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GUIDEBOOK 2

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FOREWORD

About the Guidebook

The Vermont Geological Society is pleased to present the second in its series of Guidebooks, containing field trip guides of trips sponsored by the Society in its 14 year history. Each guide has the date of the original field trip on the first page, but the contents of each represents the most recent field work done in that area. Not all leaders of Vermont Geological Society field trips have been willing and/or able to publish guides, but, in the future, when enough manuscripts for a guidebook have accumulated, we will publish another volume.

Credits

Thanks to Charlotte Mehrtens and Rolfe Stanley for reviewing one field trip guide. Camera-ready copy was prepared by Jeanne Detenbeck.

About the Society

The Vermont Geological Society was founded in 1974 for the purpose of:

1) advancing the science and profession of geology and its related branches by encouraging education, research and service through the holding of meetings, maintaining communications, and providing a common union of its members;

2) contributing to the public education of the geology of Vermont and promoting the proper use and protection of its natural resources; and

3) advancing the professional conduct of those engaged in the collection, interpretation and use of geologic data.

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October 6, 1984

STRATIGRAPHY AND STRUCTURE OF THE CAMELS HUMP GROUP ALONG THE LAMOILLE RIVER TRANSECT, NORTHERN VERMONT

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FIELD TRIP GUIDE C

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INTRODUCTION

The purpose of this excursion is to present the stratigraphic and structural relationships of the Camels Hump Group in northern Vermont. The excursion crosses the Georgia Mountain anticline and Hinesburg thrust in the Milton 7 1/2 minute quadrangle, the Enosburg Falls - Fletcher anticline of the Gilson Mountain quadrangle and the Richford syncline of the Jeffersonville 7 1/2 quadrangle. The eastern part of the Jeffersonville quadrangle lies just to the west of the Green Mountain anticlinorium (Fig. 1.)

Significant problems to be addressed on this trip are the correlation of stratigraphy and comparisons of structure west and east of the Enosburg Falls - Fletcher anticline, correlation of the Underhill Formation with the established western stratigraphy, and the origin of the Richford syncline exposed in the adjacent Jeffersonville 7 1/2 minute quadrangle to the east.

MAPPING AND ACKNOWLEDGMENTS

The mapping conducted by the author in the Gilson Mountain and Jeffersonville quadrangles during the period of 1985 to the present has been funded by the Vermont Geological Survey under the direction of Charles Ratte'. Mapping was done directly on 1:5000 orthomap aerial photographs and compiled at 1:12000. The author has benefited from able assistance in the field by students completing field camp projects in the Gilson Mountain and Jeffersonville quadrangles. The early projects were conducted on 1:12000 enlargements of the topographic base and include the following students: Dave Marshall, Chris Miksic and Todd Worsfold (1982); Debra Merrill, Steve Schope, Scott Schulein, and Dan Dowling (1983); Dave Iseri, Doug Friant, Doug Graham, Robert Myers and Jeffrey Slade (1984). Later projects were completed on the 1:5000 base by Dave Greenwalt, Greg Koop and Hugh Rose (1985) and Michael Landsman (1986). Peter Thompson's (1975) work in the Enosburg Falls and Jeffersonville quadrangles is shown in part in the northeastern corner of Figure 3B. The author has also benefited greatly by ongoing studies by his graduate students Maurice Colpron and Bill Dowling whose work in the Oak Hill Group in Quebec has helped to focus on the regional relationships discussed in this report.

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REGIONAL SETTING OF THE CAMELS HUMP GROUP

The Camels Hump Group was first defined by Cady (1956) in the Camels Hump quadrangle of Vermont and subsequently subdivided into formations by Doll and others (1961). The Camels Hump Group are almost entirely rift related rocks which predate the development of the passive continental margin along the ancient "western" boundary (present coordinates) of the Iapetus Ocean (Williams and Stevens, 1975). The Group thus includes most of the rocks stratigraphically overlying the Grenville basement and underlying platform slope or rise sequence rocks of the ancient margin.

In the western outcrop belts of the Camels Hump Group and correlative lower Oak Hill Group rocks of Quebec (Clark, 1934) the sequence is overlain by rift-drift transition "basal" quartzites of the Cheshire Quartzite and drift stage dolomites, quartzites and carbonate breccias of the shelf and slope facies. To the east the rocks of the Camels Hump Group are overlain by slope and rise facies rocks of the Sweetsburg and Ottauquechee formations. Significant facies differences on both local and regional scales are noted by previous workers and should be expected in rift environments both parallel and perpendicular to the basin. Detailed stratigraphic comparisons of rocks in the rift basin have been hampered by a lack of detailed mapping within the Camels Hump Group and deformation which postdates the formation of the basin. This excursion attempts to detail the stratigraphic relationships of the Camels Hump Group in the vicinity of the Lamoille River transect.

The Lamoille River transect from the Milton to the Jeffersonville quadrangle, provides an excellent array of exposures which help to define the across strike nature of the ancient rifted basin. The western sequence of the Camels Hump Group displays a thin stratigraphic sequence with at least one erosional unconformity and rift-drift to platform cover rocks. The eastern sequence involves a greater percentage of volcanic rocks, thicker rift-clastic sequences and slope to rise cover rocks (Fig. 2). The increased rift related subsidence to the east supports the model of formation of "instantaneous" eastward facing rift basins by lithospheric stretching and Airy-type subsidence synchronous with rifting (e.g., McKenzie, 1978).

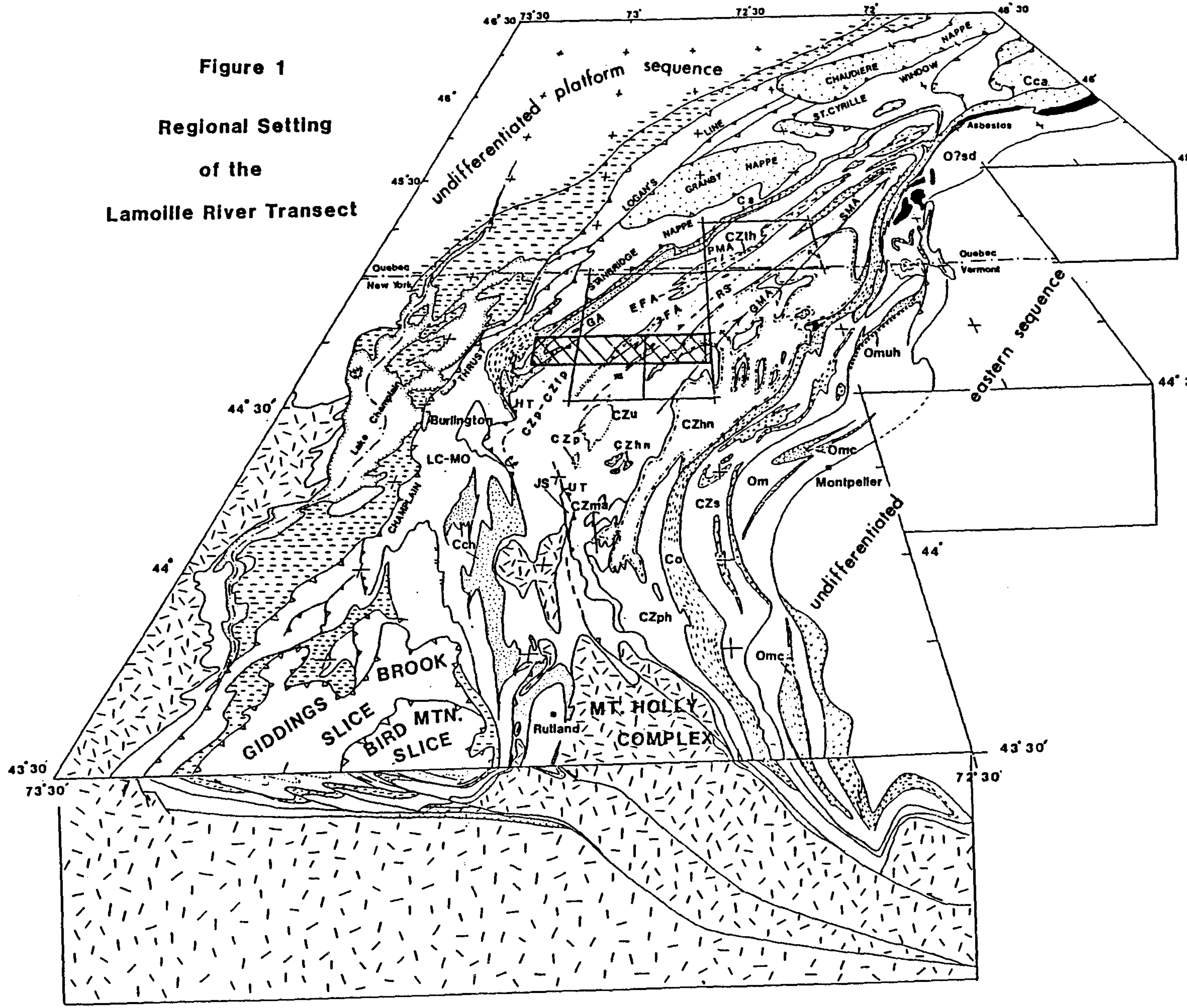
Significantly, the Lamoille River section differs in detail from that observed to the north in Quebec (e.g., Clark, 1934; Dowling and others, 1987; Colpron and others, 1987) and in the Enos-

Figure 1

Regional Setting

of the

Lamoille River Transect



burg Falls quadrangle (Dennis, 1964). To the south, in the Lincoln Mountain quadrangle, the Camels Hump Group is considerably shortened and only the westernmost and easternmost parts of the Lamoille River section are lithically comparable. The missing section to the south lies mostly within the Gilson Mountain quadrangle and the western part of the Jeffersonville quadrangle. These rocks and their cover presumably have been removed from the Lincoln Mountain area during allochthon emplacement along thrust faults rooted below and within the Underhill Slice of Stanley and Ratcliffe (1985; Fig. 1).

STRATIGRAPHY OF THE CAMELS HUMP GROUP ALONG THE LAMOILLE RIVER TRANSECT

In this section the stratigraphy of the Camels Hump Group as exposed along the Lamoille River transect is described. The formation names as originally proposed are retained wherever possible. However, in light of the facies relationships observed across the strike belt, some modification of the stratigraphic nomenclature is recommended. It is not the purpose to propose such nomenclature here since much new geologic mapping involving stratigraphic equivalent rocks described here is ongoing throughout the state. Therefore, the stratigraphic names used in this report should be considered informal.

The stratigraphy of the Gilson Mountain quadrangle is described first. Correlations with the adjacent Milton quadrangle to the west are suggested where appropriate. Secondly, the stratigraphy east of the Enosburg Falls - Fletcher anticline is discussed with the emphasis placed on correlation with the western quadrangles. A complete historical development of the stratigraphy is not outlined here, but the interested reader is referred to the excellent synopsis of the regional stratigraphy in the references under the relevant stratigraphic columns shown in Figure 2. A more formal presentation of the stratigraphy for the Gilson Mountain and Jeffersonville quadrangles is forthcoming (Doolan, in preparation; Vermont Geological Survey reports).

As shown on Figure 2, lithic correlatives of the Oak Hill Group of Clark (1934) have been traced southward into Vermont with varying degrees of success. Although Dennis (1964) and Booth (1950) recognize the general sequence from base to top of Tibbit Hill volcanics - Pinnacle Formation - White Brook - Fairfield Pond (equals the West Sutton and Frielighsburg formations of Charbonneau, 1980) - Cheshire Quartzite, the following changes in the stratigraphy are noted:

1. The Tibbit Hill is interbedded with clastic rocks in the Enosburg Falls quadrangle (Pinnacle facies of the Tibbit Hill of Dennis, 1964);

2. The White Brook dolomite is a discontinuous horizon in Vermont and when found does not precisely define the boundary between the equivalents of the Pinnacle and Fairfield Pond formations in Quebec (Booth, 1950; Dennis, 1964).

3. More coarse-grained conglomeratic facies are found in the Pinnacle Formation of Vermont and an overall increase in thickness of the Pinnacle is noted southward (Dowling and others, 1987).

4. The Call Mill Slate, and West Sutton Formation as mapped in Quebec are only locally definable in Vermont (Booth, 1950; Dennis, 1964).

These differences in the stratigraphic sequence noted in the Enosburg Falls quadrangle become even more apparent in the Gilson Mountain quadrangle.

THE STRATIGRAPHY ALONG AND WEST OF THE ENOSBURG FALLS - FLETCHER ANTICLINE

In this section the stratigraphy of the Camels Hump Group in the Gilson Mountain quadrangle is summarized for the regions along and west of the Enosburg Falls - Pinnacle Mountain anticline. The presence of faults and complex folding of the entire sequence precludes an accurate rendition of the unbroken stratigraphic succession; however, numerous topping criteria in the metagreywackes and excellent rock exposure has enabled a newly defined stratigraphic sequence to be defined.

Tibbit Hill Formation (CZth)

The Tibbit Hill Formation was first defined by Clark (1934) for a thick sequence of volcanic rocks which define the base of the Oak Hill Group in Quebec. It is separated from the overlying Pinnacle Formation in Quebec by a distinctive argillite called the Call Mill Slate. Southward, the Tibbit Hill volcanics are interbedded with metasediments and the Call Mill marker horizon is commonly not present (Dennis, 1964). Dennis (1964) notes the increase in metasedimentary material in the Tibbit Hill toward the south in the Enosburg Falls 15' quadrangle, and included the metasediments with the Tibbit Hill Formation.

[Continued on page C6.]

Figure 1. (Opposite page) The location of the Lamoille River transect is shown on this block diagram of the Quebec reentrant constructed with a single due-north vanishing point. Latitudinal cross-sections are undistorted but show south to north scale compression. 15' quadrangle boundaries are shown plus symbols throughout. The diagonally ruled area includes from west to east: part of the Milton, the complete Gilson Mountain and part of the Jeffersonville 7 1/2 minute quadrangles (see Figure 3A, 3B). The latter two quadrangles are the northern half of the Mt. Mansfield 15' quadrangle (Christman, 1959). The Enosburg Falls 15' quadrangle (Dennis, 1964) is shown immediately north. The Sutton quadrangle (Eakins, 1964) is outlined to the northeast in Quebec. The Camels Hump quadrangle (Christman and Secor, 1961; Thompson and Thompson, 1987) is directly south of the Mt. Mansfield quadrangle. Abbreviations: EFA: Enosburg Falls anticline; FA: Fletcher anticline; GA: Georgia Mountain anticline; GMA-SMA: Green Mountain-Sutton Mountain anticlinorium; HT: Hinesburg thrust; JS: Jerusalem slice; PMA: Pinnacle Mountain anticline; RS: Richford syncline; UT: Underhill thrust. Lithic designators shown on Figures 2, 3 and 4 except as follows: Єca and broad stipple pattern: Caldwell, Armagh, Granby and related rocks mostly in Quebec; Cch and fine stipple pattern: Cheshire Quartzite; CZma: Mount Abraham Formation; ЄZph: Pinney Hollow Formation; ЄZs: Stowe Formation; Om: Moretown Formation; Omc: Cram Hill Formation; Omuh: Umbrella Hill conglomerate; O?sd: St. Daniel Formation; horizontal dashes: middle to late Ordovician flysch; random dash: Grenville basement of the Adirondack Mountains, Green Mountain massif and Lincoln Mountain massif; black symbols: Thetford-Asbestos-Orford ophiolites of Quebec, Eden Mill ultramafic body of Vermont. Geology after Doll and others, 1961; Williams, 1978; Osberg, 1965; Doolan and others, 1982;

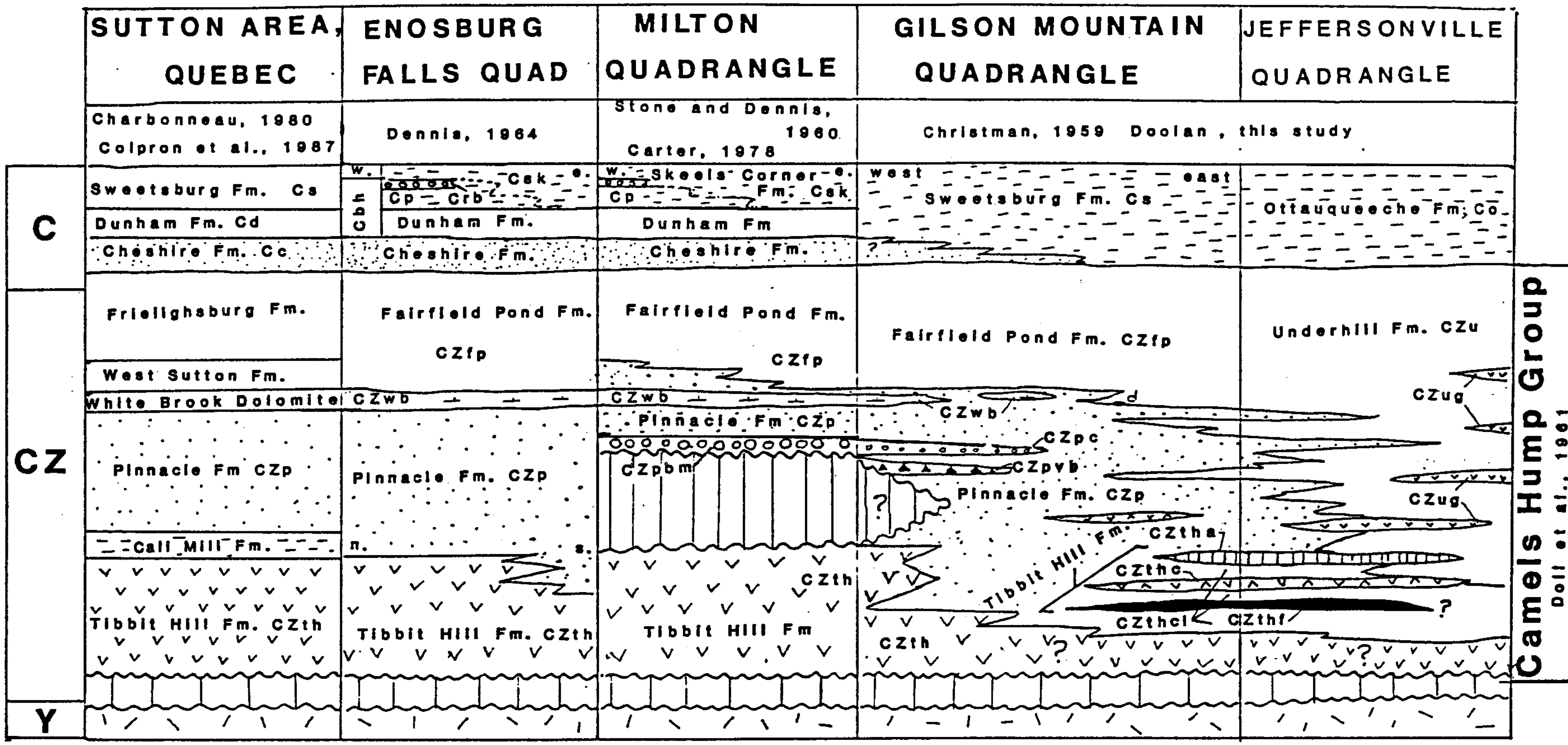
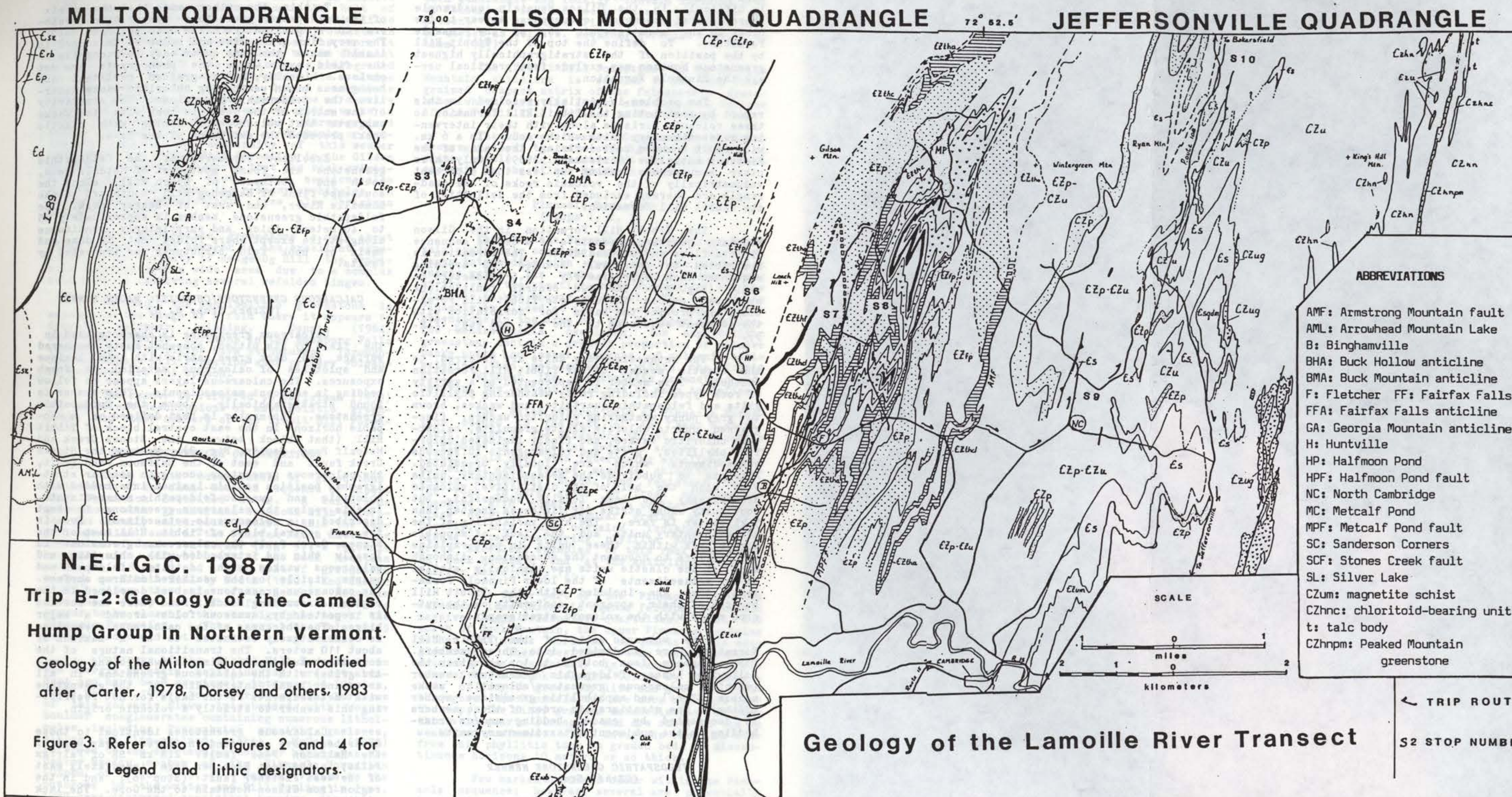


Figure 2. Stratigraphic correlation of the Camels Hump Group between Quebec and Vermont according to previous workers and this study. The stratigraphic column for the Milton quadrangle is modified from the interpretation of Carter (1979) by proposing an unconformity above the Tibbit Hill volcanics (Stop 2). The Pinnacle exposed in the Georgia Mountain anticline is stratigraphically and lithically correlative with the upper Pinnacle of the Gilson Mountain quadrangle.



N.E.I.G.C. 1987

Trip B-2: Geology of the Camels Hump Group in Northern Vermont.

Geology of the Milton Quadrangle modified after Carter, 1978, Dorsey and others, 1983

Figure 3. Refer also to Figures 2 and 4 for Legend and lithic designators.

Geology of the Lamoille River Transect

Christman (1959), confined the Tibbit Hill Formation to the metavolcanic rocks interbedded with metasediments in which amphibole and/or feldspar is identifiable in the field. This practice in effect redefined the Tibbit Hill Formation by excluding some greenstones which are interbedded with Pinnacle Formation. The problem is especially difficult in the Gilson Mountain quadrangle where several greenstone occurrences appear to lie close to the stratigraphic top of the Pinnacle Formation. To define the top of the Tibbit Hill by the position of the stratigraphically highest greenstone horizon may exclude from practical useage the Pinnacle Formation.

The problem is partially resolved in this report by restricting the Tibbit Hill Formation to three volcanic horizons along with their intervening metasedimentary units which occur in a 6 km. wide fault bounded belt defining the axis of the Fletcher anticline (Christman, 1959). Only three other occurrences making up less than 5 percent volumetrically of the volcanic rocks of the quadrangle are thereby excluded from the definition of the Tibbit Hill Formation.

The Tibbit Hill Formation in the Gilson Mountain quadrangle consists of a bedded sequence of lava flows and metasediments. The formation is restricted to four fault bounded NNE trending sequences each approximately 1-1.5 kilometers in width (Fig. 3). The two central fault block sequences expose the complete Tibbit Hill Formation, and provide the basis for correlation with the adjacent sequences to the west and east (Fig. 3).

The metasedimentary units are referred to the clastic member of the Tibbit Hill Formation (€Zthcl). This member is represented by a variety of rock types including rusty weathering phyllitic grits and feldspathic quartzites and wackes, brown to grey punky weathering chloritic wackes, fine-grained quartz chlorite albite granulite, calcareous chloritic wackes, and finely laminated chloritic phyllites, grits and volcanogenic tuffaceous metasediments which are locally brecciated. Attempts to subdivide the metasedimentary units even with the help of the distinctive volcanic stratigraphy has been locally successful, but continuity along strike for distances greater than 1 kilometer is rare. The high variability of the metasedimentary units and the lack of mappable marker beds within these units in part explains the failure to document the succession. Although many of the clastic units are lithically similar to the metasediments of the lower Pinnacle Formation, they are included with the Tibbit Hill because of their apparent conformity on the outcrop scale with the volcanic stratigraphy.

The volcanic units of the Tibbit Hill Formation are subdivided into three members. These members from bottom to top are given the following names: feldspathic greenstone member (€Zthf), calcareous greenstone/chloritic wacke member (€Zthc) and amphibolitic greenstone member (€Ztha). The stratigraphic order of these members is documented by graded bedding and/or cross-bedding in the subjacent metasedimentary units.

FELDSPATHIC GREENSTONE MEMBER (€Zthf; Stop 8)

This member is described by Christman (1959, p. 18) as follows:

> *The feldspathic greenstone is so-named because it contains lath-like remnant phenocrysts of plagioclase feldspar as much as one and one fourth inches long. These phenocrysts are seen most easily on the weathered surfaces as they weather white to light green in contrast to the fine-grained dark green groundmass. The phenocrysts have no apparent preferred orientation.*

To this, the author adds that the matrix of the feldspathic greenstone member often weathers to a distinctive olive drab to brown color. The very fine grain size and color of this rock in itself makes the unit quite easy to recognize in the field, even where the phenocrysts are not obvious. Locally, a fine-grained resistant and homogenous brown weathering chloritic wacke underlies the feldspathic greenstone. The similarity of the matrix of the volcanic rock with the wacke suggests that volcanogenic mixing with clastic rocks preceded the extrusion of the lava.

Excellent exposures of the feldspathic greenstone are found southwest of Metcalf Pond, south and southeast of Leach Hill, along the southeast flank of Hedgehog Hill southward to the Lamoille River, and west of Beaver Brook. The feldspathic greenstone member is approximately 25 to 40 meters thick and surprisingly continuous along strike except where folded out in dome and basin or hook fold structures or truncated by faults.

CALCAREOUS GREENSTONE/CHLORITIC WACKE MEMBER (€Zthc: Stops 6, 8)

Calcareous greenstone is characterized in the field by its pitted, somewhat rusty weathered surface and dark green matrix with thin laminae and splotches of calcareous material on fresh exposures. The calcareous layers appear to follow bedding in adjacent metasediments. This member is found stratigraphically above the feldspathic greenstone. The member is best defined as a mappable horizon in the east central block of Tibbit Hill (that block between the Stones Creek and Metcalf Pond faults, Fig. 3). West of the Stones Creek fault and east of the Halfmoon Pond fault, the calcareous greenstone occurs in similar stratigraphic position as thin laminae interbedded with chloritic and quartzo-feldspathic metasediments. In this region the calcareous greenstone is best described as a volcanogenic metasediment. Even in the east central block of Tibbit Hill east of the Stones Creek fault the calcareous greenstone is locally thin and interbedded with chloritic and calcareous wackes with detrital feldspar and quartz visible on the weathered outcrop surface. The calcareous greenstone is best developed in the region surrounding Metcalf Pond where the section is repeated by numerous folds around a major hinge. The thickness of the calcareous greenstone is conservatively estimated to be a maximum of about 110 meters. The transitional nature of the contact between metasediments (wackes and chloritic grits) with the calcareous greenstone in all areas and local interbedding of the greenstone with metasediments is the basis for not restricting this member to strictly a volcanic origin.

Calcareous greenstones identical to those described above are found in three areas west of the Halfmoon Pond fault: 2.8 km SE of Fairfax Falls; in the West Fletcher area immediately east of the West Fletcher fault (Stop 6); and in the region from Gilson Mountain to the Gore. The lack of other marker horizons of the Tibbit Hill Formation between the West Fletcher and Halfmoon Pond faults does not allow a direct comparison of these volcanic horizons with that of the Tibbit Hill stratigraphy to be made. Ongoing mapping will hopefully resolve the problem.

AMPHIBOLITIC GREENSTONE MEMBER (€Ztha; Stops 7,8)

The stratigraphic top of the Tibbit Hill Formation is defined here to be the amphibolitic greenstone member. This unit outcrops in massive ledges and whalebacks throughout the belt of Tibbit Hill exposures. Amphibole (actinolitic hornblende) occurs as dark green porphyroblasts in random to strongly lineated orientations where it defines an L-S fabric parallel to the dominant foliation. Locally the amphibole is retrograded to chlorite and/or chlorite/actinolite clumps pseudomorphing the earlier formed amphibole grains. Outcrops have the appearance of massive wacke and grade into calcareous chloritic wacke in some localities. Although the coarse-grained nature of this rock suggests a diabasic or gabbroic protolith, the continuity of this member throughout the Tibbit Hill Formation in the Gilson Mountain quadrangle suggests that the rock originated as a basic lava flow. Thin sections also support the view that the primary amphiboles in the rock were porphyroblasts (see also Christman and Secor, 1961).

The amphibolitc greenstone is only about 15-20 meters thick; however in the excellent exposures in the vicinity of Hedgehog Hill the amphibole outcrops over a wide area due to a complex fold pattern involving several refolded hinges.

At Stop 7 the amphibolitic greenstone is exposed in the hinge of a fold where it appears to display columnar jointing. Dennis (1964) describes volcanic horizons in the Enosburg Falls quadrangle with pillow lava structures. Throughout the Tibbit Hill Formation such primary volcanic textures are not common.

The reader is referred to Christman (1959) for excellent descriptions of amphibolitic greenstone and other members of the Tibbit Hill Formation.

OTHER OCCURRENCES OF GREENSTONE

Three other occurrences of greenstone occur west of the fault-bounded sequence of Tibbit Hill Formation described above. A calcareous greenstone with gradational contacts with chloritic and locally calcareous wackes within the mapped syncline east of the Fairfax Falls anticline and north of Sanderson Corner (Fig. 3); a volcanic breccia unit (€Zpvb) west of the Buck Mountain anticline (Stop 4) and a thin belt of greenstone mapped as Tibbit Hill by Booth (1950), and Carter (1979) along the axis of the Georgia Mountain anticline (Stop 2). The first two occurrences appear to be close to the stratigraphic top of the Pinnacle Formation. They are both thin (<15 meters) and could not be traced along strike.

Outcrops of the Tibbit Hill Formation in the Georgia Mountain anticlinorium are not abundant and have been previously interpreted (Carter, 1979; McBean, 1979) to occur as several horizons of thin volcanic flows interbedded with coarse boulder conglomerates containing numerous lithologies including Precambrian granitic gneisses, slates and fine-grained wackes. McBean (1979) referred to the spectacular boulder horizon as the Beaver Meadow member of the Pinnacle Formation (CZpbm of Figs. 2, 3, 4). Several visits to the outcrops by the writer and Bill Dowling in June, 1987 have uncovered consistent topping directions away from the Tibbit Hill volcanic outcrops toward the sharply truncating basal boulder conglomerate horizon. Based on this data and the fact that slate fragments identical to the Call Mill in the Quebec sequence are locally abundant toward the base of the conglomerate, we interpret the Beaver Meadow to be a basal conglomerate unconformably overlying the Tibbit Hill Formation. Pinnacle wackes interbedded with and overlying the Beaver Meadow boulder conglomerate are characteristic of the upper Pinnacle Formation in the Gilson Mountain quadrangle and locally contain isolated pods of dolomite. Similar occurrences of dolomite are present in the upper Pinnacle of the Sutton area of Quebec as well (W. Dowling, personal communication, 1987).

The Tibbit Hill Formation in the Georgia Mountain anticline is similar to the very fine-grained compact matrix of the feldspathic greenstone which marks the lowest stratigraphic horizon of the Tibbit Hill sequence in the Fletcher area.

The Beaver Meadow Conglomerate (€Zpbm) occurs at about the same stratigraphic position as slate pebble conglomerates and the volcanic breccia unit in the Gilson Mountain quadrangle.

Pinnacle Formation (Stops 1, 2, 3)

The Pinnacle Formation in the Gilson Mountain quadrangle consists predominantly of greywacke interbedded with rusty phyllitic grits with quartz veins. The wacke is variable even along strike and the stratigraphic succession of the Pinnacle is not known in detail. The base of the Pinnacle is placed at the top of the amphibolitic greenstone in the area of the Fletcher anticline. The metasediments overlying the amphibolitic greenstone contain a greater proportion of massive wacke units and lesser amounts of volcaniclastic material than metasediments within the Tibbit Hill. Without the amphibolitic greenstone bed, however, the contact would not be easy to define.

The major outcrop belt of Pinnacle Formation occurs to the west of the West Fletcher fault. East of the Hinesburg thrust, the Pinnacle is exposed in a repeated sequence across 4 anticlinal fold axes with intervening synclines. The anticlinal axes expose massive chloritic feldspathic greywackes which form resistant ridges. The rock is dark green on fresh surfaces and speckled with numerous white weathering plagioclase detritus and angular grains of quartz. Locally, this unit contains isolated angular clasts of tan weathering phyllite and phyllitic grit. Excellent examples of this rock type occur on Buck Mountain and at the anticlinal hinge at Fairfax Falls (Stop 1).

The upper part of the Pinnacle contains "cleaner" less chloritic quartz feldspar wacke and massive flaggy feldspathic quartzites. West and north of Buck Moutain, the upper Pinnacle contains rather continuous horizons of quartz pebble conglomerate. These resistant rocks contain calcareous matrices especially in proximity with horizons of White Brook dolomite (Stops 2, 3). Interbeds of grey to green phyllite and grit are found throughout the Pinnacle but are more abundant near the top. The finer grained metasediments display thin vein quartz stringers and are rusty to tan on weathered surfaces. The interbeds range in size from thin phyllitic tops on graded beds to discontinuous horizons 10 meters or so thick.

Few marker horizons occur within the Pinnacle sequence; however, several are potentially important horizons to further refine the Pinnacle stratigraphy in the area. These include a volcanic breccia horizon (€Zpvb; see Stop 4), a phyllitic conglomerate (€Zpc; Fig. 3), and a single horizon of calcareous greenstone (see above). These units all appear to be towards the top of the Pinnacle Formation (Figs. 2,3).

White Brook Formation (Stop 3)

The White Brook Formation of Clark (1934, 1936) only sporadically occurs in the Gilson Mountain quadrangle. Where present it occurs as discontinuous buff to cream colored massive dolomite and dolomitic sandstone. The quartz content contributes to a resistance from erosion and the common occurrence of quartz vein stringers. Dolomite is not restricted to a single stratigraphic horizon. The lowest dolomite is found at the top of the coarse-grained greywacke sequence which is taken to be the top of the Pinnacle when dolomite is absent. In the Milton quadrangle where the Pinnacle Formation overlies a coarse boulder conglomerate, dolomite occurs below the top of the Pinnacle section. In the region north of Buck Hollow school, dolomite also occurs sporadically above laminated grits and phyllites which stratigraphically overly the defined top of the Pinnacle. These stratigraphically higher dolomite occurrences are in contact with homogeneous grey to black slate. It is uncertain if these homogeneous slates correspond to West Sutton Formation of Clark (1934). The interval of dolomitic horizons is about 25 to 30 meters above the top of the Pinnacle Formation. Because of the sporadic, thin and discontinuous occurrences of dolomite in the Gilson Mountain quadrangle, the White Brook is not separated on the accompanying map; however, dolomite occurrences are located by "d" on Figure 3.

Fairfield Pond Formation (Stops 3, 5)

The Fairfield Pond was introduced to Vermont stratigraphy by Dennis (1964) as a member of the Underhill Formation. As defined by Dennis it includes the thinly laminated quartzites, grits and phyllites of the lower Cheshire Quartzite and the more homogeneous slates and phyllites of the West Sutton Formation. The Fairfield Pond is directly correlated with the West Sutton and Frelighsburg formations of Charbonneau (1980) in the Sutton quadrangle of Quebec. It may prove useful to future detailed mapping in pre-Cheshire rocks to subdivide the Fairfield Pond Formation into Frelighsburg and West Sutton facies. Examples of these facies will be shown at Stop 3.

As defined, all of the rocks stratigraphically above the Pinnacle and below the massive quartzite beds of Cheshire are included with the Fairfield Pond Formation except where the White Brook dolomite occurs. The contact between the Fairfield Pond and the Pinnacle closely corresponds to the lowest position of dolomite associated with quartz pebble conglomerate and wacke of the Pinnacle in the Gilson Mountain quadrangle. Argillaceous wackes, phyllites and grits below the contact are thin and discontinuous as are wacke and conglomeratic facies which sporadically occur above the contact. Grey to black phyllites mapped around the Buck Hollow anticline (Fig. 3) within the Fairfield Pond Formation are lithically similar to the West Sutton of Clark (1934) but occur at a higher stratigraphic position relative to the Pinnacle. Mapping along the expected strike path of these phyllites has not been completed.

A second outcrop belt of thinly laminated argillaceous quartzites and grits which stratigraphically overly Pinnacle wacke are found between the Armstrong Mountain and Metcalf Pond faults within the Fletcher anticline to the east. These rocks are identical to the Fairfield Pond lithologies mapped to the west but appear to have more phyllitic to schistose foliation surfaces. These rocks have been previously referred to as Underhill Formation but because of their close association with Pinnacle and their lithic similarity with the rocks to the west they are included here with the Fairfield Pond. Contacts between the Fairfield Pond lithologies and the amphibolitic greenstone member of the Tibbit Hill to the west of the Armstrong Mountain fault (Fig. 3) have not been observed, but a fault contact is suspected on stratigraphic grounds.

The Fairfield Pond Formation in the Milton quadrangle was restricted by Carter (1979) and Dorsey and others (1983) to argillaceous rocks east of the Hinesburg thrust. Rocks surrounding the Pinnacle in the Georgia Mountain anticline west of the thrust were referred to by these authors as the lower Cheshire. The lower and upper Cheshire of these authors corresponds to the Gilman Formation of Booth (1950). Stone and Dennis (1964) defined the Fairfield Pond to include the lower argillaceous quartzite of the Gilman Formation as well as the West Sutton Formation of Clark (1934). Consequently Dennis' map for the region defines the argillaceous rocks surrounding the Pinnacle in the Georgia Mountain anticline as the Fairfield Pond Formation. The use of both lower Cheshire and Fairfield Pond appears to be in violation of the existing stratigraphic nomenclature. On Figure 3, the lower Cheshire of Dorsey and others (1983) is designated as the Fairfield Pond Formation in keeping with the previous mapping of Stone and Dennis (1964). Lithically the rocks are similar on both sides of the Hinesburg thrust and differ only in the abundance of thin quartzite laminae in the argillaceous matrix. As noted by previous workers, the transition from the Fairfield Pond to the Cheshire is gradual. The apparent occurrence of "upper" Fairfield Pond close to the Pinnacle suggests that the Fairfield Pond Formation is considerably thinner on the west side of the Hinesburg thrust compared with the east side.

STRATIGRAPHY EAST OF THE ENOSBURG FALLS - FLETCHER ANTICLINE

As noted above, Fairfield Pond and Pinnacle lithologies occur east of the main belt of Tibbit Hill volcanic members which define the Enosburg Falls - Fletcher anticline (Fig. 3). East of this Fairfield Pond - Pinnacle sequence is an elongate belt of upper Tibbit Hill Formation (amphibolitic greenstone member and associated metasediments). The Armstrong Mountain fault at least locally separates the Tibbit Hill from the metasediments to the east and it is possible that the entire eastern amphibolitic greenstone west of the Armstrong Mountain fault is fault bounded.

From the amphibolitic greenstone eastward to the black slates and phyllites of the Sweetsburg Formation, no Tibbit Hill units are found and the entire sequence consists of quartz-rich argillites and phylittes and schistose phyllites interbedded with wackes identical to those of the Pinnacle Formation. The tan weathering argillaceous metasediments are variable in quartz, chlorite, and white mica content but locally display excellent laminae characteristic of the Fairfield Pond Formation. These rocks have previously been referred to as Underhill Formation and that useage is followed in Figure 3. On the basis of lithic similarity and association, this belt of rocks is correlated with the Fairfield Pond-Pinnacle sequence to the west. Locally, dolomite horizons associated with slightly calcareous quartz wackes are present west of the Sweetsburg lithologies. This supports the view that White Brook lithologies are only sporadically present in this part of the Camels Hump Group. Importantly, the White Brook horizons along the east side of the Enosburg Falls-Fletcher anticline are associated with wacke characteristic of the Pinnacle Formation - a situation similar to the White Brook occurrences on the west side of the anticlinorium. Dennis (1964) interpreted black slate horizons near the Underhill-Pinnacle contact on the east side of the Enosburg Falls - Fletcher anticlinorium as White Brook. These black slate occurrences are similar to rocks mapped as Sweetsburg in this report.

East of the outcrop belt of Sweetsburg Formation, the Underhill consists of a heterogeneous sequence of quartz-albite-chlorite-muscovite /sericite schists and phyllites. Magnetite is locally abundant and modal proportions of the minerals are variable. Greenstone lithically and chemically similar to the amphibolitic and calcareous greenstones of the Tibbit Hill are locally interbedded. Pinnacle lithologies are not common but where present appear as marker horizons which are discontinuous along strike. Garnet is not recognized in the field in these rocks but several localities of abundant chloritic pseudomorphs after garnet are present.

The stratigraphy and structural setting of the rocks lying between the Enosburg Falls-Fletcher anticline and the Richford syncline is similar to the Mansville phase of the Oak Hill sequence first observed by Clark (1934, 1936) east of the Pinnacle Mountain anticline in the Sutton quadrangle of Quebec (Eakins, 1964; Clark and Eakins, 1968; Rickard, 1965). The Mansville phase includes rocks representative of the complete but structurally shortened or thinned Oak Hill sequence. East of the Sweetsburg outcrops defining the Richford syncline in Quebec, the Bonsecours Formation occupies a similar position as the correlative Underhill Formation rocks east of the Richford syncline in Vermont. M. Colpron (personal communication, 1987) reports that the Mansville phase involves a complex fold and fault history which juxtaposes stratigraphically older sequences against younger rocks from west to east. It appears that the geology east of the Enosburg Falls - Fletcher anticline in northern Vermont may have a similar origin to the Mansville phase rocks in Quebec.

Cover rocks to the Camels Hump Group

Cover rocks to the Camels Hump Group undergo significant facies changes across the Lamoille River transect. These changes are consistent with the different drift stage environments recorded in the cover rocks of the Camels Hump Group. In the Milton quadrangle the cover is "basal" quartzite of the Cheshire Quartzite overlain by the Dunham Dolomite representative of the Lower Cambrian platform. The Dunham is either overlain by Skeels Corner Slate or the Parker - Rugg Brook - Skeels Corners sequence (Stone and Dennis, 1964; Dorsey and others, 1983; Fig. 3). The grey to black laminated and carbonate-bearing slates are representative of shelf edge to slope deposits which further west correlate with the Monkton, Winooski, Danby and Clarendon Springs formations of the middle to upper Cambrian platform (Rodgers, 1968; Dorsey and others, 1983; Fig. 4).

Post-Fairfield Pond cover rocks are again found west of the West Fletcher fault (Stop 6) and in the Richford syncline where black to grey pyritiferous slates correlated with the Sweetsburg Formation are found (Fig. 3). These rocks are described below.

Sweetsburg Formation
(Cs; Stops 6, 9, 10)

Two separate outcrop belts of Sweetsburg Formation rocks are found in the transect. The western belt occurs west of the West Fletcher fault (Stop 6). Here, the Sweetsburg rocks display gradational contacts along the eastern side of the outctop belt with grey and rusty phyllites and laminated chloritic quartzose grits and phyllites of the Fairfield Pond Formation. The Sweetsburg Formation in this locality consists of interbedded rusty and non-rusty carbonaceous to grey pyritiferous black slates and phyllite interbedded with quartzite and thin beds of marble. Numerous fold hinges suggest that the isolated occurrence may occupy an infolded inlier overthrust by the Tibbit Hill sequence along the West Fletcher fault.

A second belt of Sweetsburg Formation rocks first mapped by Christman (1959) outcrops in a discontinuous northeasterly trend across the Jeffersonville quadrangle from Cambridge to the northern limit of the quadrangle east of Route 108 (Fig. 3). Dennis (1964) mapped similar rocks along the strike belt in the Enosburg Falls quadrangle to Richford, Vermont, which correlates with the Sweetsburg Formation mapped by Osberg (1965) and earlier by Clark (1934, 1936) in the Sutton quadrangle of Quebec. Osberg (1965) traced the Sweetsburg lithologies through a series of infolds into the serpentine belt of the Eastern Townships where it is referred to as the Ottaquechee Formation (Fig. 1). Although Canadian workers prefer to restrict the useage of "Sweetsburg" to rocks outcropping above the Dunham formation along the west flank of the Pinnacle Mountain anticline in Quebec, the writer prefers to extend the Sweetsburg to the rocks in the Richford syncline for the following reasons: 1. The black carbonaceous phyllites in the Richford syncline are interbedded with carbonate-bearing horizons as well as white to grey quartzites. Carbonate layers in the Ottauquechee are not common in central and southern Vermont where the Ottauquechee was first described (Perry, 1929); 2. The Ottauquechee Formation may undergo revision in definition with ongoing mapping underway in the Lincoln Mountain quadrangle (see Stanley and others, 1987).

The Sweetsburg lithologies mapped in the Richford syncline do however bear strong resemblance to the Ottauquechee as mapped in the serpentine belt of Quebec and northern Vermont (e.g., Doolan and others, 1982) and the correlation of Osberg (1965) between Sweetsburg and Ottauquechee is likely correct. Ongoing studies of these lithologies on both flanks of the Green Mountain anticlinorium will hopefully define a more detailed stratigraphy within these cover rocks. If the stratigraphic relationships of the Cambrian platform are correct (Mehrtens, 1987; Dorsey and others, 1983) the continental slope rise sequence represented by the Sweetsburg and Ottauquechee formations could span the entire Cambrian period.

The map pattern evolving for the Sweetsburg Formation in the Jeffersonville quadrangle is apparently controlled by early west northwesterly trending folds refolded by folds associated with the dominant foliation. Faults parallel to the dominant foliation are suspected by map pattern offsets and locally intense shear zones. Earlier faults may be present at least on the east side of the Sweetsburg outcrop belt as evidenced by isolated blocks of amphibolitic and calcareous greenstone along the contact (Rose, in progress). Contact relationships between the Sweetsburg Formation and the adjacent Underhill are at least locally conformable. For example in the East Fletcher area, the sequence of black slate and phyllites, rusty quartzose grits and phyllites,

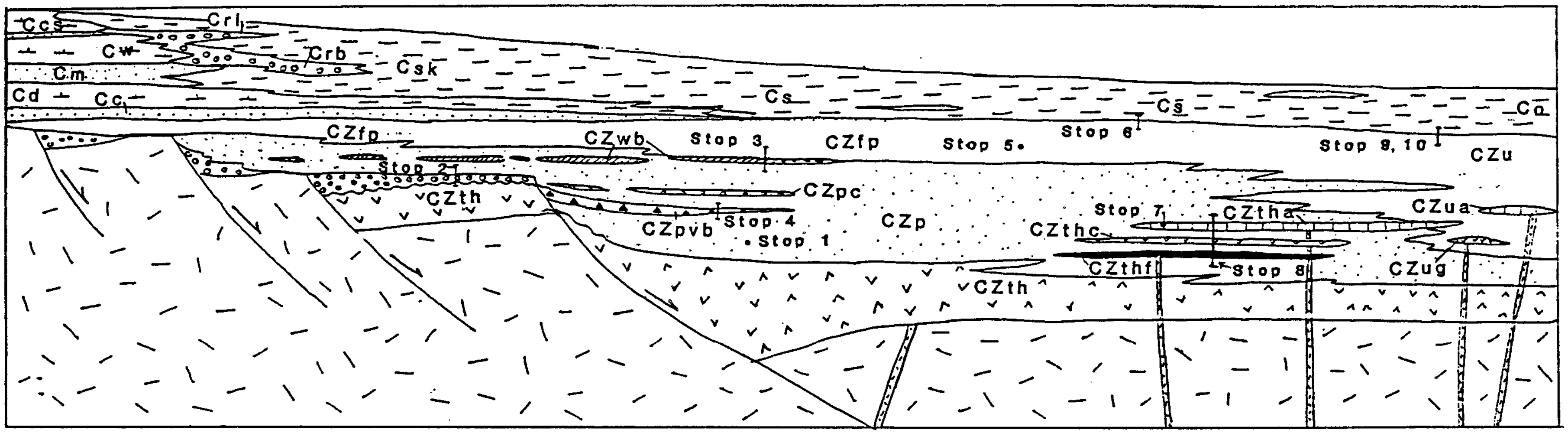


Figure 4. Interpretative restoration of the ancient rift to drift stage passive margin of North America in the vicinity of the Lamoille River transect. The Camels Hump Group rocks (Tibbit Hill, CZth; Pinnacle CZp; White Brook, CZwb; Fairfield Pond, CZfp; and Underhill, CZu) are considered to be wholly developed in a rift environment. Variations in the stratigraphy and thickness of the various units reflect the effects of syn-rift volcanism, erosion and deposition. These effects are profound near regions of exposed basement such as proposed for the Milton quadrangle and western part of the Gilson Mountain quadrangle. Cover rocks to the Camels Hump Group are either rift-drift transition rocks (Cheshire to Dunham) or drift stage platform, slope or rise deposits. The reconstruction suggests that the shelf-slope break, formed during the drift stage, occurred close to the basinward limit of rift stage erosion.

albitic schists and magnetite schists can be followed across a dome and basin geometry. Lack of adequate outcrop control combined with a very complex fold and fault deformation history and the heterogeneous nature of the Underhill Formation does not allow further elaboration of the original stratigraphic relationships to be made at this stage.

Hazens Notch Formation

Along the west flank of the Green Mountain anticlinorium in the eastern part of the Jeffersonville quadrangle, the Underhill is structurally intermixed with the rocks of the Hazens Notch Formation. The Hazens Notch which presently is included within the Camels Hump Group by Doll and others (1961), is not visited on this excursion. The palinspastic restoration of the position of the Hazens Notch Formation with respect to the other formations of the Camels Hump Group and associated cover is a problem of great importance to Vermont geology. The Hazens Notch is host to Vermont's only blueschist occurrences (Laird and Albee, 1975; 1981), as well as numerous serpentinite and talc bodies which record all the observed deformations of the host rocks. The Hazens Notch as presently exposed is likely not part of the stratigraphy of the passive margin formation involving rift to drift stage sedimentation. The blueschist/serpentine/greenstone associations with the Hazens Notch metasediments argues in favor of formation within a subduction melange formed during the closing stages of Iapetus. The lithologies present in the Hazens Notch suggest that most of the "North American" protolith for the melange involved Sweetsburg and Ottauquechee cover rocks as well as the underlying Camels Hump Group.

For further discussion of the Hazens Notch Formation the reader is referred to Thompson (1975) and Thompson and Thompson (1987).

DISCUSSION OF THE STRATIGRAPHY

The stratigraphy of the Camels Hump Group in northern Vermont is discussed in terms of the palinspastic reconstruction shown in Figure 4. The Camels Hump Group is interpreted in this reconstruction to represent clastic sedimentation and volcanism occurring synchronous with rifting of the Grenville basement prior to the development of a passive margin bordering the Iapetus Ocean. The age of rifting is unknown but based on regional relationships in the maritime and central Appalachians, a 600 - 620 m.y.b.p. age is inferred.

The oldest rocks known in the Camels Hump Group are the Tibbit Hill volcanic rocks as seen in the Pinnacle Mountain anticlinorium of Quebec and the Enosburg Falls anticlinorium of northernmost Vermont (Fig. 1). In the vicinity of the Lamoille River transect, these rocks are only exposed in the core of the Georgia Mountain anticline in the Milton quadrangle. The thickness of the basal flow of the Tibbit Hill is not known but could be considerable if it represents continental flood basalts formed on rapidly extending and subsiding continental crust.

The lowest member of the Tibbit Hill Formation (€Zthf) in the Fletcher anticline is considered to be somewhat or at best only slightly older than the youngest Tibbit Hill exposed in the Pinnacle Mountain - Enosburg Falls anticline to the north. This is based on the observation of interbedding of Pinnacle wacke and related clastic rocks with the Tibbit Hill of the Fletcher area which is not observed in Quebec, and the similarity of the volcanic rocks at the eroded top of the Tibbit Hill in the Georgia Mountain anticline with the lowest volcanic horizon of the Tibbit Hill in the Fletcher anticline.

Erosion of the flood basalts occurred in regions not extensively thinned or loaded during the early rift stage as evidenced by the Beaver Meadow conglomerate unconformably overlying the Tibbit Hill in the Georgia Mountain anticline (Stop 2). This unconformity is considered to have developed during a second rifting stage which supplied a variety of rounded clasts from Precambrian sources subsequent to Call Mill time in Quebec. This second rifting resulted in more rapid subsidence and extension to the east and perhaps coeval with flows seen today in the Tibbit Hill of the Fletcher anticline. Breccia units near the top of the Pinnacle Formation in the western part of the Gilson Mountain quadrangle may be coeval with the second rifting.

The transition from Pinnacle to the Fairfield Pond is interpreted to result from more distal source areas of clastics and a widening estuary capable of dolomite production (White Brook Formation). Locally, such as in areas located near clastic source terrains, dolomite production was impeded by continued influx of coarse-grained wacke horizons. The Fairfield Pond and Underhill Formation are considered to be correlative but the Pinnacle horizons found within the Underhill suggest that some of the finer grained protoliths of the Underhill may be coeval with coarse grained Pinnacle wacke to the west. The age of the volcanic horizons in the Underhill are considered to be slightly younger than those of the Tibbit Hill assuming that the axis of volcanism continues to migrate eastward with time in the axis of greatest lithospheric thinning.

The distribution of rock seen in the Camels Hump Group across the Lamoille River transect is interpreted to reflect the evolution of an extending and subsiding Grenvillian lithoshpere. The thickness of the lithosphere continued to provide sedimentologic controls on the distribution of cover rocks during the drift stage of passive margin development. Cheshire Quartzite and Dunham Dolomite were deposited on the filled rift basins located above slowly subsided and relatively thick lithosphere. These platform sediments may not have extended much across the palinspastic position of the Gilson Mountain quadrangle where subsidence due to loading and extension was presumably higher. During the drift stage, thermal subsidence controls dominated over the earlier rift-related subsidence causing the platform-slope transition to migrate considerably westward from the eastern limit of Dunham Dolomite (Fig. 4).

STRUCTURAL GEOLOGY

INTRODUCTION

The Camels Hump Group is polydeformed and metamorphosed. The deformation involves three periods of foliation development each of which is associated with folding. The map pattern structures are controlled largely by the second deformation structures including upright to overturned folds and steep faults subparallel to the second foliation which is usually the dominant foliation observed in outcrop. Kink folds postdate the development of the dominant foliation and appear more common where superposed on fold hinges related to the second generation folds.

Major structures along the transect as defined by previous workers include the Georgia Mountain anticline, the Dead Creek syncline, the Hinesburg thrust, the Fletcher anticline, the Richford syncline and the Green Mountain anticlinorium. This study has resulted in refining the structural detail to the regions east of the Hinesburg thrust and especially within the Fletcher anticline. Work to the east of Fletcher anticline is still in early stages of mapping, so structures in this area will not be discussed in detail.

As will be discussed more fully below, the writer wishes to retain the use of the Fletcher anticline as first proposed by Christman (1959) for the region exposing the Tibbit Hill Formation in the Gilson Mountain quadrangle. The lack of documented continuity between the Fletcher anticline with the Pinnacle Mountain - Enosburg Falls anticlinorium to the north combined with the unique Tibbit Hill stratigraphy recorded in rocks of the Fletcher anticline justifies that a distinction be made at this time.

The structural geology of the Camels Hump Group and cover rocks along the Lamoille River transect is discussed here in three parts. Part 1 discusses the geology west of the Fletcher anticline; Part 2 discusses the fold and fault structures within the Fletcher anticline; and Part 3 discusses the structural geology of the rocks within the region of the Richford syncline. A discussion of the structural evolution of the area in the context of regional geology concludes this section.

STRUCTURAL GEOLOGY EAST OF THE HINESBURG THRUST

Major structures in the Gilson Mountain quadrangle between the Hinesburg thrust and the Fletcher anticline include axial traces of anticlines and synclines which trend between N10E and N30E parallel to the dominant schistosity in the rocks. The Pinnacle Formation and overlying White Brook and Fairfield Pond formations in the northwestern quadrant of the quadrangle define four upright to overturned anticlines and associated synclines which generally plunge northerly. The axial traces of four anticlines are shown in Figure 3 based on observed bedding-cleavage relationships and fold hinges. From west to east, these anticlines are referred to as follows: Buck Hollow anticline; Buck Mountain anticline; Fairfax Falls anticline and Coombs Hill anticline. The Coombs Hill anticline is not well defined but is inferred by the mapped synclinal trace to the west and several east facing topping directions to the east of the axis. The lack of distinctive marker horizons in areas of good exposure make it difficult to further define the fold geometry in this region.

The fold pattern is predominantly controlled by an F2 fold event whose axial plane cleavage is the dominant schistosity (referred to as Sn) in the rocks. An early F1 fold event is inferred on the basis of an ubiquitous Sn-1 cleavage, numerous quartz veins folded by the dominant schistosity and a dome and basin map pattern believed to be derived by interference by the early fold events. A late fracture cleavage (Sn+1) is common in all areas of the quadrangle.

Fold geometry appears in the field and in map pattern to be asymmetric and west verging to the Buck Hollow and Buck Mountain anticlines. The Fairfax Falls anticline and the Coombs Hill anticline appear to be more upright but data is not as abundant compared with the more westerly situated anticlines. Wavelength from anticlinal crest to anticlinal crest is approximately 1200 meters; however, numerous smaller anticlines and synclines are noted on the limbs of the larger structures. Some of these smaller fold structures mapped around the Buck Mountain anticline have wavelengths of about 100 meters. Synclinal axes are not commonly observed because of the propensity of the overturned anticlinal limbs and hinges to form west facing cliff faces; shear zones observed along the overturned limbs parallel the axial surfaces of minor structures and the Sn foliation. Shearing thus appears to be a consequence of the F2 folding.

Structural relief does not appear to be great between the Hinesburg thrust and the Coombs Hill anticline; the stratigraphic units are repeated within the limits of lower Pinnacle to the upper(?) Fairfield Pond Formation. From the Coombs Hill anticline to the West Fletcher fault the structural relief increases bringing rocks as young as Sweetsburg in close proximity to Tibbit Hill volcanic rocks (Fig. 3). In the vicinity of the West Fletcher fault, the Sweetsburg-Fairfield Pond contact is characterized by a zone of rusty weathering grey to light green phyllites with intercalations of black slate. Shear zones are noted in several localities where the Sweetsburg-rusty phyllite contact is exposed. Bedding-cleavage relationships along the eastern contact of the Sweetsburg suggest that the overturned limb, an anticline, is sheared out along the contact. The Sweetsburg outcrop belt along the west side of the fault is interpreted to be a synclinal infold modified by subsequent faulting contemperaneous with the development of the dominant foliation.

STRUCTURAL GEOLOGY IN THE FLETCHER ANTICLINE

The Fletcher anticline is an elongate NE trending structure underlain by the Tibbit Hill Formation along the Lamoille River transect. It extends across the full north-south extent of the Gilson Mountain quadrangle northward into the southern end of the Enosburg Falls quadrangle mapped by Dennis (1964). The southern limit is not known but mapping by Christman (1959) suggests that the structure continues at least to Jericho, Vermont in the southwestern corner of the Mount Mansfield 15' quadrangle. Christman and Secor (1961) report exposures of feldspathic, calcareous and amphibolitic greenstone in the Richmond and Huntington area in the southwest corner of the Camels Hump quadrangle. These occurrences suggest that the rocks found within the Fletcher anticline in the Gilson Mountain quadrangle may extend as far south as the Lincoln Mountain quadrangle (Fig. 1).

The "anticline" nature is based on the supposition that the rocks are stratigraphically older than the Pinnacle and Fairfield Pond formations observed on either side of the Tibbit Hill Formation. The structure of the Fletcher anticline is far more complicated than the name implies and is more properly referred to as a nappe involving at least three periods of deformation. The pre-Sn deformational history associated with the Fletcher anticline appears to be more complicated and pervasive than the deformation in rocks to the west. This suggests that the Tibbit Hill rocks of the Fletcher anticline have been transported westward during pre-Sn deformation time relative to the Camels Hump Group rocks and cover situated in the western part of the Gilson Mountain quadrangle.

The structural complexities of the nappe can in part be resolved because of excellent exposure and a detailed volcanic stratigraphy discussed previously. Restoration of the folded stratigraphy is hampered, however, by a lack of fabric data (especially pre-Sn fold data) associated with the early deformation and the intense transposition of the earlier structures by the folding and faulting during the development of the Sn cleavage.

A fold generation model is schematically shown in Figure 5 based on the present map pattern. The model is based on the assumption that the amphibolitic greenstone constitutes a single horizon and that the syn- to post-Sn faults which transect the structure are relatively minor compared to the pre-Sn structure associated with the nappe formation and emplacement. The actual fold pattern is probably more complex than the model which is simplified by showing all fold generations as coaxial. The model, however, shows that the dominant foliation imposed on the Fletcher anticline is superposed on an already tightly folded nappe structure. As shown on Figure 5 C,D, the Fletcher nappe/anticline is interpreted as a "pop-up" structure removed from its root during the backfolding/backthrusting stage. This interpretation would resolve the problem of tracing the Tibbit Hill rocks exposed in the Fletcher anticline to the north or south. Critical areas to further test the model are in the western part of the Jeffersonville quadrangle and the southern part of the Enosburg Falls 15' quadrangle. Neither area has yet been subjected to detailed mapping.

The Fletcher anticline appears to be fault bounded and imbricated. Five faults are shown in Figure 3 with a spacing of about 1300 meters. From west to east these faults are designated: the West Fletcher fault (WFF); the Halfmoon Pond fault (HPF); the Stones Creek fault (SCF); the Metcalf Pond fault (MPF); and the Armstrong Mountain fault (AMF). The WFF is suspected on the basis of noted shearing and intense shortening of the Fairfield Pond Formation to the west of the fault (see section on stratigraphy) and the abrupt stratigraphic juxtapostion of Tibbit Hill lithologies to the east and Fairfield Pond - Sweetsburg lithologies on the west. In the southern limit of the area, greenstone similar to the calcareous greenstone at West Fletcher occurs in close proximity to White Brook dolomite breccia; this is the only basis for the extension of the West Fletcher fault southward.

The HPF is defined by sharp map pattern truncation of all the members of the Tibbit Hill Formation along the projected trace of the fault. The fault is poorly constrained in the region east of Gilson Mountain. The fault can be directly observed just north of the Lamoille River 800 meters ESE of Sand Hill by truncation of the amphibolitic greenstone member of the Tibbit Hill Formation (Fig. 3).

The MPF is defined by zones of intense shear along its trace, isolated pods or slivers of amphibolite along the contact and by truncation of the Tibbit Hill Formation rocks. Isolated occurrences of black slate of unknown origin are found east of the fault east of Metcalf Pond.

The AMF is only suspected to be a fault of any significance. Intense shear zones which separate the amphibolitic greenstone and magnetite-bearing schists of the Underhill Formation are observed on the northeast side of Armstrong Mountain in the westernmost part of the Jeffersonville quadrangle. The same boundary cannot be documented as a fault surface along the full extent of the Tibbit Hill Formation in this area.

The SCF is a suspected fault, based on the map pattern of the Tibbit Hill on each side (Fig. 3). Along most of its length in the Stones Creek valley, outcrop control is poor. The best evidence for the fault is an isolated occurrence of amphibolitic greenstone west of Metcalf Pond and shearing out of tight isoclinal folds east of Hedgehog Hill north of the Lamoille River (Fig. 3).

STRUCTURAL GEOLOGY EAST OF THE FLETCHER ANTICLINE

Our knowledge of the geology east of the Fletcher anticline is based on the work presently underway by Mock in the East Fletcher area of the Richford syncline, the work of Thompson (1975) in the northeastern part of the Jeffersonville quadrangle and contiguous areas to the north, and several field camp projects completed by University of Vermont undergraduates. In addition, Hugh Rose and Greg Koop conducted Senior Research studies at the University of Vermont in the areas north of Jeffersonville. Rose mapped the relationships of the metasediments and greenstone of the Underhill Formation and the Sweetsburg Formation east of Route 108 and Koop mapped the Sweetsburg - Underhill - Pinnacle sequence west of Route 108 and ESE of North Cambridge (Fig. 3).

These studies have concluded the following with regard to the structure of this area:

1. The orientation of the dominant cleavage (Sn) systematically changes from steep easterly dips east of the Fletcher anticline, to vertical dips in the Richford syncline to west dips east of the Richford syncline (Table 1). Thompson (1975) reports an average dip of 45 degrees to the east for Sn with progressive shallowing (83 to 28 degrees) in approaching the Green Mountain axis.

2. Rose has documented several faults with west over east sense of motion and orientation parallel with the dominant schistosity of the area. Mock also has mapped steep shear zones within the Richford syncline parallel to the dominant foliation of the area.

3. Slivers of amphibolite and wacke identical to the rocks mapped to the east as Tibbit Hill and Pinnacle, respectively, are found by both Koop and Rose which predate the formation of the dominant foliation. The faults are tentatively included with the first deformation with east over west sense of motion based on the location of the stratigraphically older slivers east of stratigraphically younger rocks along the fault (Fig. 3).

4. The Hazens Notch Formation structurally underlies the Underhill Formation as a result of eastward vergent folding related to the west dipping dominant foliation (Thompson, 1975).

5. The Sweetsburg Formation map pattern mapped by Rose and Mock includes dome and basin and hook structures supporting the view that the structure involves the interference of at least two periods of deformation involving folding.

6. The Sn+1 fracture cleavage, referred to as the Green Mountain cleavage because of its regional association with the Green Mountain anticlinorium, is present throughout the region east of the Fletcher anticline but appears not to play a strong control in the map pattern.

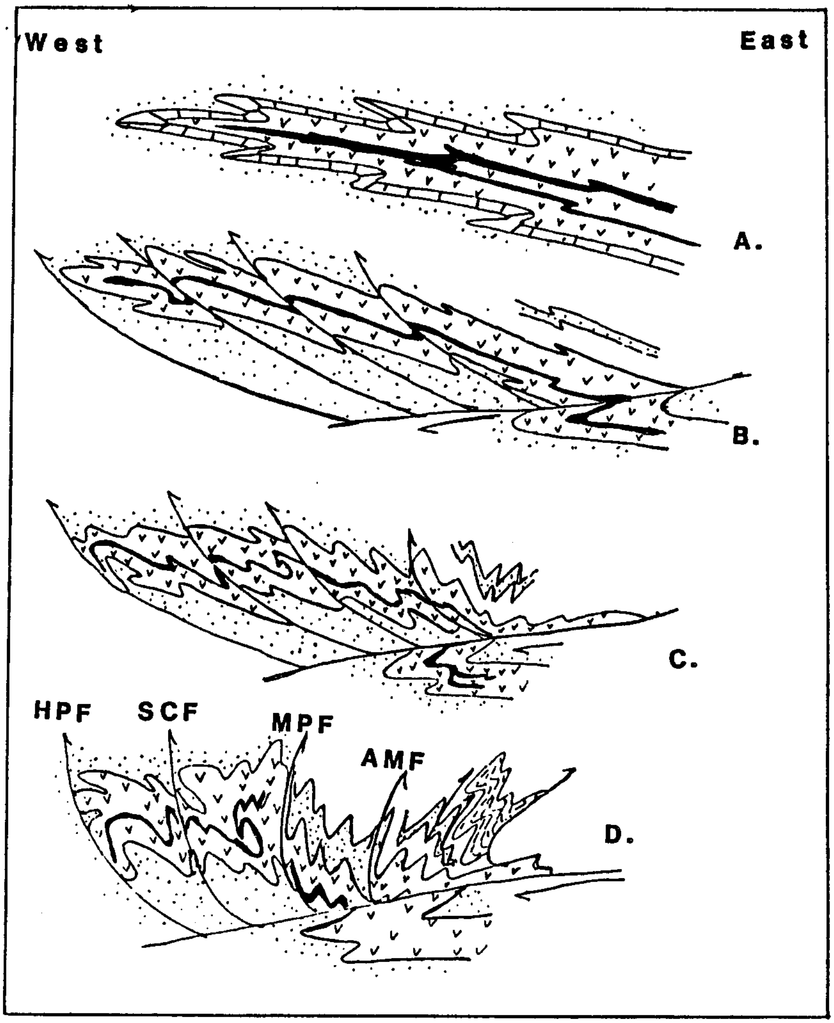


Figure 5. The evolution of the Fletcher anticline involves an early stage of tight to isoclinal folding not recognized in the map pattern of more westerly rocks. This is represented in A by the emplacement of a nappe during Sn-1 cleavage development. The refolding of the nappe is interpreted as a combination of west verging folds and faults which steepen during a back folding stage (B, C, D). The lack of volcanic rocks east of the Fletcher anticline to the Richford syncline suggests that the rocks in this vicinity represent the upper limb cover rocks to the Tibbit Hill Formation. Unpublished Vibroseis profiling along the western part of this schematic cross-section support the flat lying nature of the Fletcher anticline and the west dipping fault "delaminating" the previously emplaced nappe. The folding dash pattern in Figure 5D represents the Sweetsburg lithologies of the Richford syncline (Fig. 3B). Figure 5B, C, D are all considered to be synchronous to the development of the dominant foliation, Sn. HPF: Halfmoon Pond fault; SCF: Stones Creek fault; MPF Metcalf Pond fault; AMF: Armstrong Mountain fault. See text and Figure 6 for further elaboration.

TABLE 1. REPRESENTATIVE Sn DATA ACROSS THE LAMOILLE RIVER TRANSECT

Georgia Mtn.	Gilson Mtn. Quadrangle		Jeffersonville Quadrangle		
	west	east	west	central	east
N10E,80E	N26E,80E	N19E,85E	N24E,90	N9E,64W	N18E,45W
Carter (1979)	This paper				Thompson (1975)

DISCUSSION

The stratigraphy and structure of the Lamoille River transect is briefly discussed here with the use of a model shown in Figure 6 which summarizes our present understanding of the geology. The metamorphic history of the area, while not discussed in this report, is not incompatible with the model and indeed may provide a quantitative test in the future for the P,T trajectories suggested for various segments of the transect. The model for the hinterland evolution of the Camels Hump Group and cover rocks is similar to the delamination model recently proposed for the southeastern Canadian Cordillera by Price (1986).

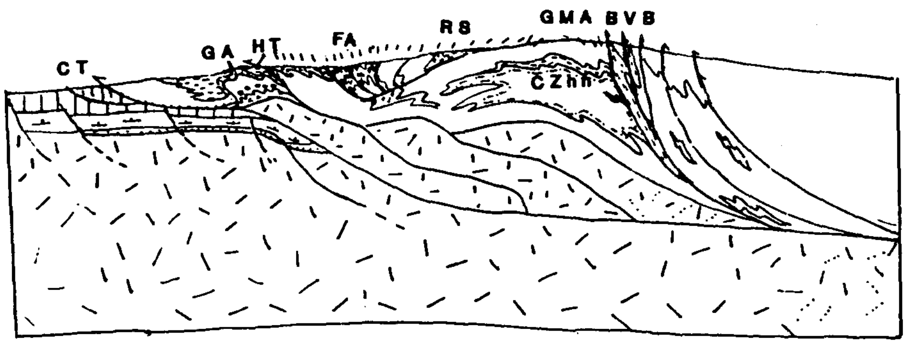
The palinspastic reconstruction for the Camels Hump Group (Fig. 4) is incorporated into the reconstructed passive margin undergoing subduction in the late Cambrian to early Ordovician time (Fig. 6A). The obduction of oceanic lithosphere occurred at this stage. The Thetford - Asbestos ophiolite belt of Quebec represents the emplacement at high structural levels and the Eden Mills ophiolite aureole in Vermont was emplaced at mid to deep crustal levels (Doolan and others, 1982). The high pressure blueschist assemblages reported by Laird and Albee (1981) also originated at this stage. Cover rocks to the Camels Hump Group (Ottauquechee Formation) are subducted beneath the ophiolite and mix with underlying Camels Hump Group and ophiolitic slivers to form the Hazens Notch Formation.

Continued subduction-related compression in the early to mid Ordovician results in tectonic thickening of the supracrustal rocks (Fig. 6B). Tectonic wedging of deep crustal oceanic lithosphere aids in delaminating the supracrustal rocks causing the North American margin rocks to be directed eastward over the oceanic lithosphere. The Ottauquechee Formation is shown as an example of a high crustal level cover rock to the Camels Hump Group being directed eastward over Hazens Notch and earlier emplaced lithosphere. At deeper and more cratonward positions in the orogen, east over west directed faults and nappes represent the first phase deformations observed in the Lamoille River transect. Tectonic thickening in the outboard regions of the Camels Hump Group results in pervasive and maximum metamorphic recrystallization of Taconic age. The Richford syncline is schematically shown as the region representing the cover rocks for the westerly directed Fletcher nappe overridden to the east by the more thoroughly recrystallized rocks of the Underhill Formation. The Fletcher nappe overrides the cover rocks of the Camels Hump Group as evidenced by the Sweetsburg Formation rocks observed west of the West Fletcher fault (Stop 6).

Basement slivers detached from the craton shown in Figure 6C are a consequence of continued subduction of the North American margin below the developing accretionary prism. These slivers aid in bringing deep crustal rocks to more shallow levels and to serve as tectonic wedges for further delamination and backfoldfng of rocks situated at higher structural levels in the crust. The second and most pervasive deformation of the rocks west of the Richford syncline occurred at this stage. The Fletcher nappe is backfolded along with cover rocks resulting in a steep structurally thinned "Mansville Phase" fold/fault sequence to the east of the Fletcher anticline. The Stowe Formation, considered to be the most distal equivalent of the Camels Hump Group (Coish and others, 1985), and metamorphosed at deep crustal levels, is thrust westwards onto the backfolded Ottauquechee-Hazens Notch and ocean remnants. The uplifted Stowe consequently served as a source area for at least part of the unconformably overlying Moretown Formation locally marked by the Umbrella Hill Conglomerate (Figs. 1, 6C; Doolan and others, 1982).

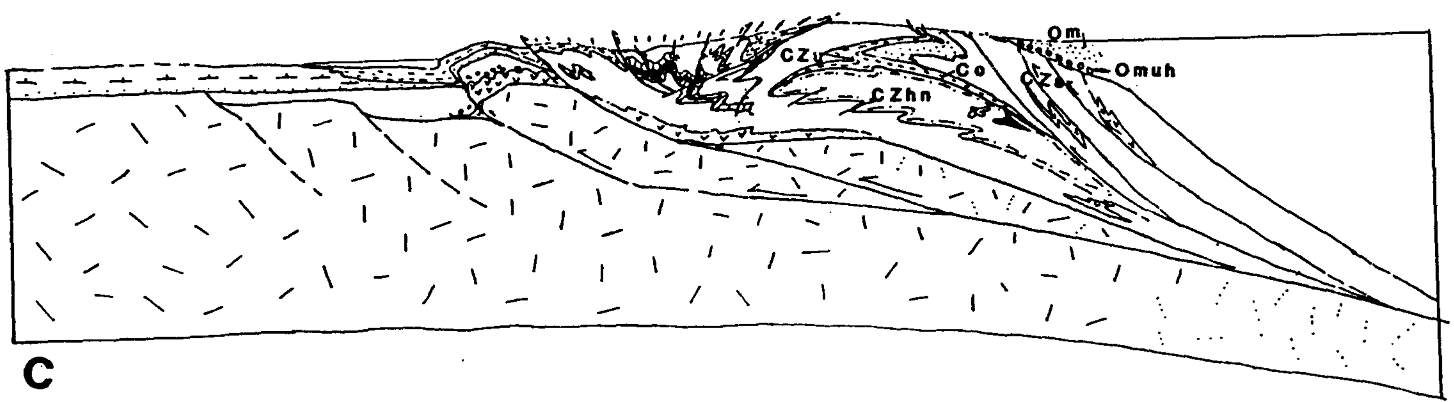
The stages represented by Figure 6A, B and C are subduction related processes associated with accretion tectonics in a subduction complex. Allochthon emplacement in the Taconics and Quebec occurred during these stages. In Quebec the allochthons involved rocks which were predominately cover rocks to the Camels Hump Group (e.g., Stanbridge Nappe, Levis Nappe) (St. Julien and Hubert, 1975). The Charney/Granby/Armagh/Caldwell rocks may also be equivalent to the cover rocks in age (post lower Cambrian) but their original stratigraphic position with respect to the Camels Hump Group and the correlative lower Oak Hill Group of Quebec is not established. The lower allochthon slices of the Taconic Mountains include stratigraphic correlatives of both the Camels Hump Group and cover rocks in the Lamoille River transect. The higher Taconic slices are composed of correlatives of the lower parts of the Camels Hump Group and basement. The accretion tectonics of the Quebec Appalachians in terms of both the obduction phase and allochthon development represents considerably higher structural levels than is observed in the central and southern part of the New England Appalachians. The Lamoille River transect represents a transitional level of crustal involvement between the Quebec and southern Vermont Appalachians.

The Taconic orogeny in the northern Vermont section can be explained by subduction related accretionary tectonics. The Moretown and St. Daniel basins formed outboard of the forearc region created by the subduction zone complex. Collision stage tectonics involving active island arcs and/or already accreted terrains outboard of the interarc basin strongly modified the Taconian structures especially in promontory regions of the Grenvillian margin. Figure 6D is a schematic cross section of the northern Vermont orogen at the conclusion of the collision stage. With regard to features seen on this excursion, this stage was responsible for the formation of the Sn+1 cleavage, the Green Mountain anticlinorium



D

Figure 6. The cartoon reconstruction of deformational events of the Taconic Orogen is schematic and not to scale. The degree of shortening shown and the relative thickness of the units involved is drawn within the constraints of clarity and space. Lithic designators are previously defined and/or shown in Figure 1. Although the reader should refer to the text for discussion of this figure, the diachronous nature of deformation and the important influence of west over east shortening by folds and faults in the upper part of the crust should be noted. Abbreviations in Figure 6D: CT=Champlain thrust; GA=Georgia Mountain anticline; HT=Hinesburg thrust; FA=Fletcher anticline; RS=Richford syncline; GMA=Green Mountain anticlinorium: BVB=Baie Verte-Brompton line.



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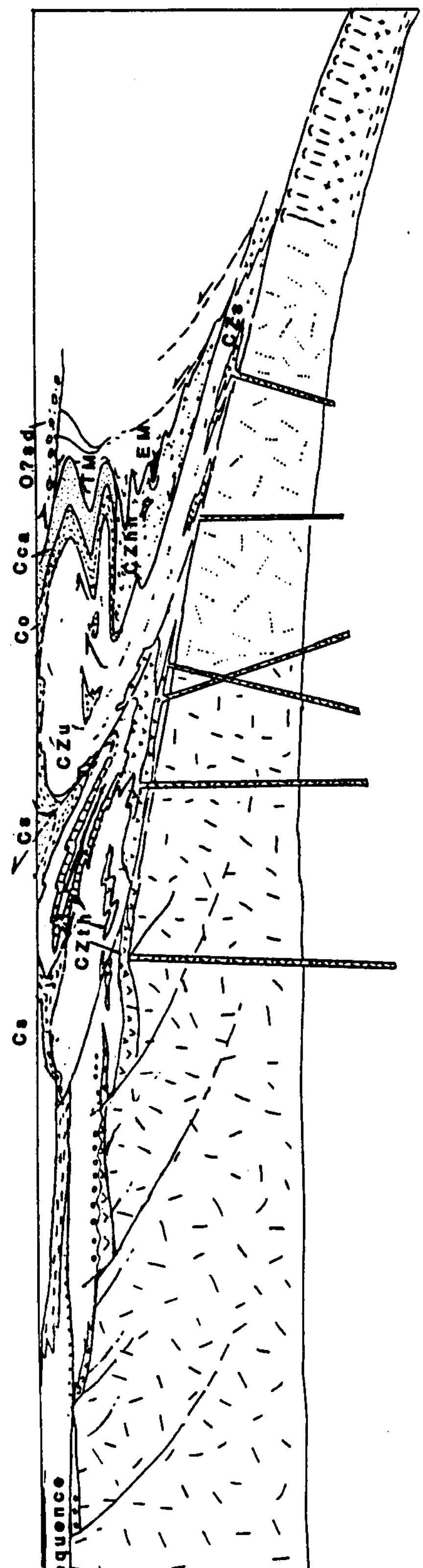
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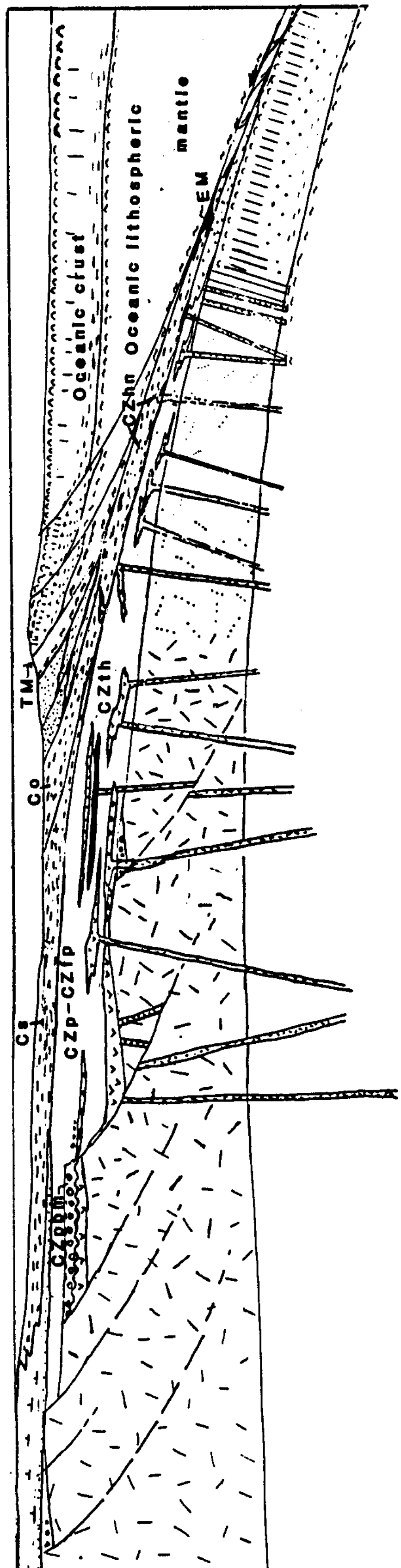
B


Cs
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CZp-CZfp
Cs
Co
TM
CZth
CZhn
Oceanic crust
Oceanic lithospheric
mantle
EM

A

and further westward imbrication of the foreland region. The possibility exists that much of the post-Moretown deformation in northern Vermont is associated with the Acadian deformation which strongly modified Taconian geology along the Vermont - Quebec serpentine belt (Baie Verte - Brompton zone of Figure 6D).

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ROAD LOG

The trip log begins at a commuter parking lot on the south side of Route 15 in Cambridge, Vermont. Lunch can be purchased at one of two village markets in Cambridge because the trip does not pass near stores at lunchtime. Consolidate participants into the smallest number of vehicles since several of the stops require parking on private property.

PLEASE NOTE: BE SURE TO OBTAIN PERMISSION FROM LANDOWNERS TO TRESPASS ON THEIR LAND.

0.0 Leave commuter parking lot and proceed west (left) on Route 15.

0.3 Intersection of Route 15 and Route 104. Bear right on Route 104.

0.6 Hills to the north (right) are underlain in part by massive amphibolitic greenstone of the Tibbit Hill Formation. They mark the eastern limit of the Tibbit Hill Formation in this area (Fig. 3).

2.4 Road intersection. Proceed straight on Route 104 west.

3.0 Hill to left underlain by feldspathic and amphibolitic greenstone members of the Tibbit Hill Formation. The Fletcher anticline crosses the river near this point (Fig. 3).

4.5 Road intersection on left; proceed straight on Route 104 west.

5.5 Road intersection on left; continue west on Route 104.

5.8 Pull off on the north side of Route 104 just past the intersection with road leading over the Lamoille River.

Stop 1. Fairfax Falls Power Station - Lower Pinnacle Formation

Excellent examples of the lower part of the Pinnacle Formation as seen in the Gilson Mountain quadrangle outcrop on both sides of the road and on both sides of the river below the dam. We will confine our stop to the roadcuts. Topping directions and bedding/cleavage relationships suggest this stop is near or at the axis of an upright anticlinal hinge referred to as the Fairfax Falls anticline on Figure 3. The massive chloritic wacke contains abundant quartz-feldspar detritus. Thin sections show the development of stilpnomelane. Numerous north dipping kink bands strike both in east northeasterly and west northwesterly directions. The dominant foliation strikes about N20E and dips steeply to the east. Sn+1 have similar strikes but dip steeply to the west. Large clasts are rarely found; however, a coarse conglomeratic facies occurs as a rather consistent horizon to the east of the mapped trace of the Fairfax Falls anticline (see Fig. 3).

5.8 Continue west on Route 104.

6.4 Entering the Milton quadrangle.

6.7 Junction with Route 128; bear to right on Route 104.

7.4 Cross the Lamoille River and enter the village of Fairfax.

7.55 Blinking yellow light; bear left on Route 104 west.

9.1 Outcrop on the north side of the road, just past road intersection is typical example of the Fairfield Pond Formation.

9.15 Intersection of Route 104A on left; bear right and continue on Route 104.

Outcrops of quartz-rich dolomite of the upper Dunham Dolomite and the Cheshire Quartzite on the east side of Route 104 as you climb the hill. The Hinesburg thrust juxtaposes rocks of the Fairfield Pond Formation against these lithologies (Carter, 1979). The Dunham occurrences along the fault mark the axis of the Dead Creek syncline of Booth (1959) and were interpreted as a syncline on the Centennial Map (Doll and others, 1961).

10.1 To the west are hills underlain by the Pinnacle Formation coring the Georgia Mountain anticline.

12.0 Road intersection on west; continue straight.

12.1 Road intersection on east; continue straight.

12.6 Road intersection on west. This road leads to Beaver Meadow, the type locality of the conglomerate to be visited at Stop 2; continue straight.

13.0 Farm on west side of road; pull into the driveway and ask permission at farmhouse.

Stop 2. Beaver Meadow Conglomerate

Walk south to the outcrops on the west side of Route 104. Exposures of medium to coarse grained wacke of the "upper" Pinnacle are exposed adjacent to the road to the south. These rocks contain pods of buff weathering dolomite which characterize the upper Pinnacle in the Oak Hill Group of Quebec (W. Dowling, personal communication, 1987). To the west the wacke is interbedded with several horizons of conglomerate composed of slate and previously deformed granitic gneiss and perthite (McBean, 1979). The igneous boulders are well rounded and range in size from 5 to 40 cm. The gneissic foliation of the clasts is randomly oriented with respect to the schistosity of the matrix. Besides the slate, sedimentary clasts include wacke and arkosic sandstone. The slate fragments are more abundant to the west where the boulder conglomerate is in sharp contact with the fine grained dark green rocks of the Tibbit Hill Formation. The hematiferous slate fragments are lithically similar to the Call Mill slate which stratigraphically overlies the Tibbit Hill in Quebec (W. Dowling, personal communication). This and other evidence is discussed in proposing that the Beaver Meadow Conglomerate marks an unconformity over the Tibbit Hill Formation. See Figure 7 for outcrop map.

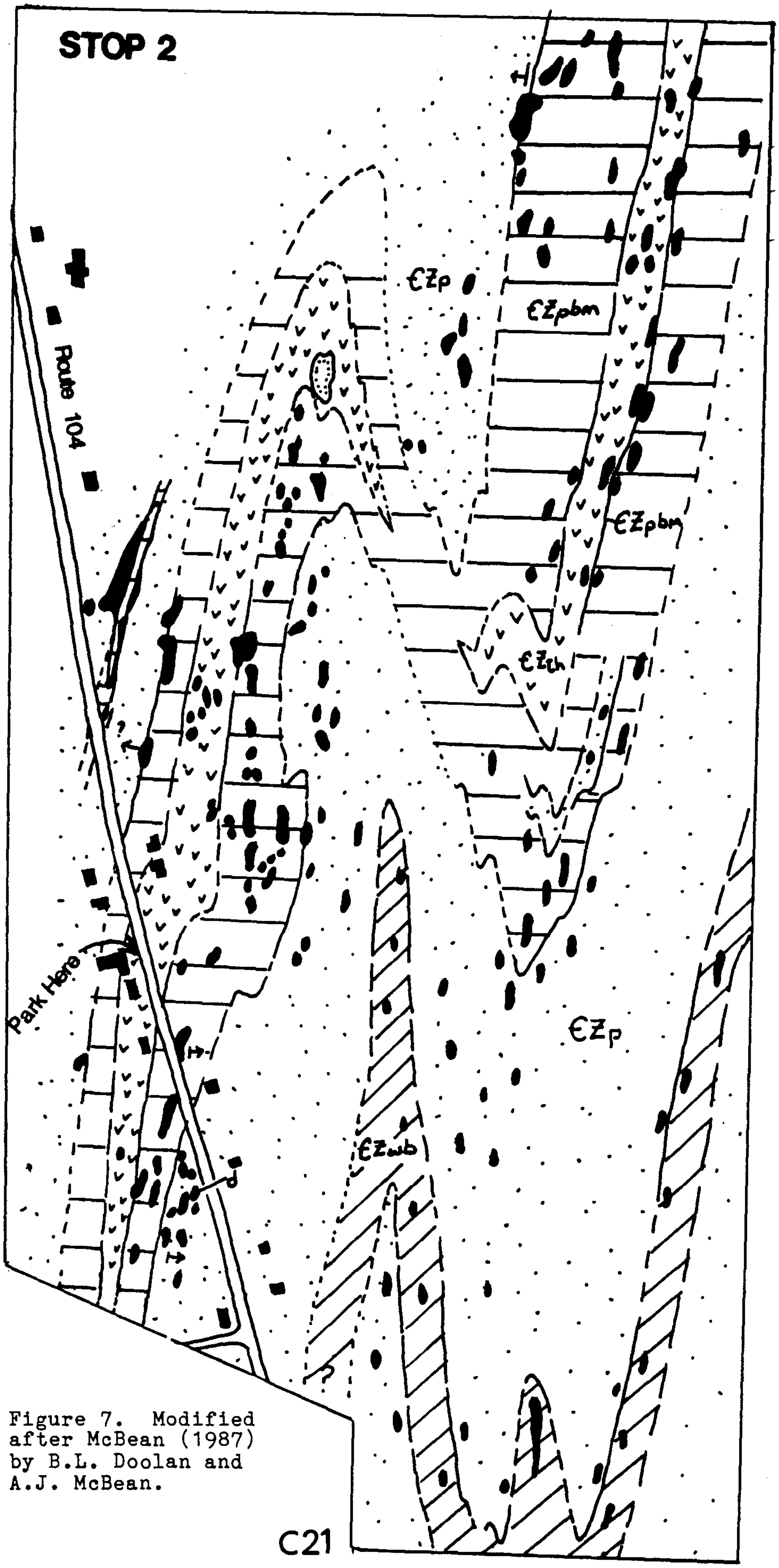

Figure 7. Modified after McBean (1987) by B.L. Doolan and A.J. McBean.

13.0 Proceed southerly on Route 104 to the unnamed dirt road which is the first left turn.

13.8 Intersection of road to west; turn left onto this unnamed dirt road.

14.4 T-intersection with another unnamed dirt road; excellent exposures of the White Brook dolomite outcrop in low ridges on the west side of the road to the north; we will proceed to the south (right turn).

15.3 Y-intersection close to the contact of the Hinesburg thrust as mapped by Carter (1979); bear left at the intersection.

15.6 Hill in front of you is Buck Hill underlain by Pinnacle Formation rocks in the Gilson Mountain quadrangle.

16.4 Crossroads. Proceed north (left) on paved road known as Buck Hollow Road (unmarked). We have now entered the Gilson Mountain quadrangle.

16.7 Farm on the east side of the road. Pull into driveway and ask permission at the farm house. Proceed east along path in the back of the cow barn to the outcrops in the pasture.

Stop 3. Pinnacle and Fairfield Pond contact.

This stop examines the contact relationships between the upper Pinnacle and the lower Fairfield Pond formations. The contact is not typical in that dolomite occurs along the contact at this locality. Such occurrences are not common in the Gilson Mountain quadrangle. The uppermost Pinnacle is a quartz pebble conglomerate which locally is highly calcareous. Proceed southward, along the ridge to outcrops of the argillaceous quartzites and thinly laminated argillites of the Fairfield Pond Formation. Contact relationships with the quartz pebble conglomerate will be seen. The structure and outcrop distribution suggests a basin or dome structure cored by the younger Fairfield Pond Formation. If time permits, cross Polly Brook to the east to a synclinal hinge in the thinly laminated Fairfield Pond Formation outcropping east of the brook. The syncline separates the two ridges displaying the Fairfield Pond/Pinnacle contact. The Fairfield Pond is discussed in terms of its correlatives in Quebec (see Fig. 2). See Figure 8 for outcrop map.

16.7 Proceed southward on the Buck Hollow Road to the crossroads.

17.1 At the crossroads, turn west (left) onto unmarked dirt road. Cliff faces on the west side of Buck Mountain ahead are of coarse facies of the upper Pinnacle Formation.

17.8 Road intersections; continue straight.

18.1 Pull off road at mobile home on east side of road and ask permission at the mobil home. Park and observe glacially polished pavement outcrop kindly stripped of cover by the landowner. **NO HAMMERS AT THIS OUTCROP.** Excellent hammer exposures are located just off the west side of the road.

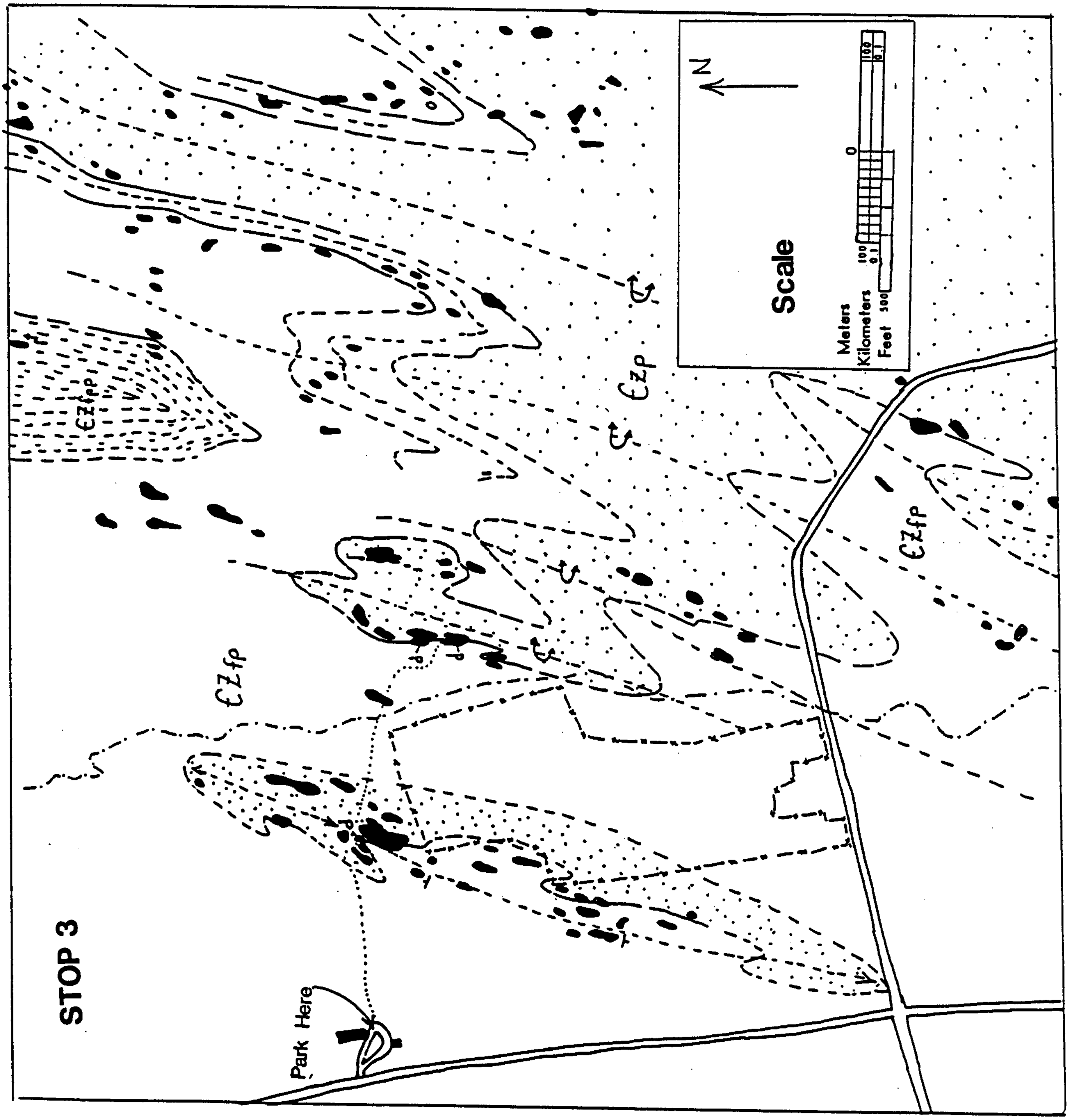

Figure 8. Geology by B.L. Doolan.

Stop 4. Volcanic breccia of the Pinnacle Formation

This unit (of the Pinnacle Formation (CZpvb on Figs. 2, 3, 4)) is only found in a single horizon approximately 60 meters thick which is folded into a series of north plunging synclines and anticlines on the east side of the Buck Hollow anticline. The dominant foliation dips steeply eastward with a strike of N35E40E. The rock breccia appears homogenous both in composition and size of the clasts; however, some variation is seen in the larger exposures on the west side of the road. The matrix also appears to be volcanogenic in origin and is tentatively interpreted as a tuff breccia. On the west side of the road a thin horizon of tuff(?) without clasts is seen on the western contact of the breccia but this horizon is not continuous along strike. The Pinnacle wacke exposed near the contact is more chloritic than the wackes found near the top of the formation. The position of the breccia in map pattern suggests that it does not occur far from the top of the Pinnacle Formation. The relationship between this volcanic breccia with the unconformity overlying the Tibbit Hill Formation at Stop 2 is discussed.

18.1 Continue south on the dirt road to Huntville.

Outcrops on the east side of the road are massive wackes and interbedded argillite of the Pinnacle Formation.

19.3 Crossroads mark the location of Huntville. Turn onto the dirt road heading to the east (left).

20.4 T-intersection. Proceed north (left) on dirt road.

20.9 Ask permission at the mobile home on the west side of road. Again, a beautifully exposed and clean outcrop provided by the landowner. Park in the driveway and proceed to the outcrop north of the mobile home.

Stop 5. Fairfield Pond Formation

This stop is of the Fairfield Pond Formation as mapped in most of the Gilson Mountain quadrangle. Numerous minor structures are preserved by the well bedded nature of this formation. Fold axes of several generations plunge predominately northward with west over east rotation. This suggests that the outcrop lies west of the synclinal axis separating the Buck Mountain and Fairfax Falls anticlines. Fold axes vary between N5W and N45E and plunge shallowly northward. The dominant foliation dips moderately to the east and strikes N30E45E. Sn+1 is steeper and more northerly. NNW trending kink bands dip northerly.

21.0 Proceed north on the dirt road.

21.4 Road intersection. Proceed east (right) onto paved road.

22.3 Approximate location of the trace of the Coombs Hill anticline.

22.5 Road intersection; proceed straight ahead on paved road.

22.8 Road intersection. This is West Fletcher! To seek permission for Stop 6, turn left and stop at first house on left side of road. Return to intersection at 22.8 miles. Proceed straight, up the hill.

23.1 Road intersection; bear left on the paved road.

23.3 Private road which is easily missed on the north (left) side of road. Turn left just past the tennis court.

23.5 Stop just short of metal gate on farm road; back up and park in clearing along the side of road and walk to the gate.

Stop 6. West Fletcher fault

Purpose of this stop is to examine the rocks on either side of the West Fletcher fault. Just to the north of the iron gate on the east side of the road are excellent exposures of argillite and wacke exposed in a fold hinge. The rocks are interbedded with a horizon of calcareous greenstone which is found about 200 feet north. Continue up the road about 1300 feet to Y-intersection with smaller woods road. Bear left and walk about 450 feet to a small brook. Proceed downstream to excellent rock exposures. The 500' traverse crosses rocks of the Fairfield Pond Formation (85') into tan weathering rusty phyllites (100') which have a gradational contact with black carbonaceous phyllites and quartzites typical of the Sweetsburg Formation to be visited in Stops 9 and 10 (280'). To the west, rusty phyllites (80') grade into the black phyllites. Along strike the rusty and tan phyllites appear to be in bedded contact with the coarse pebbly wackes of the Pinnacle Formation to the west. Use the outcrop map of Stop 6 (Fig. 9) as basis of discussion of alternative interpretations of the black slate lithologies and their contact relationships with adjacent rocks.

23.5 Return to the paved road.

23.8 At the intersection with paved road, proceed east (left).

25.3 Intersection with dirt road on north; proceed north (left) on the dirt road.

26.6 At the farm on the east side of the road south of the road intersection, pull to the side of the road and seek permission at the farmhouse.

Stop 7. Amphibolitic greenstone member of Tibbit Hill

Pavement outcrops on the west side of the road of amphibolitic greenstone member of the Tibbit Hill Formation (CZtha). Walk about 1400' west across the pasture to the sugar house on the hill. Large ledge to the south of the sugar house displays three dimensional exposures of structures interpreted to be columnar jointing. Such features have not been reported in the rift volcanics of the northern Appalachians. Alternative explanations welcome! See Figure 10 for sample locations of Tibbit Hill and Underhill greenstones, and Tables 2 and 3 for analytic data (Rose, 1986).

26.6 Proceed north on the dirt road.

27.9 Road bends sharply to right (east).

28.1 T-intersection. Turn south (right) on dirt road.

Large outcrops on east side of road are typical massive exposures of the amphibolitic greenstone member of the Tibbit Hill Formation.

28.7 Cross small brook.

28.8 Turn left at mailbox and proceed uphill along private drive. Seek permission at house at top of drive. Proceed east from clearing at the end of the driveway to the trailhead.

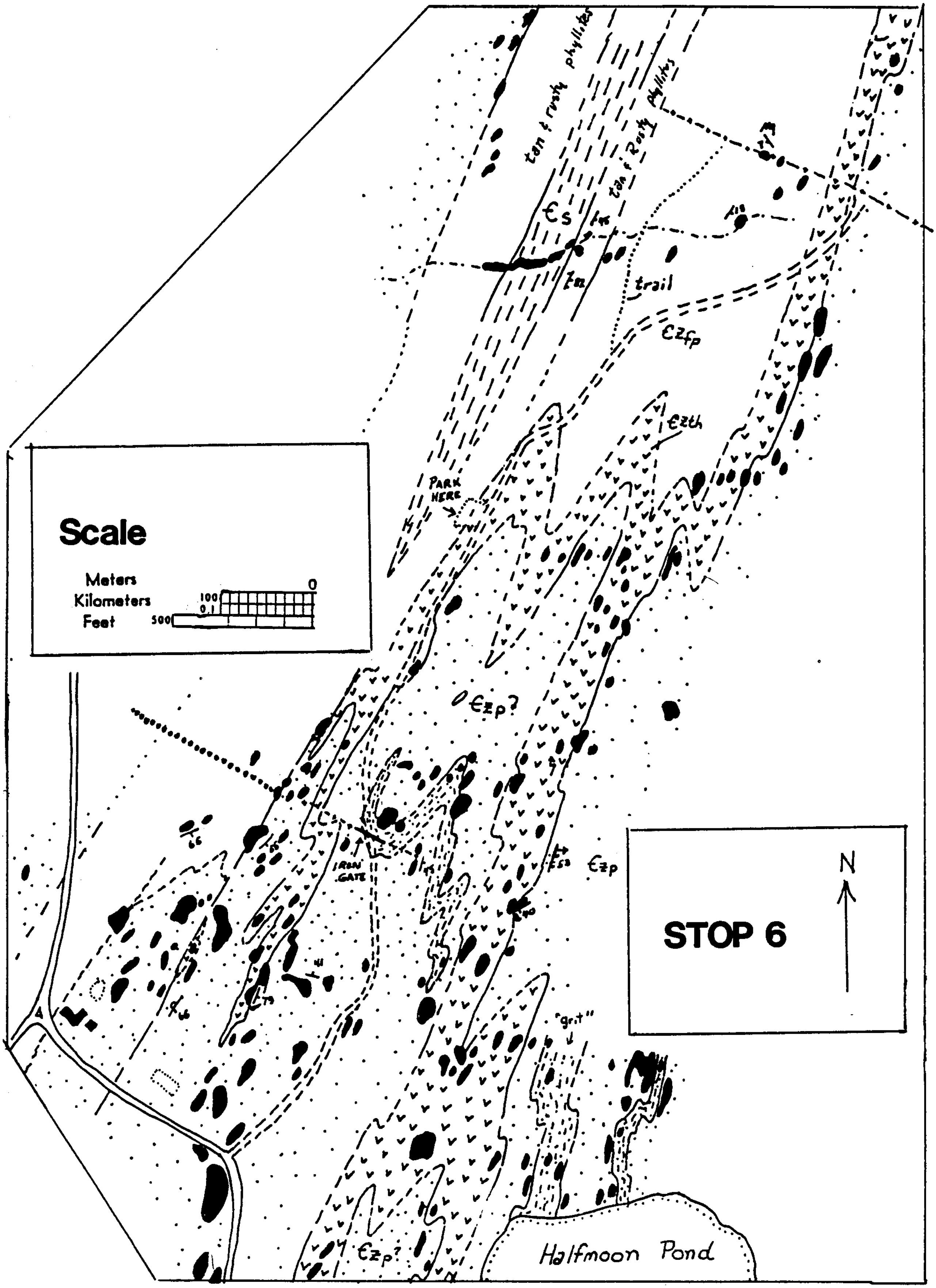

Figure 9. Geology by B.L. Doolan and David Greenwood.

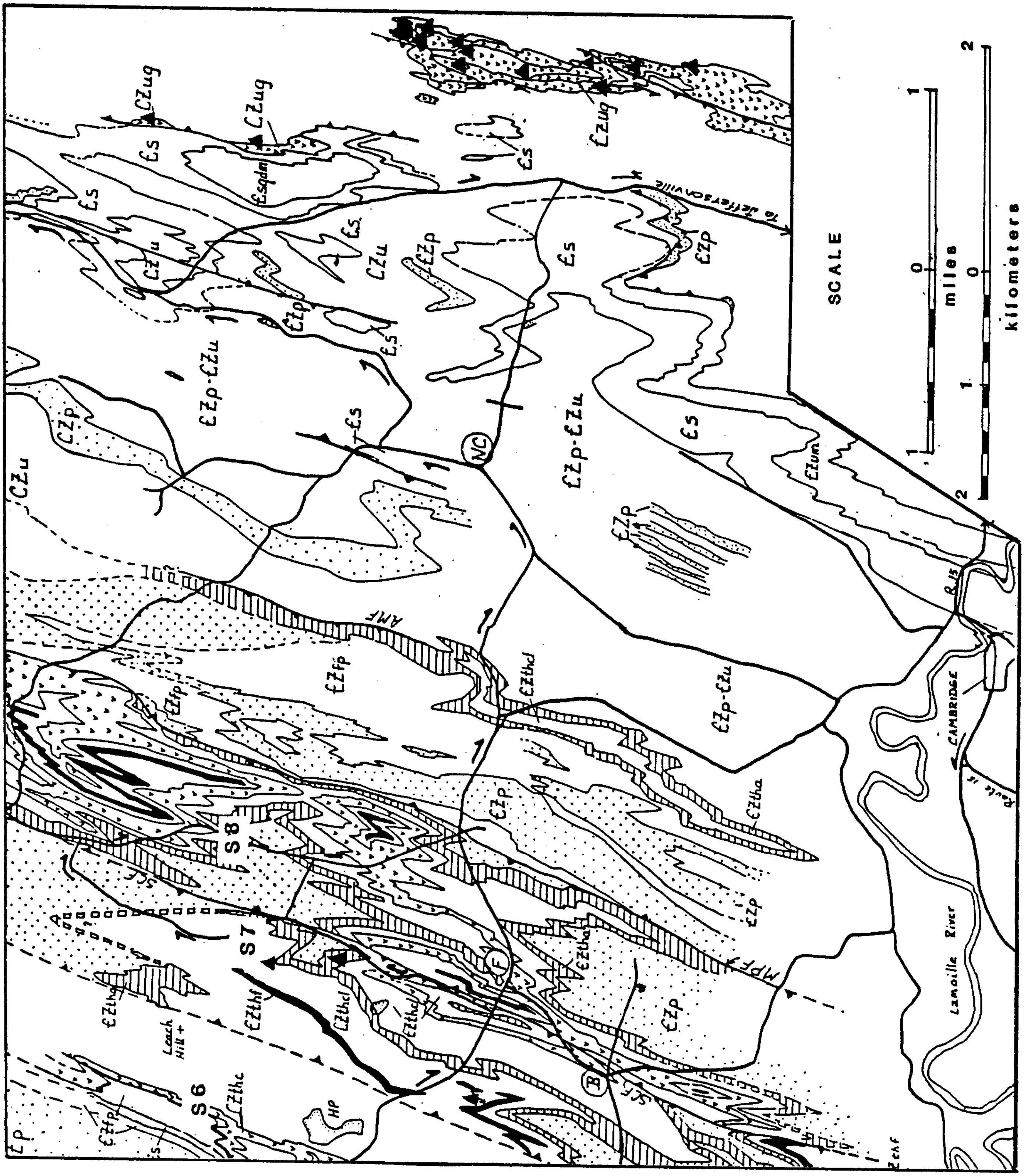

Figure 10. Location of greenstone analyses samples (Rose, 1986). Map is portion of Figure 3, this paper.

TABLE 2. MAJOR AND TRACE ELEMENTS BY X-RAY FLUORESCENCE ANALYSIS OF 13 SAMPLES OF UNDERHILL VOLCANICS FROM THE STUDY AREA, AND 4 TIBBIT HILL SAMPLES FROM 2 MILES EAST OF THE STUDY AREA, IN FLETCHER, VERMONT

Formation	Sample #	SiO_2 %	TiO_2 %	Al_2O_3 %	Fe_2O_3[3] %	MnO %	MgO %	CaO %	Na_2O %	K_2O %	P_2O_5 %	V ppm	Cr_2O_3 ppm	Ni ppm	BaO ppm	L.O.I. %	Total %
Ꞓs[1]	BP	57.74	0.93	17.17	5.81	0.08	1.30	7.76	5.56	0.53	0.15	142	57	<10	159	3.59	100.66
ꞒZug	B-2	50.50	2.89	15.16	11.95	0.15	7.16	4.20	3.44	2.73	0.29	362	97	50	769	2.37	100.96
ꞒZug	B-3	48.23	2.97	13.72	15.04	0.19	6.96	5.61	3.52	0.04	0.31	393	70	39	80	3.67	100.32
ꞒZtha[2]	HR-M1	47.40	2.72	14.85	13.48	0.18	5.94	8.92	3.96	0.43	0.56	287	180	44	176	2.38	100.89
ꞒZtha[2]	HR-M3	46.51	2.58	15.16	13.99	0.20	6.63	7.35	4.30	0.33	0.29	368	303	307	68	3.55	101.00
ꞒZthf[2]	HR-M4	45.73	3.60	12.86	15.70	0.29	4.62	7.86	3.00	1.21	1.17	238	54	11	440	4.84	100.95
ꞒZug	HR-M7	40.42	2.93	16.54	14.45	0.19	4.77	7.56	5.50	0.23	0.25	281	91	70	275	8.11	101.11
ꞒZthf[2]	HR-M4PH	47.07	4.01	15.93	12.84	0.18	3.07	8.31	3.23	1.83	0.62	329	48	25	747	3.57	100.78
ꞒZug	HR-45	46.89	3.44	15.56	17.81	0.22	3.74	7.83	1.61	0.07	0.32	432	98	146	<10	3.09	100.65
ꞒZug	HR-124	43.40	3.41	14.02	15.35	0.23	6.77	6.11	3.30	1.22	0.49	325	63	41	264	6.78	101.14
ꞒZug	HR-128	63.21	1.14	18.83	4.44	0.04	0.31	0.84	10.62	0.34	0.26	77	31	<10	88	0.65	100.71
ꞒZug	13-11	46.43	3.97	13.73	15.82	0.19	5.25	7.53	4.00	0.45	0.54	344	28	13	326	2.65	100.62
ꞒZug	14-28	53.09	4.10	11.53	19.07	0.23	3.32	4.44	0.28	0.37	0.44	473	23	27	157	3.48	100.42
ꞒZug	19-34	50.05	2.96	13.40	16.73	0.18	4.14	6.51	5.99	0.06	0.34	421	97	93	19	0.81	101.24
ꞒZug	19-49	49.68	2.73	14.64	14.11	0.16	4.64	7.57	5.67	0.10	0.28	439	44	25	93	1.20	100.84
ꞒZug	19-53	48.43	2.87	13.81	15.30	0.22	6.03	7.28	3.98	0.79	0.33	388	3372	3231	314	1.80	101.57
ꞒZug	20-58	47.95	2.62	15.92	15.26	0.17	6.92	3.15	3.59	0.06	0.51	332	111	32	69	4.41	100.62
ꞒZug	20-67	51.09	2.20	16.30	13.38	0.22	3.41	6.82	3.27	0.15	0.86	194	30	19	90	2.72	100.44

[1] Not greenstone.

[2] Tibbit Hill samples.

[3] All iron present has been calculated as Fe_2O_3. In most rocks, however, iron occurs mostly as FeO: about 90 percent of the total iron present. That is why in some cases totals are high, depending on the actual ferrous/ferric ratio.

TABLE 3. XRF ANALYSIS OF 13 SAMPLES OF UNDERHILL VOLCANICS FROM THE STUDY AREA, AND 4 TIBBIT HILL SAMPLES FROM TWO MILES EAST OF THE STUDY AREA IN FLETCHER, VERMONT.

Formation	Sample #	Nb ppm	Zr ppm	Y ppm	Sr ppm	Rb ppm	Pb ppm	Th ppm	U ppm
€$_s$[1]	BP	9	275	25	55	133	29	13	7
€Zug	B-2	20	204	36	169	61	14	7	12
€Zug	B-3	23	210	37	102	14	15	10	11
€Ztha[2]	HR-M1	25	187	35	365	22	17	8	18
€Ztha[2]	HR-M3	21	162	35	351	23	16	<5	12
€Zthf[2]	HR-M4HOMO	31	272	63	543	36	17	7	16
€zug	HR-M7	22	227	41	182	17	19	7	16
€Zthf[2]	HR-M4PH	34	319	49	548	49	13	<5	11
€Zug	HR-45	22	184	49	274	16	18	6	14
€Zug	HR-124	28	256	43	178	45	14	5	12
€Zug	HR-128	19	396	48	146	19	16	12	16
€Zug	13-11	32	284	50	207	21	10	<5	14
€Zug	14-28	25	230	53	79	25	15	7	14
€Zug	19-34	24	230	41	107	15	17	9	17
€Zug	19-49	22	214	43	190	15	18	7	16
€Zug	19-53	23	218	38	196	26	15	7	16
€Zug	20-58	25	169	39	75	15	13	6	15
€Zug	20-67	37	407	112	298	16	25	8	13

[1] Not greenstone.

[2] Tibbit Hill samples.

NOTE: Detection limit is 5 ppm.

Stop 8. Tibbit Hill Formation

This stop will entail a 1/2 mile walk along a marked trail which crosses an anticlinal limb of the stratigraphy of the Tibbit Hill Formation. Comparisons of the stratigraphic section on adjacent flanks of the anticline will be made. Of particular interest are the interbedded clastic rocks of the Tibbit Hill Formation which clearly differentiate the Tibbit Hill of the Fletcher anticline from the Tibbit Hill along the Pinnacle Mountain-Enosburg Falls anticline to the north. Figure 11 is a 1:5000 outcrop map of this traverse.

Return to cars and proceed back to the dirt road.

28.9 Intersection with dirt road; proceed south (left).

29.5 Road intersection on right; continue straight.

30.3 Intersection with paved road (Cambridge-Fletcher Road). Turn east (left).

30.9 Y-intersection; bear left onto dirt road.

31.2 Outcrops of amphibolitic greenstone mark the eastern limit of the Tibbit Hill Formation.

31.9 T-intersection; proceed north (left) on dirt road toward North Cambridge.

32.5 Road intersects on east side. This is North Cambridge! Proceed straight ahead.

33.0 Intersection on right after crossing bridge. Pull off to side and park. Proceed to outcrops on east side of the stream.

Stop 9. Underhill Formation

The large outcrop of lithologies of the Underhill Formation (CZu) display excellent minor structures and well developed schistosity surfaces. This outcrop is compared with Sweetsburg Formation rocks located on the west side of the road just northwest of the intersection. If the group is of reasonable size and/or willing, proceed down steep embankment to the northwest of the Sweetsburg outcrop to fault melange of Sweetsburg Formation lithologies. As of this writing, the continuation of the fault has not been mapped. These Sweetsburg lithologies mark the western limit of the Richford syncline in this area; outcrops of typical Pinnacle wacke containing detrital blue quartz grains outcrop to the west.

33.0 Turn east (right) at the road intersection.

34.8 Cross bridge followed by railroad tracks.

35.0 T-intersection with Route 108. Turn north (left) onto 108.

37.7 Y-intersection with dirt road on west side of Route 108. Bear left on the dirt road and park on side of road south of the bridge over Kings Hill Brook.

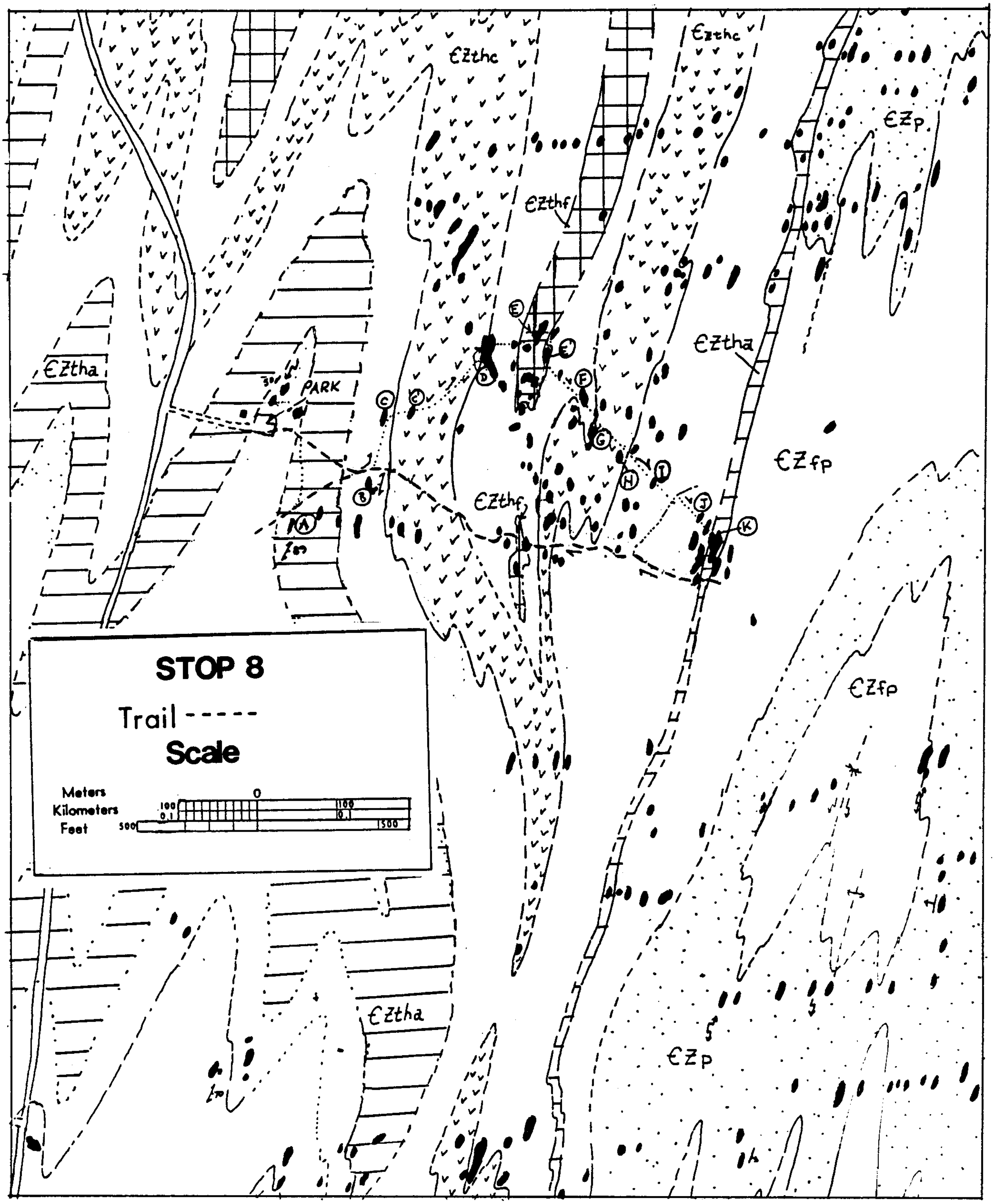

Figure 11. Geology by B.L. Doolan.

Stop 10. Sweetsburg Formation

Figure 12 is a 1:5000 outcrop map of this stop. Work done here is part of an M.S. thesis in progress by Tim Mock. The rocks on either side of Route 108 are examined and include rocks identical to those of the Pinnacle Formation, the Underhill Formation and the Sweetsburg Formation. The Sweetsburg Formation at this locality displays interbedded marble and thin bedded quartzite along with the more typical carbonaceous phyllite.

Return to cars and proceed north to the intersection with Route 108.

37.8 Intersection with Route 108. Turn south (right) and head toward Jeffersonville.

44.9 Intersection with Route 109 on west side of Route 108. Proceed straight on Route 108.

45.2 Cross the Lamoille River

45.4 Intersection with Route 15; turn south (right) on Route 15.

45.7 Bear right at the blinking yellow light. Cross the Lamoille River.

47.7 Flashing yellow light at the "wrong way bridge"; proceed left at the light, crossing the Lamoille River and staying on Route 15.

47.9 Return to the commuter parking in Cambridge.

END OF TRIP

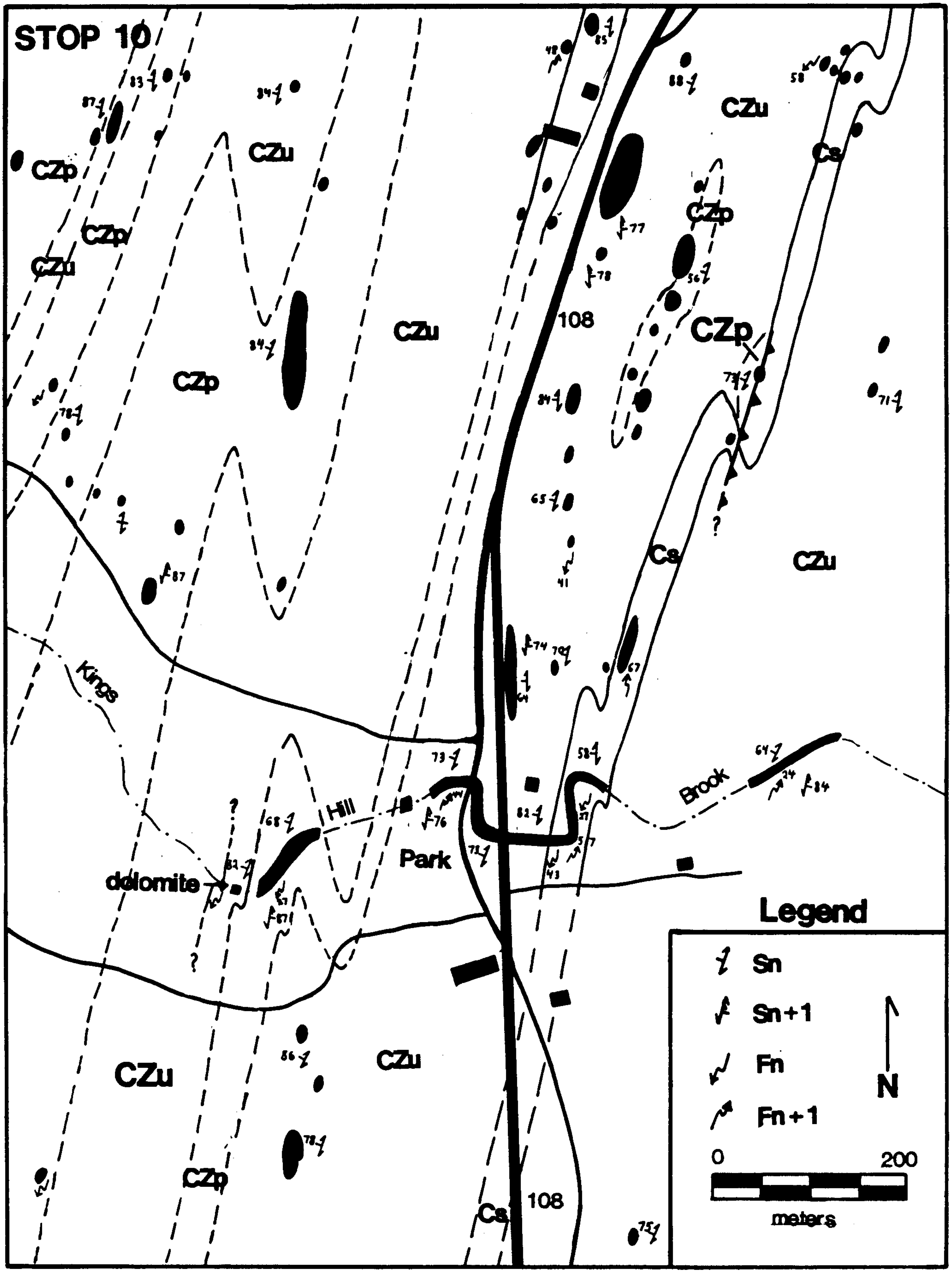

Figure 12. Geology by Mock (1987, unpublished map and field notes.)

July 12, 1986

THE BEEKMANTOWN GROUP IN THE CENTRAL CHAMPLAIN VALLEY

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VERMONT GEOLOGY VOLUME 5

VGS GUIDEBOOK 2

FIELD TRIP GUIDE F

STOP PAGE

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INTRODUCTION

The Beekmantown Group, originally the Calciferous Sandrock (Eaton, 1824), is the early Ordovician carbonate-clastic sequence of the Appalachian miogeosyncline. In the central Champlain Valley (the type locality) the stratigraphic limits of the group extend beyond the temporal limits of the early Ordovician. Also, despite considerable work by several different geologists over the space of more than a century, the details of the internal stratigraphy in the central Champlain Valley are still in dispute. This field trip will visit some localities where the stratigraphic and temporal relations can be well illustrated.

Although the Calciferous was defined by Eaton in 1824 (in the Mohawk Valley) and was mapped around the edges of the Adirondacks by Emmons (1839, 1842), it was not until Augustus Wing, an amateur geologist, undertook the task of deciphering the age and structure of the central Champlain Valley that the importance and internal stratigraphy of these strata were recognized (Wing, 1858-75; Dana, 1877a,b). The full results of Wing's investigations were never published (only Dana's [1877a,b] summary and a map in Cady [1945]), so it was only with the publication of Brainerd and Seely's (1890a,b) work, which was based to a great extent on Wing's mapping, that a formal stratigraphy was established for the Calciferous. Their stratigraphic section from eastern Shoreham has served as the basis for all subsequent work. The name change to Beekmantown, initiated by Clarke and Schuchert (1899) over the objections of several contemporary stratigraphers (Seely, 1910; Rodgers, oral communication, 1985), preserved this internal stratigraphy.

It is interesting to note that most of the points of disagreement over the details of the internal stratigraphy since the turn of the century have centered around those points where Brainerd and Seely (1890a) differed from Wing (1858-1875), and that recent work by Fisher and others (Fisher, 1966, 1977, 1984; Fisher and Mazzullo, 1976; Mazzullo and others, 1978) to the south have tended to return to Wing's original position. To a great extent our views of the stratigraphy continue this trend.

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Brainerd and Seely (1890a, b) included all of the strata between the Potsdam Sandstone and "Trenton" in the Calciferous. They differentiated five separate lithologic packages (now considered formations), labeled A-E from the base upward, and subdivided two of these, C and D, into four members each, labeled 1-4 from the base upwards. Because of the obvious Cambrian faunal assemblage in A, Clarke (1903) dropped it from the Beekmantown, a practice which has been followed by most subsequent workers. Recent detailed paleontologic analysis of trilobite assemblages by Taylor and Halley (1974) shows that the actual Cambrian-Ordovician boundary lies within B, as Rodgers (1937) had first suggested. Detailed work by Chisick (Chisick and Friedman, 1982; Chisick and Bosworth, 1984) has now confirmed Wing's (1858-1875, notebook 5, p. 8-10) belief that the top of the Early Ordovician lies within E. Thus, the Beekmantown Group in the central Champlain Valley, which is its type locality, actually extends from Late Cambrian to earliest Middle Ordovician.

STRATIGRAPHY

The stratigraphic nomenclature for the central Champlain Valley is somewhat confused because Cady (1945) correlated the strata east of the Orwell thrust line with units named by Keith (1923, 1932), but Welby (1959, 1961) generally used the standard terminology from the New York side of the lake, although some of the names are of more recent origin. We will generally use the New York terminology since we feel that the stratigraphic correlations proposed by Cady (1945) are, at best, not sufficiently established. For portions of the Beekmantown section, we will propose a modified stratigraphy based on our own experience (Fig. 1).

TICONDEROGA DOLOSTONE
(Rodgers, 1955; Welby, 1959, 1961)

The Ticonderoga Dolostone corresponds to Brainerd and Seely's (1890a) Division A, which they described (p. 2) as:

> *Dark iron-grey magnesian limestone, usually in beds one or two feet in thickness, more or less silicious, in some beds even approaching a sandstone. Nodules of white quartz are frequently seen in the upper layers, and near the top large irregular masses of impure black chert, which, when the calcareous matter is dissolved out by long exposure, often appears fibrous or scoriaceous. Thickness...310 ft.*

STRATIGRAPHIC SUMMARY

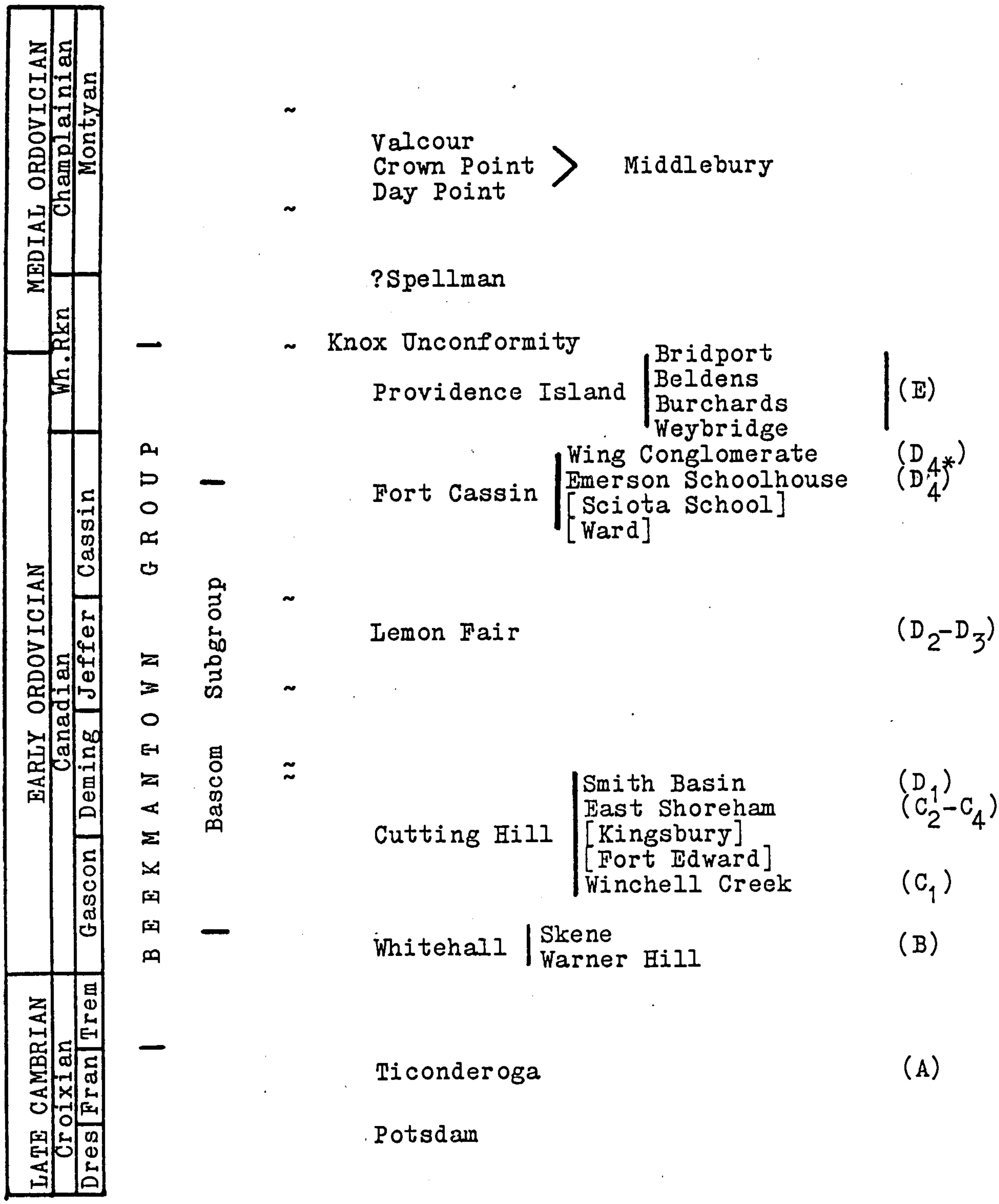

Figure 1. Summary of the stratigraphy of the Beekmantown Group in the central Champlain Valley. Brainerd and Seely's (1890a) stratigraphic designations for the various formations and members are given in parentheses.

Wing (1858-1875; Dana, 1877a) considered this unit to be "Upper Potsdam" rather than part of the "Calciferous". Despite Brainerd and Seely's (1890a) inclusion of this sequence, most subsequent authors (starting with Clarke, 1903, and Cushing, 1908) have tended to exclude it because of its Late Cambrian age. Cady (1945) correlated this unit with Keith's (1932) Clarendon Springs Dolomite and applied this term for the exposures in the Shoreham area. We agree with Keith (1932, 1933), however, that the correlation of strata between the eastern and western thrust sheets is not well constrained and that it is better to apply stratigraphic names from the autochthon to correlative strata in the western thrust sheets.

WHITEHALL DOLOSTONE
(Rodgers, 1937)

The Whitehall Dolostone corresponds exactly to Brainerd and Seely's (1890a) Division B, which they describe (p. 2) as:

> *Dove-colored limestone, intermingled with light grey dolomite, in massive beds; sometimes for a thickness of twelve or fifteen feet no planes of stratification are discernible. In the lower beds, and in those just above the middle, the dolomite predominates; the middle and upper beds are nearly pure limestone; other beds show on their weathered surfaces, raised reticulating lines of grey dolomite. Thickness...295 ft.*

The Shoreham outcrops are anomalous in the predominance of limestone, all other sections in the Champlain Valley containing relatively little limestone (with one exception - the Fort Ann), the dolostone being the principal lithology.

Ulrich and Cushing (1910) and Rodgers (1937) placed the boundary between the Ticonderoga and Whitehall above the lithologic boundary where they perceived the Cambrian-Ordovician boundary as lying. Although virtually all other workers (including the present authors) place the stratigraphic boundary at the lithologic break, Taylor and Halley (1974) have shown that the period boundary does indeed occur within the Whitehall, although the exact stratigraphic level has not been determined.

Cady (1945) correlated the Shoreham exposures with Keith's (1923, 1932) Shelburne Marble and applied that name. Despite the lithic similarity (ignoring metamorphic effects) between the Shelburne and Shoreham exposures, we feel that the application of the term "Whitehall" to the Shoreham exposures is preferable since the correlation is much more definite and Keith (1923) reported Early Cambrian fossils in his type section of the Shelburne. Also, the interfingering between the dolomitic and calcitic facies can be seen occurring within the Shoreham duplex.

Skene Dolostone Member
(Wheeler, 1941)

The majority of the Whitehall consists of light gray to tan, coarse (sugary) to medium-grained, massive dolostone. These grade vertically and laterally into the light gray weathering, fine-grained, dark gray dolomitic limestones of the Warner Hill. These nondescript dolostones are unfossiliferous and most of the original textures and fabric have been obliterated by pervasive dolomitization. Mottled in part, the dolostones are silty, cross-stratified and fetid. Fenestrae and pseudobreccia textures are common in these rocks, and burrows and bird's-eye structures, intraformational conglomerates, channels, desiccation cracks, finely laminated graded beds also occur frequently. Oncolites, cryptalgal-laminates and flat-pebble conglomerates are associated with the desiccated dolostone beds. Round-pebble conglomerates admixed with skeletal-hash compromise the basal lag-concentrates within channels and are overlain by cross-stratified silty dolostones. The grain size of the detrital quartz and feldspar components is 80u. The channels are interpreted as tidal channels by analogy to the fining-upward modern tidal-channel sequences described by Friedman and Sanders (1967).

Warner Hill Limestone Member
(Fisher, 1977)

Distributed throughout the Whitehall are massive, light gray weathering, medium gray, very fine-grained dolomitic limestones with occasional black chert nodules and intraformational breccias. Locally, stromatolitic (Crytozoan) mounds with flanking coarse-grained limestones and oolites are common. As shown by Taylor and Halley (1974), trilobites abound. The general stratigraphic succession of burrow-mottled limestone, thin- to medium-bedded limestone, thin-bedded limestone with shale partings indicates an upward increase in clastics. The partings occur along argillaceous-rich or dolomitic surfaces. These surfaces are typically flat and parallel, rarely exhibiting starved ripples or parting lineations. Typically the thin- to medium-bedded limestones are characterized by medium to dark gray limestone with coupling orange-weathering silty dolostone laminae partings. Each is interbedded and intergradational with the overlying unit.

Bascom Subgroup
(new designation)

The Bascom Subgroup is defined as the entire carbonate clastic succession extending from the base of the Winchell Creek Sandstone to the top of the Wing Conglomerate (top of the Emerson Schoolhouse Member where the Wing Conglomerate is absent). This sequence represents a very continuous and repetitive sequence forming the "core" of the Beekmantown strata. It is this sequence which led Eaton (1824) to coin the term Calciferous sandrock. The Bascom name is appropriate for the entire sequence since the primary ledges of the "Bascom Ledges", especially those around the old Bascom Farm and along Bascom Road, consist of various portions of this sequence.

CUTTING HILL FORMATION
(modified from Cady, 1945)

The Cutting Formation, as originally defined by Cady (1945), corresponds exactly to Brainerd and Seely's (1890a) Division C. We will follow Wing's (1858-1875) lead, recently adopted by Fisher and Mazzullo (1976; Mazzullo and others, 1978) in the southern Champlain Valley, and include Brainerd and Seely's (1890a) Division D member 1 in the Cutting (which we then rename the Cutting Hill to distinguish it from Cady's original formation). Although Brainerd and Seely (1890a) divided Division C into four members which Welby (1961) claimed to be able to distinguish throughout the central Champlain Valley, the differences among the upper three are so very slight that differentiation is difficult where large portions of the section are not exposed. Thus C-2, C-3, and C-4 are combined into a unified East

Shoreham dolostone Member of the Cutting Hill Formation (roughly corresponding to the Ft. Edward Limestone of Fisher and Mazzullo, 1976, and the Kingsbury Limestone of Fisher, 1984). We will use Fisher and Mazzullo's (1976) terms Winchell Creek Sandstone and Smith Basin limestone for the C-1 and D-1 members that lie at the base and top of the Cutting, respectively. In redefining the Cutting Hill in this way, we make it the exact equivalent of the Great Meadows Formation to the south and make it correspond to Wing's definition of "the Ophileta beds" in the Shoreham area.

Winchell Creek Sandstone Member
(Fisher and Mazzulo, 1976)

The Winchell Creek Sandstone is Wing's (1858-1875; Dana, 1877a) Scolithus sandstone, which Brainerd and Seely (1890a) made member C-1 and described (p. 2) as:

> *Grey, thin-bedded, fine-grained, calciferous sandstone, on the edges often weathering in fine lines, forty or fifty to the inch, and resembling close-grained wood. Weathered fragments are frequently riddled with small holes, called Scolithus minutus by Mr. Wing...60 ft.*

The two diagnostic features included in this description, fine lamination and Scolithus burrows, are most prominent in the Shoreham-Cornwall area. To the north, this member becomes a breccia, with angular carbonate blocks set in a silty matrix. To the south, cross-bedding becomes more prominent. In both cases, Scolithus burrows decrease in abundance, being quite rare at Thompson's Point and south of Whitehall.

East Shoreham Member
(new name)

The East Shoreham dolostone Member, as stated above, includes Brainerd and Seely's (1890a) members C-2, C-3, and C-4 which they described (p. 2) as:

2. *Magnesian limestone in thick beds, weathering drab....100 ft.*
3. *Sandstones, sometimes pure and firm, but usually calciferous or dolomitic ....70 ft.*
4. *Magnesian limestone like no. 2, frequently containing patches of black chert ....120 ft.*

Although Brainerd and Seely are correct that sandstone beds are generally more common in the central portions of this sequence (which allowed Welby [1961] to distinguish these three members throughout the central Champlain Valley), they also occur in the upper and lower portions. The central portion also consists primarily of dolostone, not sandstone. It is the opinion of the authors, therefore, that this entire section should be considered a single member consisting primarily of dolostone with scattered sandstone layers, especially common in the middle of the section. The name is derived from the East Shoreham cemetery which lies atop Cutting Hill in eastern Shoreham where Brainerd and Seely (1890a) defined Division C and Cady (1945) derived the formation name. We consider this member to be roughly equivalent to the upper Fort Edward Dolostone Member of Fisher and Mazzullo (1976).

Smith Basin Member
(Fisher and Mazzullo, 1976)

The Smith Basin limestone is Brainerd and Seely's (1890a) member D-1, which they describe (p. 3) as:

> *Blue limestone in beds one or two feet thick, breaking with a flinty fracture; often with considerable dolomitic matter intermixed, giving the weathered surface a rough, curdled appearance; becoming more and more interstratified with calciferous sandstone in thin layers, which frequently weathers to a friable, ocherous rottenstone...80 ft.*

The sandstones Brainerd and Seely report near the top of the unit are rare, the lithologic break with the overlying strata generally being quite abrupt. The reasons for including Smith Basin in the Cutting Hill, rather than in Division D as Brainerd and Seely (1890a) and Cady (1945) did, are its strong association with the paleokarst unconformity atop the Cutting Hill and its strong faunal connection with the underlying dolostones. Much of the known exposures of the Smith Basin are preserved in paleokarst solution pits within the East Shoreham dolostone. The faunal correlation led Wing (1858-1875) and Fisher and Mazzullo (1976) to draw the stratigraphic boundary in the same place that we have. The inclusion of the Smith Basin in the Cutting Hill is also reinforced geographically, as the best outcrops in the central Champlain Valley are on the east side of Cutting Hill just north of East Shoreham.

As noted by Wing, the Smith Basin limestone pinches out rapidly to the west, disappearing completely within the Shoreham duplex (Fig. 2). In fact, the only good exposures of this member in the Champlain Valley are in eastern Shoreham, along the Cornwall Ledges, and in West Haven. Whether it occurs on Thompson Point is unknown since the interval in which it would be expected is within a covered interval.

LEMON FAIR FORMATION
(new name)

Brainerd and Seely (1890a) differentiated four members in their Division D. The middle two of these, D-2 and D-3, are a carbonate-siliciclastic sequence which they describe (p. 3) as:

2. *Drab and brown magnesian limestone, containing also toward the middle several beds of tough sandstone....75 ft.*
3. *Sandy limestone in thin beds, weathering on the edges in horizontal ridges one or two inches apart, giving to the escarpments a peculiar banded appearance. A few thin beds of pure limestone are interstratified with the silicious limestone...120 ft.*

Based on our experience, the two lithofacies, D-2 and D-3, can not be separated vertically or laterally on a regional basis as they apparently interfinger, so we propose lumping them together into a single unit. The name comes from the most prominent outcrops of Lemon Fair strata in the central Champlain Valley which occur in the cliffs along the sides of the Lemon Fair Valley in Shoreham, Cornwall, and Weybridge. Although fossils are rare and generally not definitive in

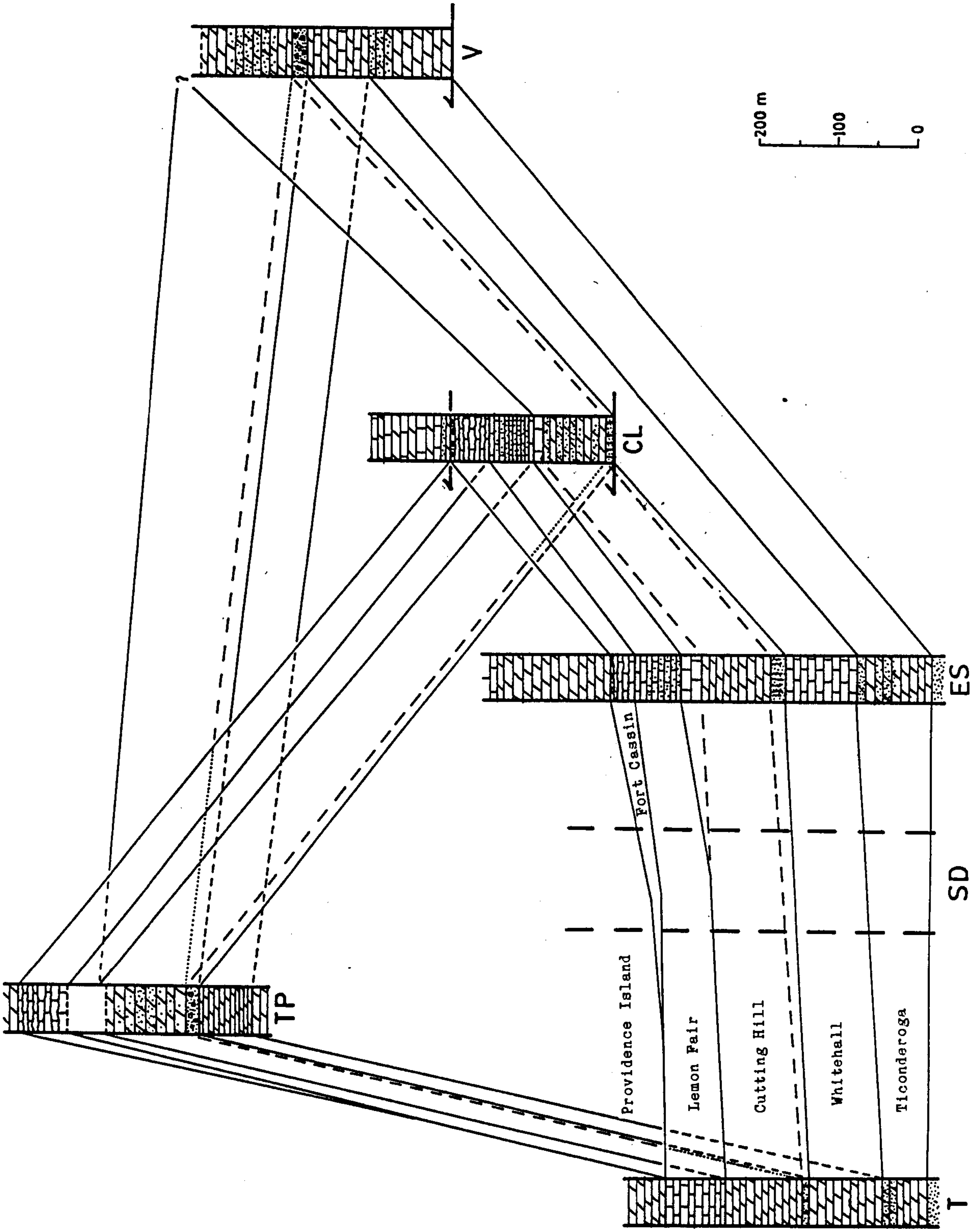

Figure 2. Correlation of the most complete sections of Beekmantown strata in the central Champlain Valley. TP - Thompson Point, Charlotte (Brainerd and Seely, 1890a; Welby, 1961); V - Vergennes thrust sheet (Welby, 1961); CL - Cornwall ledges about 2 km south of Stop 1 (Welby, 1961); ES - East Shoreham (Brainerd and Seely, 1890a); SD - the Shoreham duplex (Washington, 1985); T - Ticonderoga (Rodgers, 1955, published in Welby, 1961). The Vergennes thrust sheet has been restored to a predeformational position.

these strata, our detailed analysis (Chisick, unpublished) of conodont assemblages indicates that the Lemon Fair Formation is part Demingian and part Cassinian with the entire Jeffersonian stage missing.

FORT CASSIN FORMATION
(Whitfield, 1890)

The original definition of the Fort Cassin Formation (Whitfield, 1890) included all of Beekmantown D-3, D-4, and E. As Brainerd and Seely (1890a, p. 20-21) pointed out, however, the strata on Fort Cassin are from the upper parts of Division D and the fauna are most completely represented therein.

Actually, it was Wing (1858-1875) who most fully recognized the stratigraphic relations when he distinguished between the "Trilobite beds", the "Conglomerate", and the "Rhychonella beds". His trilobite beds correspond to the Lemon Fair and Fort Cassin formations (the Lemon Fair is generally devoid of distinctive fossils), his Rhychonella beds correspond to Brainerd and Seely's (1890a) Division E, and the conglomerate is a unit that occurs between them in eastern Shoreham, Cornwall, and Weybridge, but not at Fort Cassin, Thompson's Point or Beekmantown where a paleokarst removed it. The Fort Cassin itself pinches out to the west, thinning greatly within the Shoreham duplex and being absent at Ticonderoga (Fig. 2).

Lithologically the Fort Cassin is an interbedded bioarenitic limestone and clayey limestone with minor dolomitic laminae and subordinate dolostone. The horizontal calcitic laminae show an undulating relief interlaminated with the dolomitic laminae. The latter laminae commonly are graded or cross-laminated and the graded dolostones consist of couplets of silty dolorudite and skeletal-debris doloarenite. The beds are generally 5 cm or less thick. Limestone layers commonly contain abundant fossils, occasionally consisting almost entirely of fossil hash. Slightly dolomitized burrow-mottled limestones contain abundant fauna, dominated by gastropods and cephalopods. These limestones are carbonate mudstones or peloidal wackestone to packstone. Somewhat argillaceous and pyritic, these rocks have very fine, parallel continous laminations with rare lenses of peloidal or bioclastic grainstone. Micritization and recrystallization are the two major processes inferred to have produced the laminated crust in the dolostone. Horsetail stylolites and solution breccias and laminated crusts abound. Roehl (1967) described similar features and attributed them to subaerial leeching. Supratidal-upper intertidal, brine-enriched algal-flats such as formed the Fort Cassin are known to encourage dolomitization (Friedman and Sanders, 1967).

Ward Siltstone Member
(Fisher, 1977)

A thin sequence of thin to medium bedded, laminated and cross-laminated, sometimes cross-bedded, tan to light gray, calcerous and dolomitic siltstones with some sandstone occurs sporadically (most commonly in the southern Champlain Valley) as a gradational boundary between normal Fort Cassin and underlying Lemon Fair strata. Planar laminae, small-scale cross-laminae and very thin graded beds commonly form herringbone patterns. Erosional surfaces are overlain by basal lags of imbricated, subrounded intraclasts with concave-upward cross-laminae. Some bedsets are high-angled with very thin graded beds of quartz sand and dolomicrite. Rarely, ripples and convoluted laminae are observed. Klein (1970) suggested that herringbone bedding indicates deposition by tidal currents having approximately equal ebb and flood velocities. Braun and Friedman (1969) show beds with erosional surfaces overlain by basal lags composed of interclasts of tidal-flat lithologies. Sand grains are abundant. Peloids and oncolites are present but less common in these intraclastic laminated dolocalcarenites. Intertidal to shallow subtidal bars and channels are the presumed setting. Although few trace fossils are seen, U-shaped worm-tubes are locally abundant.

Sciota Schoolhouse Member
(Fisher, 1977)

The central portion of the Fort Cassin, especially in the southern Champlain Valley, commonly consists of thin to thick bedded, massive, light gray weathering, dark gray to black dolomitic limestone with laminated limestone and light gray coarse grained limestone interbeds. The Sciota Schoolhouse is very fossiliferous with numerous straight and coiled cephalopods (Flower, 1964). The earliest ostracodes and many high-spired gastropods are also reported. Contorted limestone and limestone breccia facies are conspicious and grade into shallow-water platform carbonates. The limestones have papery partings with thickness increasing as the carbonate content increases. Intrasparite and intramicrite commonly fill shallow tidal-channels characterized by low-angle planar cross-stratification. Single laminae within cross-stratified sets are graded from intrasparite in the toeset to intramicrite in the foreset. Locally higher-energy conditions existed as indicated by lenses of biopelintramicrite.

Emerson Schoolhouse Calc-Dolostone Member
(Welby, 1961)

In the central Champlain Valley much of the Fort Cassin is thin to thick bedded, massive, mottled medium to light gray weathering, medium to dark blue-gray to dark gray dolomitic limestone and calcitic dolostone with buff weathering. The nodular nature of the limestone suggests syndepositional compaction and stretching that produced sedimentary boudins. The slight overlapping of nodules in the same horizon and the folding-contortion of some nodules may indicate minor down-slope movement. Most units of this type are characterized by a basal detachment surface that is typically planar and parallel with underlying beds. Locally the surface is undulating and may truncate underlying units. Fine to medium grained black cherty dolostone layers, locally very vuggy with frequent cauliflower calcite nodules, are interspersed throughout this member. These vugs probably were sulfate nodules with the calcite nodules being what Chowns and Elkins (1974) described as pseudomorphs after sulfate minerals. Within some nodules the geopetal-fillings are probably collapse-features dissolved at a greater rate. Tarr (1929) reported these features along with doubly terminated quartz in nodules from the Galena Group in Missouri. Stromatolites, abundant oncolites and trace fossils are very common. It is the Emerson Schoolhouse Member that holds the bulk of the Fort Cassin fauna as described by Whitfield (1886; 1890). We feel that Welby's (1961) differentiation of the Thorp Point Member is artificial, the distinction being based on varying proportions of constituient lithologies. The Thorp Point is thus lumped together with the Emerson Schoolhouse Member.

Wing Conglomerate
(Seely, 1906)

Along the top of the Fort Cassin, except where it has been removed by a later paleokarst, lies a very thin (<2 m) polymictic conglomerate of flat pebbles, oncolites, and siltstone clasts with alternating sandy siliciclastic-rich carbonate and minor fine grained siltstones and silcretes. The micritic matrix has been recrystallized to microspar and pseudospar. Interbedded are thin lenses of limestone with flat- and round-pebble conglomerates, desiccation cracks, small channels, thin couplets of graded micrite and biopelmicrite, and occasional lenses of finely laminated limestone. The intraclasts of the Wing Conglomerate are set in a matrix of biopelmicrite and dolopelmicrite. The intraclasts are composed of pelmicrite, micritized intraclasts of dolopelmicrite, biopelmicrite, or fine alternations of these lithologies.

PROVIDENCE ISLAND FORMATION
(Ulrich and Cooper, 1938)

The Providence Island Formation corresponds exactly to Brainerd and Seely's (1890a) Division E which they described in eastern Shoreham (p. 3) as:

> *Fine-grained magnesian limestone in beds one or two feet in thickness, weathering drab, yellowish or brown. Occasionally pure limestone layers occur, which are fossiliferous, and rarely thin beds of slate. Thickness...470 ft.*

In the Shoreham area, the limestone layers are most abundant in the lower middle portion of the unit. To the east (in central Cornwall), however, the character of the strata changes abruptly to interbedded limestones and brown-weathering dolostones (beds generally between 10 cm and 1 m thick), with occasional layers of sandy limestone. Cady (1945; Cady and Zen, 1960) has termed the interbedded limestone-dolomite facies the Beldens, the sandy limestone the Weybridge, the standard dolomitic facies the Bridport and the pure limestones the Burchards. The entire formation he renamed (Cady and Zen, 1960) the Chipman Formation, but since the Providence Island name has precedence, we prefer to use it. This also makes our nomenclature consistent with that presently in use in New York (Fisher and Mazzullo, 1976).

Bridport Dolostone Member
(Cady, 1945)

West of the ledges in Cornwall and Weybridge, the Providence Island consists primarily of massive, thick-bedded, tan to gray weathering, very dark gray siliceous dolostone with some thin interbeds of argillaceous limestone that is dark-gray to black very fine-grained (sublithographic). It generally weathers to a light bluish-gray with pronounced fretwork. Locally, white to bluish-white chert is common. Fossils are extremely rare.

Weybridge Siltstone Member
(Cady, 1945)

Near the base of the Providence Island, especially along the Cornwall Ledges, there is commonly a light to medium gray, buff to tan-orange weathering, thin- to medium-bedded, quartzitic siltstone. A light tan coarser-grained, fine sandstone generally starts this member. Cross-bedding is common but subdued.

Beldens Dolostone Member
(Cady, 1945)

From the Cornwall Ledges eastward the Providence Island is primarily a sequence of alternating layers (.5 to 2 m thick) of blue to gray, light gray to buff weathering limestone and blue to blue-black, brown to buff weathering dolostone. The limestones are primarily micrite with occasional thin shale laminae. Although generally unfossiliferous, locally (especially along the Cornwall Ledges) fossils abound. The dolostones are generally quite silty. According to Kay (1958), the Beldens is definitive "Upper Canadian", but analysis of conodonts (Chisick, unpublished) indicates that the strata are actually White Rockian.

Burchards Limestone Member
(Kay and Cady, 1947)

Distributed through the lower portions of the Providence Island are zones of mainly calcirudite and calcilutite, These zones, which are of limited lateral extent, are quite destinctive because they are generally brown weathering, medium-bedded, with off-white thin beds. White marble-like calcilutite is often interbedded in lower sections with sandy Weybridge-type siltstones. Mottling is common but not pervasive. Fossils are generally very sparse.

BRIEF SEDIMENTATION HISTORY

During the Late Cambrian, Early Ordovician, and early Middle Ordovician, a shallow tropical sea extended into Vermont and adjacent New York as an epicontinental embayment. Sedimentation occurred under marginal-tidal to epeiric conditions in an extensive belt separated from normal marine tidal conditions by a seaward barrier complex of ooid and peloid shoals. Channels through this barrier extended modest tidal effects to the periphery of an epeiric region where a patchwork pattern of semi-emergent shoals was transitional with marginal emergent pavements. Indurated pavements, firmgrounds, algal mounds, and shoals of carbonate detritus constituted a varied but low relief bathymetry.

Sedimentation was cyclic as a result of repeated shoaling and progradational offlap during storms, prevailing epeiric or tidal conditions, as well as small eustatic oscillations. Increasing salinities, evaporitic and emergent conditions resulted.

Furthermore, sabkhas developed on the emergent shoals and tidal-flats with gypsum precipitation within and dolomitization of the sediments. Under more extreme temperatures and salinities within the sabkhas, anhydrite nodules precipitated at the expense of metastable gypsum. Subsequently, near-surface chertification and dedolomitization occurred. Dissolution of sulphates, creation of karsts and calcretes resulted where emergence was prolonged.

These carbonate suites have been modified by interaction with hypersaline fluids throughout diagenesis, resulting in superimposed features of marginal-marine sabkhas, pervasive dolomitization, minor mineralization, and even dedolomitization. The various depositional environments were sensitive to sea-level fluctuations and change of shelf slope. Thus, the Beekmantown environments migrated seawards across the shelf during progradational offlap as a response to a slow drop in relative sea-level. With subsequent rise in relative sea-level, transgressive deposition occurred until

Figure 3a. Schematic diagram of paleoenvironments and the resulting lithologies for early Beekmantown time.

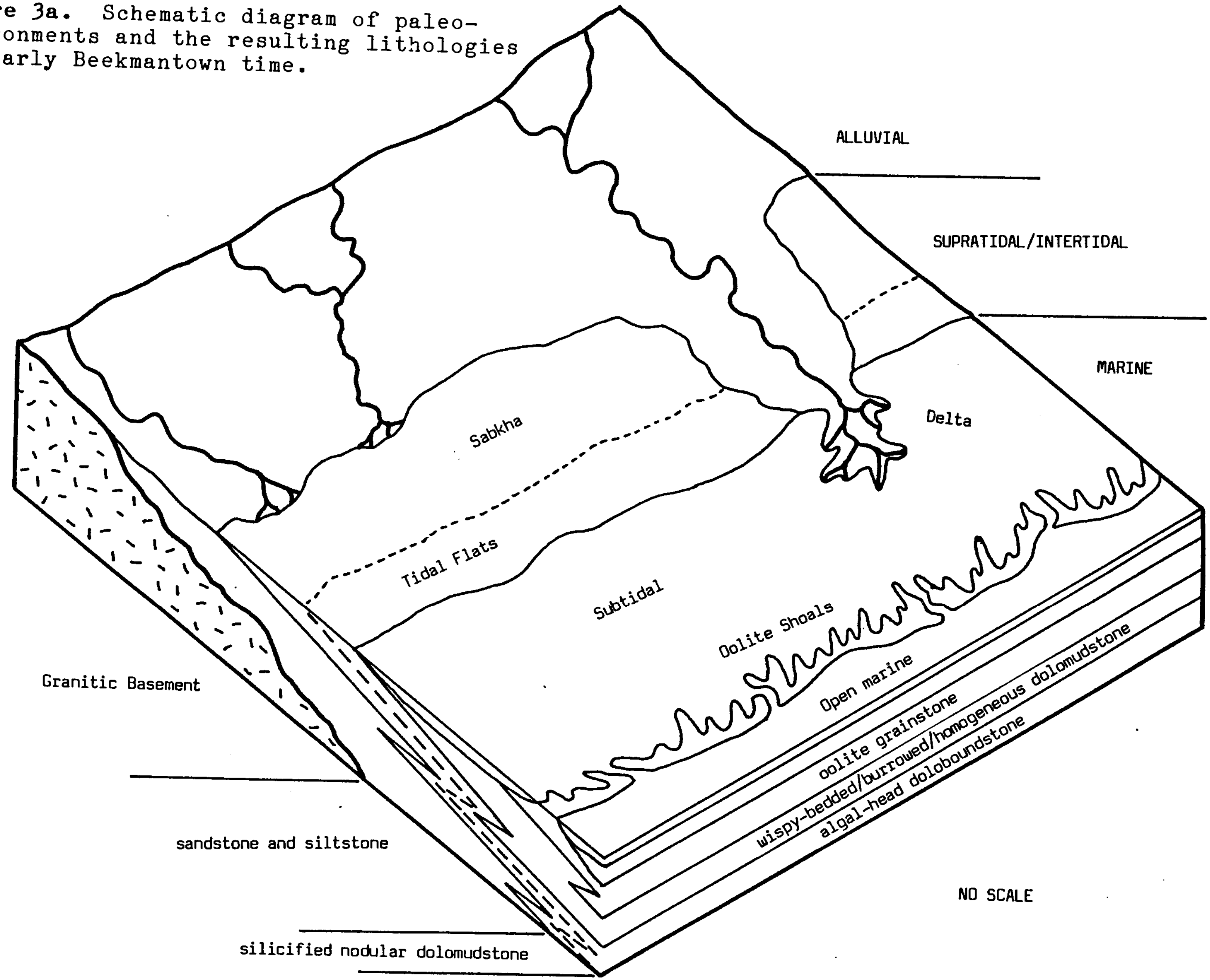

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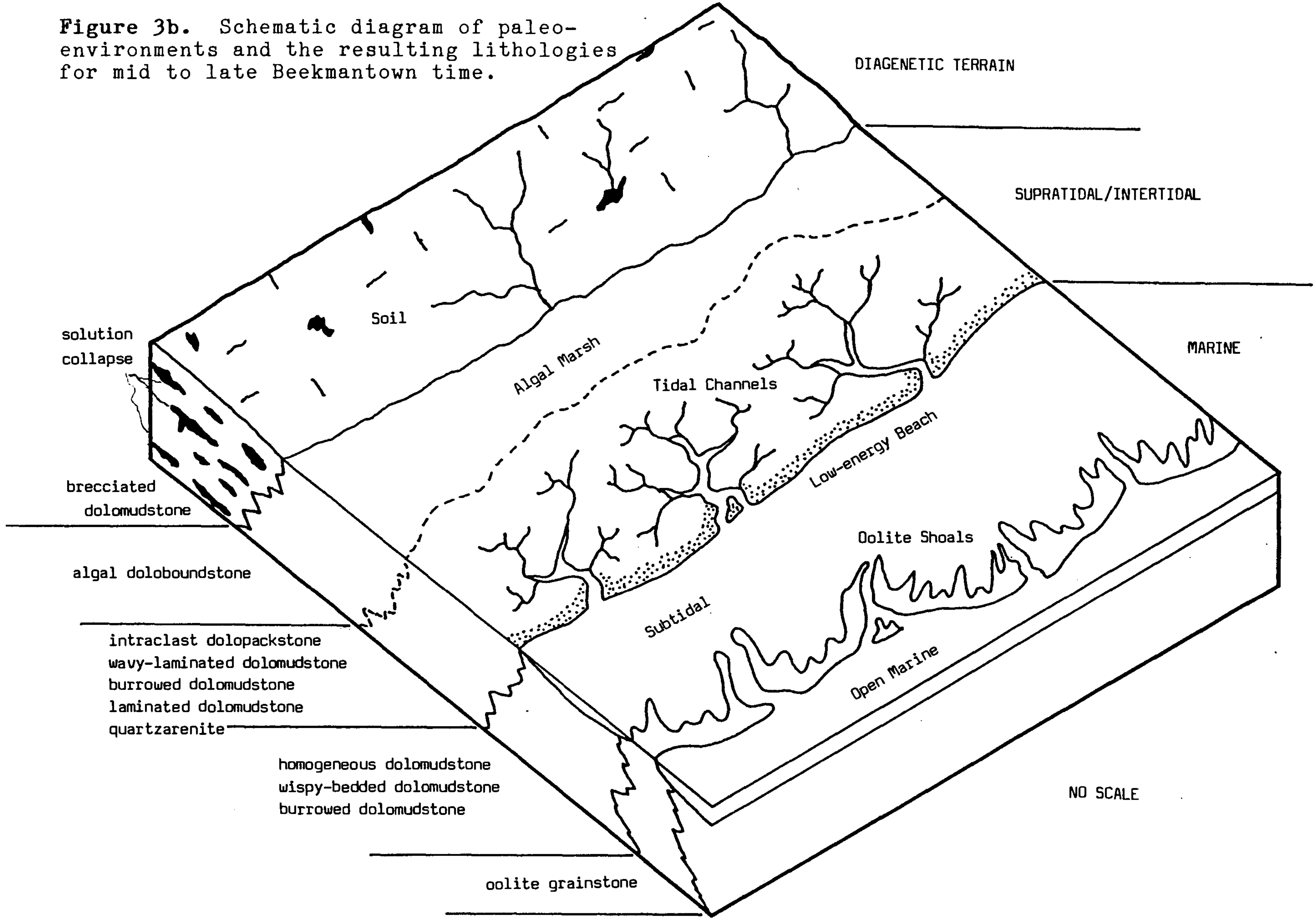
Figure 3b. Schematic diagram of paleo-environments and the resulting lithologies for mid to late Beekmantown time.



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excessive sediment accumulation rates again produced offlapping progradation of lithofacies.

The mega-sequences are interpreted as recording cycles of tectonic subsidence of the sedimentary basin and related upheaval of the source region which produced pulses of clastic input. These sediments range from 80u to 300u in size. The initial subsidence of the basin was followed by a gradual progradation of a sand-plain over a tidal-flat (Fig. 3a). Originally these deposits were ephemeral river sediments reworked by the wind. The eolian sands/silts were driven over the tidal-flats and sabkhas by strong offshore winds or were remobilized by an influx of fresh-water flooding down the ephemeral rivers. Subsequently, a period of quiescence and tectonic subsidence of the basin created a major transgression. Repetition of this pattern resulted in the cyclic nature of the preserved sections.

Gradually, however, the relief diminished and slower transgression resulted (Fig. 3b). The rather abrupt lower boundaries of the Beekmantown formations can be correlated with regional and local unconformities. The source regions of the sandstones/siltstones were indicated by paleocurrent data from foreset dip interpretations and from the local facies patterns. This interplay of sand-transporting offshore winds and flood-gorged ephemeral rivers with modifications of these deposits by waves, balances with the observed units.

REGIONAL STRUCTURE

The structure of the central Champlain Valley has long been considered relatively simple, with autochthonous rocks by the lake, a narrow line of thrusts (Champlain, Orwell, etc.) running down the western edge of the Red Sandrock range, and a large recumbent synclinorium between there and the Green Mountain front. Recent detailed mapping is finding that this picture is not totally correct. In particular, the synclinoral structure proposed for the allochthon by Wing (Dana, 1877a, b) and promulgated by Cady (1945) does not exist. As Keith (1932, 1933) recognized, this area is a thrust belt, any stratigraphic repetition across the "core of the synclinorium" resulting from the occurrence of successively older strata in the thrust sheets to the east of Middlebury. Keith (1932, 1933) also recognized, although he did not work out the details, that the line of thrusts along the western edge of the Red Sandrock range is not as simple as generally thought.

The plateau around Middlebury (extending westward to the Cornwall ledges - STOP 1) is a complex thrust belt containing only Providence Island and younger strata. It is bounded on the east by the main marble belt and a complex dolostone-sandstone-metapelite sequence, both of which abut the plateau structural belt along major thrusts. Although the correlation has not been proven because of a dearth of fossils, both Keith (1932) and the present authors feel that the complex eastern sequence is equivalent to the Monkton Quartzite and related rocks of the Red Sandrock range.

Furthermore, the Champlain thrust, which has generally been considered to floor the synclinorium, can now be shown to have originally overlain the structure. The Snake Mountain and Buck Mountain massifs, instead of representing the western edge of a synclinorium, are structurally preserved remnants of an original roof thrust sheet that overlay the Middlebury region. Although these massifs might be termed klippen, they actually are bounded on their eastern sides by thrusts that rise from within the plateau structure and breach the original Champlain thrust surface (Keith [1932, 1933] called these thrusts the Weybridge thrust). Thus, their trailing edges were overlain by the leading edge of the plateau (a relation which caused previous workers, except Keith, to assume the Champlain thrust lay beneath the "Middlebury Synclinorium").

The pattern of roof breaching continued westward, isolating another remnant of the Champlain thrust sheet to produce the ridge extending south from Vergennes (i.e., the Vergennes thrust sheet). When reconstructed, it becomes apparent that the Champlain thrust sheet totally overlay the Orwell and Middlebury plateau thrust sheets. We postulate that the Champlain thrust at Snake and Buck Mountains and the Vergennes thrust are westward extensions of the "New Haven" thrust (our name) which is the western boundary of the complex dolostone-sandstone-metapelite sequence east of Middlebury (Fig. 4).

In view of this structural synthesis, all of our stops are below the original Champlain thrust system. Stop 1 is in the upper plate of the penetrating thrust (called the Weybridge thrust by Keith [1932, 1933]) along the back of the Snake Mountain massif. The facies changes from carbonate-siliciclastic to predominately pelitic in the Champlain thrust sheet make only the western-most portion of the sheet easily correlated with the "classic" Beekmantown of the Shoreham area.

Over the last few million years (or possibly less), the Adirondacks have domed up, causing a number of high-angle faults to form along the western edges of the Champlain Valley. Although not as important for the map pattern along the edge of the allochthon as Welby (1961) claimed, they do bring a number of additional exposures of Beekmantown strata to the surface along the edges of the Adirondacks. Included in these are the name locality for the Beekmantown Group at East Beekmantown, New York, (the stratigraphic type locality is still eastern Shoreham) and Whitehall, New York (STOP 6).

STRATIGRAPHIC CONTROLS ON STRUCTURE

The structure of all thrust belts is controlled to a great extent by the vertical and horizontal arrangement of the various lithologies, and the thrust belt of the central Champlain Valley is no exception. In the Middlebury area, the shaly facies of the Fort Cassin Formation provides an easy gliding surface which was used as a detachment for the main deformation of the Providence Island and younger strata (locally, this detachment appears to lie within the Providence Island above the Burchards Member). Toward the southwest, however, the Fort Cassin pinches out, so this detachment horizon is no longer available. This is expressed in the Shoreham duplex by the abrupt migration of the glide horizon about 40 m. up into the Providence Island to the top of the series of limestone beds.

Between Middlebury and the Cornwall Ledges, the Providence Island and overlying strata are involved in a duplex (Washington, 1981a, b, 1982). The floor thrust, although not directly exposed, is quite evident at the Cornwall Ledges (Stop 1) where the Providence Island, except for some Burchards Limestone at the base, is highly deformed and imbricated, but the Bascom Subgroup is virtually undeformed except for some cleavage

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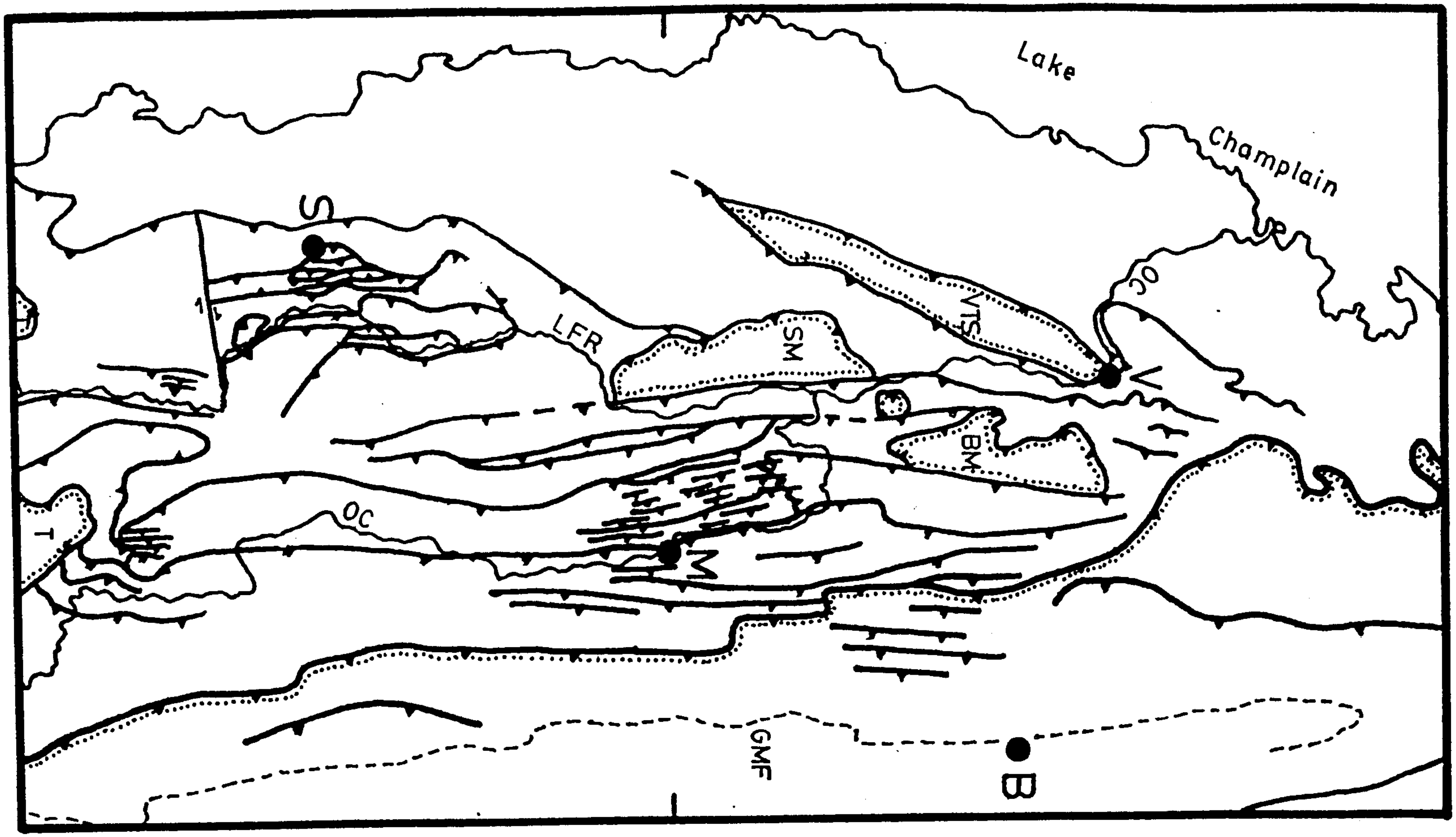

Figure 4. Structure map of the central Champlain Valley high-lighting the various portions of the Champlain thrust sheet. B - Bristol; M - Middlebury; V - Vergennes; S - Shoreham; VTS - Vergennes thrust sheet; SM - Snake Mountain massif; BM - Buck Mountain massif; T - Taconic allochthons; LFR - Lemon Fair River; OC - Otter Creek; GMF - Green Mountain front.

and minor drag folding in the less competent Fort Cassin strata.

The top of the Smith Basin Member of the Cutting Hill also serves locally as a detachment surface, with flats often occurring at this level. Very few of these flats extend for any great distance, however. Within the Shoreham duplex, the limestone pinches out and the associated flats also disappear

The boundary between the Cutting Hill and Whitehall formations acts as a regionally important detachment, serving as the base of the Orwell thrust sheet over large areas. It is not as important as the sub-Providence Island surface, as shown by the occurrence of ramps bypassing the sub-Cutting detachment, as can be seen at STOP 3. (No known ramps bypass the sub-Providence Island detachment where the Fort Cassin strata are well developed).

The final detachment within the strata originally assigned to the Beekmantown occurs in the Ticonderoga. Evidence of this surface is only seen in the Shoreham duplex (STOP 4) and near Vergennes (not directly exposed), and the exact stratigraphic level has not yet been determined.

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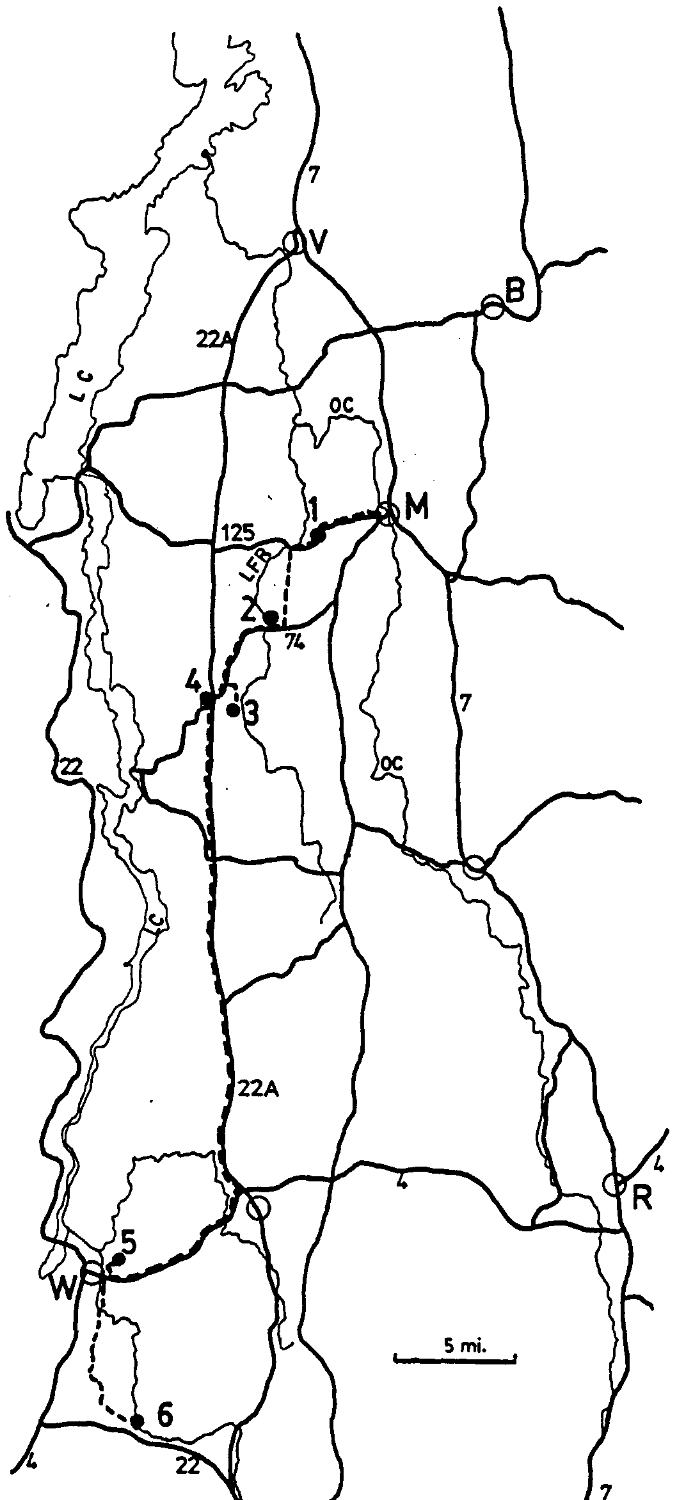

Figure 5. Route of field trip. M - Middlebury; R - Rutland; W - Whitehall, New York.

ROAD LOG

Mileages are cumulative from indicated starting points.
Roads are shown in Figure 5.

0.0 START at base of hill on College Street (Route 125) by Middlebury College Science Center. Go west on Route 125.

2.6 Park at roadside.

Stop 1. Cornwall Ledges

This is an excellent exposure of the Bascom Subgroup except for the lower part of Cutting Hill (Winchell Creek and lower East Shoreham). The top of the ledges contains deformed Providence Island within the duplex underlying the plateau between here and Middlebury. Note that the deformation is restricted to the strata above the Fort Cassin. The roadcuts and cliffs include a virtually complete section of Fort Cassin and Lemon Fair, the contacts with the overlying Providence Island (specifically the Burchards Member) and underlying Cutting Hill (Smith Basin Member) being exposed.

Continue west on Route 125.

4.6 TURN LEFT; if you cross bridge over Lemon Fair, you have missed turn. The very south end of the Snake Mountain massif, the southernmost point on the Champlain thrust as mapped by Cady [1945]), is just across Lemon Fair to the north.

7.9 TURN RIGHT onto Route 74.

8.7 **Please park across bridge** and walk back to farm road on north side of road.

Stop 2. Lemon Fair Cliff

This is an excellent example of the lithology that led Eaton (1824) to apply the term Calciferous Sandrock, the original term for the Beekman town. These strata are lower Lemon Fair (very near the contact with the Cutting Hill). This outcrop is in upper plate of Pinnacle thrust, trace of thrust runs along base of cliff.

Continue west on Route 74.

11.6 Potsdam Sandstone in field on right. Core of Shoreham anticline.

12.6 TURN LEFT onto dirt road.

12.9 SHARP RIGHT.

Stop 3. Scolithus Ledge

13.2 Park. Follow logging road into woods for about 1/3 mile (Fig. 6). Where logging road turns up hill into field, continue along hillside about 50 m. (downhill side of ledge).

A small cliff provides one of the few cross-sectional views through the "scolithus sandstone" (Winchell Creek of Fisher and Mazzullo [1976]). Note the alternation of scolithus burrowed and cross-bedded sandstones. That this is the base of the Cutting is shown by the Whitehall outcrops just down the hill to the southeast.

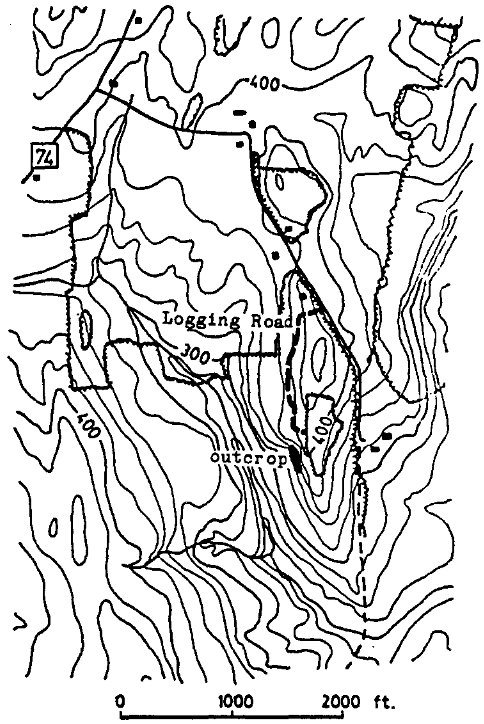

FIGURE 6. Location map for Stop 3, Scolithus Ledge.

RETURN to cars. TURN AROUND - dead end road.

CONTINUE.

12.2 TURN LEFT onto Route 74.

12.7 Stop, TURN LEFT onto Route 22A.

Stop 4. Shoreham Historical Society Building

13.0 Shoreham Historical Society Building. **LUNCH.** (Lunch fixings can be purchased at either of the two "corner stores" - one about 100 yds. north and the other about 200 yds south.

Just behind the building along the east bank of the little brook is an outcrop of the Shoreham thrust with middle Ordovician Crown Point Limestone thrust atop Ticonderoga Dolostone. Potsdam lies just to the west on the next little hill. The thrust is difficult to identify because of the lithologic similarity between the underlying Ticonderoga and the overlying Crown Point (indicating that sedimentary environments were very similar). This is the leading thrust of the Shoreham duplex, of which the rocks between here and the Pinnacle thrust are part. The Shoreham thrust breached the roof thrust and carried the leading edge of the duplex onto a portion of the roof sheet, producing a younger-over-older thrust very similar to the Weybridge thrust at Stop 1.

After lunch, GO SOUTH on 22A for 20.5 miles to Route 4.

0.0 RESTART MEASURED MILEAGE at entrance to Route 4.

7.1 TURN RIGHT at first road past Quarry. Quarry contains Cutting Hill.

8.0 TURN LEFT

8.2 TURN RIGHT

8.8 TURN RIGHT

9.2 PARK

Stop 5. Whitehall Quarry

This is the type locality for the Whitehall Dolostone and contains excellent examples of silicretes, algal mounds and mats, tidal channels, flat-pebble conglomerates, and assorted emergent features.

CONTINUE - turn around.

9.6 TURN LEFT

10.2 TURN LEFT

10.4 TURN RIGHT

11.3 Route 4 - TURN RIGHT

11.9 TURN LEFT at traffic light onto Upper Turnpike Road.

13.4 BEAR RIGHT (straight)

15.0 BEAR LEFT

17.0 You are presently riding along the Taconic frontal thrust surface. The hill slope is essentially the footwall thrust surface.

17.6 BEAR LEFT

18.0 BEAR LEFT

18.1 Park in lot.

Stop 6. Mettawee Falls

On the west side of the river below the falls is an excellent exposure of the Fort Cassin Formation. Included in this exposure are algal mats, dessication cracks, cauliflower nodules, ripples, and paleokarst features.

END OF FIELD TRIP

September 29, 1979

TACONIC GEOLOGY
NEAR FAIR HAVEN, VERMONT

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VERMONT GEOLOGY
VOLUME 5

VGS GUIDEBOOK 2

FIELD TRIP GUIDE G

VERMONT GEOLOGICAL SOCIETY
BOX 304
MONTPELIER, VERMONT 05602

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GEOLOGIC SETTING

The Fair Haven area (Fig. 1), west of Rutland, Vermont, is situated in the western part of the Taconic allochthon, which consists of thrust slices of folded deep-water argillites and associated interbeds of carbonate and terrigenous material. The assemblage is now emplaced on an ancient continental platform sequence of folded Cambrian and Ordovician carbonates that include three quartz-sandstone sheets. The youngest two of these can be correlated with the Taconic sequence, as discussed below.

The Taconic rocks have provoked a variety of controversies. The history of the older ones is outlined by Zen (1967). The recent geologic interests in the allochthon include: 1) establishing the stratigraphy of the Taconic sequence, 2) mapping the structure, and 3) determining the timing and method of emplacement of the Taconic allochthon. W.S.F. Kidd and his students have re-mapped and up-dated the work of Zen (1961) (for example, see Fisher, 1984). Most recently, Stanley and Ratcliffe (1985) have presented a thorough and exhaustive synthesis of the structural evolution of the Taconic allochthon.

The Fair Haven area is in the westernmost of the several thrust sheets that comprise the Taconic allochthon, and the rocks here are least metamorphosed -- probably because they suffered the least tectonic burial. The well-displayed sedimentary features provide information on the source areas, environments of deposition, and even the nature of the Taconic orogeny. The commercially produced green and purple slate of the Fair Haven area occurs about in the middle of the Taconic sequence.

This guidebook is a major modification of the one prepared for the 1979 VGS field trip. It is an expansion of the guidebook prepared for DNAG (Baldwin and Raiford, 1987).

Figure 1 of this report was compiled mostly from Rowley (1983). The part southeast of Fair Haven is from Wright (1970), and the distribution east of Lake Bomoseen is inferred from Zen (1961). Figure 1 maps the oldest black slate (Browns Pond Formation) and the youngest green slate (Middle Granville Formation) as a single transition unit that separates the main mass of green and purple slates from the overlying black and gray slates.

Most of the stops are associated with the Mount Hamilton syncline and the West Castleton syncline (Fig. 1, Cross-section A-A').

GEOLOGIC EVOLUTION

REGIONAL TECTONICS

The stratigraphic age of the Taconic sequence ranges from 650 Ma, with the initiation of rifting, through the opening and closing of the proto-Atlantic ocean, until about 450 Ma, with the onset of the Taconic orogeny. The orogeny was due to collision of the proto-North American continent with an offshore volcanic arc (Chapple, 1973; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985).

The Taconic strata originated as deep-water sediments, deposited on the continental slope and rise on the east margin of proto-North America. With the closing of the proto-Atlantic ocean (Iapetus), these sediments (and the accretionary prism to the east) moved onto the continental platform.

Our understanding of how this happened has matured in the last few years. Zen (1961) and Bird and Dewey (1975) inferred that the Taconic sediments were emplaced on the platform sediments by gravity sliding westward from an uplifted belt. Baldwin (1971) had inferred that the sediments were not yet buried tectonically and so were not yet lithified.

Now, it is evident that the sediments of the Taconic allochthon were emplaced on the platform sequence by obduction, as the leading edge of the continent entered the east-dipping subduction zone. Rowley and Kidd (1981) concluded that the Taconic thrusts stacked successively westward on the ocean floor and then the pre-assembled stack was emplaced on the continental platform. Stanley and Ratcliffe (1985) used data from a larger region to document a continuing sequence of deformation and emplacement; this included renewed movement on some faults.

The following outline of events is based largely on Stanley and Ratcliffe (1985). Rifting had stretched the continental crust, so that in the post-rifting stage, the continental slope and rise were developed on attenuated continental crust. The oldest Taconic sediments, not seen around Fair Haven, are rift facies, whereas those in the Fair Haven area are deep-water sediments (Stop 8) that accumulated on the slope or adjacent rise. Closing of the ocean basin may have started in the Late Cambrian, but the accretionary prism and volcanic arc still lay some distance east of the site of Taconic sedimentation. In time, these sediments, scarcely 1500 m thick and still not lithified, began to underplate the accretionary prism, and some tectonic stacking occurred. Soft-sediment deformation (Stops 2, 7) probably dates to this time. Meanwhile, the continental platform began to subside rapidly and muds (the Hortonville Slate) accumulated on former shallow-water sediments (Baldwin, 1983).

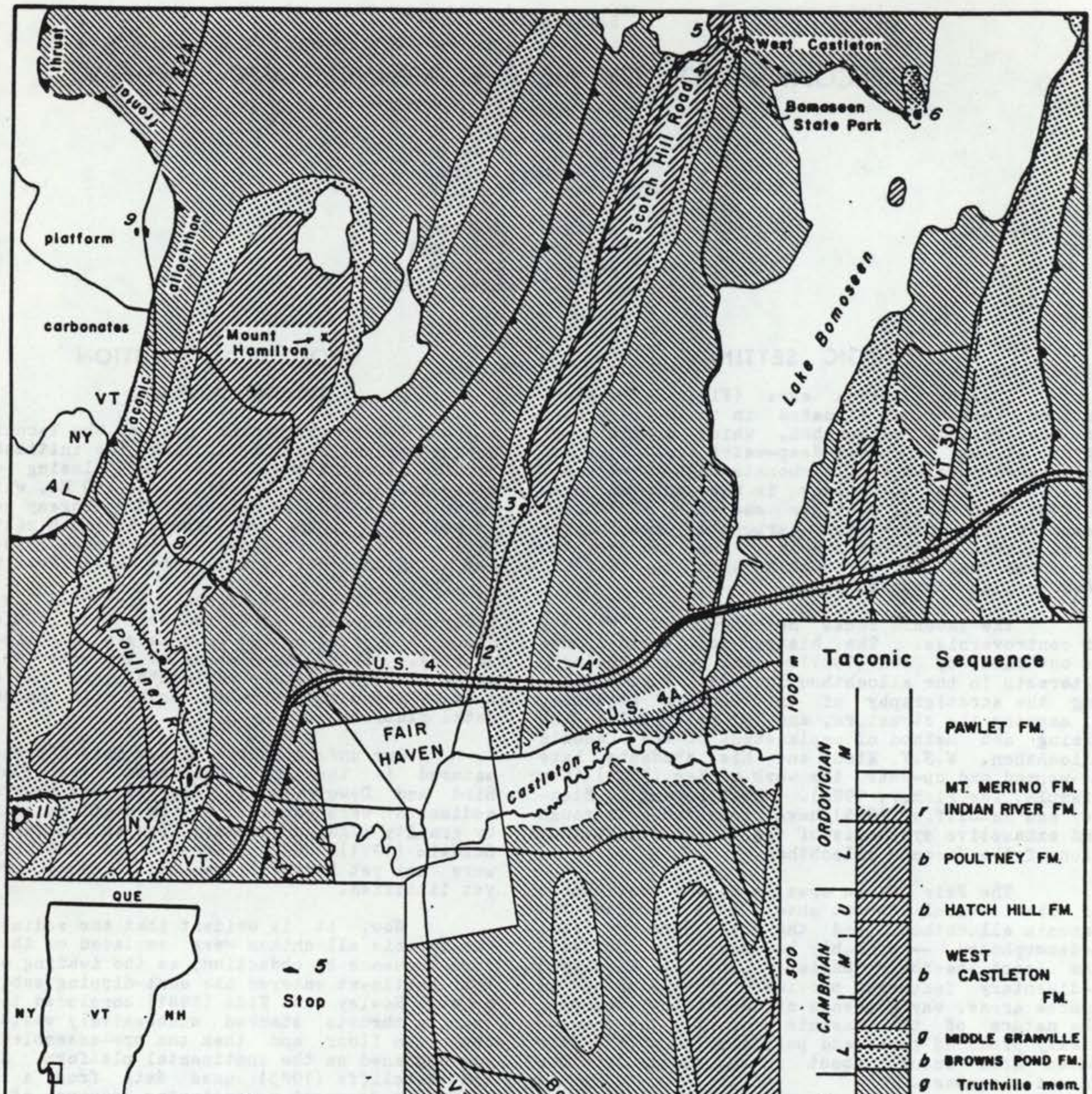

FIGURE 1. Geologic map of the Fair Haven area (modified from Rowley, 1983; Zen, 1961; Wright, 1970). In cross-section A-A', MH, Mount Hamilton syncline; WC, West Castleton syncline.

Soon the stacking dewatered and lithfied the Taconic sequence. This preserved the primary sedimentary features of the Giddings Brook slice, so well displayed at Stop 8. Now more rigid, the Taconic sequence, with some overlying slices, moved onto the platform because of the push from the larger stack to the east. This emplacement of the Giddings Brook slice, about 445 Ma (Baldwin, 1982), was marked by deposition of clasts as Zen's "wildflysch-type conglomerate" or Forbes Hill conglomerate in the Hortonville muds on the former continental platform (Stop 9).

With continued compression, according to Stanley and Ratcliffe (1985), the Giddings Brook slice developed nearly recumbent folds with axial surface slaty cleavage (Stops 1, 5, 6). To the east, deformation brought slivers of platform carbonates into the Taconic pile (the Group 2 slices of Stanley and Ratcliffe, 1985). The next slices (Group 3) show the effects of syntectonic metamorphism that post-dates the main emplacement of the Group 2 slices. Finally, the continued deformation involved the Precambrian basement in such faults as the Champlain thrust. This phase caused folding of the Middlebury synclinorium and warping of the Taconic allochthon. The frontal thrust of the Taconic allochthon (Fig. 1) represents this late-stage deformation that folded the emplacement thrust and dragged with it some parautochthonous slices of platform carbonates and overlying shales.

Later, the underthrust edge of the continental plate rose isostatically (Rowley, 1980; Baldwin and Butler, 1982) creating the mountains of the Taconic orogeny.

STRATIGRAPHY

The strata of the Taconic sequence are believed to be equivalents of strata found east of the Green Mountain anticlinorium (Stanley and Ratcliffe, 1985), but the Taconic sequence has its own array of stratigraphic names (Theokritoff, 1964; Zen, 1964).

The lower half of the Taconic sequence is greenish and purplish in color, whereas the upper half is black and dark gray. If these colors extended eastward across the ocean floor, it would imply that the older sediments east of the Green Mountain front were also green and the younger sediments were black. That, in turn, would suggest that the central Vermont repetition of green and black argillites is due to tectonic duplicating of a simple pair of sedimentary sequences.

Table 1 adopts the nomenclature of Rowley (1983), though the Hatch Hill Formation is used here in a narrow sense, as noted below. As Rowley (1983; Rowley and others, 1979) demonstrated, there is a brief stratigraphic flip-flop (black up to green and purple) between the underlying green and purple slates and the overlying black slates; here, these beds are labeled "transition zone". Figure 1 shows the approximate structural (outcrop width) thickness. Because of thinning on the limbs of folds, the stratigraphic thickness could be twice as much (Rowley, 1983).

ORIGINS OF SEDIMENTS

The emphasis in this field guide is on how the sediments originated. The sediments include a variety of conglomerates, a variety of colors, and a variety of lithologies.

Three distinct kinds of deep-water conglomerates occur, and field evidence provides a basis for inferring their different origins. 1) A limestone "conglomerate" (Stop 2) in the middle of the sequence (in the Browns Pond Formation), is really an interval of black slates and thin limestone beds (Stop 3). The soft-sediment disruption of the beds into clasts, some of which are bent, occurred before the deep-water sediments were lithified, and so this evidently happened before the sediments were thrust onto the continental platform. 2) In the Hatch Hill Formation, a dolomite-cemented quartz sandstone, with abundant sandstone clasts to 20 cm, was derived from continental platform sediments and accumulated as a 4-meter bed (Stop 8; Fig. 3). It, and associated turbidites, formed a submarine fan (Baldwin, 1971, p. 298). 3) In the autochthonous Hortonville Slate (= Normanskill), a "wildflysch-type conglomerate" (Stop 9) is debris shed from the Taconic sequence, as the allochthon advanced westward across the mud-covered subsiding platform.

The several types of sediment (Table 1) pose questions on their source and method of deposition. What caused changes in oxidation levels of the ocean floor? Green and purple slates of the Bull Formation, which characterize the lower half of the Taconic sequence, represent non-reducing conditions. Soon after the start of the Cambrian, ocean-floor conditions became reducing, as indicated by black slates of the West Castleton to Poultney formations, which were deposited through the Early Ordovician. Moreover, the change from green-purple to black was not unique, because there is a transition zone in which the first black slates (Browns Pond Formation) are overlain by more green-purple slates (Middle Granville Formation).

Which material settled through ocean water, and which flowed down the continental margin as density currents? The Taconic muds accumulated at scarcely 10 mm/m.y. (solid-grain thickness). This figure is modified from Baldwin (1983), who had estimated the rate to be 4 mm/m.y. The modification is based on Rowley (1983), who noted that the outcrop width of vertical beds is a structural thickness that may be less than half of the stratigraphic thickness. Some mud turbidites are recognized, but the accumulation rate was most probably controlled by the supply of mud particles to the continental slope and rise (Baldwin, 1983).

What was the source of the limestones that are interbedded with black slate in the Browns Pond Formation and in the West Castleton Formation? Graded bedding is rarely if ever seen in the limestones; some of the limestones are interbedded with green slate. The Browns Pond is 50 feet thick in some places and only 15 feet thick in others (Campbell, 1983); Wright (1970) showed it to be absent in places north of the Eureka quarry. Are the limestones local turbidites that mostly coincided with deposition of the black slates? If so, why the coincidence?

In the Taconic argillites are interbeds of dolostone, some associated limestone, and quartzose sand. These strata pose fewer questions. They are turbidites and obviously formed submarine fans on a continental slope or rise. The Mudd Pond quartzite of the Nassau Group is correlated with the Cheshire Quartzite of the platform sequence (Zen, 1967).

In the post-Nassau section of the Taconic sequence described by Baldwin (1983), only two prominent intervals of interbeds occur. Admittedly, half of the total section is covered. None-

Table 1. The Taconic Sequence
(Modified from Rowley, 1983)
Names of formations seen on these trips are **boldfaced**

WILLARD MOUNTAIN GROUP
Pawlet Formation: graywacke and black slate
Middle Ordovician

Mt. Merino Formation: black slate, dark chert

MOUNT HAMILTON GROUP
Indian River Slate: orange-red and green slate .
Middle Ordovician (Chazy)

Poultney Slate: black and dark gray slate, with some interbeds of dolostone and dolomitic quartzite
Upper Cambrian to lower Middle Ordovician

Hatch Hill Formation: quartzites and black slate
Upper Cambrian

West Castleton Formation: black slate with some limestone interbeds
Lower and Middle Cambrian

"transition zone"
Middle Granville Slate: purple and green slate; quarries;
Lower Cambrian

Brown's Pond Formation: black slate, siltstone, thin limestones (North Brittain "conglomerate" of Zen, 1961)
lowermost Cambrian

NASSAU GROUP (includes also Biddie Knob Formation and Rensselaer Grit)
Bull Formation: green and purple slate (Mettawee facies) with sandstones and limestones:
Truthville slate member: green, some purple slate; quarries; this is the Mettawee above the Bomoseen
Mudd Pond quartzite (= Cheshire Quartzite of the platform sequence)
Bomoseen member: graywacke, grit
Zion Hill member: quartzite and graywacke
Eocambrian to lowest Cambrian

theless, the very slow accumulation rate of the argillites, as noted above, suggests that the rate was fairly constant. Assuming that this is the case, the two sets of quartzose interbeds correlate satisfactorily with two quartz-sand sheets of the platform sequence. Thus, it is inferred that the distal quartz sands of the lower Middle(?) Cambrian Monkton Quartzite fed turbidites of the West Castleton Formation, and those of the Upper Cambrian Danby Formation fed turbidites of the Hatch Hill Formation (Baldwin, 1983).

It should be noted that Rowley (1983) and Fisher (1984) use the Hatch Hill in a broader and less discriminating sense -- a Lower to Upper Cambrian formation that includes the West Castleton as a member.

The correlation of quartzose deep-water and platform sediments seems to justify the assumption that argillite accumulated at a fairly constant rate. If so, the interbeds were deposited infrequently -- every 25,000 to 6,000 years (Baldwin, 1983).

Rowley (1979) had suggested that the quartz-sand sheets prograded during times of sea-level lowering. However, Mehrtens (1985) showed that the carbonate and terrigenous (quartz-sand) sediments had the same environments of deposition in the same places, meaning that the prograding terrigenous sheets reflected increased supply rather than eustatic lowering.

Using a different model, quartzose interbeds in the fans would not correlate with particular formations on the platform (C.J. Mehrtens, personal communication). Instead, quartz sands moved in tidal channels to the shelf-slope break, bypassing the dolostone of the shallow-water platform. The sands spilled over at irregular intervals to form turbidites on the continental slope and rise. In this model, argillites with interbeds were deposited as interchannel facies of the turbidites; thus, these argillites would not have accumulated at a constant rate.

The limestone "conglomerate" of the Browns Pond poses a different question. In the Fair Haven area, the "clasts" are disrupted beds of limestone, commonly related to west-verging folds and thrusts (Campbell, 1983). Some clasts are bent, so the disruption probably coincided with soft-sediment deformation seen in some black slates. The solid-grain thickness of the Taconic sequence is only 1000 or 1500 m, so that the solidity (complement of porosity) was only about 80 percent (Baldwin and Butler, 1985) before tectonic burial. That leaves enough pore water to permit soft-sediment deformation during early-stage stacking of the thrust sheets. But the stacking evidently dewatered and lithified the Taconic sediments before initial emplacement on the platform.

A similar question of pore water relates to another lithology in the Browns Pond Formation. Round unstrained clear quartz sand grains float in a massive gray muddy siltstone, which also has floating clasts of fine-textured black slate. This is the "calcareous quartz wacke" of Rowley (1983). The grains and clasts may have become mixed during submarine flowage of continental rise sediments, either during sedimentation or during stacking.

The Indian River Formation, exposed in this area on Mt. Hamilton, is a Chazy-age orange-red slate that probably represents erosion of soils (terra rossa) formed during up-bulging of the continental platform just before the margin entered the subduction zone (Bird and Dewey, 1970). The Pawlet Formation, seen by the bridge just upstream of Poultney (on Vermont 30, about 7 miles south of U.S. 4), Vermont, consists of slates and graywacke turbidites; the latter are immature sands derived from an arc-related source to the east (Rowley and Kidd, 1981, p. 207).

ACKNOWLEDGMENTS

Charlotte Mehrtens and Rolfe Stanley, of the University of Vermont, patiently questioned some assumptions of an earlier version, and we have endeavored to accommodate their thinking in our work. Lucy Harding, of Middlebury College, was a helpful editor.

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ROAD LOGS

(Estimated time away from vehicles is given for each stop.)
(Though there are excellent outcrops on U.S. 4 between Vermont 30 and Vermont 22A, parking on U.S. 4 is illegal.)

Each trip starts at the village green in Fair Haven (see map). TRIP A (Stop 1) is to the Eureka slate quarry, southeast of Fair Haven. **Phone for permission**, Vermont Structural Slate, 802+265-4933. TRIP B (Stops 2 to 6), to West Castleton, shows the Browns Pond disrupted limestones, the type locality of the West Castleton Formation, and two localities where synclines can be photographed. TRIP C (Stops 7 to 9), northwest on VT 22A, includes several localities to show sedimentary features -- soft-sediment deformation, excellent exposures of slates with turbidite interbeds, and a small outcrop of wildflysch. TRIP D (Stops 10 to 12), are west of Fair Haven, to see a mud turbidite, the window in the thrust sheet at William Miller Chapel, and the contact between parautochthonous carbonate rock and autochthonous slate with clasts (wildflysch). Stop 12 is off the map (Fig. 1).

TRIP A, EUREKA QUARRY

Mileage

0.0 Leave the southeast corner of the village green and drive south; go up the hill (0.2 miles) on VT 22A; U.S. 4A bears right.

0.7 Somewhere Restaurant on the left.

1.3 Turn left (east) onto Bolger Road.

2.9 Take the left fork.

3.1 Park near the entrance to the Eureka quarry of Vermont Structural Slate.

Stop 1. Eureka Quarry 20 minutes

(**Get permission** from Vermont Structural Slate, 802+265-4933. Only visit this during working hours; **do not trespass.**)

Check in at the splitting shed (nearest the road). If the foreman is available, visit the inactive (northern) part of the upper (eastern) quarry. It is usually possible to watch the slate-splitting operation.

The Eureka quarry exposes the east-dipping, upright sequence of the upper part of the Nassau Group and the lower part of the Mount Hamilton Group (Fig. 2). The main quarry is in the Truthville slate member of the Bull Formation, and the Browns Pond Formation constitutes the top 15 feet of the main quarry wall. This part of the section must be viewed from the north rim, near the splitting shed. The orange-red slate in some pallets nearby is from the Indian River Formation, from the Granville area in New York. Some loose blocks of purple slate show green ellipsoids, which represent reduction spheres that were later stretched (Wood, 1974).

In the upper quarry, the lower 20 or 30 feet is the Middle Granville Formation. The upper half of the upper quarry face is black slate and thin limestone of the West Castleton Formation. A fault, about parallel to the contact, extends along the quarry face.

Return to northeast corner of village green; cross railroad tracks slowly.

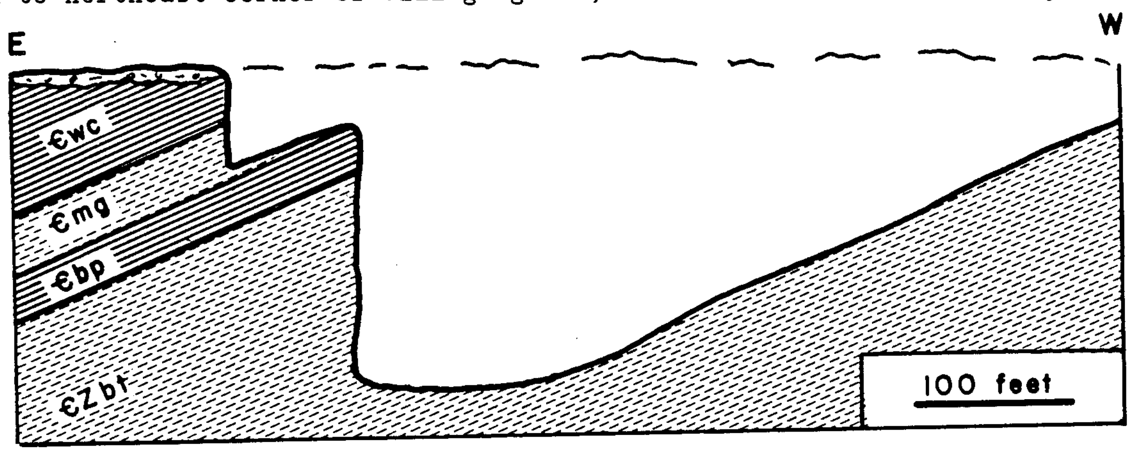

FIGURE 2. Stratigraphy of the Eureka Quarry. Ꞓwc, West Castleton Formation; Ꞓmg, Middle Granville Formation; Ꞓbp, Browns Pond Formation; ꞒZbt, Truthville member of Bull Formation.

TRIP B, SCOTCH HILL ROAD

0.0 Continue north on U.S. 4A.

0.5 At the blinker light, bear left (north) onto Scotch Hill Road; U.S. 4A turns right. Cross the overpass above U.S. 4.

1.1 Park on the east shoulder, across from the long low outcrop.

Stop 2. Limestone "Conglomerate" 20 minutes

At the north end, the outcrop shows interbedded black slate and limestone. For most of the road cut, however, the limestone is disrupted into a "conglomerate", with some bent clasts of limestone. This is Zen's (1961) North Brittain conglomerate, and is part of the Browns Pond Formation.

Continue north on Scotch Hill Road.

1.9 (from the village green) park on a left curve near a driveway to the east.

Stop 3. Ribbon Limestone 15 minutes

Walk 100 feet northwest on the north side of a slate pile, and then 150 feet north to the south side of a deep abandoned quarry. At the top of the distant east wall, the 10 feet of ribbon limestone and black slate of the Browns Pond is here not disrupted into a "conglomerate". In the loose quarry blocks of purple slate, note the green (reduced) ellipsoids about an inch across; these have been used to determine the amount of bulk strain within these deformed rocks.

Continue north on Scotch Hill Road.

4.7 Park at a turnout on the left.

Stop 4. Type Section of West Castleton Formation 10 minutes

This section, along the road and also continuing up the ridge, farther south, was described by Mertz (1969) and by Garver (1983). Painted unit 9 (Mertz, 1969) includes limestone turbidites. The abandoned quarry just to the west of the road is in the Middle Granville Formation (Rowley, 1983). Garver (1983) found 66 m of green and purple slate in the Poultney Slate, above the West Castleton black slates.

4.9 Park on the right; this is the former community of West Castleton.

Stop 5. West Castleton Syncline 10 minutes

This is **Private Property** and the owner (Roy Whitman, Glen Lake, VT) does not want the outcrop defaced. **Please get permission** from the house across the road; **no hammers**. The south-facing cliff exposes a much-photographed fold in slate and dolomitic slate. Zen (1961, Pl. 3, Fig. 2) identified this as the West Castleton Formation; according to Mertz (1969) and Rowley (1983), this is his Mount Hamilton Formation (Poultney Slate).

Continue eastward.

5.05 Road to left leads north; bear right.

5.2 Near entrance to Bomoseen State Park, turn left onto dirt road to the Cedar Point quarry.

6.5 Park.

Stop 6. Cedar Mountain Quarry 40 minutes

Walk up the trail onto the waste pile of the Cedar Point quarry (abandoned) in the Middle Granville Formation. Note the limestone "conglomerate" of the Browns Pond in the cliff. From the top of the highest pile, there is a good view of underside of a syncline (Zen, 1961, Pl. 3, Fig. 1) in the Middle Granville purple slates.

Return to car, drive back to Fair Haven village green.

TRIP C, VERMONT 22A

0.0 Northwest corner of the village green. Drive northwest on Vermont 22A, past the U.S. 4 overpass.

1.3 Zen (1961) mapped the roadcut as Bomoseen grit, and the thin quartzites as Mudd Pond quartzite member (= Cheshire Quartzite of the platform sequence).

1.7 Park on the right shoulder just past Sheldon Road.

Stop 7. Browns Pond Formation 20 minutes

The 300-ft outcrop is the Browns Pond Formation, occurring on the overturned east limb of the Mount Hamilton syncline. Pacing from the southeast end, several lithologies are seen: 0 to 90 ft, soft-sediment deformation, including floating clasts of black slate and clasts of folded slate; near 65 ft, clear black grains of round unstrained quartz floating in a gray muddy siltstone; at 85 ft, a disrupted sandstone; 90 to 230 ft, black slate, with a sandstone at 150 ft; 230 to 270 ft, the limestone conglomerate; 270 to 300 ft, green-gray slate.

The material at 65 ft. is the "calcereous quartz wacke" of Rowley (1983). The grains and clasts may have become mixed during submarine flowage of continental rise sediments, either during sedimentation or during stacking.

Zen (1961, Pl. 2, Fig. 2) showed a low ledge on the northeast side of the highway (destroyed when the highway was widened). The photograph is on its side(!); the hammer handle is down, and the overturned limestone beds and "clasts" dipped southeast.

Continue northwest on Vermont 22A.

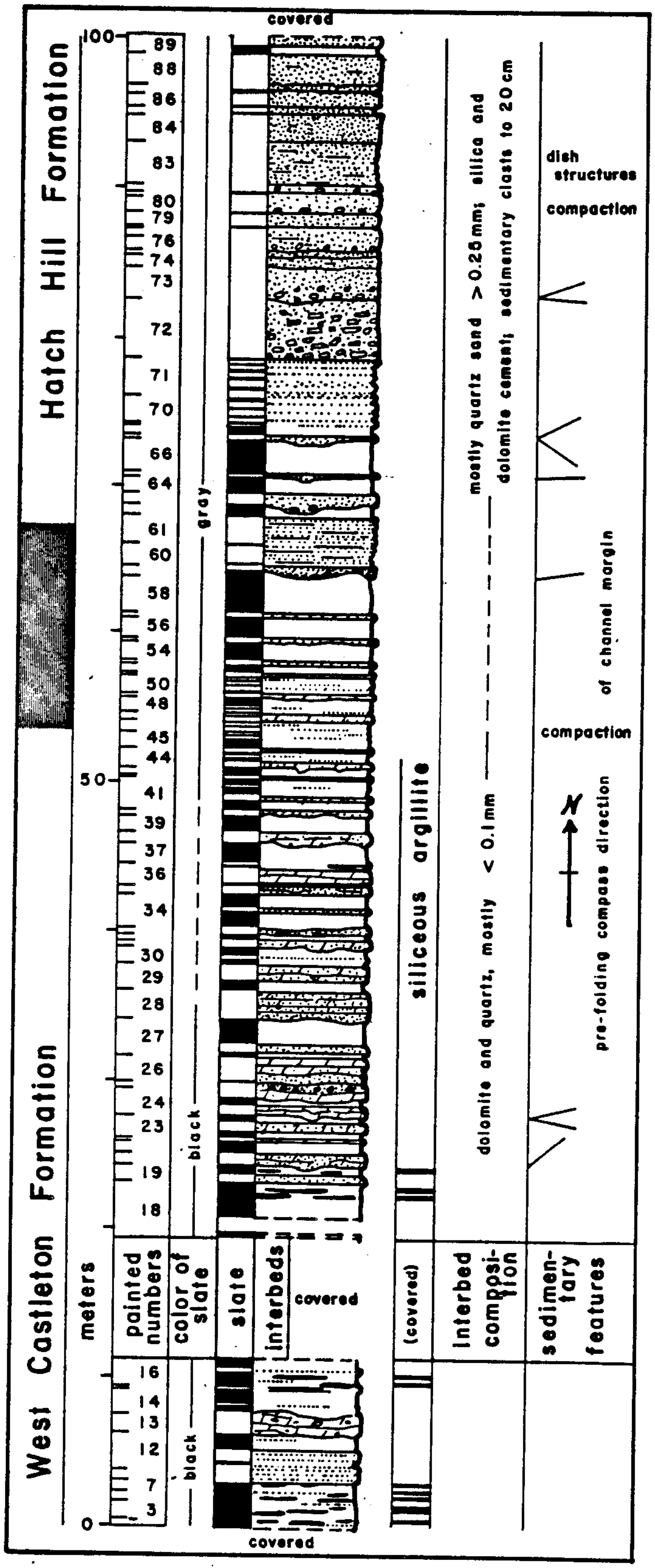

FIGURE 3. Strata exposed at Stop 8, on Poultney River.

1.9 Just before the next right-hand curve, park by the Edward J. Morris farmhouse and **ask permission.**

Stop 8. Deep-Sea Fan 90 minutes

The 0.5-mile walk from the highway takes 10 minutes. Leave the gates the way you find them -- open or closed. Go south through the gate to the bottom of the hill. Continue south in the pasture, going up the hill on a farm trail, between two fences. Go through a gate and up to the knoll where a transmission line crosses. From here, leave the trail; continue due south, obliquely to the left down the slope, to the north end of several evergreens. From here, scramble down the open 40-foot slope eroded in glacial deposits, to the extensive outcrops by Poultney River.

This is Section 3 of Baldwin (1983); the painted numbers identify arbitrarily selected units. **Please do not sample near the painted numbers.** The beds are about vertical and strike N10E.

Figure 3 illustrates important features of the 330-ft section of glacially polished vertical beds. Above unit 19, the slate and interbeds represent a submarine fan (Baldwin, 1971); unit 72 is a channel fill. The decompacted geometry of units 46 and 80 was illustrated by Baldwin (1971).

The source of the carbonate and quartzose interbeds lay to the west. The quartz sandstone and dolostone clasts in unit 72 and other channel breccias must have come from a stable continental platform. Channel margins are within 40° of vertical; if the vertical beds are unfolded about a horizontal cylindrical axis, the channel margins are rotated into an angle within about 40° of S80E (Fig. 3). On the river bank, an interbed in unit 71 shows cross-bedding directions oriented upward (and so eastward).

The West Castleton/Hatch Hill boundary is here put somewhere between units 45 and 61. Below this boundary, the interbeds have much dolomite and the quartz is fine (Fig. 3). The upper interbeds have abundant medium quartz sand, and these are taken as the off-shelf equivalent of the Upper Cambrian Danby Formation (Baldwin, 1983). Thus, they are Hatch Hill Formation and the lower beds are the West Castleton Formation; overlying beds in outcrops downstream are of the Poultney Slate. Curiously, Theokritoff (1964) named the Hatch Hill Formation but did not recognize its presence here. As noted earlier, others (Rowley, 1983; Fisher, 1984) use the Hatch Hill in a broad and less discriminating sense to include the West Castleton unit, because their work did not establish a mappable distinction between the two.

Return to vehicle and continue north on Vermont 22A.

2.2 Roadcut shows slates with thin graded beds (Zen, 1961, Pl. 2, Fig. 5).

2.6 Slate quarry, now mostly filled in, on right. On southeast side, the roadcut exposes the boundary between green and purple slate, overlying black slate mapped by Rowley (1983) as the Mettawee - Browns Pond contact.

3.6 (north of the village green) the West Haven road joins from the left.

3.8 Ordovician limestone with some black slate (fault slices?) forms the roadcut on the west side.

4.0 Where the road widens to three lanes, park on the left (west side) in the parking area.

Stop 9. Wildflysch-Type Conglomerate 25 minutes

This locality is just beneath the base of the Giddings Brook thrust sheet. At the "No Littering" sign at the south end of the parking area, step over the fence; walk west 100 feet down the faint trail, and then northwest through brush 80 feet to the outcrop in the clearing. This is a "wildflysch-type conglomerate" in the autochthonous Hortonville Slate (mid-Ordovician; Trenton Group). The type section of this Forbes Hill conglomerate (Zen, 1961) is in the northwest corner of Figure 1; Zen inferred that the clasts of foreign rocks fell off as submarine slumps when the Taconic sheet slid west onto the muds of the former platform.

The rocks in the roadcut across from the parking area are fault slices in the upper part of the platform-sequence limestones and shales. This deformation occurred during the final movement along the Taconic frontal thrust (Bosworth and Rowley, 1984; Stanley and Ratcliffe, 1985). Bosworth and Kidd (1985, p. 129-131, Fig 9) describe this in some detail.

The limestones and slates are fossiliferous (bryozoans, trilobites, brachipods, pelmatozoan stems). Toward the north end, some slates have quartzite clasts. Other slates have limestone "conglomerate".

Return to Fair Haven village green.

TRIP D, WEST OF FAIR HAVEN

0.0 Northwest corner of village green. Drive west on West Street.

0.8 Cross over U.S. 4.

1.0 Stop at farm house on left (R.L. Weatherbee) **for permission** to go to the outcrops of Stop 10.

1.3 Park on left just before the bridge across Poultney River.

Stop 10. Mud Turbidites 20 minutes

Walk up the road about 200 feet and then go into the field to the north. The large glaciated outcrop of light gray, greenish gray, and reddish tan slate has thin (1 cm) bands of white chert (Fig. 4). Just west of the low area, and just east of a small patch of black slate (West Castleton Formation), there is a 15-cm cluster of white chert clasts (X in Fig. 4) that represent a lag "gravel" at the base of a mud turbidite about 25 cm thick. Stanley (personal communication) named these "unifites" (Stanley, 1981).

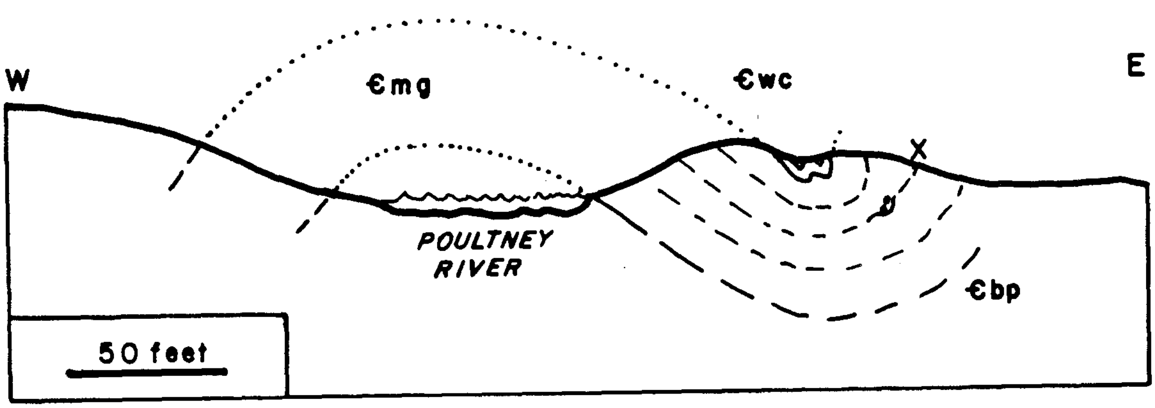

FIGURE 4. Stratigraphic relations at Stop 10. €wc, West Castleton Formation; €mg, Middle Granville Formation; €bp, Browns Pond Formation; X, chert lag pebbles.

Continue west into New York State.

1.3 (west of the village green) cross Poultney River into New York State.

1.5 Road to the north; continue straight.

1.7 Road to the south; continue straight.

2.0 Road bends left.

2.3 Park on left at William Miller Chapel

Stop 11. William Miller Chapel 20 minutes

Walk south on the path on the east wall of the chapel. About 150 feet in, the path is at the base of the surrounding slate. The clearing 200 feet in is a window in the thrust sheet, exposing the limestone underneath the Taconic frontal thrust sheet. The cream-colored, grooved rock is dolostone. The dolostone is probably a slice under the frontal thrust.

Continue west (off Fig. 1).

3.6 Turn sharp left (southwest) at road junction. The roadcuts expose Isle la Motte Limestone and slate with clasts (wildflysch).

4.6 Continue straight past a road that branches to left.

4.7 The outcrop is on the right, Continue down the hill.

4.8 Park at turnout on right, walk back up hill.

Stop 12. Wildflysch and Carbonate Rocks 20 minutes

Carbonate strata at the top of the outcrop show duplex faulting. The carbonate rests on slate with clasts (wildflysch). Thus, the outcrop shows the base of the parautochthonous carbonate slivers that are dragged across the autochthonous wildflysch by the Taconic frontal thrust. Bosworth and Kidd (1985, p. 124-127) identify the carbonate as the Middlebury Limestone.

Return to Fair Haven or continue on road about 3 miles into Whitehall.

October 24, 1981

PLEISTOCENE GLACIATION AT LAKE WILLOUGHBY, VERMONT

VERMONT GEOLOGY
VOLUME 5

Ballard Ebbett
Lyndon State College
Lyndonville, Vermont 05851

VGS GUIDEBOOK 2

FIELD TRIP GUIDE H

STOP		PAGE

VERMONT GEOLOGICAL SOCIETY
BOX 304
MONTPELIER, VERMONT 05602

PLEISTOCENE GLACIATION AT LAKE WILLOUGHBY, VERMONT

Ballard Ebbett
Lyndon State College
Lyndonville, Vermont 05851

INTRODUCTION

The southeast trending Lake Willoughby rock basin occurs within a granodiorite pluton (Wg) and the adjacent schist (Dw or Dg) (See Fig. 1). The Vermont Bedrock Geologic Map (Doll and others, 1961) shows that most lakes in northeast Vermont occur in granitic plutons near or at the contact with the surrounding schist. Three observations may help explain this relation between bedrock and lakes. (1) Fractures in granitic rock seem to be more closely spaced near granite-schist contacts than they are elsewhere in northeast Vermont. (2) In most glaciated regions of the world, granite is more susceptible than schist to glacial block removal (Embleton and King, 1975, p. 197-198). (3) The deepest lake basins surveyed by Mills (1951) occur where one set of fractures more or less trends in the same direction as the flow of the last ice sheet at its maximum extent. (See Hughes and others, 1985, for surface flow directions in the Last Great North American Ice Sheet.)

The glacial geology of New England, according to Hughes and others (1985), reflects flow patterns and basal thermal regimes of former ice sheets during their maximum extent, which they reached up to seventeen times over the last two or three million years (Chorley and others, 1984, p. 510), as well as the pattern of disintegration of the last ice sheet, which broke up 14,000 years ago in the Willoughby region. Hughes and others (1985) suggest that quarrying within a basal freezing zone of the Laurentide Ice Sheet created an "arc of exhumation" (White, 1972), which is reflected, in part, by the ring of deep lakes and estuaries that surround the Canadian Shield. Radial extensions of the "arc of exhumation" follow major stream valleys that were more or less parallel to glacial flow, e.g., Hudson-Champlain Valley and Gulf of St. Lawrence. Along such major valleys, ice streams of faster flowing ice caused generally greater glacial erosion than slower moving ice in the intervening regions. Willoughby Gap may be a segment of a minature radial extension of the "arc of exhumation".

The pattern of disintegration of the last ice sheet in the Willoughby region may have been similar to that in the Androscoggin Valley region north of the Presidential Range in New Hampshire as described by Gerath and others (1985). As the south-southeast flowing Laurentide Ice Sheet thinned, it separated over the northeast-trending Willoughby Range. This led to ice stagnation southeast of the range. During the last stages of ice disintegration, an active valley ice lobe, sustained by regional ice in the upper St. Francis River basin, extended through Willoughby Gap, across the regional drainage divide in the valley just southeast of Lake Willoughby, and down the West Branch of the Passumpsic River to West Burke Village. Eventually the ice over the valley divide became too thin to nourish the portion of the backwasting valley ice lobe southeast of the divide. This led to stagnation and downwasting. A similar sequence of events in the northern Appalachian Plateau is described by Fleischer (1986). The relation of the active front of the ice lobe to the detached stagnant portion fits Koteff's (1974) fluvial non-ice-contact morphologic sequence (morphosequence). As the active valley ice lobe continued to backwaste through Willoughby Gap, proglacial Lake Willoughby formed between the ice margin and the valley divide (Stuart and MacClintock, 1969, p. 143).

ACKNOWLEDGMENTS

Former students of Lyndon State College who assisted with the mapping of these landforms and the search for their history are Tim Gomo, Dave White, Doug Reilly, Dean Herdman, Mike Powell, and Bernadette Cooney.

Herbert Hawkes is the first person that I know who referred to the till ridges at Stop 5 as moraines.

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Figure 1. Topographic and geologic map of the Lake Willoughby area showing locations of field trip stops and portions of the successive margins of a backwasting valley ice lobe. Base map is taken from U.S.G.S. topographic contour maps of 15 1/2 Lyndonville and Memphremagog Quadrangles, 1951 and 1953. Bedrock within area enclosed by hachured lines is mostly Willoughby granite (Wg). Bedrock outside area enclosed by hachured lines is mostly pelitic schist (Dg) and calcareous schist (Dw). Field trip stops are labeled 1 through 8. Contour interval = 20 feet.

20 Dip and strike of foliation in Dw or Dg

⊕ Horizontal foliation

Glacial Striations.

Friction Cracks.

7 Proposed Valley Ice Lobe Margins. Numbers refer to till ridges at STOP 5.

E E′ Location of topographic profile.

Meltwater Channels. adjacent to till ridges 3, 6 and 7.

Mount Pisgah or Wheeler trail.

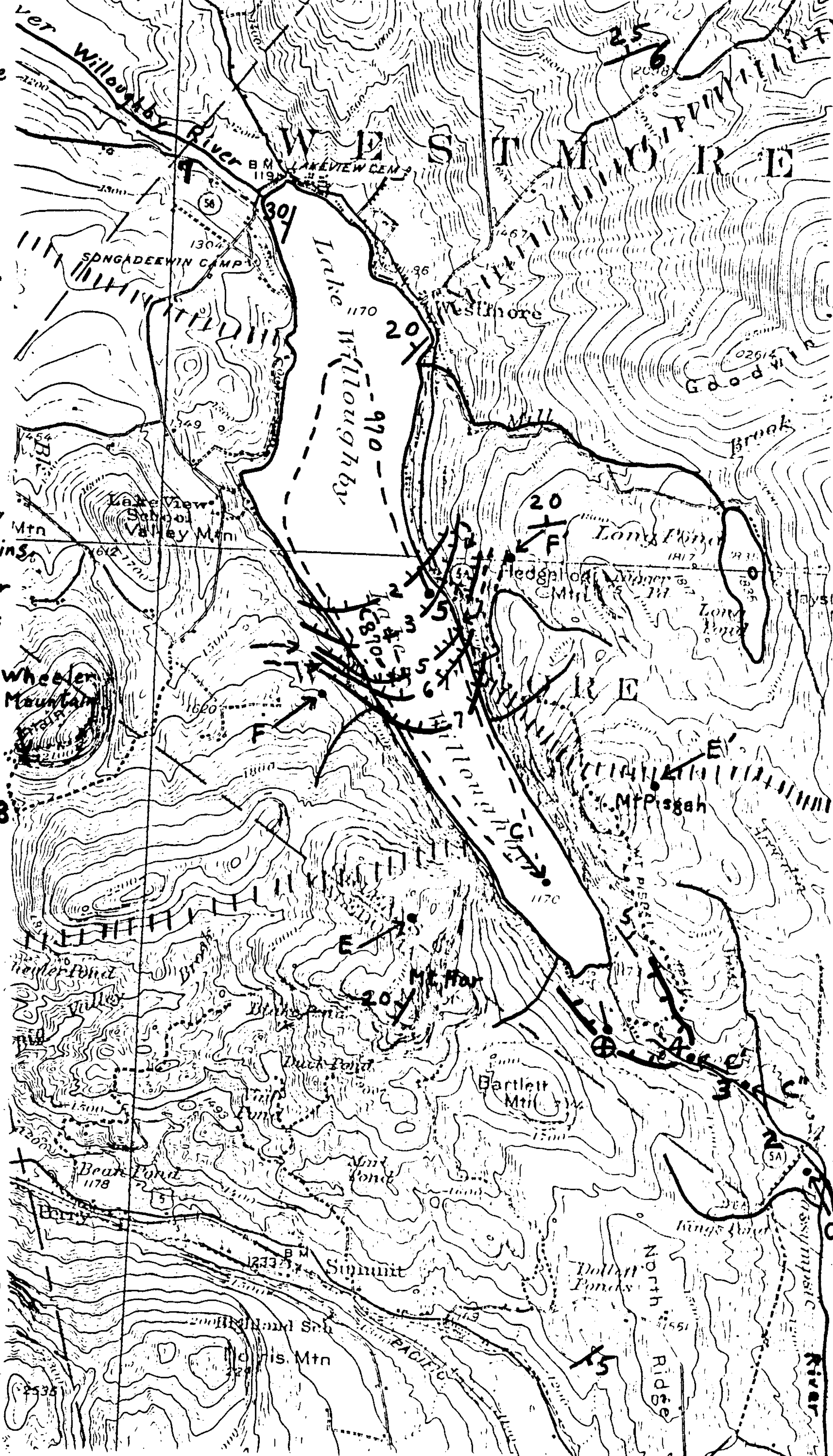


FIGURE 1

TABLE 1. MILEAGE BETWEEN FIELD TRIP STOPS

LOCATION	DISTANCE IN MILES
Junction of routes 5 and 5A in West Burke Village to STOP 1, a small quarry on right side of road opposite Geremia Farm.	3.1
STOP 1 to STOP 2, a pull-put on left side of road, just beyond gate to Westmore Town Dump.	1.7
STOP 2 to STOP 3, a logging road on right side of main road, nearly opposite a road cut in bedrock.	0.4
STOP 3 to STOP 4, a pull-out on right side of road at new south Pisgah Trail head.	0.5
STOP 4 to STOP 5, a pull-out on right side of road at new north Pisgah Trail head.	2.9
STOP 5 to Westmore.	1.8
Westmore to STOP 6, a fork in road near crest of Clover Hill.	2.0
Westmore to beach at north end of Lake Willoughby.	1.0
Northeastern corner of beach to STOP 7, Westmore Town Garage on left side of road.	0.6
Southwestern corner of beach to STOP 8, Wheeler Mountain Trail head. Go three-fourths of a mile up trail.	~5

ROAD LOG

Stop 1. Saprolite on East Flank of Upper Passumpsic Valley Trough.

Saprolite is exposed in a cut bank, twenty-five feet high. A and B soil horizons (defined by Birkeland, 1984) with a combined thickness of one foot occur at the bank top. Nearly unweathered schist is at the foot of the bank. Two vertical, relatively unweathered pegmatitic dikes, a few inches wide, cut through the saprolite. The schist (Dw) is fine grained and contains quartz, mica, calcite, and carbonaceous matter. The horizontal foliation of the schist is clearly visible in the friable saprolite. Carbonate content of the saprolite decreases upward and thin films of clay and iron or aluminum oxides on mineral grains become more continuous upward. Horizontal slabs, a few inches thick, of less friable non-carbonate schist occur, intermittently, a few inches below the soil horizons. Soil creep is indicated by pieces of dike rock that extend one foot down-slope from the dike in the saprolite along the clear soil-saprolite contact. Occasional boulders, mostly of Wg, are found on the surface in the vicinity of the saprolite exposure. Other exposures of saprolite occur in the banks of a small stream 700 feet to the north and in cut banks 1000 and 1500 feet farther to the north.

A similar but much larger area of deeply weathered bedrock with sporadic erratic boulders on the surface occurs on the down-glacier flank of the Appalachian watershed divide in western Gaspe, Quebec. The saprolite at Stop 1 is in a similar position with respect to the Appalachian divide. David and Lebuis (1985) suggest that the location of this large area of deeply weathered bedrock is related to the basal thermal regimes of the Laurentide Ice Sheets at their maximum extents. Basal meltwater began to freeze as the ice sheet slowed down against the up-glacier side of the Appalachian divide and became completely frozen when it reached the down-glacier side. Where a glacier bed is completely frozen no erosion takes place because flow occurs above the basal regelation ice layer. The eventual melting of the regelation ice layer during deglaciation supplied the sporatic erratics. LaSalle and others (1985) suggest a similar explanation for the preservation of some saprolites that occur in "less protected areas" of New England and adjacent Canada.

VIEW

The view from Stop 1 includes the broad trough of the Upper Passumpsic River Valley and the irregular topography of the Willoughby Range. The Willoughby Range probably owes its relief to the Willoughby granodiorite (Wg) that underlies most of it. This is consistent with the nearly unweathered condition of the granitic dikes in the saprolite. The unusually smooth profile of West Sutton Ridge across the valley from Stop 1 is reminiscent of those of the Finger Lakes region in New York which White (1972) attributes to glacial erosion. Horizontal rock structures in carbonate rocks are common to both areas. The straightness of the broad valley and the downstream increase in gradients of deeply incised tributaries suggest glacial erosion of the valley flanks. This is not consistent with the hypothesis for the preservation of the saprolite at Stop 1.

The poorly drained valley floor and low hummocky lateral terraces are underlain mostly by till whose clasts are locally derived schist (Dw) with lesser amounts of Willoughby granodiorite (Wg). The scarcity of meltwater deposits in the valley and their abundance in the Passumpsic Valley south of West Burke is consistent with the hypothesis that a segment of a valley ice lobe stagnated between West Burke and the valley divide just southeast of Lake Willoughby. Koteff and Pessl (1981) argue that if the bulk of the material carried by a continental ice sheet is at its base, then the continuous process of foreward-moving ice supplying rock debris to the snout of the ice sheet is necessary to account for the large volumes of gravel, sand, silt and clay found in many valleys throughout New England. However, Gustaveson and Boothroyd (1987) conclude from observations made on the Malaspina Glacier that meltwaters in englacial and subglacial channels of stagnating ice masses can carry large volumes of sediment to the glacier's snout.

Stop 2. Hummocky Till Terraces, Outwash Fan, Esker, and Floodplain southeast of Valley Divide

INTRODUCTION

The sequence of landforms that lies between Stops 2 and 3 has some characteristics of Koteff's (1974) fluvial non-ice-contact sequence. Meltwater from an active ice margin and a narrow stagnant outer border fed sediments from the up-glacier side of a divide to an outwash plain on the down-glacier side (see Figs. 2 and 3). This situation has some similarities to the Casement Glacier Snout, Southeast Alaska, between 1948 and 1963 (Price, 1973, p. 143-151). In 1948 a stagnant ice mass with developing eskers at its base occurred to the lee of ice-covered bedrock ridges. Upstream from the bedrock ridges was the active snout of Casement Glacier with an extensive cover of supraglacial till.

VALLEY FLOOR MARGINS

The view from Stop 2 includes a pitted outwash fan that is flanked on the west by a lateral, hummocky, till terrace, 30 feet above the fan, and on the east by a flat valley floor 20 feet below the fan surface. Trees hide the esker just to the northwest.

The Westmore Town Dump, 1500 feet southwest of Stop 2, lies near the crest of a hummock on the till terrace. Usually exposed in a freshly dug fifteen foot trench is till with many of the characteristics Boulton (1971) ascribes to supraglacial melt-out tills that have been redeposited as debris flows of relatively low pore water content. The till's matrix is light-tan silty sand; maximum clast size is six inches; clast shape is angular to subrounded; one-third of the clasts are Wg and two-thirds are variably weathered Dw or Dg; deformed sand inclusions are up to one foot across and some contain ripple cross-laminations and planar cross-beds.

Erratics up to twelve feet across tend to occur in clusters on the till terrace surface. The terrace surface south of the dump is cut by three flat-floored channels, 50 to 100 feet wide, that may be scars left by debris flows with relatively high pore water pressure. Similar landforms are discussed by Boulton and Eyles (1979, p. 22) and by Lawson (1982, cited in Chorley and others, 1984, p. 450-452).

Horizontal layers of sand and gravel are exposed in five to fifteen foot walls of an extensive excavation in the outwash fan. Cross-bedding in pebbly sand indicates water current directions were toward the south and southeast. Clast size increases and sorting generally decreases toward the esker.

A shallow closed depression 100 feet across occurs on the outwash fan 300 feet west-northwest of Stop 2, and another, 1500 feet south-southeast, contains a small pond (see Fig. 2).

A sinuous ridge that rises ten to twenty-five feet above the outwash fan surface has many characteristics of eskers described by Price (1973, p. 140-160) and Embleton and King (1975, p. 471-484). Its crest line undulates but generally decreases in elevation toward the southeast. Occasional ten to twenty foot erratics along the crest may have melted out of regelation ice in the roof of an ice tunnel. Near the esker's northwest end is an overgrown gravel pit with an exposure of a generally clast supported gravel. Maximum clast size is three feet; clasts are rounded to subrounded; the matrix is coarse, poorly sorted sand.

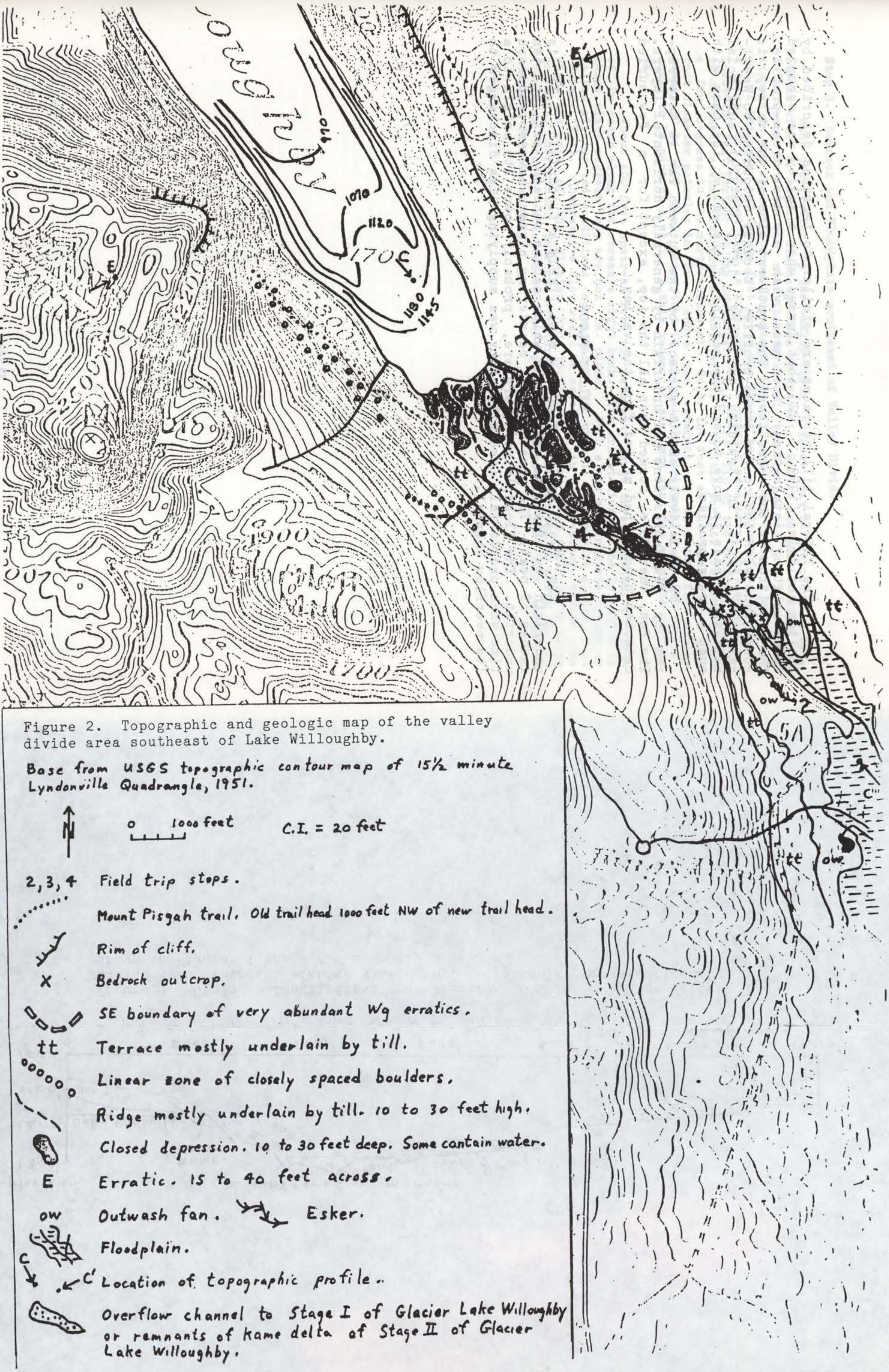


Figure 2. Topographic and geologic map of the valley divide area southeast of Lake Willoughby.

Base from USGS topographic contour map of 15½ minute Lyndonville Quadrangle, 1951.

N

0 1000 feet

C.I. = 20 feet

2, 3, 4 Field trip stops.

Mount Pisgah trail. Old trail head 1000 feet NW of new trail head.

Rim of cliff.

X Bedrock outcrop.

SE boundary of very abundant Wg erratics.

tt Terrace mostly underlain by till.

Linear zone of closely spaced boulders.

Ridge mostly underlain by till. 10 to 30 feet high.

Closed depression. 10 to 30 feet deep. Some contain water.

E Erratic. 15 to 40 feet across.

ow Outwash fan. Esker.

Floodplain.

C C' Location of topographic profile.

Overflow channel to Stage I of Glacier Lake Willoughby or remnants of kame delta of Stage II of Glacier Lake Willoughby.

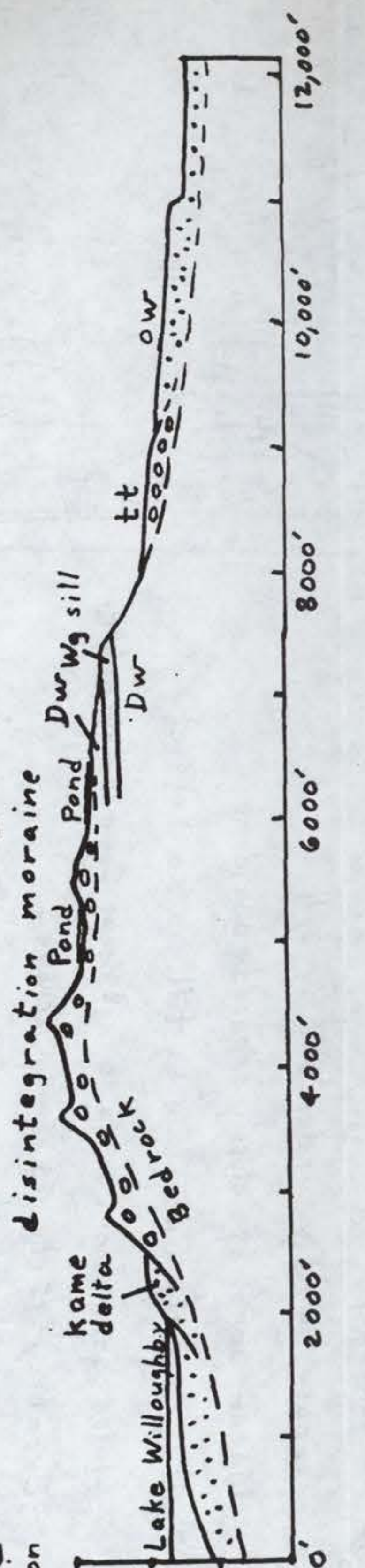


Figure 3. Geologic longitudinal cross-section CC'C''C''' of the valley divide area southeast of Lake Willoughby. Location of cross-section is shown in Figures 1 and 2. V.E. = x5.

The lateral till terraces are similar to terraces described by Boulton and Eyles (1979, p. 72) that formed when a valley was occupied by an ablating (stagnating) ice mass with a convex cross-profile. The position of the esker toward the southwest side of the valley is consistent with the convex cross-profile, because according to Weertman (1972, cited in Embleton and King, 1975, 328-329), subglacial meltwater tends to move toward the margins of the valley ice mass with such a cross-profile.

The relatively smooth surface of the outwash fan is interrupted by few depressions and underlain by generally horizontal layers. The fan, therefore, formed when all the ice, except for a few buried detached blocks, had melted from the valley.

VALLEY FLOOR

The flat valley floor is most likely a floodplain. It is underlain by fine to pebbly sand which occurs beneath six to ten inches of peat. The present small meandering stream, however, appears underfit. The floodplain, therefore, probably owes its origin to a time when proglacial Lake Willoughby was supplying high discharges to its overflow channel.

Stop 3. Bedrock and Floodplain southeast of Valley Divide.

Many observations on the valley floor suggest that it was occupied by a stream with a higher discharge than at present. Schist and granite dikes and sills crop out in and near the stream bed (see Figs. 2 and 3). The bedrock exposed in the roadcut across from Stop 3 is fifteen feet below the northwest end of the esker and twenty feet above the stream bed. Upstream from Stop 3 is a thirty foot cataract that is part of a low bedrock escarpment that extends at least 100 feet farther to the northeast. Below the lip of the cataract bedrock is carbonate schist (Dw) with a few interbeds of pelitic schist. Foliation dips a few degrees to the north-northwest. Below the foot of the cataract are two-foot diameter potholes in Dw. Above the cataract is an almost flat valley floor, 50 to 100 feet wide, that is underlain by horizontally jointed Wg presumed to be part of a sill. A thin veneer of sand and very few erratics overlie Wg.

About 400 feet upstream from the cataract, large three-foot thick slabs of Dw occur along the stream's right bank. The orientation of the foliation in these slabs suggest they are not in place. Such slabs may indicate frozen or melted glacier bed conditions accompanied by upward compressive glacial flow (Moran, 1971). Compressive flow would be expected in the valley divide area where the sole of the glacier had to flow up slope. Slip could occur along unfrozen, horizontal bedrock fractures beneath a frozen bed or along bedrock fractures beneath a melted bed where pore water pressure was high enough to overcome frictional drag.

Stop 4. Lateral and Disintegration Moraines in Valley Divide Area.

INTRODUCTION

Bedrock surfaces on the steep valley sides show some features that can be explained by glacial abrasion and plucking. Most of the topography and sediments that lie between the steep valley sides are characteristic of disintegration moraines. An overflow channel to proglacial Lake Willoughby cuts through this topography.

The downslope increase in steepness of the valley sides (see Figs. 2 and 4) may be a reflection of greater ice velocities within the confines of Willoughby Gap. Boulton (1974) demonstrated that one of the variables that increases the rate of abrasion and plucking by a glacier is greater ice velocities. Converging ice flow from the northwest into Willoughby Gap would result in greater ice velocities.

The only evidence for glacial abrasion in the area of Stop 4 is S15^{o}E trending shallow troughs, a few inches wide, that occur on a gently sloping Wg surface, 2000 feet west-northwest of Stop 4 (see Fig. 2).

If it is assumed that glacial plucking produced the steep cliffs in Willoughby Gap, then the many joints and dikes parallel and subparallel to the direction of glacial flow must have made the plucking process more effective in the Gap than elsewhere in the region. The bedrock of the cliffs is Dw with nearly horizontal foliation and a network of mostly vertical dikes and horizontal sills of Wg up to twenty-five feet thick. Vertical dikes that are parallel to a cliff face low on the valley side, 2000 feet west-northwest of Stop 4, and high on the valley side, 3000 feet north (see Fig. 2), occur with most of their wall rock on the valley side missing. Some closely spaced fractures in Wg dikes subparallel to the cliff face may have formed with the release of horizontal stress on the dike as rock on the valley side was removed by glacial plucking or as ice melted that once pressed against the cliff. Eyles (1983, p. 104) suggests that large rockfalls and slides that often occur immediately following glacial retreat can be accounted for by these causes for the release of horizontal stress. Evidence for such large rockfalls and slides in the vicinity of Stop 4 are: (1) the large blocks, up to fifty feet across, of Dw with Wg dikes and sills found cropping out along the highway 0.7 and 1.1 miles northwest of Lake Willoughby's southeast end and (2) the bouldery till terrace and linear bouldery zones that are discussed later.

The lobate shaped boundary that separates an area of very abundant and large (five to twenty-five feet across) erratics from an area of less abundant and generally smaller erratics (see Fig. 2) may mark a brief stillstand of the front of a debris laden snout of a valley ice lobe. The change in abundance of surface erratics across this boundary is in most places obvious. Another change is the noticeably larger ratio of Wg to Dw erratics northwest of the boundary than southeast of it. This proposed stillstand was caused, at least in part, by (1) the slope northwest of the valley watershed divide that faced the glacier and (2) the down-glacier convergence of the steep valley sides (see Figs. 1 and 2). Both these topographic features reduced the frontal velocity of the ice-lobe. This, in turn, accentuated compressive flow which causes slip planes in ice that raise basal debris to the surface of the glacier. These processes are reviewed by Chorely and others (1984, p. 438) and Koteff and Pessl (1981).

VALLEY FLOOR MARGINS

Many characteristics of the till terraces along both sides of the valley divide area (see Fig. 2) suggest they formed at the laterofrontal margins of a valley ice lobe. A break in slope occurs where both terrace surfaces meet the steep valley sides. Their outer margins are irregular with slopes of varying steepness. That both terraces had an ice core, at least in part, is suggested by the uneveness of their surfaces and the presence of shallow closed depressions.

The size, shape, composition, and abundance of the boulders on the till terrace along the northeast side of the valley suggest they have a supraglacial origin. It is unlikely that most of the boulders fell from the cliffs immediately above them since most of them are composed of Wg and the cliffs are mostly Dw. The abundance, large size (five to twenty-five feet across), and subangular shape of most of the boulders are some of the characteristics that Boulton and Eyles (1979, p. 12-14) use to distinguish supraglacially derived debris. These boulders, therefore, must be derived from Wg bedrock exposed on the northwest flank of Mt. Pisgah (see Fig. 2) from which loosened blocks fell onto the lateral margin of an active, backwasting, valley ice lobe and were transported by the ice to their present location.

The terrace along the southwest side of the valley, however, is composed of till that may be mostly subglacially derived. Fewer and rounder boulders litter the terrace surface. Three till exposures, two of which are in the till terrace, have many characteristics of subglacially derived material as described by Boulton and Eyles (1979, p. 12-14). One exposure occurs 275 feet northeast of Stop 4 in a narrow trench at the outer margin of the till terrace; another occurs at the east end of the till terrace in a recent logging road cut; and a third occurs 200 feet east-northeast of the lower Beaver Pond in an old logging road cut. The till matrix is silty sand; the clasts are subrounded to rounded, one inch to fifteen feet across, two-thirds Wg, one-third variably weathered Dw or Dg and, rarely, Derby granite. (Derby granite bedrock occurs fifteen miles to the northwest in a pluton, six miles in diameter. A description of Derby granite is given by Doll, 1951, p. 41-42.) Vague striations occur on one faceted boulder at the old logging road exposure. A moderately developed soil (defined by Birkeland, 1984, p. 7 and 24) that is exposed at the recent logging road cut is an expression of the relatively abundant silty sand matrix. This is in contrast to the near lack of a fine matrix exposed near the surface of the till terrace along the opposite side of the valley. If this till terrace is underlain by mostly subglacially derived till, then it is likely a product, at least in part, of compressive flow, as previously described, near the front of the valley ice lobe.

The outer margin of the till terrace on the northeast side of the valley is in part coincident with a fifty to seventy-five foot wide linear zone of closely spaced Wg boulders (see Fig. 2). Boulton and Eyles (1979, p. 14) ascribe the origin of similar features to the fillings of ice-marginal, supraglacial gullies or cravasse fills along which melt water has winnowed out finer debris. Above the till terrace on the opposite side of the valley occur two shorter, fifteen feet wide, linear bouldery zones. A more diffuse (3000 feet long, 300 feet wide) linear bouldery zone occurs along the steep slope southeast of the cliffs on Mount Hor. The largest boulder is thirty by forty feet. The origin of this bouldery zone may be similar to that of the till terrace on the northeast side of the valley divide area. A terrace may not have formed here, however, because the bedrock slope was too steep or there were not enough boulders. Observations at an exposure in the southern stream gully crossed by the trail to the foot of the cliffs on Mount Hor are consistent with the supraglacial origin for this linear bouldery zone. The stream gully above the trail is partially filled with large subrounded to subangular Wg boulders. The boulders lie on top of a till with characteristics that suggest it is subglacial in origin. The till has a sandy silt matrix with one-half to three foot subrounded clasts of Dw or Dg. The subglacial till may have acquired its clasts and matrix when the base of the Ice Sheet at its maximum extent was melting in Willoughby Gap. During deglaciation large rockfalls from the cliffs on Hor and the steep slopes to the north of Hor accumulated along the margin of a southeastward flowing ice lobe in Willoughby Gap.

VALLEY FLOOR

The topography of the lower ground between the two till terraces has characteristics of a disintegration moraine as defined by Gravenor and Kupsch (1959). Depressions five to twenty-five feet deep are enclosed by till ridges along which there are occasional knolls (see Figs. 2 and 3). Gravenor and Kupsch conclude that such depositional forms owe their origin to material that fell into crevasses, subglacial channels, and irregularly shaped hollows under and in stagnant ice. Clayton and Moran (1974, p. 107) give two explanations for the tendency of depressions to be rimmed by till ridges: (1) the amount of debris slumping or flowing from an ice-cored hummock into a hole may not have been enough to completely fill it; (2) thick deposits on the bottom of a hole in stagnant ice may have caused the hole to invert to an ice-cored hummock down the sides of which the deposits slumped. The southeast elongation of most of the closed or partially closed depressions suggest that a system of crevasses or channels more or less perpendicular to the ice lobe front widened and filled with till as the stagnant ice downwasted.

Soon after these crevasse, channel, or hollow fills lost their ice walls, mud flows from over-steepened till ridges partially filled the adjacent depressions. Observations that can be linked to these mud flows are: (1) crude two or three foot high terraces on some depression floors, (2) fewer erratics on depression floors than till ridges, and (3) perched giant erratics on and near the crest of many till ridges.

That the bedrock valley floor slopes up from the lake and then flattens in the vicinity of the present valley divide is deduced from the slope of the depression floors (see Fig. 3). On the lake side of the valley divide, depression floors slope away from the valley divide and on the southeast side depression floors are horizontal or slope very gently away from the divide.

The floors of three elongate depressions that extend to the shore of Lake Willoughby appear graded to the lake's present level. One elongate depression ends fifteen feet above the shore of Lake Willoughby. Two of these depressions meander slightly, three have flat-floors, and three have steep head ends. A high knoll occurs at the upper end of two till ridges that separate these depressions (see Fig. 2). The former Willoughby Inn stands at the crest of one of these knolls. A perhaps "outrageous hypothesis" considers these till ridges former southeast facing ice front gully fillings and the knolls scree at the foot of the ice front gullies. Such features are described at the front of recent glaciers by Boulton and Eyles (1979, Figs. 4 and 5). Melting and evaporation of underlying and intervening ice resulted in the reversal of the gully filling's slopes and the formation of the elongate depressions that flank them. Post-glacial erosion and deposition graded the floors of the depressions to the present level of Lake Willoughby.

There are four explanations for the disintegration moraine to be lower topographically than the lateral till terraces. These explanations are not mutually exclusive. (1) A thicker ice core existed for glacial debris in the center of the snout than at its margin. This probably was related to bedrock topography. (2) Boulton and Eyles (1979, p. 20) suggest that more debris is concentrated toward the lateral margins of a snout than toward its center line because there is usually a component of flow within a valley ice lobe that is away from its center line. (3) The till terraces represent a slightly earlier ice-marginal stagnation zone than the disintegration moraine does. (4) Along the lateral margins of the valley ice lobe, more subglacial erosion took place and more debris fell onto the lobe than toward its center line.

The overflow channel for Stage I of proglacial Lake Willoughby (Stuart and MacClintock, 1969, p. 143 and Fig. 18) may have been in part a series of depressions that were linked together by being partially filled with pebbly sand (see Fig. 4). The poor drainage within the overflow channel floor suggests that the ice core of the disintegration moraine had not yet completely melted during Stage I time.

The campground at the southeast end of Lake Willoughby is on the former site of a kame terrace that was underlain, at least in part, by pebbly gravel and sand foreset beds (observation by B. Ebbett in 1961). The terrace surface was about 1220 feet elevation in 1960 (see Fig. 2). A remnant of the terrace sediments crops out on the road side of the campground. The elevation of this terrace is consistent with Stuart and MacClintock's (1969, p. 146 and Fig. 19) Stage II of Proglacial Lake Willoughby.

[Stop 5 begins on page H22.]

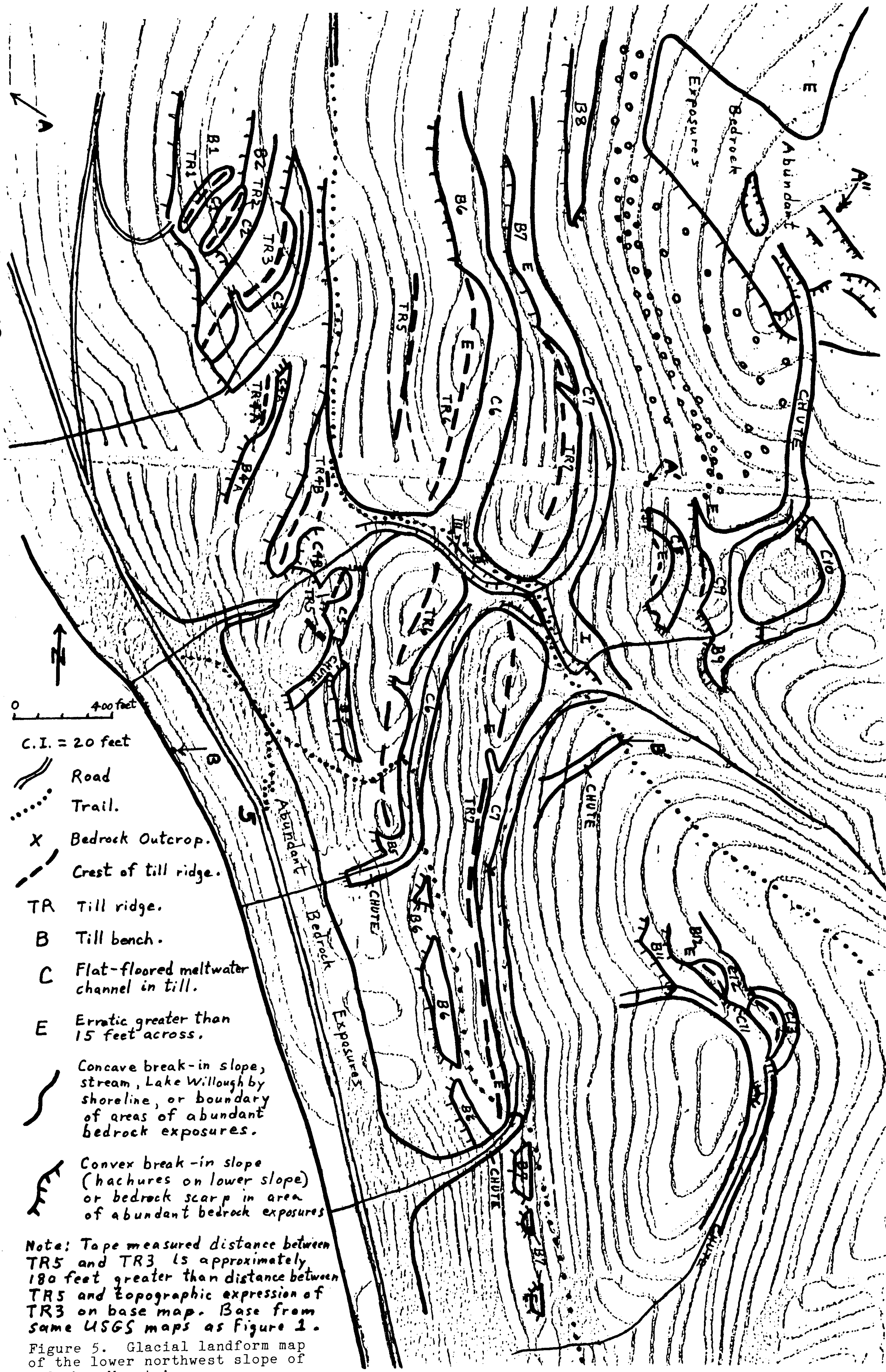


Figure 5. Glacial landform map of the lower northwest slope of Hedgehog Mountain.

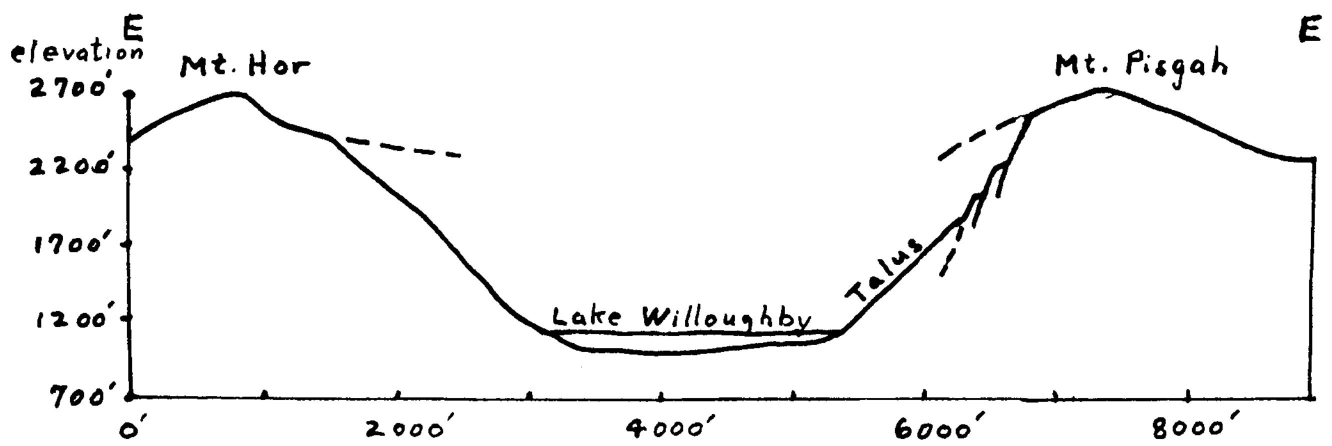


Figure 4. Transverse topographic profile EE' of the southeast portion of Willoughby Gap. Location of profile is shown in Figures 1 and 2. V.E. = x1.1.

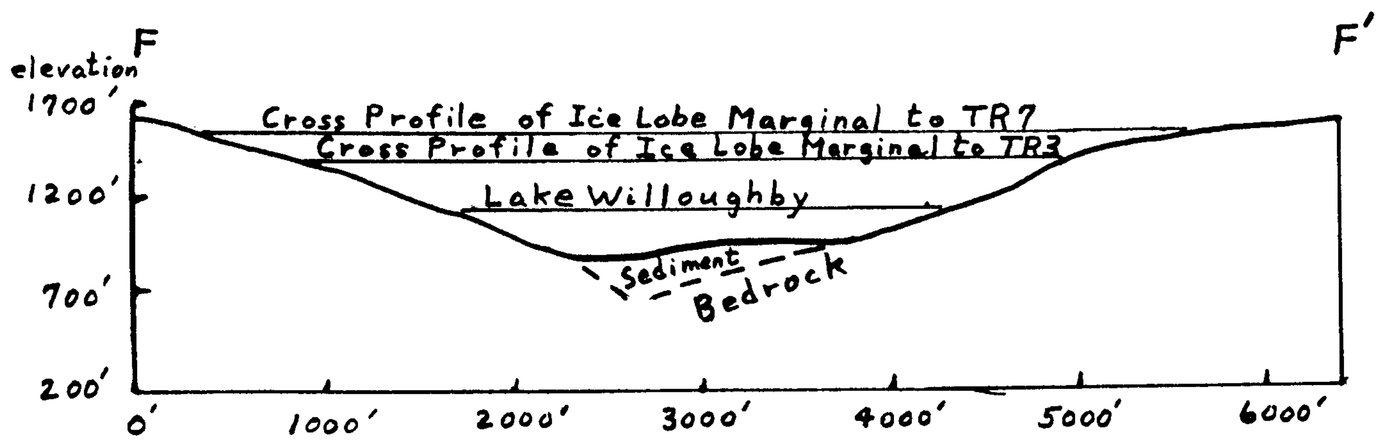


Figure 6. Transverse topographic profile FF' of the northwest portion of Willoughby Gap that also shows ice lobe surface levels. Location of profile is shown in Figure 1. No V.E.

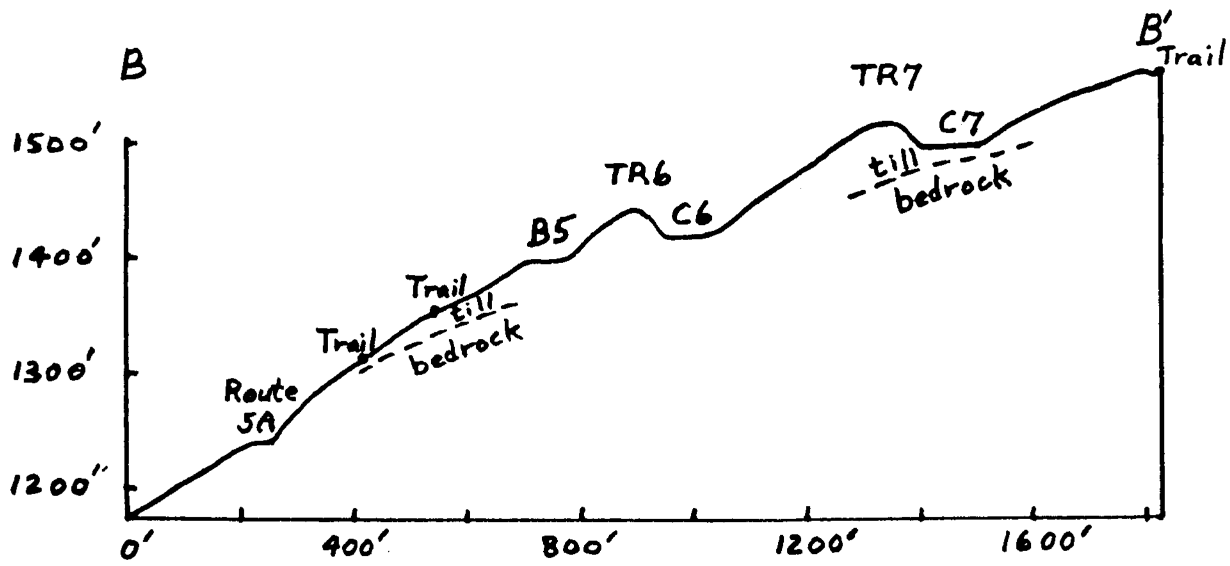
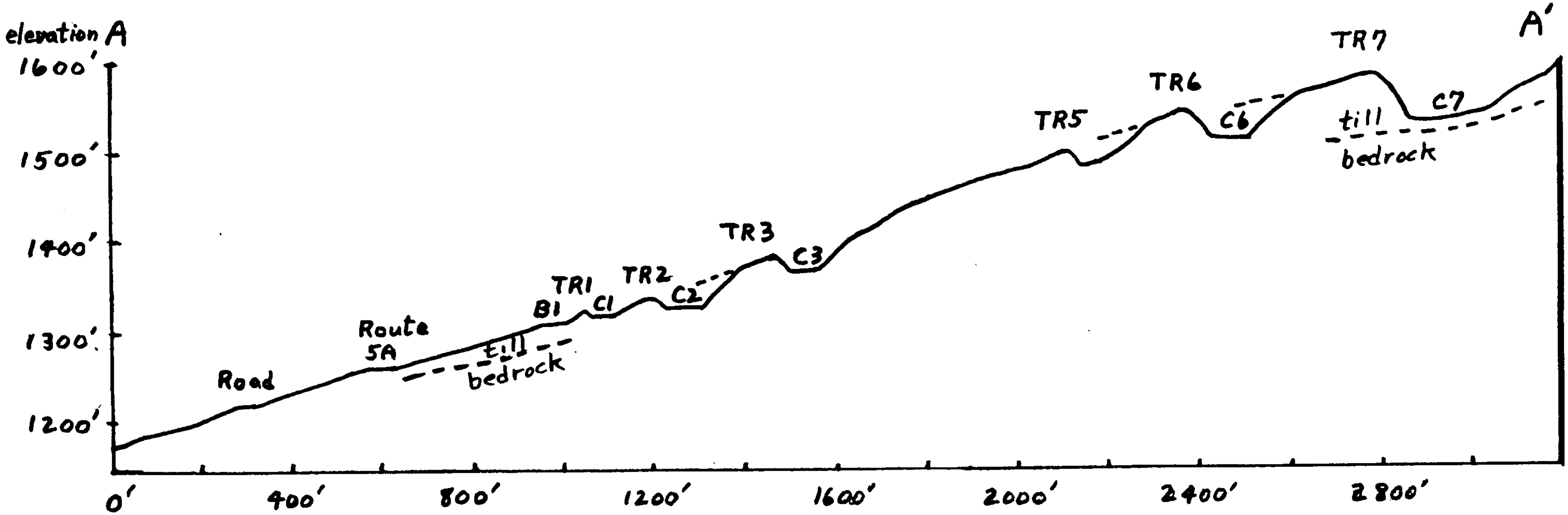


Figure 7. Transverse topographic profile BB' of till ridges TR6 and TR7. Location of profile is shown in Figure 3. V.E. = x2.



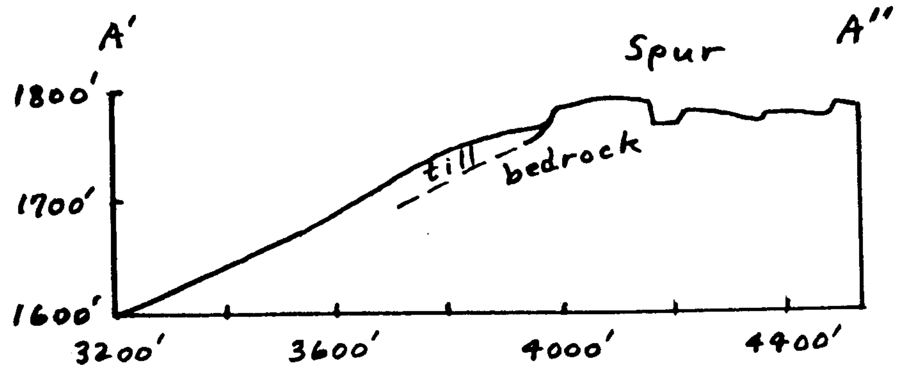


Figure 8. Transverse topographic profile AA'A'' of till ridges TR1 through TR7 and lower northwest spur of Hedgehog Mountain. V.E. = x2.

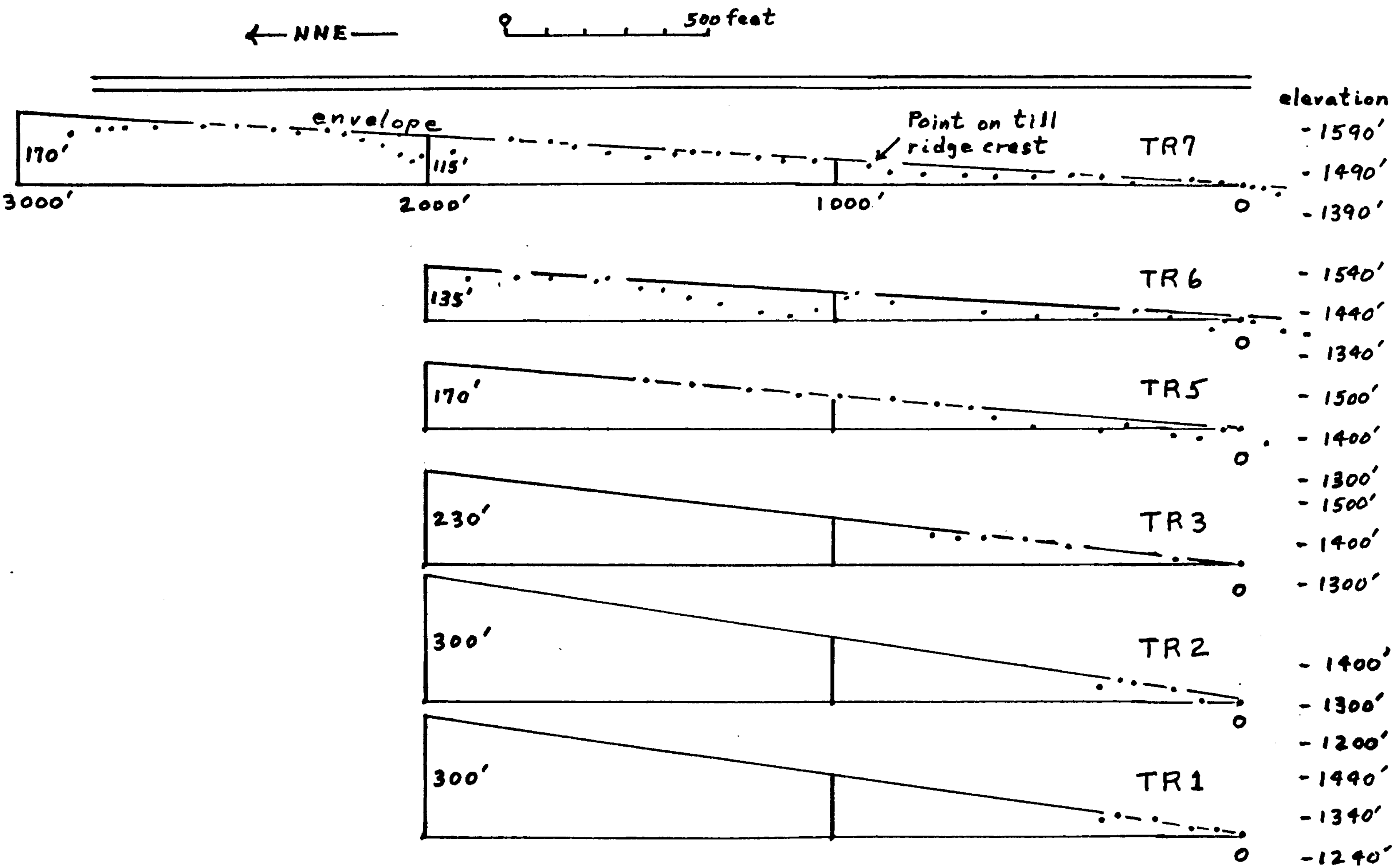


Figure 9. Longitudinal profiles of till ridges presumed to be laterofrontal moraines.

H16

FIGURE 11

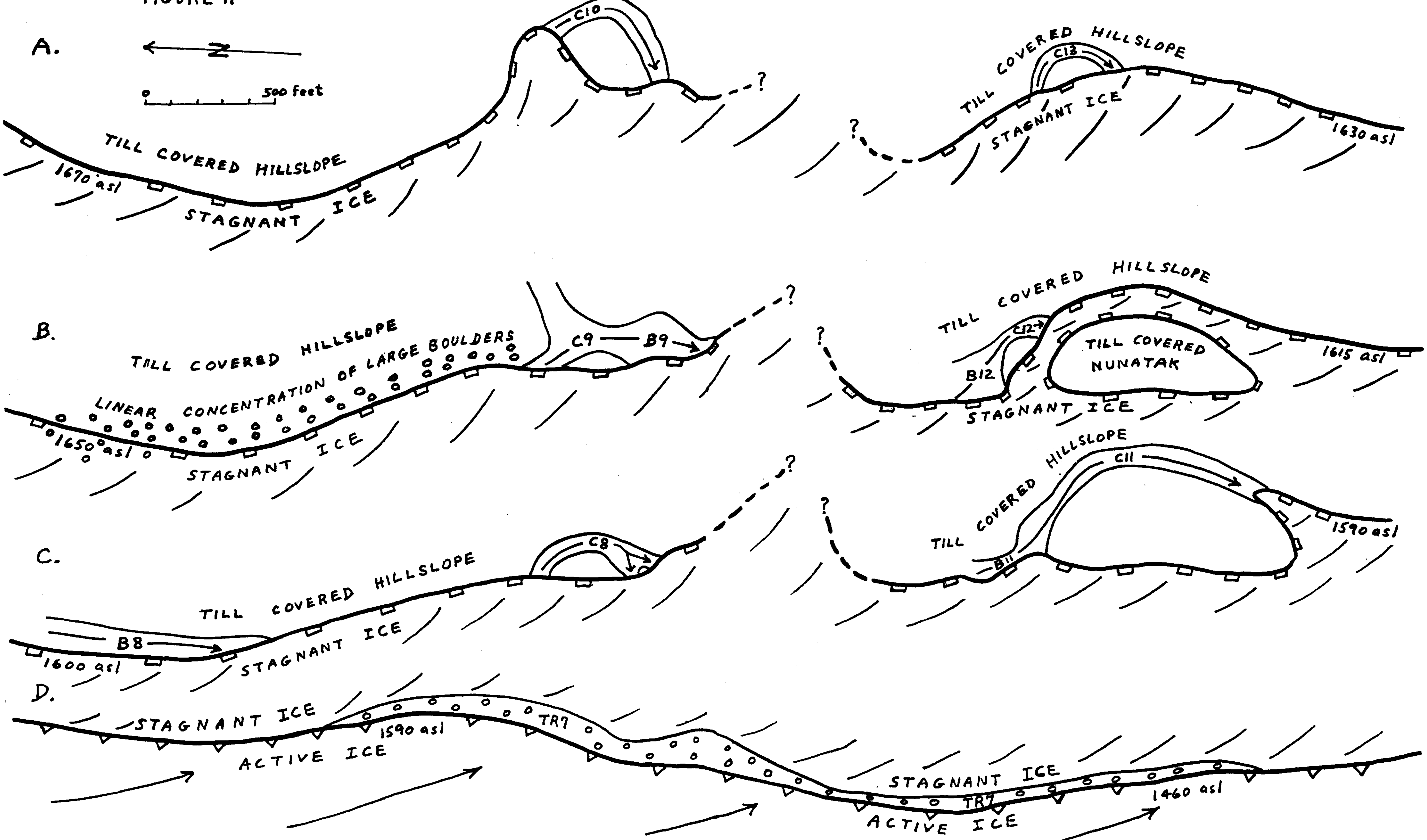
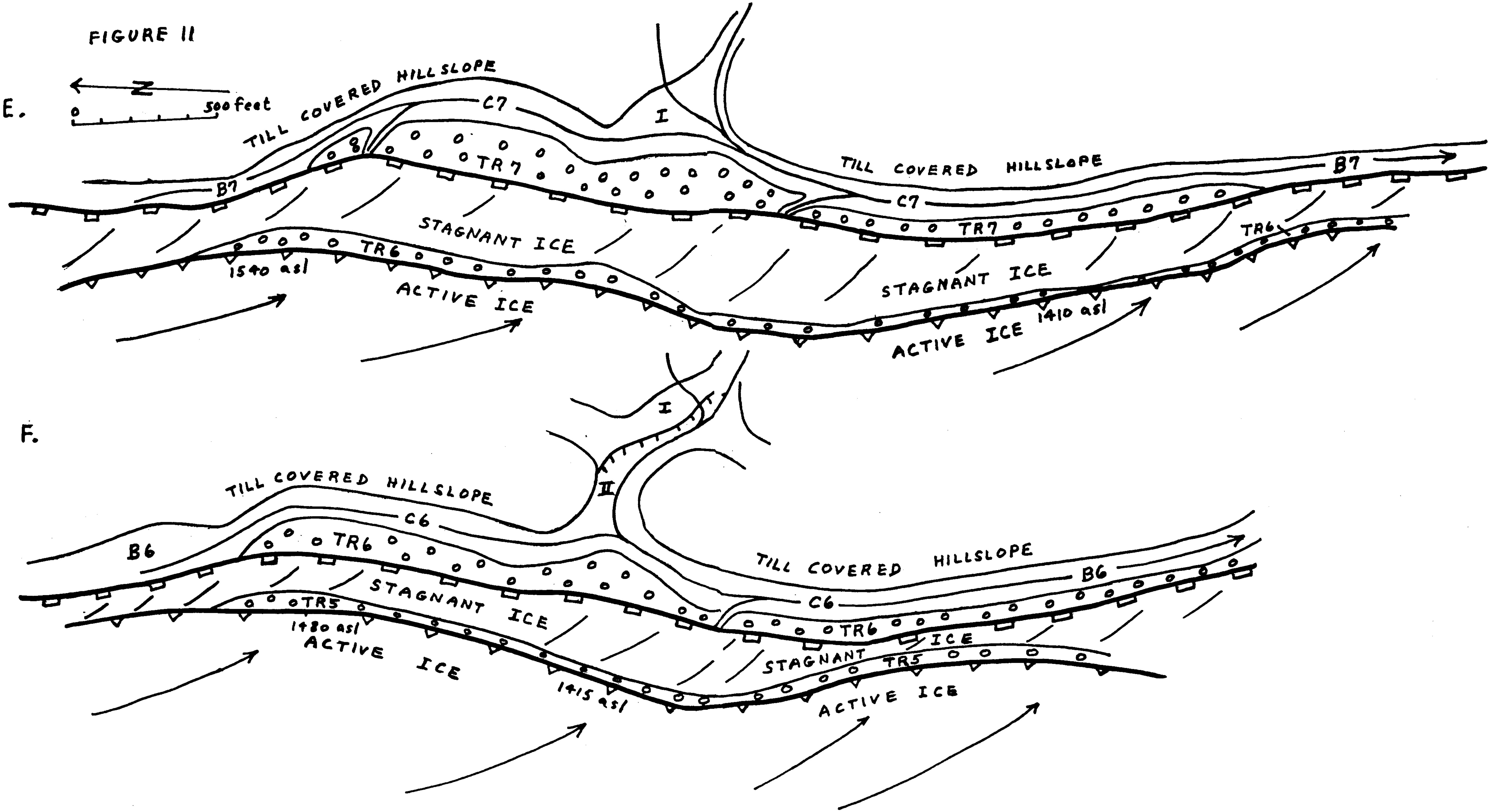

A.
N
0
500 feet
C10
C13
TILL COVERED HILLSLOPE
1670 asl
STAGNANT ICE
TILL COVERED HILLSLOPE
STAGNANT ICE
1630 asl
B.
TILL COVERED HILLSLOPE
LINEAR CONCENTRATION OF LARGE BOULDERS
C9
B9
1650 asl
STAGNANT ICE
TILL COVERED HILLSLOPE
C12
B12
TILL COVERED NUNATAK
1615 asl
STAGNANT ICE
C.
C8
TILL COVERED HILLSLOPE
B8
1600 asl
STAGNANT ICE
TILL COVERED HILLSLOPE
C11
B11
1590 asl
D.
STAGNANT ICE
TR7
1590 asl
ACTIVE ICE
STAGNANT ICE
TR7
1460 asl
ACTIVE ICE

FIGURE 11


E.
N
0
500 feet
TILL COVERED HILLSLOPE
C7
I
8
TR7
B7
TILL COVERED HILLSLOPE
B7
C7
TR7
TR6
STAGNANT ICE
TR6
1540 asl
ACTIVE ICE
STAGNANT ICE
1410 asl
ACTIVE ICE
F.
I
II
TILL COVERED HILLSLOPE
C6
B6
TR6
TILL COVERED HILLSLOPE
B6
C6
STAGNANT ICE
TR5
1480 asl
ACTIVE ICE
TR6
ICE
STAGNANT
TR5
1415 asl
ACTIVE ICE

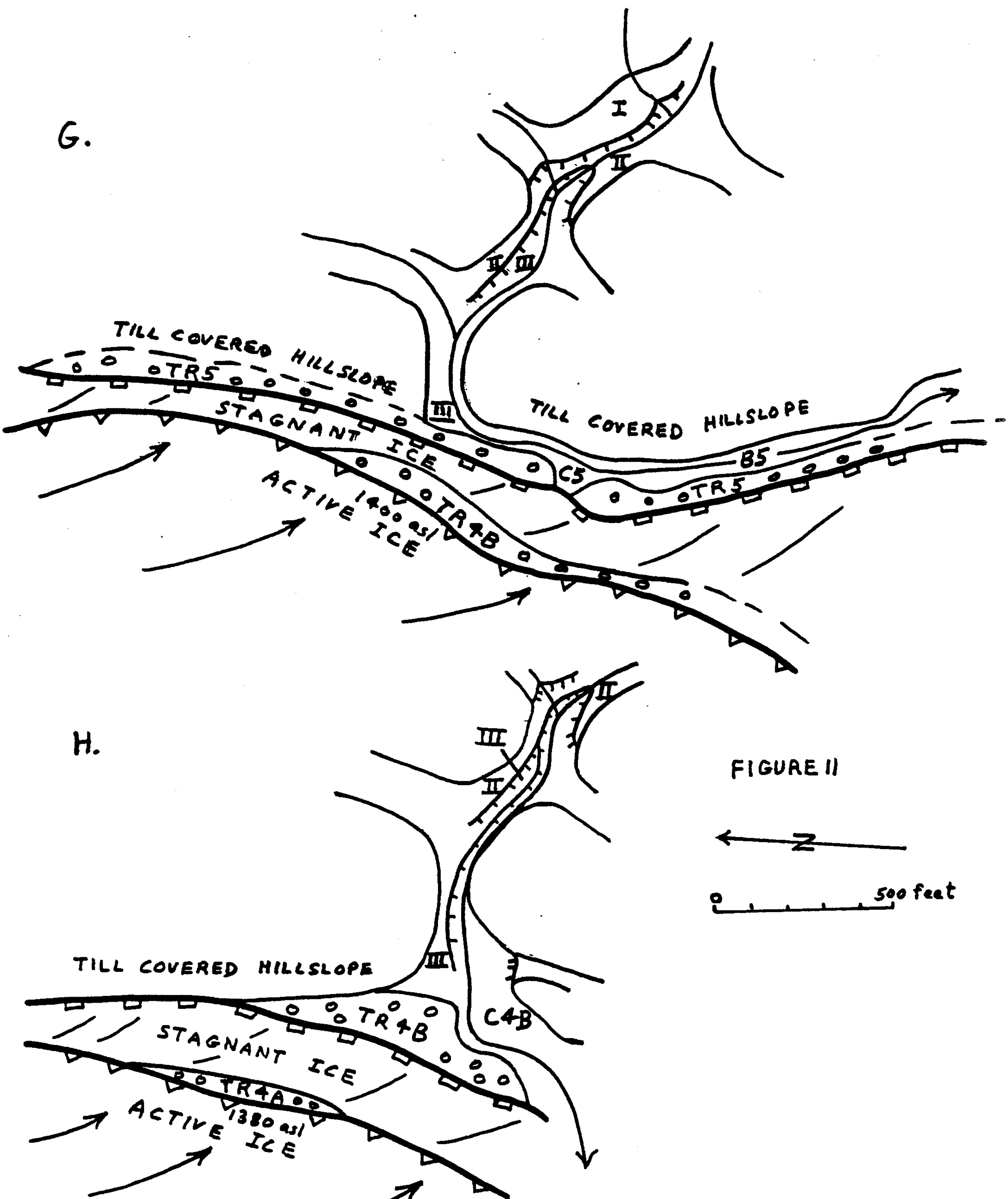
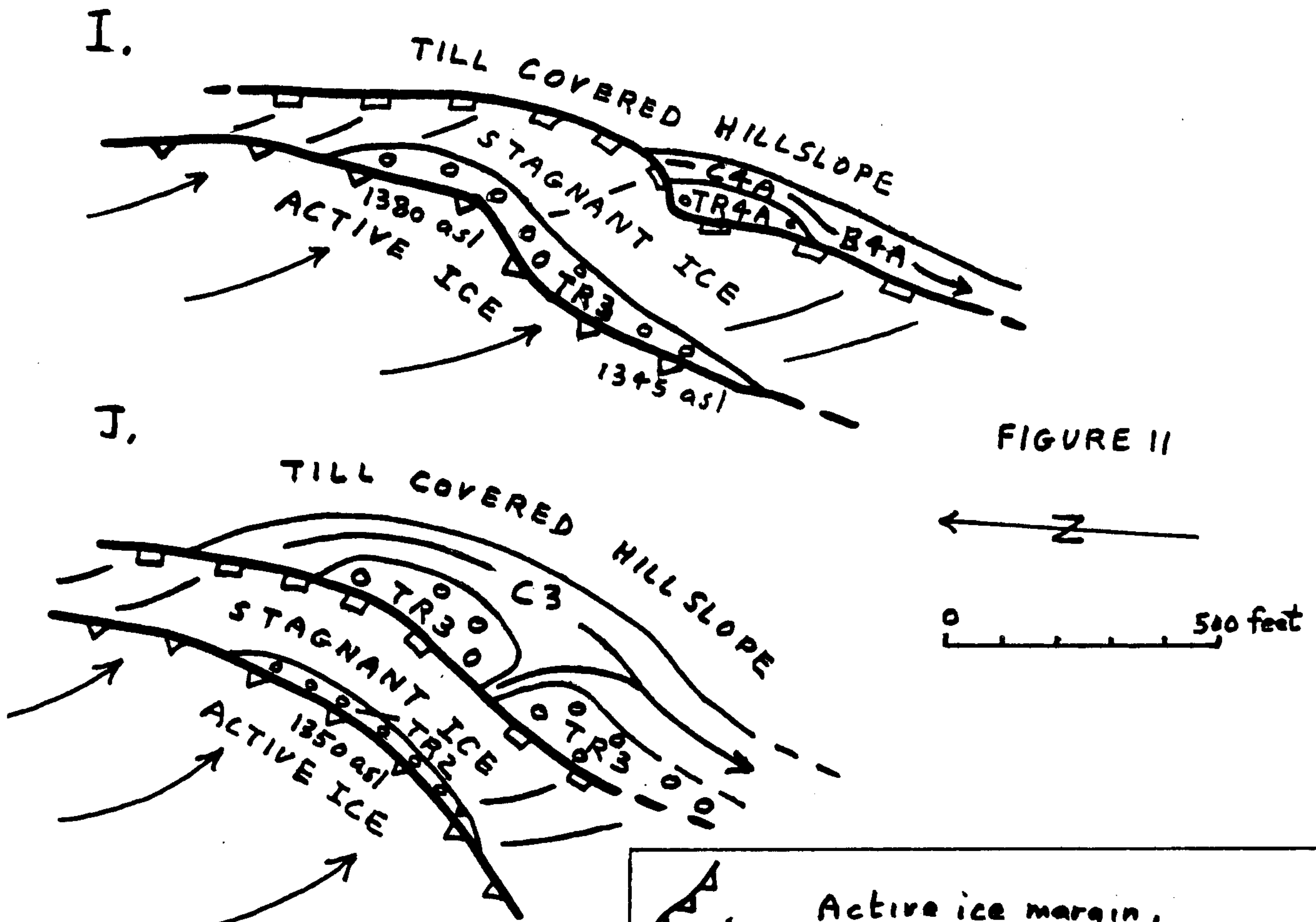
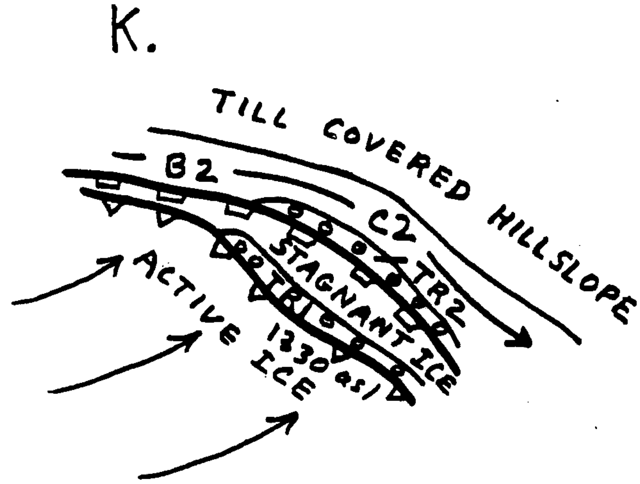

G.
I
II
II III
III
TILL COVERED HILLSLOPE
TR5
TILL COVERED HILLSLOPE
STAGNANT ICE
B5
C5
TR5
ACTIVE ICE
1400 asl
TR4B
H.
III
II
III
TILL COVERED HILLSLOPE
TR4B
C4B
STAGNANT ICE
TR4A
1380 asl
ACTIVE ICE
N
0
500 feet

FIGURE 11





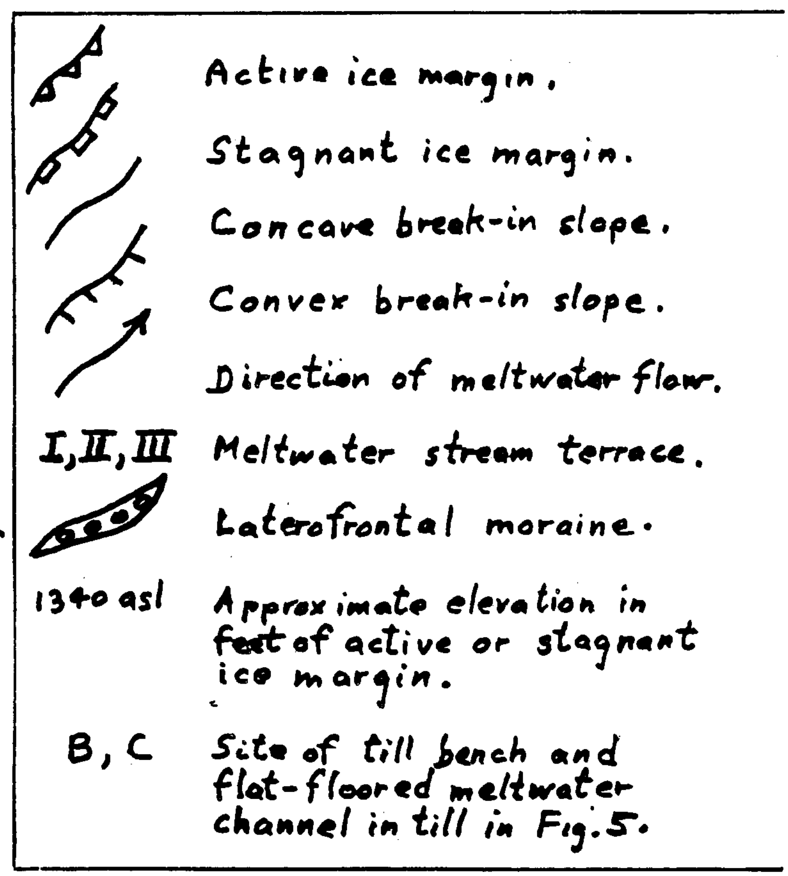


Figure 11. A series of maps derived from Figure 5 that show the nature of the successive margins of a backwasting ice lobe in Willoughby Gap.

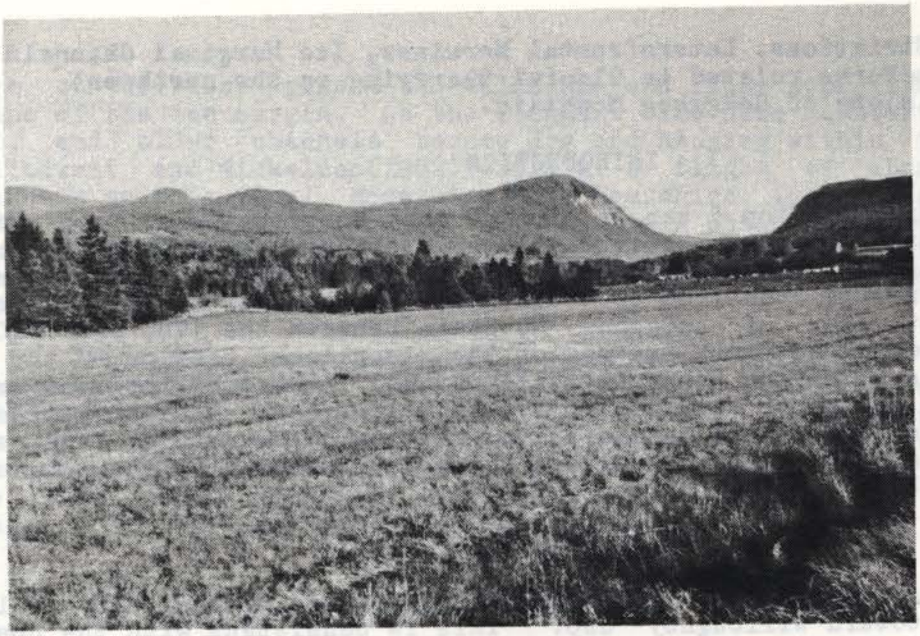


Figure 12. View southeastward from just north of the road intersection northeast of Valley Mountain. Lake Willoughby is visible at the foot of the talus slope below the cliff on Mount Pisgah. The slope described at Stop 5 is below Hedgehog Mountain, the next mountain north of Pisgah. The generally gentler northwest facing slopes and steeper southeast facing slopes of the Willoughby Range is shown by all three mountains east of the Willoughby trough.

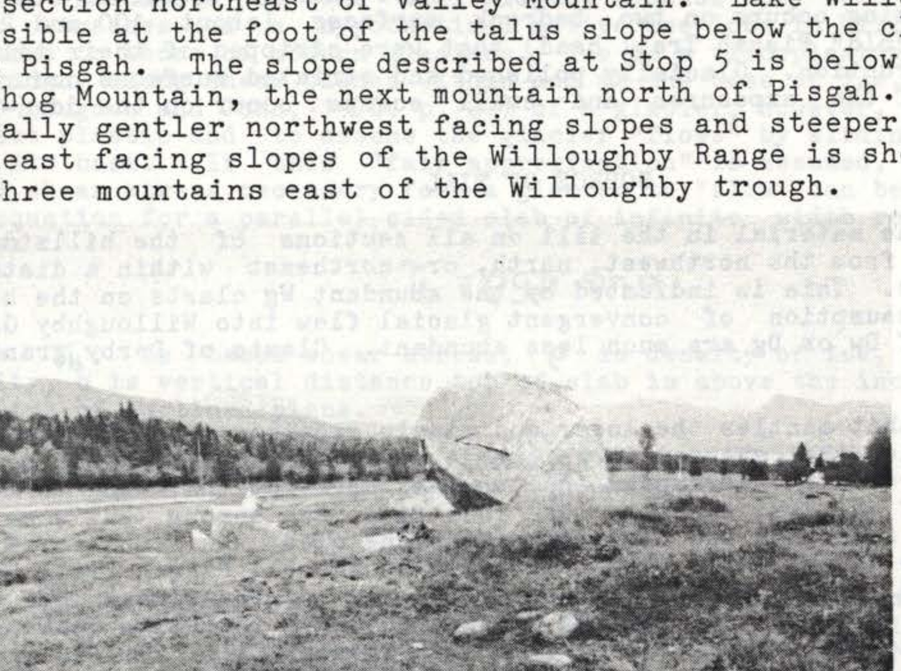


Figure 13. View southwestward from Stop 6. Lake Willoughby is visible at the foot of the slope to the right of the large rounded boulder of Derby granite. The flat side of the large boulder and the angular block to the left of the large boulder are products of quarrying by man. The asymmetry of the northwest and southeast facing slopes of the Willoughby Range is well shown by Wheeler Mountain at the right margin of the picture.

Stop 5. Glacial Striations, Laterofrontal Moraines, Ice Marginal Channels and Landforms related to Glacial Quarrying on the northwest facing slope of Hedgehog Mountain.

INTRODUCTION

The hillside at Stop 5 can be divided into four sections (see Figs. 1 and 5) of contrasting geomorphology. The lower section extends from the road to the 1300 foot contour; the middle section extends to the 1600 foot contour; the upper section extends to the 1700 foot contour; and the fourth section includes the lower northwest spur of Hedgehog Mountain between the 1700 and 1800 foot contours. Some bedrock surfaces exposed on the steep lower section of the hill-slope show evidence of having been glacially abraded. Most of the landforms in till of the middle and upper sections of the hillslope have characteristics that suggest they originated at the laterofrontal margins of an active, backwasting, valley ice lobe and along narrow zones of stagnant downwasting ice marginal to this active ice lobe.

LOWER SECTION

The lower section of the hillslope steepens where glacial flow is likely to have converged toward Willoughby Gap. This is consistent with the observation that glacial velocities and, therefore, glacial erosion (Boulton, 1974) increase where glacial flow is convergent. More direct evidence for glacial abrasion and plucking occurs on two bedrock surfaces (about 100 and 200 feet southeast of the "old" Pisgah Trail head) that were stripped of their till cover during road construction. Glacially polished and striated surfaces occur on the up-glacier side of the exposures and small scarps occur on the down-glacier side.

SOURCE OF TILL

Most of the material in the till on all sections of the hillside above Stop 5 is derived from the northwest, north, or northeast within a distance of two or three miles. This is indicated by the abundant Wg clasts on the hill and based on the assumption of convergent glacial flow into Willoughby Gap (see Fig. 2). Clasts of Dw or Dg are much less abundant. Clasts of Derby granite are rare.

The till that mantles the lower and middle sections of the hillslope has some characteristics that suggest it is subglacially derived. The surface of the till is littered with subrounded to rounded erratics that tend to occur in clusters. Most of the erratics are less than five feet across, but some are up to twenty feet. A faceted and striated greenstone clast was found in a silty sand till exposure about 200 feet northwest of the old Pisgah Trail head. A compact silty sand till is exposed in a bank 500 feet up the stream near the old Pisgah Trail head. Most clasts are Wg, less than three inches in diameter, and rounded. A two foot thickness of a loose, more gravely till overlies the silty sand till.

MIDDLE SECTION

The middle section of the hillside at Stop 5 contains at least seven till ridges (TR1 through TR7 in Fig. 5) and the same number of intervening flat-floored channels (C1 through C7) that trend at small angles to the contours of the hillside. That the till ridges do not have a bedrock core is suggested by the lack of bedrock exposures, except for a lone exposure in the bed of a small stream on the floor of channel C7. The break-in slope on portions of the lake-ward flanks of till ridges TR7, 6, and 3 shown in Figure 8 suggests that the channels have been incised into a relatively smooth slope underlain by at least a twenty to thirty foot thickness of till.

Similar sets of channels in the Glacier Bay area of southeast Alaska are associated with downwasting stagnant glaciers that were cut off from their accumulation zones by the thinning of ice over buried ridges. These channels, as described by Goldthwait and Mickelson (1982), are most common in till of the lower one-half and up-glacier side of hillslopes. Water emerges from a glacier cave or crevasse against the hillslope and flows approximately parallel to the slope of the ice margin. As the stagnant ice mass downwastes, new channels are cut, and older channels become dry and hanging within two to five years. Goldthwait and Mickelson(1982) explain a flight of abandoned, flat-floored channels on Bois (Boy) Mountain, New Hampshire, in the light of these observations at Glacier Bay. The Bois Mountain channels, however, differ in some respects from those above Stop 5. They form more of an anastomosing pattern and there is no single distinct ridge crest between them (Field notes of B. Ebbett, 1987).

The shape, trend, spacing, and crestal gradient of the till ridges that lie between the supposed ice-marginal channels within the middle section of the hillslope can be shown to be consistent with the hypothesis that the till cover was not a smooth sheet prior to the incision of the channels, but a series of laterofrontal moraines that formed at the active margins of a backwasting valley ice lobe. The channels, therefore, were likely marginal to a narrow zone of stagnant ice that was contemporaneous with active ice lower on the hillslope.

Crestal Gradients of Till Ridges

The best data, perhaps, in support of laterofrontal moraines is the crestal gradients of the till ridges, because they lead to a way that basal shear stresses of a hypothetical active ice lobe in Willoughby Gap can be compared to those of modern glaciers. Ice is a viscoplastic material that flows by sliding along shear planes within and at the base of ice. A "fair approximation" (Sugden and John, 1976, p. 27) of a glacier, however, is to treat ice as a perfect plastic and to assume the glacier "flows" by sliding along shear planes near its base. If this "fair approximation" is assumed, then the approximate basal shear stress necessary for a glacier to "flow" can be determined by using the equation for a parallel sided slab of infinite width resting on an inclined plane:

$$\tau_b = \rho g h \sin \alpha$$

where τ_b is basal shear stress, ρ is density of ice, g is acceleration of gravity, h is vertical distance top of slab is above the inclined plane, and α is slope of inclined plane.

A modification of this equation for its application to a valley ice lobe whose bed slope is less than twenty degrees is:

$$\bar{\tau}_b = \rho g \frac{A}{P} \frac{\Delta t}{\Delta x}$$

where $\bar{\tau}_b$ is basal shear stress averaged over a transverse cross-section of a valley ice lobe, t is thickness of ice lobe at its center line in the transverse cross-section, x is horizontal distance of the transverse cross-section from the front of the ice lobe, A is area of the transverse cross-section, P is perimeter of the glacial bed in the transverse cross-section, and $\Delta t/\Delta x$ is surface slope of glacier at the transverse cross-section (see Schilling and Hollin, 1981, p. 207-208).

If till ridges TR1 through 7 were formed at the laterofrontal margins of an active, backwasting, valley ice lobe, then their crestal gradients should approximately equal the surface slope ($\Delta t/\Delta x$) of the ice lobe. A longitudinal topographic profile of each till ridge was obtained by (1) locating on U.S.G.S. topographic map a point near the southeast end of a till ridge crest that can be easily recognized on the ground; (2) measuring with a tape measure, "Suunto" clinometer, and "Brunton" compass the ground distance between points along the crest that are seldom more than 100 feet apart, the angle of elevation or depression between these points, and the direction between these points; (3) making a map of the till ridge crests from the data in (2); (4) projecting the points along a till ridge crest onto a straight line drawn between the first and last point; (5) constructing longitudinal topographic profiles of the projected points (see Fig. 9); (6) drawing envelopes of the longitudinal profiles and noting that they are straight lines; (7) determining Δt and Δx graphically.

The area bounded by a cross-profile of Willoughby Gap in the vicinity of the till ridges up to a level line at the elevation of the midpoint of a till ridge should approximately equal A, the cross-sectional area of the ice lobe. The length of the cross-profile up to the elevation of the level line should approximately equal P, the perimeter of the glacier bed (see Fig. 11). A and P for till ridges three and seven were determined graphically from Figure 6. It was found that if A/P = ct, then ct = 0.5 for both till ridges. This data made it possible to calculate A/P for the other till ridges by assuming A/P = 0.5t.

Table 2 shows the values of the variables substituted into the modified basal shear stress equation and the solution to the equation for each till ridge.

Similar calculations that have been done to determine basal shear stresses averaged over the transverse cross-section ($\bar{\tau}_b$) of many modern valley glaciers results in a range of values between 50 and 150 kPa (Schilling and Hollin, 1981, p. 208). The calculated basal shear stresses of the hypothetical backwasting valley ice lobe in Willoughby Gap fall well within this range (see Table 2). These values are consistent with the approximate yield stress of 100 kPa determined for ice in the laboratory. This value is approximate because ice is not perfectly plastic.

An estimate of the distance between cross-profile AA' (see Fig. 11) and the front of the valley ice lobe for each till ridge is obtained by integrating the modified basal shear stress equation in the form:

$$\Delta x = \frac{\rho g \frac{A}{P} \Delta t}{\bar{\tau}_b}$$

between the cross-profile AA' and the front of the lobe (Schilling and Hollin, 1981, p. 207-208). This results in:

$$x = \frac{\rho g \left(\frac{A}{P}\right)^2}{2\bar{\tau}_b}$$

This equation assumes the bed slope of the ice lobe is less than twenty degrees and that ice is a perfect plastic. The equation indicates that the longitudinal profile of a glacier should approach that of a parabola. This is confirmed by the profiles of many modern glaciers whose active margins are on land (Paterson, 1981, p. 153-161; Drewry and others, 1985, p. 10-12). Table 2 shows the solution to this equation for each till ridge. The actual distances to the front of the ice lobe are probably somewhat less, because glacial margins at proglacial lakes usually form cliffs rather than ramps.

Trend and Spacing of Till Ridges

Successive margins of the backwasting valley ice lobe in Willoughby Gap are plotted in Figure 1 on the basis of the above data. These plotted positions of the ice lobe explain the observation that the amount of change in crestal gradient and trend of adjacent till ridges is directly proportional to the map distance between the till ridges (see Fig. 10). The increase of crestal gradients down the flight of till ridges reflects the parabolic surface slope of the backwasting ice lobe. The change in trend down the flight of till ridges from south-southwest to southwest reflects the convergence of the lateral margins of a backwasting valley ice lobe toward its front.

The increase in average basal shear stress of the ice lobe ($\bar{\tau}_b$) down the flight of till ridges is directly proportional to the map distance between them (see Fig. 10). This may reflect decreasing basal water pressures toward the front of the backwasting ice lobe and a frozen toe. Basal water pressures would be expected to decrease toward the front of the lobe if the hydrology of the lobe was similar to that of the Malaspina Glacier, Alaska. The water table within the Malaspina Glacier lies close to the surface and increases in elevation from a few feet above sea level near the southern margin of the glacier to 2000 feet above sea level 20 miles to the north (Gustavson and Boothroyd, 1987).

TABLE 2. Basal shear stress ($\overline{\tau}_b$) for the backwasting valley ice lobe at each till ridge assumed to be a laterofrontal moraine, horizontal distance of front of ice lobe from each till ridge (x) and values used to calculate both of them.

Till Ridge	A		P		t		$\Delta t/\Delta x$	$\overline{\tau}_b$	x
	ft^2	m^2	ft	m	ft	m		kPa	km
TR7	2.6×10^6	2.4×10^5	6.6×10^3	2.0×10^3	820	250	0.06	63	1.00
TR6	*	*	*	*	770	235	0.07	72	0.85
TR5	*	*	*	*	730	223	0.08	79	0.67
TR3	1.6×10^6	1.5×10^5	4.6×10^3	1.4×10^3	650	198	0.11	104	0.49
TR2	*	*	*	*	580	177	0.15	117	0.29
TR1	*	*	*	*	570	174	0.15	115	0.29

See Stop 5 for explanation.

Constant values used: $\rho = 9\times10^2$ kg m^{-3} and $g = 9.8$ m sec^{-2}.

*A/P = 0.5t

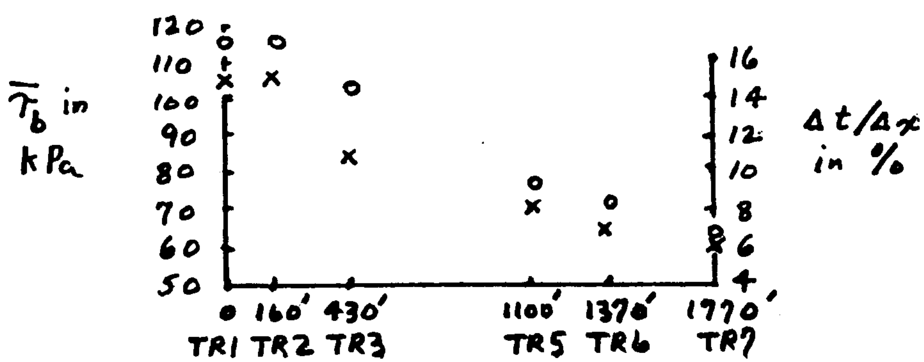


Figure 10. Relation of map distances between presumed laterofrontal moraines at Stop 5 to their crestal gradients ($\Delta t/\Delta x$) and the average basal shear stresses ($\overline{\tau}_b$) of the backwasting valley ice lobe adjacent to them. Map distances between till ridges are those shown in Figure 3. x is plot of $\Delta t/\Delta x$ versus map distance and o is plot of $\overline{\tau}_b$ versus map distance.

Shape of Till Ridges

Transverse cross-profiles of the till ridges, which were surveyed with tape measure, Suunto compass, and Brunton compass (see Figs. 7 and 8), are more similar to those of push moraines than they are to disintegration (ice-cored) moraines or dump moraines as described by Andrews (1975, p. 147-156). The break-in slope on the ice-contact flank of till ridges TR 3, 6, and 7, which previously was mentioned as evidence for a continuous till sheet prior to the incision of the meltwater channels, may instead be evidence of meltwater erosion on the lower portions of the ice-contact flanks of push moraines.

If these till ridges are laterofrontal moraines, then why are relatively clearly defined segments of such moraines not obvious elsewhere in the region? (Two short till ridges occur just above the "oversteepened" slope directly across the lake from Stop 5, between the 1500 and 1600 foot contours.) Boulton and Eyles (1979, p. 20) give two reasons for the larger concentrations of debris to emerge from the ice in the laterofrontal zone than elsewhere. (1) Near the front of an ice lobe, flow is away from the center line. (2) In a given time interval smaller displacements occur along the lateral margins of a backwasting valley ice lobe than at its front. The role of slope steepness in preserving these till ridges is demonstrated by the observation that all the till ridges occur within the middle section of the hillside, which has a gentler downslope gradient, overall, than the lower or upper sections.

Meltwater Channels

If the till ridges were constructed at the active margins of a backwasting valley ice lobe, then the meltwater channels must have been constructed at or near the margins of narrow zones of stagnant ice that bordered active ice. This situation is known for modern valley glaciers (Sugden and John, 1976, p. 241-243; Schytt, 1956, cited in Price, 1973, p. 112). The maps that show successive stagnant and active ice margins on the middle section of the hillslope above Stop 5 (see Fig. 11) assume meltwater channels C1 through C7 are ice marginal, but they could be submarginal. The landform map of the hillslope at Stop 5 (see Fig. 5) shows that these marginal meltwater channels are interrupted by channels that trend directly downslope (chutes). There are some arguments that support the view that these channels (chutes) formed subglacially.

These observations considered together lead to the conclusion that the meltwater channel floors are surfaces of erosion. The 50 to 100 foot width and straightness of the channels suggest high volumes of fast flowing water (Chorely and others, 1984, p. 298, p. 303-304; Mackin, 1948, cited in Derbyshire, 1962, p. 1117). Erratics on the floors of the channels are less numerous and usually more etched out than on the adjacent till ridges. (Clusters of erratics on the channel floors may reflect their original distribution in the till.) Local boggy areas on the channel floors suggest former pool sites. The single bedrock exposure of Wg on the floor of channel C7 and bedrock exposures where the north Pisgah trail crosses the southwest end of channel C7 suggest that till thickness on the channel floors is thin.

A rough estimate of the gradients of channels C6 and C7 was determined from long profiles of the entire length of these channels constructed from U.S.G.S. topographic maps at a scale of 1:62,500 and a 20 foot contour interval. The 6 percent gradient for C7 and 7 percent for C6 are a little steeper than some researchers referred to by Sugden and John (1976, p. 312) would expect for ice-marginal channels. These gradients, however, are approximately the same as those for the crestal gradients of the adjacent till ridges.

The supposed ice-marginal channels often grade to till benches at their upslope or downslope ends (see Fig. 5). This suggests that the till benches are extensions of meltwater channels whose valleyward flanks were stagnant ice rather than till ridges. (Similar benches are reviewed by Price, 1973, p. 111-112 and Derbyshire, 1962, p. 1116-1117.) Since these benches do not extend very far beyond the upslope end of the till ridges, their outer edges may be where meltwater flowed out of the ice onto the hillside.

The hanging upslope ends of channels C3 and C4 suggest places where meltwater flowed off the glacier and onto the hillslope. Five small gullies that cut across the till ridges (see Fig. 5) have more or less hanging upslope ends and join on-grade meltwater channels on their downslope ends. The acute angle formed between these small gullies and the meltwater channels is consistent with the direction of supposed meltwater flow. The hanging ends of these gullies, therefore, are places where considerably lower volumes of meltwater flowed off the glacier than at the upslope ends of the main meltwater channels.

The head end of chutes, i.e., channels that "fall precipitously downhill" (Derbyshire, 1962, p. 1124), interrupt ice-margin channels in at least three places. These may be sub-glacial channels that formed in the summer when ice-marginal meltwaters would more likely find a way beneath the ice directly downslope. Two of the chutes terminate at the 1300 foot contour, the level of proglacial Lake Willoughby. A similar situation is used by Sissons (1958, cited in Sugden and John, 1976, p. 314) as evidence of a water table in the glacier, which in this case was controlled by the level of proglacial Lake Willoughby.

The terraces along a west-northwest trending valley that crosses channels C6 and C7 can be explained in-the-light-of the deflection of ice-marginal meltwater to a subglacial path along a preglacial valley that was more directly downslope. Deflection of marginal meltwater is likely here, because this valley is aligned with a deep valley in the upper section of the hillslope that may have brought warmer water to the ice margin. Similar situations have been described for modern glaciers in southheast Alaska by von Engeln (1912, cited in Sugden and John, 1976, p. 312-313). Terraces I and II record the level of the cross-cutting valley floor when meltwater in C7 and C6, respectively, flowed across the valley. Terrace III, perhaps, records the level of the cross-cutting valley floor when meltwater in C6 was deflected down it.

UPPER SECTION

The upper section of the hillslope is also mantled with enough till to prevent exposures of bedrock. Landforms within this section are: a linear zone of closely spaced giant boulders similar to that described on the steep slope southeast of Mount Hor, three till benches, three chutes, and six in-out or col channels (see Fig. 5). These landforms were all constructed under or lateral to downwasting stagnant ice, The lack of obvious laterofrontal moraines may be related to the generally steeper slope of the upper section and a flow direction at the lateral margin of the valley ice lobe that was more nearly parallel to the hillslope.

The in-out or col channels (see Price, 1973, p. 107 and p. 113-117 for definitions) C8 through C13 included with their till bench extensions are hanging at both ends. The tendency for these channels to have flat boggy floors and steep sides is particularly noticeable in channels C8 and C11. The short till ridges on the valley side of these channels are fifteen to forty feet above the channel floors. Channels with hanging ends immediately above the floors of lower channels prove that the upper channels formed previously to the lower ones.

The floors of the chutes are generally narrower than the channels. At the head of the chutes is a col, i.e. a notch across a ridge.

SPUR OF HEDGEHOG MOUNTAIN

On the lower northwest spur of Hedgehog Mountain occur numerous exposures of bedrock in knobs, low ridges, low cliffs, and smooth gentle slopes on which there is, in places, a thin veneer of till or an occasional erratic. The contrast between a thick till cover on lower sections of the hillslope and little or no till on the spur (see Fig. 5) may be explained in-the-light-of the basal thermal regimes of steady state Laurentide Ice Sheets over the last three million years. (The slopes on the spur do not seem steep enough for postglacial mass wasting to account for the lack of a thick till cover.)

Hughes (1981, p. 231) believes relatively little quarrying and abrasion of bedrock was done during the disintegration of the Laurentide Ice Sheets, because: (1) disintegration flow lasted only about one tenth as long as steady state flow, i.e., flow at the maximum extent of the ice sheet; (2) steady state flow is stable and orderly, whereas disintegration flow is unstable and chaotic. The source, therefore, of glacial drift deposited at or near the margins of the last disintegrating Laurentide Ice Sheet was mostly debris, quarried and crushed during long intervals of steady-state flow. The presumed push moraines of the middle section of the hillslope, therefore, are composed mostly of subglacially derived debris that was plucked and crushed during steady-state flow and "reworked" during disintegration flow.

A reconstruction of basal thermal regimes of the last ice sheet at its maximum extent is given by Hughes and others (1981, p. 303). This reconstruction shows that the transition between a basal freezing zone on the north and a basal melting zone on the south in part coincides with the Appalachian Watershed Crest (Hughes and others, 1985, p. 141), which passes through the Willoughby region. This transition zone is where the ratio of frozen to thawed bed reaches a maximum and begins to decrease southward (Hughes and others, 1985, p. 142). Fluctuations in the boundary between frozen and thawed patches are likely to occur in such a transition zone. For example, frictional heat generated by ice shear at a frozen patch may lead to thawing, Thawing reduces the amount of frictional heat generated, which in-turn causes the bed to freeze again. Such freeze-thaw cycles may fracture bedrock and lead eventually to the plucking of bedrock by the glacier if it can exert a force great enough to overcome frictional drag (Hughes, 1981, p. 225-227; Boulton, 1974, p. 69). Frictional drag might generally be less over the Appalachain Watershed Crest because of thinner ice. A review of present research by Choreley and others (1984, p. 450, 443-445), however, leaves in doubt the conditions necessary for a glacier to erode its bed under an ice thickness of 8,250 feet, the thickness of ice over Willoughby range when the last ice sheet was at its maximum extent (Hughes and others, 1981, p. 305-306).

The change from a generally thick till cover on the lower sections of the hillslope above Stop 5 to a thin till cover on the lower spur northwest of Hedgehog Mountain may in part be related to the area's location near the boundary between Hughes' basal freezing and outer basal melting zones referred to previously. At this boundary, melting would likely occur in the valleys with freezing patches in the uplands. Thick till would accumulate where the base of the ice sheet was melting and little or no till would accumulated where the base was freezing (Hughes, 1981, p. 224-227).

Glacial Quarrying

The abundance of Wg clasts in the tills of subglacially derived debris on the lower and middle sections of the hillside above Stop 5 is consistent with Hughes' (1981, p. 227) conclusion that quarrying of bedrock can occur in a basal melting zone of the steady state Laurentide Ice Sheet. If the direction of flow of the steady state Laurentide Ice Sheet in the Willoughby region was south-southeast (Hughes and others, 1985), then the Wg clasts must have been plucked from Wg bedrock on the lower slopes and the lake floor within a mile to the north and northwest where the base of the ice was most likely melting (see Fig. 1).

Quarrying of bedrock can also occur in a freezing patch (Hughes, 1981, p. 227) which is presumed to have existed over the northwest spur of Hedgehog Mountain because of its thin till cover. The bedrock cliffs, knobs, ridges, and rock basin on the spur (see Fig. 5) all can be related to glacial plucking by a south-southeastward moving ice sheet.

The bedrock knobs and cliffs in this area reflect on a small scale the larger scale topography of the Willoughby Range. The ten to twenty-five foot cliffs that cross the spur are parallel to the N20°W trending cliffs that border the southeastern half of Lake Willoughby. The long dimension of low ridges and one rock basin more or less parallel those of Valley, Bartlett and Hedgehog Mountains, and Lake Willoughby, Long Pond and Nigger Pond (see Fig. 1). Since all these bedrock forms trend more or less parallel to the steady state flow of the Laurentide Ice Sheets, they may have been produced or modified by them.

Knobs on the spur tend to have gentle northwest and steep southeast facing slopes as do Wheeler Mountain, Mount Hor, the mountain between Hor and Wheeler, and Barton Mountain (see Figs. 12 and 13). The orientation of the asymmetry of these forms can be related to bedrock structure as well as to the direction of steady state ice flow. Over fifty percent of the exposed bedrock on the the spur is Wg, which occurs as vertical dikes up to fifteen feet wide and sills up to two feet wide. The wall rock is Dw with foliation that dips fifteen degrees to the northwest. Northwest dipping Wg sills could have formed erosionally resistant asymmetric strike ridges that predate the Laurentide Ice Sheets (see Dennis, 1956, p. 14). Rock would be more easily plucked by the south-southeastward flowing ice sheets from the already relatively steep southeast facing slopes of the strike ridges, because frictional drag that the ice sheet would need to overcome in order to pluck rock would be less on these slopes (Hughes, 1981, p. 225-227; Boulton, 1974, p. 69).

Lithology of Till Clasts Between Derby and Willoughby Plutons

The lithology of clasts on the generally lower northwest slopes of the Willoughby Range give further support to the hypothesis that the base of the Laurentide Ice Sheet in the Willoughby region was in part melting and in part freezing. At Stop 7, 2,500 feet northwest of Lake Willoughby, occurs a possible subglacial lodgement till with abundant clasts of Dg or Dw and a few of Derby granite. At Stop 6, near the crest of a ridge northeast of Westmore and scattered over the slopes between Stop 6 and the lake, are abundant erratics of Derby granite. South of Mill Brook, Wg erratics become more abundant than those of Derby granite. Basal freezing conditions of the Laurentide Ice Sheet over the Derby Pluton and perhaps the highlands just south of the Pluton (see Doll and others, 1961) caused plucked Derby granite clasts to become encased in regelation ice and then Dw clasts were plucked and encased. As the ice sheet continued to move farther south-southeastward, its base began to melt. Dw and Dg bedrock continued to be plucked and the resulting clasts were crushed, abraded, and lodged. Eventually enough regelation ice melted for Derby granite clasts to be released in abundance from the ice and become lodged on the lower slopes of the northwest flank of the Willoughby Range. South of Mill Brook most of the regelation ice with abundant Derby granite clasts had melted. Basal melting conditions changed to basal freezing conditions as the base of the ice sheet ascended to the lower spur northwest of Hedgehog Mountain. Similar explanations for the mapped distribution of clast lithologies in tills are given by Newman and others (1985) for northeastern Maine and by David and others (1985) for western Gaspe.

Col Channel

The N20°W trending bedrock cliffs across the lower spur northwest of Hedgehog Mountain may also be walls of a col channel (see Price, 1973, p. 111 for definition) caused, at least in part, by meltwater erosion. This hypothesis is consistent with the location of a chute, an extension of channel C9 that may be the down-glacier extension of the col channel. The "humped" long profile of the col channel suggests that meltwater was able to flow uphill. This could only occur if meltwater was flowing subglacially under hydrostatic pressure (Derbyshire, 1962, p. 1118; Price, 1973, p. 122). The alignment of the col channel more or less parallel to the flow of the ice sheet suggests that the subglacial meltwater occurred at the base of active ice (Sugden and John, 1976, p. 303-304).

Stop 6. Glacial Erratics at the crest of a ridge just north of Willoughby Pluton.

A large cluster of rounded erratics, that are mostly one to three feet across, occur on a ridge nearly one thousand feet above Lake Willoughby. Most of the erratics are Derby granite and a few are Dw or Dg. An explanation for the shape and lithologies of these erratics is given at Stop 5. A prominent rounded erratic of Derby granite is twenty feet in diameter (see Fig. 13). Bedrock of Dg or Dw is exposed two hundred feet west of the big boulder.

A rounded erratic, one foot in diameter, of coarse-grained pink feldspar gneiss occurs 50 feet east of the big boulder. Such gneisses are common in the bedrock of the Canadian Shield, one hundred miles to the north. A rounded erratic, two feet in diameter, of greenstone breccia occurs about one hundred and fifty feet east of the big boulder. Thirty miles to the north-northwest occur similar greenstone breccias in bedrock in the vicinity of Owls Head Mountain on the west shore of Lake Memphremagog. The bedrock belt in which greenstone is abundant occurs along the east flank of the Green Mountains and their extension into Canada, which can be seen on the Western horizon from Stop 6.

Stop 7. Lodgement Till and an overlying Kame Delta just north of Lake Willoughby.

A deposit with many characteristics of a lodgement till as described by Shaw (1985, p. 30-38) is exposed in a series of gullies on a steep bank 200 feet southwest of the Westmore Town Garage. The matrix of the till is bluish gray, compact silty clay with a horizontal foliation. The clasts are faceted and striated, up to one foot across, 95 percent Dw or Dg, five percent Derby granite, and occasionally greenstone or amphibolite.

The clasts in the overlying sandy gravel are 50 percent Dw or Dg and 50 percent Derby granite. The coarse sand that overlies the gravel contains abundant white feldspar. The difference between the lithologies of the clasts in the lodgement till and the overlying sands and gravels may be explained in the light of the proposed basal thermal regimes of the Laurentide Ice Sheet between the Derby Pluton and Willoughby Pluton described at Stop 5. The clasts of the lodgement till are mostly locally derived because the melting base of the Laurentide Ice Sheet did not reach regelation ice that contained abundant Derby granite clasts. During deglaciation, however, regelation ice with abundant Derby granite clasts melted.

The sand and gravel underlie a kame delta at about the level of Stuart and MacClintock's (1969, p. 145-146) Stage II of proglacial Lake Willoughby. When the sand pit is being worked, foreset beds of sand are exposed. These are overlain by cross-bedded topset beds of pebbly sand. The interface between the foreset and topset beds slopes moderately valleyward and in one place is broken along nearly vertical faults. This observation suggests that at least part of the delta was deposited on ice.

Figure 14. Crescentic gouges with shallow groves on Wheeler Mountain that trend N15°W toward the viewer. A pocket knife lies near the margin of a pegmatite dike that crosses the gouges.

Stop 8. Friction Cracks, Talus and Lateral Moraines at Wheeler Mountain.

Wheeler Mountain has a relatively gentle northwest facing slope and a steep east-southeast to south facing slope (see Fig. 1). It is underlain by granitic rock that is cut by occasional dikes of pegmatite and aplite. An eight foot wide vertical dike of a fine-grained mafic rock cuts the granitic rock at the extreme northeast portion of the Wheeler Cliffs. The strike of this dike is approximately parallel to that of Lake Willoughby.

Seven occurrences of friction cracks have been located, six in the upper half of the south facing slope on Wheeler Mountain and one on the north-west facing slope near its juncture with the south facing slope. At five of the friction crack localities are crescentic gouges and at the other two are crescentic fractures (see Embleton and King, 1975, p. 189 for definitions). The trends of the axes of these sets of repeated lunate gouges or fractures are within a few degrees of N15°W.

The two localities with the best display of friction cracks are on the eastern portion of the first large open slab traversed by the Wheeler trail. Both localities occur a few tens of feet downslope from the trail. A twenty foot long series of at least fifteen crescentic gouges up to eight inches across occurs along shallow two inch wide grooves (see Fig. 14). Three intersecting series of crescentic fractures occur just above where the dip on the slab steepens. One of these series of crescentic fractures consists of approximately fifteen lunate fractures up to four feet across.

Two localities of friction cracks occur on the lower portions of the large open south facing slab on Wheeler Mountain. This suggests that at least part of the time that the ice sheet flowed over Wheeler Mountain, no leeside cavity existed which some reseachers believe aids glacial plucking of bedrock at the top of the leeside of hills (Chorley and others, 1984, p. 447). Some observations at Stop 5 were shown to be consistent with the change in conditions at the base of the ice sheet from melting to freezing with increasing elevation. If a freezing patch of ice existed over Wheeler Mountain, then this area would likely be a zone of stick-slip glacier sliding with the production of friction fractures during periodic slip phases (Chorley and others, 1984, p. 447).

Three distinct zones of boulders, each of which are approximately 200 feet wide, occur at the foot of the east-southeast facing portion of the Wheeler Cliffs. The upper zone is normal talus. It lies on a steep slope of angular granitic blocks that are generally smaller than two feet across. The middle zone is on a less steep slope that consists of subangular granitic blocks, two to five feet across. The lower zone is on a gentle slope that consists of subangular to subrounded granitic blocks, 25 to 40 feet across. The origin of the two lower zones may be related to the lateral margin of a downwasting ice lobe.

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STRUCTURAL CHARACTER OF THE "PRE-SILURIAN" AND "SILURIAN" ROCKS AND THE NATURE OF THE BOUNDARY BETWEEN THEM IN CENTRAL VERMONT

VERMONT GEOLOGY VOLUME 5

VGS GUIDEBOOK 2

FIELD TRIP GUIDE I

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INTRODUCTION

Passing through central Vermont, parallel to the regional north-south structural fabric, is a tectonic boundary referred to informally by New England geologists as the Richardson Memorial Contact (RMC). This boundary has recently been formally named the Taconian Line (Hatch, 1982) and its location is shown in Figure 1. It is a line which separates predominantly non-calcareous, argillaceous-arenaceous rocks on the west side from calcareous argillaceous rocks on the east side (Currier and Jahns, 1941; White and Jahns, 1950). Hatch (1982) suggests, as have others before him, that rocks west of the line have experienced both the Taconic and Acadian orogenies whereas those east of the line have experienced only the Acadian orogeny. This author's studies have not produced any evidence that more orogenic events are recorded in the rocks on one side of the line than on the other.

Mappers working in Vermont during the 50's and 60's assigned Ordovician or older ages to rocks west of the RMC, based on correlation with fossiliferous rocks to the north in the Province of Quebec. These Ordovician rocks in central Vermont include the Missisiquoi Formation of Doll and others (1961) which includes rocks of the Moretown, Harlow Bridge and Cram Hill members, as well as a mapped "carbonaceous slaty phyllite". Stanley and Ratcliffe (1985) continue to assign an Ordovician age to these rocks which they refer to as the Moretown Formation (after Cady, 1956).

The rocks to the east of the RMC have historically been assigned Silurian and Devonian ages. This was based primarily on the combination of fossils found in the Shaw Mountain Formation along the western edge of this belt and certain assumptions about the stratigraphic relations between adjacent units. The conventional stratigraphy of these rocks (Doll and others, 1961) which make up the Connecticut Valley synclinorium is the Shaw Mountain Formation at the base, with the Northfield, Waits River and Gile Mountain formations above bringing the section well up into the Devonian. The age of all rocks "above" the Shaw Mountain rocks is based on the assumption that the regional structure is synclinal and that the rocks "young" to the east. All of these ideas are currently being re-evaluated.

Richardson and Camp (1918) assigned the Northfield and Waits River formations to the Ordovician period based on graptolite ages. Wallace Bothner of the University of New Hampshire has recently recollected samples from some of Richardson's localities and has determined that the rocks are Middle to Upper Ordovician (Bothner and Finney, 1986; Bothner, 1987). This new information, coupled with discoveries in Quebec of Ordovician fossils in rocks thought to be correlative with those of the Gile Mountain Formation (Bothner and Berry, 1985), raises major questions about the nature of the major structures in the rocks of central Vermont. A model calling for a simple synclinal structure for eastern Vermont is no longer tenable.

Examination of structural elements in rocks of the Moretown Formation (west of the RMC) and the Northfield and Waits River formations (east of the RMC) reveals early isoclinal folds which have been isoclinally refolded. The styles of these folds are very similar as can be seen in Figures 2 and 3 which show folds from the Moretown Formation at Stop 1 and the Northfield Formation at Stop 6, respectively. Younger crenualtion and small-scale asymmetric folds with moderate west-dipping spaced cleavage are also common to both of these assemblages. Since these lithotectonic units are separated by rocks of the Shaw Mountain Formation (Silurian), it has previously been assumed that the structural elements common to them were produced during Devonian time by the Acadian orogeny. Since the isoclinal folding events have not been observed in rocks of the Shaw Mountain Formation, and since the rocks now found on both sides of that formation may have been present as far back as the Ordovician period, the time of deformation of rocks in the Connecticut Valley trough is uncertain.

Recent mapping by this author in the Northfield - Montpelier section of the Dog River Valley has revealed that the RMC is a complex zone of shearing. This zone separates the readily recognizable rocks of the Moretown Formation from those of the Northfield and Waits River formations. Lithologic units (formations) are generally bound by faults, and the rocks often are stacked with faults more closely spaced than the distance between outcrops, making certainty in mapping impossible. Despite the inability to always trace contacts at a detailed scale, some major relationships can be seen in Figure 1. The pattern of contacts along the RMC is one of braiding. Slices of different formations have been transported within the fault zone and stacked together.

The map pattern of braiding and stacking can be seen at the scale of thin sections in which small lithons have been transported and stacked. Another characteristic of this style of deformation is that the rocks in the interior of a transported slice tend to remain relatively undeformed by the shear. The strain appears to be taken up within discrete zones between the transported "wedges". In some cases, this results in undeformed fossils being preserved in a mylonitized shear zone (Stop 4).

Since rocks of Shaw Mountain Formation are confined to the zone of intense and pervasive shear that has been named the RMC and the Taconian Line, using fossils from that unit to assign ages to adjacent rocks is inappropriate.

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*A more thorough discussion and review of this subject can be found in Westerman (1987).

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ROAD LOG

The field trip starts at Montpelier High School, located on Route 2 just south of the bridge over the Winooski River at the west edge of town.

0.0 Head west on Route 2 (toward Middlesex).

2.8 Park at the east end of the long roadcut which exposes a variety of granulites and phyllitic schists of the Moretown Formation as mapped by Cady (1956).

Stop 1. Moretown Formation

Included in these exposures are "carbonaceous slaty phyllites" which Cady mapped as a separate unit. Walk to the west end of the section (not quite half a mile), noting the structures in these rocks as you work your way along. Bedding is preserved locally and is folded in various styles. Near the western end of the section there is an easily recognized synform of a thick quartzite bed. Close examination of the folded bedding surface shows a well-developed cleavage parallel to the bedding, both of which have been folded and then cut by the regional axial plane cleavage.

As you return eastward along the outcrop, note the locally well-developed cleavage which dips less steeply westward than the predominant cleavage. This younger cleavage is associated with crenulations indicating west-over-east movement. Younger still than the crenulation cleavage is a locally prominent cleavage associated with shear. This youngest fabric parallels the upright cleavage but has obliterated the crenulations. It is particularly well-developed near lithologic contacts, especially those of the rusty, black phyllite.

If time allows, cross to the south side of Route 2 and examine the glaciated pavement exposed between Route 2 and I-89. Abundant small-scale, assymmetric folds are well exposed there. Also exposed are pinstripe textures which probably resulted from pressure-solution during dynamothermal metamorphism.

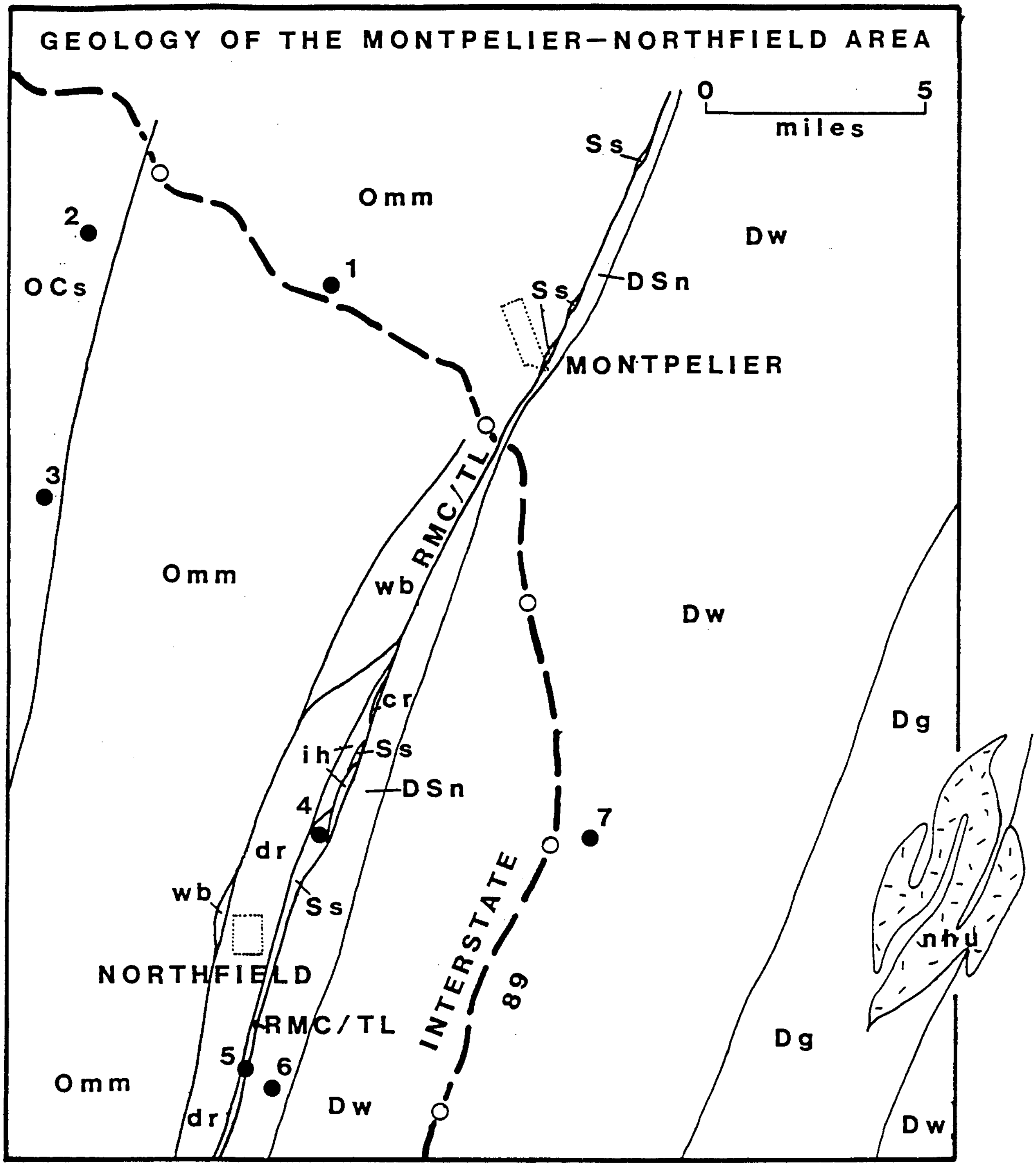


Figure 1. General geology of the Montpelier - Northfield area, modified from Doll and others (1961). Symbols are as follows: **OCs** = Stowe Formation, **Omm** = Missisquoi Formation, **wb** = West Berlin formation, **dr** = Dog River formation, **cr** = Crosstown Road formation, **ih** = Irish Hill Road formation, **Ss** = Shaw Mountain Formation, **DSn** = Northfield Formation, **Dw** Waits River Formation, **Dg** = Gile Mountain Formation, **nhu** = Barre Granite, RMC/TL = Richardson Memorial Contact / Taconian Line.

Figure 2. Isoclinally folded beds and early cleavage(?) in the Moretown Formation at Stop 1. The axis of the fold is oriented 42,N25E and the axial plane cleavage is N36E,79W.

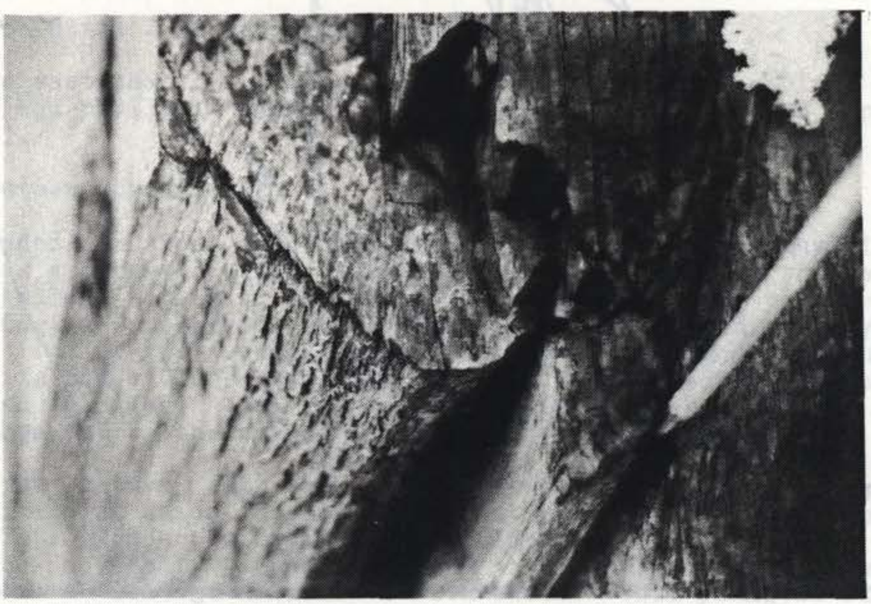


Figure 3. Isoclinally folded beds and early cleavage(?) in the Northfield Formation at Stop 6. The axis of the fold is oriented 45, N14E and the axial plane cleavage is N23E,79W.

Continue west on Route 2 to the town of Middlesex.

6.1 Turn left (south) on Route 100B headed for Moretown.

8.2 Cross the Winooski and following 100B as it winds its way up into the Mad River Valley. Cross the first bridge over the Mad River (you're now on the west side).

8.4 Park at the first roadcut.

Stop 2. Moretown/Stowe Formation contact zone

These rocks are located essentially at the Stowe (west)/Moretown (east) contact as mapped by Tom Ray (personal communication). They contain no primary structures at the eastern end of the outcrop where phyllonites have formed during the cataclasis of these rocks. Phyllonites are rocks which have textures like phyllites but the textures were produced by mechanical grain size reduction during shear. The quartz veins which are quite common in rocks of the Stowe Formation have here been totally dismembered by shear. Comparison with the structural history as seen at Stop 1 would suggest that this fabric correlates with the youngest major deformation, namely an episode of shear parallel to the most steeply-dipping surfaces. Here these surfaces trend N17E, 80W and kinematic indicators suggest a west-over-east sense of motion.

Continue south on Route 100B until you have crossed back to the east side of the Mad River and have come to Moretown.

13.0 Turn left up the gravel road opposite the clapboard mill (the Northfield Road, which will become the Cox Brook Road on the other side of the high ground).

13.2 Bear right at the top of the first rise and continue.

13.9 Go straight through the four corners and continue up onto and across the "deltaic" deposits being quarried for gravel.

14.7 After entering the woods again, bear right at the first fork.

14.9 Bear right at the second fork and start up a long steep grade which takes you to the Moretown Gap in the Northfield Range.

15.7 Park along the side of the road and walk toward Northfield (the direction your car is facing).

Stop 3. Moretown Formation

Various granulites and phyllitic schists of the Moretown Formation are exposed here along with a lamprophyre dike which trends about N75W, V. This dike is most likely Cretaceous in age and postdates all the major structures in the rocks which it intrudes. The chemical composition of the dike is as follows (after correction for 4.62 weight percent loss on ignition):

Major Elements (wt%)	Minor Elements (ppm)
SiO_2 = 47.8	Ni = 96
Al_2O_3 = 15.6	Cr = 40
CaO = 8.07	Rb = 50
MgO = 5.22	Sr = 890
Na_2O = 3.69	Y = 40
K_2 = 1.98	Zr = 320
FeO^* =10.7	Nb = 80
MnO = 0.14	Br = 590
TiO_2 = 3.00	
P_2O_5 = 0.88	

* Total Fe

The purpose of this stop is to review the textural and structural characteristics of the "pre-Silurian" rocks before we head east. Small faults with both normal and reverse senses of motion are abundant here, locally juxtaposing pinstripe gneiss and greenschist. Associated with some of these faults are asymmetric folds, locally with planar limbs and acute hinges (chevron style). The wide variety of small-scale structures exposed here and the apparently conflicting indications of the senses of motion make this a particularly useful location for learning about ambiguities.

We will be passing a water well just a quarter mile from this location which has been drilled to 3,400 feet in the hope of finding a supply for raising trout. The hole recently had to be abandoned due to problems encountered when the workers tried to cement the sides of the hole below the casing (which goes to 2,400 feet). On the advice of a clairvoyant, a new hole will be sunk this winter to a depth of 3,500 feet, "no more than 4.5 feet NNW of the first hole".

Continue east on this road.

18.2 Cross bridge.

19.6 Road turns to tar. Continue through three covered bridges.

21.8 Junction with Route 12. (You may wish to purchase lunch material here at the Northfield Falls General Store). Turn right (south).

21.9 Turn left onto Davis Avenue. Work your way eastward.

22.3+ Park under the power line.

Stop 4. "Cram Hill"/Shaw Mountain Formation contact zone

Rusty and buff weathering micaceous quartzites and bluish gray phyllites of what I refer to as the Dog River formation (not capitalized) are exposed under the powerline and to the west of it further up hill. Walk up the woods road to the north, noting the large cliff of quartz-pebble conglomerate of the Shaw Mountain Formation. Continue north along the road and powerline until the powerline bends easterly and the road splits off due east. Take the road to the east (up slope). You will find mylonitized limestones exposed in the road bed (in a gully and so soft your hammer sinks to your hand) and unstrained white limestones just 20m off to the north in a small quarry. Crinoids are present on weathered surfaces of the limestone. A bit further up hill to the east, an outcrop of conglomerate crosses the road. Off to the south a few paces into the woods is a small quarry of the conglomerate in which varying degrees of strain can be observed. All of the textures and structures in this section of the Shaw Mountain Formation seem to be either primary or directly related to shear. No folding features have been seen.

Return to the cars and retrace your steps back to Route 12.

22.8 Turn left (south) and proceed through town and past Norwich University.

25.5 Bear right (southwest) on Route 12A.

26.3 Turn left (east) on Lover's Lane which parallels Sunny Brook and park.

Stop 5. "Cram Hill/Shaw Mountain/Northfield formations contact zone

We will be spending a fair amount of time here, walking eastward from Route 12A toward Route 12. Exposures along Route 12A just south of the turn-off are mapped as Cram Hill Formation by earlier workers and are what I have called the Dog River formation for convenience. They are rusty weathering quartzites and phyllites like those exposed along the western side of the Shaw Mountain Formation as far north as Riverton and for an as yet undetermined distance to the south. Note the upright lithic layering and well-developed cleavage which parallels it. This cleavage is cut by a crenulation cleavage which is also present in the rocks which we will see further east in the Northfield Formation.

Walk east noting the appearance and structures of the Cram Hill rocks. After the last exposure of rocks which are certainly still part of the same lithologic package (containing the two key rock types), the next exposure is in a new logging road on the south side of Lover's Lane. Here, deeply weathered green schists with transgressive biotite(?) alternate with saprolitic pink-weathering limy schists. These rocks are thought to be attenuated slivers of lithologies which occur within the RMC zone further to the north back at Stop 4. Uphill along Winch Hill Road, fossiliferous limestones have recently been shown by K.E. Denkler and A. Harris (written communication, 1986) to contain Silurian-Early Devonian morphotypes conodonts.

Continuing east, note the boulders of Shaw Mountain conglomerate in which the quartz pebbles show varying degrees of strain. The next outcrop exposed along the south side of the road is mylonitized quartz pebble conglomerate where locally, strain has been so extensive that no pebbles are left. The rocks formed under these conditions have a greenish-white cream-colored tint, but they are so sulfidic that the outcrop is a deep rust color and broken pieces of fresh rock smell. Just a few paces east, a lense of coarse conglomerate has been preserved between shear surfaces. Continue east to the next corner and an outcrop of slates of the Northfield Formation is present.

Since no powerlines have been run along this road, it is a convenient place to work with magnetic and electromagnetic surveying equipment. If time permits we can experiment with the usefulness of these instruments in locating changes in lithology and the presence of mineralized faults which produce a conducting horizon.

Return to the cars and drive east to Route 12.

26.9 Entrance road to quarry of Northfield Formation on the left (north).

27.0 Go straight across Route 12 and park at the entrance to the new access road to I-89 (General Harmon Highway).

Stop 6. Northfield Formation

Walk up Route 12 and examine the rocks in the fresh roadcut on the east side. In the central position of the outcrop and near the top, an isoclinal synform occurs in these phyllitic slates. This fold provides an opportunity to see if an early cleavage is folded along with the bedding or whether these folds are, in fact, the first generation. Note also the crenulation cleavage and how its orientation is the same as was seen in the rocks of the Cram Hill Formation back at the start of Stop 5.

Continue east up the new access road to I-89. Big road cuts of the Northfield Formation occur on both sides. Megacrenulations can be seen on the north side, as well as abruptly terminating and crossing(?) basalt dikes. The outcrops end temporarily one half mile up the road. On the south side are rocks interpreted to be transitional between the Northfield and Waits River formations. Bedding is thin and very well preserved as triplets which face predominantly west (although at the west end of the outcrop they appear to face east). The bases are thin "choclate-weathering" calcite rich sands. These are overlain abruptly by less calcareous sands which grade quickly to phyllite tops. Cleavage is well-developed and parallels the beds.

30.0 Head north.

33.9 Turn east off the interstate at the next exit, heading east toward South Barre.

34.3+ Park and cross to the north side of the road.

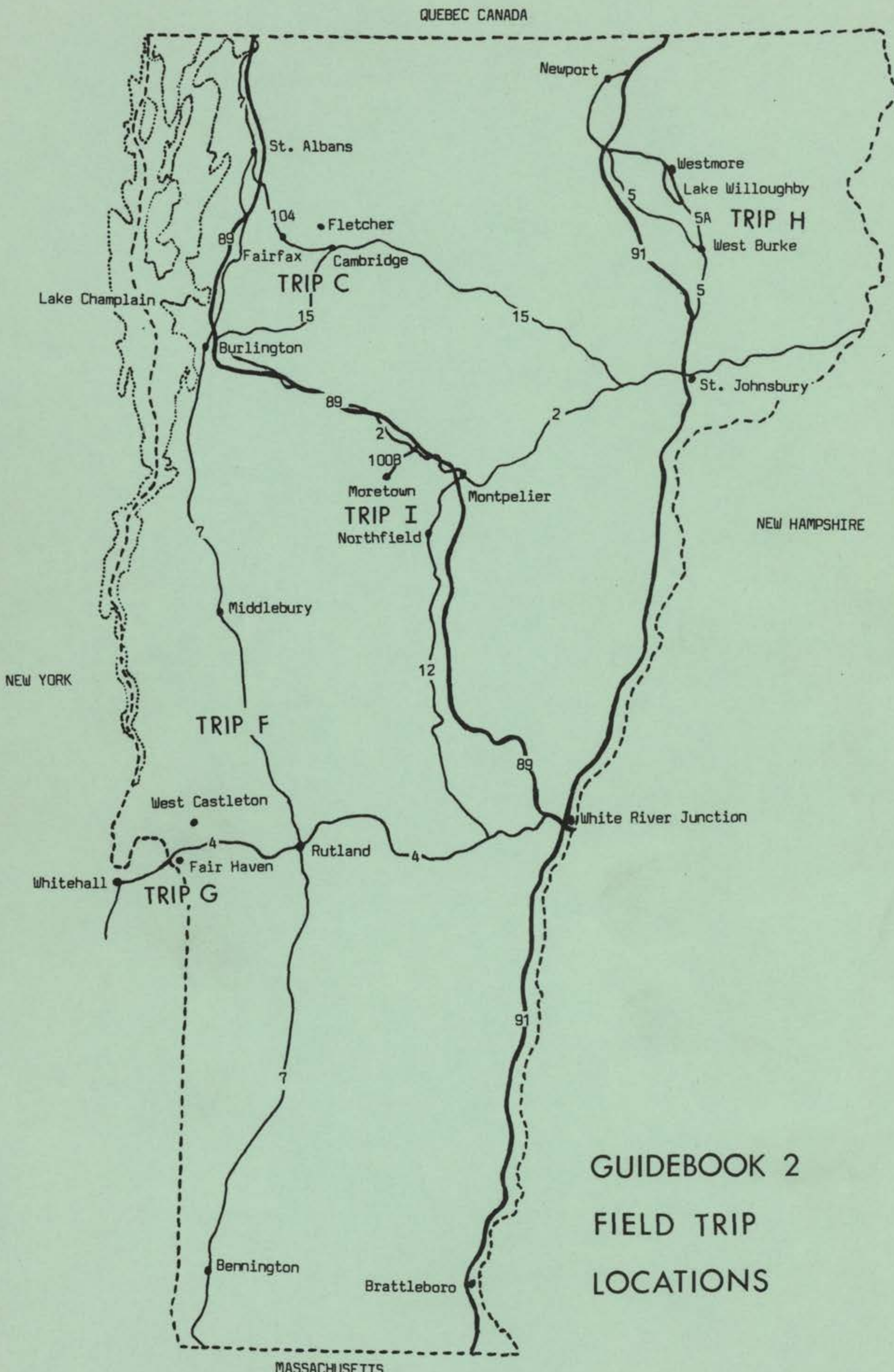
Stop 7. Waits River Formation

The calcite content of these rocks has allowed them to behave more plastically than we have seen at some other locations. Evidence for at least three episodes of folding can be documented here; the earliest may have been a soft-sediment deformation.

In a quarry to the south, a spectacular example of asymmetric folds is exposed on a vertical wall.

End of trip.

Turn around and return to the Interstate 89. Turn north to get to Montpelier or south if you want to retrace your route to Northfield. Head for Sambel's on the Common in downtown Northfield for a social hour followed by dinner at 6 P.M.



GUIDEBOOK 2

FIELD TRIP

LOCATIONS