Trip C

BEDROCK GEOLOGY OF THE SOUTHERN PORTION OF THE HINESBURG SYNCLINORIUM

by

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INTRODUCTION

Purpose

The geology of west-central Vermont has attracted the attention of geologists since the early work of Hitchcock, et al. (1861), and Logan (1863), to mention only two. During this century notable contributions have been made by Perkins (1910), Keith (1923, 1932, 1933) and, in particular, Cady (1945, 1960). The mapping of Welby (1961) west of the Champlain thrust, and Stone and Dennis (1964) in the Milton quadrangle just north of Burlington, completes the present state of knowledge of the Hinesburg synclinorium and immediate surrounding areas (Fig. 1, inset map). These works are incorporated in the Centennial Geologic Map of Vermont (Doll, et al., 1961).

Among several major problems still remaining to be resolved is a clearer understanding of the effects of the Acadian and subsequent orogenies or disturbances on the structures of west-central Vermont. Inasmuch as rocks of Upper Ordovician through Devonian age are not present in western Vermont, it becomes most difficult to evaluate the relative importance of Taconic and younger events.

Answers to this central problem with its many corollaries can hopefully be provided by detailed studies involving a combination of stratigraphic, petrologic, geochronometric, and structural approaches. It is the purpose of this trip to examine the geology of the Hinesburg synclinorium and to show how recent work may be helpful in answering these problems in the future.

Acknowledgements

It goes without saying that the results of the many studies in western Vermont and Quebec must be acknowledged in forming the background for the present and future work in the Hinesburg synclinorium and adjacent areas. Although each of these cannot be mentioned individually, under-

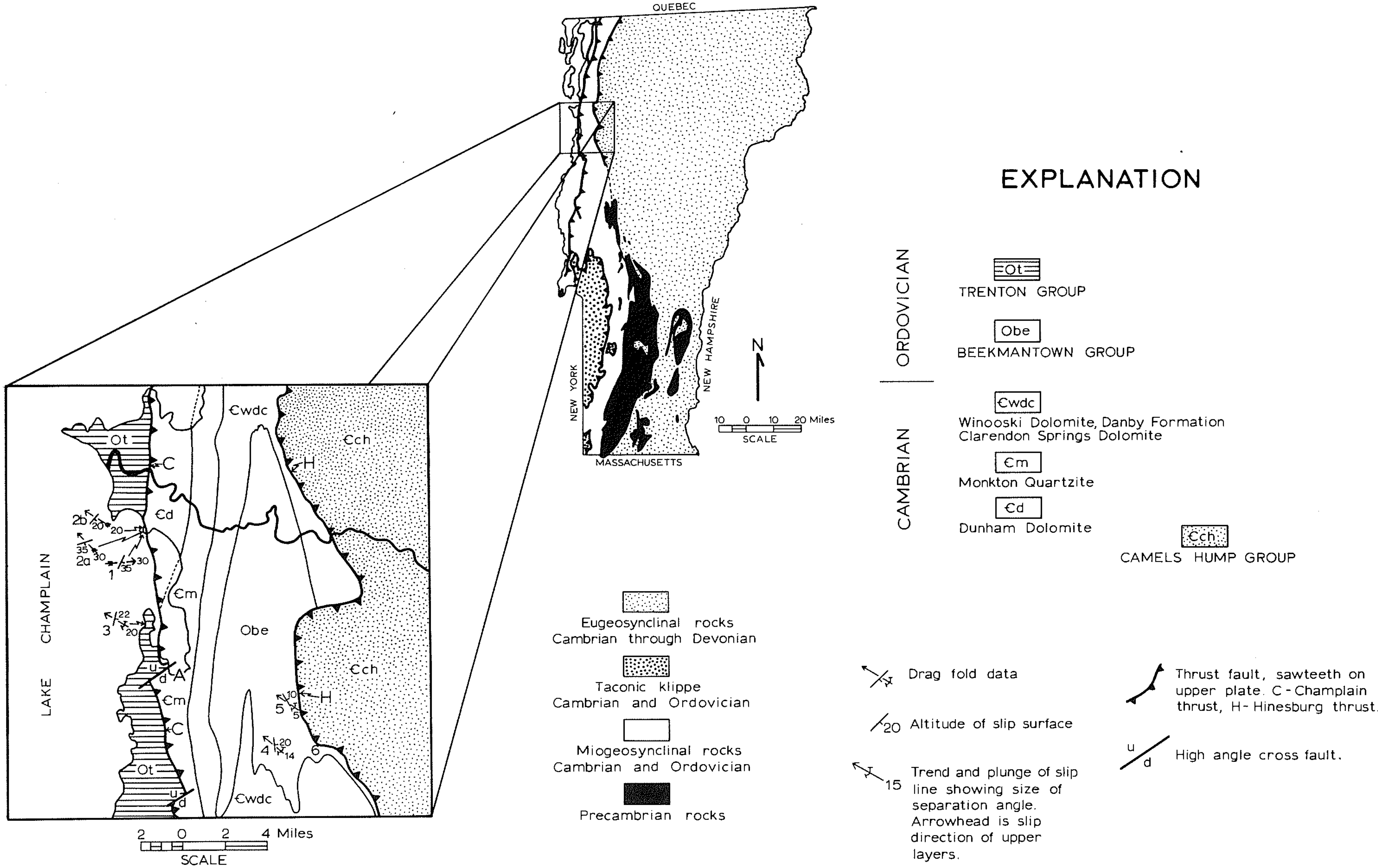


Figure 1 Generalized geologic map of Vermont and the Hinesburg synclinorium. Slip line directions at localities 1-6 are shown in more detail on figure 3.

standing of this area has been greatly assisted by syntheses of Cady (1945, 1960, 1968), Doll, et al. (1961), Rodgers (1968) and Zen (1967, 1968).

Recent work in the southern portion of the Hinesburg synclinorium has been carried out by various students at the University of Vermont in conjunction with my own work in the area. Information on the Shelburne access area is largely drawn from an unpublished research report by Charles Rubins. Quartz deformation lamellae studies have been done, in part, by Charles Rubins, John Millett, Edward Kodl, Robert Kasvinsky, Arthur Sarkisian and Evan Englund, whereas the drag fold data from localities 4 and 5 (Fig. 1) was collected by John Pratt. Data at stop 4 was provided by James Dinger. Charles Doll, State Geologist of Vermont, and E-an Zen kindly reviewed this paper and made several important suggestions. The geologic map of the Burlington quadrangle (available only to participants) was kindly provided by the Vermont Geological Survey. Mistakes in interpretation of the existing knowledge of the area are my responsibility.

REGIONAL GEOLOGY

The geology of western Vermont forms the northern extension of the Ridge and Valley provinces of the Appalachians. Carbonates and quartzites with minor amounts of shale, except in northwestern Vermont, characterize the autochthonous Cambrian and Lower Ordovician section and represent shallow-water shelf sediments of the miogeosynclinal belt. The Lower and Middle Cambrian part of this section wedges out to the west with rocks of Upper Cambrian age resting with angular discordance on the Precambrian rocks of the Adirondack dome. To the east in Vermont the Cambrian and Ordovician section thickens drastically and changes facies to graywacke and shale which represent the western part of the eugeosyncline belt of New England. Volcanic rocks, which are totally absent in the miogeosynclinal belt, are present here and become more abundant farther to the east in New Hampshire. In most places the change in sedimentary facies has been complicated by subsequent deformation, but in such areas as Milton (15 miles north of Burlington) it is relatively undeformed and appears abrupt. Rodgers (1968) has recently suggested that this change represents the eastern edge of the carbonate shelf which dropped off rapidly into the eugeosyncline to the east much like the present eastern edge of the Bahama Banks.

The upper part of the autochthonous section is composed of several thousand feet of shale with thin beds of silty dolostone and limestone, commonly graded and current rippled (Hawley, 1957), which represent the westward advance of the eugeosynclinal or transitional zone over the carbonate shelf during Middle Ordovician time (Cady, 1960; Zen, 1968). At present these rocks underlie most of the western part of the Champlain Valley whereas the older carbonate and quartzite underlie the central part of west-central Vermont (Cady, 1945).

The structure of the autochthonous part of the Champlain Valley consists of a series of east-dipping, imbricate foreland thrusts and west-facing asymmetrical or overturned folds somewhat typical of the Ridge and Valley province although the belt here is much narrower than in the central and southern Appalachians. Major culminations near Monkton and near Milton divide the belt into three synclinoria which are known from south to north as the Middlebury, Hinesburg and St. Albans synclinoria (Cady, 1945). Allochthonous rocks exist in both the Middlebury (Taconic klippen) and the St. Albans synclinoria (8 small isolated outliers) whereas no such structure has been recognized in the central Hinesburg synclinorium.

Although numerous thrusts of varying extent have been recognized in the Champlain Valley, the Champlain and Hinesburg thrusts are the largest and bound the three synclinoria on the west and east respectively (Fig. 1.). The Champlain thrust (Keith, 1923), first recognized as such by William Logan (1863) and hence part of Logan's line, extends for a distance of approximately 120 miles and places Lower Cambrian dolostone and quartzite (Dunham Dolomite north of Burlington, Monkton Quartzite to the south) of the Rosenberg slice (Clark, 1934) on Middle Ordovician shale to the west. North of Shelburne this fault dips eastward at less than 15° but to the south it is gently folded and offset by high angle faults (Welby, 1961). The Hinesburg thrust (Keith, 1932) is the easternmost thrust in the Champlain Valley and extends for approximately 60 miles. It places the Lower Cambrian Cheshire Quartzite and the older Underhill Formation (schist and phyllite) of the western edge of the eugeosyncline (transitional sequences of Zen, 1968) on the eastern limb of the Hinesburg and St. Albans synclinoria. Although the stratigraphic throw is probably not as great as the Champlain, which is estimated at 9000 to 10,000 feet (Shaw, 1958, Stone and Dennis, 1964), the Hinesburg brings to the surface older rocks than the Champlain thrust. In contrast to the gentle eastward dip of the Champlain thrust, the Hinesburg dips both east and west at various angles and in such areas as the southern part of the Milton quadrangle definite evidence of folding of the fault surface has been found (Stone and Dennis, 1964).

Minor thrusts and high angle faults of either normal, reverse, or strike-slip displacement further complicate the geology in west-central Vermont. Such thrusts as the Muddy Brook north of the Winooski River and the Monkton south of Lewis Creek disrupt the stratigraphic continuity within the Rosenberg slice (Doll, et al., 1961) in the Champlain Valley. West of the Champlain thrust in northwestern Vermont a series of parallel thrusts (Hawley, 1957, Fisher, 1968) repeat the Ordovician section and can be interpreted as imbricate faults or possibly offshoots to the eastern master faults. Southwest of Middlebury a series of imbricate faults also mark the southern end of the Champlain thrust (Cady, 1945, Welby, 1961). Steeply dipping longitudinal and cross faults of either normal, reverse, or wrench origin further complicate the geology between the Champlain thrust and the Adirondacks (Welby, 1961, Doll, et al., 1961).

The rocks of western Vermont have been regionally metamorphosed with the grade increasing from chlorite in the Champlain Valley to biotite or higher in the Taconic and Green Mountains. The pattern of isograds accords with those of the rest of central New England in which Middle Paleozoic rocks are involved and hence is thought to be related to recrystallization during the Acadian orogeny (Thompson and Norton, 1968). However, isotopic age dates older than 400 m.y. in Quebec (Richard, 1965), along the Sutton-Green Mountain anticlinorium in northwestern Vermont, (Cady, personal communication; cited by Richard, 1965, p. 530) and in the Taconic allochthon (Harper, 1967) strongly suggest that pre-Silurian metamorphism and deformation are responsible for the fabric in that part of the Appalachians.

Igneous rocks are limited primarily to minor east-west dikes and are predominantly bostonite and lamprophyre (Welby, 1961). A small laccolith of syenite underlies Barber Hill in Charlotte (Migliore, ms; Dimon, ms). These bodies are clearly post orogenic inasmuch as they cut the folded and faulted country rocks. K-Ar determination on biotite from one such lamprophyre on Grand Isle yielded a radiometric age of 136 $\pm$ 7 MY (Zartman et al., 1967) and thus suggests a late Mesozoic age for its emplacement.

REGIONAL GEOLOGIC HISTORY

Although the interpretation varies somewhat, the geologic history of western Vermont has been well portrayed in recent papers by Zen (1967, 1968) and Cady (1960, 1967, 1968). Interested readers are referred to these and other papers (mentioned by these authors) for more information on the subject. As an aid to the reader, however, a brief summary follows of those events which are perhaps best exhibited in the Middlebury synclinorium where the Cambrian and Ordovician section is most complete.

During the Cambrian and Lower Ordovician most of western Vermont was the site of shallow-water deposition on a carbonate bank whose eastern edges dropped off steeply into the deep-water facies of the eugeosyncline. The first signs of tectonism appeared in the late Early or Middle Ordovician when, according to Zen (1967, 1968), high angle faults apparently divided the shelf into a series of grabens and horsts. Erosion ensued, accompanied by a short interval of carbonate deposition. Gradually the carbonates in the post-unconformity, Middle Ordovician section became more shaly, heralding the wave of argillaceous material that was spreading westward from the now uplifted former eugeosyncline. During this time (Zone 13 of Berry, 1960) the first segments of the Taconic allochthon were emplaced. The remaining slices (a total of six comprise the Taconics, Zen, 1967) followed during the Ordovician and were possibly in place before the formation of the three synclinoria in the autochthon

TABLE I.

COMPOSITE STRATIGRAPHIC SECTION FOR THE HINESBURG SYNCLINORIUM AND AREA WEST OF THE CHAMPLAIN THRUST AND EAST OF THE HINESBURG THRUST

(Cady, 1945; Doll, *et al*., 1961)

Middle Ordovician

Hathaway Formation*

Missing but present north of the Lamoille River.

Iberville Formation

Noncalcareous shale, rhymthically interbedded with thin beds of silty dolomite and in the lower part with calcareous shale.

Stony Point Formation

Calcareous shale that grades upward into argillaceous limestone and rare beds of dolomite.

Cumberland Head Formation

Missing but well exposed in Grand Isle County.

Glens Falls Formation

Thin bedded, dark blue-gray, rather coarsely granular and highly fossiliferous limestone.

Orwell Limestone

Missing but present south of Charlotte.

Middlebury Limestone

Missing but present south of Charlotte.

*Only those formations encountered in the course of the trip will be described. Kindly refer to the Centennial Geologic Map of Vermont for other formation descriptions.

Lower Ordovician

Chipman Formation

Missing in the Hinesburg synclinorium but present south of Charlotte.

Bascom Formation

Beds of light bluish gray calcitic marble with laminae and thin beds of siliceous phyllite. Beds of brown-orange weathering dolomite 1 to 3 feet thick. The upper part becomes more phyllitic and is mapped separately as the Brownell Mountain Phyllite. Contact against typical Cutting dolomite is gradational.

Cutting Dolomite

Massive whitish to light grayish weathering dolostone, dark gray on the fresh surface. Sand-size quartz grains present in places especially near the base where sandy laminae are more abundant and brecciated in places. Black chert nodules are found in the upper part. Contact is sharp with sandy dolomite above and white calcitic marble of the Shelburne Formation below.

Shelburne Formation

Massive whitish gray weathering calcitic marble that is white on the fresh surface. Laminae of green phyllite present in the eastern part of the Hinesburg synclinorium. Contact is sharp with typical calcitic marble above and gray dolomite with quartz knots below.

Upper Cambrian

Clarendon Springs Dolomite

Massive, gray weathering dolomite with numerous knots of white quartz crystals. Black chert commonly found in the upper part. Contact gradational with the Danby Formation.

Danby Formation

Beds of gray sandstone interlayered with beds of dolomite 1 to 2 feet thick and sandy dolomite 10 to 12 feet thick. Sandstones are cross laminated. Beds 1 to 3 feet thick. Contact gradational.

Middle Cambrian?

Winooski Dolomite

Beds of very light gray to buff weathering dolomite, gray to light pink on the fresh surface. Thin siliceous partings characteristically separate beds of dolomite which range in thickness from 4 inches to 2 feet. Contact with the Monkton Quartzite is gradational and is placed above points where quartzite beds over 1 foot thick are separated by beds of dolomite less than 25 feet thick (Cady, 1945, p. 532).

Lower Cambrian

Monkton Quartzite

Beds of fine to coarse grained pink and brick red quartzite interlayered with minor beds of pink and red dolomite and thin laminae of red, green and purple shale. Beds of dolomite are more numerous in the lower part than the upper. Contact with the Dunham Dolomite is gradational and is placed where quartzite beds 1 foot thick are separated by beds of dolomite less than 25 feet thick (Cady, 1945, p. 532).

Dunham Dolomite

Massive, buff weathering dolomite that is pink and cream mottled or buff to gray on the fresh surface. Siliceous veins and laminae are present in places. Grains of sand size material common in places. Basal contact with the Cheshire Quartzite is gradational. Upper part of the Dunham north of Burlington is sandy and has been mapped separately as the Mallett Member.

Cheshire Quartzite

Massive, very thick bedded, white quartzite. Lower part is brown weathering and is more argillaceous and less massive. East of the Hinesburg thrust the Cheshire is more argillaceous. There the contact with the Underhill Formation is gradational and placed above "the frankly phyllitic, dark gray to greenish phyllite and below the rather characteristic mottled gray silty impure quartz schist of the Cheshire" (Stone and Dennis, 1964, p. 26).

Cambrian

Underhill Formation

Fairfield Pond Member: Predominantly green quartz - chlorite - sericite phyllite. Quartz grains common.

White Brook Member: Chiefly brown-weathered whitish, tan and gray sandy dolomite.

Pinnacle Formation

Schistose graywacke, gray to buff, with subordinate, quartz-albite-sericite-biotite-chlorite phyllite. Includes the Tibbit Hill Volcanic Member.

TABLE II

Age: System	Series	Group	Location: Saratoga Springs, New York Mohawk Valley (Fisher 1965)	West-central Vermont (Doll et. al. 1961)	West Limb of the St. Albans Synclinorium (Doll et. al. 1961) (Shaw 1958)	Lincoln Mtn.- Enosburg Falls Anticline East of Hinesburg Thrust (Doll et al 1961) (Stone and Dennis 1964)
ORDOVICIAN	Middle	Trenton	Schenectady Shale Canajoharie Shale Shoreham Limestone Larrabee Limestone	Hathaway Formation Iberville Formation Stony Point Formation Cumberland Head Formation Glens Falls Formation	Morse Line Formation	
		Black River	Amsterdam Limestone	Orwell Limestone		
		Chazy		Middlebury Limestone		
	Lower	Beekmantown	Chuctanunda Creek Dolostone Gailor Dolostone	Chipman Formation Bascom Formation Cutting Dolomite Shelburne Formation	Highgate Formation	
CAMBRIAN	Upper		Hoyt Limestone Mosherville Sandstone Thesesa Formation Potsdam Sandstone	Clarendon Springs Dolomite Danby Formation	Gorge Formation Sweetsburg Fm.: Hungerford Sl., Rockledge Congl., Skeels Corner Sl., Mill River Congl., St. Albans Sl.	
	? Middle ?					
	Lower			Winooski Dolomite Monkton Quartzite Dunham Dolomite Cheshire Quartzite	Rugg Brook Formation Parker Slate Dunham Dolomite ← Champlain Thrust	Dunham Dolomite Cheshire Quartzite
CAMBRIAN?				Mendon Formation	Not Exposed	Underhill Formation Pinnacle Formation
PRECAMBRIAN			Metamorphic Rocks of the Adirondack Dome	Mount Holly Complex		Mount Holly Complex

of western Vermont.

It is still not clear, particularly in the eastern part of western Vermont, to what extent the synclinoral folding and foreland thrusting continued into the Silurian or Devonian, possibly blending with some of the events in the Acadian orogeny. On the western edge of the allochthon, however, an upper age can be assigned to the Taconic orogeny. At Becraft Mountain in the Hudson Valley relatively undeformed Lower Devonian rocks (Manlius Limestone) rest with angular discordance on rocks of both the Taconic allochthon and shales of the autochthon (loc. 16 cited by Pavlides, et al., 1968, p. 69). To the east the termination of Taconic orogeny is marked by the unconformity separating the Silurian and Devonian rocks from the Cambrian and Ordovician in central New England (for series or stage assignments see Pavlides, et al., 1968). Here, however, deformation and metamorphism of the Acadian orogeny has overprinted and greatly obscured the structures of the Taconic orogeny. Herein lies the problem of dating the structural and metamorphic events in western Vermont. Specifically, how far west does Acadian deformation and metamorphism extend and, conversely, how far east does Taconian deformation and metamorphism persist? Present investigators are divided on the problem. Cady (1945, 1960) and Crosby (1963) believe that most of the foreland thrusts and folds are Acadian, whereas Clark (1934) and Richard (1965) believe similar features in Quebec are Taconian. Zen (1967, 1968) favors a pre-Acadian age but admits that Acadian movements could have modified the structures. The fact that the metamorphic isograds are not deformed in western Vermont as they are in central New England (Thompson, et al., 1968) suggests that the structures are older than Acadian metamorphism. Isotopic age dates from the Taconics (Harper, 1967, 460-445 my) and the Sutton Mountain anticlinorium in Quebec (Richard, 1965, 440-420 my) also favor a Taconian age for the deformation in western Vermont.

The degree to which post-Taconian or post-Acadian events (normal faulting and regional tilting) have further complicated the picture is not known.

GEOLOGY OF THE HINESBURG SYNCLINORIUM

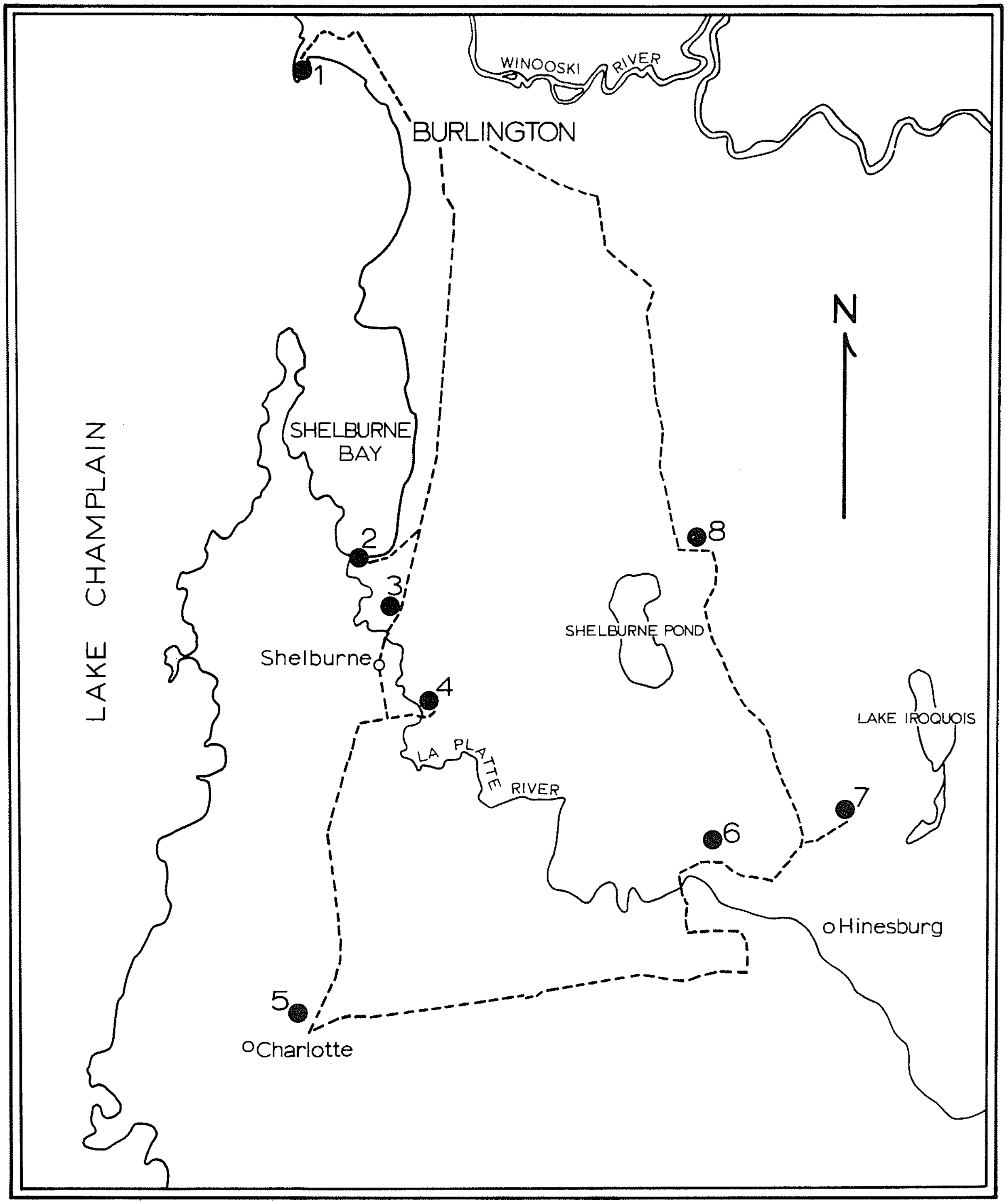
The Hinesburg synclinorium is separated from the Middlebury and St. Albans synclinoria by the Monkton cross anticline (Cady, 1945, p. 562) and the Milton culmination respectively. As seen on the geologic map of Vermont the smaller anticlines and synclines of the southern part are not mirrored by the single synclinal hinge of the northern part. This may be due to poor exposure to the north but it is probably a result of the configuration of the synclinorium and the fact that the eastern part is covered by the upper plate of the Hinesburg thrust. Minor faults, such as the Muddy Brook to the north and the Shelburne Pond to the south, partially disrupt the fold continuity of the synclinorium. These relations are well shown on the Geologic Map of Vermont (Doll, et al., 1961) and generalized in figure 1.

A composite stratigraphic section and correlation chart for the area is shown in Table 1 and Table 2 respectively. The rocks east of the Hinesburg thrust form the base of the composite section and mainly consist of metagraywacke and phyllite of the Pinnacle and Underhill formations. Isolated slivers of the Cheshire Quartzite (more argillaceous than to the west) appear just east of the Hinesburg thrust. Between the Winooski River and the Canadian border, outcrops of the Cheshire Quartzite form part of the upper plate of the thrust. The rocks in the synclinorium itself range in age from Lower Cambrian to Lower Ordovician (Dunham Dolomite through the Bascom Formation) and are characterized chiefly by carbonates which are predominantly dolomitic in the Cambrian and calcitic in the Lower Ordovician.

Quartzite beds occur in the Monkton Quartzite and the Danby Formation, and shale or phyllite is present as thin beds in the Monkton, Shelburne, and Bascom formations. The rocks west of the Champlain thrust are predominantly shales of the Stony Point and Iberville Formations and represent the argillaceous wedge that advanced westward on the carbonate shelf of western Vermont during the Middle Ordovician.

Current Work

In order to clarify the chronologic problems in the tectonic history of western Vermont, a long term project of detailed mapping and fabric analyses has been undertaken in the Hinesburg synclinorium and areas to the west, south and east of it. During the last four years students in Field Geology at the University of Vermont have been remapping parts of the synclinorium near Charlotte and Hinesburg. Although these studies are just a beginning in the long-term effort of refining the tectonic history of the area, it is already apparent that the structural relations in such areas as along the Champlain thrust and in the central part of the Hinesburg synclinorium are different from those shown on existing maps. For example, at least two generations of minor folds have been delineated in the southern part of the synclinorium and beneath the Champlain thrust. Although the styles of these folds differ in the shale and marble, the youngest generation of folds beneath the Champlain thrust shows the same northwesterly slip direction as does the older generation of folds in the Hinesburg synclinorium. The axial surfaces of these folds are in turn folded about northward-plunging axes of the younger generation. This suggests that the youngest movements on the Champlain thrust were followed by subsequent folding in the eastern part of the Hinesburg synclinorium. Furthermore, analysis of fractures, such as described at stop 2, indicates a sequence of events that can be related to the Champlain thrust. The inferred orientation of the principal planes of stress, however, does not



STOP MAP

----- Route

● Stops

1 0 1 2 3 4 Miles

SCALE

Figure 2

coincide in symmetry with other localities on the Champlain thrust. Elaborating and clarifying these relationships, as well as answering questions to other related problems, will hopefully culminate in a sequence of structural events for western Vermont that can be related to the orogenic history of the region.

STOP DESCRIPTIONS

General

The trip will consist of eight stops which are located on figure 2 and plate 1. The geologic relations along the route of the trip can be followed on figure 1 and plate 1.

Stop 1. Champlain thrust at Long Rock Point, Burlington - (Note: The Episcopal Diocesan Center has been kind enough to allow us to visit this locality. Please do not litter.) This locality is perhaps one of the finest exposures of the Champlain thrust in Vermont and Canada. Here the Dunham Dolomite (Conners facies) of Lower Cambrian age overlies the Iberville Formation of Middle Ordovician age. The thrust contact is sharp and marked by a thin discontinuous zone of breccia in which angular clasts of dolostone are embedded in a highly contorted matrix of shale. Slivers, several feet thick, of limestone are found along the fault and may represent pieces of the Beekmantown Group (Beldens Member of the Chipman Formation?). The undersurface of the Dunham Dolomite along the thrust is grooved by fault mullions which plunge 15^{o} to the southeast (Fig. 3 diagram 1 and 2a). The average southeasward dip of the thrust is 10^{o}.

A variety of minor structures are found in the Iberville Formation whereas joints are the only structures in the Dunham Dolomite. Faults and joints are oriented in a number of attitudes in the shale, but they have not been analyzed as yet. Many of these fractures are filled with calcite and grooved with slickensides. The minor folds in the shale are numerous, and are easily grouped into two age generations. The early folds have a well developed slaty cleavage which offsets the bedding and attests to differential flow parallel to the axial surface. Although only a few fold hinges are present, the slaty cleavage forms the dominant layering and is concordant to the undersurface of the Dunham Dolomite.

The younger generation consists of asymmetrical drag folds with short gently curved hinges and rather open profiles. The axial surface is rarely marked by cleavage but when it is developed, fracture cleavage, which is commonly filled with calcite, is typical. These folds deform the slaty cleavage of the older generation and hence are younger in age. The orientation of 59 axes with their sense of rotation is shown in diagrams 1, 2a and 2b of figure 3.

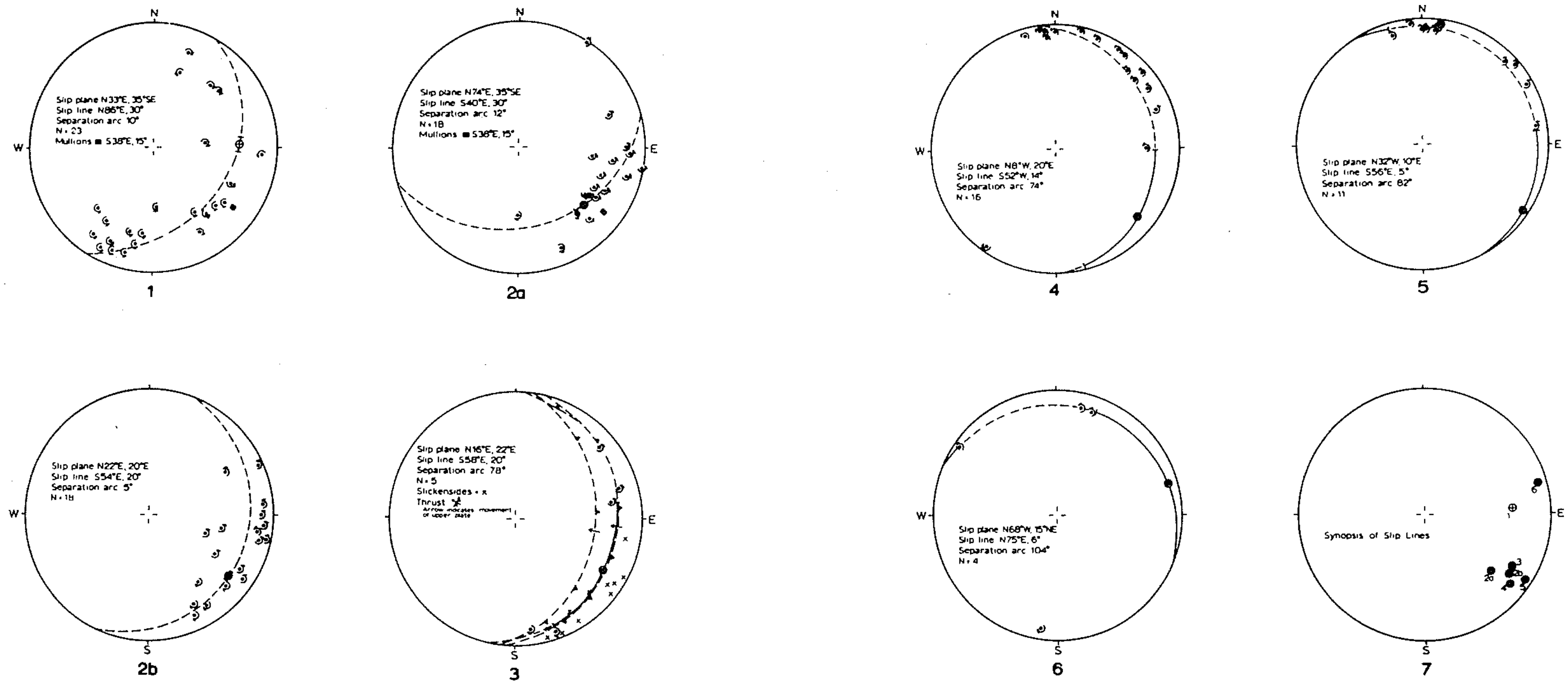


Figure 3 Slip line orientations determined from drag folds. Data shown on lower hemisphere equal area projection. Solid dots represent fold axes. Semicircular arrows show sense of rotation of asymmetrical drag folds. Dashed great circles are slip surfaces each of which contains a slip direction shown by a circle with solid dot (movement up of upper beds) or a circle with cross (movement down of upper beds). Numbers correspond to location shown on figure 1.

Slip line orientations The drag folds of the younger generation can be used to determine a slip surface and direction according to the method described by Hansen (1967, p. 390-397). The analysis assumes that these folds are in the truest sense drag folds and have formed by movement of one layer or zone over another. In this locality the drag folds in the shale presumably were formed by Dunham Dolomite moving over the Iberville Formation. Because the shale is highly anisotropic, consisting of thin layers of compact shale separated by thinner layers of extremely fissile shale, the folds are confined to zones several feet in thickness and are disharmonic in profile.

A slip plane and line were determined for three separate localities at Rock Point (Fig. 3, diagrams 1, 2a, 2b). For each place the 18 to 23 fold axes with their respective senses of rotation were plotted on a lower hemisphere equal area net. The great circle that best approximates the spatial distribution of axes defines the slip plane. In all the diagrams of figure 3, clockwise folds cluster along one part of the slip plane whereas counterclockwise folds cluster along the other. The bisector of the arc separating the two groups of folds is the slip line. The location of clockwise and counterclockwise arrows on either side of the separation arc determines whether the upper layers moved up or down along the slip line. In diagrams 2a and 2b (Fig. 3) the upper layers moved to the northwest along a line striking N 40 W for 2a, and N 54 W for 2b. In contrast the upper layers moved downward along a line striking N 86 E for the southern part of the Champlain thrust at Long Rock Point. (diagram 1, Fig. 3).

Discussion of Results The kinematic basis for this analysis has been worked out in such geologic environments as tundra and sod slides, glacial lake clays, lava flows and metamorphic rocks of all grades (Hansen, 1967; Hansen, et al., 1967; Howard, 1968). It can be shown that the separation arc contains the movement direction of slip line and that the drag folds which locate this direction are a product of one movement event. One can further deduce the likely position of the principal axes of stress (σ_1, σ_2, σ_3)[1] by analogy to slip systems. σ_2 would lie in the slip plane perpendicular to the slip direction. σ_1 and σ_3 would define the deformation plane which is perpendicular to the slip plane and parallels the slip direction.

$\sigma_1$would be oriented approximately 45° from the slip plane in a direction permitted by the sense of shear indicated by the drag folds. The slip line would then parallel the direction of maximum resolved shear stress.

[1]In all subsequent discussions σ_1 is the maximum compressive stress, σ_2 is the intermediate compressive stress and σ_3 is the minimum compressive stress.

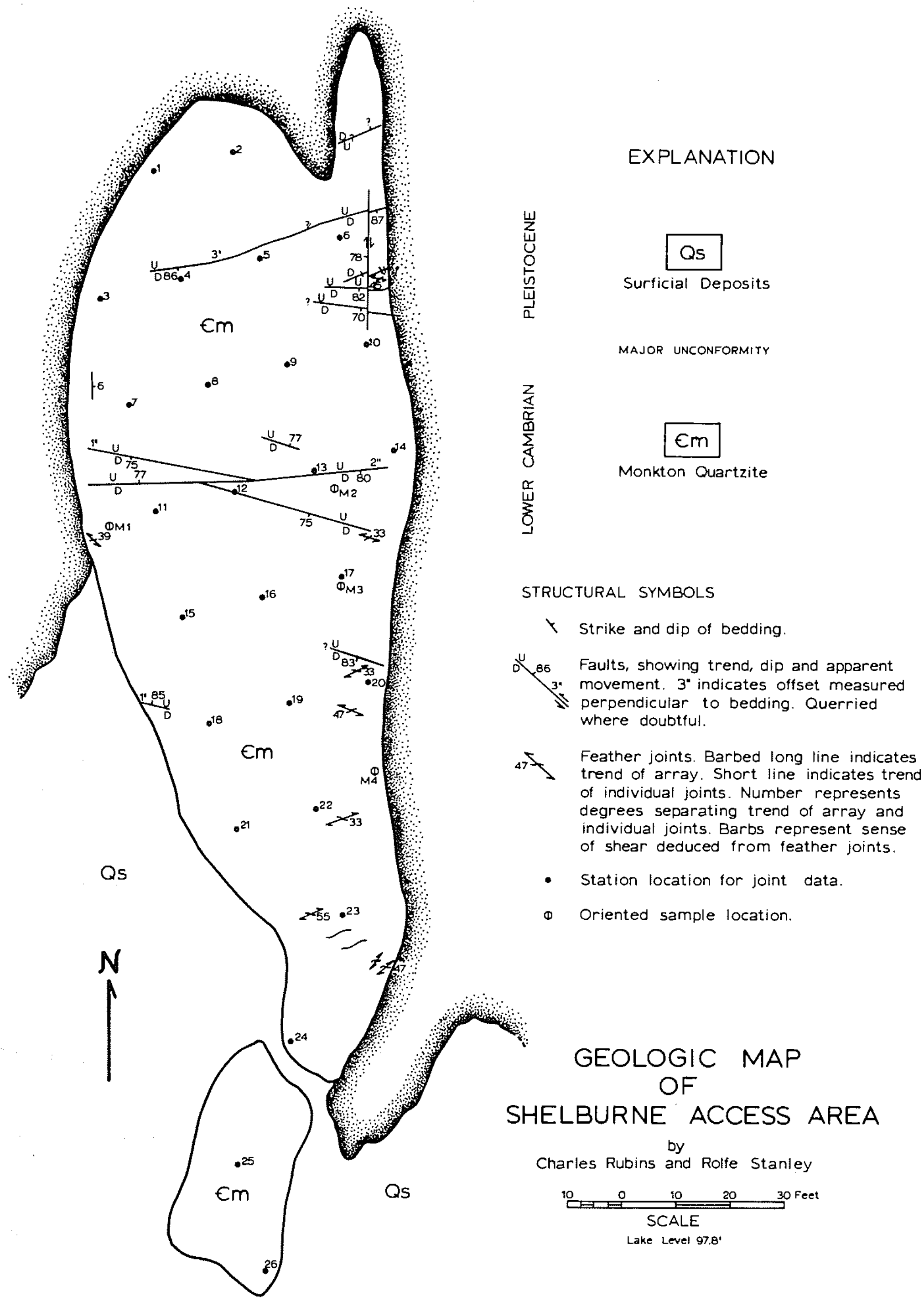


Figure 4 Structures in the Monkton Quartzite at the southern end of Shelburne Bay, Vermont.

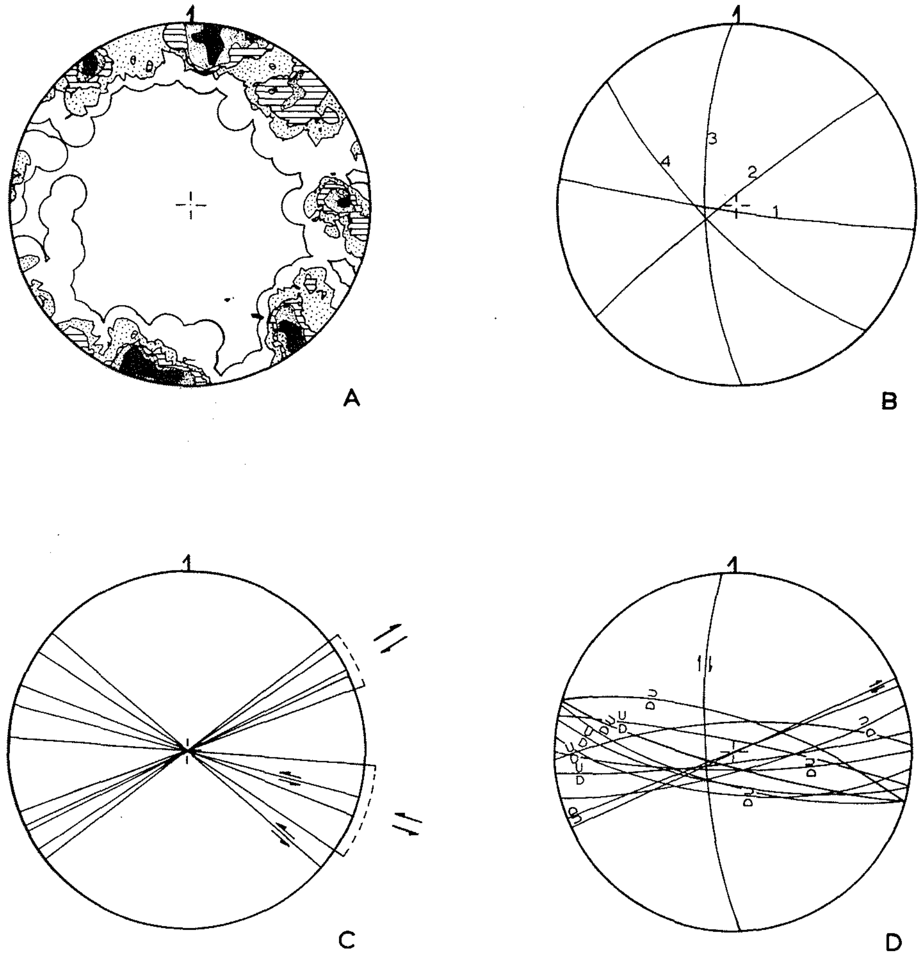


Figure 5 Lower hemisphere equal area projections showing fractures present in the Monkton Quartzite at the Shelburne Access area (fig. 4), Diagram A shows 248 poles to joints. Contour intervals are 0.4, 1.2, 2.0, 2.8, 3.6 respectively per 1 percent area. Diagram B shows planes corresponding to the 3.6 percent maxima of diagram A. Diagram C shows the trend and sense of shear of feather joints. Diagram D shows the orientation and apparent movement of 13 faults. Modified from an unpublished research report of Charles Rubins.

Diagrams 1 through 3 (Fig. 3) were determined at localities in the Iberville Formation beneath or just west of the Champlain thrust. With the exception of diagram 1, the slip directions indicate movement to the northwest (see diagram 7, Fig. 3 for synopsis). This direction is approximately parallel to fault mullions at Long Rock Point and slickensides on calcite-veneered surfaces at Shelburne Point (diagram 3) and, therefore, suggests that the deformation plane containing σ_1 and σ_3 for this phase of movement did trend northwesterly. A unique explanation for the easterly slip direction in diagram 1 is not known at present, but several possibilities will be discussed.

Stop 2. Shelburne Access Area This locality is one of several places in the Monkton Quartzite that displays a sufficient number of joints, faults and feather joints for dynamic analysis. The location and orientation of faults and feather joints are shown on the geologic map (Fig. 4). At each numbered station the orientation, relative abundance, and surface features of approximately 10 joints were recorded. Diagram A of figure 5 shows the poles to 248 joints and diagram B shows four planes corresponding to the maxima in diagram A. The trend of each of the 10 feather joint arrays and their sense of shear are shown in diagram C.

Diagram D shows 13 faults with letters or arrows indicating the apparent movement. All faults trend approximately eastward (hereafter called cross faults) except for one that is striking to the north. As indicated on the map the dip slip displacement of each fault is only several inches whereas the strike slip displacement on one of these faults is 9 feet (fault 13,[2] Fig. 4). Feather joint arrays adjacent to several cross faults further attest to strike slip displacement. This relationship is further emphasized by comparing diagram C and D (Fig. 5). Although Welby (1961, p. 204) assumes that all cross faults are normal faults, the above data definitely supports a wrench fault interpretation. Petrofabric analysis of quartz deformation lamellae in three samples (M1, M3, M4) further confirms this conclusion (Fig. 6).

Dynamic Analysis Discussion of the stress configurations for each structure is based upon techniques and principles summarized by Friedman (1964). In stress analysis only the direction and relative magnitudes of the principal stresses can be evaluated. Furthermore, it is assumed that the manner in which these structures formed in nature is similar to the way analogous structures are formed in the laboratory.

Joints: The planes in diagram B (Fig. 5) corresponding to the maxima of diagram A can be interpreted in several ways. The acute angle between joints 1 and 3 is 80 degrees and the acute angle between joints 2 and 4 is 83 degrees.

[2]Faults are identified by the station number nearest them on the geologic map (Fig. 4).

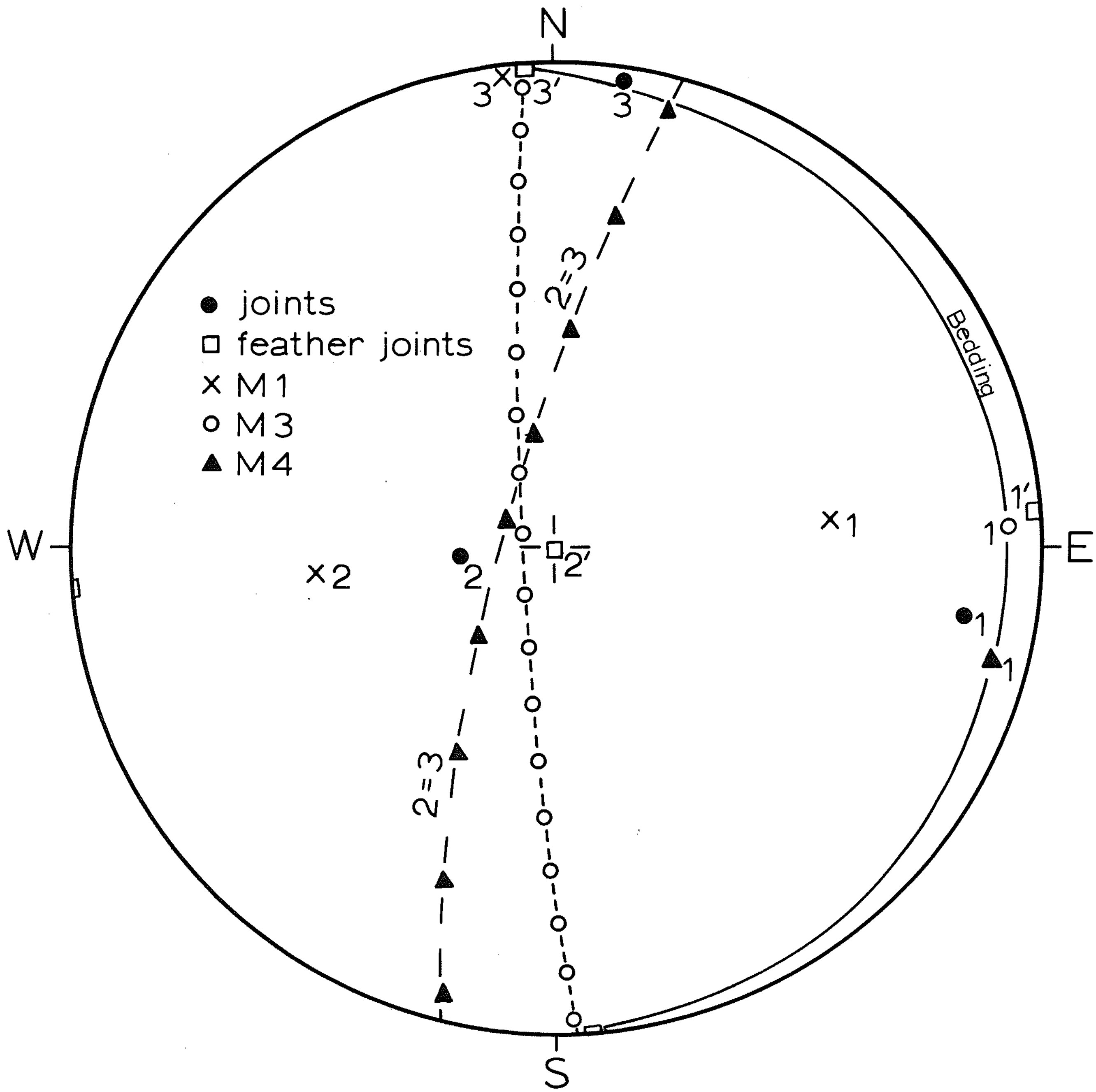


Fig. 6: Synoptic diagram of principal stress directions ($\sigma_1,\sigma_2,\sigma_3$) deduced from joints (n = 248), feather joints (n=11), and deformation lamellae in quartzite samples M1 (n=204), M3 (n=119), M4 (n=62). Numbers 1, 2, 3 refer to σ_1 (maximum compressive stress), σ_2,σ_3 (minimum compressive stress) respectively.

Hypothesis 1: Joints 2 and 4 formed in the shear position. Joints 1 and 3 formed as extension and release joints respectively.

Hypothesis 2: Joints 1 and 3 formed in the shear position. Joints 4 and 2 formed as extension and release joints respectively. In both hypotheses σ_2 is defined by the intersection of all joints. In hypothesis 1, σ_1 would parallel joint 1 and trend to the east whereas in hypothesis 2 σ_1 would parallel joint 4 and trend to the southeast. Hypothesis 1 is preferred because joint 1 is commonly filled with calcite. Furthermore, this configuration is similar to the stress configurations deduced from the other structures (Fig. 6).

Feather joints: The 10 feather joint arrays in diagram C (Fig. 5) with their respective senses of shear indicate that σ_1 is oriented east-west. σ_3 is oriented north-south and σ_2 is oriented vertically. Because the dip of the feather joint array could not be measured, the inclination of the principal stress axes could not be pinpointed. Although the principal stress directions are not as accurately known for feather joints as they are for other structures, their symmetrical relationship to the cross faults support a wrench fault interpretation and thus suggest that those faults trending north of west are left lateral whereas those trending north of east are right lateral.

Wrench Faults: The apparent vertical movement on all faults trending north of west is up to the north. This movement can be realized on left lateral wrench faults if σ_1 dips more steeply to the east than the bedding. Those faults trending north of east that are downthrown to the north would then be right lateral wrench faults.

The apparent vertical movement of faults 4 and 13 trending north of east is up to the north. These faults are right lateral wrench faults if the movement on 4 is assumed to be the same as the displacement of fault 12 by 13. The horizontal and vertical components of the actual movement at these faults could be caused by σ_1 trending westward and dipping eastward more gently than the bedding.

North-South Fault: This fault cuts several of the wrench faults and hence is younger in age. It could result from the following two stress configurations:

Hypothesis 1: If the fault is a normal fault downthrown to the west, then σ_1 would be vertical or steeply inclined to the north so as to produce the right lateral movement. σ_3 would trend eastward or dip gently westward. σ_2 would trend northward.

Hypothesis 2: If the fault is a wrench fault, then σ_1 would trend northeasterly. σ_3 would trend northwesterly and σ_2 would be approximately vertical.

Quartz Deformation Lamellae: Approximately 380 deformation lamellae from 3 oriented samples have been measured from this outcrop. For each sample 100 (50 for M4) quartz grains were measured from each of three mutual perpendicular thin sections. For each grain the orientations of the c axis and deformation lamellae (if present) were measured. The results were analyzed using methods described by Carter and Friedman (1965), and Scott, et al., (1965). The deduced orientation for σ_1, σ_2, and σ_3 are shown in figure 6. In M3 and M4, σ_1 lies in the bedding and σ_2 appears to be equal to σ_3 in magnitude. In M1 σ_1 is inclined 40^o to the east, σ_2 dips 50^o to the west and σ_3 trends northward and is horizontal. Although these orientations are not parallel in all samples, they are consistent with the stress positions deduced from the megascopic structures and support the conclusions on the cross faults and joints.

Structural and Stress History Based on the above information it is possible to develop a structural history for this outcrop. This sequence is divided into the following three phases.

Phase 1: During this time all the wrench faults formed except fault 13, 4 and 10 (north-south fault), σ_1 was inclined more steeply to the east than the bedding so that faults trending north of west are upthrown to the north and faults trending north of east are downthrown to the north. σ_3 was oriented northward and σ_2 was inclined steeply to the west.

Phase 2: During this time faults 13 and 4 developed with σ_1 either horizontal or inclined more gently eastward than the bedding. This orientation permitted the right lateral faults (13 and 4) to be upthrown to the north. It should be emphasized at this point that the change in orientation of σ_1 relative to the bedding in the Monkton Quartzite can equally be attributed to a rotation of the bedding within a stress system of constant principal stress orientation.

Phase 3: During this time the north-trending fault developed and can be interpreted as a normal, wrench, or reverse fault.

The feather joints and deformation lamellae are thought to have formed during phase 1 and 2, although definite evidence for their timing is lacking at present. The formation of the joints shown in diagram A and B in figure 5 could well have spanned the first two phases.

Relationship to the Champlain Thrust Wrench faults are commonly associated with thrust faults. Both can be related to the same σ_1 direction and only require a switch of σ_2 and σ_3 in the stress configuration during thrusting to develop wrench or tear faults. In the same manner the small wrench faults in the Monkton Quartzite are thought to bear the same relationship to the Champlain thrust and as such suggests that other cross faults shown on the Geologic Map of Vermont (Doll, et al., 1961), may also be wrench faults.

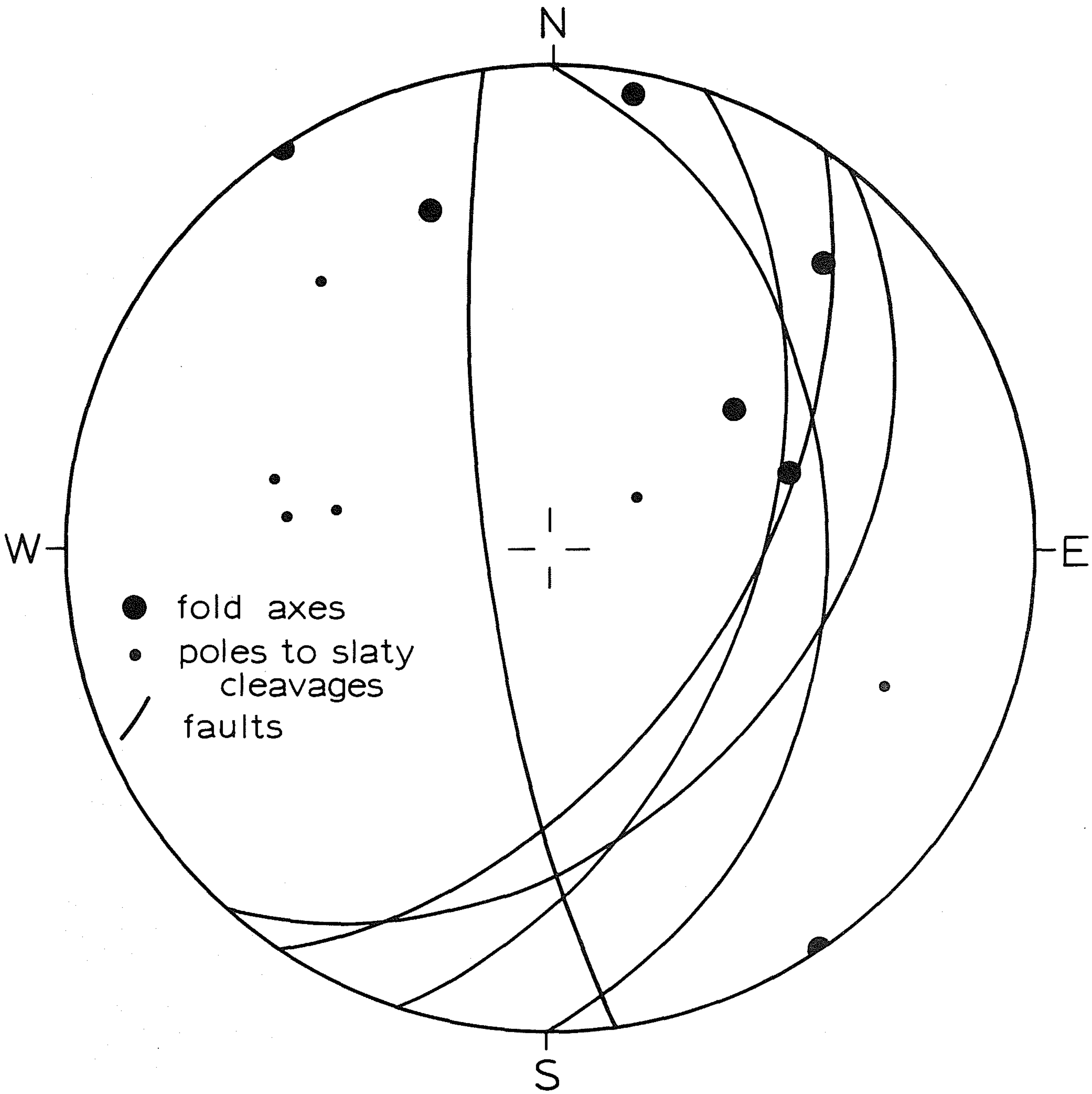


Figure 7 Lower hemisphere equal area projection showing the orientation of fold axes, slaty cleavage and faults in the Iberville Formation on the west side of Jones Hill.

The different stress orientations deduced at Long Rock Point (stop 1) and the Shelburne access area suggest several explanations. It is possible that they formed under the same stress system but have assumed their present orientation by subsequent deformation. Alternatively, they could have formed at different times under stress systems of different orientations. Future work will resolve this problem.

Stop 3. This locality is a fine exposure of the Winooski Dolomite in fault contact with the upper portion of the Monkton Quartzite. The bedding in both these units dips gently eastward. The most conspicuous of several northeastward-trending faults is downthrown to the north placing the Winooski Dolomite in fault contact with the upper portion of the Monkton Quartzite south of the fault.

Stop 4. Shelburne Falls just southeast of the village of Shelburne The Danby Formation dips gently eastward at this locality and is cut by several sets of joints. Ripple marks of varying sizes, cross beds and fossil burrows are well displayed in the quartzite and sandy dolomite on the west bank of the river. Current directions based on current ripples in the upper layers along the southern part of the exposure indicate flow to the southeast (S 10 E to S 40 E) whereas current directions based on cross bedding in the lower layers exposed near the falls to the north indicate flow to the north and west.

Stop 5. Jones Hill This stop is located just west of the Champlain thrust and shows several large folds and minor thrusts exposed in the Iberville Formation. A lamprophyre dike, striking N 78 W and dipping 75 degrees to the northeast, is exposed in the southern part of the outcrop. The laccolith of syenite on Barber Hill is located approximately a mile southeast of here near the village of Charlotte.

The orientation of fold axes, slaty cleavage and faults is shown on figure 7; most of the faults are thrusts dipping eastward or southeastward at gentle to moderate angles. As shown on plate 1 the trace of the Champlain thrust is sinuous and offset at several places in the area. Welby (1961, Plate 1a) shows normal faults at each of these places but the orientation of bedding, particularly between Jones Hill and Pease Mountain, suggests that the eastward reentrants can be explained equally well by anticlines plunging gently to the east. Minor folds with this orientation are present on the north side of Jones Hill.

Stop 6. This stop will consist of a short traverse through the fields north of the road. The Clarendon Springs Dolomite, the Shelburne Formation, and the Cutting Dolomite will be crossed. If time permits the traverse will be extended northward into the Bascom Formation. At least two generations of folds are present in this area. Axes of the earlier generation plunge to the southeast, east and northeast and possess a well developed axial surface cleavage which commonly dips eastward except when it is deformed by folds of the younger generation. The slip line orientations of diagrams 4 and 5 (Fig. 3) were determined from drag folds of the earlier fold generation at localities 4 and 5 on figure 1. These slip lines are almost parallel to those to the west of the Champlain thrust discussed under stop 1 and, therefore, strongly suggest that they may have formed contemporaneously as a result of the same stress system. The younger generation of folds plunges northward and folds the cleavage of the earlier fold generation.

Stop 7. Hinesburg thrust northwest of Mechanicsville This locality is one of the finest exposures of the Hinesburg thrust. The Bascom Formation forms the lower plate and the argillaceous facies of the Cheshire Quartzite forms the upper plate. Minor folds can be found in both these units. Inasmuch as only 4 folds could be found in the Bascom Formation little significance can be attached to southwestward movement direction shown in diagram 6 (Fig. 3).

Stop 8. The Bascom Formation crops out on several small hills just west of Brownell Mountain. To the east the Bascom grades into the Brownell Mountain Phyllite, a member of the Bascom Formation only recognized in the Hinesburg synclinorium. This stop may be cancelled if insufficient time is available.

End of Trip. Return to Burlington.

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