STRUCTURE AND STRATIGRAPHY
OF
WEST HAVEN, VERMONT

A thesis presented to the Faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of
Master of Science

College of Science and Mathematics
Department of Geological Sciences

Christoph K. Steinhardt

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ABSTRACT

Mapping and structural investigations in an area bounded to to the west by by Grenvillian basement and to the east by allochthonous rocks emplaced during the medial Ordovician Taconic Orogeny led to division of the area into a Western Undeformed Zone and an Eastern Deformed Zone. The former consists of a thin, undeformed shelf sequence of upper Cambrian to lower Ordovician (Canadian) clastics and carbonates, which unconformably overlie Grenvillian basement and dip gently to the east.

The Eastern Deformed Zone consists of early to medial Ordovician carbonates folded about northnortheast trending axes with east to southeast dipping axial surfaces, overlain by likewise folded and sheared medial Ordovician shales. The rocks in the Eastern Deformed Zone form three eastward dipping, imbricated thrust sheets, from west to east the West Haven, the Forbes Hill, and the Carver's Falls Thrust Sheet. Evidence for east over west thrusting is presented. The initiation of folding predates thrusting, but further folding has probably occurred during the transport of the thrust sheets. Folding and thrusting are interpreted as early and late stages of tectonic movements caused by the emplacement of the Taconic Allochthon.

The consistent similar orientation of thrusts and axial planar cleavage and the large size of the carbonate sheets led to this new interpretation which contrasts with the existing one which viewed the Eastern Deformed Zone and its continuation along strike as an olistostrome.

Displacement estimates based on shelf geometry yield results on the order of 100 km for the thrust sheets in the field area. It is proposed that the boundary between the undeformed and the deformed part of the field area constitutes one of the southern continuations of the Champlain Thrust, the

location of which was hitherto unknown that far to the south. Post-thrusting normal faulting in the field area may possibly be an expression of reactivated rift related basement faults in geologically recent time.

ACKNOWLEDGEMENTS

The writer is indebted to Winthrop D. Means for his encouragement to undertake degree studies after a semester as an exchange student in Albany. I sincerely thank William S.F.Kidd for his guidance and interest in my work during the various stages of the preparation of this thesis. He also suggested the area chosen for this study and introduced me to the field. George W. Putman helped with discussions about recent tectonics in the Champlain Valley. Kevin C.A.Burke provided insightful comments and some financial assistance and Dave B.Rowley helped with field observations and discussions of the regional geology. I want to thank my fellow graduate students for their help and support, especially Mark Jessell,Mauricio Roma, Ricardo Lopez-Torrijos, Tim Kusky, Suzanne Baldwin, Steve Tanski, Dave Bonner and Michelle Aparisi, without whose friendship this thesis would not have been possible. The same is true for the typing efforts of Sharon Poissant.

I owe greatest gratitude to the townspeople of West Haven for their open welcome and interest in my work. Especially I want to thank Bill Bishop and his family for their generous hospitality, and Ruth Best for long conversations and maple walnut icecream.

Financial support for this thesis was received through a Quadrille Ball scholarship and a fellowship from the "Studienstiftung des deutschen Volkes".

Last but not least I want to express my deepest thankfulness to my parents, for the many ways in which they have helped me and for their love.

"Man erblickt nur was man schon weiss und versteht"

J.W.v.Goethe, April 24th, 1819

State University of New York at Albany

College of Science and Mathematics

The thesis for the master's degree submitted by

CHRISTOPH K. STEINHARDT

under the title

STRUCTURE AND STRATIGRAPHY

OF WEST HAVEN, VERMONT

has been read by the undersigned It is hereby recommended for acceptance by the Faculty with credit to the amount of six semester hours

(signed) W. Means

(date) 6th December 1983

(signed) N. D. Means

(date) Dec. 6, 1983

(signed) G. W. Putman

(date) Dec. 6, 1983

Recommendation accepted by the Dean of Graduate Studies for the Graduate Council

(signed) ______________________

(date) ______________________

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CHAPTER I

INTRODUCTION

1.1 Opening Remarks

The opening quote, "We see only what we already know and understand", states a problem that is common to all natural sciences. In my opinion it is particularly true for the earth sciences, as proven by the numerous and sometimes saddening disputes between different schools, that have hindered progress in geology, e.g. the Neptunist-Plutonist debate, or more recently Fixist versus Mobilist views of the earth's crust.

I think the quality of scientific work depends on the extent to which one overcomes the limits of one's knowledge and preconceived notions of how things should look, while making observations. If geology is to become a science we must make as many observations as possible (and even more) and clearly distinguish them from the conclusions we derive from them. For the conclusions are more prone to error and fashionable thinking than the data.

With this idea in mind, I have studied an area of about sixty square kilometers in some detail. It is located around 73°20'W, 43°40'N in the town of West Haven, Vermont and parts of the adjacent townships of Whitehall, New York, Hampton, New York, and Fair Haven, Vermont (Fig. 1.1). To the west, the area is bounded by the crest of the "Vermont Adirondacks" and a line parallel to and about one kilometer east of the southernmost tip of Lake Champlain. The northern boundary coincides with the town line of West Haven; Route 22A marks more or less the eastern limit of the area and a line east to west approximately one and a half kilometers south of Carver's Falls on the Poultney River the southern border.

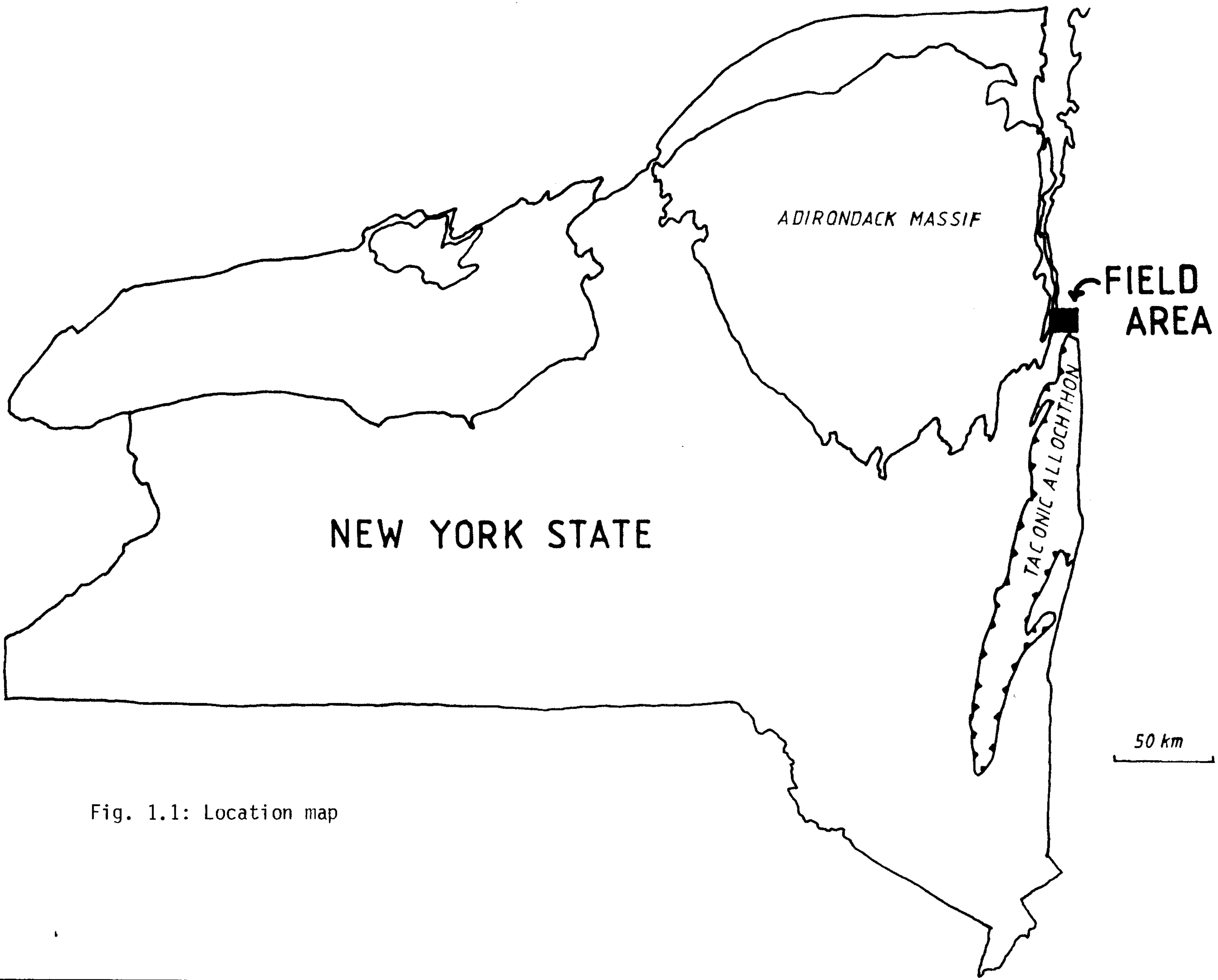


Fig. 1.1: Location map

Various localities outside the area outlined above have been studied to gain insight into local and regional relationships. A SUNYA fieldcamp in Whitehall, New York for which I was a teaching assistant, as well as various fieldtrips with SUNYA faculty, provided excellent opportunities for that purpose. They also led to acquaintaince with the West Haveners and their town. Today, a decreasing number of dairy farmers form the core of West Haven's 260 inhabitants. Before the advent of the Delaware and Hudson Railroad in Fair Haven, however, West Haven was a much larger town, having seven school districts, approximately 700 inhabitants, and sizeable papermills, sawmills, and even a steel mill (R. Best, personal communication, 1982). The Poultney River was the major access route to Carver's Falls from Whitehall and Lake Champlain, and along it most business was concentrated. The river had already been travelled by Indians as arrowheads found by B. Bishop and other townspeople demonstrate.

Topographically, the study area is characterized by soft, rolling hills, with heights varying from 100 feet at Carver's Falls, to 550 feet on Forbes Hill. The hills, which contain most of the outcrop, are generally covered by woods, whereas pastures and, where flat enough, corn and some wheat fields occupy the lowlands. This landscape is mainly due to glaciation during the last ice age, as evidenced by glacial striations (trending about 006°) on several outcrops, and numerous erratic boulders of "foreign" lithology (e.g. Monkton Quartzite) on the hills. Areas lower than 250 feet are generally underlain by sands and clays deposited in a periglacial Lake Champlain, which followed the retreat of the glaciers (Chadwick, 1935; DeSimone, 1983).

In this setting, I mapped in the summer and fall of 1982 for three months on a scale of 1:10,000, using enlargements of 7.5 minute topographic maps as well as airphotos with mylar overlays as a base. On several east-west traverses across strike, about 850 outcrops were recorded (the bigger ones in detail). The exposure in the area is well below 1% due to glacial and postglacial cover, and Holocene sedimentation. Outcrop is concentrated in elevations above 250 feet (85m). This height represents the maximum sedimentation of Lake Champlain, and therefore large portions of the field area have no outcrop, except where swift rivers have cut down to the bedrocks.

Fortunately, the Poultney River produces very good outcrop across strike along its steep banks between Carver's Falls and Coggman Pond, thus providing a good cross-section through three quarters of the area studied. Because it cannot be reached by road, it has not been studied in its entire length by previous workers, which has led to some quite erroneous interpretations, as I shall discuss later.

1.2 Previous Work and Remaining Problems

The first(and last) published study that included the study area and described it in any detail is the one by J. Rodgers (1937). He rightly recognized that the rocks of the area were deformed, and outlined the main geologic features in three sketch maps (Fig. 1.2 compiles the three). He did not, however, publish a detailed geologic map (personal communication, 1983).

Portions of the area were studied by Zen (1961), who named a mélange unit after the highest elevation in it (i.e., Forbes Hill Mélange) (Fig. 1.3).

Fig. 1.2:

Structural provinces, normal and thrust faults in a portion of the Whitehall Quadrangle, after Rodgers (1937, his figs. 1, 3 and 4)

Key to symbols:

Stippled: Rocks of the Taconic Province

Dashed: Rocks of the Adirondack Province

No symbol: Rocks of the intermediate Province

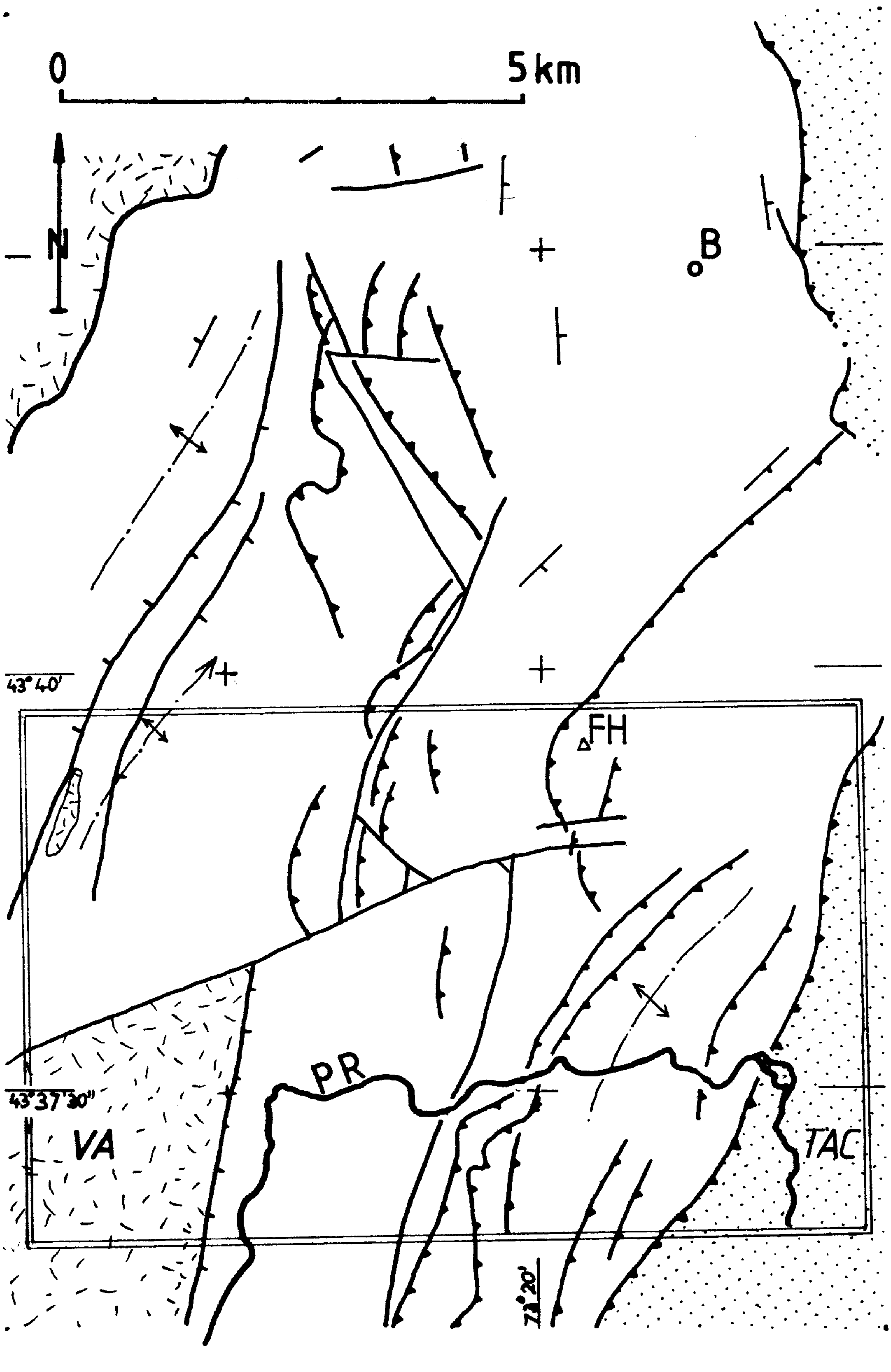
VA: Vermont Adirondacks

TAC: Taconic Allochthon

PR: Poultney River

B: Village of Benson

FH: Summit of Forbes Hill


0
5 km
N
B
43°40'
FH
43°37'30"
VA
PR
TAC
73°20'

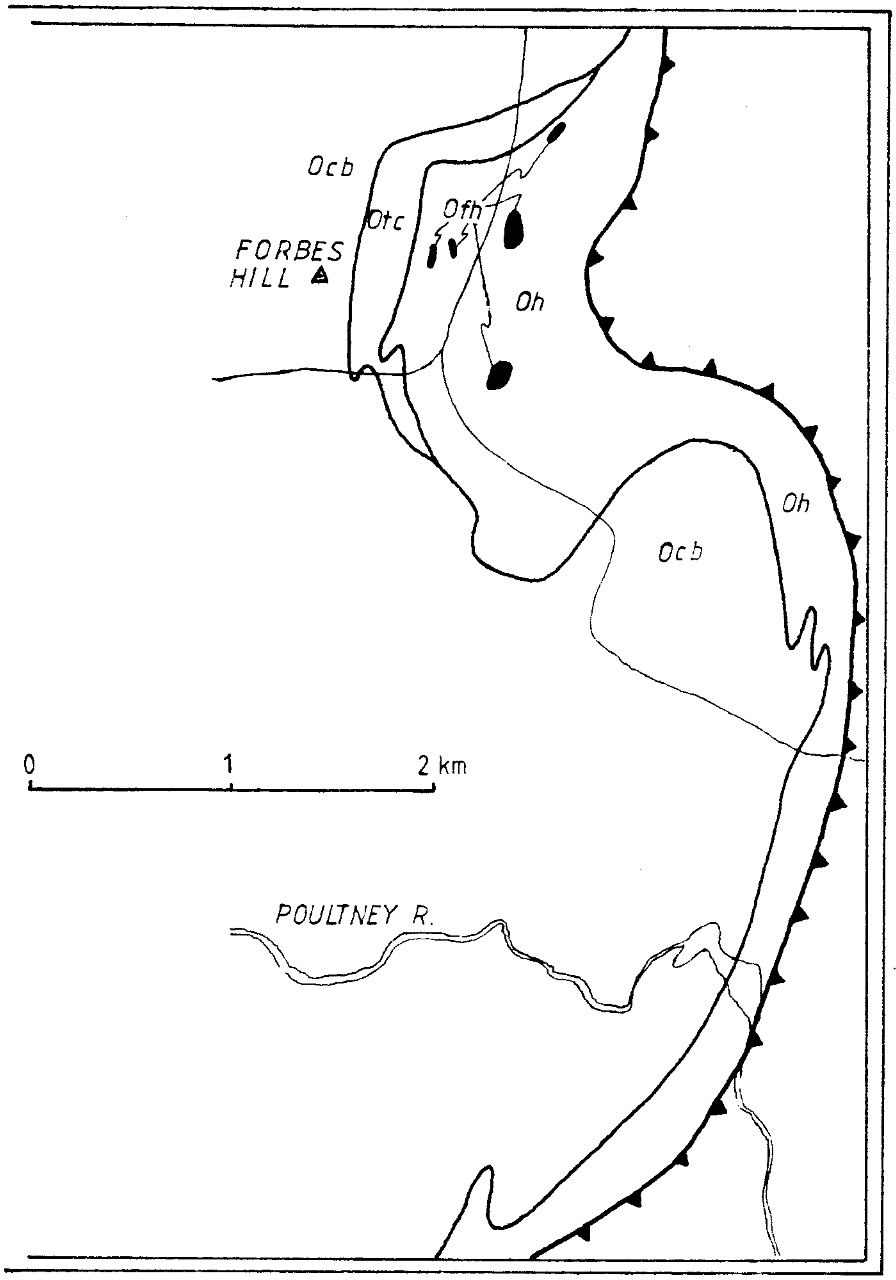


Fig. 1.3: Parts of the field area (outlined by frame) as mapped by Zen (1961) Oh= Hortonville Formation, Ofh = Forbes Hill Conglomerate, Otc = Mid-Ordovician limestone (Middlebury, Orwell and Glens Falls), Ocb = Beldens member Chipman Formation (corresponds to the Providence Island Dolostone)

The only published geologic maps covering the whole study area thus far are the Centennial Map of Vermont (Doll et al., 1961), and the Geological Map of New York State (Fisher et al., 1970). However, both are at a scale of 1:250,000. Fig. 1.4 is an enlargement from Doll et al. (1961) which contains the field area. W. Cady, in 1961, visited various localities for the compilation of this map, but did not draft a detailed map (personal communication, 1983; field notes, 1961). He did, however, investigate a large region immediately to the north (Cady, 1945) and published a map on a scale of 1:62,500 and cross-sections of the Upper Champlain Valley, showing structures that can be correlated with the ones I observed. Welby (1965) dated limestones from Carver's Falls as Chazyan in age, but made no attempt at mapping the extent of the dated stratigraphy.

Only neighbouring areas have been studied more closely in recent years: south of the Poultney River, D. Fisher (in press) mapped similar lithologies for the new Geologic Map of Whitehall Quadrangle. His results are shown in Fig. 1.5. Being a stratigrapher and paleontologist, however, he paid little attention to the structural geology, which has the greatest influence on the map pattern and its interpretation. These structural considerations are so important because the deformation in the eastern part of the area has obliterated sedimentary fabrics and fossils almost completely. In addition, dolomitization is known to vary laterally and therefore a stratigraphy based on lithologic distinctions may be problematic in places.

S. Chisick has studied the sedimentology of rocks in my field area, but at the time of this writing has not published a map (personal communication, 1983).

Fig.1.4: Geologic map of West Haven, Vt., enlarged from Doll et al. (1961) Key to symbols: HR = Hubbardton River, PR = Poultney River, FH = Forbes Hill, VA = Vermont Adirondacks, TAC = Taconic Allochthon

Map units: 1 - Danby (= Potsdam) Formation, 2 - Clarendon Springs/Ticonderoga Fm., 3 - Whitehall Fm., 4 - Cutting Dolomite (corresponds to the Great Meadows Formation), 5 - Bascom Fm. (corresponds to the Ft. Ann and Ft. Cassin Fms.) 6 - Bridport Dolostone (= Providence Island Dolostone) 7 - Middlebury Limestone, 8 - Orwell Limestone, 9 - Root Pond Quartzite member, 10 - Glens Falls Limestone (not present in shown portion of map), 11 - Hortonville Shale

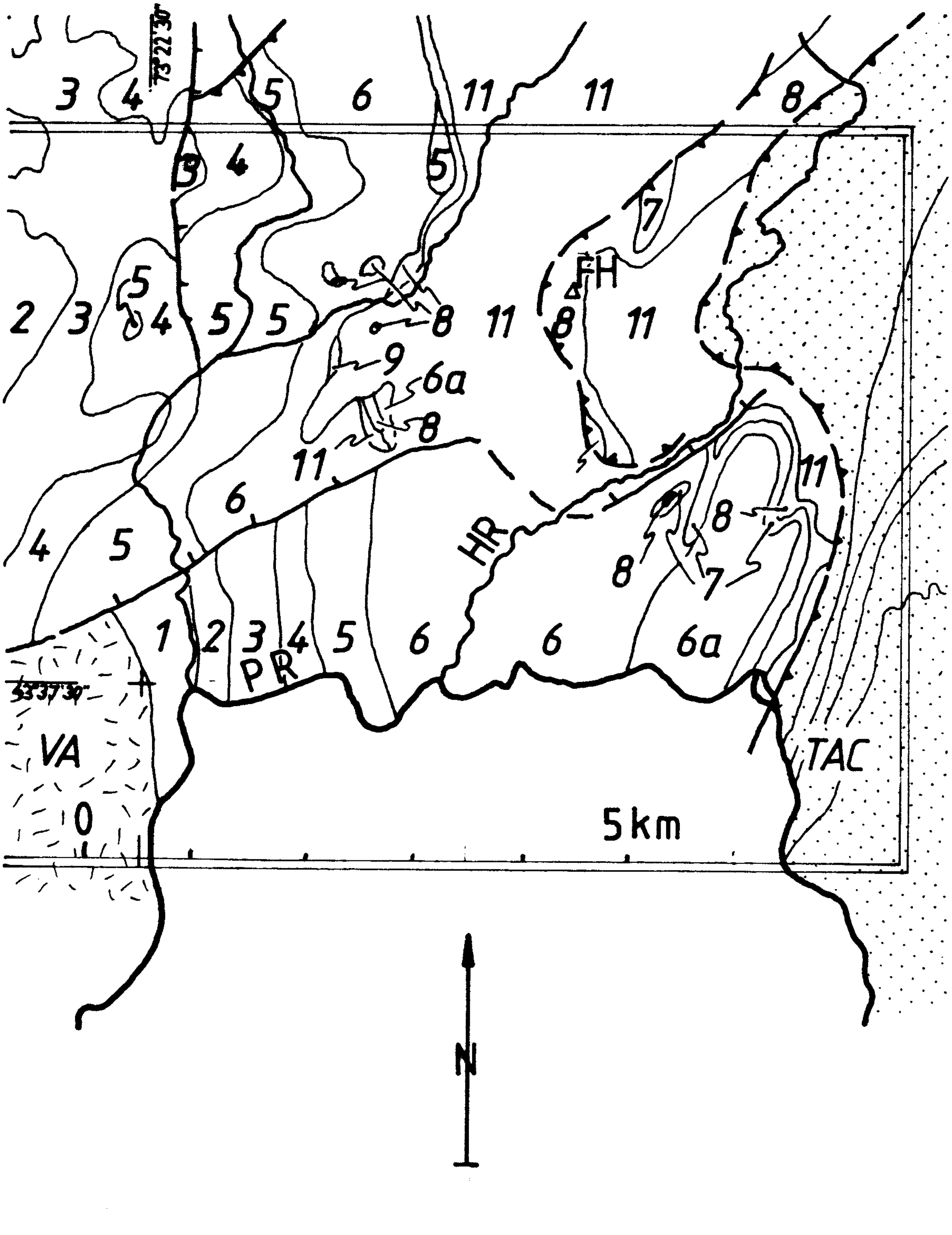

73°22'30"
3
4
5
6
11
11
8
3
4
5
7
FH
5
2
3
4
5
5
8
11
8
11
9
6a
8
11
6
HR
11
8
4
5
8
7
1
2
3
4
5
6
6
6a
PR
53°37'30"
VA
0
TAC
5km
N

Fig. 1.5: Geology of part of the Whitehall 15 min. Quadrangle as mapped by Fisher (in press)

Key to symbols:

Brick symbol = Shelf carbonates

Dashes = Grenvillian basement

Stipples = Taconic Allochthon (greywackes and slates)

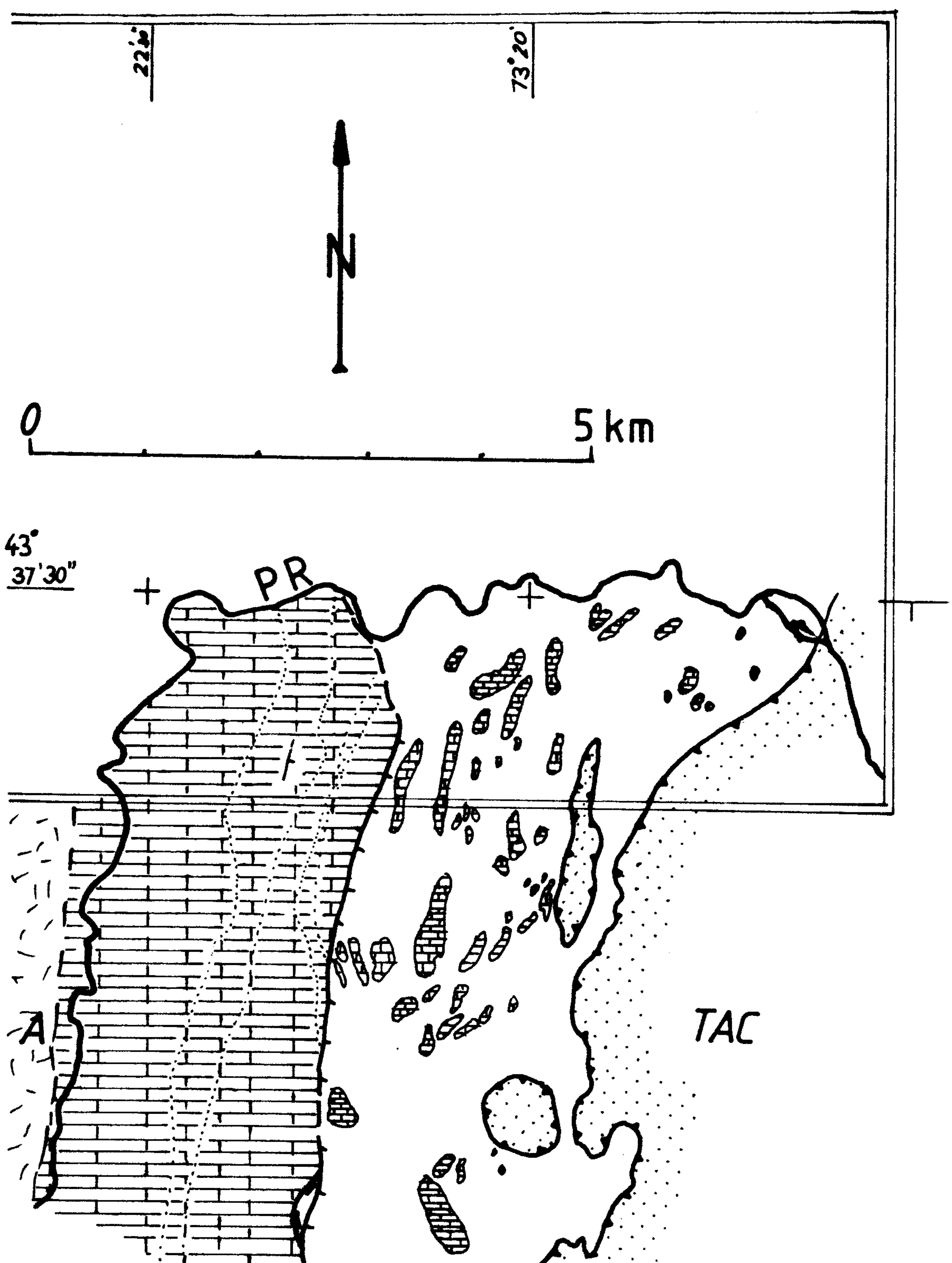
No signature = Forbes Hill Melange containing large carbonate blocks

Double Line = Limit of field area of this study

A = Vermont Adirondacks

TAC = Taconic Allochthon

PR = Poultney River


22'30'
73°20'
N
0
5 km
43°
37'30"
PR
A
TAC

1.3 Purpose of Study

From the above it is clear that producing a detailed map is an end in itself, especially since the geology there is rather complex and cannot be inferred by extrapolation of previously mapped structures along strike.

Other purposes of this study are:

1. To determine the structural effects of the emplacement of an allochthonous complex of continental rise and slope sediments (i.e., Taconic Allochthon) on top of a carbonate shelf sequence comprising most of the area mapped. The area chosen was particularly promising for this purpose, because there the allochthonous mass has come closest to the autochthonous basement (i.e., the Adirondacks) and therefore, any effects due to the transport should be most clearly expressed in the structure of the rocks.
2. In particular, a deformation history of the area was sought. Points of interest were:
   -- the shape of the thrust surfaces (folded or not?)
   -- the timing of cleavage development and folding (pre-, syn-, or post-thrusting?)
   -- structural variations across strike (can structures in the east be correlated to those in the west?).
3. To localize the continuation of a major thrust fault known to the north of the field area; the Champlain Thrust.
4. To re-examine the extent, nature and generation of a mélange unit (Forbes Hill Conglomerate, Zen, 1961) thought to occupy

large portions of the area to be mapped.

1.4 Methods of Examination

Structural analyses of the area on different scales have been carried out. Structures were examined in detail in some of the better outcrops. For this purpose some of the larger outcrops were mapped on a large scale (e.g., Plate IV), and numerous fabric elements were measured.

The map-scale structure was derived from the map pattern (Plate I), stereographic air photo examination, and interpretation in the cross-sections (Plate II). The local and regional framework as described in the available literature and discussions with students of New England geology at SUNYA and elsewhere were helpful.

The lithologies and their spatial distribution were further defined by examination of well log data obtained from the State of Vermont (Department of Water Resources, 1966-1983).

Correlation and dating of litho-units was achieved by comparison of the rocks found with those described in the pertinent literature, in particular Fisher (in press) and Cady (1945 and unpublished field notes, 1960).

Crucial field relationships were re-checked in April 1983 and during the SUNYA fieldcamp May - June 1983 comparisons to immediately adjacent areas to the south provided further confirmation of the data presented.

CHAPTER II

REGIONAL GEOLOGY

2.1. Facts

Fig. 2.1 shows the generalized bedrock geology of western New England and eastern New York. Going from northwest to southeast the following rocks crop out:

1. Precambrian metamorphosed gneisses and anorthosites of the Adirondack massif, deformed and metamorphosed during Grenvillian time, (about 1 Ba ago) representing a stable craton since the late Precambrian (Symbol #7).

2. An early Precambrian to early middle Ordovician shelf sequence. It begins with basal conglomerates and sandstones, e.g., Cheshire Quartzite (Walcott, 1891), and Potsdam Formation (Emmons, 1842). Bird and Dewey (1970) showed that the basal conglomerates prograde westward with time, indicating that rifting and subsidence started somewhere east of the present Green Mountains and migrated westward.

   Above the basal transgressive series follows a carbonate sequence of at least 500 meters thickness (Fisher, in press) in which the lithofacies and several disconformities document a typical shallow marine shelf environment. The lithologies range from mainly dolomites in the lower part to more dominantly limestones in the upper part (Doll et al., 1961; Fisher, in press) (Symbol #4).

3. Overlying and to the east of the shelf sequence, pelitic and fine grained psammitic rocks occur (Symbol #2). This is the Hudson River Group of Mather (1840), parts of which were dated using graptolites (Hall, 1847), as Trentonian in age (about 445 Ma ago). Ruedemann established the biostratigraphy of the Ordovician of New York State and today the graptolite zones of Riva (1974) are

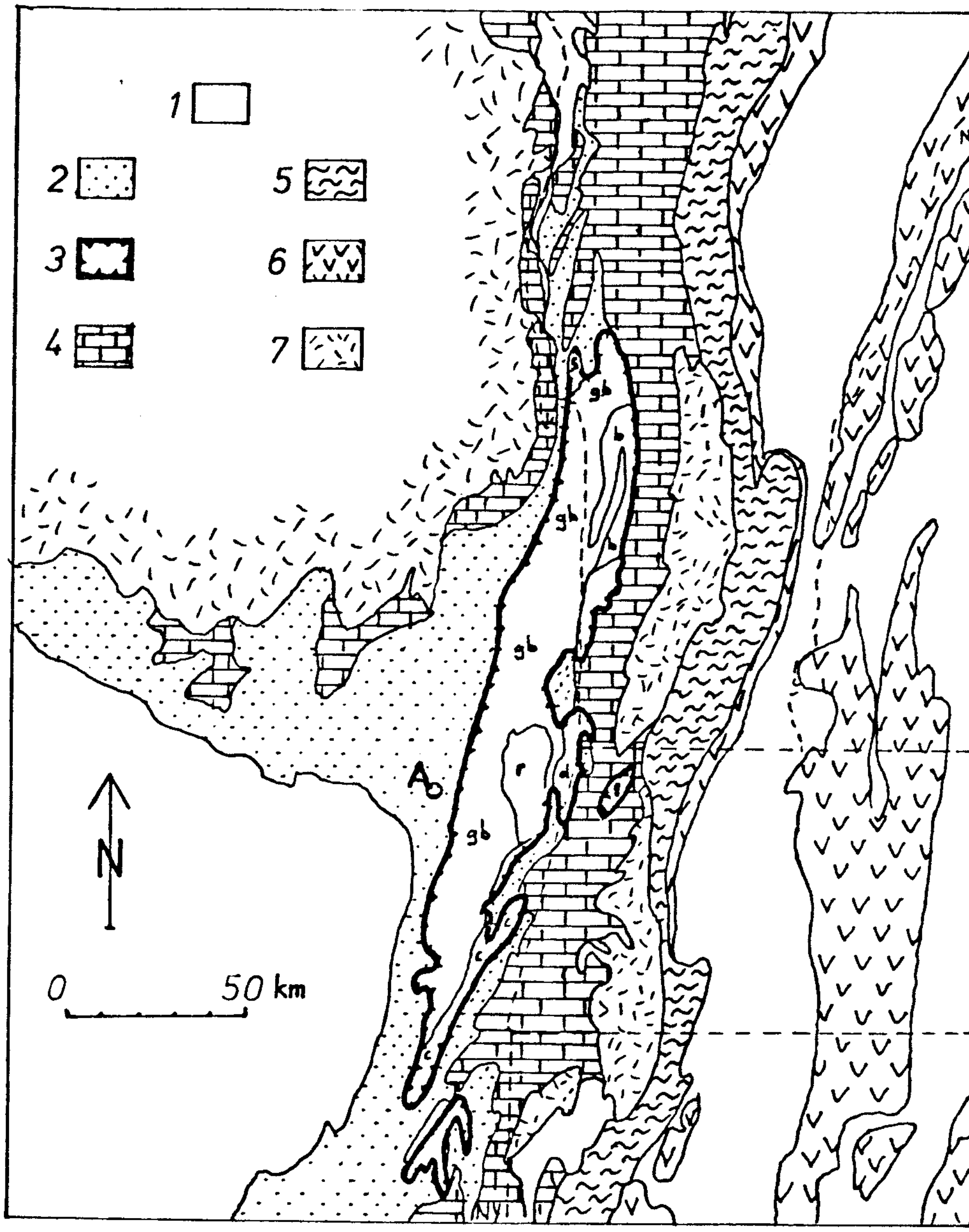


Fig.2.1: Generalized bedrock geology of western New England and eastern New York (after Rowley & Kidd 1981). A = Albany, gb,r, d,c,s,b = slices of the Taconic Allochthon. Symbol # 1 = post-Ordovician sediments. Other symbols are explained in text.

the standard time scale. Various other names have been used in different localities for this sequence and the degree of confusion increases towards the east, where it contacts the Taconic Allochthon. In the eastern part, this group becomes increasingly deformed and disrupted and gives rise to a mélange zone immediately adjacent to the Taconic Allochthon. This is the Taconic Mélange of Fisher et al. (1970), the Forbes Hill Conglomerate of Zen (1961) and it has also been given various other local names (see Vollmer, 1981; Bosworth and Vollmer, 1981 for a discussion). Early on in the investigation it was recognized that this mélange unit represented a major overthrust, which brought the Taconic rocks above the Hudson River Group.

4. The Taconic Allochthon (Symbol #3) consists mainly of argillites, greywackes and minor quartzites and carbonates of Cambrian to medial Ordovician age (Rowley, Kidd, and Delano, 1979). These rocks were deposited in deep water and are now believed to represent a continental rise sequence which is coeval with the autochthonous shelf sequence (Bird and Dewey, 1970; Zen, 1972; Rowley, Kidd and Delano, 1979; Rowley and Kidd, 1981). The whole of the Taconic Allochthon can be subdivided into seven slices (Zen, 1967). These slices form an imbricate stack, the contacts between them dipping east, so that the structurally lowest slices, Giddings Brook and Sunset Lake crop out in the westernmost part of the Allochthonous mass, where they overlie the middle Ordovician flysch (i.e., the Taconic Mélange or the Forbes Hill Conglomerate of Zen).

However, similar flysch units, also of medial Ordovician age are found at the top of the Taconic stratigraphic sequence in the

Giddings Brook slice, i.e., the Pawlet Formation of Zen (1961), and the Austin Glen member of Ruedemann (1942). The base of some slices carries large "slivers" of parautochthonous shelf carbonate, e.g., in the Dorsett slice (Thompson, 1967; see Rowley and Kidd, 1981 for discussion).

5. To the east of the Taconic Allochthon, metamorphosed remnants of shelf sequence rocks (Symbol #4) are preserved in front of and overlying parautochthonous massifs of Grenvillian basement, such as the Berkshire Massif (Ratcliffe, 1979) and the Green Mountain Massif (Doll et al., 1961) (Symbol #7). These massifs were probably deformed and transported during the medial Ordovician Taconic Orogeny (Rowley and Kidd 1981).

6. East of the rocks just described, a zone of polyphase deformed, strongly metamorphosed rocks follows (Symbol #5). Of unknown thickness, those rocks consist of metavolcanics, metamorphosed volcaniclastics, and schists locally grading into gneisses (Rowley and Kidd, 1981; Doll et al., 1961). Slivers of serpentinized mafics and ultramafics also occur in this zone. They are interpreted as remnants of ophiolites marking the Taconic suture.

7. Even further to the east, rocks of volcanic origin (Ammonoosuc Volcanics, Doll et al., 1961) overlie and intrude igneous and metasedimentary gneisses of late Precambrian age (Rowley and Kidd, 1981). The Ammonoosuc Volcanics are associated with plutonic rocks that yield U/Pb ages of 490 to 440 Ma ago (Naylor, 1976; Aleinikoff et al., 1979). These rocks represent a pre-mid Ordovician volcanic arc sequence (Rowley and Kidd, 1981).

2.2 Interpretations

The dominant orogenic event in the region is of medial Ordovician age (about 440 Ma ago) and has been termed the Taconic Orogeny. Before 1970, research was focused on its internal characteristics rather than its mode of formation, with special attention to the "Taconic System" (Emmons, 1842). Its allochthonous nature had been proposed very early on in the investigation by Ruedemann (1909). It was, however, strongly debated by other workers (see Zen, 1967, p. 3-6 for discussion). Zen, in 1967, published an outstanding summary of the Taconic geology, but he wisely refrained from anything more than speculation about the actual cause of the deformation. His speculations involve a crustal collapse to form the Middlebury Synclinorium, into which the major part of the Taconic Allochthon (the Giddings Brook slice) was brought from the east by gravity sliding, followed by thrust emplacement of the upper slices.

It took the evolution of the plate tectonic concept and its application to old mountain belts to reach a somewhat better understanding of the large scale processes that caused the Taconic Orogeny. Following a suggestion by Wilson (1966), Bird and Dewey in 1970 first presented a model that explained the geology of the Appalachian Foldbelt in terms of the opening and closing of a Paleozoic Proto-Atlantic (Iapetus of Wilson, 1966). Their model, involved the conversion of an Atlantic-type margin to an active Andean-type margin. More recently Rowley and Kidd (1981) presented a model that explains the geology shown in Fig. 2.1 quite satisfactorily (Fig. 2.2). It interpreted the northwest-southeast succession of rocks described above in terms of an Atlantic-type margin-island arc collision. (See their paper and explanation of Fig. 2.2 for details).

Their model has been widely accepted today, although some problems remain open for discussion (Rodgers, 1982; Geiser, 1982). Some of these problems are:

-- Soft sediment versus hard rock emplacement of the Giddings Brook slice

-- Timing of slaty cleavage development: before or after emplacement of Taconic Allochthon?

-- Involvement of underlying (par-) autochthonous shelf sequence in Taconic deformation.

These problems will be addressed in later chapters of this thesis.

PLATE TECTONIC EVOLUTION OF THE TACONIC OROGENY (after ROWLEY & KIDD 1981)

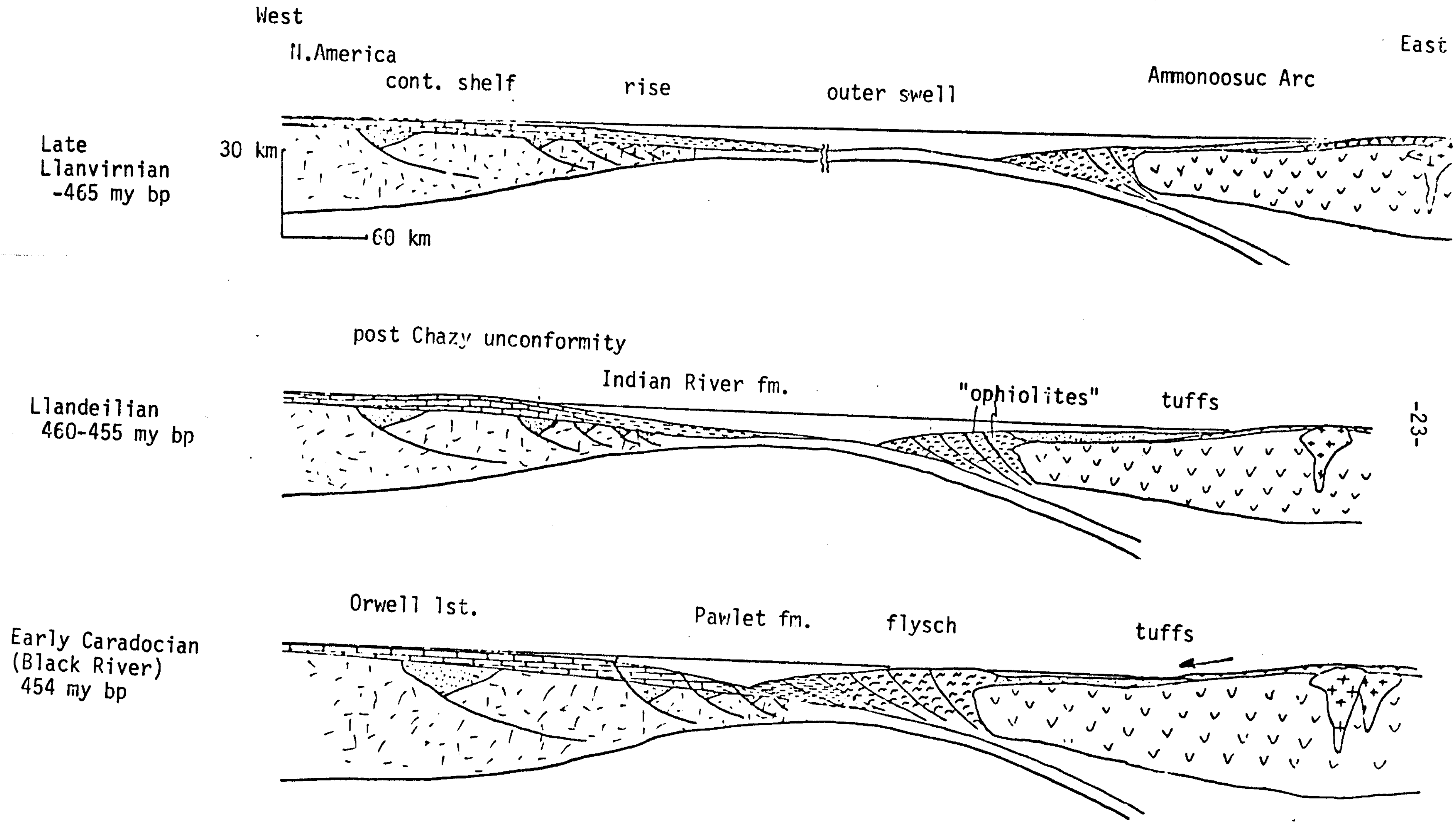


Fig.2.2: Plate Tectonic Model for the Taconic Orogeny

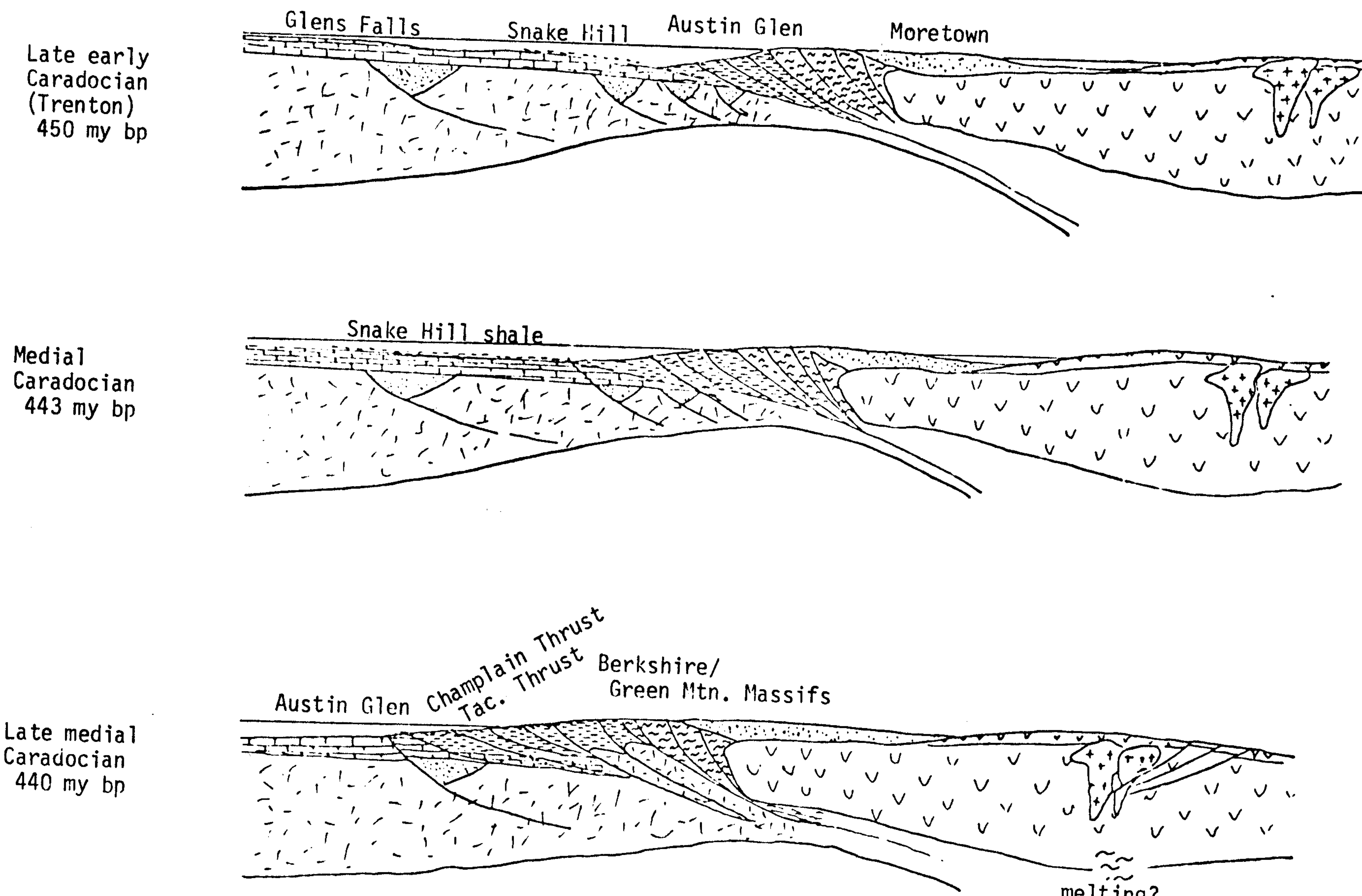

Late early
Caradocian
(Trenton)
450 my bp
Glens Falls
Snake Hill
Austin Glen
Moretown
Medial
Caradocian
443 my bp
Snake Hill shale
Late medial
Caradocian
440 my bp
Austin Glen
Champlain Thrust
Tac. Thrust
Berkshire/
Green Mtn. Massifs
melting?

Fig. 2.2 cont'd

CHAPTER III

STRATIGRAPHY

3.1 Introduction

The rocks in the field area are part of a sequence of Paleozoic carbonate rocks forming an elongate north-south trending belt in the Champlain and Vermont Valleys in New England and the Hudson River Valley in New York where they are increasingly covered to the south by Silurian and Devonian clastics, which are not treated in this study. Plate III gives a summary of the evolution of stratigraphic names and correlations. After early investigations by Hall, Emmons and others, Brainerd & Seely in 1890 published the first stratigraphic nomenclature and rock description that covered the broad equivalents to rocks preserved in the field area. Their column is based on the study of various widely spread locations around Lake Champlain, with some concentration around Ft. Cassin (Welby, 1961). Their work formed a basis for most further investigations, which refined their stratigraphy in specific areas (e.g., Cady 1945), assigned new names (e.g., Rodgers 1937) and defined ages more precisely. Welby in 1961 published a comprehensive study of the Paleozoic carbonate sequence in the Central Champlain Valley which to this day forms a key reference (Plate III, column 5). His work (together with that of others in other parts) formed a base for the choice of formations for the Vermont Centennial Map (Doll et al., 1961) which my mapping has found to be a very dependable source (Fig. 1.4). Comparing carbonate rocks of similar age from different localities is made difficult by lateral facies variations, especially where dolomitization has occurred. This problem will be discussed at the end of this chapter.

The choice of mappable units in this study was based on observational differences wherever possible. Unfortunately, about 70% of the rock column consists of dolomitic rocks which are characterized by a distinct lack of determinable fossils. Lithologic criteria were used in separating these formations. Some of the contacts appear to be gradational, so that the map boundaries have to be taken as representing this and not sharp contacts. In the field no attempt was made to name units in terms of existing classifications. Instead, unit boundaries were drawn to join outcrops of similar lithologies.

During the drafting of the map the litho-units so found were compared to previous work particularly that of Fisher (in press), and the similarity was close enough to use his nomenclature. This, however, is only so for the pre-Chazyan part of the sequence, exposed in the western half of the area. In the eastern half rocks are strongly deformed and displaced and the effects of deformation on the observable lithology may be great. Also, Fisher's interpretation of that part of the area and the one corresponding to it in the south is different from the interpretation in this study, so that other names have been used. In particular, his subdivision of tectonized limestones of inferred Black River and Trentonian age could not be followed, especially since other workers have assigned those very rocks a Chazyan age based on fossil evidence (e.g., Welby, 1965; Chisick, personal communication, 1983). The units are described in their stratigraphic or structural order from lowest to highest. Differences in lithology always cause differences in the quality of outcrop, so that for some of the units (e.g., Winchell Creek Siltstone) the observations and their interpretations are more complete than for others (e.g., Ticonderoga). Wherever a unit is characterized by a distinctive feature it will be described in some detail. In the undeformed part of the field area some of these characteristics could be used as valuable guides to establish boundaries in an otherwise often very monotonous dolostone sequence of several hundred feet thickness. A good example for this is the cross bedding in the Winchell Creek Siltstone.

To describe rocks I tried to use non-genetic terms, such as the classification proposed by Friedman & Sanders (1976) for carbonate rocks. Definitions for other rock types and general terms are taken from the Glossary of Geology, 2nd edition, 1982.

At the beginning of each unit description a general correlation and reference to previous work is given and at the end some genetic considerations are made, based on observations of the sedimentary fabric.

It must be borne in mind, however, that the distinction of stratigraphic units is mainly confined to the undeformed rocks in the western half of the area, and in the east may be very obscured by the deformation observed. This is all the more true for paleoenvironmental interpretations based on sedimentary structures.

3.2 Potsdam Formation

The Potsdam Formation was first named by Ebenezer Emmons (1838) and comprised rocks along the northernmost shore of Lake Champlain. In my field area this unit underlies the westernmost part and forms some of the cliffs that make up the shore of Lake Champlain. Its lower boundary is seen nowhere in the area studied, but can be observed about 6 kilometers to the west on Route 22 about 6 kilometers south of the town of Ticonderoga. There this formation overlies Grenvillian basement. The contact is marked by an impressive transgression conglomerate with pebbles derived from the underlying highly metamorphic basement rocks. The Potsdam is the first unit preserved on the Grenvillian basement in the vicinity of Lake Champlain.

In the study area the formation consists of light grey to tan, medium to coarse grained sandstones and quartzites. Arkosic sandstones in places with a calcareous matrix do occur (Fig. 3.1). The quartzites form only thin beds in the lower part of the observed column, protrude on outcrop surfaces because of their greater resistance to weathering and vary in color from grey to a yellowish darker grey.

Medium to coarse grained sandstones with an argillaceous or carbonate matrix predominate. Their colours vary from tan to grey and they weather to a light yellowish brown. In places concentrations of iron hydroxide give the rock a rusty brown colour, usually near cavities or very porous zones. This suggests that parts of the original matrix were carried away by solution processes. The iron is distributed unevenly, forming spots and concretions as well as bands and lenses parallel to bedding and seems to be associated with coarser grained and cross-bedded strata.

The upper half of the observed column consists mainly of fairly clean white to grey quartzites in beds that are on the order of one meter thick. A lamination, sometimes emphasized by differential weathering, usually marks bedding. It is probably the result of compositional changes of the material arriving at the site of deposition at different times. It is not clear whether these quartzites formed

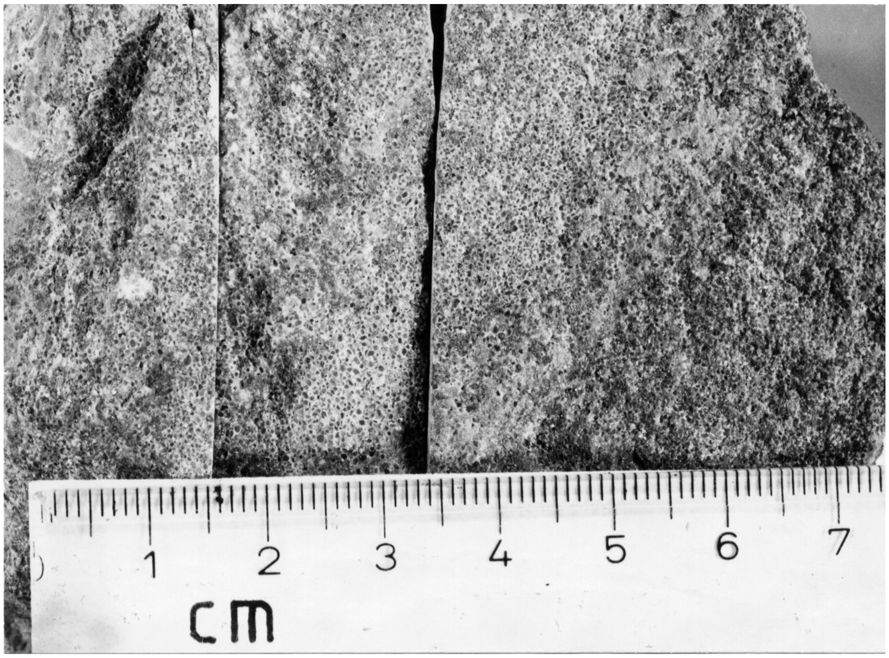


Figure 3.1: Arkosic sandstone from the Potsdam Formation. Kline Hill [21], 5 km west of West Haven town center.

as such during diagenesis or whether they have been silicified by silica-rich pore fluids at some later time during the past 500 million years.

The thickness of the observed section is 50 meters (170 feet), but neither the upper nor the lower boundary are seen. For the construction of the map the boundary to the overlying Ticonderoga was put at 15 meters above the top of the observed section. The grain size of the lowest part of the observed section is much smaller than that seen at the basal contact. Therefore, it is likely that there are some tens of meters of Potsdam unexposed. A total minimum thickness of the formation is on the order of one hundred meters. This is consistent with the thicknesses assumed by Fisher (in press), Cady (1945) and Rodgers (1937). Their maximum values are 75 meters, 160 meters and 135 meters, respectively. Thickness variations in the Potsdam are very likely for the following reasons:

First the Potsdam was deposited unconformably on pre-Cambrian gneisses as can be seen elsewhere. These rocks probably showed some kind of relief prior to the deposition of the Potsdam sediments.

The grain size of the Potsdam (upward of 0.5mm) indicates high energy currents with velocities on the order of about 50 cm s^{-1} (1 mi/hour) if the Hjulstrom curve is a valid criteria (e.g., Seibold & Berger, 1983, p. 81). Currents moving at such speeds have also the capability to erode and while deposition may occur in one place erosion may take place elsewhere.

Another cause for varying thickness is sketched in Fig. 3.2. It shows a good example of a growth fault. It comes from an outcrop of

Figure 3.2: Growth fault. Potsdam Sandstone, 5 km south of Whitehall, New York.

Potsdam Sandstone outside the field area, about 5 kilometers due south of Whitehall.

If the area was faulted during deposition of the Potsdam then sedimentary thicknesses will be higher on downfaulted blocks and smaller on the relatively higher blocks.

The coarse grains, the dominance of arenaceous material, the presence of arkoses and iron oxide dots all point to the Potsdam being a very shallow marine deposit. It may be a beach/barrier bar deposit.

From fossils found by other workers outside the field area, the Potsdam Sandstone has been assigned an upper Cambrian (Croixian) age (Rodgers, 1937; Brainerd & Seely, 1890).

3.3 Ticonderoga Formation

Conformably overlying the Potsdam, the Ticonderoga was named by Rodgers (1955) (in Welby 1961).

In the field area it is restricted to a mere two outcrops one of which is an old quarry [1] that exposes rocks essentially identical to those in the type section given in Welby 1961, p. 232.

The observed lithologies correlate with units 40 to 33 in Rodgers' type section. At the bottom a massive thick bedded unit of blue-grey and yellowish weathering medium grained dolostones is exposed. In the yellow weathering beds, joints perpendicular to bedding are prominent, whereas in the blue-grey weathering beds the only structure is the bedding parallel lamination. The blue beds are morphologically harder and protrude from the weathered surface.

At the top of this about 6 meter thick unit lies a very distinct grey quartzite of slightly less than a meter thickness. It is laminated and cross-laminated, weathers light blueish grey and shows a light grey - grey banding on the fresh surface.

It is overlain by a chert-mottled and banded dolomitic sandstone of rather fine grain and a yellowish weathering colour. It grades upward into purer dolomitic sandstone and thick beds at the bottom give way to thinner beds at the top of the outcrop. The sequence above the quartzite bed is approximately 5 meters thick. I believe that the quartzite bed can be correlated with unit number 38 in Rodgers' type section and have used this as a criterion to establish the upper and lower boundaries of the Ticonderoga Formation. This extrapolation was necessary because of the lack of outcrop. Apparently, the Ticonderoga is less resistant to weathering than the underlying Potsdam or the Whitehall Formation above and therefore less exposed. It is not surprising since most of the unit is a sandstone with dolomite and lesser quartz grains, and a carbonate matrix. The quartz grains are generally well rounded and make up less than 10 per cent of the rock volume, except in the quartzite bed.

The thickness of the Ticonderoga had to be taken from the literature, since neither the top nor the bottom are exposed in the field area. Rodgers' type section from Ticonderoga Village (about 20 kilometers northwest of the area studied) has a measured thickness of 62 meters (185 feet). His unit 38, which is presumably exposed in the field area, occurs about 13 meters (40 feet) above the contact with the underlying Potsdam Sandstone. Fisher (in press) gives thicknesses between 50 and 80 meters (150 to 240 feet) for the Ticonderoga in the Whitehall quadrangle and this is assumed also in my map.

The age of the Ticonderoga is latest Cambrian (Trempeleau) according to Rodgers (1937), Rodgers & Fisher (1969) and Fisher (in press). No fossil was observed in my two outcrops.

It seems to me that the sequence described above is one of very shallow marine supra- to intertidal environment. The following points support this interpretation:

- The dominance of dolomite as major constituent of the rocks.
- The existence and paucity of quartz sandstone.
- The existence of lamination and cross-lamination showing that material was deposited or reworked by fairly strong currents.
- The absence of ferric hydroxides from the Ticonderoga, which distinguishes it from the Potsdam Formation.
- The generally smaller grain size of the clastic portion of the rock. It indicates a more distal position with respect to the source area.

It is not clear however, when the rocks actually became dolomitic. It could have been before diagenesis through magnesium-rich interstitial fluids ("hypersaline brines" in the sense of Friedman & Sanders, 1967). According to Friedman & Sanders such syngenetic supra-/intertidal dolostones would have just about the characteristics found in the Ticonderoga (p. 287 loc. cit.). Indeed they cite rocks of similar age and stratigraphic position from New York State as an example (p. 315 loc. cit.).

Another mode of formation would be post-diagenetic (epigenetic in terms of Friedman & Sanders, 1967, 1976; Fairbridge, 1967, 1983). This assumes that the rocks present after lithification were limestones (probably with a high Mg content) and were subsequently transformed into dolostones by metasomatism at depth. In the case of the Ticonderoga Dolostone the first mode seems more likely.

3.4 Whitehall Formation

The name Whitehall Formation was added to geologic nomenclature by J. Rodgers in his 1937 paper. In the field area this formation is restricted to four outcrops and the correlation is not very satisfying. In all cases the rocks are coarse grained, grey to light grey, massive dolostones, with weathering colours varying from dark grey to light grey. Some lamination and some dark grey mottles are observed, as well as calcite veins and crystals. The bedding is poor, fractures perpendicular to bedding are the most prominent structure. With no fossils and little outcrop at hand the full description of this unit is somewhat difficult. It seems quite likely that there is considerable lateral variation of this unit in thickness as well as lithology.

Fisher (in press) distinguishes several limestone members within the Whitehall (e.g., the Warner Hill, Steve's Farm and Rathbunville School Limestones) with a total thickness varying between 4 and 37 meters. Rodgers (1937) citing Brainerd & Seely also mentions limestone occupying the middle and upper parts of the formation (p. 1576/7).

Yet in my field area, less than 10 kilometers north of the type section, and along strike, none of these limestones are seen. Also the thickness must be less than the 295 feet (almost 100 meters) that Rodgers cites from Brainerd & Seely. At Whitehall a thickness of at least 80 meters is observable at Skene Mountain and Warner Hill. But north of the Poultney River my mapping has yielded the conclusion that the Whitehall cannot be much more than 30 meters thick and this is the value adopted for the construction of the map. The constraints for the thickness are set by easily recognized formations below (Ticonderoga) and above (the Great Meadows Formation with the distinctive Winchell Creek Siltstone). Since the dip of the rocks is consistently 10° to the east the thickness of the unit between can be calculated by simple trigonometry. In order to account for a thickness twice as big as the one assumed in the map, one has to allow a dip of 20°, and to reach the value of Rodgers (100 meters or 295 feet) the dip would have to be 30°, which is unrealistic. The derived thickness loss could be explained in several ways:

1) Hidden fault
2) Lensing out of unit - lateral change of units.
3) Mistake in assigning units
4) Disconformity

1) With the lack of outcrop it is quite possible that under the glacial cover a fault has escaped mapping that displaces the top (i.e., eastern) unit boundary relatively downward and therefore causes a smaller apparent thickness.

2) Lateral variation in thickness and/or facies. Steno's law says that sedimentary layers are of limited extent and that they thin towards their limits if these are sedimentary. This could be the case for the Whitehall Formation between the type locality and the study area.

3) Lateral unit variation. It is common that units change their lithologic character laterally. It doesn't require much imagination to accept this all the more for dolomitized rocks be they syn- or epigenetic. Hence what looks like Whitehall dolostone in Whitehall, may look like Ticonderoga in my field area. In the absence of fossil witnesses one cannot even assign an age for either formation. Therefore, the question cannot be conclusively answered.

4) Disconformity at the top of the Whitehall. Fisher (in press), Cady (1945) and Rodgers (1937) all assume a disconformity on top of the Whitehall (or in Cady's case the corresponding Clarendon Springs Dolostone). A disconformity requires uplift or sea level drop and implies erosion. Depending on how long erosion prevailed in a given place more or less of the now Whitehall material could have been carried away, i.e., in the study area more than at the type locality.

3.5 Great Meadows Formation

The name Great Meadows Formation was created by Flower (1964) and is now used by Fisher (in press; and also Fisher and Rodgers, 1969) for rocks overlying the Whitehall Formation in northeastern New York State. It correlates with the Cutting Formation of Cady (1965) and the Calciferous C of Brainerd & Seely (1890).

In the legend to the New York State Geological Survey's maps of the area (Fisher in press) four map units are distinguished. From bottom to top they are Winchell Creek Siltstone, Kingsbury Limestone, Fort Edward Dolostone and Smith's Basin Limestone. The two limestone units are not seen in the area under consideration, so that only the Winchell Creek Siltstone and the Fort Edward Dolostone are distinguished and described below. The thicknesses given by Fisher (in press) are from zero to 12 meters for both limestones and even in areas where the two limestone units are seen, they may lens out along strike.

In my field area, some outcrops of the Great Meadows Formation are calcareous but nowhere is the calcite portion of the rock high enough to name it a limestone. At the base of the Great Meadows Formation lies the Winchell Creek Siltstone (Fisher, 1977). According to Fisher (in press) there is a disconformity between the top of the Whitehall and the Winchell Creek. In the field, however, no clear evidence has been found. The contact appears to be rather gradational showing a transition from the massive grey dolostones typical of the Whitehall to better bedded and laminated dolomitic sandstones typical of the Winchell Creek. These sandstones are generally medium grained and the name siltstone doesn't seem justified. Colours are blueish-grey to grey on the fresh surface and yellowish grey to buff on weathered surfaces. Beds are from 10 to 50 centimeters thick and usually clearly distinguishable. The most characteristic feature of the unit is its well developed cross-bedding. It is best seen on weathered surfaces as shown in Fig. 3.3. Although the figure shows sets of lateral extent up to a meter, they are generally somewhat smaller. The individual laminae are from one millimeter to half a centimeter thick and created by a denser packing of sand grains and a decrease of the carbonate matrix. This explains

Figure 3.3: Crossbedding. Winchell Creek Siltstone, 1 km south of Poultney River [22].

also why the crossbedding is best seen on weathered surfaces. While dolomite or calcite is dissolved with the aid of carbon dioxide present in the atmosphere, the quartz grains remain in place and form a relief. Fig. 3.4 shows another example of the cross lamination. In this case the individual laminae are much smaller.

On fresh surfaces the cross lamination is often hardly visible, because there is very little colour variation between quartz rich and carbonate rich laminae. Not all of the unit is crossbedded as can be seen from Fig. 3.3. Instead, varying amounts of homogeneous, even massive grey dolostone occur, that weathers light grey and shows a "sugary" texture on fresh surfaces, diagnostic of a thoroughly recrystallized dolostone. In some outcrops a bed of solid quartzite up to one meter thick was seen.

It is the crossbedding, however, that makes this unit so useful in the field. Since above and below it are rather massive dolostones that resemble each other closely the Winchell Creek Siltstone serves as a guide horizon. Its thickness varies, especially since the contact to the overlying Fort Edward is not very sharp and had to be placed arbitrarily in places. For the construction of the map a thickness of 24 meters (80 feet) has been assumed.

Its age according to Fisher (in press) is early Ordovician and it represents the lowermost unit of the Beekmantown group.

Above the Winchell Creek Siltstone and with a gradational contact given by the decrease of crossbedded strata and the increase of massive dolostone lies the Fort Edward Dolostone. It is generally a thick bedded to massive, medium to coarse grained dolostone. It

Figure 3.4: Crossbedding in handsample of Winchell Creek Siltstone. From valley south of Warren Hollow Fault.

weathers yellowish grey and is blue-grey on fresh surfaces. It shows a "sugary" recrystallized dolomite texture. Bedding tends to be lost in this lithology and subvertical joints become the most prominent feature. Another common sedimentary structure are chert nodules of irregular shape usually black or dark grey in colour. Often the nodules give way to dark grey patches in the blue-grey rock, giving the rock a mottled appearance. If these patches are of the same origin as the pure chert nodules they are not a synsedimentary feature, but due to diagenesis. There is a second lithology within the Fort Edward consisting of finely laminated, fine-grained blue-grey dolstone with minor amounts of quartz sand and silt.

In a study of the Great Meadows and the overlying Fort Ann Formations of New York State, Mazullo & Friedman (1977) have interpreted these laminated rocks as formed by the build-up of algal mats in a supratidal hypersaline environment. In evidence they found desiccation cracks and curled polygonal clasts. Organic rich darker grey laminae are interpreted as the remainders of the algal filaments. The mottled facies is explained as the result of the burrowing action of herbivore invertebrates that could only exist in the intertidal zone where salinities are normal. This is quite an interesting hypothesis, but in the field, the evidence for the hypersaline character of the laminated facies is lacking. It is possible, however, that most of the sedimentary structures have been wiped out by subsequent dolomitization, so that the algal mat origin of the laminated dolostones cannot be excluded. More detailed petrographic work is needed to decide the question.

The thickness of the Fort Edward Dolostone is about 30 meters (100 feet) as an average value. The whole of the Great Meadows Formation is then 54 meters (180 feet). This value is somewhat less than the 66 meters given by Fisher (in press). It is clear, that the Great Meadows Formation was laid down in a shallow marine shelf environment. Evidence for that is seen in the crossbedding, the lateral variation of thickness and the presence of algal mats. The presence of abundant sand size quartz grains in the Winchell Creek Siltstone suggests that it cannot have been too far away from the shore so that there was sediment input by rivers bringing in clastic material derived from the continent.

3.6 Fort Ann Formation

The Fort Ann Formation (Flower, 1964) corresponds to the lower half of the Calciferous D of Brainerd & Seely (1890), the Bascom Formation in the Vermont State Centennial Map (Doll et al., 1961), and the lower half of Cady's (1945) Bascom Formation.

It is distinguished from the underlying Great Meadows by the greater lithologic variety, in grain size as well as composition and the presence of two sets of joints one parallel and one roughly perpendicular to bedding. These joints are frequently filled with quartz and/or calcite which protrude on weathered surfaces and give the rock a "fretwork" look.

In the lower part the formation consists of medium to coarse grained massive dolostones. On weathered surfaces, they are yellowish grey and on fresh surface, blueish-grey. Where massive, joints perpendicular to bedding are most visible. Some chert nodules have been seen but they are much less common than in the formation below. Quite often centimeter sized cavities are filled with euhedral calcite crystals up to one centimeter across. That the rock is quite porous can also be seen from the fact that the yellow weathering colour penetrates the rock several millimeters deep.

The upper portion of the Fort Ann consists of much finer grained dolostones and limestones. The dolostones there are less massive and show good bedding. Sometimes beds are only 5 centimeters thick. On fresh surfaces they display a fine discontinuous lamination which is bedding and probably an algal mat structure. The same lamination is found in the limestones contained in the upper section of the Fort Ann Formation. The limestones are rather thick bedded, fine grained lime mudstones which also show a grey-dark grey lamination. Their weathering colour is yellowish grey - more like that of dolostone thus indicating a rather high magnesium and iron content.

In the upper section the quartz and calcite filled joints are present which cause the previously described fretwork-look on weathered surfaces.

The thickness of the Fort Ann Formation has been determined in the field to be 27 meters (90 feet). This value falls in the upper end of the range from 15 to 35 meters given by Fisher (in press).

The age of the Fort Ann deduced from its stratigraphic position is medial - early Ordovician or Canadian.

3.7 Fort Cassin Formation

Sciota Limestone

The Sciota Limestone (Fisher, 1977) overlies the Fort Ann disconformably (Fisher, in press). It correlates with the Fort Cassin Formation of Welby, 1961, the Beldens member on the Vermont Centennial Map and divisions D3 and D4 of Brainerd & Seely's Calciferous Formation.

The Sciota Limestone is the first unit, going stratigraphically upward from the Potsdam, in which fossils have been found in the field area. It is thereby quite easy to distinguish from the underlying rocks and was used as guide horizon. However, the fossils are concentrated in thin, conglomeratic layers, which make up less than 10% of the total thickness.

In the nonfossiliferous parts, Sciota Limestone occurs in rather thick and sometimes massive beds, is usually dark blue-grey and weathers light grey to yellowish. Grain size is usually so small that the rock must be called a micrite. The dark colour comes from a high clay content, and clay seams are sometimes found along bedding planes. In one outcrop a fine lamination was observed, expressed by a colour change between yellowish grey and blueish-grey in adjacent laminae. The fossiliferous layers in the Sciota, however, look quite different. They display a much coarser grain size, with clasts in places up to 10 millimeters across, which are often fossil fragments. The fossils are mostly cephalopods and brachiopods, but not very well preserved and many seem to have been fragmented before deposition (Fig. 3.5). They are best seen on bedding planes and are not prominent in outcrops of which the face is a section through bedding. In sectional view, however, one outcrop shows a very good example of a flat pebble conglomerate (Fig. 3.6).

The "flat pebbles" are rectangular pieces of individual beds up to one centimeter thick which consist of the same material as the surrounding rock. It is likely that they were formed by curling up and drying out of the lime mud during a time of low water. When the water rose again it could have pulled individual pieces from the substrate and transported them somewhat. Thus it has to be assumed, that the Sciota Limestone was deposited in an extremely shallow marine environment, which at times lay dry. The fragmented character of the fossils and their occurrence in thin but coarse grained sheets indicates that these have been produced by wave action probably during storms or, perhaps, times of strong tidal currents.

A fine grained calcarenite occurs in several outcrops at the base of the unit. It is showing fine laminations on weathered surfaces. Its thickness seems to vary greatly along strike. In one traverse the apparent thickness was on the order of 20 meters (65 feet) whereas in a traverse 500 meters farther north it was only seen in one outcrop with about one meter thickness. This facies is probably what Fisher (in press) separated from the Sciota Limestone and called Ward Siltstone member. In my view the occurrence of these calcarenites is

Figure 3.4: Fossiliferous layer in Sciota Limestone. View onto bedding plane. Outcrop 1200 m southsoutheast of West Haven town center.

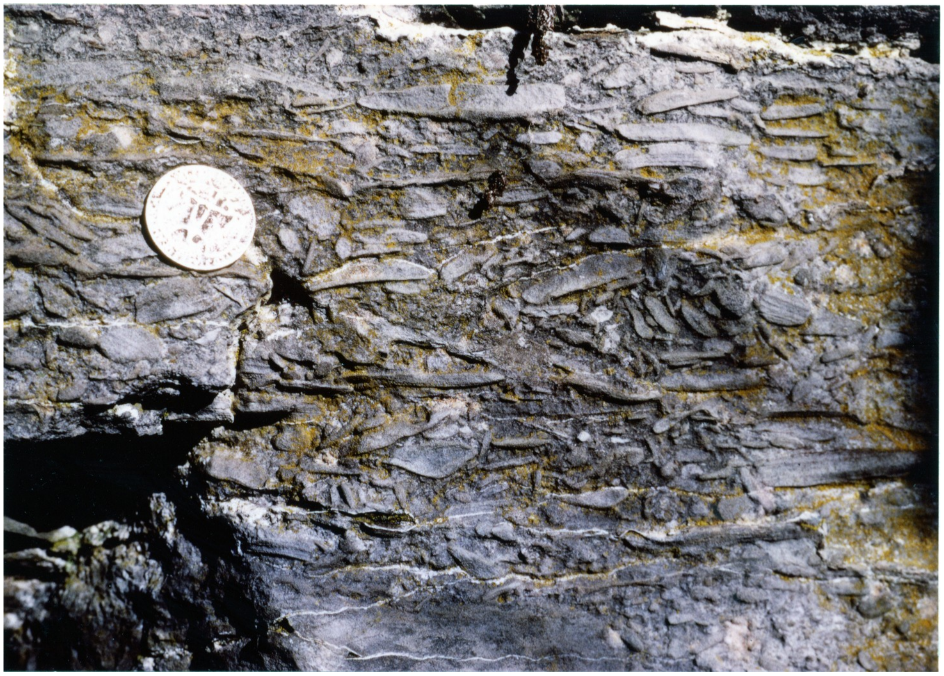


Figure 3.6: Flat pebble conglomerate in Sciota Limestone. Roadcut on the west side of Sciota School.

too sporadic to map them as a separate unit. The whole of the Sciota Limestone is at least 100 meters (330 feet) thick. It is possibly even more, because the upper boundary is of tectonic nature and defined by the thrust fault that marks the beginning of the eastern, deformed area. The fossils in Sciota Limestone yield an upper Canadian age (Fisher, in press).

<u>Providence Island Dolostone</u>

The Providence Island Dolostone was named by Ulrich (1938). It is called Bridport Dolostone on the Vermont Centennial Map (Doll et al., 1961).

The outcrops of this unit occur mostly in the eastern, strongly deformed zone of the mapped area, and the effects of deformation on the lithology are quite significant. The base of the Providence Island Dolostone, however, is exposed in a river section north of Sciota School in the western, less deformed part of the field area [2]. I have used this section for the description of the lower part. The top part is exposed in a beautiful outcrop at Carver's Falls [3], in the deformed part of the area studied. The outcrop there is on the eastern limb of a big anticline and there are 66 meters of continuous section exposed including its upper contact with the overlying limestones (Fig. 3.7). Despite its situation in the deformed areas, the sequence is not internally faulted and dips consistently at about 40° to the southeast. I think that these two outcrops are representative for the Providence Island and use them as general reference. Together they give 76 meters (250 feet) of stratigraphic thickness as a minimum value but the true value is more likely twice that much, judging from the cross sections (Plate II). At the bottom, the Pro-

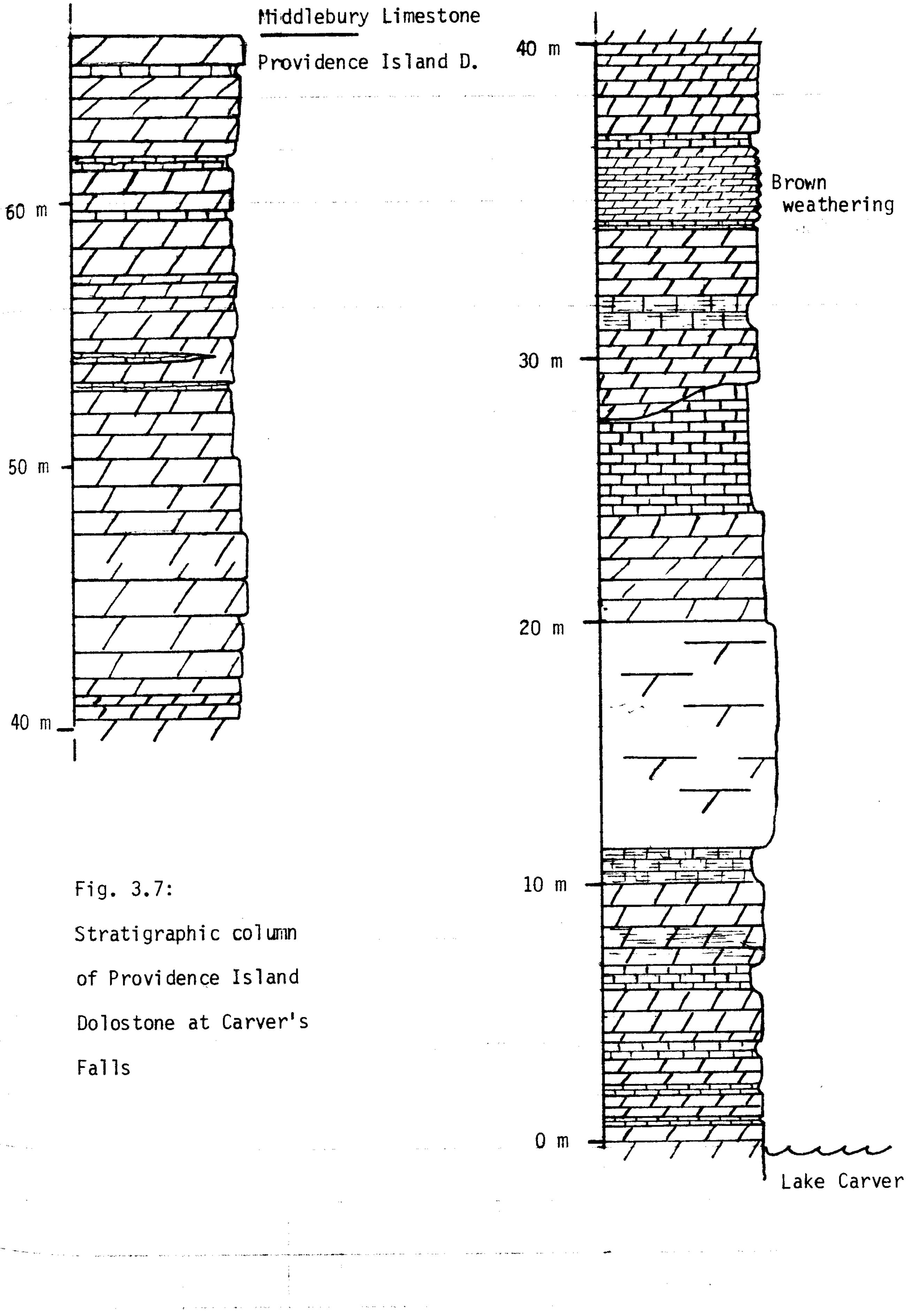


Fig. 3.7:

Stratigraphic column of Providence Island Dolostone at Carver's Falls vidence Island overlies the Sciota conformably with a marked change in lithology from the dark grey clay-rich Sciota Limestone to a thin to medium bedded light grey dolostone which weathers yellowish brown. On exposed bedding surfaces the characteristic egg-carton structure is seen which is caused by preferential weathering into joint sets which are perpendicular to bedding and intersect each other at about 45 , the acute angle generally being around the strike direction (Fig. 3.8). This criss-crossed pattern on weathered bedding surfaces, along with the intensive yellowish brown weathering colour, are the two most useful features for identification. The internal features of the rock are rather inconspicuous. The grain size is generally very small and uniform. No sedimentary structures are left, except sometimes a dark grey light grey lamination that could indicate that these rocks were produced by algal mats. More often than not, the beds are internally completely homogenous and on a fresh surface there is nothing within that indicates bedding.

The sequence exposed at Carver's Falls is generally similar to the one described above. Here, too, the beds are very homogenous fine-grained dolostone of grey colour. In the first 20 meters of section, the weathering colour is more of a whiteish-grey and the rocks have a greater amount of calcite in them. Also, interbeds of very pure blue-grey limestone occur which are between 10 and 20 centimeters thick and comprise up to 5% of the section. These limestones are extremely fine grained, weather to a light blue-grey and show a dark lamination in them.

Above, there is 13 meters (43 feet) of completely massive, white

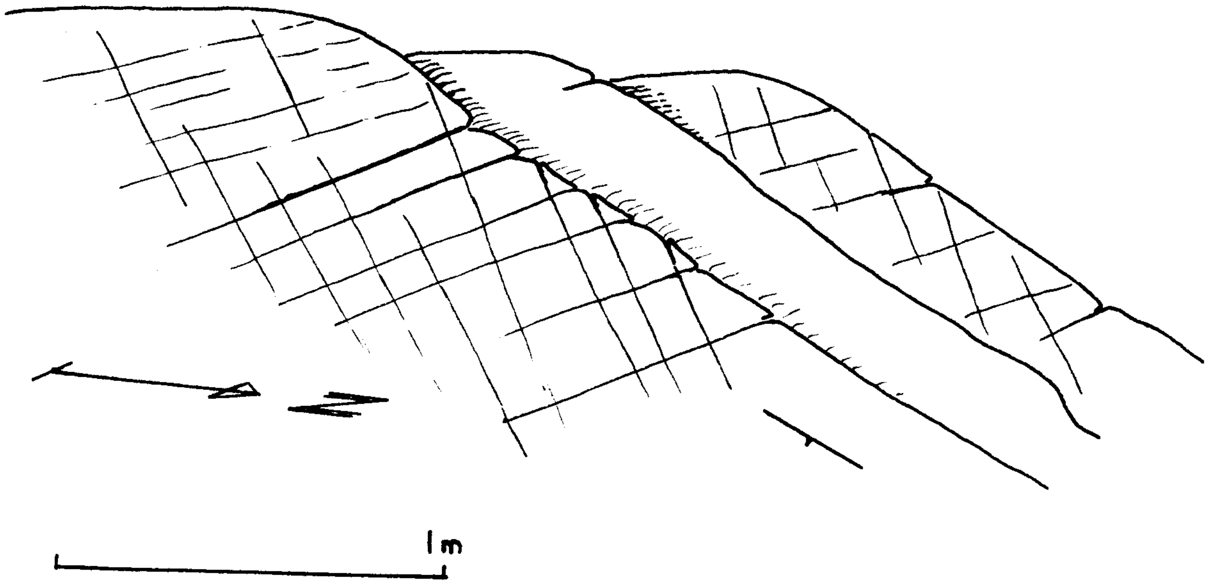


Fig. 3.8: Joint pattern in the Providence Island Dolostone. Beds dip moderately towards the lower left-hand corner. Seen at Carver's Falls.

weathering, light grey dolostone which is strongly veined by very thin calcite veins of irregular orientation. It is these veins that cause a very rough non-planar weathering surface, which looks the same no matter how it is oriented with respect to the attitude of the bedding. Bedding in this subunit cannot be seen at all and it must be assumed that it is parallel to the one seen above and below it. The origin of this peculiar lithology is not clear, and there is little hope that it can be unravelled by field methods, since no structures are seen apart from the late joints and fractures. It appears that the material originally laid down was thoroughly recrystallized during the process of dolomitization. Above, up to the contact with the overlying limestones, there are 42 meters (140 feet) of well bedded dolostone with a few limestone beds. Beds vary in thickness from 5 centimeters to 1.5 meters. Both dolostones and limestones are very fine grained and some of the limestone beds show fine laminae. Where the beds are thin, the weathering colour in the dolostone turns more brown and the fresh surfaces acquire a dark grey to almost black tone. This is due to a higher content in clay-rich material causing also the better developed bedding fissility.

Toward the top of the Providence Island Dolostone, light blueish grey limestone occurs that crosscuts beds of the dolostone (Fig. 3.9). These limestone plugs are up to 50 centimeters thick as they cut through the dolostone beds. Often they seem oriented parallel to the joint system described above. This limestone clearly was deposited in a hole that existed in the dolomite. This hole was probably created by weathering out of the joints through a karst process. This implies,

Figure 3.9: Limestone crosscutting beds of Providence Island Dolostone. Jointed surfaces dipping towards viewer are bedding planes. Limestones are oriented subvertically. From Carver's Falls.

that after diagenesis and before deposition of the limestone fill-ins, the rocks that are now Providence Island Dolostone must have been exposed subaerially for some time, before they returned below sea level again, where limestone was produced to fill the karst holes. This view is supported by the fact, that at the top of the Providence Island Dolostone lies a disconformity which cuts out parts of the Providence Island (Rodgers & Fisher, 1969). This disconformity is what can be observed in the Carver's Falls outcrop.

The age of the Providence Island Dolostone is determined by its stratigraphic position since there are no fossils within it. Going downstream along the south bank of the Poultney River from Carver's Falls, one traverses at least 60 meters of dolomites which are mostly massive and very thick bedded. Laminations are quite common and at one location [4] where the rocks are not as strongly dolomitized, stromatolite heads are recognizable.

Although I have mapped these rocks under the heading Providence Island Dolostone it is not certain whether they are indeed the correlatives of the type Providence Island found in Lake Champlain. While the section at Carver's Falls is lithologically quite similar to the one in the undeformed zone and the descriptions in previous work (Cady, 1956; Welby, 1961; Fisher, 1969; Fisher, in press) it does show some differences. The most striking is the occurence of the massive closely fractured white weathering dolomite in the middle of the section. It has not been reported by any of the previous workers. The existence of limestone beds also is a fairly uncommon feature elsewhere. The section stratigraphically below the exposures on Carver's Falls seems to be even more different from the description given in the previous work. In the field area, these lower parts have a high calcuim carbonate content and are sometimes dolomitic limestones e.g., at the stromatolite locality described before. It must be said however, that because of its dull lithology and the absence of fossils, the Providence Island Dolostone has never been described by anyone in more than a few sentences. At present, there is a study in progress in this department with the aim to obtain a type section of the unit. From samples collected by M. Roma from areas of undeformed Providence Island Dolostone I could see sedimentary characteristics that are not found in the field area. These include horizons full of very small pyrite crystals and banding caused by a change in grain size rather than colour. Another feature typical of type Providence Island are polygonal mudcracks. None of these things have been seen in the area of this study.

Another difference is the thickness of the unit. According to M. Roma who is studying the Providence Island Dolostone in its outcrops around Lake Champlain, the thickness there does not exceed 65 meters (200 feet). In my area the thickness is at least 200 meters (650 feet).

Two explications for these differences are imaginable.

1. The rocks mapped as Providence Island in the eastern deformed part of the area are really in part belonging to formations below it. Most of these sections consist of more or less uniform dolomites and the differences between them are often very subtle. In the folded and thrusted area a distinction between various types of dolostones could

not be made, and stratigraphic position cannot be used as a criterion where the structures are more complex.

2. The eastern zone consists of several thrust sheets of early Ordovician carbonates that override mid-Orovician shale. This requires a big lateral displacement and thus these rocks could have come from an area of entirely different sedimentary history. If the emplacement was from the east, as is most likely (see Chapter IV), then the source area was farther away from the Paleozoic North American continent. The Cambrian/Ordovician carbonates are rocks produced on a shelf and it is known that sediments on shelves generally thicken seaward. This would explain also the greater thickness, if the rocks in the thrust sheets are a time equivalent of the type Providence Island.

Another problem related to this question is that of dolomitization, i.e., whether it is of primary or secondary nature. This is not known for the undeformed Providence Island Dolostone, as Cady pointed out (1945, p. 546). Indicators for both are observed there (Roma, personal communication). Experimental work by Chilingar et. al. (1979) has revealed the role of temperature in the dolomitization process. They found that at temperatures above 60°C, dolomite may form from high magnesium calcite (their Fig. 7.17). These temperatures are almost certainly not achieved in any sedimentary environment, yet it is quite likely that they exist in an orogenic zone at relatively shallow depth. I therefore, favor a secondary dolomitization for the "Providence Island" in the deformed area.

Hence, the name Providence Island should be understood as a litho-unit only, which is used to designate dolomites of similar lithologies in stratigraphic positions immediately below the Chazyan/Canadian boundary.

3.8 Middlebury Limestone

The Middlebury Limestone (Cady, 1945) is the equivalent of the Chazy Limestones, which are found on the western side of the Champlain and Orwell thrusts. Whereas there they can be subdivided into the Day Point, Crown Point and Valcour members this is not possible in the east. A complete stratigraphic description of the Middlebury and the overlying units is impossible from field observations, because these rocks are restricted to the eastern edge of the deformed zone where folding and thrusting are most intense and have great influence on the observed lithology.

The best section in the area under study is found at Carver's Falls, disconformably overlying the Providence Island Dolostone [3]. It has been dated by Welby (1965), who found a new tabulate coral there, as early Chazyan in age. This is the only instance of a confirmed fossil age within my field area.

The Middlebury Limestone at Carver's Falls is a blueish, dark grey, blue-grey weathering clay rich limestone, of fairly fine grain with a few scattered intraclasts of fossil material. At the bottom bedding is marked by dolomitic buff-weathering beds of one to two centimeters thickness that are 10 to 15 centimeters apart. The dolomitic beds make up about 10% of the section, but decrease in number upward, where the rock becomes more and more shaly and at the very top of the Carver's Falls outcrop bedding is marked by millimeter thin clay seams 5 to 10 centimeters apart.

This top part of the outcrop contains the coral found by Welby (located about 15 meters below the dam and just next to the water pipes along the southern edge of the outcrop). It has also several thin coarse biomicrite beds which are rich in fossils. On bedding surfaces, straight and coiled cephalopods can be seen, as well as brachiopods, but they are very recrystallized and not too well preserved. The fossiliferous beds weather forming little round buttons on bedding planes of about one centimeter diameter. These are probably pellets produced by organisms inhabiting the environment of Middlebury deposition. The observed thickness of Middlebury in continous outcrop is 26 meters (85 feet), but probably it is more than double that. Cady (1945) estimates a thickness of "not more than 600 feet". This figure probably refers to the total physical thickness rather than the stratigraphic thickness. It can be much bigger because of sequence repetitions caused by thrusting and intense folding, which are both very common in the Middlebury. I, therefore, estimate the thickness of the unit to be about 100 meters (330 feet).

3.9 Orwell Limestone

The Orwell (Cady, 1945) has also been called Isle LaMotte by Fisher (in press) on the basis of its lithologic similarity with the rocks on that island in northern Lake Champlain. Where it can be distinguished in the field it is a light grey weathering, blue grey rather fine grained, well recrystallized limestone that shows no or very little bedding. Where the rocks are folded a fracture cleavage may obliterate the massive appearance of the Orwell Limestone, making it hard to separate it from the underlying Middlebury Limestone.

The stratigraphic thickness of the Orwell Limestone is unknown because in the map area its observed contacts are all of tectonic nature. It has been mapped separately on the hill north across the road from the West Haven School and west of Route 22A and there the structural thickness is on the order of 170 meters (570 feet). Its age is Black Riverian (medial Ordovician) (Welby, 1961).

3.10 Glens Falls Limestone

The Glens Falls Limestone (Ruedemann, 1912) has been described by Cady, 1945 as a thin bedded, rather coarse grained, light grey weathering, dark grey limestone with characteristic shaly interbeds of several centimeters thickness. It overlies the Orwell Limestone conformably.

The only place where rocks of this type are seen in the field area, is an outcrop on the east side of Route 22A about 3 miles north of Fair Haven, Vermont [5] , shown in Plate IV. They overlie massive limestones in a maximum thickness of 5 meters (17 feet) before they are in turn overlain by black shale which is strongly tectonized. Because of its limited extent the Glens Falls Limestone is not separated from the Orwell and/or Middlebury on the map.

Its lithology expresses the deepening of water at the time of its deposition, by the presence of a substantial portion of shale.

Fisher (in press) and Cady (1945) place the Glens Falls within the Trenton stage.

3.11 Hortonville Shale

The name Hortonville Shale (Keith, 1932) refers to rocks found on top of the Black Riverian and Trentonian limestones described above. In New York State, rocks in a similar position have been referred to as Snake Hill Shale, Utica, or Canajoharie (Rickard & Fisher, 1973), and Normanskill and Austin Glen greywacke where they are coarser grained. It is dangerous to assign a stratigraphic or chronologic value to these units partly because of the paucity of fossils and their obliteration through deformation and partly because these sequences are demonstrably time transgressive (Bosworth & Vollmer, 1980), i.e., they become progressively younger to the west. The name chosen is therefore merely a lithostratigraphic term taken from the Vermont Centennial Map (Doll et al, 1961).

In the area studied these "shales" occupy a large portion of the total area and underlie most of the low lying areas where they are usually covered by glacial till and/or Quaternary lake deposits. Outcrop is generally restricted to roadcuts and beds of swift brooks. They are also seen below ledges of carbonate, where they are near thrust fault contacts and show well developed deformation structures.

The name shale is inappropriate for these rocks because they display a very good, in some instances slaty, cleavage. In most places, however, it is not very regular, but changes its orientation and is crosscut by numerous faults so that the rocks are generally not suitable for roofing slates.

In the field area, three different facies of the Hortonville Shale are recognized:

- A normal variety
- The Forbes Hill Conglomerate
- A (tectonic) mélange associated with thrust faults

In the normal variety, Hortonville is a black shale for the most part. It weathers greenish or yellowish brown. On cleavage planes the colour sometimes shines violet to bronze, like gasoline on a rainy street. This is a staining caused by manganese oxides and hydroxides. The dark colour is due to a high content of organic matter, and occasionally minute pyrite crystals can be observed on cleavage planes. Wells drilled into these shales yield water with a sulfurous taste (B. Bishop, personal communication, 1982). The shales also contain a considerable amount of siltstones and fine grained greywackes and only where these occur can bedding be confidently recognized. The siltstone beds are usually very thin (about 1 centimeter $\pm$) and less dark than the shales. Their weathering characteristics are the same as those of the shales except for a slightly less well developed fissility.

The greywacke beds are very fine to medium grained, show a brown colour on fresh surface and weather greenish brown. They are usually not traceable along any distance in the outcrops, instead they are cut off by one of the many discontinuities visible in every outcrop. These discontinuities are usually at low angles to the cleavage and curved, thus dividing the outcrop surface in an array of phacoid-shaped lenses. Where indicated by greywacke layers, bedding forms a very low angle with cleavage. Sometimes the younging direction can be determined by graded bedding in the coarser greywackes, but it is never consistent even in one single outcrop. This and the bedding cleavage relationship indicate that the rocks are not only faulted but folded also. Proof of folding is seen in numerous fold hinges that are seen in the greywackes. A particularly big and instructive example is seen in Fig. 3.10 which also shows the tight character of most of these folds. Typically, only isolated hinges and or limbs are seen which can be explained by the strong internal faulting of the whole unit so that limbs are cut off from the hinges and vice versa.

Amother important feature of the Hortonville Shale is the presence of pebbles and lumps of black and rare tan quartzite. The tan quartzite was found once as a very big flat boulder about 3 meters in the longest direction, which is aligned parallel to cleavage. It is situated in an outcrop where the normal facies of the Hortonville Shale is quite different from that observed elsewhere. Greywackes and siltstone there make up at least two thirds of the rock. The weathering and fresh colours are lighter than those of standard Hortonville Shale. It is possible, that in this outcrop, which directly underlies one of the thrust faults, there is a rock that is of different origin and age than the Hortonville. In its lithology it resembles some parts of the Taconic Sequence, which now overlies the limestones and shales of the field area structurally.

In the light of the strong internal deformation and the vicinity of large thrust faults, it does not seem impossible that Taconic rocks were tectonically mixed with those below the basal Taconic thrust.

The black quartzite pebbles mentioned above are found quite often in the Hortonville. They are generally not more than 10 centimeters

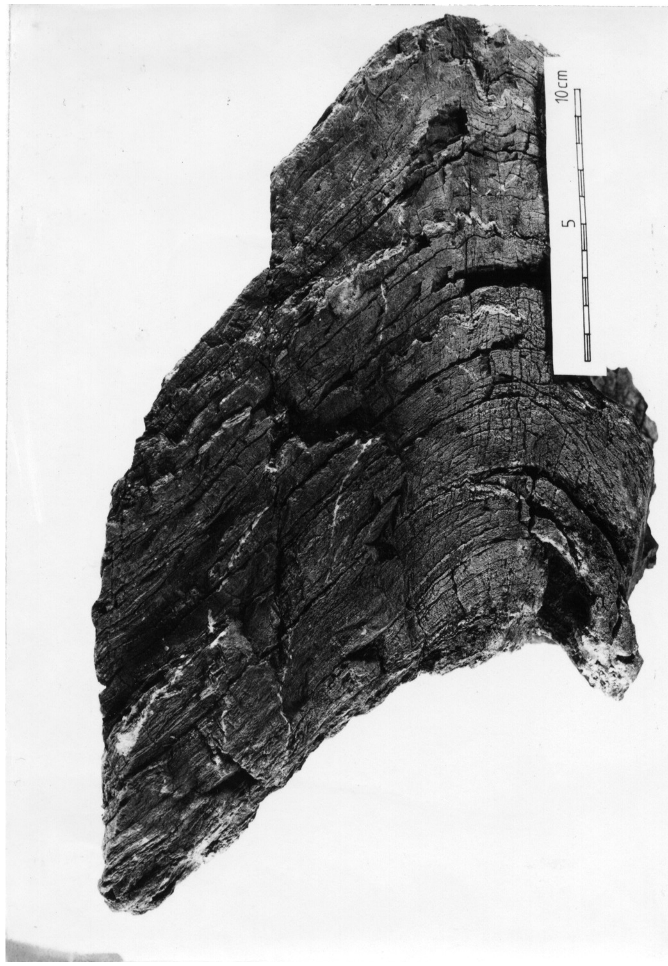


Figure 3.10: Fold hinge from the Hortonville Shale. 600 m southeast of summit of Forbes Hill.

across, are lens shaped and aligned with the cleavage. The latter seems to wrap around them, indicating that these pebbles were deposited simultaneously with the shales and later deformed together. Fig. 3.11 shows a sketch of one of these quartzite pebbles. They are different from the greywackes in that no bedding and no planar fabric can be seen in them, even under the microscope. Also their grains are much less rounded and bigger than in the greywackes.

However, the greywacke beds are often disrupted into such short pieces, that they appear as flat pebbles. In most cases, they show slickenside striations on their interface with surrounding shale, which distinguishes them from the black quartzite pebbles as a product of deformation.

The Forbes Hill Conglomerate (Zen 1961)

This variety of the Hortonville crops out at various localities within the shales in isolated occurrences. The type locality falls in the field area about one kilometer due east of the summit of Forbes Hill [14] . It consists of silicified black shales with a strongly phacoidal cleavage and numerous quartz veins and tension gashes. Fig. 3.12 is a photograph of a slab of this lithology. The Forbes Hill Conglomerate contains exotic blocks up to one meter in diameter. They consist mostly of dark grey to black medium to coarse grained quartzite. Only a few block sized clasts have been seen e.g., 200 meters west of the outcrop of limestones and shales on Route 22A [5] shown in Plate IV. Most of the lumps are smaller - on the order of 10 centimeters across. The smaller clasts are generally sigmoidal in section and lens shaped in three dimensions.

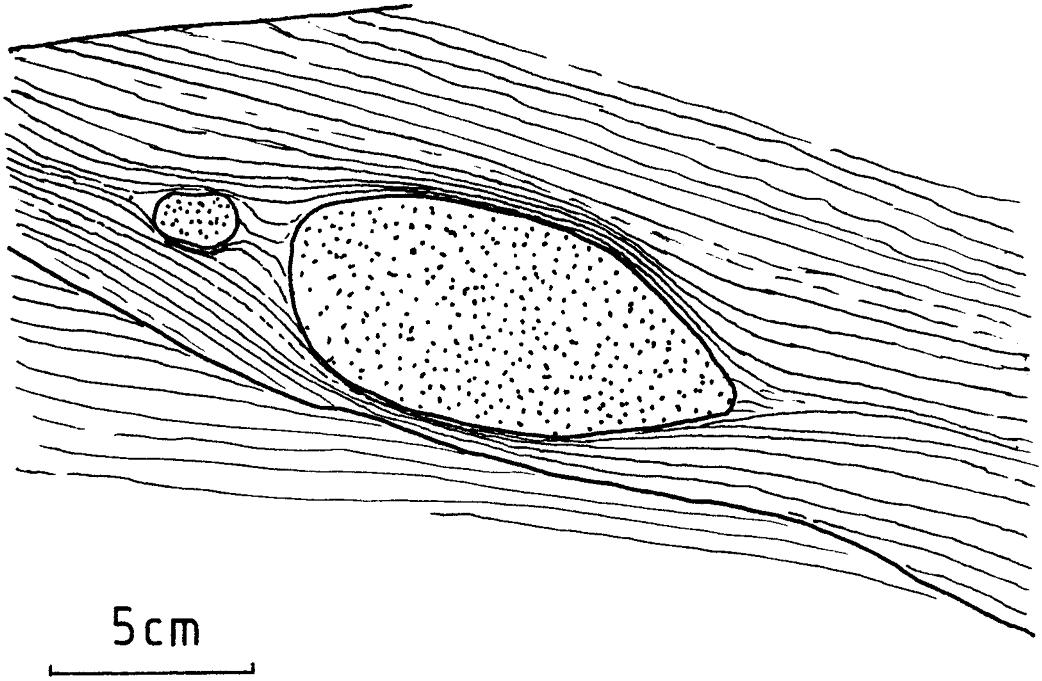


Fig. 3.11: Pebbles of black quartzite in Hortonville Shale. Cleavage wraps around pebble, which lies with its longest axis parallel to it. From outcrop 200 m west of Route 22 A, 7 km north of Fair Haven.

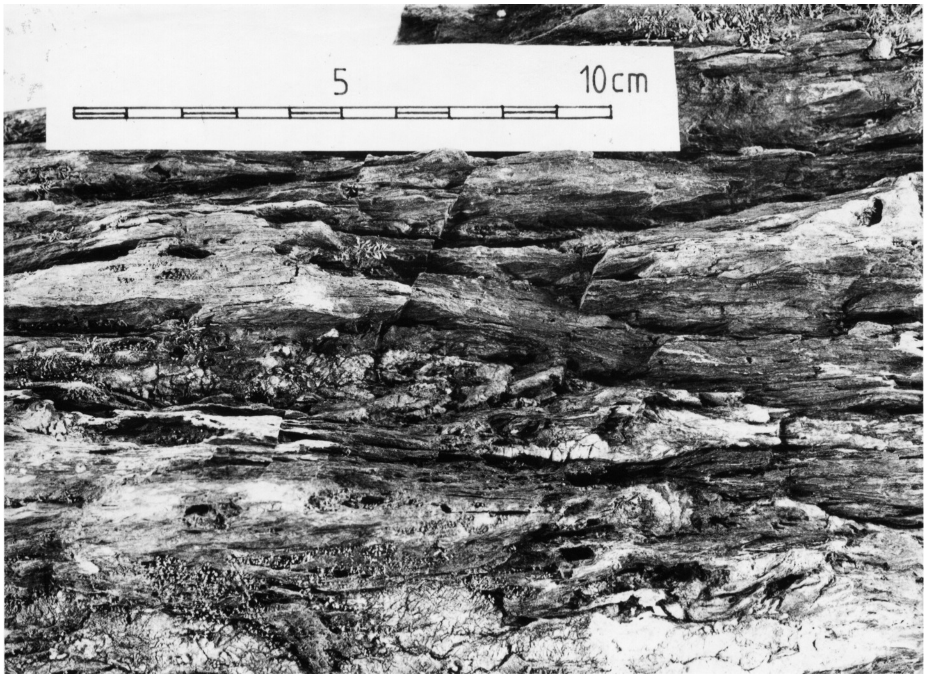


Figure 3.12: Handsample showing phacoidal cleavage in Hortonville Shale. Eastern Forbes Hill.

In thin section the quartz grains show signs of internal deformation like undulose extinction and the formation of subgrains. It can therefore be concluded that the pebbles achieved their flattened form at least in part by deformation. However, the deformation visible in the microstructure can also have occurred before the quartzite pebbles were eroded and incorporated in the Forbes Hill Conglomerate.

Besides quartzite, limestones and greywackes are also found as lumps within the shale matrix, but they are much less abundant. One limestone clast seemed to consist of a central circular part and two flanges adding to a sigmoidal shape. It is thought that they could have been formed by pressure solution.

Fault Mélange

This is a pitch black, soft but very fissile shale found in immediate contact with carbonates at thrust faults. Its colour stems from a high content of organics and pyrite, which can often be seen in crystals covering cleavage planes. It weathers very quickly and the pyrite produces a rusty weathering colour on parts of the outcrop.

It is never thicker than 5 meters and only seen in contact with limestone. A good exposure of this rock can be studied in a roadcut on the east side of Route 22A about 5 kilometers north of Fair Haven [5]. Plate IV is an outcrop sketch of this locality.

The clasts in this black mélange are derived from adjacent limestone units and are very tabular shaped. They are always subparallel with the cleavage.

This rock is certainly a product of tectonic movements between the "normal" Hortonville Shale and the overriding carbonates. On several surfaces slickenside striations were found. With respect to its internal deformation it is difficult to assign a thickness to the Hortonville Shale. The values from the literature (e.g., 660 meters, Welby 1961; 700 meters Fisher in press) are probably much too high and represent structural rather than stratigraphic thickness.

CHAPTER IV

STRUCTURAL GEOLOGY

4.1 Introduction

The purpose of this study was mainly to investigate structural features. In particular, the amount and style of deformation of rocks immediately underlying the Taconic Allochthon was to be clarified. In the Geologic Map of New York State (1970), the basal Taconic thrust is underlain by a unit called Taconic Melange with an outcrop width of 3 - 10 kilometers. In a more detailed map (Fisher, in press) the same unit will appear.

This unit supposedly consists of mainly black shales and siltstones which contain exotic fragments of Taconic and mainly, shelf lithologies. Some of these are big enough to be shown on the New York State Map (1:250,000 scale). Despite their enormous size, they have been interpreted as olistostromic blocks, that is as sedimentary in origin (Fisher, 1970; Rodgers, 1982; Bird, 1970; Zen, 1961). Fig. 1.5 shows an enlargement of Fishers' recent work in the southern part and to the south of my field area. One of my aims was to test his "blocks in shale" idea by examining structures within both the "blocks" and the "shale" and the nature of their contacts.

For description, my field area is divided into two domains:

1. Western Undeformed Zone
2. Eastern Deformed Zone

The boundary between these two corresponds to the westernmost low angle fault shown on the map (Plate I) and separates the Sciota Limestone from overlying Canadian carbonates. (Ofcs and Ofcp respectively on map).

The two terms for the area are meant to be just names, they do not exclude the occurrence of deformation in the west or the local absence of the latter in the east. In general terms, however, it can be said that deformational structures are much more prominent in the eastern half of my field area.

4.2 Structural Associations in Different Lithologies

Before describing and analyzing the structures in detail, a few remarks about general structural features in the different rocks are necessary.

First, there is a striking difference in outcrop structures between carbonates and shales/slates and to a lesser extent between limestones and dolostones. This difference is expressed in both primary structures and tectonic structures. In the shale-slate sequence bedding is usually not prominent. The cleavage is the dominant planar structure in these rocks. It is often not possible to determine if this is a shaly parting or a genuine slaty cleavage. In the carbonates, however, the bedding is dominant and only in more strained positions such as tightly folded zones or close to faults does it become overprinted by cleavages, or fractures.

Fractures and joints are in turn the more prominent planar features in the dolostones, especially where they are poorly bedded or massive. Sedimentary structures in the shale are only recognizable in places where layers with coarser grain size exist. There compositional differences express themselves by colour changes, differential weathering and, in places, graded bedding. Since these beds are rare and in general not traceable even over a single outcrop an internal stratigraphy of these shales cannot be defined.

In the carbonates, however, primary structures (bedding, cross-bedding, algal laminations, some fossil layers, grains and cement) are fairly well preserved, especially in less deformed parts. They allow the establishment of a clear stratigraphic sequence of carbonates grading from mainly dolostones below to pure limestones and argillaceous limestones at the top of the sequence. (See Chapter III for details). The amount of deformation is quite clearly visible by means of deformation on individual beds and I have been able to determine in one place a numerical value for strian, based on deformed fossils. (This figure is, of course, only locally valid, and should never be taken as representative for the whole area.)

Whereas folds in the carbonates do not disrupt the layering, dismembered folds are a common occurrence in the shales. The shales, especially in the vicinity of faults are thus stratigraphically and structurally discontinuous, and the deformation in them must have been much greater than in most of the overlying carbonates.

4.3 Faults

Faults are easily the most prominent feature among the structural elements. The topography is greatly influenced by them and some unusual relationships can only be solved by fault contacts. Both low angle and high angle faults can be demonstrated in the field.

Low Angle Faults Placing Carbonates over Shale

The low angle faults are most easily recognized where they place Beekmantown Carbonates over black shale. The great difference in resistance to weathering produces ledges often a hundred feet or higher, where limestones or dolostones form the top and black shale is found where the ledge gives way to a gentler slpe, usually made up of carbonate debris and the underlying shale (Fig. 4.1). These contacts are those that bound Fishers' "blocks" from shale surrounding them.

In his view, these surfaces are in essence the surface of oversize pebbles, making them spherical structures enclosing a volume, namely that of the block within them. In a map they should form closed rings around the outcrop of carbonate blocks, as shown on the New York State Map and its revised version (shown as Fig. 1.5 in this thesis). However, in my mapping I found the limestone-shale contacts only on the western side of the carbonate "blocks" (see map Plate I) and the ledges that expose them all face westward. The shale carbonate contacts on the eastern sides of the blocks are seldom exposed and never form ledges. All the low angle contacts seen are dipping toward the SE as can be seen from Fig. 4.2 showing 16 poles to the low angle fault contacts from 12 different locations. The circles in the stereogram represent contact orientations deduced from cleavage in the melange directly underlying the contact. Fig. 4.3 shows in cross sectional view how the cleavage directly below the contact parallels the latter. These measurements are probably more representative of the general orientation than those taken directly, because the contact was found to be quite wavy with wave amplitudes on a meter scale or even bigger.

In outcrop these contacts are readily recognized as structural and stratigraphic discontinuities. They are very sharply defined with a very small zone of mixing (less than one meter in general). The con-

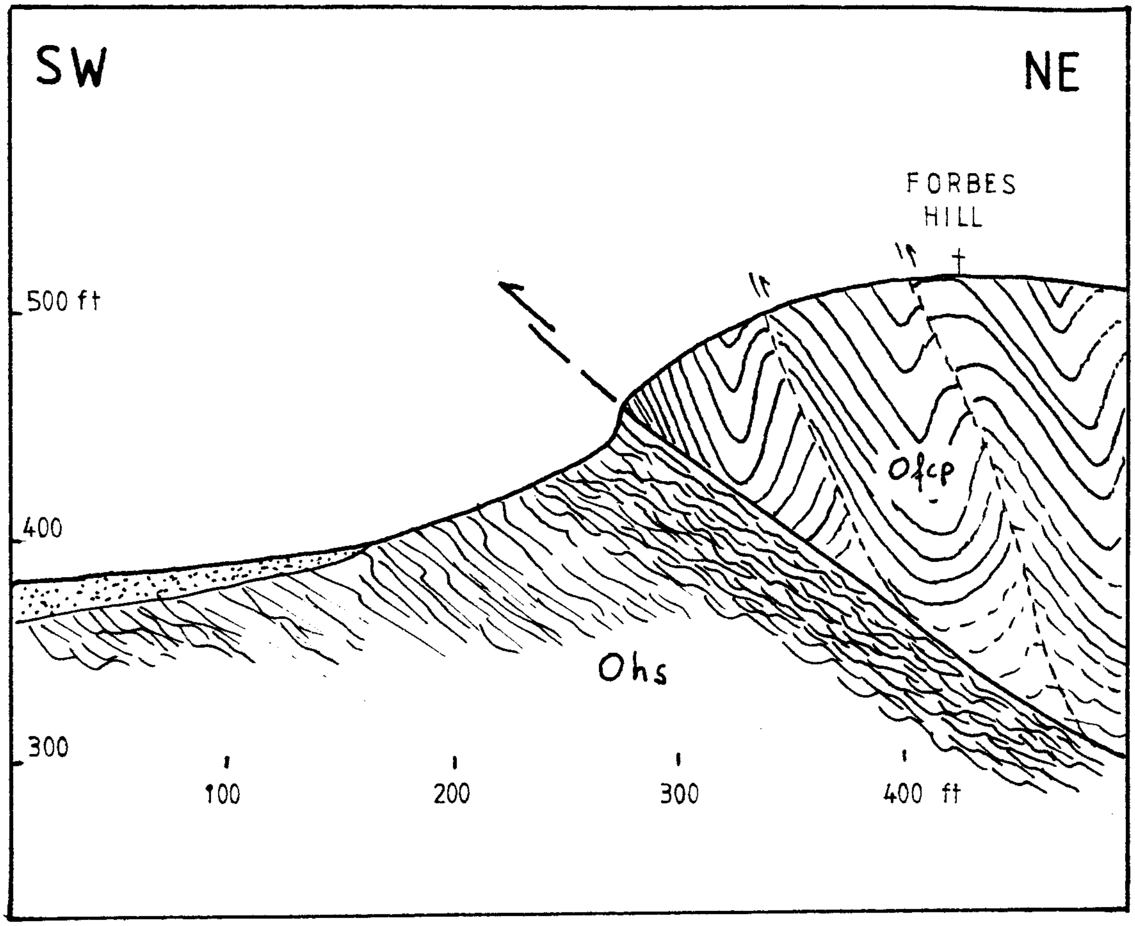


Fig. 4.1: Typical fault scarp topography for cases where carbonates are thrust over shale. Ohs = Hortonville Shale, Ofcp = Providence Island Dolostone.

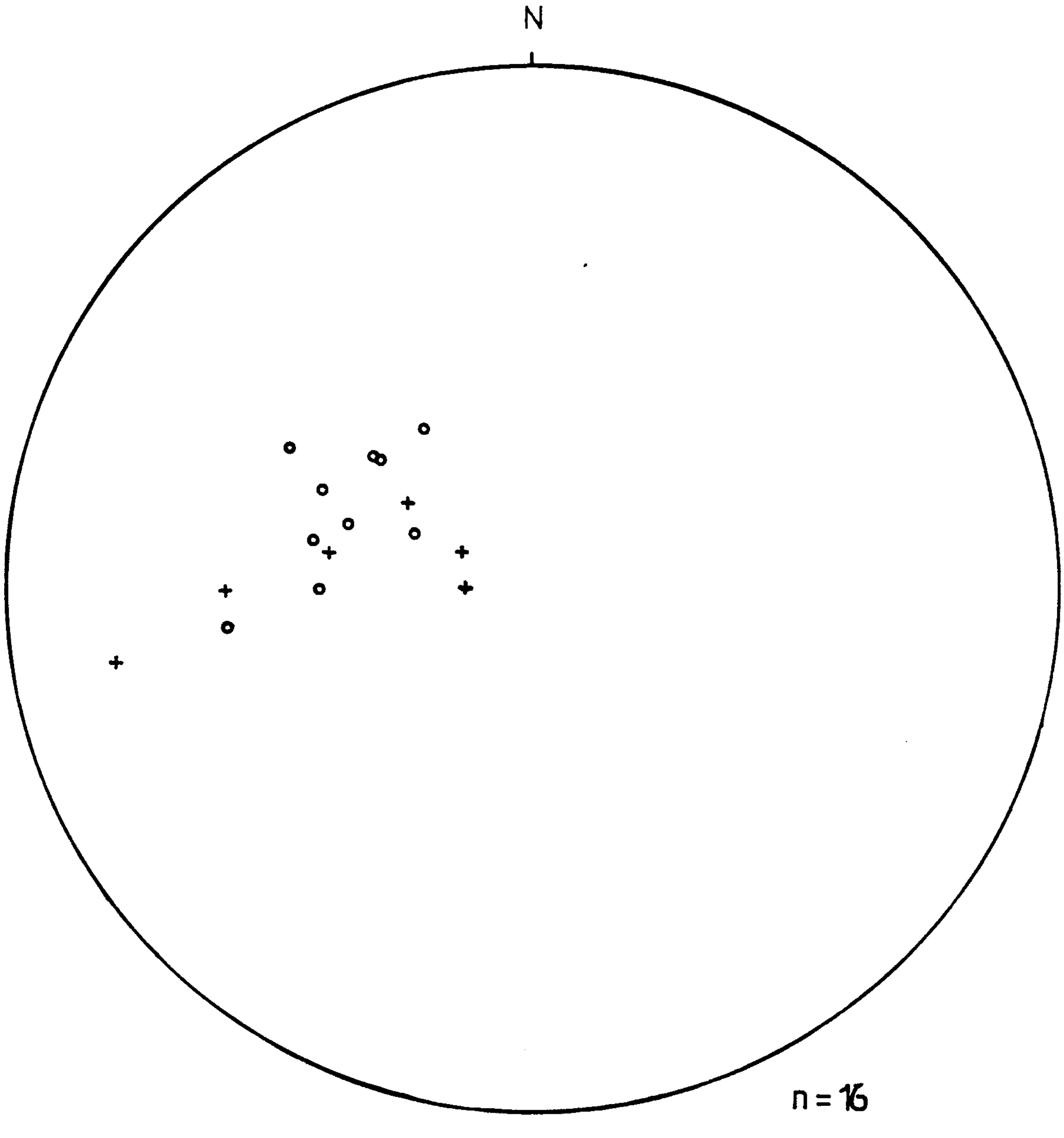


Fig. 4.2: Orientation of low angle faults placing carbonates over shales. Crosses: poles to fault plane. Circles: contact orientation inferred from cleavage. Equal area stereographic projection.

Figure 4.3: Reorientation of cleavage parallel to low angle fault, suggesting right to left (i.e. east over west) movement of the overlying carbonates. Outcrop 2.25 km southsoutheast of Sciota School.

tact surface undulates in places and separates a few small carbonate blocks from the main mass, surrounds them by dark grey to black shale, with a cleavage wrapping around these blocks. On Fig. 4.4 it can be seen, that the contact is not always parallel to bedding in the carbonates, and that the latter are folded. This is very characteristic for all of the observed shale-limestone contacts. (The implications of this and other observations are discussed at the end of this chapter.)

In rocks immediately above and below the contact, slickensides and slickenside striations are common, in many cases made up of fibrous calcite. The fault zone itself is usually too weathered to show any clear microstructures. The observed slickensides in the limestones are connected with high angle faults of lesser importance and may not directly relate to movements on the master fault. In the shales below it, they lie within the dominant cleavage where they separate "pseudo-stretched pebbles" (Hills, 1953, p. 134) of dark quartzite and sandstones from black shale surrounding them (Fig. 4.5). These psammites are always oblate slabs of material in three dimensions and can usually be lined up with each other following the cleavage trace on the outcrop surface. Slickensides and the lateral continuity of the slabs point to them as being tectonically disrupted beds of the shale sequence.

One structure found immediately at the carbonate-shale contact in one particularly well exposed contact on the northern bank of the Hubbardton River 1700 meters south of the summit of Forbes Hill [6] is a lineation on cleavage planes of the "fault rock". It parallels the local dip direction of the cleavage with a trend of 80° and a plunge

Figure 4.4: Low angle fault placing Orwell Limestone over Hortonville Shale. Outcrop south of field area, about 5 km northeast of Whitehall, New York.

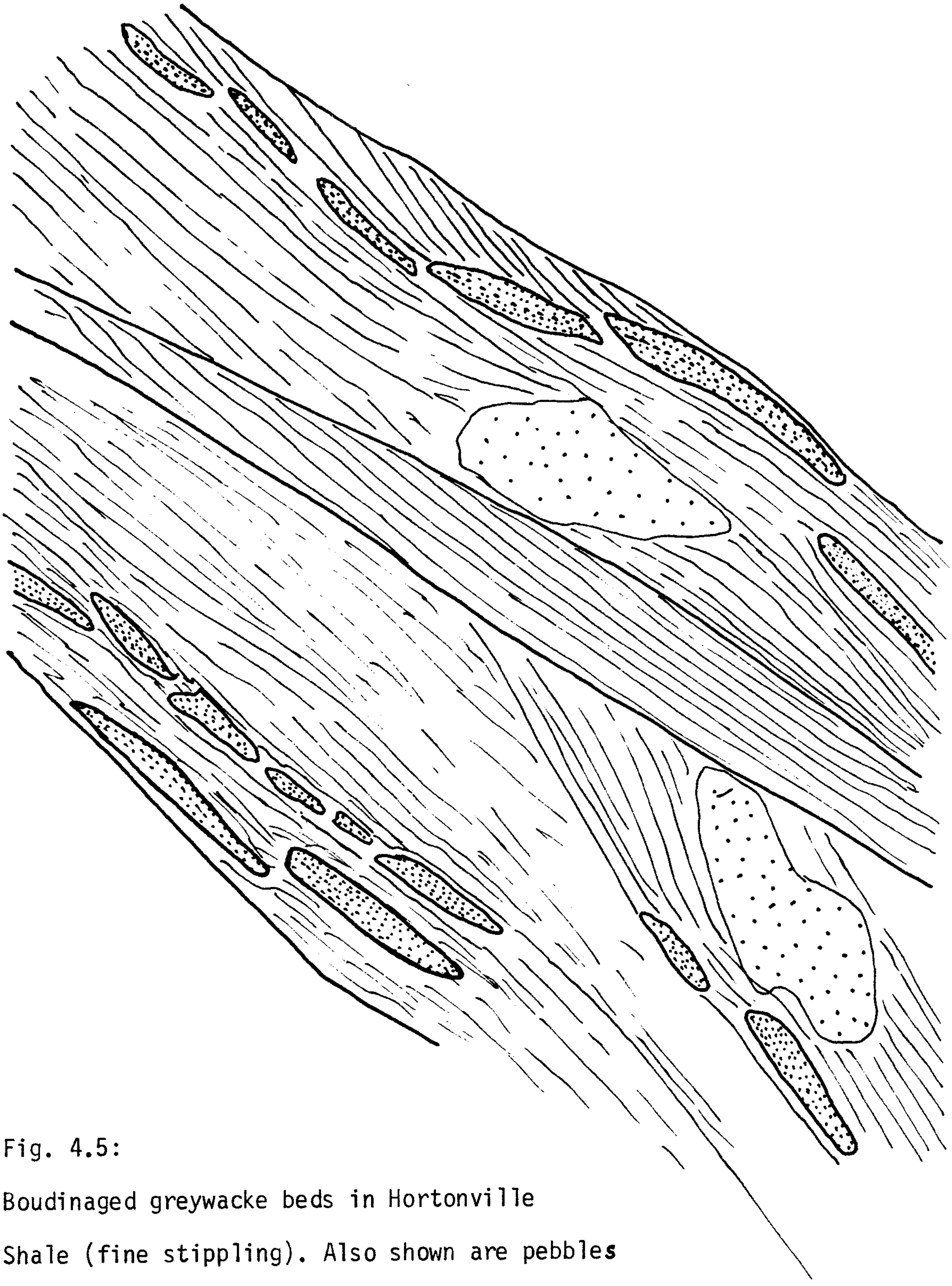


Fig. 4.5:

Boudinaged greywacke beds in Hortonville Shale (fine stippling). Also shown are pebbles of black quartzite (coarse stippling).

Same locality as fig. 4.3., 7 meters downsection.

69°. The fault rock is a dark black, very soft flakey slate with no visible grains, except for minute euhedral pyrite crystals seen with a hand lens. Often this rock is stained in a rusty colour on weathered surfaces, which probably stems from oxidation of the iron contained in pyrite. The iron sulfide could explain the deep black colour of the rock. (Another possibility is a high carbon content, which may be responsible for the softness of the material.)

Another structure that is found close to these faults are quartz veins and tension gashes in the overlying carbonates (Fig. 4.6). Sometimes they form en echelon arrays and show usually a nearly vertical dip on outcrop surfaces which are at high angles to cleavage. However, their orientations are by no means regular enough to fit the textbook interpretations. As can be seen in Fig. 4.6 they are in places truncated and offset by minor faults in the carbonates. In a few cases, they offset and deform each other (Fig. 4.7).

The offset and deformation are left lateral in a south facing outcrop, demonstrating thus an east over west movement within the limestone.

Apart from the few indicative examples, it is more the presence of tension gashes and veins than their orientation that can be related to the faults. Their composition varies from pure calcite to pure quartz and combinations of these two.

It was noted that wide veins (more than a centimeter thick) are restricted to limestones, whereas in the dolostones they are of millimeter size, are a more penetrative feature, and no obvious tension gash sets are seen there.

Figure 4.6: Offset tension gashes immediately above thrust contact of Middlebury Limestone above Hortonville Shale. North side of Hubbardton River [6].

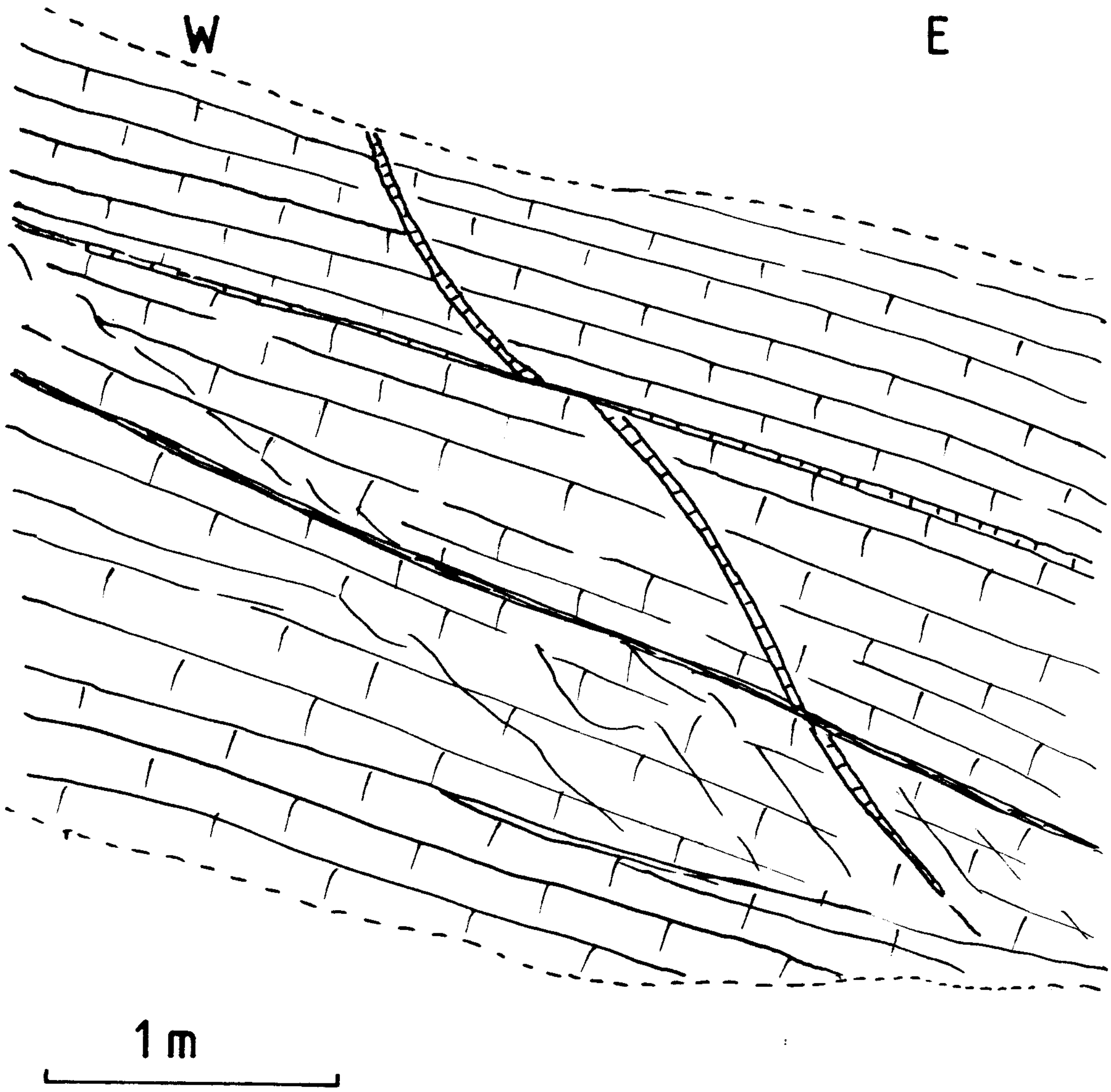


Fig. 4.7: Left lateral offset of veins in limestone immediately below thrust contact. Roadcut 700 m south of West Haven Town Center.

One very important observation from the limestones directly above fault contacts with black shale has to be mentioned here. In several places on Forbes Hill and elsewhere the fabric of such rocks changes locally in the vicinity of faults. The change seen is a strengthening of the cleavage to a degree where it completely wipes out the sedimentary layering and gives the rocks a highly foliated appearance. The foliation is then a compositional layering of dark grey, grey and white layers about a millimeter thick, but with varying thicknesses as drawn in Fig. 4.8 (from a polished slab obtained from Forbes Hill). The lensing and coalescence of various layers excludes a sedimentary origin of their present geometry and leads me to conclude that these are structures formed by tectonic processes.

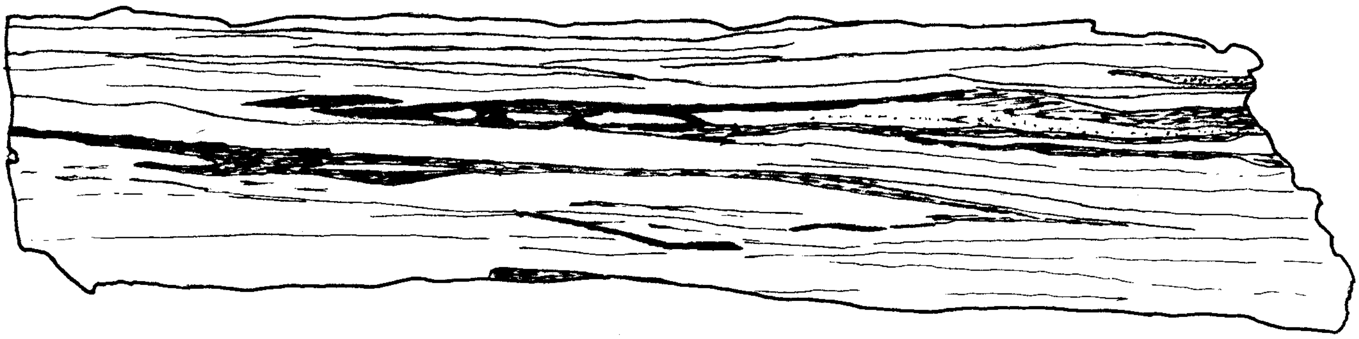
The following observations support the view that the low angle boundaries are faults (the last 5 are evidence for thrust faults):

1. The discontinuity of stratigraphy and structure at the contact: it cuts through bedding in the carbonates and reorients cleavage in the underlying shales (Fig. 4.3).

2. The consistent southeastward dip of the contacts: this is in favor of a failure of rocks in a regular way in response to a directional stress field, rather than flow of blocks in a soft shale on a submarine slope.

3. The existence of tension gashes, extensional quartz and calcite veins, slickensides and slickenside striations and disrupted beds: these features are commonly observed next to fault zones and are related to them.

4. The dark grey to black pyritic fault gouge observed in many contacts.



TRUE SCALE

Fig. 4.8: Strongly foliated limestone from Forbes Hill [15]. Possibly a mylonite.

5. The fact that veins crosscut and offset each other with an east over west sense: it indicates that deformation was going on after lithification of the carbonates and after extensional veins had formed and that this deformation was a shear affecting the whole body of the rock.

6. The fact of cleavage swinging around toward parallelism with the fault in the underlying shales. If this is caused, as I believe, by movement of the overlying carbonates it requires an east over west transport of the latter.

7. The presence of folds in the overlying unit requires compression therein. The underlying unit shows no folds, so that the difference in lateral extension must be taken up by a fault between them. The consistent westward vergence of these folds can best be explained by a shear component toward the west with repect to the east dipping fault plane. This leads to an updip movement on it, i.e., a thrust fault.

8. It is known from the regional geology that my field area is immediately adjacent to a major overthrust namely the Taconic Basal Thrust, along which large east over west movements occurred in a time following the deposition of the middle Ordovician limestones in my area.

9. Finally a stratigraphic argument: The shales underlying the carbonates are known to be younger than the carbonates (see Plate III, Fisher, in press; Welby, 1961; etc.). If the carbonates are found above those it requires that they be moved upsection and that is likely to involve thrusting (see Hobbs, Means & Williams, 1976, pp. 301-309.) Hobbs, Means & Williams (1976, p. 309) give a well known method of determining the minimum displacement as the distance between windows

and klippen of the same thrust sheet.

In my field area several thrust sheets have been identified. Assuming the dip of the thrust surface to be less than 45° and the dip of the top surface of the structurally underlying limestone to be no less than half that (22.5°), simple geometrical considerations as shown in Fig. 4.9 yield that the displacement on these thrusts is of the same order of magnitude as the distance between the thrust and the next limestone, or more. The displacement estimate is only correct, however, if the frontal limestone sheet is still parallel to the décollement horizon, as shown in the figure. This is believed to be true, for the Frontal Thrust Sheet overlies the undeformed sequence and its basal thrust follows the dip of the top of the Sciota Limestone, the top of which represents the décollement horizon.

Low Angle Faults within Carbonates

Due to the similar weathering resistance of limestones and dolostones, faults within those lithologies are unlikely to have a marked effect on topography like to ones described above. The similarity of lithology on either side of these faults makes it very easy to overlook them.

The erosional activities of the Poultney River have provided me with one big outcrop [8] where a low angle fault can be demonstrated within carbonates. It is located 2 kilometers downstream of Carver's Falls and 300 meters north of the river bed. Fig. 4.10 a and b show pictures and a line drawing of the observed relationships.

The contact separates dolostone above and limestone below. The dolostones are medium bedded, typically yellowish grey weathering and

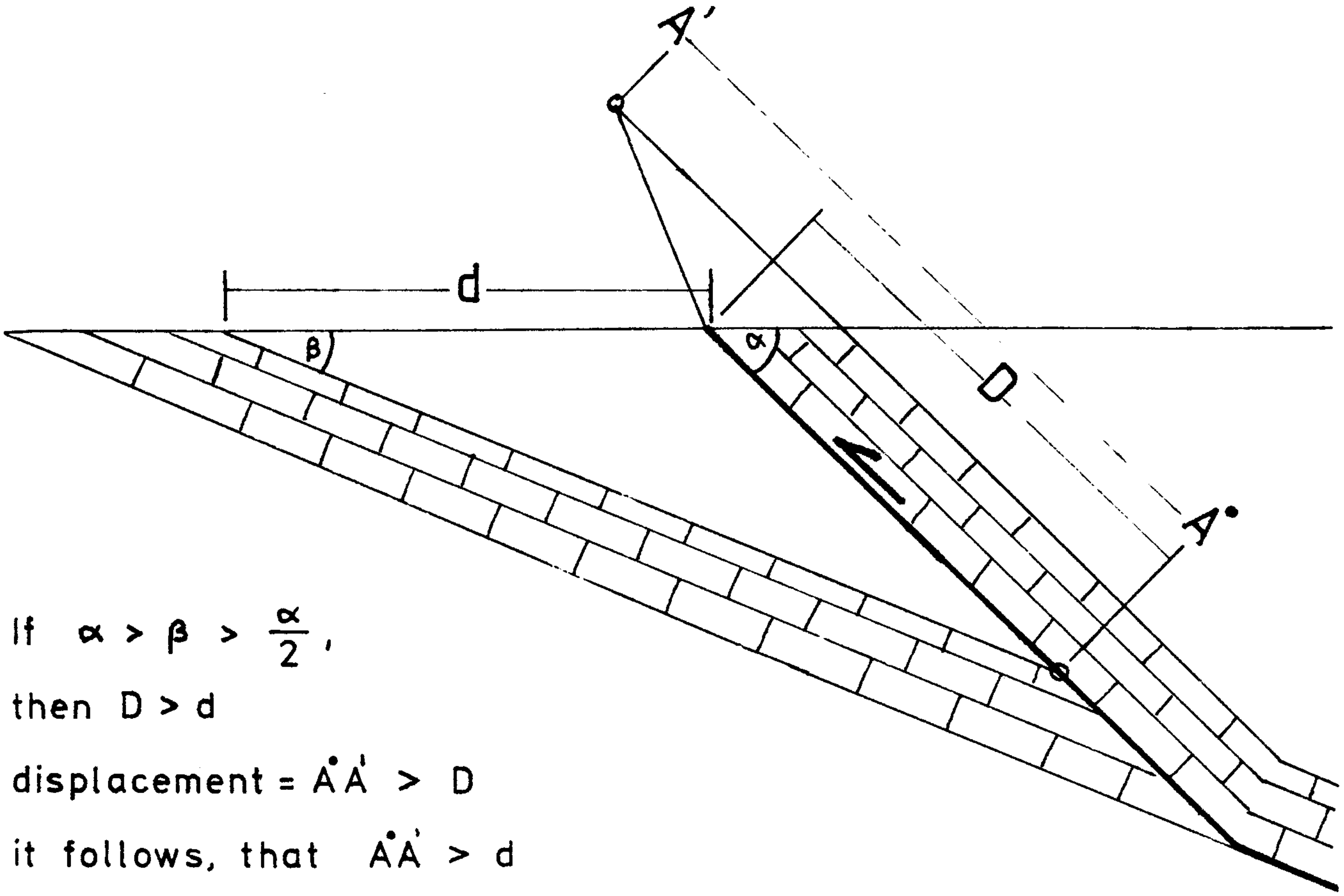


Fig. 4.9: Method to estimate a minimum fault displacement value for thrust faults that dip no more than two times steeper than the decollement horizon.

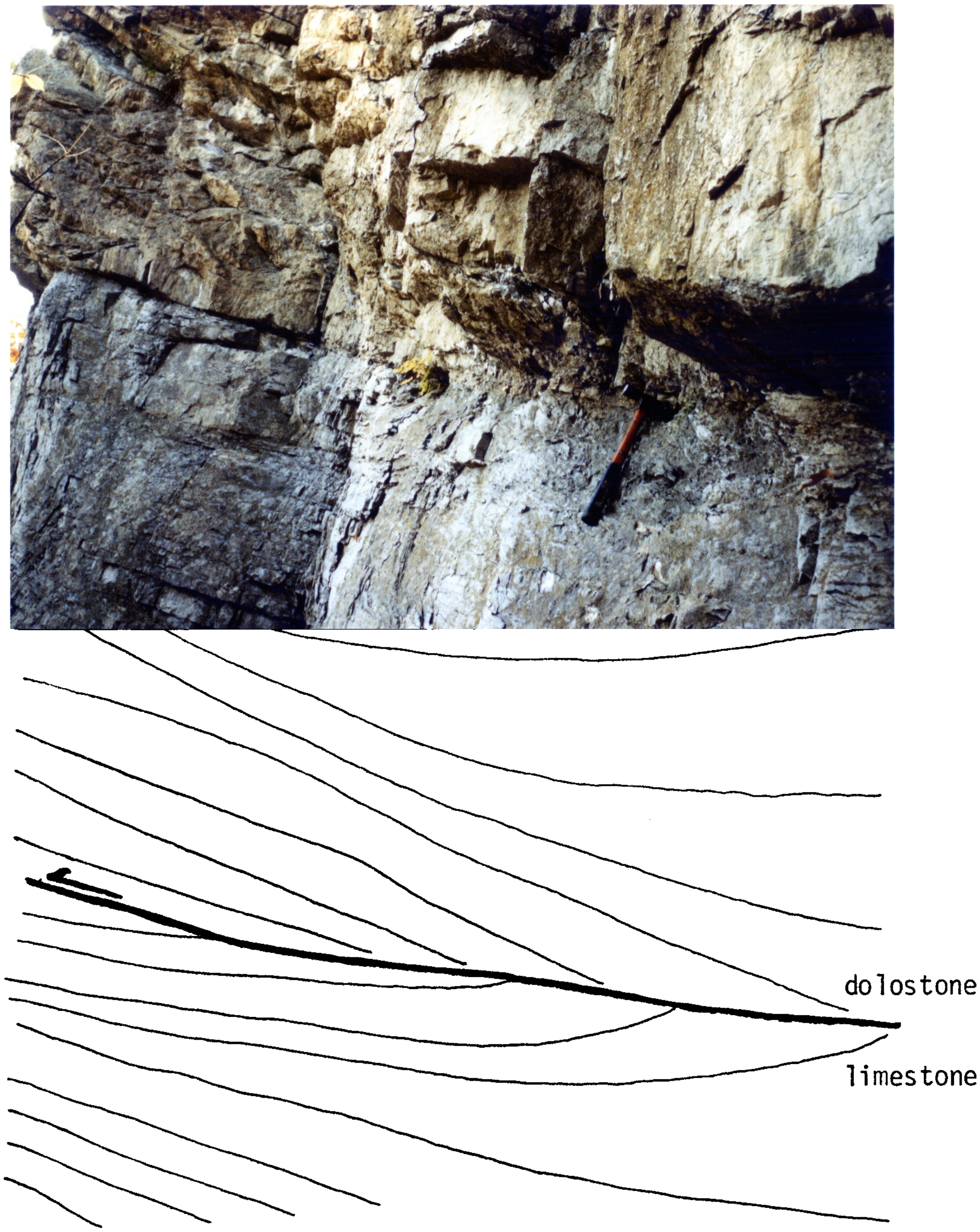


Figure 4.10a: Dolostone thrust over limestone. Note truncated bedding on either side of thrust contact. North bank of Poultney River [8].

Figure 4.10b: "Swing" of fracture cleavage towards fault plane in vicinity of thrust. Pencil indicates cleavage orientation and its top the position of the thrust. Same locality as fig 4.10a.

apart from bedding no sedimentary structures can be observed. They are apparently completely recrystallized. They dip at 35° toward 120° immediately above the contact. The dip varies across strike and in other parts of the outcrop folds with about 10 meters wavelength and broken hinges in anticlines are observed (Fig. 4.11).

The underlying limestones weather blueish-grey, have thin beds that are separated by clay-rich dark layers. The beds dip 30° toward 062° . Apart from bedding which is rather coarsely defined, but clearly visible, a set of regularly spaced joints at approximately 10 centimenter intervals is prominent. These fractures are more or less perpendicular to bedding and strike at about 170° . In the vicinity of the fault contact, they swing around toward bedding as shown in Fig. 4.10 b. At first sight, this indicates a sinistral movement in the south facing outcrop plane, and therefore east over west thrusting. Since the joints are only seen in the limestones and swing around and are terminated by the fault, they probably existed prior to the faulting. Unless the orientation change in the joints was there before faulting it seems reasonable to connect it with the latter. It could be an expression of simple shear affecting the rock immediately prior to faulting.

The following observations support the notion that the contact observed is a thrust fault:

1. The orientation of initially bedding-perpendicular joints in limestones gradually changes toward parallelism with the contact in a 50 centimeter wide zone immediately below it.

2. The fault cuts the bedding of both the overlying dolostone and the underlying limestone.

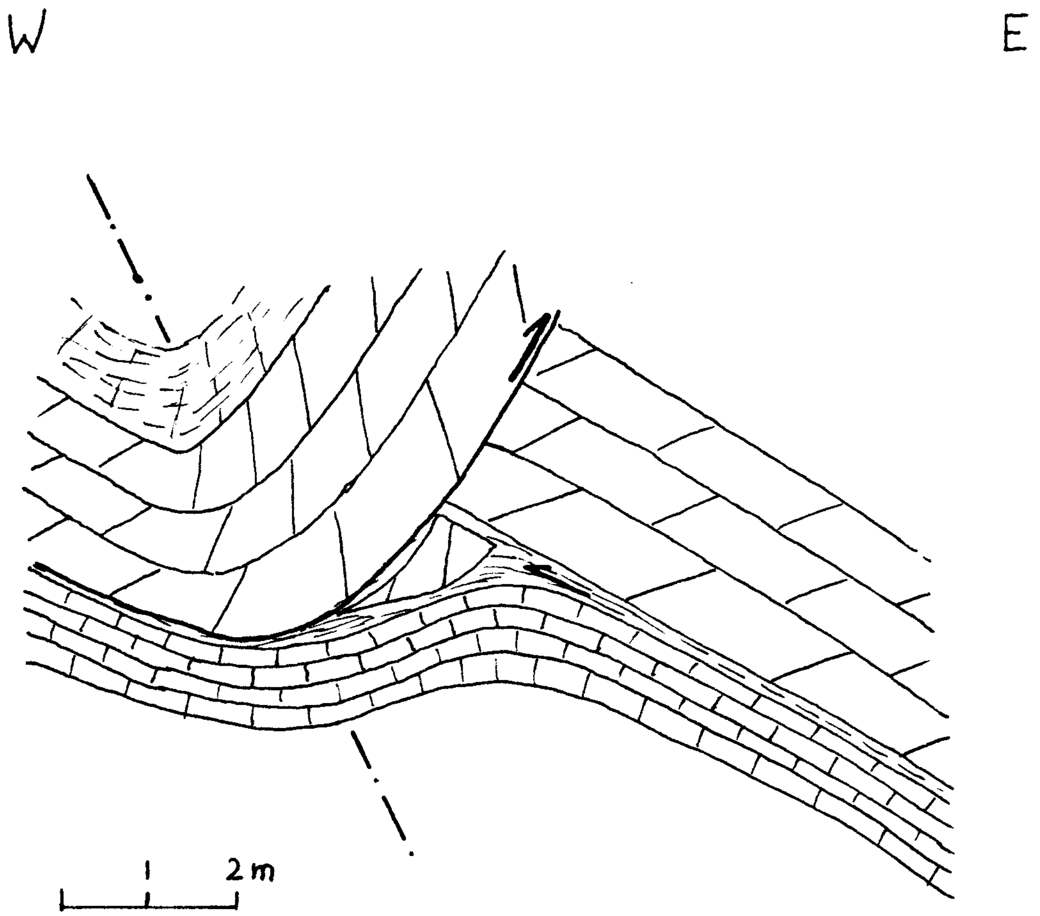


Fig. 4.11: Syncline-Anticline pair with disrupted fold hinge in the anticline. Situated above thrust within carbonates. North bank of Poultney River [8].

3. The dolostones are stratigraphically below the limestones in the unfaulted sequence. Yet in the outcrop they structurally overlie the limestone directly. This relationship requires a thrust fault.

4. Dolostones in the upper half are seen folded with west vergent asymmetry (Fig. 4.11). This speaks for an east over west shear component causing the observed structures. In the case of a dip slip movement, one would expect structures caused by extensional forces in the overlying unit, and that is not the case.

Faults within Shales

Faults within the Hortonville Shale are very difficult to determine on a map scale. This is due to the absence of markers such as bedding or a clear stratigraphy. In individual outcrops they are, however, clearly visible. There they are recognized by the fact that they interrupt the cleavage and place different orientations of the cleavage next to each other. Also, they are often lined by quartz veins, which are usually of massive appearance. Their shape is usually curved and they form an anastomosing pattern on the outcrop surfaces.

The variation in fault orientation is large, but they always dip toward the east. Fig. 4.12 shows a stereoplot of 16 of these faults together with cleavage attitudes in their vicinity. It can be seen that their general attitude is quite similar to that of the cleavage. Since they cross cut cleavage they must have developed after it. In many cases the bending of the cleavage increases toward the faults, and this indicates, that prior to faulting there was folding.

It may or may not be that the folds seen in outcrop are connected

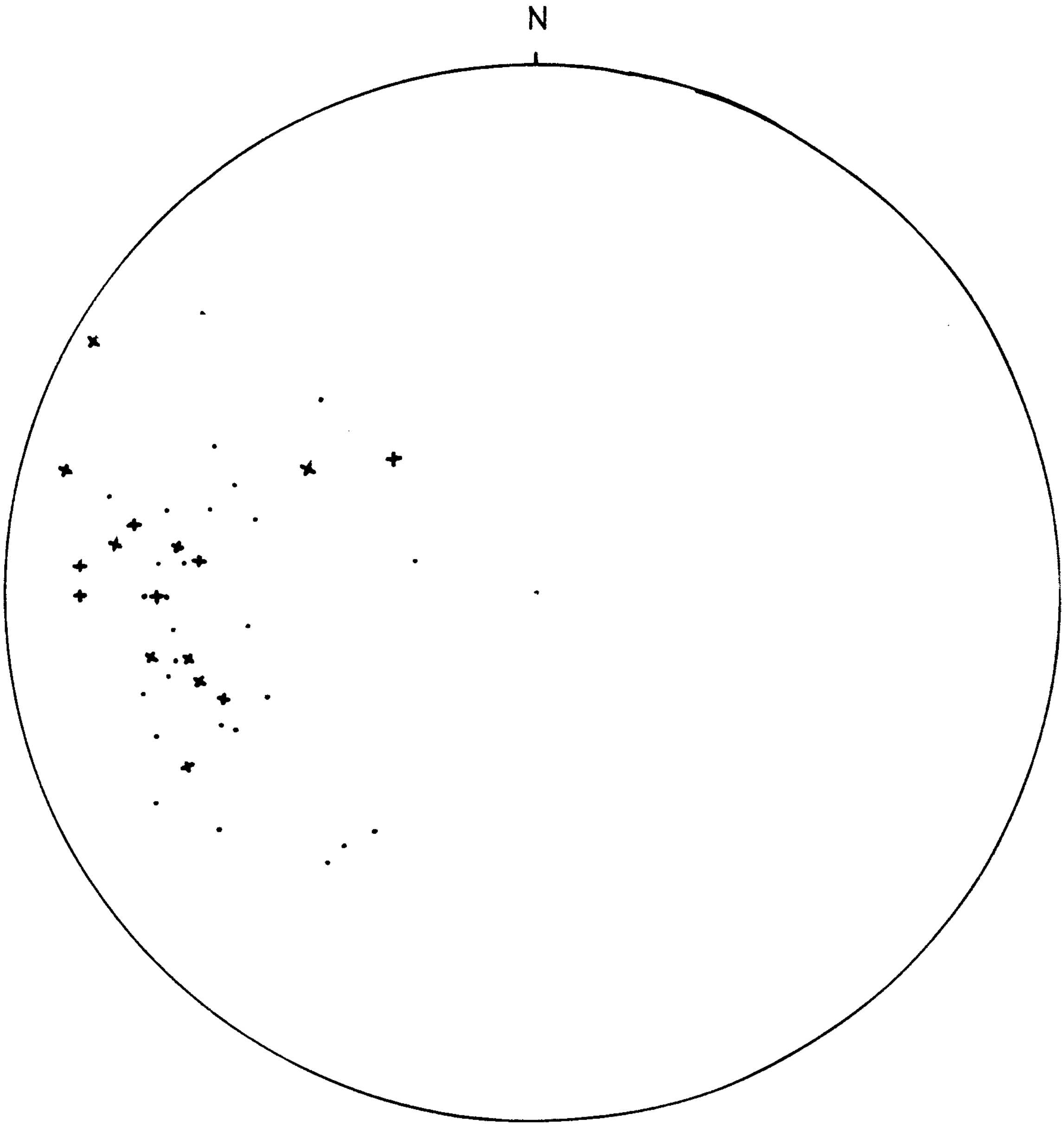


Fig. 4.12: Equal area stereographic projection of poles to fault planes (crosses) and cleavage (dots) from the Hortonville Shale. Faults n=16, cleavage n=27

beyond a single outcrop and represent one group of faults. Their attitude close to the cleavage and their eastward dip is however, consistent with thrusting demonstrated for the carbonate rocks.

It may be that within the carbonates shear was accomplished on one planar surface whereas within shales, the zone of shear deformation was wider and the anastomosing faults seen on an outcrop scale might be internal structures of such a larger shear zone.

Normal Faults

Several normal faults of importance have been mapped in the field area. They fall into two groups: those oriented north-south with a west side down displacement and those oriented northeast-southwest with the hanging wall on the north western side of the fault.

The displacement of the faults has been determined mainly by stratigraphic evidence and therefore the displacement is only determinable in the western part of the map area, where the stratigraphy is established. The larger normal faults are also prominently expressed in the topography.

The southernmost of the soutwest-northeast trending faults follows the Poultney River approximately and has therefore been named after it. It strikes about 80°. The exact location of the Poultney River Fault is unknown, because it is buried under Quaternary sands that form the base of the river valley. Its existence is derived from the fact that the boundary between the Whitehall and Great Meadows Formations is offset in an apparent left lateral sense by 550 meters. This left lateral offset is, however, only apparent and caused by the eastward dip of 10°

of the sequence both north and south of the fault. The field relations are shown in Fig. 4.13. It shows that the fault displacement in that location measured vertically is 105 meters, the hanging wall being north.

The Warren Hollow Fault lies from 1 to 2 kilometers north of the Poultney River and strikes 063°. It is marked by a deep valley that extends over two thirds of the width of the field area and is narrow in carbonates and wide where it runs through shale. One outcrop in its immediate vicinity shows the effects of faulting on adjacent rocks. It is situated about 300 meters west from where the Warren Hollow Fault crosses another strongly cut-out valley which runs north-south [9]. Neither of those valleys has a stream in it.

The rocks in the outcrop [9] are beds of Sciota Limestone, that have been fractured and folded. In places faulting has produced typical fault breccias, recognized by angular clasts cemented by sparry calcite filling the gaps between them that were opened by the fracturing. In the whole outcrop calcite veins are abundant and they have no regular orientation. Perhaps the best evidence for the fault and its sense of displacement is the orientation of the beds. They dip at 58° toward 345° (north northwest). Normally, the Sciota Limestone dips toward the east or souteast at 10°- 15°.

The change in bedding in this outcrop next to the fault is best explained by a bending of the strata about an axis parallel to the trace of the fault before the rocks failed in a brittle way. However, it is also imaginable that the rocks were bent by a slip-stick mechanism during the movement of the fault. This is possible because faults are

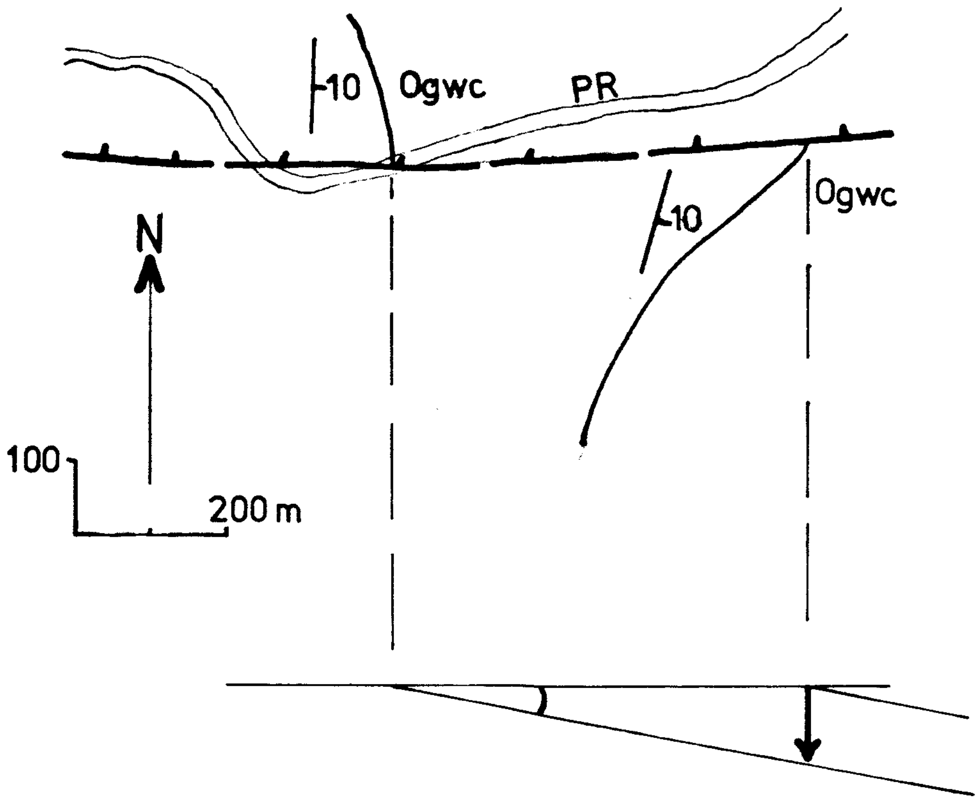


Fig. 4.13: Displacement calculation for the Poultney River Fault from apparent left lateral offset of the basal contact of the Winchell Creek Siltstone (Ogwc). PR = Poultney River.

rough surfaces and move over long periods of time, and the process of faulting is unlikely to be synchronous everywhere and steady through time. Otherwise, there would be no earthquakes.

The displacement on the Warren Hollow Fault can be calculated from the offset of the frontal thrust that forms the eastern edge boundary of the undeformed domain. The calculation assumes that this thrust is parallel to the bedding of the Sciota Limestone that underlies it. With a dip value of 10° and a lateral apparent offset of 950 meters, the vertical displacement comes to 165 meters as a first approximation. Again, the hanging wall is on the northern side of the fault. The same sense of motion is also indicated by the apparent left lateral offset of stratigraphic contacts across the Warren Hollow Fault. This apparent offset is greater than that of the thrust, indicating a westward increasing amount of displacement. Going east along the fault into the deformed zone it was found that it offsets a thrust contact just west of Forbes Hill in an apparently right lateral fashion. Since the thrust contact dips demonstrably eastward, a different sense of motion on the Warren Hollow Fault must be postulated. Thus a pivoting fault might be present with its pole of rotation between the area of apparent left lateral and that of apparent right lateral offset. This pole must be sought in the low zone of shales covered by Quaternary sediments west and southwest of Forbes Hill and no outcrop is present there that could clarify the question.. The change in the sense of the apparent lateral offset along the trace of this fault is very good evidence against it being a strike slip fault. Only a pivoting motion can produce such a change.

A westward increase of displacement on the Warren Hollow Fault is supported by the fact that on its western end it juxtaposes the Whitehall Formation on its northern side to Grenvillian gneisses on its southern side. That means that the movement was at least equivalent to the combined thicknesses of the Potsdam (100 meters), the Ticonderoga (50-80 meters) and part of the Whitehall Formation (30 meters) plus a height difference of 20 meters between the Whitehall on the downthrown side and the Grenvillian gneisses which form an entire mountain on the southern side of the fault. Thus the total vertical displacement there is on the order of 200 meters.

A third northeast-southwest striking fault is the Coggman Creek Fault. It is well expressed in the topography and strikes approximately 045°. Its sense of offset as determined from the stratigraphic boundaries is the same as that of the previous two, i.e., north side down. The amount of offset cannot be determined with confidence, because the outcrop control is very limited. Outcrops in its vicinity on either side are of the same stratigraphic unit, (Great Meadows Formation) which indicates that the amount of vertical offset is probably smaller than on either the Poultney River or the Warren Hollow Faults. The effects of movements on rocks close to it are seen in a series of outcrops about 2.2 kilometers west of West Haven Town Center [10] . Again, the beds dip toward the northeast whereas away from the fault and in the outcrops south of it they dip invariably toward the east or southeast. They also display internal faults and intense calcite veining, in the outcrops next to the fault. Their orientations are, however, not indicative of the movement on the main fault or of its orientation.

Another set of normal faults which strike north-south is recognized in the study area. The most prominent of these is one crosscutting the Warren Hollow Fault about one kilometer east from the road going south from the Town Center to the Poultney River. It is strongly expressed in the topography by a valley that is even narrower than that of the Warren Hollow Fault. The sides of this valley are steep cliffs made up of the Great Meadows Formation with the cross bedded Winchell Creek Siltstone at its base. With the aid of this guide horizon the displacement across the fault could be determined as 45 meters. On the western side of the fault the rocks are relatively down. The fault itself is not seen in outcrop since the bottom of the valley that follows its trace is covered with blocks and debris derived from the cliffs that form the valley shoulders.

Also, there is no sign of deformation in the outcrops on either side of the valley such as bending of beds in connection with or prior to faulting, or presence of calcite veins.

North of the Warren Hollow Fault, this north-south trending fault faults down a lobe of the thrust sheet that lies to the east, which is added proof for its sense of motion. It also constrains the timing of these faults.

It is clear that both sets of normal faults were originated after the thrusts. Their timing with respect to each other cannot be determined since the data is not sufficient to determine their dips and mutual offsets.

Two more normal faults with north-south strike are recognized in the area between the Poultney River and Warren Hollow Faults. Both of them have a west down displacement which could be calculated, in the case of the more easterly fault, as 45 meters. The western north-south fault is deduced from an apparent thickening of the Great Meadows Formation by about 70 meters. This can be accounted for by a fault of similar displacement, which displaces the updip side relatively downward and therefore widens the outcrop of the faulted unit.

4.3 Folds

The existence of folds in the area studied was not well known in previous work. In particular, Fisher (in press) did not mention or show folds in his map. This is understandable because the area where folds are present is the eastern deformed domain, which has been mapped by him as a zone of olistolithic carbonate blocks in black shale. During mapping of this study, folds have been found in the better exposures. Where the outcrops are not as good, especially in the terrane underlain by dolostones, the presence of cleavage and its orientation to bedding demonstrate the presence of folds, even though some may not be mappable in detail.

An angle between bedding and cleavage is present in the whole eastern part of the area studied, indicating that folds are not a local phenomenon but common in the whole Eastern Deformed Zone. Some of the better examples of folding that have been observed are described and illustrated below. It is reasonable to take these examples as representative for the whole of the deformed part of the area studied. A distinctive difference in the style of folding has been noted between various lithologies and therefore the description of folds is separated into lithologic groups.

Folds in Shale and Greywacke

In the Hortonville Shale, folds are not easily recognized because in those lithologies, bedding is often not recognizable and the cleavage whether phacoidal or slaty, is the dominant structure. Where greywacke beds are present, however, folds can be seen.

In all cases these folds are isolated occurrences and seldom have amplitudes bigger than half a meter. They are always bounded by faults that "cut off" their limbs so that only the hinge region is left. The folds seen display a wide variety of shapes. They range from gentle folds to isoclinal and they range from concentric to similar in general geometry. The tight to isoclinal folds are generally approximately similar in shape, whereas the open folds have a more concentric form with a constant bedding thickness. Fig. 4.14 shows the orientation of 23 of those folds in a stereographic projection. The poles to the axial planes are clustered and an average value for the attitude of axial surfaces can be estimated which dips toward 070° at 55°.

The fold axes show a greater scatter, but it can be seen that their distribution fits reasonably well to the trace of the average axial surface. Most of the fold axes trend eastward to southeastward and plunge at moderate to steep angles.

The generally uniform orientation of the axial surfaces is parallel to that of the average orientation of the cleavage. The fold axes lie more or less within that plane yet the fold axes are not coaxial. One explanation for this is non-cylindrical folding. In that case, the hinge line of a given fold instead of being a straight line becomes a

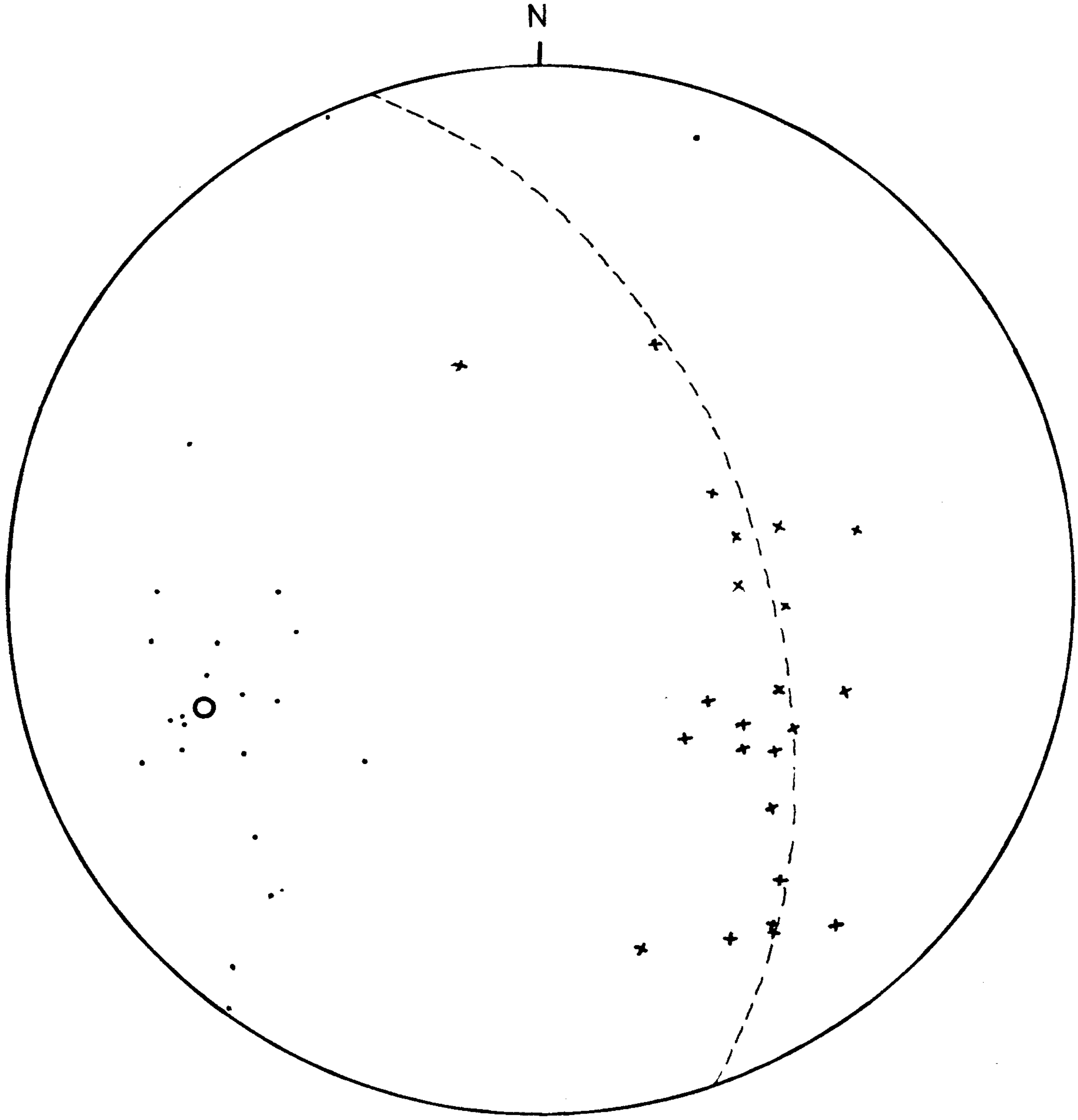


Fig. 4.14: Poles to axial planes (dots) and fold axes (crosses) from folds in the Hortonville Shale. Equal area stereographic projection. Fold axes n=23, axial planes n=22.

a curve. In folds with planar axial surfaces this curve lies on the axial plane. Measurements taken at various points along the curved hinge would then plot in different positions on the great circle representing the axial plane.

In case of a non-planar axial surface the points representing the fold axes will not fall directly on the great circle but scatter around it.

A second explanation implies a second deformational event by simple shear. Pre-existing foliations and lineations are reoriented parallel to the extension direction with increasing simple shear, as long as they are not parallel to the plane of shear at the beginning of deformation (Turner & Weiss, 1963, pp. 98-101; Escher & Watterson, 1974). If the axial plane of folds in the Hortonville Shale was not parallel to the plane of shear and their hinge lines did not lie in that plane than they should have been reoriented to lie within that plane now. It is likely that folds formed by buckling of competent layers in response to subhorizontal shortening prior to thrusting. Their axial planes would then be perpendicular to the shortening direction and at a high angle to the shear plane if the latter is defined by the orientation of the low angle thrust faults. A slight deviation of the hinge lines is enough to have them suffer reorientation towards the shear direction as Escher & Watterson (1974) pointed out.

This is quite likely to have existed before initiation of simple shear, but it could also be introduced by a small heterogeneity in the shear deformation itself. On a stereogram, the fold axes would be distributed on the great circle representing the axial plane. The greater the amount of shear, the more intense the concentration of fold axes around the shear direction becomes. This distinguishes the two step process from the one described before. In the former case of non-cylindrical folding a random distribution of axes orientation would be expected.

The field data shown in Fig. 4.14 display a definite concentration of fold axes around an estimated average trend of 125° and a plunge of 40°. It does support the shear hypothesis and indicates a shear direction parallel to that average orientation, i.e., southeast-northwest. However, the limited amount of data makes this argument weak. Evidence for shear is also found in the outcrop by the presence of isolated fold hinges and limbs. Fig. 4.15 and 4.16 show typical examples of such folds drawn from the outcrop. In no case was it clear what the sense of shear was because one could not find matching sequences along the fault traces. It must be assumed therefore, that the displacement along the faults limiting the folds is bigger than the size of these folds.

It is impossible to give an estimate of the displacement, its amount or direction from my outcrop observations. All that can be said in the light of the presence of innumerable faults at low angles to the cleavage is that there was displacement and its total amount across this zone was probably of significant size.

Folds in Dolostones

Fig. 4.17 is a list of those folds that were actually seen in outcrop. Their locations are given and the numbers in square brackets

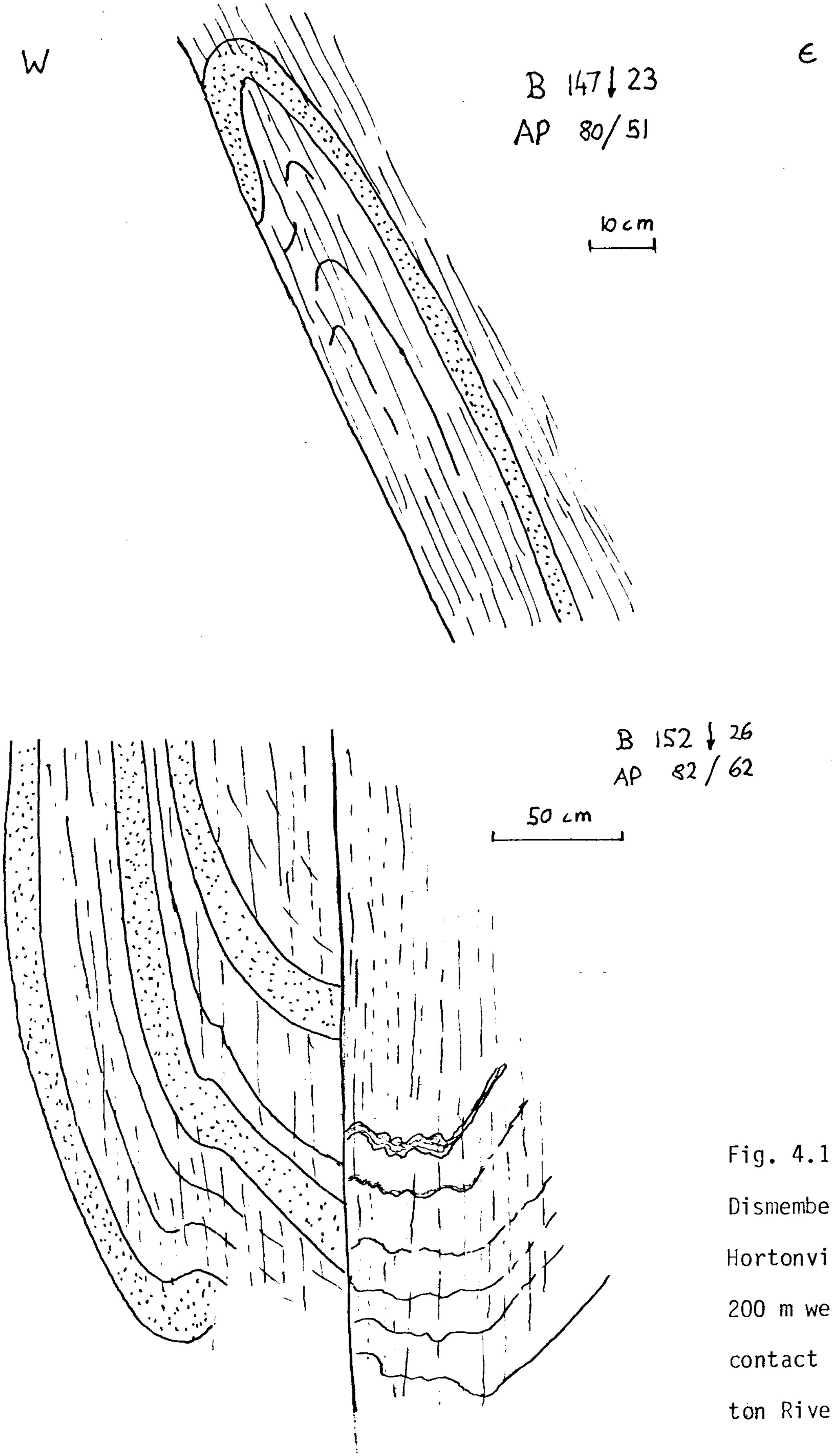


Fig. 4.15:
Dismembered folds in Hortonville (?) Shale. 200 m west of thrust contact north of Hubbardton River [6].

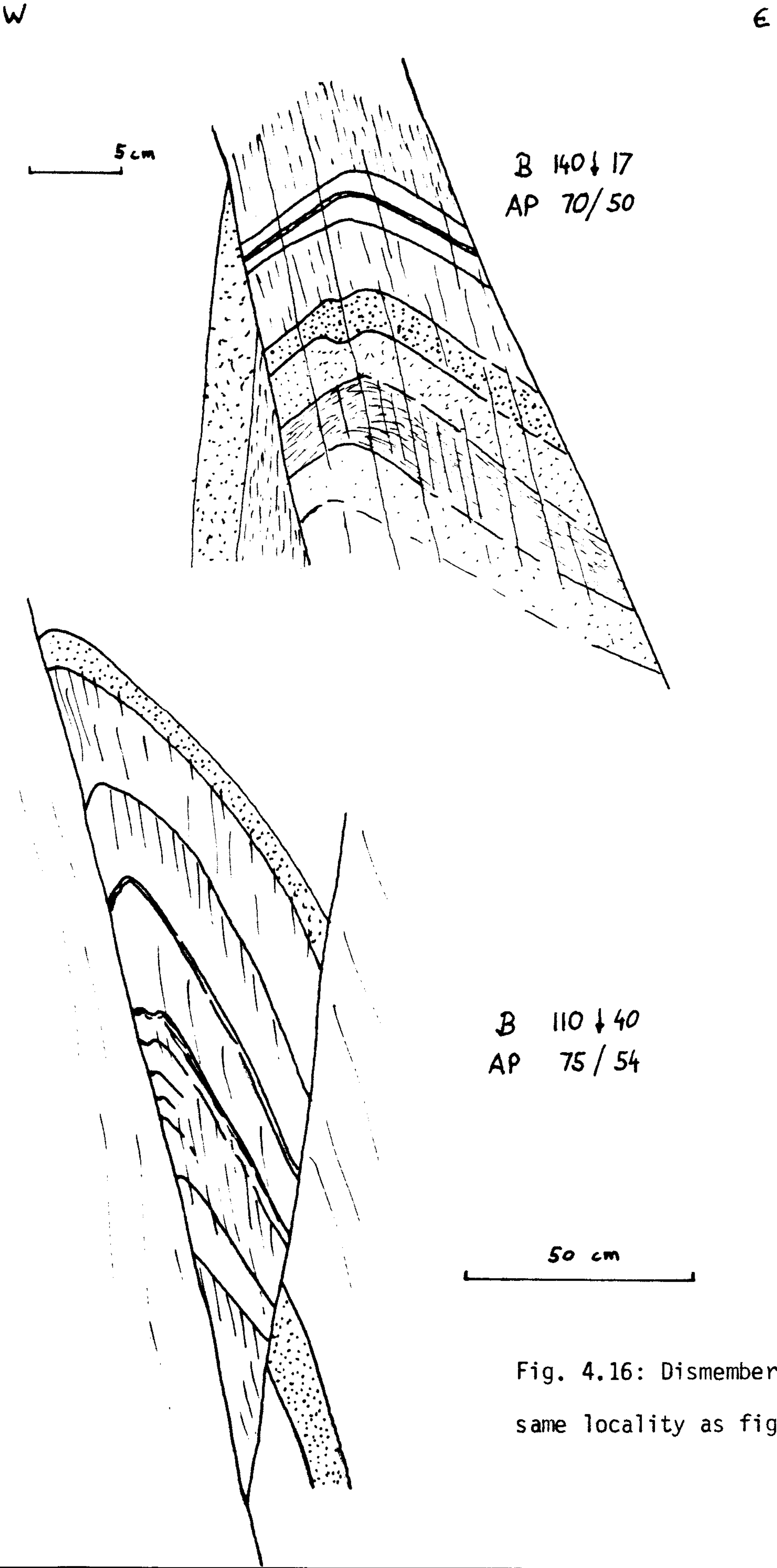


Fig. 4.16: Dismembered folds same locality as fig. 4.15.

#	Location # on Plate I	Subarea on fig. 4.26	Axial plane orientation	Fold axis orientation	Half wavelength (meters)	Amplitude (meters)	Opening angle ψ(degrees)
1	3	III	135/85	55↓12	5	1.5	160
2	8	III	095/21	50↓22	1	-	50
3	11	III	125/47	32↓17	10	6	90
4	11	III	098/50	25↓25	12	6	90
5	11	III	115/62	38↓25	10	10	60
6	12	III	122/47	37↓14	15	7	90
7	12	III	122/60	55↓20	13	-	80
8	12	III	132/75	45↓10	8	(8)	60
9	12	III	125/40	35↓08	10	4	100

Fig. 4.17: Observed folds in dolostones. Axial plane orientation is given as dip direction and dip

refer to outcrops numbered on the geologic map.

The orientation of the axial surfaces is given as dip direction followed by the dip. The half wavelength is an estimate for the distance between inflection points on the limbs of a given fold. The amplitude is an estimate for the distance from the inflection point on one limb to the hinge, taken parallel to the axial surface. The opening angle is the dihedral angle between the limbs of the fold.

It can be seen from the list that the folds are on the order of ten meters wide and about 5 to 10 meters high. Their axial surfaces have a remarkably uniform orientation with an average strike of about 30°. Axial surface dips vary however, between 21° and 85° with the majority of them between 40° and 60°. All measured fold axes trend northeast and plunge gently northeastward. The shape of the folds is generally open to tight with the exception of the one from Carver's Falls which is very gentle. Fig. 4.18 shows a photograph of the most completely exposed of all of these folds. It is found about 500 meters west of the parking lot on Route 22A in the entrance to "Boss Hogg's Quarry" [11]. It shows a synform in slightly calcareous thin bedded dolostone. It can be seen how the thickness of the beds changes as one goes from the eastern (left) limb of the fold to the western limb. The change in thickness is continuous at the hinge and visible in all beds. This shows that the eastern limb has thickened relative to the western limb through folding. Individual beds have exactly double thickness on the eastern, west dipping limb as compared to the western, east dipping limb. The axial planar cleavage almost parallels the thin limb of the fold.

Figure 4.18: Syncline in dolostone at entrance to Boss Hogg's Quarry [11], looking southeast. Fold #3 in fig. 4.17.

This orientation of the axial surface and the different bed thickness make the fold asymmetric, which is a typical characteristic of all the folds seen in dolostones. No indicators of inverted bedding are visible in the synform shown in Fig. 4.18 and taking the open nature of folding into account, it seems most likely that the fold is facing upwards or in other words a synformal syncline. Fig. 4.19 shows an antiform just opposite to the synform discussed above. The style of this antiform, which shares its eastern limb (right) with the western limb of the synform of Fig. 4.18, is markedly different from the former. Whereas the syncline had a clearly and continuously folded hinge this is only so for the core of the anticline. The outermost beds show very little folding but instead a break in the hinge region and an abrupt turn of these beds from a position parallel to the spaced cleavage to one almost perpendicular to it.

Again, the west dipping limb shows thicker beds than the east dipping limb. This antiform with a broken hinge is typical of most of the antiforms observed in the dolostones. In the case illustrated in Fig. 4.19 the core is made up of calcareous dolomite whereas the outer bed consists of purer dolostone. It is seen that the difference in composition corresponds to a different behaviour in the process of folding. The calcareous core is folded continuously whereas the dolomitic outer bed deforms in part by fracturing. The shape of the exposed core of the fold is similar (although inverted) to that of the adjacent syncline. The fracture in the outer dolostone layer parallels the (rather weakly developed) cleavage and is approximately parallel to the axial surface of the fold.

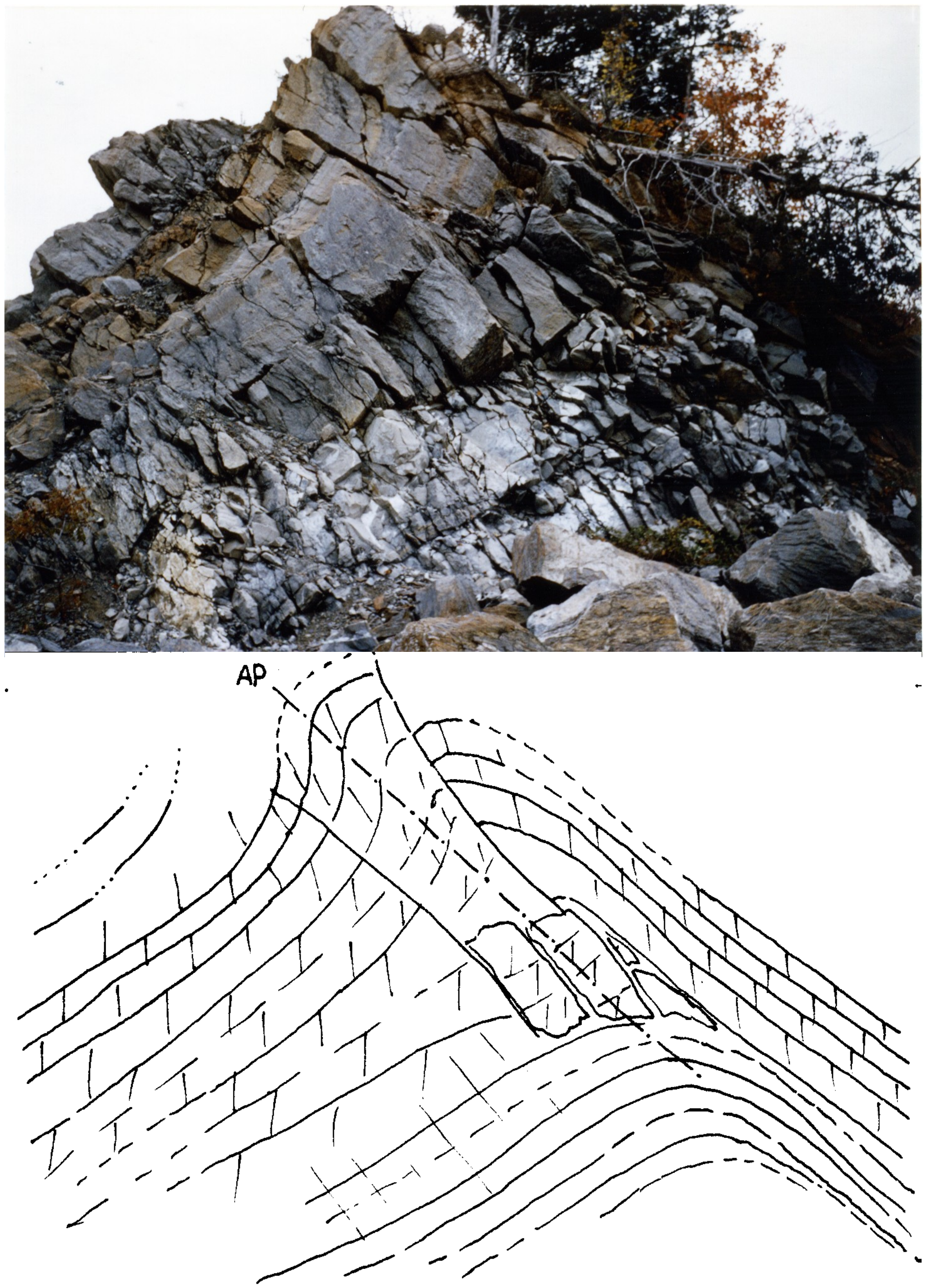


Figure 4.19: Broken anticline in dolostone at entrance to "Boss Hogg's Quarry" [11], looking northeast. Fold #4 in fig. 4.17.

I believe that the difference in the deformational style between dolostone and calcareous dolsotone represents a difference in strength. This has been demonstrated in experiments by Tullis (1980) who found that independent of pore fluids, temperature or pressure, critical resolved shear stresses in dolomite are one order of magnitude higher than those of calcite. His data suggest minimum values of 100 bar for calcite and about 1 kilobar for dolomite. They refer to the occurrence of twinning in weakly deformed samples under uniaxial compression and extension and are therefore not fully applicable to field situations. However, they give support to the hypothesis that dolomite may be significantly less ductile than limestone.

Folds in Limestones

Fig. 4.20 is a list of the folds actually seen in the field area, organized in the same manner as the one given for the dolostones. It records four folds from the southern part of the Forbes Hill thrust sheet and 10 folds from the Carver's Falls Sheet in the southeastern part of the map area. It can be seen that the first group of folds has an almost uniform orientation. The axial surfaces dip toward 061° at 56° on average. The average trend of the fold axes is 136° and the plunge 17°.

For the ten folds from the Carver's Falls Sheet the average dip direction of the axial surface is 111° and the average dip 56°. Fold axes trend 43° and plunge 14° on average.

It is clear that within each thrust sheet the orientations of the folds are consistent but different thrust sheets have differently oriented folds. In both groups the trend of the fold axes roughly

#	Location # on Plate I	Subarea on fig. 4.26	Axial plane orientation	Fold axis orientation	Half wavelength (meters)	Amplitude (meters)	Opening angle (degrees)
1	3	III	128/56	40!18	1	1.5	-60
2	3	III	130/70	40!20	(1)	-	-
3	3	III	118/71	36!(0)	(1)	(2)	60
4	3	III	?	35!o8	-	-	-
5	6	II	050/45	133!30	1.0	1.6	20
6	6	II	071/62	129!25	0.3	1	16
7	15	II	059/57	137!20	1.5	2	60
8	15	II	065/62	150!05	2 - 5	(2)	90
9	18	III	127/21	47!32	1.4	1	30
10	18	III	082/37	175!(0)	1	2	30
11	18	III	128/62	63!30	10	1	140
12	19	III	087/63	18o!10	1	1.5	15
13	19	III	090/90	180!(0)	1	0.2	140
14	20	III	110/30	50!20	1.2	-	(30)

Fig. 4.20 Folds observed inlimestones. Axial planes as dip direction and dip

parallels the nearby trace of the thrust limiting the sheet on its western edge.

All axial surfaces measured are dipping toward the east or southeast and so is the cleavage wherever it can be observed.

The folds in the limestones are generally smaller than those in dolostones; their width varies around one meter, with the exception of one gentle fold which is 20 meters wide. The height of the folds is on the same order of magnitude (about one to two meters).

Fig. 4.21 shows a characteristic example of a fold pair in limestone at Carver's Falls [3] . The orientation of the outcrop surface at a low angle to the fold axis tends to distort the picture of the folds by "stretching" its profile parallel to the axial surface trace. The picture has therefore been taken from a view along the fold axis to obtain a more accurate view of the cross-sectional shape of the fold. Standing on the outcrop the folds appear to be isoclinal, but the picture with its down plunge view shows that they are not; they are tight folds with an opening angle of approximately 60°.

The bedding in the example shown is marked by clay seams, as is typical for the Middlebury Limestone that contains this fold. The syncline-anticline pair shows a marked asymmetry with the short, west dipping limb between them connected to long, east dipping limbs. This is the same asymmetry as in the fold pair in dolostone described earlier. The picture shows also that the beds below the fold pair are unfolded and those above it decrease in fold intensity away from the axial surface. It is likely that the fold is not continuous but that it dies out down section along its axial surface as well as along its axis. If that is the case, the fold is only of local extent and

Figure 4.21: Fold pair in Middlebury Limestone. Note that beds below fold are unfolded. Fold #1 in fig. 4.20. Carver's Falls.

must be non-cylindrical. The shape of this fold pair is approximately similar, marked by a thickening of beds in the hinge region. Fig. 4.22 is a sketch which tries to explain one mode of formation of such folds. It requires a certain amount of bedding parallel slip as well as thickening perpendicular to bedding to account for the observed geometry. The form of the folds in such a model is largely dependent on the amount of slip between individual beds: the more displacement occurs along a bedding plane, the bigger the amplitude of the fold will be. This model could in the case of the fold discussed account for an east over west shearing displacement acting on the rocks while they were folding. Local variations in the shear strains would produce a change in the amplitude of folding. The folds would then represent inhomogeneities of a simple shear deformation imposed on the thrust sheet as a whole.

Folds like the one at Carver's Falls are best recognized in outcrops where clay seams mark bedding. In Orwell lithologies where the bedding is not as prominent to begin with and not traced by clay seams, it becomes quite difficult to recognize folds, as well as to distinguish bedding from cleavage. Fig. 4.23 shows a case where both cleavage and bedding can be seen. Bedding can be distinguished by its wider and more irregular spacing. Bedding surfaces are not as planar as those produced by cleavage. Both are essentially fracture surfaces and therefore not always easily distinguished. In such a case the information that still can be obtained is the orientation of the intersection line between bedding and cleavage. If one assumes that folding is cylindrical and the cleavage is parallel to the axial plane of the folds then this intersection must be coaxial with the fold axis.

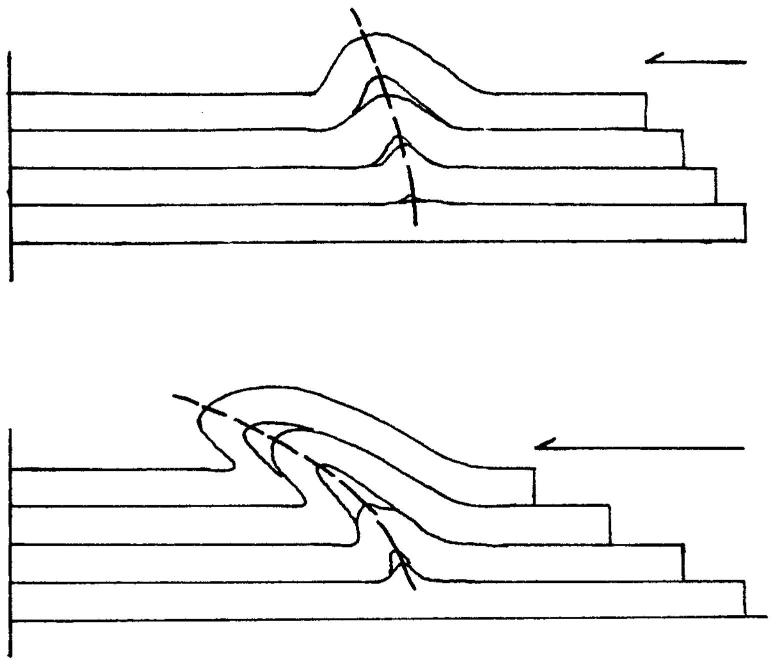


Fig. 4.22: Possible mode of formation of the foldpair of fig. 4.21. Model requires bedding parallel slip and flow of material towards the hinge region.

Figure 4.23: Bedding and cleavage at high angles to each other. Hammer handle parallels bedding. View is northward. From Forbes Hill, approximately 300 m north of summit.

Considering the measuring accuracy and the limited outcrop information it is useful to make those assumptions, and at least obtain a first approximation for the orientation of structural elements in the area studied.

Fold Orientations in Carbonates

To obtain information about the orientation of the folds in areas where they are present but not visible in outcrop, the orientation of cleavage-bedding intersection has been recorded wherever possible. The results are presented in Fig. 4.24 and 4.25. The orientation data has been plotted for four different subareas which are shown in Fig. 4.26. Fig. 4.24 presents measurements obtained from the northern part of the Forbes Hill Thrust Sheet (points) and the southern part of the same sheet (crosses). Fig. 4.25 is collected from measurements of the Carver's Falls Thrust Sheet (points) and the West Haven Thrust Sheet (crosses).

Most measurements stem from limestones where both bedding and cleavage are much better developed than in dolostone lithologies. This is the reason why data from the West Haven Sheet are so limited, because its outcrops consist almost exclusively of dolostone, as can be seen from the map.

Within each subarea the orientations of the bedding-cleavage intersection show a remarkable uniformity. For the northern Forbes Hill the average trend is 052° and the average plunge about 10°. The one measurement that does not fit the general picture (175° plunging 45°) was obtained in immediate vicinity of the frontal thrust and may reflect a local reorientation caused by the thrusting.

The measurements from the southern part of Forbes Hill average

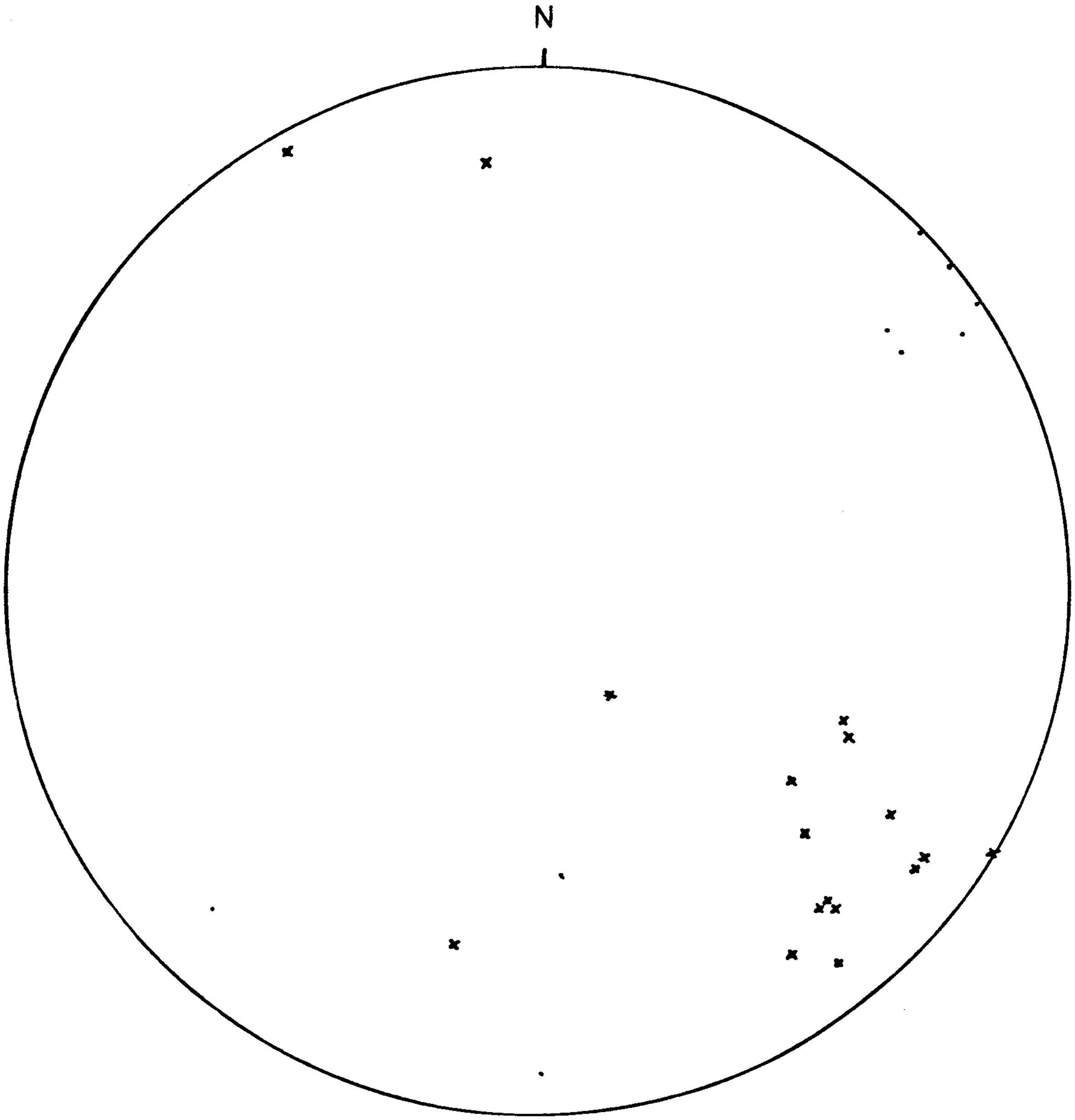


Fig. 4.24 Equal area stereographic projection of bedding-cleavage intersections. Dots: Measurements from northern Forbes Hill (subarea I), crosses: measurements from the southern Forbes Hill Sheet (subarea II). Subarea I n=9, subarea II n=11.

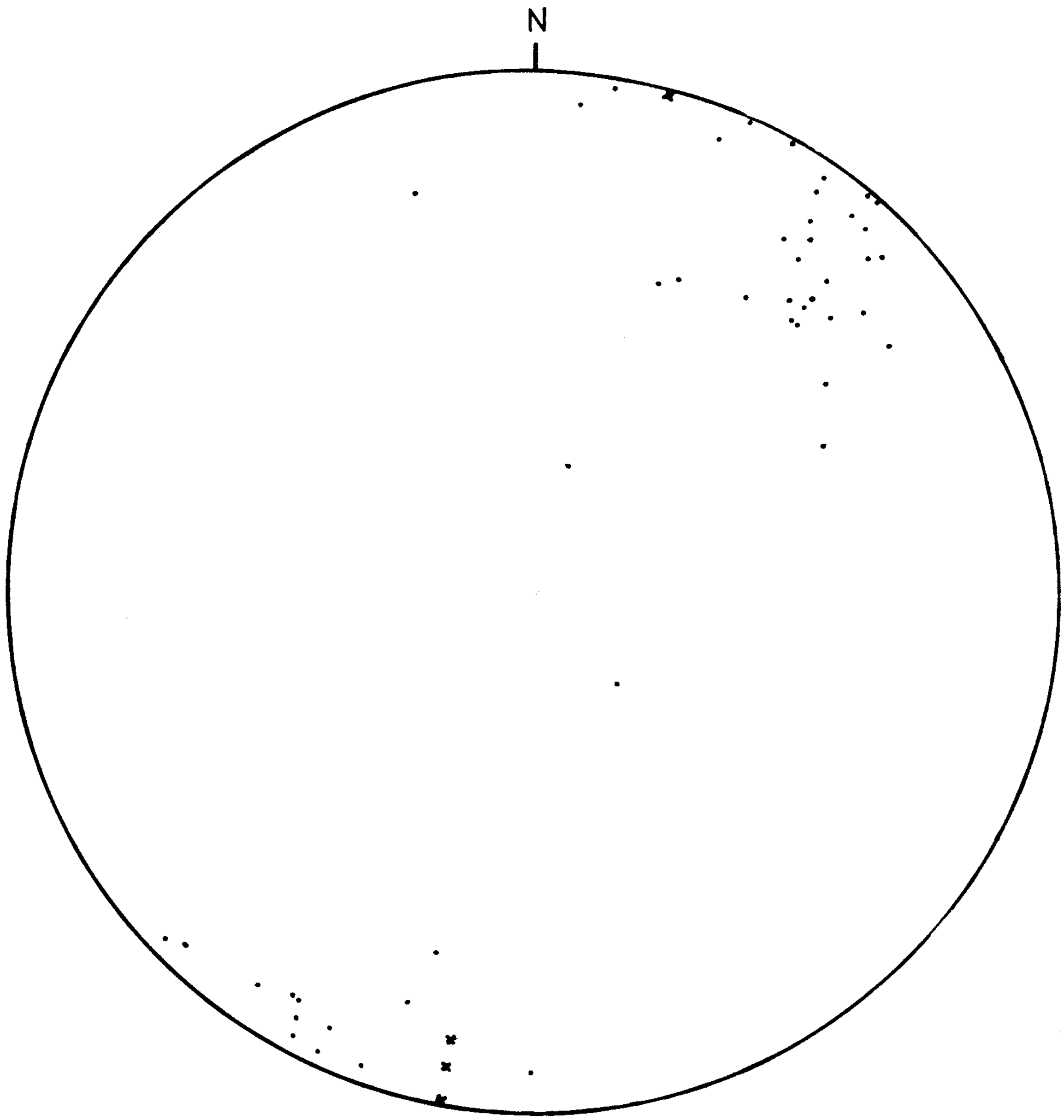


Fig. 4.25: Equal area stereographic projection of bedding-cleavage intersections. Dots: Carver's Falls Sheet (subarea III) n=47; crosses: Frontal Sheet (subarea IV) n= 4.

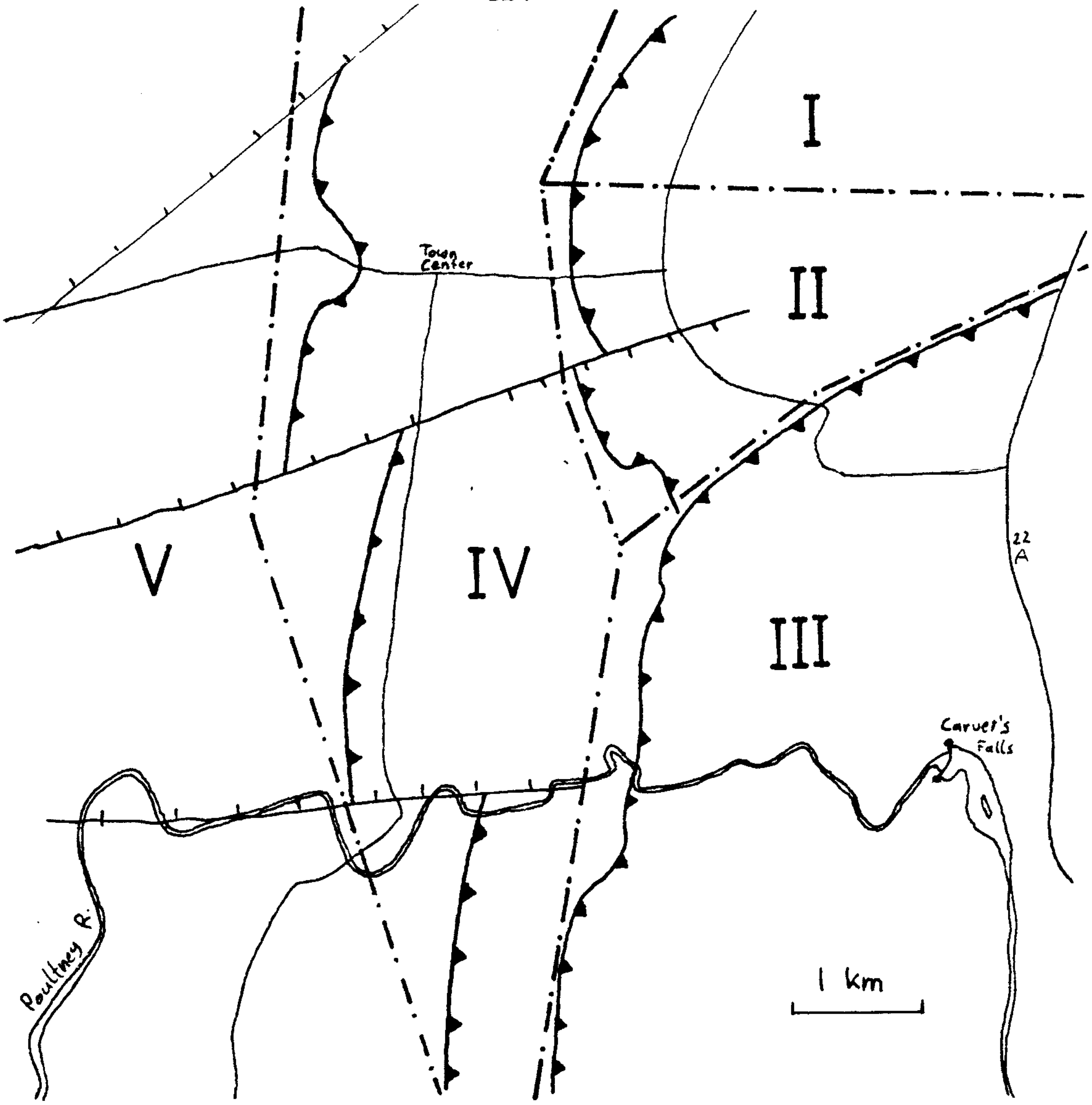


Fig. 4.26: Subareas of the Western Deformed Zone.

-.-.- = Subarea boundaries

I = Northern Forbes Hill Thrust Sheet

II = Southern Forbes Hill Thrust Sheet

III= Carver's Falls Thrust Sheet

IV = West Haven Thrust Sheet

V = Western Undeformed Zone

to a trend of 127° and a plunge of 24°. This conforms with the mean value for the four fold axes measured in the better outcrops along the road across Forbes Hill. This value was a trend of 136° and a plunge of 17°.

For the Carver's Falls Sheet (southeastern subarea) the estimated average is 042° trend and 17° plunge. It will be noted that 13 of the 47 intersections measured show a plunge toward the southwest. 11 of those 13 measurements come from outcrops along the Poultney River or further south of there. The vast majority of the measurements in the northern part of that sheet plunge to the northeast.

If it is correct to consider the bedding-cleavage intersection as approximately coaxial with the fold axis, one can conclude that the structures in the southern part plunge to the south whereas they plunge to the north in the northern part of the Carver's Falls Sheet.

This is consistent with the map pattern south of the Poultney River. There the Middlebury Limestone widens its outcrop considerably at the expense of the underlying Providence Island. This is the effect of a gently southwest plunging anticline in the form surface of the contact between the two units.

Only four measurements could be obtained from the West Haven Sheet, so that information about fold orientation within it is limited. (Fig. 4.25 cross symbols.) However, they indicate a trend in axial direction of 190° and a probable plunge of no more than 10° in that direction.

The generally consistent orientation of the bedding-cleavage intersection lines in each subarea is a good argument for regular deformation. It can be concluded that in a statistical sense the folding can be considered cylindrical. In individual examples, however, the folding may well by non-cylindrical and this is expressed in the scatter of measurements on the stereoplots.

Another observation that emerges from the orientation data when compared with the map is that the hinge lines and bedding-cleavage intersections closely parallel the strike of the thrust faults that terminate each sheet on its western side. Three explanations for this phenomenon are possible:

1. The basal thrusts acted as guides which directed the orientation of folding parallel to their strike. This requires some sort of stick-slip mechanism, that caused the actual bending of the beds around axes parallel to the fault trace. This involves faulting prior to folding. A problem inherent in the idea is that a thrust movement acutally releases stresses by allowing the rock pile above it to move as a whole. It is therefore unlikely that the stresses would have to be taken up internally by folding after the faults have formed. Also folds created by stick-slip would only be present immediately adjacent to the thrusts. In the field area however, folds are found throughout the entire area of the thrust sheets and not only close to the thrusts. Therefore, such a mechanism seems unlikely as mode of formation of the majority of the observed folds.

2. Folds may have formed as ramp folds that is by the bending of the thrust sheet as a whole above bends in the thrust surface, especially steps in the décollement horizon (Suppe & Namson, 1979). The principle is shown in Fig. 4.27. Such a model produces one relatively open large anticline, with axes parallel to the strike of the thrust and only one syncline above the line where the under-

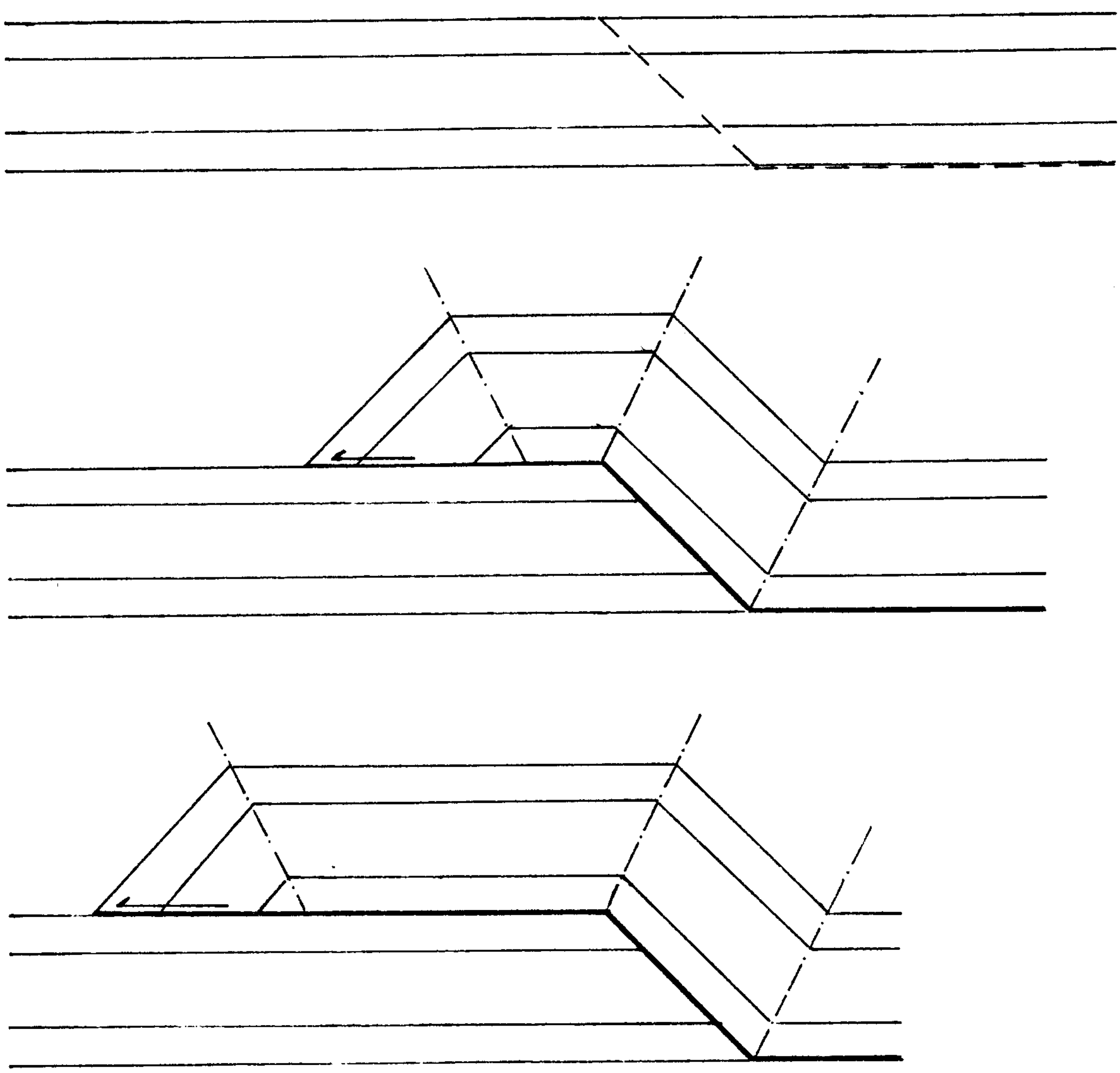


Fig. 4.27: Ramp fold development, schematic. Deformation by interbed slippage. After Suppe and Namson (1979).

lying thrust begins to step up to a higher stratigraphic horizon. As the fault sheet continues to move the width of the anticline becomes wider and the amplitude stays equal. The beds between the front of the thrust sheet and the last bend of the basal thrust are undeformed in this model. This cannot be combined with field observations which show that the thrust sheets are consistently folded throughout and that on a much smaller scale than could be expected from a ramp folding process. It is however, possible that on a large scale folds created by movement of the thrust sheet over a ramp are present. This cannot be demonstrated from field evidence, however.

3. Shortening in the limestones may have resulted in buckling and thus generation of folds. As these folds became tighter they created an anisotropy in the strength of the rock e.g., by the formation of an axial planar cleavage. With continued deformation the rocks yielded in a brittle manner and thrusts formed in an orientation guided by this anisotropy.

In this latter case, the folding would have initiated before thrusting. Field observations support this. In various localities folds have been seen to be truncated by thrust faults (e.g., Fig. 4.11, 4.15, 4.16 and 4.19). Folding may also have continued during movement of thrust sheets, possibly caused by shear between the overlying and underlying thrusts.

The orientation of bedding-cleavage intersections in the Forbes Hill Thrust Sheet shows a swing of about 75° in the trend. In the northern part the trend is 052° and in the southern part it gets to a value of 127°. This change is paralleled by the strike-controlled topography as well as the trace of the underlying thrust.

A good explanation for this is a deformation of the whole thrust sheet after its emplacement. The present configuration could have been produced by folding of the whole sheet about an axis of easterly trend and plunge to produce a synform plunging to the east of the size of that whole sheet.

Alternatively the late deformation of the Forbes Hill Sheet might be related to movement on the northeast trending normal faults. The idea is that movement on the fault planes may reorient the structures at high angles to it. On the Coggmann Creek Fault with its north side down offset, the reorientation of the strike should be in a counterclockwise sense, but the strikes swing around clockwise as one approaches the fault. On the Warren Hollow Fault where the footwall is on the southern side, the reorientation of structures by drag or stick-slip should be clockwise. The structures deviate from their strike in the middle of the sheet in a counterclockwise sense. Therefore, late faulting cannot be the direct cause for the reorientation of structures seen on Forbes Hill.

4.5 Structural Synthesis

In this section an attempt will be made to summarize the structural elements presented above and establish time and space relationships between them. The aim is to give an account of the structural history of the area as far as it could be determined and discuss the geometry of the major structural elements.

Except for the part on normal faults all that is said below refers to the eastern deformed part of the field area. In the western half, apart from faults, signs of deformation are virtually absent. The rocks in that part are not folded and dip consistently at approximately 10° to 15° toward the east.

The eastern zone however, knows no such regularity. Instead, it is underlain by large thrust sheets consisting of dolostones, limestones and shales in that structural and stratigraphic order. Three main thrust sheets have been identified:

1. The West Haven Sheet
2. The Forbes Hill Sheet
3. The Carver's Falls Sheet

The West Haven Sheet lies directly above the undeformed stratigraphic succession and its basal thrust marks the boundary of the Eastern Deformed Zone. It is sometimes underlain by a tectonic mélange consisting of strongly deformed black shale, but elsewhere the contact is without such a "glide horizon". The basal thrust of the West Haven Sheet is not very well defined, and had to be inferred in areas where the lack of outcrop did not allow it to be closely localized. Its dip is assumed to parallel that of the underlying Sciota Limestone. This is probable because the Sciota Limestone may act as a décollement horizon for thrusts, and the underlying dolostones are more rigid, so that a bedding parallel thrust would most likely develop in that comparatively weaker unit.

It is a well known phenomenon that thrust faults follow bedding surfaces in incompetent units and climb upsection in more competent beds. (Ramsay et. al., 1983; Boyer & Elliot, 1982). I believe that this is true for the West Haven Thrust in my field area. However, this is only an assumption made in analogy to observed geometries elsewhere. The structurally upper boundary of that sheet is given by the thrust that place the two eastern thrust sheets over it.

The West Haven Sheet is divided by the Warren Hollow Fault into a northern and a southern part. These two parts show a different lithologic succession when one traverses them from west to east. While in the north the sequence consists of dolostone then limestones and then shale, the limestone is missing in the southern part (See map Plate I).

The missing limestone poses a geometrical problem. One possibility is that it is hidden in the zone of no outcrop that has been shown as shale on the map. But this is unlikely for topographic reasons. There is a pronounced steepening where the outcrop begins and that is most likely caused by a change in lithology from soft shales to hard carbonates.

It is inferred that the medial Ordovician limestones have not been deposited on top of the dolostones, because epigenetic dolomitization is not known to have occurred anywhere else in Chazyan and higher medial Ordovician limestones (Kidd, personal communication).

Seismic profiling by the oil industry north and south of the field area has indicated a horst and graben morphology of synsubsidence origin below the Taconic Allochthon (Kidd, personal communication). If this is true in the field area, the local absence of the mid Ordovician strata above the Beekmantown group might be explained. The presence of karst features on top of the Providence Island also supports the idea of local non-deposition of Chazyan Limestones.

A third way to account for the missing limestone is a thrust contact between dolostones and shales, such that the dolostones are overlying the shales. The West Haven Sheet would then be a klippe with respect to the shales below. The only indicator for that is the steepening of topography with the start of carbonate outcrops on its eastern side. On the other hand it is not nearly as steep as in places where thrust contacts between dolostones and shales are observed. Therefore, a thrust contact on the eastern side of West Haven Sheet seems unlikely.

The Forbes Hill Sheet

The Forbes Hill Sheet is situated structurally above the West Haven Sheet and below the Carver's Falls Sheet. Its basal thrust places dolostones on top of shale and is well defined by both topography and structures such as mylonitic tectures in carbonates. It is characterized by a pronounced arcuate pattern of both structures and lithologies that must be caused by refolding after thrusting (see Chapter 4.3 Orientation of Folds). It also shows that the basal thrust is not always in the dolostone unit but cuts upsection into the limestones as can be seen at its southern end, north of the Hubbardton River in an outcrop described earlier [6].

The Forbes Hill Sheet probably has some internal thrusts as is indicated by the presence of some tens of meters of shales in between limestones on Forbes Hill. In the part of the Forbes Hill Sheet that is occupied by shale, the type locality of the Forbes Hill Conglomerate [14] is situated; this unit contains pebbles and lumps of Taconic lithologies. It has been explained by Zen (1967) as having formed through the shedding of olistoliths off the scarp of an advancing thrust fault nearby to the east. In one outcrop along the Hubbardton River farther northeast [17] Taconic lithologies may be present. It is not impossible that there is indeed another thrust present where Taconic rocks are placed over Hortonville Shale. Zen (1967) has put the Taconic Frontal Thrust through there and connects it with the Sunset Lake slice situated to the north of the field area. Since there is only one outcrop and the lithology is not proven to be of Taconic origin, Zen's interpretation has not been followed on the map.

Following the trace of the thrusts one recognizes that the structural thickness of the carbonate part of the Forbes Hill Sheet decreases away from its center. That means that the form of this sheet is changing along strike and that it may and probably does terminate laterally within a few kilometers. The geometry of this sheet is that of a lens bounded by its basal thrust on the bottom and the Taconic Frontal Thrust on its top in the north and the basal thrust of the Carver's Falls Sheet to the south. It is therefore most probably what has been termed a horse by Boyer & Elliot (1982).

The Carver's Falls Sheet

This is the biggest thrust sheet recognized in the area and occupies the whole southeastern part of the field area. It is entirely overlain by the Taconic Allochthon and in turn lies on top of the Forbes Hill Sheet and the West Haven Sheet in its northern and southern part, respectively. It actually consists of at least two thrust sheets the lower of which does not contain shales in its upper portion.

It is characterized by a consistent trend of fold axes and bedding-cleavage intersections, and a change in the plunge of those structures; in the north they plunge northeastward, in the south they plunge southeastward. This could indicate a refolding of this sheet similar to that of Forbes Hill. In this case the late fold would be an antiform, while Forbes Hill has a superimposed synform. The primary structure of the Carver's Falls Sheet is a big northeast-striking anticline which is truncated by the thrust that separates the upper from the lower part of the sheet.

The upper boundary of the Carver's Falls Sheet in its northern part does not consist of a single thrust separating it from the Taconic rocks above but, at least in one outcrop [5] on Route 22A, several thrusts. Plate IV, scaled drawing of this outcrop, shows five repetitions of a sequence consisting of Orwell-type massive limestones at the base, thin bedded limestones in the middle with upward increasing shale interbeds (Glens Falls Limestone) and finally a black shale above. Outcrops following further south on Route 22A which are structurally above the one shown in Plate IV continue to repeat this sequence. This repetition is best explained as a stack of duplexes developed between the two major thrust sheets. They are much smaller than the three big thrust sheets (only ten to a hundred meters as an estimate) and may indicate the last movements of the Taconic thrust before the bulk of thrust deformation was transferred below the Carver's Falls Sheet.

Relation Between Folding and Thrusting

In various outcrops in the field area quite large folds are seen. They are, however, restricted to the area of the thrust sheets. Thus folding and thrusting are apparently connected. One important question that was to be answered in this study is that of the timing of different deformational events.

In several outcrops in the field area the relationship between folds and thrust faults could be studied. One example is an outcrop about 500 meters east on the road from Sciota School [13] . There Providence Island Dolostone overlies black shale in a thrust contact. The Providence Island is openly folded with bedding dips from 17° to 40° E. The contact however, is almost planar and dips at a maximum of 15° to the east. The folds there are clearly truncated by the thrust surface and therefore they predate the thrusting. The same relationship can be observed on the frontal thrust of the West Haven Sheet in a series of outcrops [14] about 1 kilometer west-southwest from the West Haven Town Center. Again dolomites are seen folded above an almost planar contact and the axial planes of the folds and the thrust plane crosscut each other.

A third example is seen on Plate IV. At meter 35 an open fold in thin bedded limestone is truncated by the thrust and the black shales below.

Fig. 4.4 also shows how folds and their axial planes are crosscut by the thrust fault below them. In the internally deformed shales there is abundant evidence for folding prior to faulting. The most striking is the presence of isolated fold hinges that are limited by faults (Figs. 4.15 and 4.16). In addition, the fact that cleavage is almost everywhere crosscut by faults is a very clear argument that folding which produced the cleavage occurred before the internal thrusting of the shales. It is difficult to prove that the internal faulting in the shales and the formation of new thrusts were, after all, contemporaneous events. Both the faults and the internal faulting in the shales and greywackes have about the same attitude (see Fig. 4.2 and 4.12). This indicates that they were produced by similarly oriented stresses. The thrusts have the same general orientation as the Frontal Taconic Thrust that borders the field area to the east. On the basis of this similar attitude it is concluded that they are expressions of the same tectonic event i.e., the emplacement of the Taconic Allochthon. The folding in both shales and carbonates is then an event prior to the thrusting and it was probably the first step in the deformational sequence.

All folds observed in carbonates show a strong asymmetry expressed in a thickening of the west dipping limbs and a relative abundance of the east dipping bedding attitudes, where folds are not seen in their entirety. The asymmetry expresses itself also in a vergence of the folds toward the west. In one model of folds formation, it is assumed that the folds initially formed were symmetrical and produced by an east-west shortening in front of the advancing Taconic Thrust Sheet. When the stresses became higher or when the folds had become tight and cleavage had developed, the rocks failed and thrusts formed in the carbonates of the shelf sequence and the overlying shales.

With continued movement of the Taconic Sheet over the carbonates the latter ones were subject to simple shear. This may have caused a change in the geometry of the pre-existing folds to their present asymmetric shape.

It is equally possible, however, that the asymmetry of the folds was achieved prior to thrusting. In that case, the folding would have been a response to shearing stresses acting before the thrusts formed. With increasing deformation, the folds would become tighter and their axial planar cleavage more pronounced. Finally the folds would break on surfaces close to the cleavage planes and thrusts would form. Observations in the field support this model. In various outcrops in the carbonates, especially in the dolostones, folds have been seen cut in half by thrust faults (e.g., Fig. 4.13). However, the main thrusts cannot be entirely parallel to the axial planes of the folds, because their dip is much shallower than that of the axial planes (See cross-sections; Plate II).

Timing of Thrusting

The three thrust sheets recognized in the field area are arranged as a hindward dipping imbricate stack that is, each sheet is structurally overlain by its eastern neighbour (See Plate II). In the classical view on fold and thrust belt geometry (Dahlstrom, 1970; Boyer & Elliot, 1980) this is interpreted as having developed toward the foreland. In that process, slices are added to the hanging wall of the thrust at its leading edge and thus the thrust sheet becomes bigger as it moves. The most frontal thrusts are younger than the ones behind, and displacement is confined to the structurally lowest and most forward thrust.

In each of the imbricate sheets, a deformational sequence is likely to develop as its basal thrust develops (Fig. 4.28). Folds will be produced by shortening and shearing. As the thrust joins the roof thrust again the enclosed duplex moves as a whole and may be folded by ramps it traverses. Then a new thrust will start to form forward

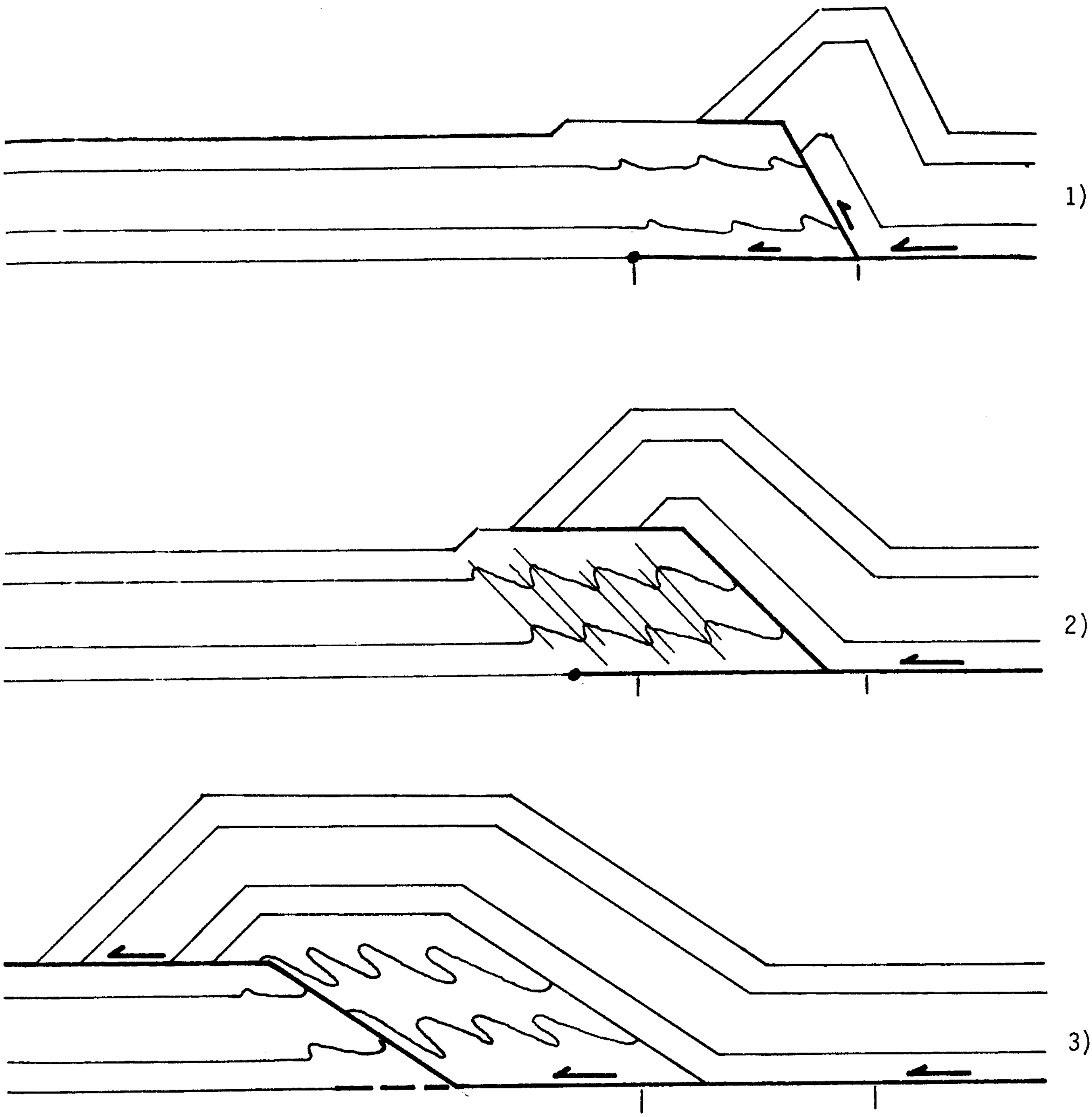


Fig. 4.28: Model explaining a possible deformational sequence in each of the thrust sheets of the field area.

1) Shortening and initiation of folds above propagating new thrust
2) Overturning and tightening of folds.
3) Basal thrust joins upper thrust by "stepping up". Duplex is now transported as a whole with overlying thrust sheet.

of the last formed duplex, again accompanied by folding until the fault has rejoined the roof thrust.

It is seen that the sequence of deformational events in each thrust sheet is the same, but the onset of each sequence is not simultaneous, even though they are part of one process. In the field the West Haven Sheet would then be the last to have formed preceded by the Forbes Hill Sheet, the Carver's Falls Sheet and the Taconic Thrust Sheet. One would imagine that while the basal thrust of the Carver's Falls Sheet was active, the future Forbes Hill thrust was propagating, accompanied by folding in the overlying units. The lower décollement horizon would be the Sciota Limestone, and the upper one the Hortonville Shale so that the imbricate rocks in between comprise the Providence Island Dolostone and, where present, the middle Ordovician limestones.

In general, this model is consistent with field observations. It can be seen from the map (Plate I) that within each thrust sheet the stratigraphic sequence is preserved i.e., Beekmantown Dolostones form the base of each sheet overlain by medial Ordovician limestones and shales.

In various locations however, this order is not present. One example is found along the road traversing the Forbes Hill Thrust Sheet, where a second limestone sliver is recognized that is separated from the main body by a zone of Hortonville Shale [15] . This requires that the thrust underlying the second sliver be younger than the basal thrust of the Forbes Hill Sheet (Fig. 4.29). That means that the thrust sequence of this particular location is younging backwards (to the east) and possibly that the younger thrust crosscuts the basal thrust of the Forbes Hill

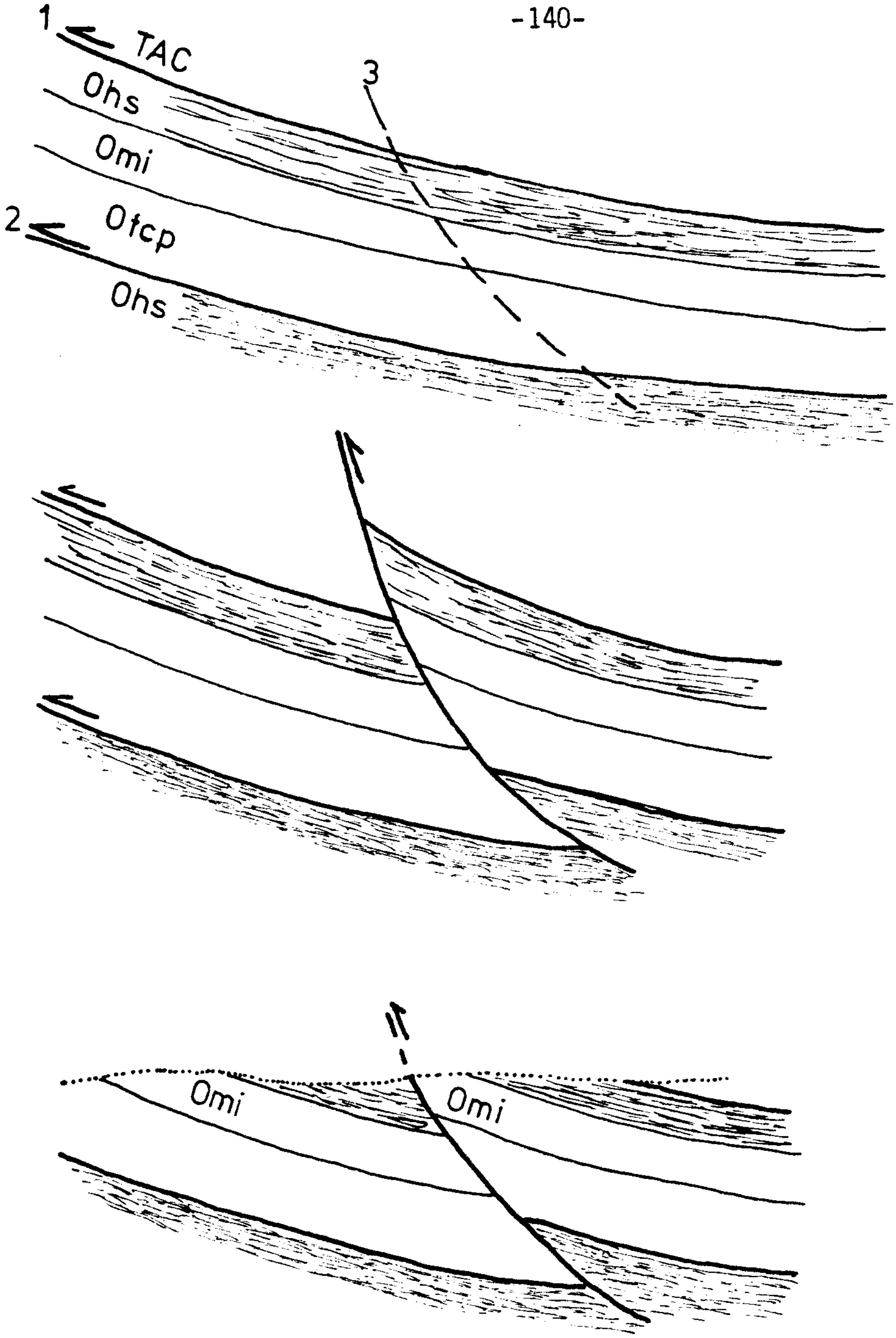


Fig. 4.29: Late thrusting explaining the omission of Providence Island Dolostone on Forbes Hill. 1 - Taconic Basal Thrust, 2 - Forbes Hill Sheet Basal Thrust, 3 - Forbes Hill late thrust.
TAC - Taconic Allochthon, Ohs - Hortonville Shale, Omi - Middlebury Limestone, Ofcp - Providence Island Dolostone.

Sheet.

A second example could be present at the thrust contact above the Hubbardton River [6], if as mentioned in the stratigraphic description the underlying shales and greywackes are Taconic lithologies. Then this thrust, which is the frontal thrust of the Forbes Hill Sheet would cut across the Basal Thrust of the Taconic Sheet and bring that older thrust in a position forward (west) of the Forbes Hill Thrust. This would put an older thrust in front (west) of a younger one.

A third example is situated at the southern edge of the field area [16]. There, a sliver of Bomoseen Greywacke, (the basal unit of the Taconic sequence), is in thrust contact with medial Ordovician limestones above and below. This also points to a late imbrication of the Carver's Falls Thrust Sheet after it had been emplaced at its present site. However, the nature of the thrust contacts is not well defined there. Instead of being overthrust by limestones on its eastern side, it could also lie structurally above those limestones and form a klippe. In the former case, the higher thrust must have crosscut the basal thrust of the Taconic Thrust Sheet to bring Taconic rocks to a position structurally below the medial Ordovician limestones. The western side of the sliver of Taconic material would then be bounded by the Basal Taconic Thrust which is older than the one following it to the east. This would be another example where the frontal thrust is not younger than those behind.

For the Giddings Brook slice of the Taconic Allochthon, which immediately overlies the carbonates in the field area, Rowley (1983) has demonstrated a hindward younging of thrust faults. He found that the basal thrust of the Taconic Allochthon is folded and crosscut by various faults including the Frontal Thrust. The examples for a similar sequence from the field area could support this observation of a two step fault history. However, the field data are not sufficient to determine whether it is really Taconic lithologies that are incorporated in the thrust sheets of the area studied.

Even though a firm conclusion cannot be reached on this matter from present data, it is clear from the observations presented that:

The three thrust sheets found in the field area are emplaced in front of and below the Taconic Thrust Sheet and have been transported together with it. The emplacement was from the east-southeast judging from the trend of fold axes and thrust traces which are perpendicular to that direction.

CHAPTER V

TECTONIC INTERPRETATION OF THE FIELD AREA

Several observations in the area studied can be interpreted in plate tectonic terms and this is the aim of the following chapter.

The stratigraphy can be explained as indicating the first stages of a "Wilson Cycle" (Dewey & Burke, 1974).

The lowest unit above the continental basement of Grenvillian gneisses, the Potsdam Sandstone is related to subsidence following the rifting of the continental crust. It is however, not the oldest known sediment in this sequence. Instead, to the east sandstones and carbonates of early to medial Cambrian or greater age are found such as the Winooski Dolomite and below it the Cheshire quartzite (Theokritoff & Thompson, 1969; Doll et.al., 1961). They, too, are widespread clastic sheets indicating the initial stages of thermal subsidence. That they are not present below the Potsdam, indicates an onlap of sedimentation continentward during the first phase of thermal subsidence after rifting. This produced what has been termed the steer's head configuration (Burke, 1981).

The rifting sequence is not easily recognized in New England, because it is now situated in the center of the Appalachian orogen where the rocks have undergone deformation and high grade metamorphism that obliterated sedimentary structures.

Rifting of the Proto-Atlantic Ocean started probably in the early Cambrian or even latest Precambrian (Bird & Dewey, 1970; Rankin, 1976) and was located in a roughly north-south axis to the east of the present day Green Mountains.

The lack of sediments from the early phase of subsidence in the field area (i.e., older than Potsdam) is consistent with its position far to the west of the rift axis.

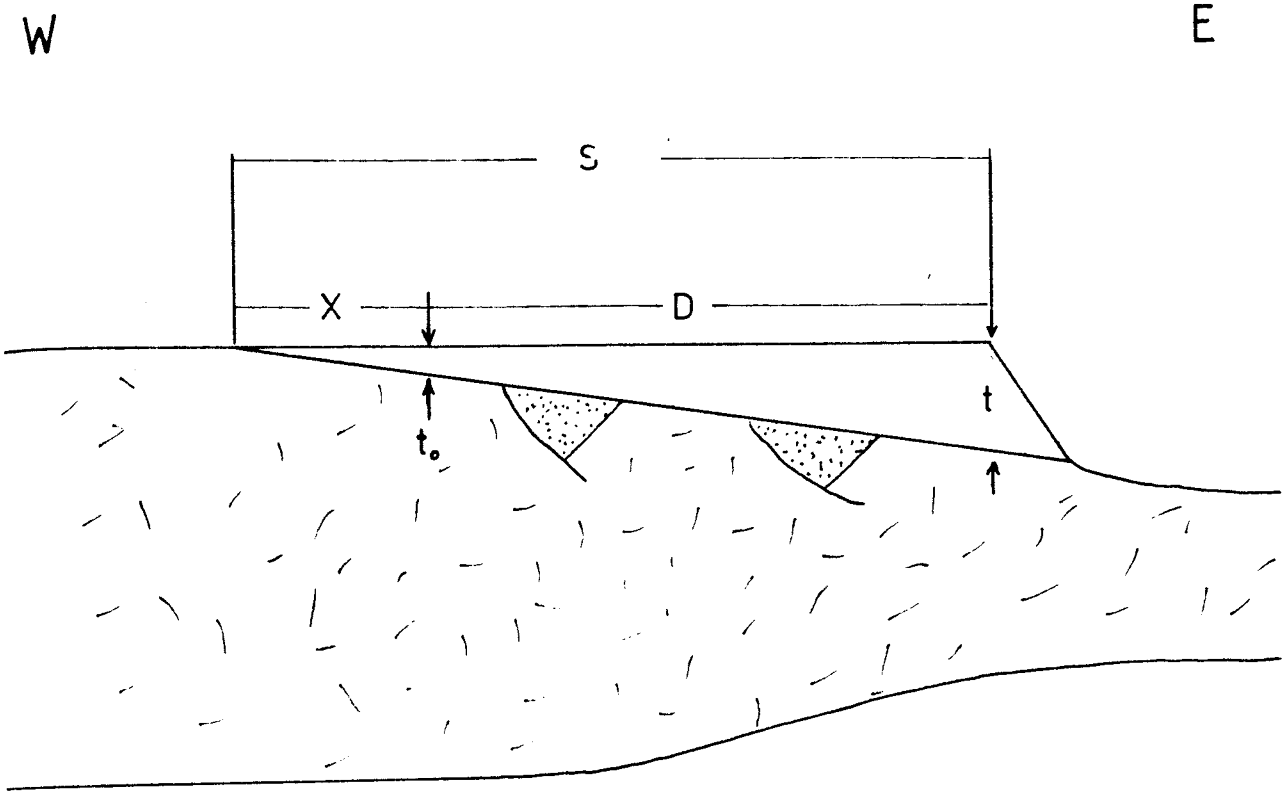
The sediments overlying the Potsdam are good examples of shallow marine shelf carbonates. They represent the opening stage of the Iapetus Ocean, during which subsidence was mainly produced by thermal subsidence caused by cooling following rifting and lithospheric stretching (McKenzie, 1978). The persistent shallow marine character of the Cambrian to early Ordovician carbonates demonstrates, that the sedimentation kept pace with subsidence. Present day examples show that passive continental margins have a sedimentary cover which is a seaward thickening wedge (Watts & Steckler, 1979; Boillot et. al., 1979) caused by increasing subsidence away from the continental margins. The maximum thickness of the post-rifting sedimentary sequence at the shelf edge is about 5-6 kilometers.

Whereas the initiation of the passive continental margin is not recorded in the sediments of the field area, its termination is visible in the lithologic change in the youngest rocks of the area mapped. The Middlebury Limestone with its coral fauna and sheets of fossil fragments clearly is a shallow marine deposit, the Glens Falls Limestone is formed at greater depths such as on the fore-reef slope and the Hortonville Shale must have been deposited deeper yet (Zen, 1967; Cisne et. al., 1982). This deepening was caused by the beginning of tectonic destruction of the passive margin in Trentonian time (Rowley & Kidd, 1981). Assuming that the Paleozoic shelf was of similar dimensions as present day Atlantic-type margins, one can arrive at an estimate of the relative position of the field area within the shelf. The sediment thickness of the Jurassic to Cenozoic margin at the shelf edge off the

North American east coast is about 6000 meters (Watts & Steckler, 1979). This margin is about 160 million years old. In the Paleozoic an Atlantic-type margin was present in New England from at least the early Cambrian to the beginning of the medial Ordovician i.e., during about 110 million years (about 570-460 million years before present). In the southern Appalachians most of the shelf width is still visible and the dimensions of this margin seem to be comparable to those of modern passive margins (Colton, 1970).

In a schematic cross-section given by Colton the maximum thickness of the shallow marine sequence of early Cambrian to medial Ordovician is 4.2 kilometers. The width of this Paleozoic shelf appear quite big-about 300 kilometers. Thus it appears as though the Paleozoic North American continent was bordered by a shelf up to twice as wide as that of the present day Altantic margin of eastern North America.

The ratio between the thickness of the shelf sequence at a given point and the assumed maximum thickness at the shelf edge should be a measure for the distance of that point from the shelf edge. In the field area, the shelf sequence from the base of the Potsdam Sandstone to the top of the Orwell Limestone is about 425 meters thick. Fig. 5.1 shows how the position of the field area can be calculated from this value and an assumed maximum thickness at the shelf edge of 6000 meters. According to this wedge model, the distance of the field area from the shelf edge is about 93% of the entire shelf width. Admittedly, this calculation makes a lot of assumptions about the shape and dimensions of the Paleozoic shelf and can therefore only give a very rough idea of the position of the field area. But stratigraphic evidence for a good



$$X : S = t_o : t$$

$$= 425 \text{ m} : 6000 \text{ m}$$

$$X = 0.07 \times S$$

$$D = 1 - 0.07 \times S$$

Fig. 5.1: Estimation of position of field area on shelf.

S = Shelf width, t_o = Thickness of shelf sequence observed in field area, t = Thickness of shelf sequence at shelf edge, D = Distance of field area from shelf edge.

part of the carbonate sequence supports the result given by the model. In particular, the presence of flat pebble conglomerates in the Sciota Limestone requires a peri-to supratidal environment. Another indicator is the sheets of coarse fossil debris in the same unit and the younger Middlebury Limestone which are ascribed to tidal currents. However, shallow water conditions are characteristic of the whole carbonate shelf, and eustatic sea level changes could have changed sedimentary conditions for short time intervals on the entire shelf. Some support for the large distance from the shelf edge to the field area comes from the stratigraphic record to the southwest of the field area. In the Central Mohawk Valley about 75 kilometers projected distance across strike from the field area the Cambrian and Ordovician sequence is extremely thin totalling about 150 meters and to a large degree only present in non-persistent lenses bounded by a number of disconformities (Fisher, 1980).

The Providence Island Dolostones, like the whole of the Beekmantown Group, is not at all present in the Western Mohawk Valley (Fisher, 1977). In the field area, the allochthonous Providence Island is at least 150 meters thick. This may be interpreted such that the Western Mohawk Valley (i.e., west of Canajoharie) represents a point past the hinge region for the subsidence of the North Atlantic margin during the lower Ordovician. With this interpretation the area studied would indeed lie relatively close to this hinge region.

This landward position of the field area is important because it allows conclusions about the amount of transport on the thrust faults that have been mapped in the field.

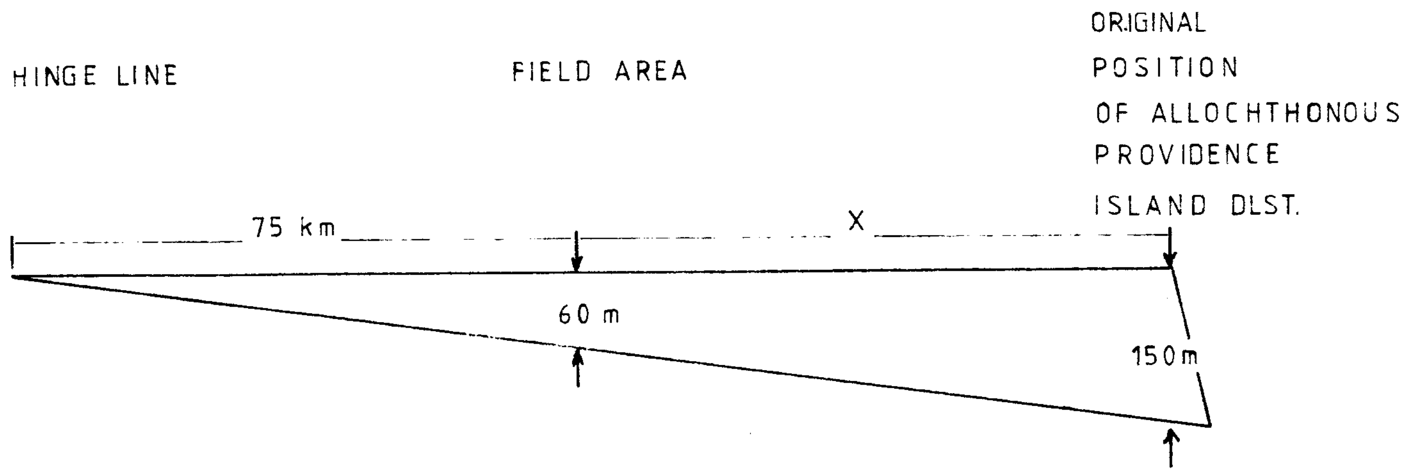
The deepening of the water in the mid Ordovician has been attributed to the onset of obduction of an island arc and associated accretionary prism onto the shelf (Rowley & Delano, 1979; Rowley & Kidd, 1981). This convergent plate motion was the cause for large scale thrusting. As part of that process the Taconic Allochthon, which consists of sediments originally deposited on the continental rise (Bird & Dewey, 1970), was thrust toward the continent, together with parts of the underlying shelf carbonates.

The Frontal Thrust of the Taconic Allochthon and several underlying carbonate thrust sheets are occupying the eastern half of the area of this study. If the estimate for the position of the field area on the shelf is correct, then the Taconic rocks (and probably the carbonate thrust slivers beneath it) must have traversed the bulk of the width of the carbonate shelf! For a shelf width similar to that of modern analogs like northern Australia, North America or South America, on the order of one hundred kilometers, the displacement on the Taconic Basal Thrust and possibly on the thrusts under the carbonate sheets would be at least 90 kilometers.

Considering the data from the Paleozoic shelf of the southern and central Appalachians (Colton, 1976), where the shelf seems to have been up to 300 kilometers wide the displacement would be estimated by the same rationale as 270 kilometers or more. This seems excessively large and reflects the large amount of uncertainty in the underlying assumptions. It is possible for example, that the stratigraphic thickness mapped in the field area is less than normal north and south along strike, which leads to an estimate of the position of the field that is too far away from the shelf edge and hence, too big an estimate for the amount of thrust displacement.

Applying a model proposed by Rowley (1982) to estimate thrust fault displacements independent of their stratigraphic throw yields more probable results. The model is based on the observation that carbonate shelves in cross section approximate triangular seaward thickening wedges. In this calculation, the Mohawk Valley around Canajoharie is taken as the "hinge region" or the top of the shelf wedge. Two values for the thickness of the Providence Island Dolostone are used to determine the original shape of the wedge.

In its autochthonous state, the Providence Island has a maximum thickness of 60 meters (Fisher, 1968; Roma, personal communication) and the thickness of the allochtohonous Providence Island as mapped in the Carver's Falls Thrust Sheet is at least 150 meters. The Carver's Falls Thrust Sheet now overlies autochthonous Providence Island Dolostone. Hence the initial horizontal distance between the point of 60 meters thick and that of 150 meters thick Providence Island is equivalent to the minimum displacement of the overlying thrust sheet. The tip of the shelf wedge lies about 75 kilometers to the west of the field area judging by data from the Mohawk Valley. Fig. 5.2 shows the geometric considerations used to calculate the displacement by this method. It yields a value of about 110 kilometers. This is certainly a very coarse approximation, but it correlates well with values obtained from different stratigraphic data by Rowley (1982) for the Champlain Thrust, situated along strike north of the field area. The displacement values there are given as 80-94 kilometers (Rowley, 1982).



$$\frac{X + 75\text{ km}}{150\text{ m}} = \frac{75\text{ km}}{60\text{ m}}$$

$$X = \frac{150\text{ m}}{60\text{ m}} \times 75\text{ km} - 75\text{ km}$$

$$= 2.5 \times 75\text{ km} - 75\text{ km}$$

$$\underline{\underline{X = 112.5\text{ km}}}$$

Fig. 5.2: Displacement calculation for thrust faults in field area based on the thickness difference between autochthonous and allochthonous Providence Island Dolostone.

The Champlain Thrust is known as far south as Cornwall, Vermont on Route 74 (Doll et. al., 1961) where it splits up into three or so poorly defined thrusts, from west to east the St. George, the Orwell and an unnamed thrust (Coney et. al., 1972). They are not mapped any further south than the town of Orwell, about 15 kilometers north of the field area.

Suggestions for the continuation of these three thrusts in the field area may now be made, with the St. George thrust underlying the West Haven Sheet, the Orwell thrust under the Forbes Hill Sheet and the unnamed thrust as base of the Carver's Falls Sheet. It is suggested from the similar displacement values, that the thrusts in the field area constitute the southern continuation of the Champlain thrust.

That the thrusts in carbonates are not short segments of only local extent is indicated also by recent COCORP seismic profiling undertaken about 15 kilometers to the south of the field area (Brown et.al., 1983). The COCORP profile shows a reflector (event N) at shallow easterly dip in the carbonate sequence west of and underlying the one reflecting the Taconic Basal Thrust (event O). I interpret this reflector as the seismic footprint of the southern continuation of the Champlain Thrust, the outcrop traces of which are the thrusts mapped in the area of this study. Further to the south, these thrusts pass upward into mid Ordovician shales and are only traced with difficulties. They are expressed in melange zones striking north-south as mapped by Bosworth & Vollmer (1981) and the underlying carbonates are only present as exotic blocks and a small thrust sliver in some of these melanges e.g., at Bald Mountain, 22 kilometers east of Saratoga Springs, New York.

The field area as interpreted in the cross-sections shows a geological profile very similar to fold and thrust belts in the foreland of the Alps or the Canadian Rocky Mountains: one side is bounded by undeformed shallow marine sediments underlain by continental basement, and it displays an increasing amount of deformation by thrusting. On the other side it is limited by the strongly allochthonous continental rise sediments of the Taconic Allochthon. The alpine counterpart of the Taconics would be the Pre-Alpine Nappes.

The difference between the field area and the well-studied portions of Mesozoic mountain belts mentioned is the size. The foreland thrust belt of the Rocky Mountains is over a hundred kilometers wide (Dahlstrom, 1970); that of the Alps is of the same order (Laubscher, 1973; Windley, 1977, Fig. 18.19). In the field area, however, the distance from the edge of the allochthonous continental rise nappes to autochthonous foreland is less than 10 kilometers. Several factors may be considered to account for this:

1. A narrower shelf
2. Transport of nappes farther onto the foreland during Taconic orogeny
3. Reactivation of thrusts during the Acadian orogeny

The width of the shelf during the lower Paleozoic does not seem to have been significantly less than that of present day examples. The Geological Map of Vermont (Doll et. al., 1961) shows possible shelf lithologies to the east of the Green Mountain Anticlinorium (e.g., Plymouth Member of the Hoosick Formation) that is 60 kilometers east of the field area. The shelf edge is assumed to have been in the vicinity of the Connecticut Valley (Bird & Dewey, 1970) about 100 kilometers east of the field area. Hence, the width of the shelf does not seem to have been narrower than normal. Data from the southern Appalachians suggest even the opposite (Colton, 1970).

The Taconic Allochthon may have been transported across the largest portion of the shelf width during the Taconic Orogeny. This would require almost 100 kilometers of transport during the medial Ordovician. Until recently this was assumed to have occurred by gravity sliding (e.g., Bird & Dewey, 1970; Zen, 1972; Rodgers, 1970).

However, Rowley & Kidd (1981) present good evidence that emplacement was a tectonic process powered by the continued convergence of the Ammonoosuc Island Arc and the continental shelf, involving polyphase deformation of already lithified rocks. The apparently complicated tectonic history may reflect such a long movement of the Taconic Allochthon and underlying carbonate sheets.

The medial Ordovician was a time of relatively high sea level, where marine sediments were deposited far inside continental areas (Cook et. al., 1975). It was the time of the maximum of the Tippecanoe transgressive sequence (Sloss, 1972). The high sea level may help to explain the large continentward displacement on the Taconic and the underlying thrusts in shelf rocks as well as the small structural thickness of the shelf thrust stack and the small number of thrusts in it compared to say that of the Canadian Rocky Mountains. In the area studied, the thrusts originated at a trench where rocks from the accretionary prism moved across recently deposited shales. The instant loading of water rich, impermeable sediments with those thrust sheets was an ideal condition for the creation of high pore pressure in the shale. A high pore pressure leads to a drastic reduction of shear strength in a material (Hubbert & Rubey, 1959; Weber, 1980). With a high sea level the conditions for high pore pressures were moving upward on the outer trench slope than in other fold and thrust belts. It seems likely that thrusting would be concentrated to the high pore pressure horizon for long distances.

The shortening required by the convergent plate motion would be accomodated in one long thrust rather than a large number of imbricate thrusts and therefore the thrust sheet as a whole would be wider and thinner than one that formed with intensive imbrication of thrusts. The relative thinness of the thrusted portion is demonstrated by seismic data (Brown et. al., 1983).

The thrusts on which the carbonates and the Taconic Allochthon moved, may have been reactivated during the medial Devonian Acadian orogeny. There is some evidence for shortening in the mid-Devonian molasse of the Catskill Mountains and the fold and thrust belt seen in the Helderburg Limestones in the Hudson Valley which is ascribed to Acadian tectonics (Geiser, 1981; Ratcliffe, 1975). It has been proposed that the Champlain Thrust was active during both the Taconic and Acadian orogeny (Stanley & Sarkisian, 1972). There is some evidence for a second deformational phase in the field area which might be related to Acadian movements, but it could just as well be of Taconic origin, since there are no rocks of post-Taconic and pre-Acadian age there that could help decide the question. Further to the south in the medial Ordovician flysch of the Hudson Valley, Bosworth & Vollmer (1981) have reported that structures and melange lithologies created by movements in the Taconic fold and thrust belt are truncated by the overlying Devonian beds. Had there been thrust movement during Devonian time, the same deformational structures should be present in the pre-Acadian Devonian strata as well. Since they are not, Bosworth & Vollmer conclude that the formation of the melange by movement of the overlying thrusts was restricted to the Taconic deformational event. The melange studied by Bosworth & Vollmer can be traced into the field area of this study. If their argument is right, then it would apply for that area as well and argue against any Acadian movement on the thrust faults found in my field area.

Late Normal Faults

The normal faults in the field area postdate all other deformations. Some of them are extremely well expressed in the topography. This may be an indicator of them being recent features. It is known that the Adirondack Mountains which border the field area to the west are presently doming or moving up (Barnett & Isachsen, 1980; Isachsen, 1975; Burke, 1976). An earthquake on October 7, 1983 of magnitude about 5.0 with its epicenter in the central eastern Adirondacks is the most recent and direct proof of such movements. It is therefore tempting to relate normal faulting in the field area to Adirondack uplift. However, the sense of offset on the north-south normal faults is west side down, yet the Adirondacks which lie west, are going up. Normal faults require extension and if the fault surfaces are listric, rotation of fault blocks will occur which might produce the observed offset. The required extension could be produced by the doming up of the lithosphere in the Adirondacks which would create extension in the upper part of the crust (Fig. 5.3). However, most of the normal faults bordering the Adirondacks on their eastern margin show an east side down sense of movement (e.g., the McGregor fault). Therefore, it is difficult to relate the west side down faults to a doming of the whole Adirondacks. It is possible that the normal faults in the field area are reactivated, rift related basement faults forming part of a graben system that defines the Champlain and Hudson Valleys.

A sub-bottom seismic profiling study of lake sediments in central Lake Champlain (Chase & Hunt, 1972) has yielded results that can best be explained by west down normal faulting in Holocene or even historic time. Their profiles show that thicknesses of Pleistocene Lake Vermont clays are much greater in the western part of the lake than in the east where they seem to abut against Paleozoic (?) bedrock (Fig. 5.4).

At the southeastern margin of the Adirondacks, normal eastward down stepping faults such as the McGregor Fault trending north-south through Saratoga Springs are believed to be in part very young (Young & Putman, 1979; Williams, Bosworth & Putman, 1983). It is possible however, that at least some of the normal folds observed in the field area are of medial Ordovician or earlier age. Cisne et. al. (1982) in the Mohawk Valley have studied the paleobathymetry of mid-Ordovician (Trentonian) limestones and shales through fossil assemblage analysis. They have found good evidence for syn-depositional block faulting and have ascribed this to extension caused by flexure of the shelf and underlying lithosphere due to the overriding arc. It is imaginable, that in the field area this flexure occurred earlier during the attempted subduction because of its situation closer to the trench. This might also be cited as an explanation for the

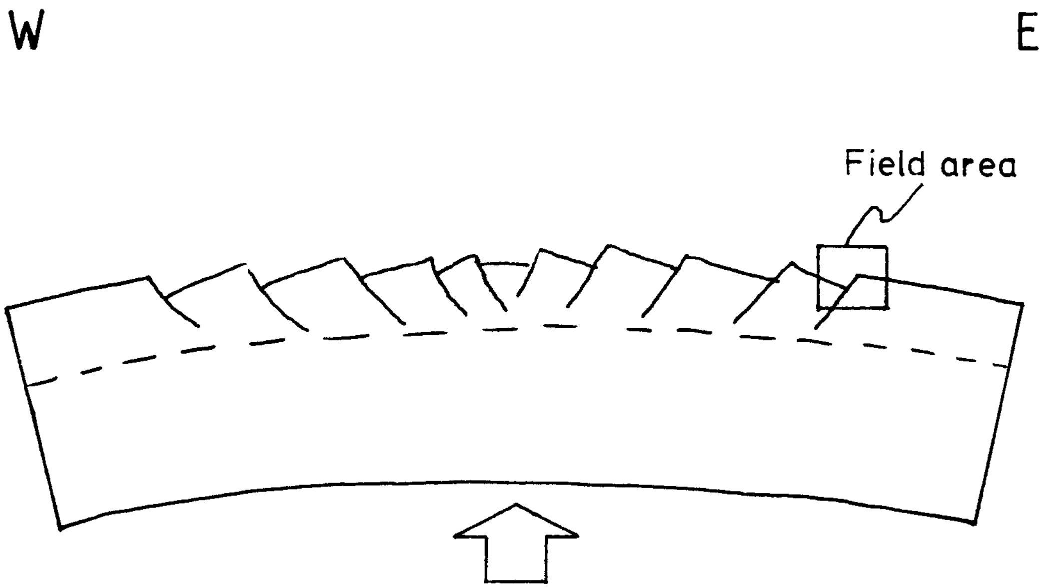


Fig. 5.3: West side down normal faults related to the uplift of the Adirondacks as extensional features in the upper crust in a doming of the lithosphere.

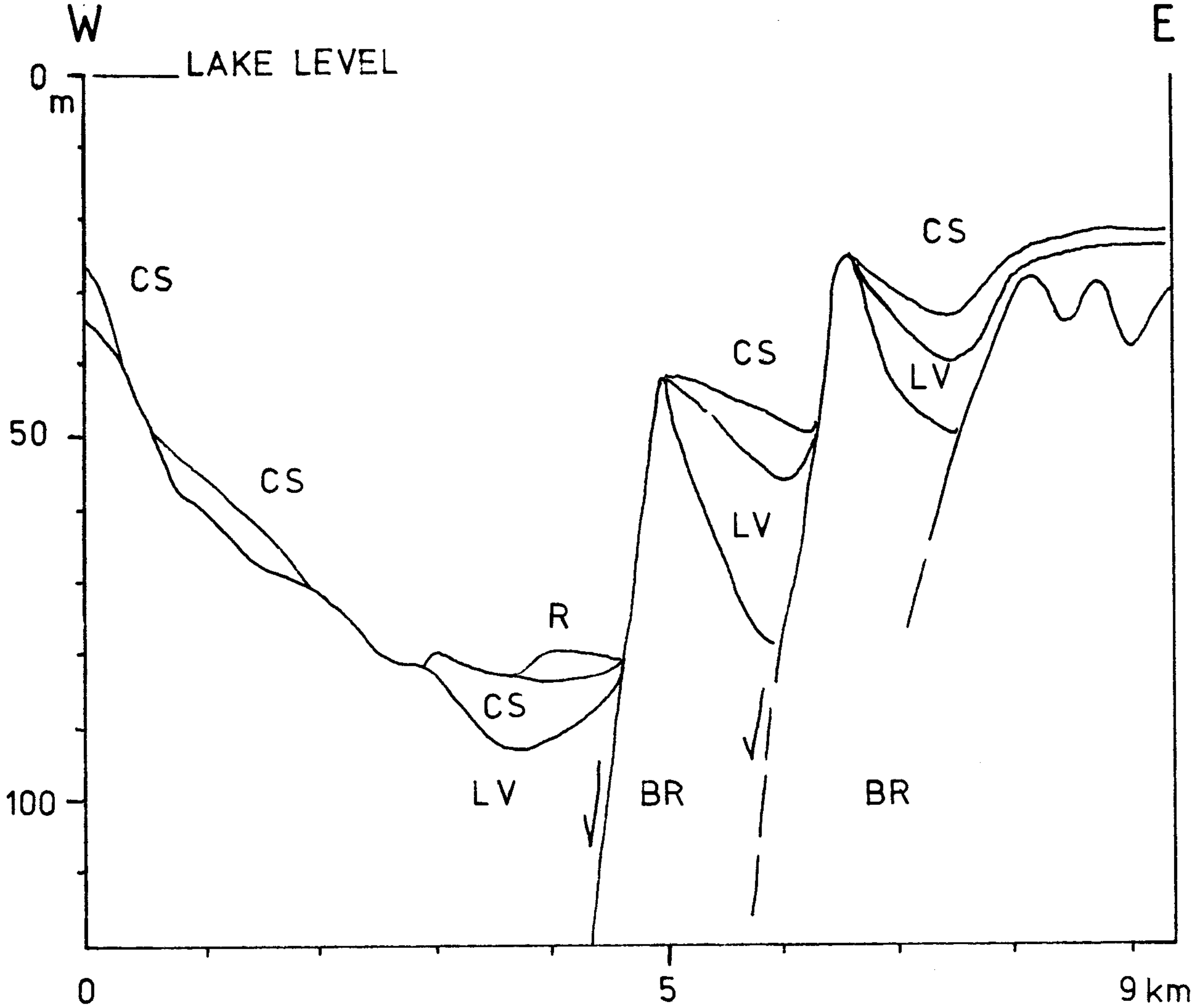


Fig. 5.4: Seismic reflection data from upper Lake Champlain, indicating recent west side down faulting. R = Recent deposits, CS = Champlain Sea sediments (less than 10 000 years old), LV = Lake Vermont sediments (10 to 11 000 years old), BR = Paleozoic or older bedrock.

After Chase and Hunt (1972)

thickness change in the mid-Ordovician limestones. Thus, some of the normal faults in the field area might have formed already in the medial Ordovician. However, they must in this area have moved after the emplacement of the carbonate thrust sheets since they cut across them demonstrably. They may indeed reflect displacements as recent as those inferred for the Champlain Valley to the north.

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