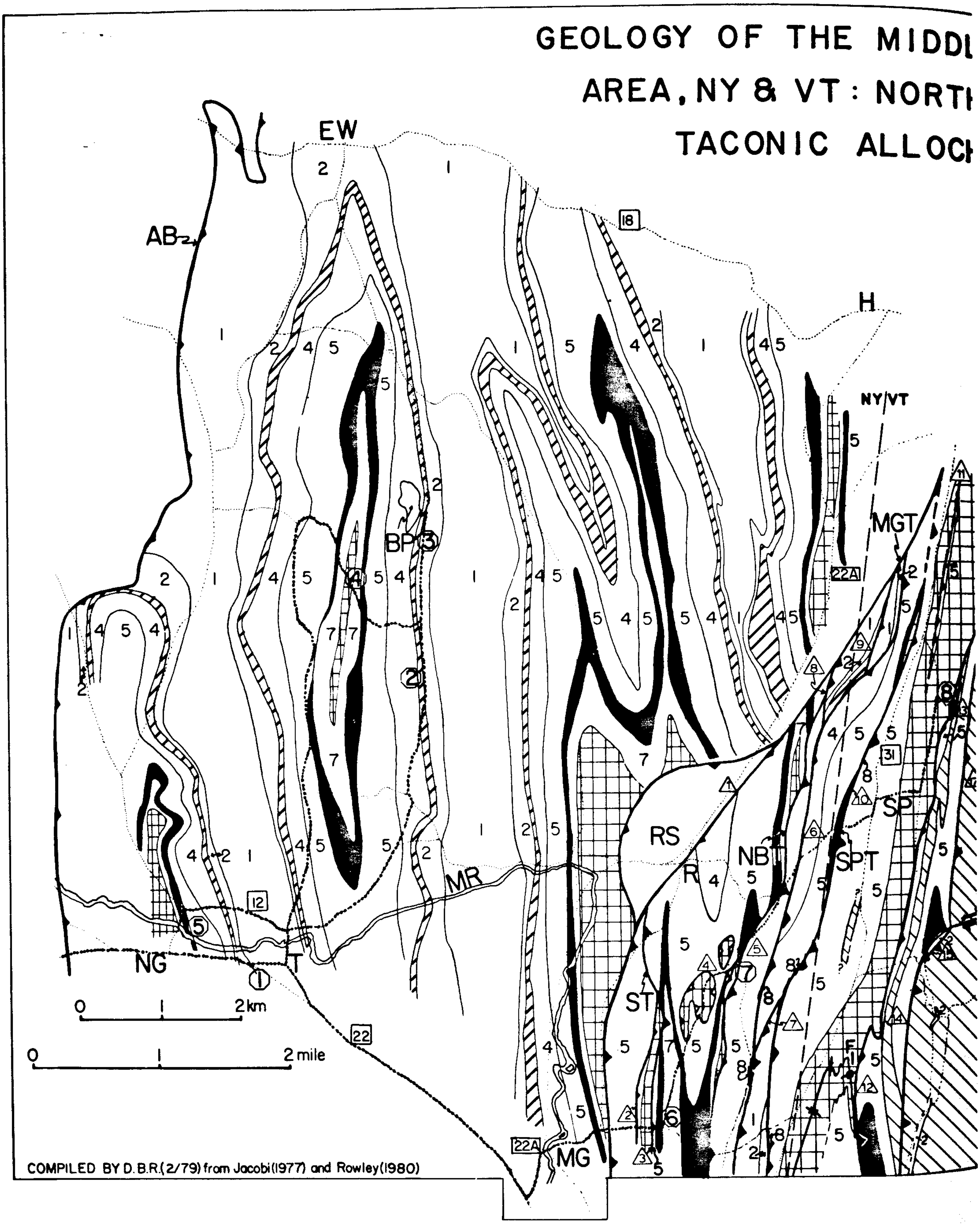
(3) Thin, discontinuously exposed Pawlet greywackes, often only a few meters wide, surrounded by older lithostratigraphic units. Bedding and facing within the greywackes are usually unrecognizable (6, 10, 13).

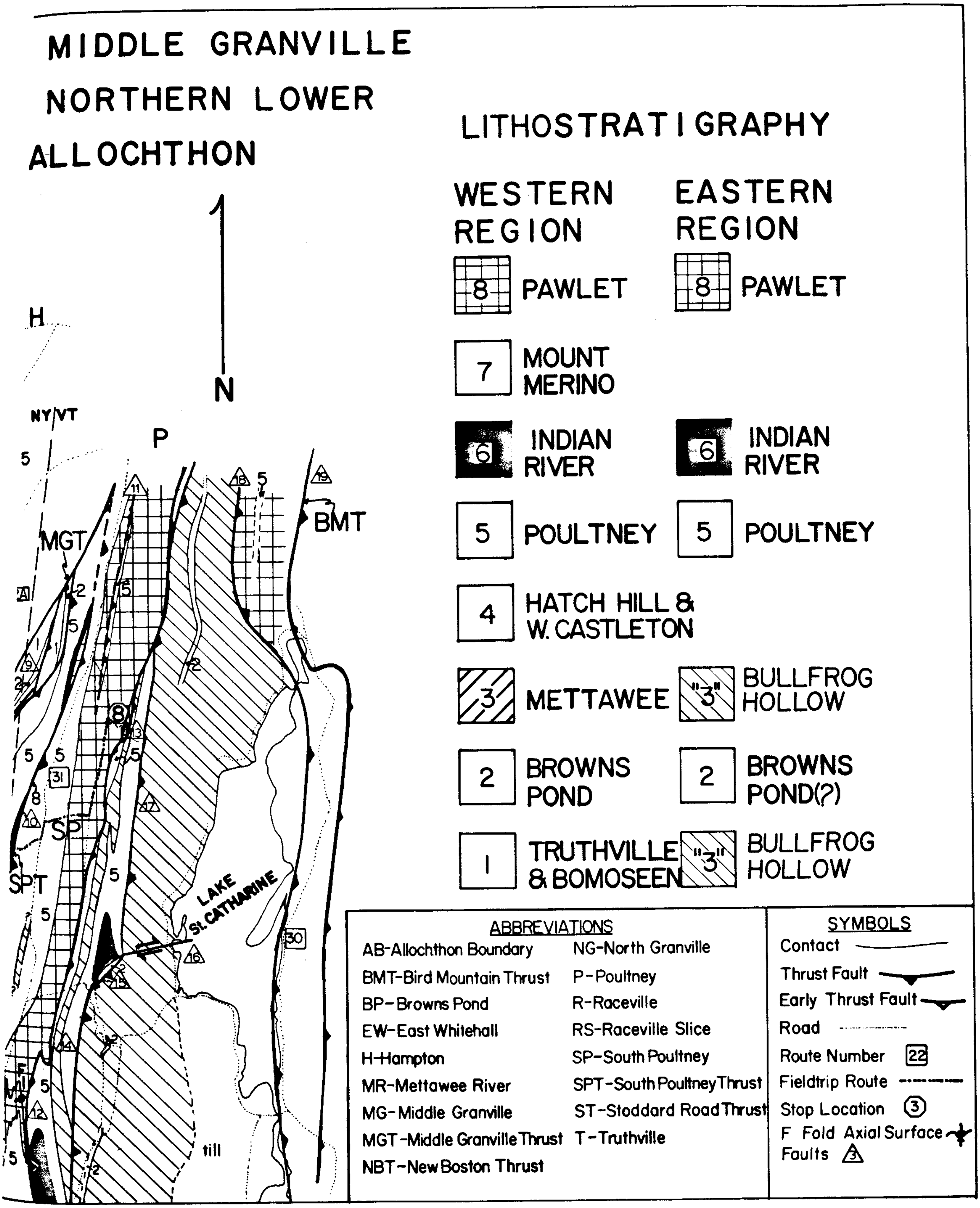
(4) Complexly interdigitated distinctive lithologies from two or more lithostratigraphic units in zones from 2 to 10 meters thick. Fault-

** Numbers refer to thrust faults shown on Figure 74

Figure 74

Compilation map labeling thrust faults from the study area. Map modified from one presented by Rowley, et al., (1979).


GEOLOGY OF THE MIDDL
AREA, NY & VT: NORTH
TACONIC ALLOCH
EW
AB
H
NY/VT
MGT
BP
RS
NB
SP
SPT
R
ST
MR
NG
T
MG
F
0 1 2 km
0 1 2 mile
COMPILED BY D.B.R.(2/79) from Jacobi(1977) and Rowley(1980)


MIDDLE GRANVILLE
NORTHERN LOWER
ALLOCHTHON
N
H
NY/VT
P
MGT
BMT
SP
SPT
LAKE St. CATHARINE
till
LITHOSTRATIGRAPHY
WESTERN REGION
EASTERN REGION
8 PAWLET
8 PAWLET
7 MOUNT MERINO
6 INDIAN RIVER
6 INDIAN RIVER
5 POULTNEY
5 POULTNEY
4 HATCH HILL & W. CASTLETON
3 METTAWEE
"3" BULLFROG HOLLOW
2 BROWNS POND
2 BROWNS POND(?)
1 TRUTHVILLE & BOMOSEEN
"3" BULLFROG HOLLOW
ABBREVIATIONS
AB-Allochthon Boundary
BMT-Bird Mountain Thrust
BP-Browns Pond
EW-East Whitehall
H-Hampton
MR-Mettawee River
MG-Middle Granville
MGT-Middle Granville Thrust
NBT-New Boston Thrust
NG-North Granville
P-Poultney
R-Raceville
RS-Raceville Slice
SP-South Poultney
SPT-South Poultney Thrust
ST-Stoddard Road Thrust
T-Truthville
SYMBOLS
Contact
Thrust Fault
Early Thrust Fault
Road
Route Number 22
Fieldtrip Route
Stop Location 3
F Fold Axial Surface
Faults 3

related lithologies are commonly found associated with these zones (5, 12, 13, 18).

(5) Apparent loss of a limb of a fold, either laterally (3, 9) or longitudinally (13).

(6) Extreme thickness variations and loss of lithostratigraphic units across repeating sequences (12, 13).

All but a few of the thrusts depicted on the map (Figure 74) are proposed from one or more of the criterion listed above. Those not included on the list (i.e., 2, 11, 17) are dotted on the map and are suspected from geometric relations, but independent evidence is not known.

Three generations of thrust faults are recognized in this area on the basis of cross cutting relations. In the following section each generation of thrust faults are described and individual examples of each generation are used to illustrate their character. The generations are treated from oldest to youngest.

Generation 1: D_1 thrusts (T_1)

Only a single example of a first generation thrust fault is recognized in this area (fault #4). This is the fault at the base of the small klippe of Pawlet near the northern end of Stoddard Road. The klippe of greywackes and slates lies in the hinge of a large F_2 anticline. The greywackes are characterized by tight, asymmetric, overturned F_2 folds and moderately east-dipping penetrative slaty cleavage (S_2). Bedding-slaty cleavage intersection lineations and mesoscopic fold hinge orientations are conformable with the regional pattern. At one outcrop, downward-facing bedding-slaty cleavage relations are observed suggesting possible F_1 folds. Locally, the greywackes are discontinuous and apparently (poor outcrop) associated with isolated fold hinges suggesting possible soft-sediment disruption or F_1 or F_2 transposition. No F_3 folds are observed and the S_3 crenulation cleavage is only locally observed, but not particularly well developed.

The time of emplacement of this small klippe is uncertain due to lack of outcrop near the basal thrust, but is most likely associated with D_1. The greywackes cannot lie unconformably above already folded units because (a) a conformable sequence from at least Poultney to Pawlet is observed on the eastern limb of this very same F_2 anticline; (b) the Pawlet greywackes of the klippe are affected by all of the regional deformations seen in the underlying Poultney and other rocks of the F_2 anticline. Syn- to late-D_2 thrusts are characterized by approximate parallelism with F_2 axial planes and not by the flatness of the basal thrust of the klippe. T_3 thrusts are more shallowly dipping than syn- to late-D_2 thrusts (T_2) but are not known to occur as klippe and more importantly do not show evidence of being folded. A D_1 timing is suggested by the following observations, (a) the outcrop pattern of the klippe thrust suggests that the thrust has been folded, most likely by F_2, (b) the restriction of downward-facing bedding-slaty cleavage relations to the klippe suggests that rocks of the klippe were isolated with respect to surrounding rocks during D_1 folding. However, since inverted structural relations were observed at only a single outcrop this relation might equally well be explained by local slump-related overturning of bedding rather than the rocks having experienced an earlier D_1 phase of deformation, (c) the present outcrop pattern suggests that the thrust was initially close to horizontal which is compatible with the apparent pre-F_2 folding orientation of the F_1 folds farther to the east. These observations suggest that the klippe represents a presently isolated piece of a nappe that developed by detachment, possibly in the hinge region of an F_1 recumbent fold in a fashion similar to that illustrated by Heim (1919; see figure 75) from his work in the Alps.

Generation 2: Syn- to late-D_2 Thrusts (T_2)

Syn- to late-D_2 thrusts are northerly-trending, moderately east-dipping structures which closely parallel the D_2 structural grain. These thrusts essentially parallel F_2 axial surfaces or slightly less steeply east-dipping and are believed to be genetically related to the F_2 folds. They are not known to be folded by F_2. Mesoscopic and microscopic evidence outlined above is consistent with a syn- or late-D_2 timing. These data include (a) fault-related rocks characterized by well-developed slaty cleavage defined by preferred orientation of chlorites and white micas and deformed quartz veins. The preferred orientation of the micas is similar to that observed in non-fault-related slates; (b) local development of a homogeneous crenulation cleavage which is similar to the regional S_3 crenulation cleavage both in terms of morphology and orientation. The fact that the crenulation cross-cuts the apparently fault-related folation suggests a pre-D_3 timing of thrusting.

The Stoddard Road Thrust (3) is the best example of a syn- to late-D_2 thrust, particularly in terms of exposure and demonstrable relationship to a F_2 fold. Additionally, thrusts 6, 7, 8, 9, 10, 11, 12, 13, 17?, 18? are considered to belong to this generation. The Stoddard Road thrust juxtaposes an easterly-facing sequence of Poultney through Mount

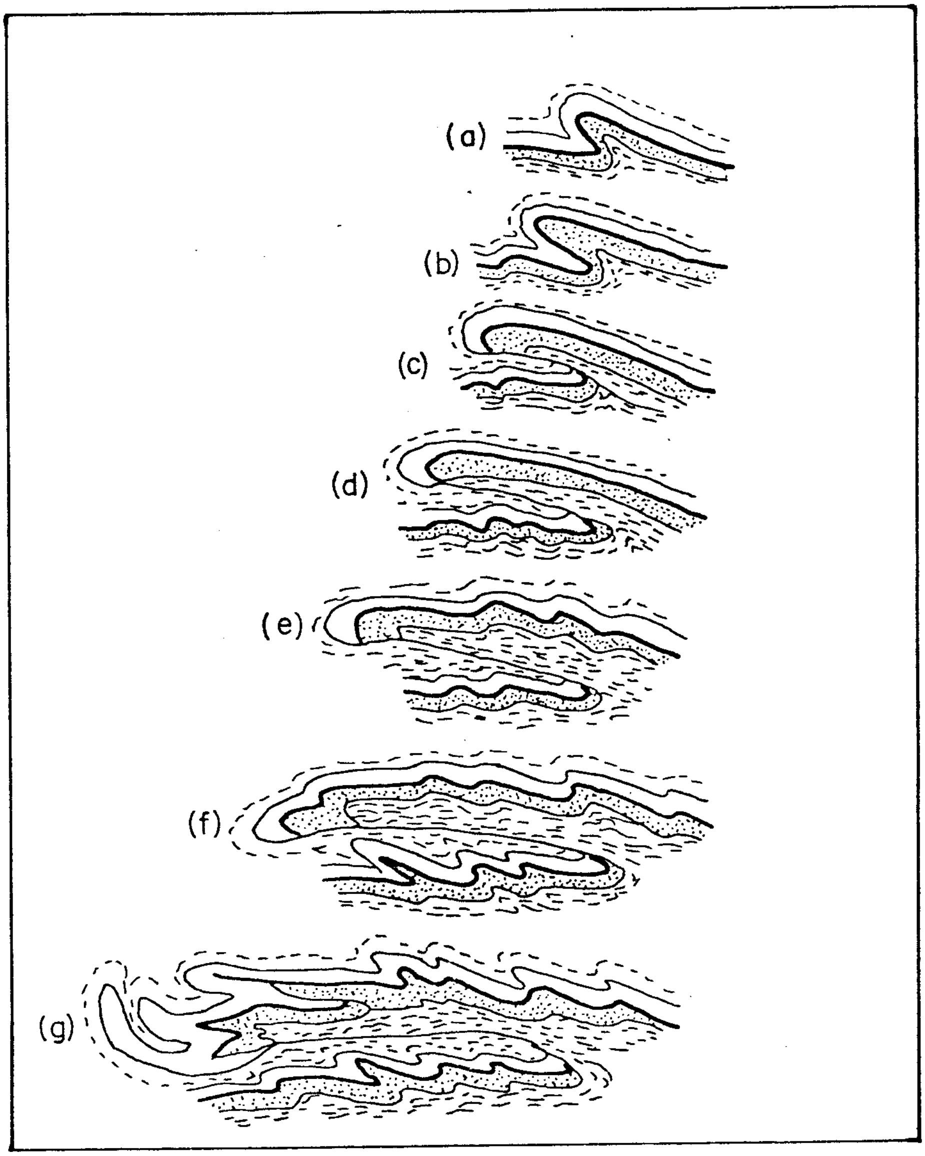


Figure 75

Diagram from Heim (1919) illustrating the progressive development of fold nappes. (a) Fold with overturned limb. (b) Overturned limb thinned. (c) Extreme thinning of overturned limb. (d) Thrust fault replaces overturned limb. Figure taken from Hobbs, et al. (1976, p. 409).

Merino in the lower limb of a large F_2 syncline, against Pawlet. Locally, the Pawlet greywackes may young east into the thrust, but outcrop is not sufficient to adequately demonstrate this. Fault-related lithologies, including slates with lensoid Poultney quartzites, are present on this fault and have been described above and in Rowley et al (1979, Stop 6). The position of this fault is well constrained to the south, but to the north beyond the northern limit of Pawlet outcrop the position is unknown and has been placed in the valley near Butler Road. These thrusts are considered to be similar to the D_1 thrusts in that they are genetically related to associated folds, and form parallel to their axial surfaces. Thrusting probably occurs when (a) shortening can no longer be accomplished by further folding and flattening, and (b) the resolved shear stress across the S_2 slaty cleavage, which is a well developed plane of weakness and anisotropy, exceeded the shear strength of the slates. Mesoscopic and microscopic observations indicate that brittle failure, associated with the development of veins occur at some time(s) during thrusting and at least some of these veins were initially highly oblique to the slaty cleavage. This was followed by ductile straining including folding, flattening, solution transfer, and homogeneous and inhomogeneous shear-strain mechanisms.

Generation 3: Late D_2 to syn-D_3 Thrusts (T_3)

The third and last generation of major thrust faults T_3 is mapped by their truncation of macroscopic F_2 folds and second generation thrusts. The map pattern of these thrusts suggest that they are fairly shallowly east-dipping structures and are oblique to a-1 fold axial surfaces. Faults considered to belong to this generation of thrusts include the Middle Granville and associated Raceville thrust (1), New Boston thrust (5), and thrusts 14 and 15 farther to the east. Fault 18 and the Bird Mountain Thrust (19) may also belong to this generation.

There is no independent evidence that provides constraints for the age of these thrusts, except that they cross-cut D_2 structures. In a qualitative way, at least the New Boston thrust is spatially associated with unusually well developed mesoscopic D_3 folds and crenulation cleavage (Figure 76). This spatial association may mean that these thrusts are temporally associated with D_3 but this need not necessarily be the case, because the other thrusts considered to be of this generation are not known to be associated with atypically developed mesoscopic D_3 structures which would support such a correlation.

The Middle Granville thrust (1) essentially defines the boundary between Jacobi's area to the west and my area to the east. It is a major cross-cutting structure which truncates F_2 folds on both sides of the thrust and is characterized by large alluvial-filled valleys that lack outcrop. Part of the Middle Granville thrust defines the eastern edge of the Raceville slice. The Raceville slice is typified by Poultney-like lithologies, but which have more carbonate and chert associated with them than typical of the Poultney in either Jacobi's or this area. The Raceville thrust truncates a large F_{J-1} syncline on the eastern edge of Jacobi's area.

Other third generation thrusts are similar to the Middle Granville thrust but are not as spectacularly cross-cutting features.

Which generation of thrusts the major allochthonous slice boundary thrusts correspond to is an interesting question, but irresolvable from the present area. The regional relations perhaps indicate that most probably they are equivalent to the third generation. The lack of

Figure 76: F_3 folds near the New Boston Road Thrust.

tectonic slivers of shelf carbonate like those seen in many places on major slice boundaries (Zen, 1967) along the Middle Granville or related thrusts make this correlation uncertain.

Part 5: Map

Mesoscopic and microscopic observations indicate that at least four phases of regional deformation and an earlier phase of soft-sediment deformation penecontemporaneous with deposition have affected the rocks in the study area. Of these deformation phases, only D_1 and D_2 influence the map geometry of lithostratigraphic units and D_2 folds and thrust faults appear to be dominant.

The discussion that follows focusses on the macroscopic structure and variations in structural style and sequence across the study area. In order to better appreciate this variation, the study area is divided into two subareas. The structural style within each subarea is similar and is separated from adjacent areas with different but related, structural style by tectonic discontinuities. Included with this discussion is a brief description of Jacobi's (1977) area. The inclusion of Jacobi's area allows the analysis of a complete transverse geologic section of the Giddings Brook Slice in terms of variation of structural style and sequence. This type of spatial analysis may yield insight into the temporal evolution of structures within lower Taconic slices.

Western Giddings Brook Slice - Jacobi's Area

Jacobi's area extends from the allochthon-autochthon boundary on the west to the Middle Granville Thrust on the east. Two phases of regional deformation, and an earlier phase of penecontemporaneous, soft-sediment deformation are evident. The structure is completely dominated by large, coherent, north-trending, west-facing, overturned folds (F_{J-1}). These folds are tight to isoclinal, with essentially similar morphology. A moderately east-dipping, penetrative, axial surface slaty cleavage (S_{J-1}) is well developed in this area. Major thrust faults only occur along the eastern and western boundaries. Local, minor thrust faults may be present, but nowhere do they grossly disrupt the map pattern. Structures associated with the second phase of regional deformation (D_{J-2}) do not influence the map pattern and are only locally observed. These D_{J-2} structures include open, rounded to angular, upright to inclined asymmetric folds (F_{J-2}) and associated vertical to steeply east-dipping axial surface crenulation cleavage (S_{J-2}), which is best developed on the lower limbs of large F_{J-1} synclines (Kidd, personal communication, 1979). F_{J-1} and F_{J-2} folds are generally coaxial. Late, steep schuppen are locally present, but their offset is never more than a few meters.

The two phases of regional deformation D_{J-1} and D_{J-2} delineated in Jacobi's area and commonly described elsewhere (Zen, 1964; Potter, 1972) are correlated with D_2 and D_3 of the present study. This correlation is based on the close similarity in terms of structural style, specifically fold morphologies and size, nature of the axial surface foliation, and metamorphism associated with cleavage development. Evidence of an earlier phase of regional deformation (D_1 of the present study and noted elsewhere, Zen (1972b), Potter (1972); Wright (1970)) has not been observed in Jacobi's area.

Study Area - Eastern Giddings Brook Slice - Western Subarea

The western subarea in the thesis area lies between the Middle Granville Thrust and the New Boston Thrust. This relatively small area is intermediate in structural style between the structural coherence of Jacobi's area and the structurally dismembered character of the bulk of the study area to the east. This western subarea is characterized by large, tight to essentially isoclinal, gently plunging, asymmetric overturned F_2 folds. Slaty cleavage (S_2) is penetratively developed and axial surface to the F_2 folds. Macroscopic F_1 folds are not recognized, but rare downward facing bedding-cleavage relations are seen in the small klippe of Pawlet near the northern end of Stoddard Road, suggesting possible F_1 folding at least in the klippe.

Major thrust faults, such as the Stoddard Road thrust begin to dismember the F_2 folds. These thrust faults are probably pre-D_3 and syn- to late- D_2 based on their geometric relationship to F_2 axial surfaces, the presence of well developed slaty cleavage defined partly by muscovite and chlorite, and the presence of a homogeneously developed crenulation fabric in the fault-related rocks. This crenulation fabric is correlated with the regional S_3 foliation based on similarity of morphology and orientation.

D_3 structures include open and rounded, to tight angular mesoscopic F_3 folds and associated steeply east-dipping axial surface crenulation cleavage (S_3). Late syn- to post-D_3 schuppen are locally present. The D_3 structures do not influence the map pattern.

The small klippe of Pawlet lying in the hinge region of the large F_2 anticline is a unique occurrence and poses an interesting and yet unresolved problem: When was it emplaced? Possibilities run the gamut from pre-D_1 to D_4 with the most likely timing being syn-D_1, or less likely syn-D_3. Arguments concerning the timing of emplacement of this klippe are elaborated in Part 4.

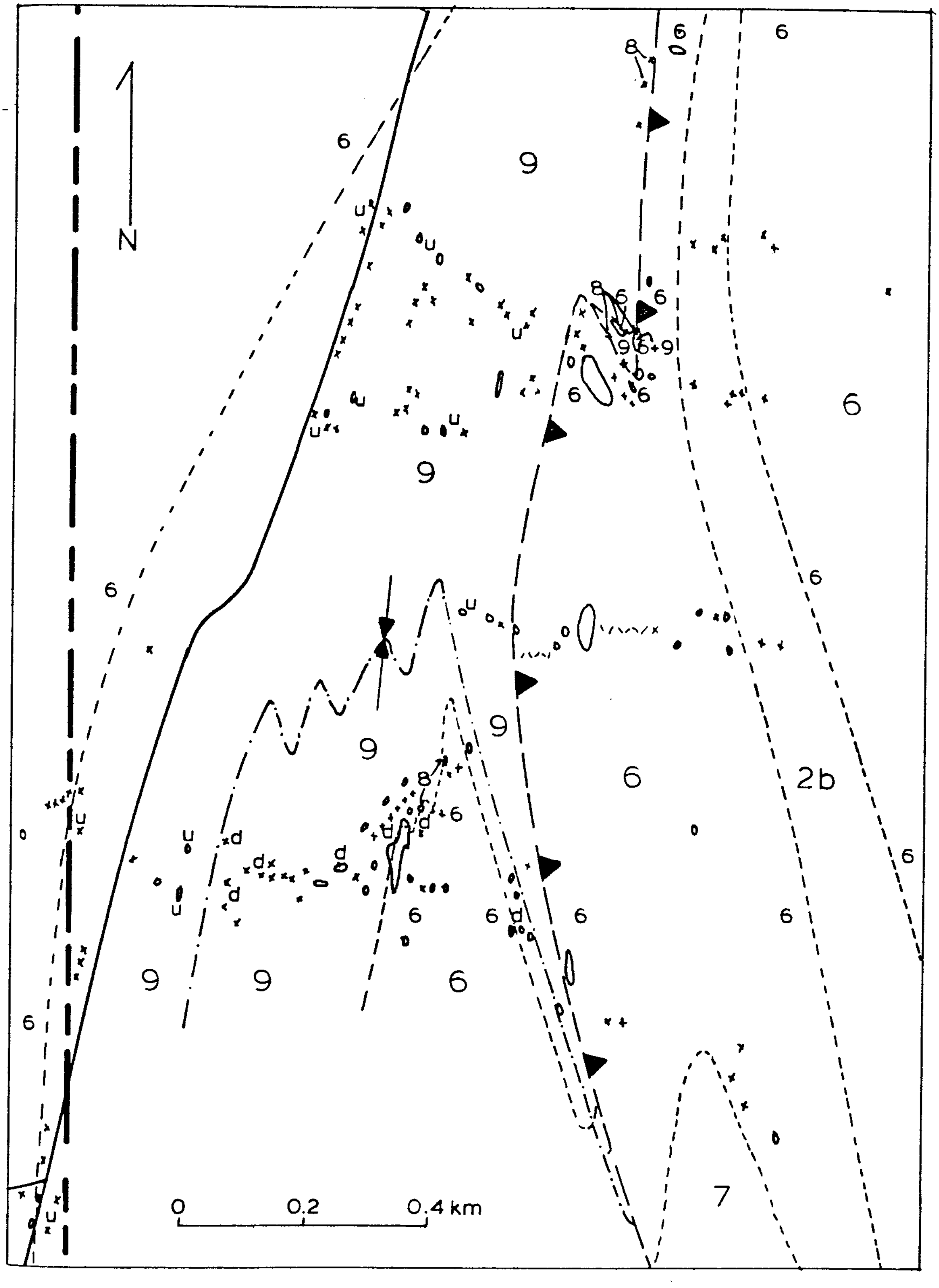
Study Area - Eastern Giddings Brook Slice - Eastern Subarea

The eastern subarea extends from the New Boston thrust on the west to the Bird Mountain thrust on the east. D_2 and to a lesser extent D_1 folds and syn- to late- D_2 thrust faults are responsible for creating the map pattern. The map pattern is characterized by relatively narrow, essentially linear outcrop belts with only a few, generally poorly defined and apparently isolated macroscopic F_1 (Figure 77) and F_2 (Plate 1) fold closures. Caution must be exercised in the macro-structural interpretation of this subarea because of, 1) the lack of distinctive marker horizons, particularly to the east of the South Poultney thrust; 2) the linearity of the outcrop belts with only a few poorly defined fold closures; 3) the presence of unequivocable thrusts; and 4) the common parallelism of bedding, most of which lacks primary facing evidence, and the S_2 slaty cleavage. These observations are compatible with macroscopic dismemberment occurring during D_2 phase of regional deformation. This dismemberment is envisioned to have resulted from tightening of F_2 folds and slicing parallel to F_2 axial surfaces. Boudinage and isolated, intrafolial folds are not commonly observed and are not demonstrable on a macroscopic scale, for this reason the term transposition is not appropriate (See Gregg, 1979b).

Only one macroscopic F_1 fold is still recognizable; it is a fairly large, and tight to isoclinal, and probably recumbent, west-facing syncline (Figure 77). Unfortunately, it is impossible to determine the areal extent of D_1 folding due to the almost complete lack of primary facing evidence in the bulk of the sequence, and the essential lack of Pawlet to the east of Thrust #12 (Figure 74). The area known to be affected by D_1 folding is delimited by the South Poultney thrust and thrust #12 or #13. Tight to isoclinal, overturned folding (D_2) associated

Figure 77

Location map showing outcrop pattern and facing evidence used to construct the position of the axial trace of F_1. u- upward facing structural relationships. d-downward-facing structural relationships. Units 2b- lithofacies-b of the Bullfrog Hollow, 6-Poultney, 7-Indian River, 8-Mount Merino, 9-Pawlet. Dot-dash line- axial trace of F_1 fold.


N
6
9
9
6
6
6
6
9
9
9
8
6
2b
6
6
6
6
6
9
9
6
7
0
0.2
0.4 km

with moderately east-dipping, penetrative, axial surface slaty cleavage, and syn- to late-D_2 thrust dismemberment have almost completely obliterated evidence of pre-D_2 deformation. The lack of primary facing evidence and intensity of D_2 deformation make it difficult to assess the extent of D_1. The presence of defineable macroscopic F_1 folds in the eastern subarea which are not evident farther to the west, may indicate that the deformation front progressed from east to west, or alternatively, that F_1 folding and associated deformation are areally restricted.

The fold and thrust geometry of D_2 is similar to fold-thrust geometries drawn for fold-thrust belts in other regions of the world such as the Front Ranges of the Canadian Rockies (Price and Mountjoy (1970), Himalayas (LeFort, 1975), Southern Appalachian Valley and Ridge (Rodgers, 1964, 1970; Roeder et al, 1979) and at least some subduction-accretion prisms both modern(Moore and Karig, 1975);(Seeley et al, 1974) and ancient (McKerrow, et al 1977).

D_3 structures, including open and rarely tight, rounded to angular, inclined folds and associated crenulation cleavage (S_3) and steep schuppen do not affect the map pattern. Northwest-trending, southeast plunging, vertical D_4 kink bands are also present, but unimportant macroscopically.

CHAPTER 6

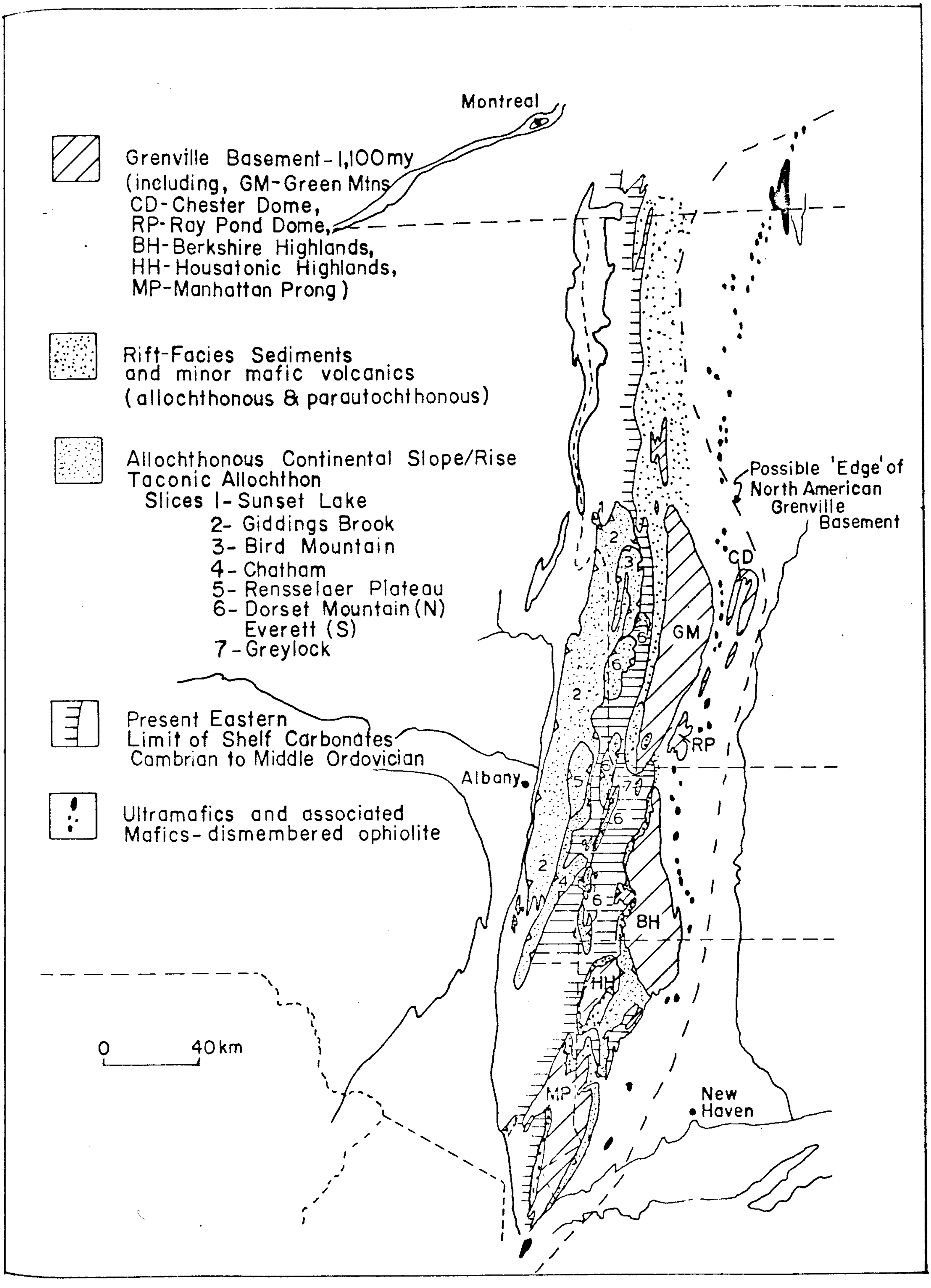
TECTONICS

Pre-Silurian rocks of the Appalachians, extending from Alabama to Newfoundland, show evidence of deformation and, at least locally, emplacement of large allochthonous sheets during the medial to late Ordovician Taconic Orogeny (Rodgers, 1970). In the northern Appalachians, the Taconic orogeny is characterized by the emplacement of stacked and imbricately thrust nappes, of lithologies with continental margin and oceanic affinities, onto a coeval carbonate platform assemblage (Dewey, 1974). The nappe sequence, when complete, usually consists of parautochthonous shelf with or without parautochthonous basement, allochthonous continental rise and slope sediments, often with intercalated slivers of shelf carbonate and/or basement, ophiolite and/or ophiolitic melange, allochthonous and parallochthonous sediments overlying the ophiolites, at least locally unconformably (Casey and Kidd, 1979) and arc terrain. In Newfoundland, a proposed back-arc region is also preserved.

Paleogeographic reconstructions suggest that the shelf-slope-rise sequences represent sediments deposited along an Atlantic-type continental margin that developed in Cambrian (?) time and was destroyed during the Taconic orogeny in medial Ordovician time. The ophiolite, ophiolitic melange, overlying sediments, and arc terrain represent an arc-fore arc terrain developed over an east-dipping subduction zone. Recent plate tectonic corollary modelling suggest that the Taconic orogeny reflects the collision of the North American Atlantic-type continental margin with the east-dipping subduction zone (Chapple, 1973; Rowley, et al 1979; Nelson and Casey, 1979; Hiscott 1978). Collision involved

Figure 78

Geological map of western New England and eastern New York (after Williams, 1978). Possible limit of Grenville basement is shown.


Montreal
Grenville Basement - 1,100my
(including, GM-Green Mtns
CD-Chester Dome,
RP-Ray Pond Dome,
BH-Berkshire Highlands,
HH-Housatonic Highlands,
MP-Manhattan Prong)
Rift-Facies Sediments
and minor mafic volcanics
(allochthonous & parautochthonous)
Allochthonous Continental Slope/Rise
Taconic Allochthon
Slices 1- Sunset Lake
2- Giddings Brook
3- Bird Mountain
4- Chatham
5- Rensselaer Plateau
6- Dorset Mountain (N)
Everett (S)
7- Greylock
Present Eastern
Limit of Shelf Carbonates
Cambrian to Middle Ordovician
Ultramafics and associated
Mafics- dismembered ophiolite
Possible 'Edge' of
North American
Grenville
Basement
CD
GM
RP
Albany
BH
HH
MP
New
Haven
0
40km

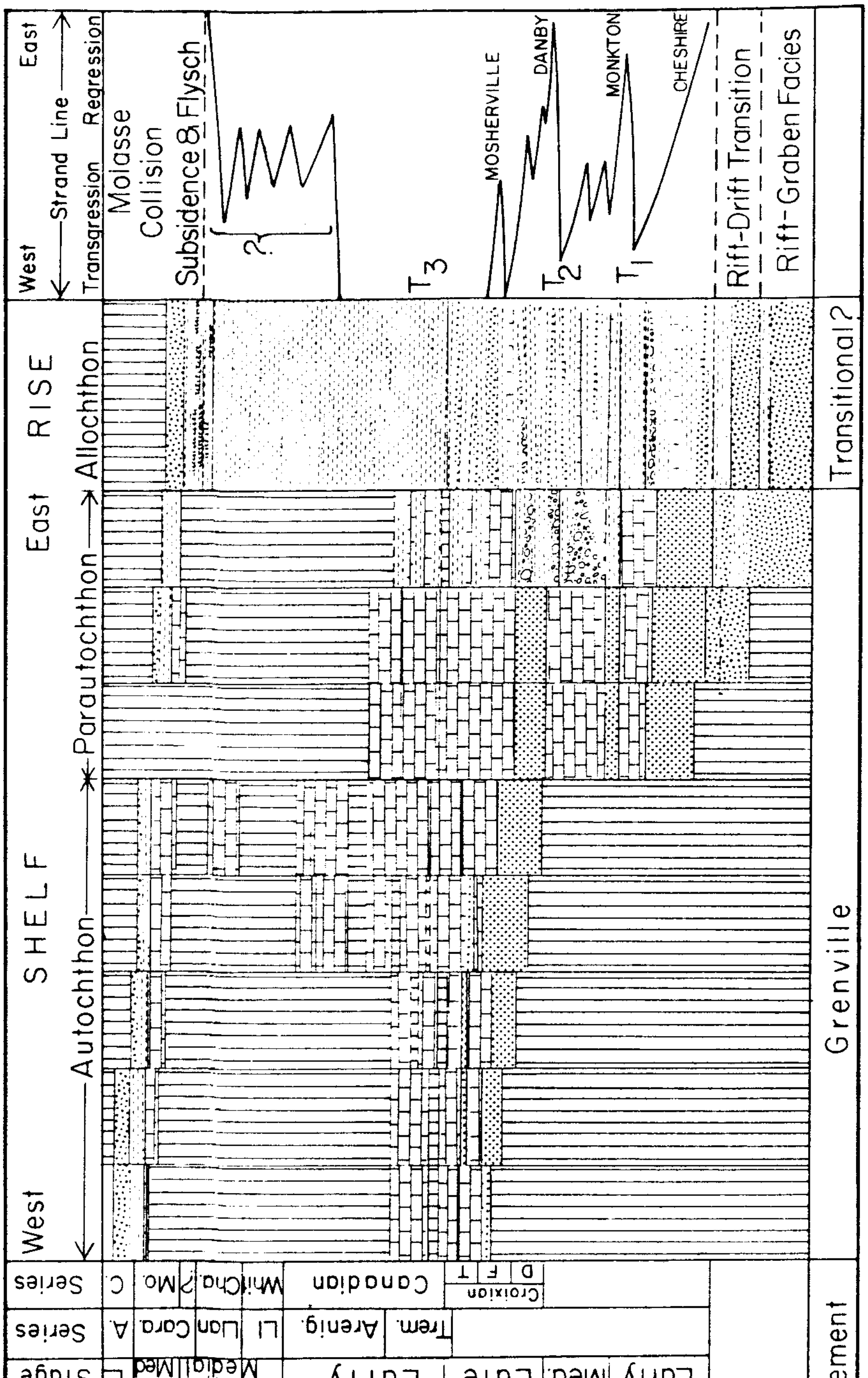
deformation and obduction of ophiolites and continental margin assemblages during the attempted subduction of the continental margin. The Taconic Allochthon is but one of the now discontinuously preserved nappes of continental rise sediments associated with this collisional event.

The purpose of this chapter is to outline briefly the depositional and structural evolution of the Taconic (continental rise) and coeval Champlain and Vermont Valley (continental shelf) (Figures 1 and 78) sequences as it reflects this history of the opening and closing of oceans. This cycle has been termed the Wilson Cycle (Dewey and Burke, 1974).

The stratigraphic sequences that develop on the shelf, slope, and rise on Atlantic-type continental margins are intimately related and strongly reflect the transgressive-regressive history of the margin (Pitman, 1978 and Vail, et al, 1977). An important aspect of this chapter is to illustrate the relationship of stratigraphic sequences developed in the shelf and rise assemblages of the early Paleozoic margin preserved in this part of northeastern North America. This is best accomplished by looking at a series of time-stratigraphic sections through the margin, drawn for different phases in its evolution. The schematic maps and sections of Figure 79 through 84 are drawn at important stages in the development of the margin, including, early rifting (Cambrian (?)) to rift drift transition (Cambrian (?)), early transgression (early Cambrian to late early Cambrian), regression and transgression-2 (late early Cambrian to medial Cambrian), regression and transgression-3 (late Cambrian to late early or early medial Ordovician), 'emergence' and rapid subsidence (medial Ordovician), collision and demise of the ocean and continental margin (late medial Ordovician).

Figure 79

Time-stratigraphic correlation chart based on Plate III. Standard symbols are used. See Plate II for explanation, unit names and location of columns.


West
Autochthon
SHELF
Parautochthon
East
Allochthon
RISE
West
Strand Line
East
Transgression
Regression
Molasse
Collision
Subsidence & Flysch
?
T3
MOSHERVILLE
DANBY
T2
MONKTON
T1
CHESHIRE
Rift-Drift Transition
Rift-Graben Facies
Grenville
Transitional?
System
ORDOVICIAN
CAMBRIAN
Є?
Basement
Stage
L
L. Med
Medial
Early
Early
Late
Med.
Early
Series
A.
Cara.
Llan
Ll
Arenig.
Trem.
Series
C
Mo.
?
Cha.
Whi
Canadian
Croixian
T
F
D

These successive time slices allow a complete iteration of the Wilson Cycle to be documented in some detail.

Early Rifting-Cambrian(?)

The oldest, post-Grenville sediments recognized in western New England are immature, arkosic and feldspathic sandstones, locally interbedded with conglomerates, mafic volcanics, and reddish, micaceous, argillites. These sediments appear to be geographically restricted, and are presently more prevalent to the east, towards the edge of the former continental margin. Rift-facies units include the Nicholville (Fisher, 1977), Dalton (Theokritoff and Thompson, 1969), Hoosac (Norton, 1967, 1976), Cavendish (Skehan et al, 1972), and Pinnacle (Doll, et al, 1961) on the shelf, and Rensselaer (Potter, 1972) and Everett Formations (Ratcliffe, 1969) on the rise. These sequences usually show evidence of shallow water deposition. Similar lithologic assemblages are commonly associated with the early phases of continental rifting (Dewey and Bird, 1970).

Rift-Drift Transition-Cambrian(?)

Overlying the early rift facies, sediments with apparent conformity are cleaner, yet still argillaceous sediments of the Mendon (Thompson, 1967), and 'lower' Cheshire (sometimes referred to as the Gilman (Stone and Dennis, 1964)) quartzites of the shelf, and Bomoseen, 'Zion Hill' wackes, and silty, mica-rich argillites of the Nassau (Potter, 1972) Bullfrog Hollow, and Truthville (Jacobi, 1977) Formations on the rise. The rocks of the rise have a deep water turbdite/slump aspect, suggesting that significant subsidence below wave base had occurred.

Analysis of Mesozoic and younger continental margins indicates that the early sedimentation history is characterized by rapid sedimentation in rapidly subsiding basins. The shelf edges of Atlantic-type margins subside at an exponentially decreasing rate (Watts and Ryan, 1976) similar to that observed for oceanic lithosphere (Sclater, et al, 1971; Sleep, 1971). Subsidence along these margins appears to result from simultaneous operation of two mechanisms, one, thermal contraction (Sleep, 1971) called the 'driving' subsidence by Watts and Ryan (1976) and the other sediment loading (for example, Watts and Ryan, 1976). This early subsidence is associated with rapid sedimentation, and sequences associated with the rifting phase of margin evolution usually constitute 30% or more of the total sections along passive continental margins.

The rift-drift transition also marks the time at which the shape of the margin is defined. The shape of the early Paleozoic margin of eastern North America is unknown, although many people believe that they are able to define it (Rankin, 1976; Thomas, 1977; Williams and Doolan, 1979). The distribution of Grenville basement (Rankin 1976) (Figure 78) and isopach data (Rickard, 1973) (Figure 81) of Cambrian units suggest that a configuration similar to that shown in Figure 78 is possible. This configuration reflects the presence of 'failed arms' along both the north (Burke and Dewey, 1973; Kay, 1975) and south New York State borders, and the presence of Grenville age basement in the Chester Dome (Faul, et al, 1963). However, the shape has probably been significantly modified due to convergence and more importantly large scale underthrusting of continental basement as suggested in the Southern Appalachians (Cook et al, 1979).

The so-called Green Mountain 'anticlinorium' plunges to the north

Figure 80

Schematic model illustrating possible rifting sequence of late Precambrian to early Cambrian history of eastern North America.

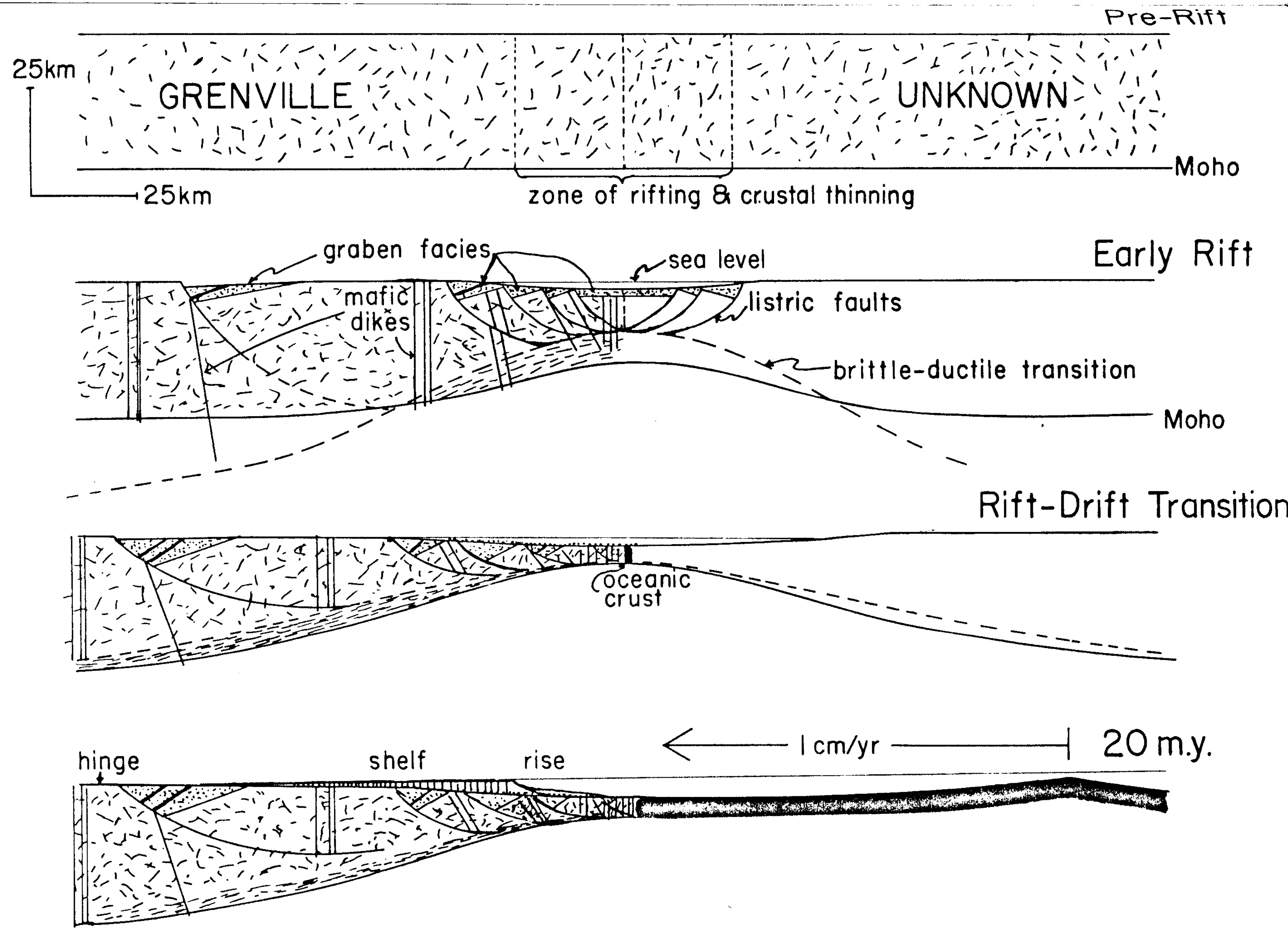
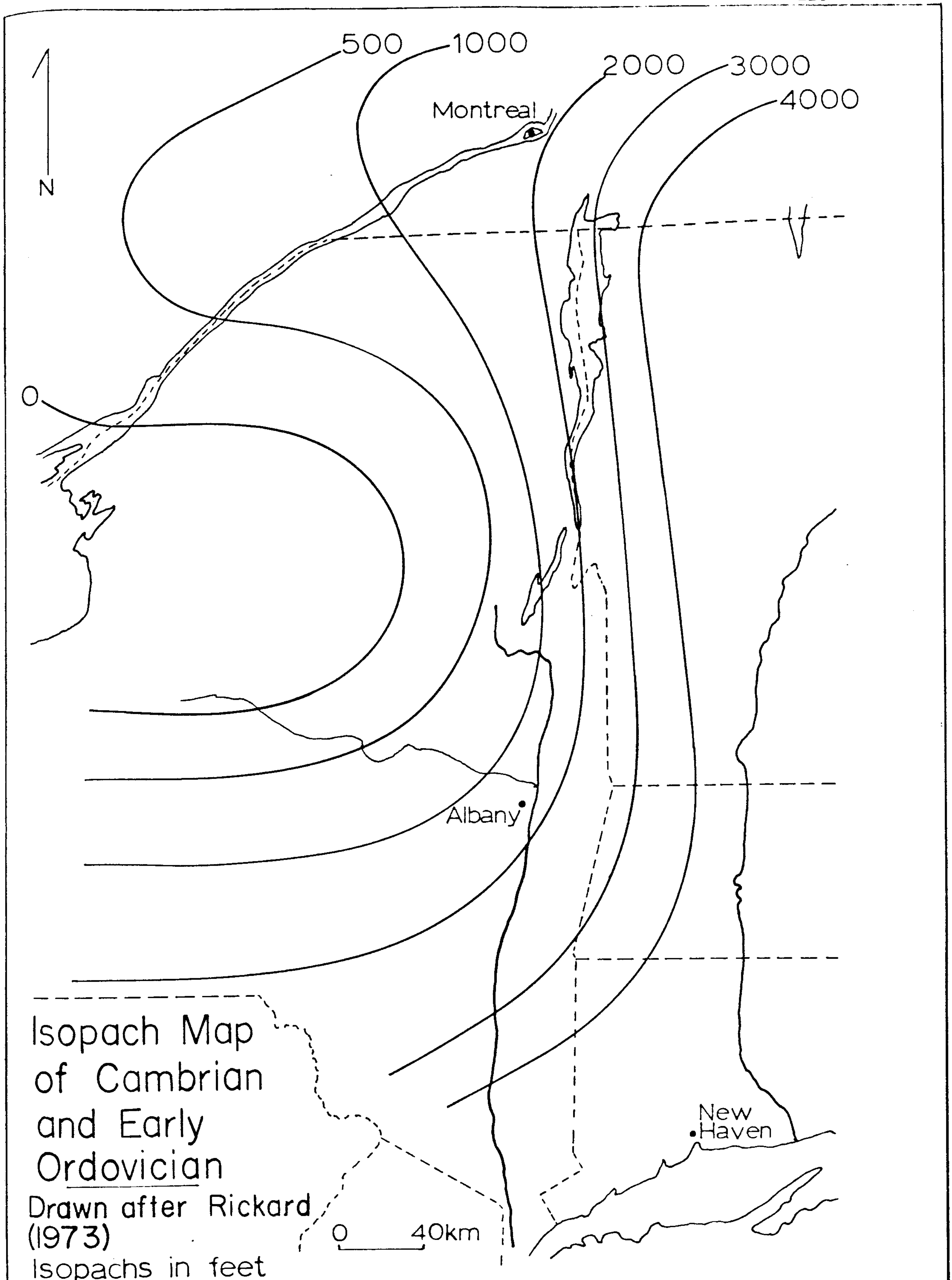

Pre-Rift
25km
25km
GRENVILLE
UNKNOWN
Moho
zone of rifting & crustal thinning
Early Rift
graben facies
sea level
mafic dikes
listric faults
brittle-ductile transition
Moho
Rift-Drift Transition
oceanic crust
1 cm/yr
20 m.y.
hinge
shelf
rise

Figure 81

Isopach map of Cambrian and early Ordovician.



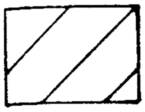
500
1000
2000
3000
4000
Montreal
N
0
Albany
New Haven
Isopach Map of Cambrian and Early Ordovician
Drawn after Rickard (1973)
Isopachs in feet
0 40km

beneath an apparently thick sequence of schists, phyllites and slates, intercalated with meta-greywackes, greenstones, minor quartzites and carbonate of the "Eastern Vermont Sequence' of Doll et al (1961). This succession, including rocks mapped as Tyson through Moretown, is completely unfossiliferous, and yet the rocks have been assigned ages ranging from Cambrian(?) through middle Ordovician (Cady, 1969; Thompson, 1967; Doll, et al, 1961). These rocks are usually described as "eugeosynclinal" and believed to lie farther east than the original site of deposition of the Taconics (Zen, 1967; Cady, 1969). An alternative interpretation is that this sequence is equivalent to the early rift facies (Doolan, personal communication, 1979) and, rift-drift transition sequence of the shelf and rise. Its present geographic position suggests that it may have been deposited in a position off-shelf to the Ottawa-Bonnachere graben (Kay, 1942), which has been interpreted to be a failed arm of a triple-rift system (Burke and Dewey, 1973; Kay, 1975). This is a location in which enhanced sedimentation rates might be expected.

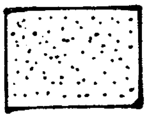
Transgression-1 Early Cambrian to late early (pre-Pagetides zone) Cambrian

The earliest transgressive sequence recognized on the shelf is composed of the Cheshire, or upper Cheshire quartzite and overlying Dunham (Rutland) Dolomite. The Cheshire lies either conformably above Mendon or other rift-drift transition facies or unconformably above Grenville basement. It contains the oldest fossils on the shelf (Theokritoff, 1964; Fisher, 1977). On the rise, the transition from rift-drift facies to 'normal' rise deposits is conformable and sharp and marked by a change from green, silty, mica-rich argillites to black argillites interbedded, in roughly ascending order, with wackes and

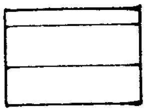
Figures 82 and 85: Lithologic symbols for schematic paleogeographic maps.



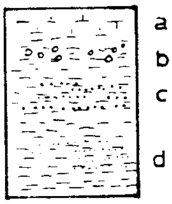
Areas undergoing erosion and non-deposition. Consist of Grenville basement and previously deposited shelf sediments exposed due to regression.



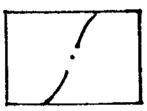
Areas of dominantly shallow marine near-shore clastic sedimentation. Includes orthoquartzites, calcareous and dolomitic sandstones, and lesser sandy carbonates.



Areas of dominantly shelf carbonate deposition. Facies transitions occur on the shelf, primarily changes in degree of dolomitization and increasingly argillaceous carbonate deposition to the east.



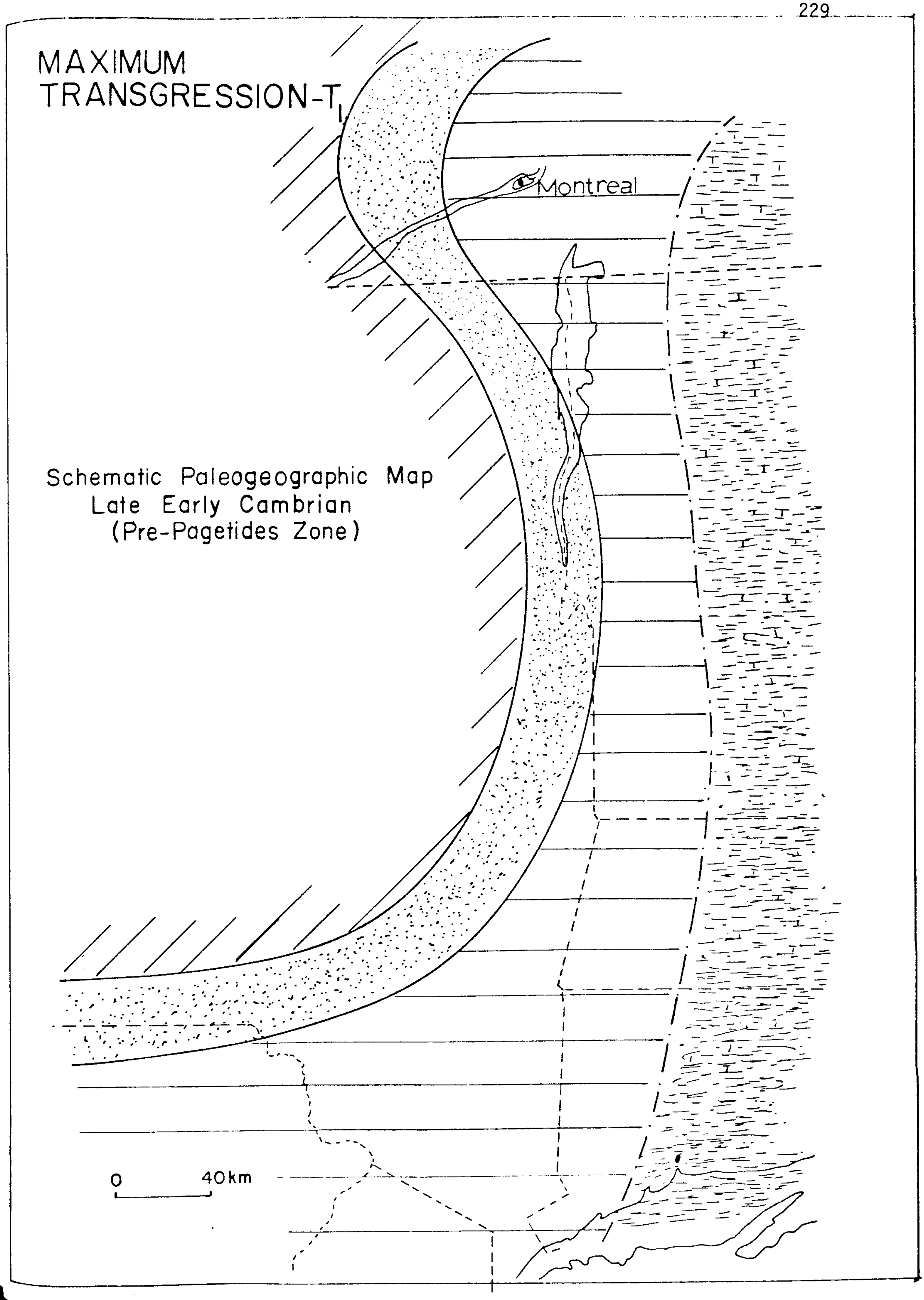
Areas of dominantly argillite deposition beyond the shelf-slope break. Horizontal dashes represent argillite deposition. a) short vertical dashes represent micritic carbonate, b) open circles represent carbonate conglomerates, c) coarse stipple represent arenites and arenaceous carbonates, d) fine stipple represent thin silty quartzites (contourites).



Shelf-slope break

Figure 82

Schematic paleogeographic map for the late early Cambrian (pre-*Pagetides* zone) during maximum transgression T_1.


MAXIMUM
TRANSGRESSION-T1
Montreal
Schematic Paleogeographic Map
Late Early Cambrian
(Pre-Pagetides Zone)
0 40km

clean quartzites of the 'Mudd Pond facies', thin, calcisiltite to calcarenite turbidites, possibly locally reworked by contour currents, and finally carbonate breccias of the Browns Pond Formation. This order mimics the sequence from Cheshire to Dunham on the shelf. In fact, the presence of carbonate within the Browns Pond indicates that carbonate was being deposited on the shelf, as all of the carbonate within the Taconic sequence is believed to be allochthonous in a sedimentary sense.

The oldest fossils within the rise sequence are found in the Browns Pond and date the development of shelf sedimentation (i.e., post-break-up unconformity) as early Cambrian Elliptocephala zone.

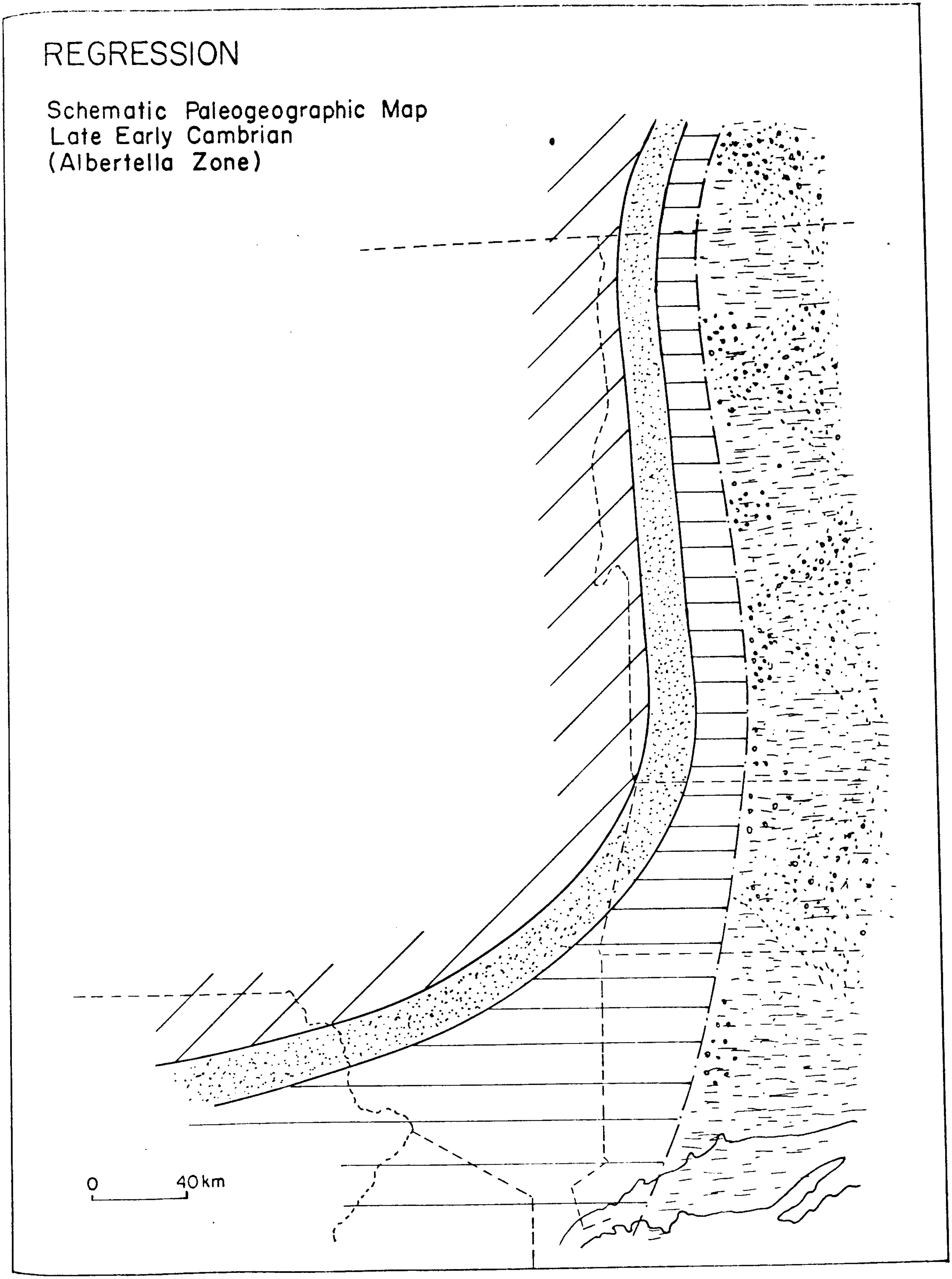
Conformably overlying the Browns Pond on the rise are variegated purple, green and gray porcelaneous slates of the Mettawee ('sensu stricto') and probably upper Bullfrog Hollow. These slates contain thin micritic to calcisiltitic layers and disseminated carbonate (Dale, 1899) and, locally, carbonate breccia (Rowley et al, 1979). The breccia also contains Elliptocephala zone faunas. The variegated slates are interpreted to have been deposited as hemipelagic clays during a period of little sediment input from the shelf to the west. Transgression on the shelf, giving rise to a relatively wide carbonate platform (Dunham) could act as significant barrier to off-shelf sediment transport.

Regression and Transgression-2: Late early Cambrian to early late-Cambrian

The Dunham Dolomite becomes more sandy towards the top (Theokritoff and Thompson, 1969) and is overlain, with apparent conformity, by the Monkton Quartzite. The Monkton consists of red and purple, dolomitic and argillaceous quartzites, interbedded with reddish and purplish argillites and sandy dolomites. Locally, mud cracks are preserved (Palmer, 1971).

Figure 83

Schematic paleogeographic map for the late early Cambrian (Albertella zone) regression.


REGRESSION
Schematic Paleogeographic Map
Late Early Cambrian
(Albertella Zone)
0
40 km

The Monkton is interpreted to be a regressive deposit (Palmer, 1971) that flooded out onto 'distal' parts of the shelf. The color and detritus of Monkton probably reflect its derivation from previously deposited Cheshire and Dunham Dolomites farther to the west. The Monkton belongsto the Pagetides zone of the late early Cambrian.

A marked change in rise sedimentation occurs above the Mettawee, that is characterized by the influx of carbonate and quartz sand, interbedded with black, pyritiferous slates, at least in the north, corresponding to the West Castleton-Hatch Hill, or lower Germantown to the south. The lower part of the Hatch Hill-West Castleton sequence is dated as Pagetides elegans zone of the late early Cambrian (Theokritoff, 1964 as reinterpreted by Rowley et al, 1979), and is equivalent to the Monkton. This suggests that input of coarse detritus on the rise reflects the regressive incursion of coarse quartz and carbonate sand on the shelf (Rowley, 1979).

Keith and Friedmann (1976) studied sedimentological aspects of some exposures of Hatch Hill-West Castleton and Germantown, in detail, and suggested that these sediments were deposited in a submarine fan environment. Their suggestion agrees well with the suggested correlation of these deposit s with regression on the shelf.

There are two aspects of this correlation that are disconcerting. One is the thinness of the Hatch Hill-West Castleton and the other is the presence of interlayered black argillites (i.e. not red or purple) as seen on the shelf. The thinness of the sequence, i.e. 130 meters during a period ranging from late early Cambrian to late Cambrian (Tremadocian), is surprising considering the depositional environment. Several possibilities can be suggested to account for this, including,

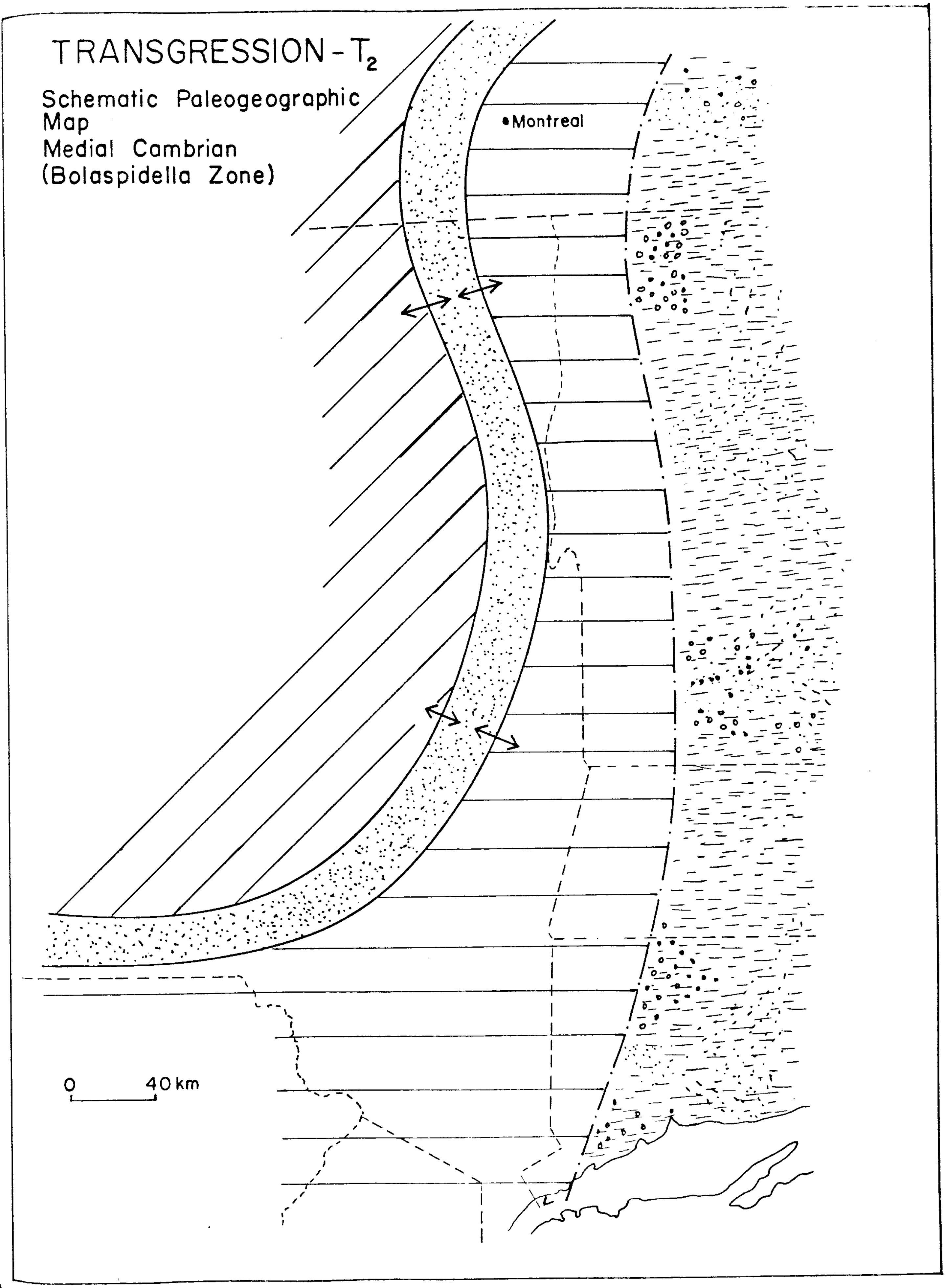
(1) the shale intervals between 'coarse' clastic beds represent long periods of quiescence and slow pelagic deposition; (2) rapid sedimentation occurred followed by either isolation from sediment source or contour current reworking and removal of section, and this may have occurred episodically, in order to account for medial Cambrian fossils (Bird and Rasetti, 1968) within the succession; (3) a third alternative is that appreciable sections of these units were episodically removed by downslope slumping events similar to that documented by the Deep Sea Drilling Project (Scientific Staff, 1976); (4) Some, possibly complex combination of these possibilities. Unfortunately, fossil data is insufficient to really test these different possibilities.

The problem of black argillite versus reddish argillite in supposed source region also poses a problem. The change from red to black reflects the reduction of ferric iron to ferrous iron. In general, sea water is not a sufficiently reducing environment to accomplish this reaction. However, at various times or places oceans have become significantly depleted in oxygen (Ryan and Cita, 1977). These periods are called oxygen minimum crises and may result from a greatly expanded zone of oxygen minimum (Ryan and Cita, 1977). At the present along the east coast of North America the zone of oxygen minimum is shallowest near the continental slope (220-300 m) deepening seaward (800-900 m) (Emery and Uchupi, 1972). Thus the presence of black and not red argillites may reflect either a time of oxygen minima crisis or that deposition of the Hatch Hill-West Castleton coincidentally occurred within the zone of oxygen minimum for that time period.

The Monkton Quartzite is also the basal unit of a transgressive sequence which includes the Winooski dolostone and possibly Parker Shales.

Figure 84

Schematic paleogeographic map for the medial Cambrian (Bolaspidella zone) during transgression-T_2. Arrows indicate smaller amplitude transgressive - regressive events during the medial Cambrian.


TRANSGRESSION - T_2
Schematic Paleogeographic Map
Medial Cambrian
(Bolaspidella Zone)
Montreal
0 40 km

The westerly extent of this transgression and the precise timing of it are unclear due to the lack of fossils in the Winooski and overlying Danby, and later erosion.

A very interesting sequence of units of late early Cambrian to medial late Cambrian (medial Dresbachian) are preserved to the north in the Hinesburg 'Synclinorium'. In the southwestern part of the synclinorium this time period is represented by the Winooski Dolomite which is characterized by dolomites, and, in places, intercalated chert nodules (Stone and Dennis, 1964). To the northeast, above the Dunham, there is a sequence of interdigitated carbonate conglomerates (Rugg Brook, Mill River, Rockledge, and lowest Gorge Formation) and shales (Parker, St. Albans, Skeels Corner, and Hungerford). The sequence from St. Albans through Hungerford contain fossils that range in age from late-medial Cambrian to early Dresbachian (Palmer, 1971), while the Parker shale contains Pagetides zone fossils. Rodgers (1968) interpreted the Rugg Brook conglomerates to be proximal bank edge deposits. This interpretation is extended to this entire sequence including carbonate breccias. Because the Parker Shales (and units above them) overlie Dunham and Cheshire, both of which are interpreted to be shelf deposits, the shelf edge must have retreated prior to Parker deposition. The sloughing off of carbonate blocks into shales may reflect oversteepening of carbonate bank edge, or minor transgressive-regressive fluctuations on the margin. Palmer (1971) cites evidence from the Southern Appalachians that suggest that such fluctuations are recorded, at least there.

Transgression-3: Late Dresbachian to Late Champlainian

The Danby Formation, a sequence of interbedded sandy dolomites, dolomitic arenites, and quartzites is believed to be the basal deposit of the last major transgression prior to the demise of the shelf in medial Ordovician time. The age of the Danby is unknown in precise terms, but is usually assigned either a late medial Cambrian (Stone and Dennis, 1964) or Dresbachian age (Palmer, 1971). The Hatch Hill, which represents the sediments of the rise in late medial to late Cambrian are similarly dolomitic arenites, interbedded with black shales. Thus another possible correlation exists between rise and shelf sedimentation.

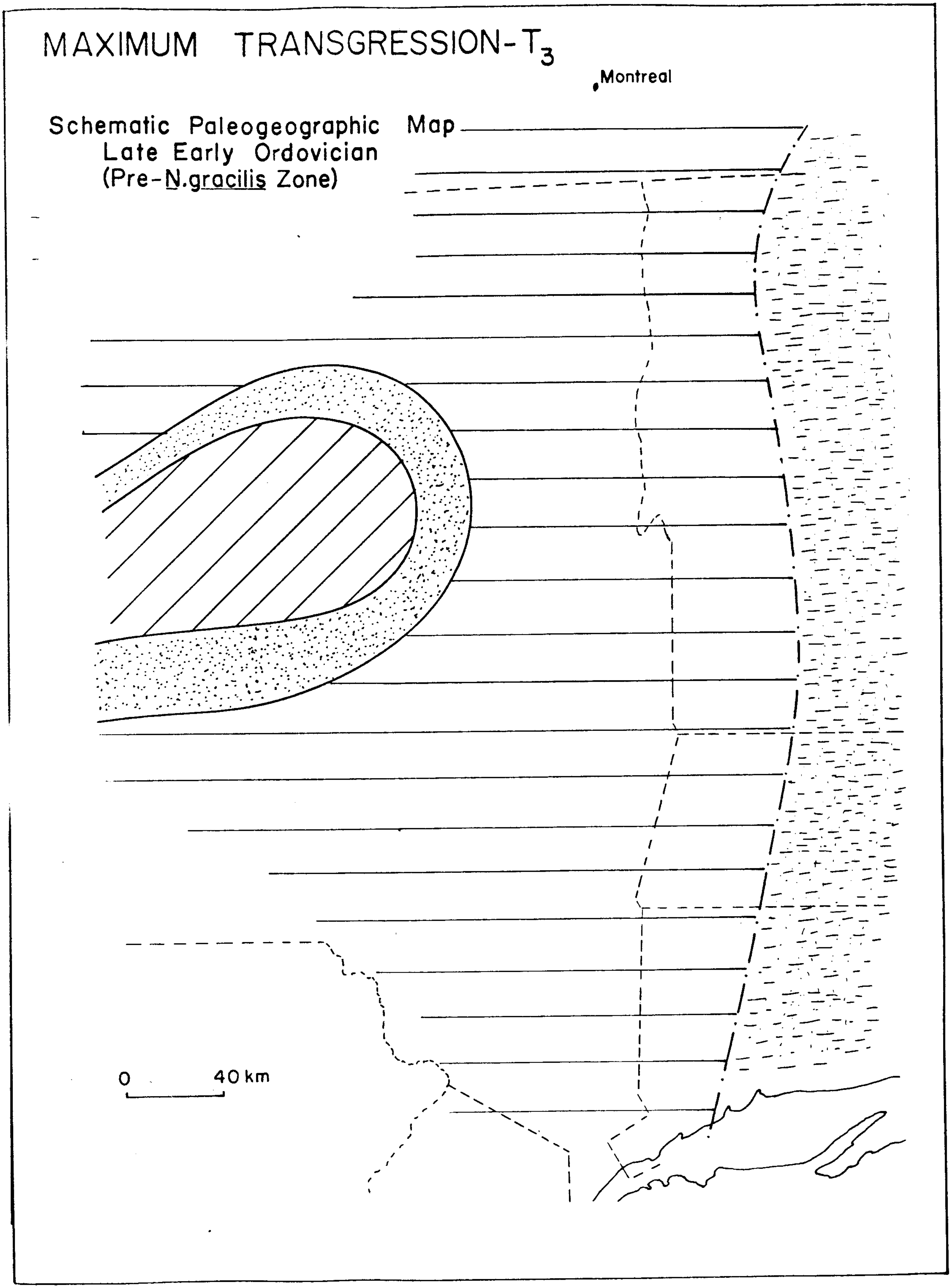
The transgression that began in the Dresbachian (Palmer, 1971) includes deposition of the Danby, Potsdam and Galway basal sands, above which were deposited a thick section of carbonates including Clarendon Springs to Bascom to the east, and Ticonderoga through at least Bridgeport to the west. This transgression extended to Canojoharie, New York by medial Franconian. Near the shelf edge, Hungerford Shales and lower Gorge Formation carbonate conglomerates are overlain by massive dolomites of the upper Gorge, suggesting the upbuilding and out-building of the carbonate shelf. The upper Gorge and overlying Highgate Formation of Canadian Age (Stone and Dennis, 1964) are probably outer shelf sediments.

The upper part of the Hatch Hill is characterized by more abundant black argillites and less frequent and thinner quartzite (Rowley et al, 1979) which passes upward into dark argillite and calarenite and/or calcisiltite of the lower Poultney. This progression is interpreted to reflect the progressive cutting off of clastic debris from the shelf as the Dresbachian transgression proceeded westward.

The bulk of the Poultney is characterized by variegated, though mostly medium gray argillite and thin, silty, very well sorted, quartzites. The medium gray argillite is interpreted to have been deposited

Figure 85

Schematic paleographic map for the late early Ordovian (pre-*N. gracilis* zone) during maximum transgression-T_3 and prior to destruction of the continental margin.


MAXIMUM TRANSGRESSION-T3
Montreal
Schematic Paleogeographic Map
Late Early Ordovician
(Pre-N.gracilis Zone)
0
40 km

as *in situ* and reworked hemipelagics. Interbedded with these argillites, often on a very fine (laminated) scale, are thin silty, well sorted quartzites suggested to be contourites (Jacobi, 1977). Together, these sediments suggest that little or no coarse sediment was being supplied from the shelf to the west. This has been interpreted to result from sediment trapping on a wide shelf (Rowley, 1979). The Poultney contains graptolites and conodonts that range from *T. approximatus* zone to *G. dentatus* zone. Rowley et al (1979) suggested that it probably extends as high as *C. angustatus* or possibly into *N. gracilis* zone, but because of unsuitable host lithologies fossils have not been found.

The post-Providence Island, pre-Chazy history of the eastern part of the shelf is not well known because of lack of units due to later erosion A disconformity is commonly placed below the Chazy to east. There is no recognizable input of clastics on the rise at this time, suggesting that a large regression did not occur during this time.

'Emergence' and Submergence: Upper Champlainian (Llandeilian)

In post-Chazy time the carbonate shelf experienced an extensive erosion and/or a disconformity. At approximately this time bright red argillites appear on the rise. Bird and Dewey (1970) suggest that these two events are intimately related by the erosion of *terra rossa* soils on the karst. The cause of this apparently rapid 'emergence' of the shelf is unknown, but several suggestions have been put forth. Chapple (1973) suggested that it represents the passage of the continental shelf through the bend fold commonly observed oceanward of oceanic trenches (Figure 86). A second possibility is that it results from rapid changes in sea level due to glaciation on the Saharan craton (Fairbridge, 1971).

Within the Indian River are thin pale gray cherty horizons interpreted to be silicic tuff bands (Rowley et al, 1979). Bentonites first appear on the shelf, within the lower Black River (D. multidens zone) (Rickard, 1973). These tuffs are interpreted to herald the approach of an volcanic arc from the east (Rowley, et al, 1979).

The Indian River passes conformably upward into black cherts and argillites of the Mount Merino, which are well dated as N. gracilis zone (Berry, 1962). The cherts of the Mount Merino have been interpreted as volcanogenic in origin by Lang (1969). On the shelf to the west, this post-Chazy period is characterized by deposition of argillaceous black limestones of upper Black River age (D. multidens). These are the Whipple, Orwell, and Amsterdam. To the east, on the outer part of the shelf, the limestone is quickly overwhelmed by black shales of the Walloomsac Formation, suggesting rapid submergence of the shelf edge. Farther west, limestone deposition continued, suggesting that submergence transgressed to the west. This 'submergence' may reflect the downbending of the shelf and rise into a subduction zone to the east or rapid rise of sea level due to waning of the glaciation on the Africa, or the combined influence of the two.

Collision: latest Champlainian to upper Mohawkian (Uppermost N. gracilis to C. Spiniferus)

Plate tectonic corollary models for the emplacement of the allochthonous continental rise (Taconics) sediments onto the autochthonous and parautochthonous continental shelf ("synclinorium") sequence are of two varieties. The earliest model, proposed by Bird and Dewey (1970, 1975) and reiterated by others suggest that the initiation of a west-dipping subduction zone led to the following related events.

(1) Development of an Andean-type volcanic arc in the outer continental rise region (Ammonoosuc volcanics), (2) uplift of the continental rise terrain to the west of the volcanic arc, (3) collapse of the continental shelf and development of the 'Normanskill Exogeosyncline,' (4) soft-sediment gravity sliding of the low Taconics in the Middle Ordovician and, (5) progressive emplacement of the high Taconics by 'hard-rock' thrusting during late Middle and Late Ordovician. According to this model the Taconics were assembled in their present location and subsequently folded with the underlying shelf sequence in the Middlebury Synclinorium, as first suggested by Zen (1961).

Chapple (1973, 1979) and Rowley and Delano (1979) proposed an alternative type of plate tectonic corollary model. They suggest that the emplacement of the Taconic Allochthon resulted from partial subduction of the Atlantic-type North American continental margin in an east-dipping subduction zone. This model is similar to that proposed by Gealey (1977) for the Oman, and Nelson and Casey (1979) for western Newfoundland. Chapple (1973)argued that the post-Chazy history of the continental shelf, involving uplift, erosion and development of a karst surface, followed by block faulting and rapid subsidence, accords well with the trajectory of crust entering subduction zones. Chapple (1973) followed previous models by suggesting that the low Taconics were emplaced as soft-sediment gravity slides prior to emplacement of the high Taconics slices by subduction accretion 'hard-rock' thrusts. Rowley and Delano (1979), however, argued that structural evidence from within the low Taconics does not support a soft-sediment gravity slide origin, but instead indicates emplacement by 'hard-rock' thrusts. Rowley and Delano (1979) and Chapple (1979) propose that most of the deformation associated with $D_{1\&2}$ occurred during subduction-accretion tectonics as the accretionary prism of the island arc overrode the continental rise. The allochthonous continental rise sediments were therefore already assembled as a series of thrust packages prior to overriding of the outer continental shelf edge. Independently, Rowley and Kidd (in prep.) and Chapple (1979) proposed that continuing convergence between the two lithospheric plates resulted in: (1) overriding of the shelf edge by the Allochthon, as a composite thrust sheet, not as a detached slide block(an argument first made by J. Sales (1971), and incorporation of carbonate slices along the base of the Allochthon (e.g., Dorset Mountain); (2) thin-skinned shortening, by folding and thrusting within the carbonates (i.e., Sudbury Nappe and early folding of the Middlebury 'Synclinorium' (Voight, 1965, 1972)), (3) late imbrication and open folding (D_3) of the Taconic Allochthon to form the presently defined major slice boundaries, as demonstrated by the presence of thin carbonate slivers and locally Grenville basement (e.g., Ghent Block, Ratcliffe and Bahrami, 1976) along all major slice boundaries, and (4) westward-directed thrusting involving crystalline basement, of the Berkshire Massif (Ratcliffe, 1965, 1969, 1977) and the Green Mountains (maps of Hewitt, 1961; and MacFayden, 1956); these thrusts progressed sequentially outward to the Champlain Thrust and other thrusts of the shelf carbonate sequence in front of and along-strike with the edge of the Allochthon (Coney, et al., 1972; Fisher in Rodgers and Fisher, 1969).

A major difference between the models of Bird and Dewey (1970, 1975) versus Chapple (1973, 1979) and Rowley, et al (1979) is the presence or absence of a suture between the volcanic arc (Ammonoosuc) and the eastern edge of the Atlantic-type North American continental margin. Bird and

Figure 86

Schematic Evolution of the Taconic Orogeny

Sections:

(A) Atlantic-type continental margin of eastern North America during early to medial N. gracilis time, the first evidence of a volcanic arc to the east is seen in thin tuff bands in the Indian River Fm. and then volcanogenic cherts (Mount Merino Fm.).

(B) Trajectory of oceanic lithosphere entering a subduction zone. Possibly also applicable to continental lithosphere. On the bulge and to the east crust is in an extensional regime. If this trajectory is applicable it could explain the sub-Orwell unconformity and the so-called Tinmouth orogeny (Chapple, 1973).

(C) Taconic rocks already incorporated in accretionary wedge due to eastward subduction of the Atlantic-type continental margin. Deposition of black argillites on the rapidly subsiding eastern and central parts of the continental shelf. Continued carbonate deposition to the west. Approximately late N. gracilis to D. multidens zone time.

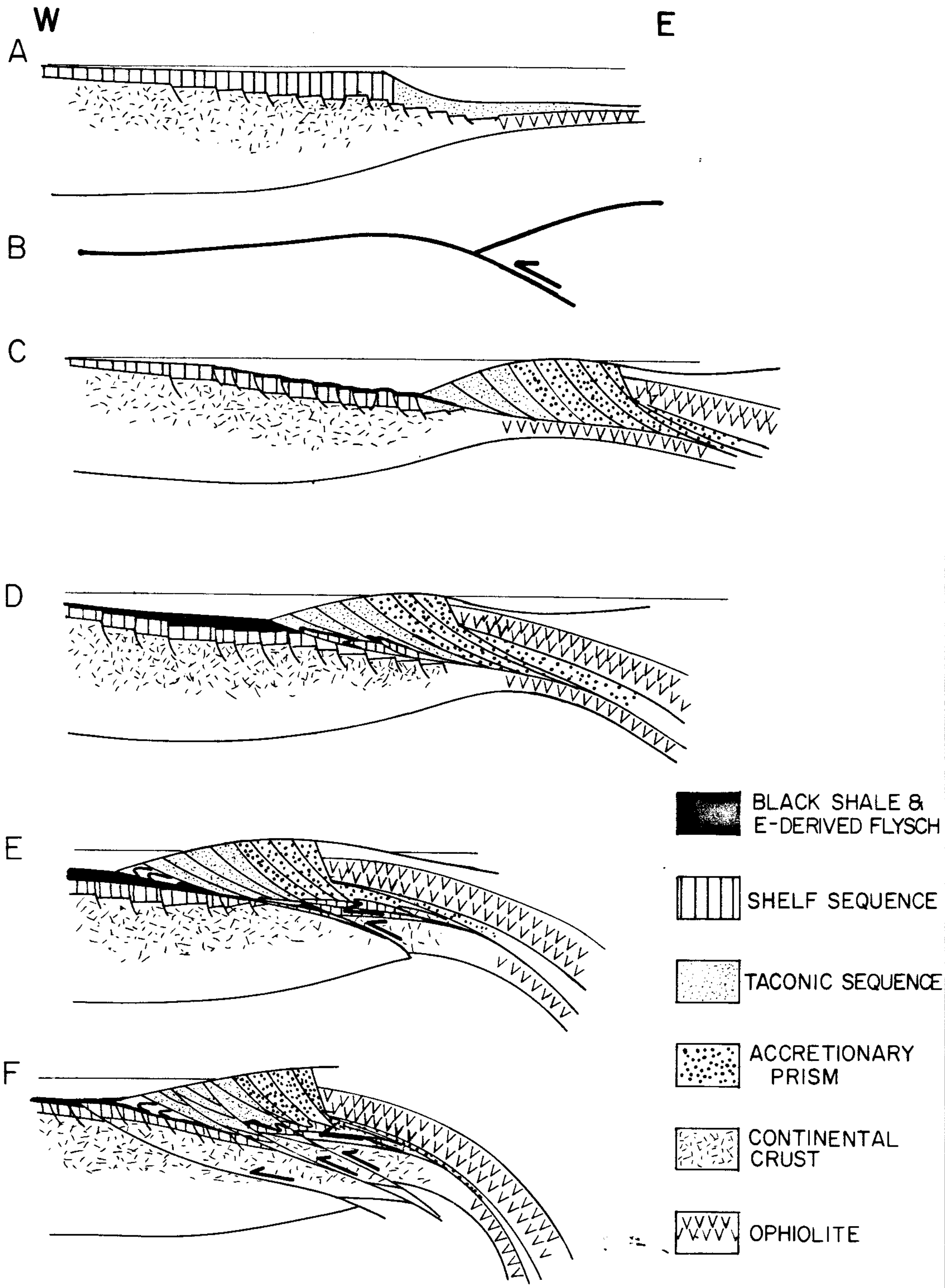
(D) Continued subduction, now involving underthrusting of continental basement and shelf. Peeling of cover sequence from basement along deeper parts of the overthrusting wedge (Sudbury Nappe, base of Dorset Mountain). Folding of early thrust surfaces as active thrust surface migrates westward and considerably less rapidly downward. Approximately D. multidens zone time.

(E) Continued convergence of lithospheric plates, shortening within the continental crust (due to buoyancy effect of attempting to subduct light bouyant continental crust) gives rise to initial imbrication of the continental crust. Subduction rates lessens as more continental crust is underthrust. Melanging of parautochthonous Austin Glen in front of advancing composite thrust wedge. Approximately C. americanus zone time.

(F) Progressive westward imbrication of the basement along listric thrust surfaces. Imbrication of the basement results in late imbrication of the overlying shelf and allochthonous Taconic sequences. Final emplacement of the Allochthon. Approximately O. ruedemanni zone time. Basement thrusts correspond to the Hoosic Thrust (Norton, 1976), Beartown Mountain Thrust(Ratcliffe, 1969), Champlain and/or Orwell Thrust (Coney, et al., 1972; Zen, 1972) in front of and to the north of the Allochthon.

Presently defined Taconic slice boundaries result from thrusting illustrated in sections D through F of Figure 85.

Schematic Tectonic Evolution of the Taconic Orogeny



D.B.R. (5/79)

Dewey accepted the 'dogma' that a continuous Cambrian (?) to Medial Ordovician, time-stratigraphic, "eugeosynclinal", sequence is present in the Eastern Vermont Sequence, equivalent to the Taconic sequence, and thus did not place a suture between the arc and continental margin. The second model requires a suture within the Eastern Vermont sequence. We prefer to place the suture along the Vermont Ultramafic belt, probably to the east of the Chester Dome in the Connecticut River "Synclinorium." This requires that at least in parts of the Eastern Vermont sequence where ultramafics are present, a continuous time-stratigraphic sequence does not (Gregg, 1975; Nisbet, 1976) and cannot exist (as pointed out in Burke, Dewey and Kidd, 1976).

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