FORELAND DEFORMATION AS SEEN IN WESTERN VERMONT

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INTRODUCTION

Western Vermont is underlain by three distinctive sequences of rocks that range in age from Late Proterozoic to Middle Ordovician and are typical of the western part of the Appalachian Mountains. The lowest most sequence, which rest with profound unconformity on the Middle Proterozoic of .the Green Mountain and Lincoln massifs, largely consists of metawackes, mafic volcanic rocks and phyllites that represent a rift clastic sequence. These rocks grade upward into siliciclastic and carbonate rocks of the platform sequence which in turn are overlain by Middle Ordovician shales of the foreland basin sequence. The boundary between the two sequences is the based of the Cheshire Formation (fig. 1). North of Burlington, Vermont the platform sequence grades into shales, breccias and conglomerates of the ancient platform margin and eastern basin. These sequences have been studied by a number of workers in the past (Cady, 1945; Cady and others, 1962; Hawley, 1957; Erwin, 1957; Welby, 1961; Stone and Dennis, 1964, for example) and are receiving current attention by Mehrtens (1985, 1987) and her students (Gregory, 1982; Myrow, 1983; Teetsel, 1985; Bulter, 1986; MacLean, 1986, 1987). Agnew (1977), Carter (1979), Tauvers (1982), DiPietro (1983) and Dorsey and others (1983) have reexamined parts of the rift clastic sequence while Doolan and his students are currently working in the same sequence in northern Vermont and Quebec (Doolan and others, 1987; Colpron and others, 1987). Figures 1 and 2 illustrate the representative stratigraphic columns for western Vermont north of the latitude of the Lincoln massif where the field conference will be held. Additional information can be found in Welby (1961) and Doll and others (1961).

The structure of western Vermont is dominated by major, north-trending folds and imbricate thrust faults which are well displayed on the Geologic Map of Vermont (Doll and others, 1961). The rift clastic and platform sequences have each been displaced westward on major thrust faults that extend through much of Vermont. The larger of the two, the Champlain thrust, extends from southern Quebec to Albany, New York and places the older platform sequence over the younger foreland basin sequence. Estimates of westward displacement range from 15 km to 100 km. The smaller of the two, the Hinesburg thrust, places transitional and rift clastic rocks over the platform sequence and as such forms a boundary between the synclinorial rocks of western Vermont and the

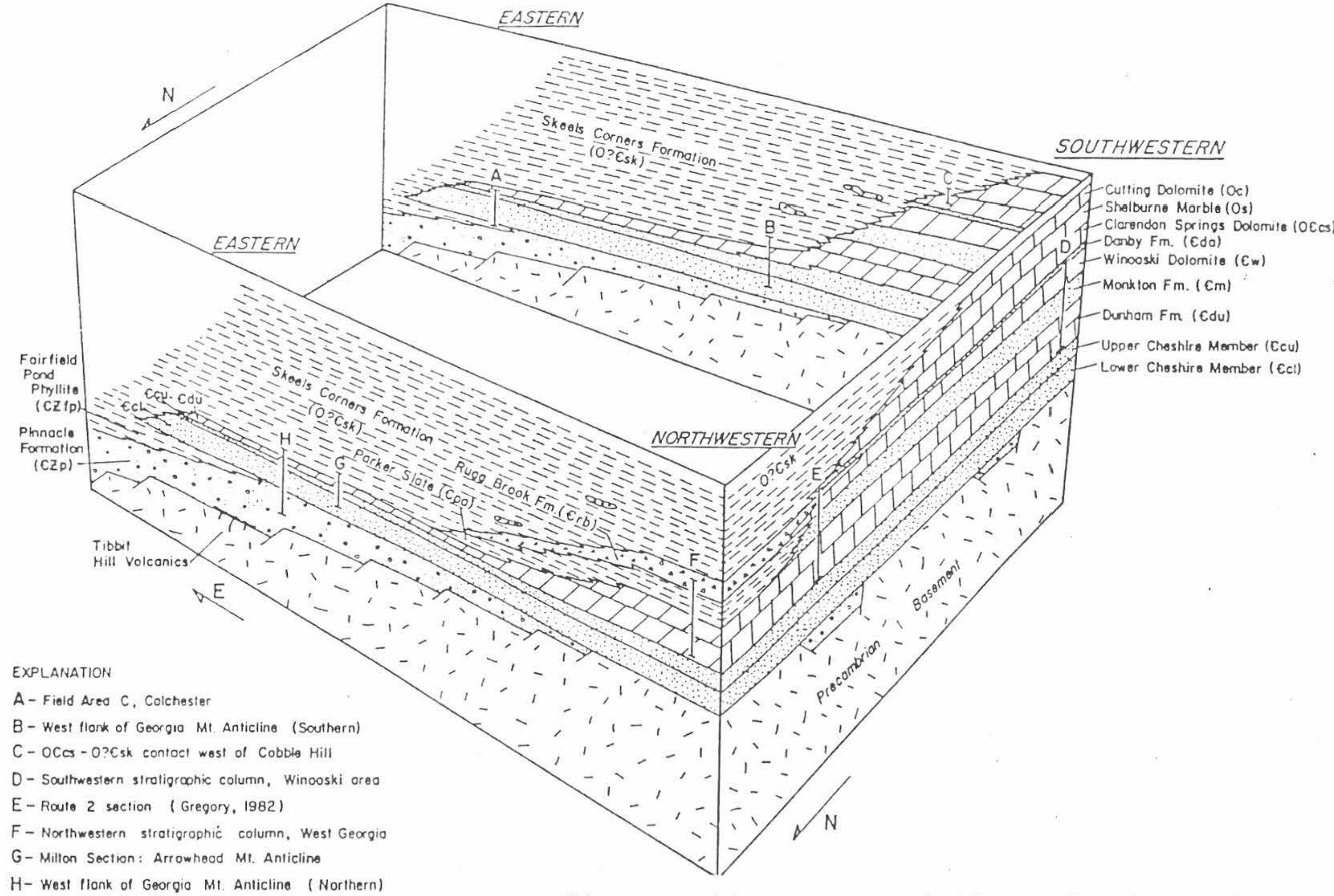


Figure 1 - Diagrammatic representation of the original depositional protolith for the Milton quadrangle reconstructed from stratigraphic information in the indicated areas.

Dorsey and others (1983)

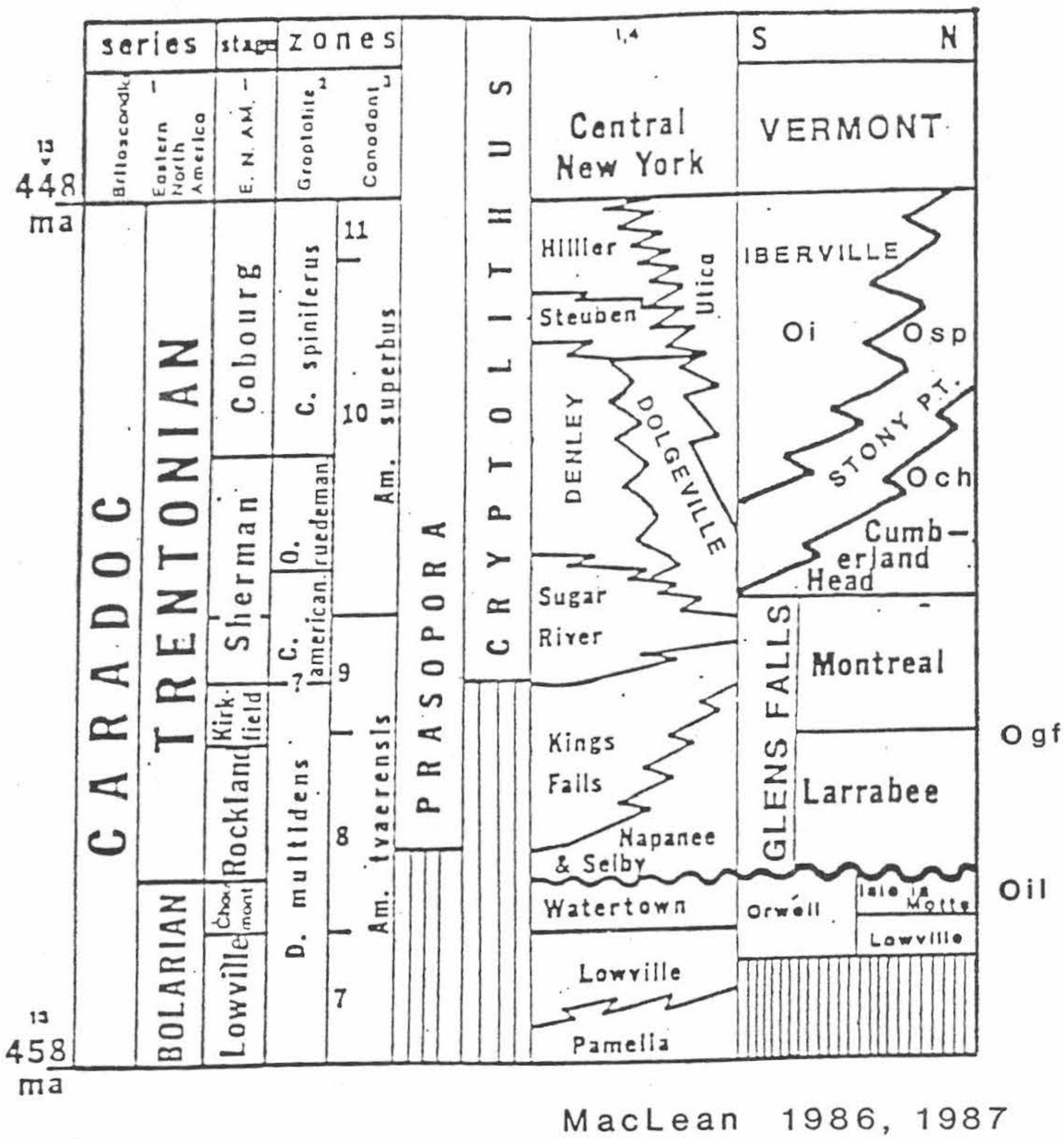


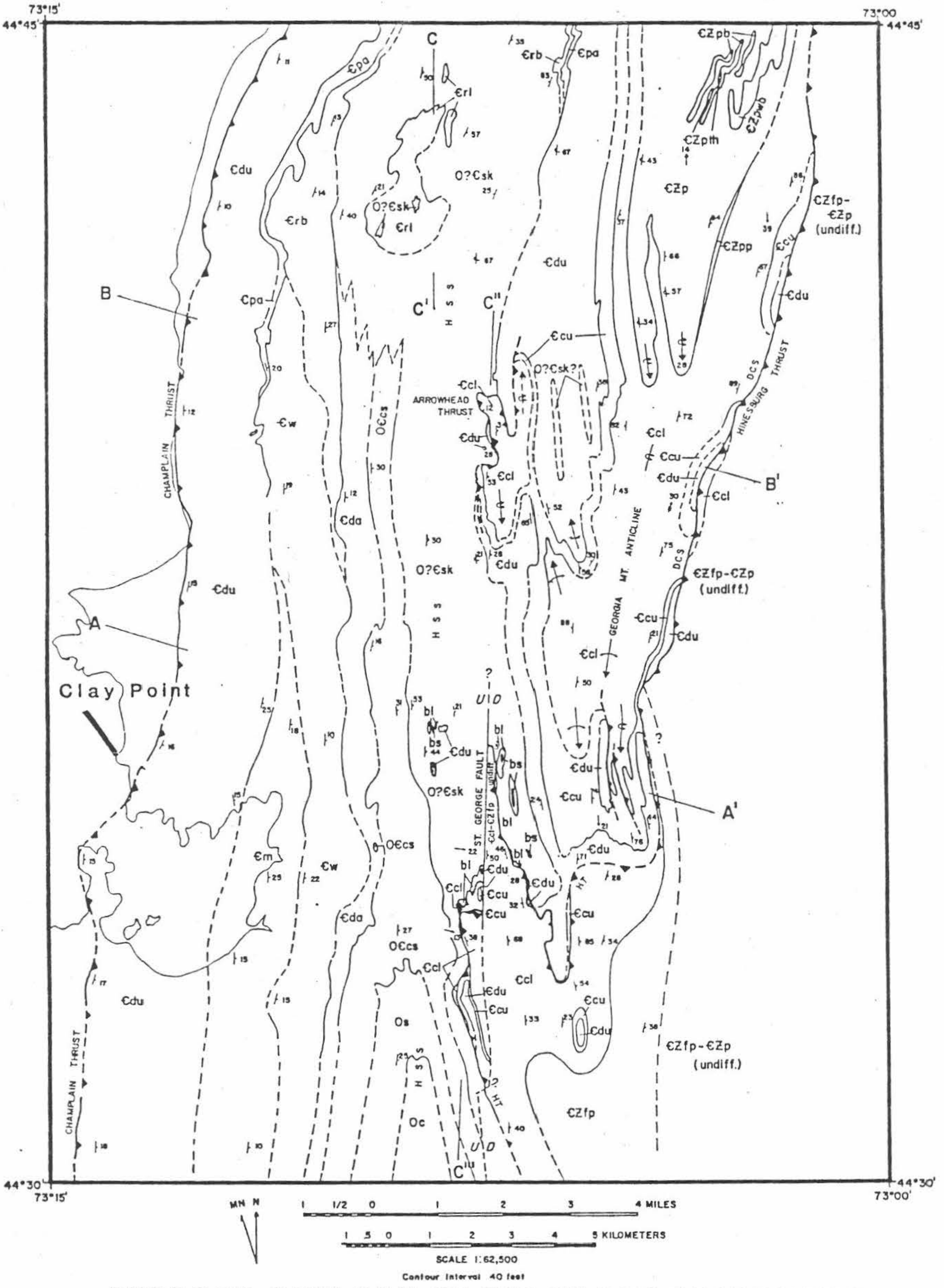
Figure 2 - Stratigraphic column for the western Vermont and Central New York. For further stratigraphic discussion read Erwin (1957) and Hawley (1957, 1972).

Green Mountain anticlinorium. Dorsey and others (1983) have demonstrated that the Hinesburg thrust developed along the overturned limb of a large recumbent fold (fault-propagation fold of Suppe, 1985) and therefore is quite different from the geometry of the Champlain thrust fault. Westward displacement is estimated to be 6 km.

Recent seismic traverses across western Vermont demonstrate that the Champlain thrust dips eastward at approximately 15 degrees beneath the Green Mountain anticlinorium and that the major folds of western Vermont are formed by duplexes and related structure on this and other thrust faults which are present both within the platform and the foreland basin sequences. High angle normal faults have been mapped along the Champlain thrust fault and in many parts of the foreland basin sequence (Welby, 1961; Doll and others, 1961). Seismic information shows that some of the faults are older than the Champlain thrust fault whereas others are younger. Stanley (1980) has shown that several of the faults that cut the eastern part of the platform sequence are Mesozoic in age.

The structure of western Vermont in the Burlington region is best illustrated by the recent work of Dorsey and others (1983) in the Milton quadrangle (fig. 3) and Leonard (1985) in South Hero Island. The cross sections for this region show that the Champlain thrust fault is essentially planar which is consistent with recent seismic studies at this latitiude. Detailed surface mapping, however, has shown that the overlying thrust faults are folded. Dorsey and others (1983) therefore suggest that these folds are related to duplexes in the Champlain slice (fig. 4). Based on this configuration the highest fault, the Hinesburg thrust, is the oldest and the Champlain thrust is the youngest. Thus the thrust sequence developed from the hinterland (east) to the foreland. Shortening as measured between the pin points in section B-B' (fig. 4) is in the order of 55 percent with 6 km of displacement on the Hinesburg thrust and 0.85 km of movement on the Arrowhead Mountain thrust. The Milton cross sections clearly show the change in structural style and fabric that occurs as one crosses from the foreland to the hinterland.

The foregoing conclusions are consistent with the character of each of the fault zones. The Champlain thrust fault is marked by gouge, welded breccias and pressure solution features (Stanley, 1987). An estimate of the confining presssure based on the stratigraphic thickness (2700 m) of the hangingwall block is 0.73 kbars which corresponds to approximately 2.5 km below the earth's surface. This assumes the fault is a Taconian feature. At this depth the temperature would be in the order of 100^{0} C assuming a geothermal gradient of 20^{0} to 30^{0} C/ km. (Strehle and Stanley, 1986, fig. 5). The gouge and cataclastic features are consistent with these estimates and suggest that if the

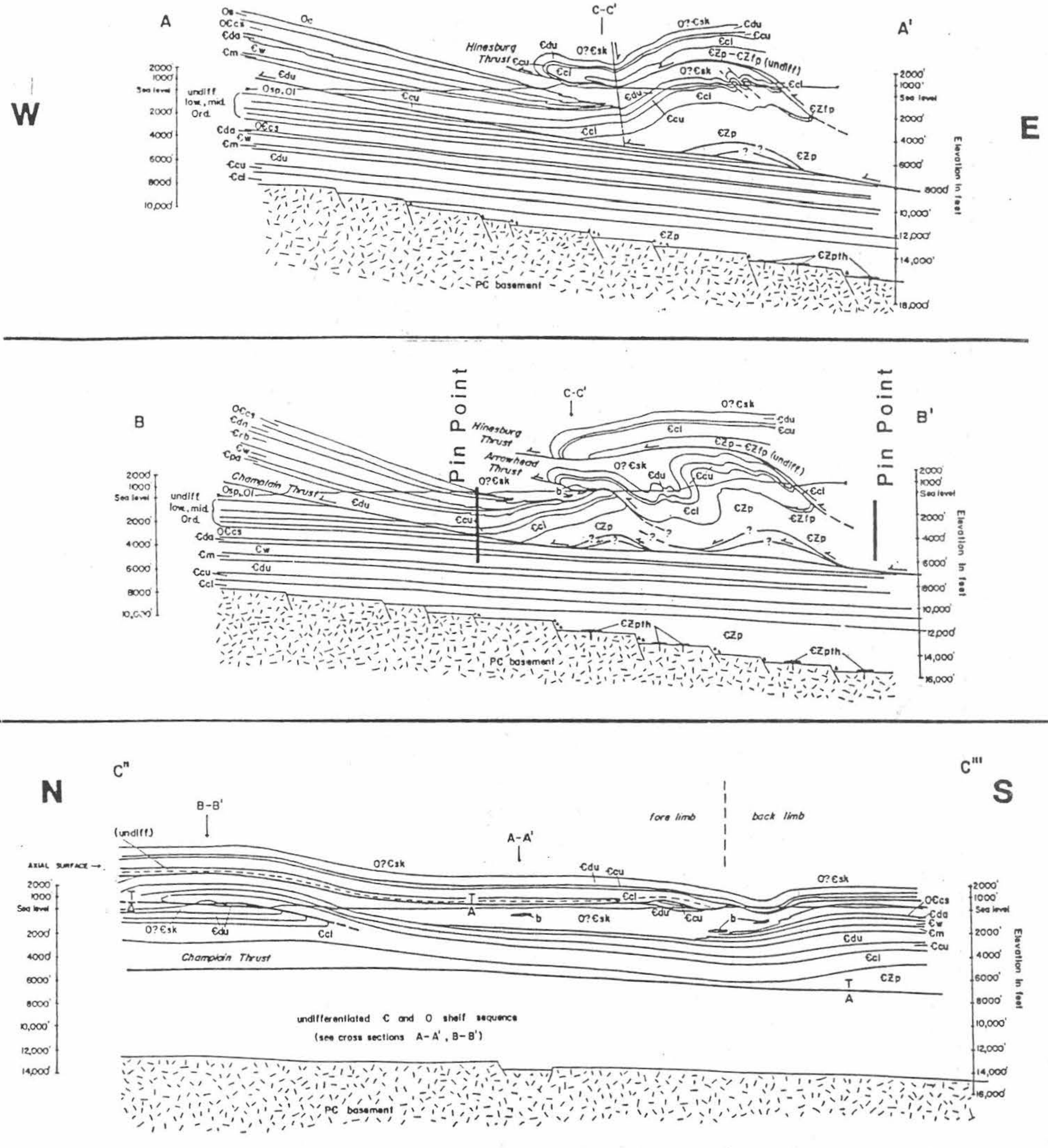


BEDROCK GEOLOGY OF THE MILTON QUADRANGLE

WESTERN VERMONT

Figure 3

Dorsey and others 1983



Dorsey and Stanley (1983)

CROSS SECTIONS OF THE MILTON QUADRANGLE WESTERN VERMONT

Figure 4

hangingwall block at the longitude of Lone Rock Point did involve a repeated section or tectonic load, the additional section was small. The Arrowhead Mountain thrust fault, on the other hand, is another story. The shaley dolostone slivers in the fault zone are welded cataclasites that contain a later, fault-related, pressure-solution cleavage. In the quartzite near the fault surface the quartz shows undulose extinction, deformation lamellae, and, limited recrystallization whereas the feldspar is fractured and twinned. Oriented sericite occurs along cleavage surfaces in the quartzite (Strehle and Stanley, 1986). Because the stratigraphy of the hangingwall block is only slightly thicker (2900 m.) than the Champlain slice the fault-zone fabric along the Arrowhead Mountain thrust fault should be similar to the Champlain thrust. The fabric at Arrowhead Mountain, however, has clearly formed at an higher temperature and pressure because it involves recrystallization of quartz and the growth of sericite. Strehle and Stanley (1986, fig. 5) suggest temperatures in the order 200^{0} C to 250^{0} C which is in agreement with oxygen and carbon isotope temperatures of $210\text{-}295^{0}$ C from calcite and dolomite assemblages in the weakly metamorphosed carbonate rocks directly west of the Lincoln massif 40 km to the south. These temperatures correspond to pressures of approximately 2.5 kbars or 7.5 - 8.0 km. These fabrics and the corresponding estimates indicate that the Arrowhead Mountain thrust must have carried a tectonic load of more than double the standard stratigraphic section. This load would correspond to the hangingwall block of the Hinesburg thrust (fig. 4).

The story for the Hinesburg thrust fault is again similar to that of the Arrowhead Mountain thrust. At Hinesburg the quartzite within 25 m. of the fault has been extensive recrystallized and mylonitic textures become progessively well developed as the fault contact is approached. An ultramylonite marks the fault surface. Feldspar grains are fractured and bent but show little signs of recystallization. Sericite, chlorite, and stilpnomelane are present along the fault - related cleavages. The deformation features in quartz and feldspar indicate that deformation occurred below 450^{0} C (Voll, 1976). Strehle and Stanley (1986; fig. 5) suggest temperatures in the order of 350^{0} C which would correspond to pressures in the order of 3.5 km or 11 to 12 km. Clearly the upper plate of the Hinesburg thrust must have carried older thrust faults that rooted in the pre-Silurian section to the east (Underhill slice of Stanley and Ratcliffe, 1985, fig 2A, pl. 1). Thus the fabrics along the Champlain, Arrowhead Mountain and Hinesburg thrust faults developed under progressive more ductile conditions as a result of higher temperatures and pressures generated by an increasing tectonic load.

Most of the deformation in western Vermont occurred during the Taconian Orogeny. This conclusion is based on regional considerations (Stanley and Ratcliffe, 1985) and the analysis of available isotopic data (Sutter and others, 1985). Mesozoic deformation in the form of extensional faults and igneous activity clearly affected the region (McHone and Bulter, 1984; Stanley, 1980). Available evidence, however, can not rule out limited Acadian or even Alleghenian deformation.

Participants of this field trip will see the change in structure and fabric as they cross the foreland from South Hero Island to the western margin of the hinterland at the Hinesburg thrust fault at Mechanicsville, Vermont.

ITINERARY

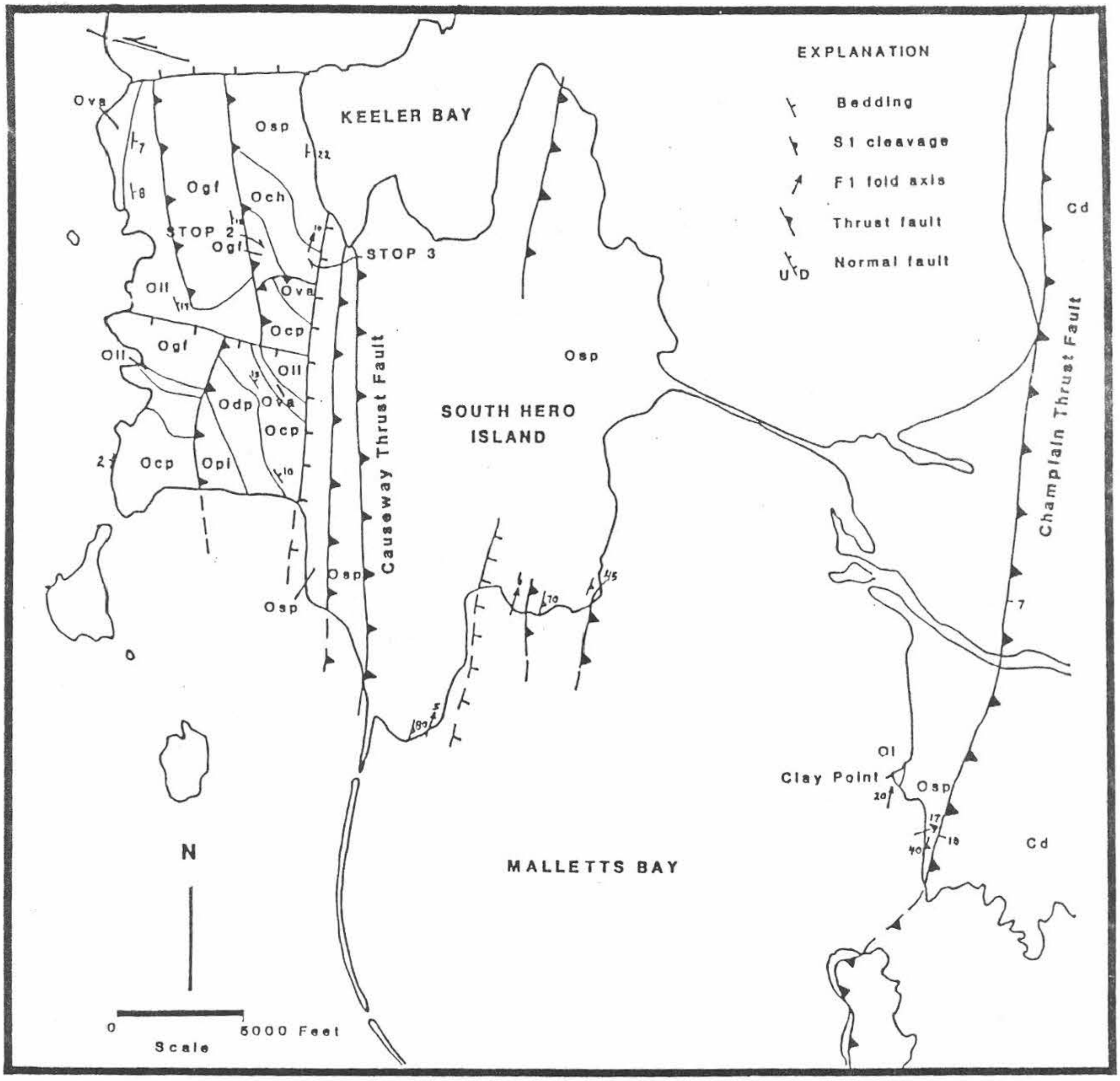
ALL STOPS ARE LOCATED ON GEOLOGIC OR TOPOGRAPHIC MAPS. A MILEAGE LOG IS NOT INCLUDED.

Assemble at the Apple Store on the south side of Route 2 in the village of South Hero at 8:30 AM. The first three stops will be on South Hero Island (fig. 5). PLEASE DO NOT USE HAMMERS ON THIS TRIP. LEAVE THEM IN THE CAR.

STOP 1 - West Shore of South Hero Island. This stop illustrates the low level of deformation that characterizes much of the west shore of the Champlain Islands and the eastern shore of New York in the area of Plattsburg. Local areas of moderate deformation, however, do exist where bedding plane faults and high-angle faults cut the bedrock. These outcrops of shale and thin micrite beds are in the Stony Point Shale. Note that the bedding, which is nearly horizontal, is cut by a poorly developed cleavage that is only developed in the shale beds. The cleavage dips very gently to the east and is interpreted to result from simple shear parallel to the bedding. Thin layers of calcite are present on some of the beds. Those layers that are marked by prominent slickenlines are bedding plane faults. As we will see at other outcrops today, the larger faults are marked by thicker layers of lineated calcite.

STOP 2 - Lessor's Quarry (fig. 6) - This quarry is located in the fossiliferous Glens Falls Limestone. The quarry contains some of the finest evidence of pressure solution in western Vermont. The cleavage (S1), which is discontinuous and wavy, is a classical pressure solution feature with well developed selvedges that truncate fossils and offset bedding. A small anticline at the south edge of the quarry contains adjustment faults at its hinge that end along cleavage zones with thick clay selvedges.

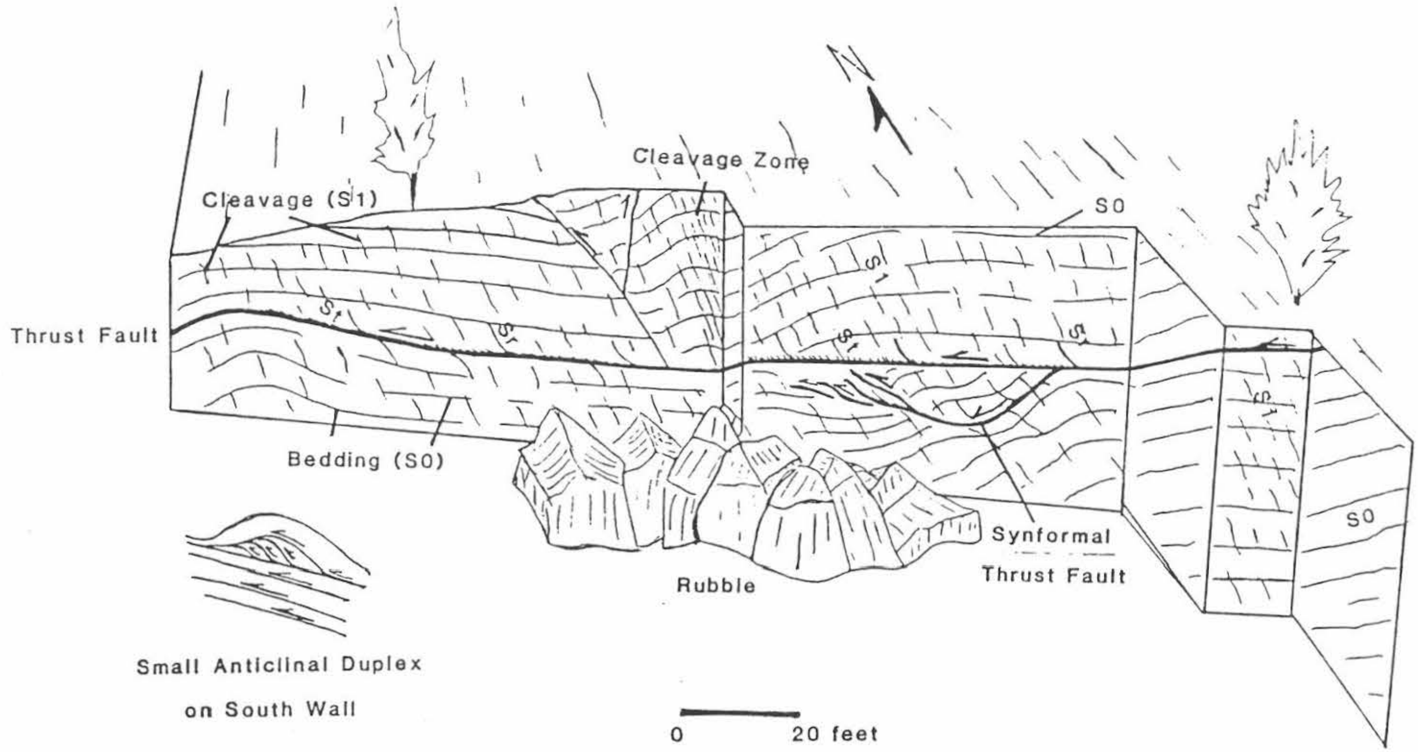
The major structures in the quarry are bedding-plane thrust faults. These faults are marked by calcite layers with west-trending slickenlines and a fault-zone cleavage (St).



Bedrock Geology of South Hero Island, Vermont

Leonard (1985), Erwin (1957)

Figure 5



Stanley (1987)

THRUST FAULTS AND RELATED STRUCTURES

AT

LESSOR'S QUARRY, SOUTH HERO ISLAND, VERMONT

Figure 6

Near the larger faults the S1 cleavage is rotated (Sr) toward the plane of the fault. Note that both St and Sr dip gently to the east and indicate that movement on the bedding faults was to the west. The St cleavage forms as a result of simple shear on the faults. The anticline along the southwall and edge of the quarry is formed from a small duplex. Unfortunately, the best evidence for this duplex has been excavated.

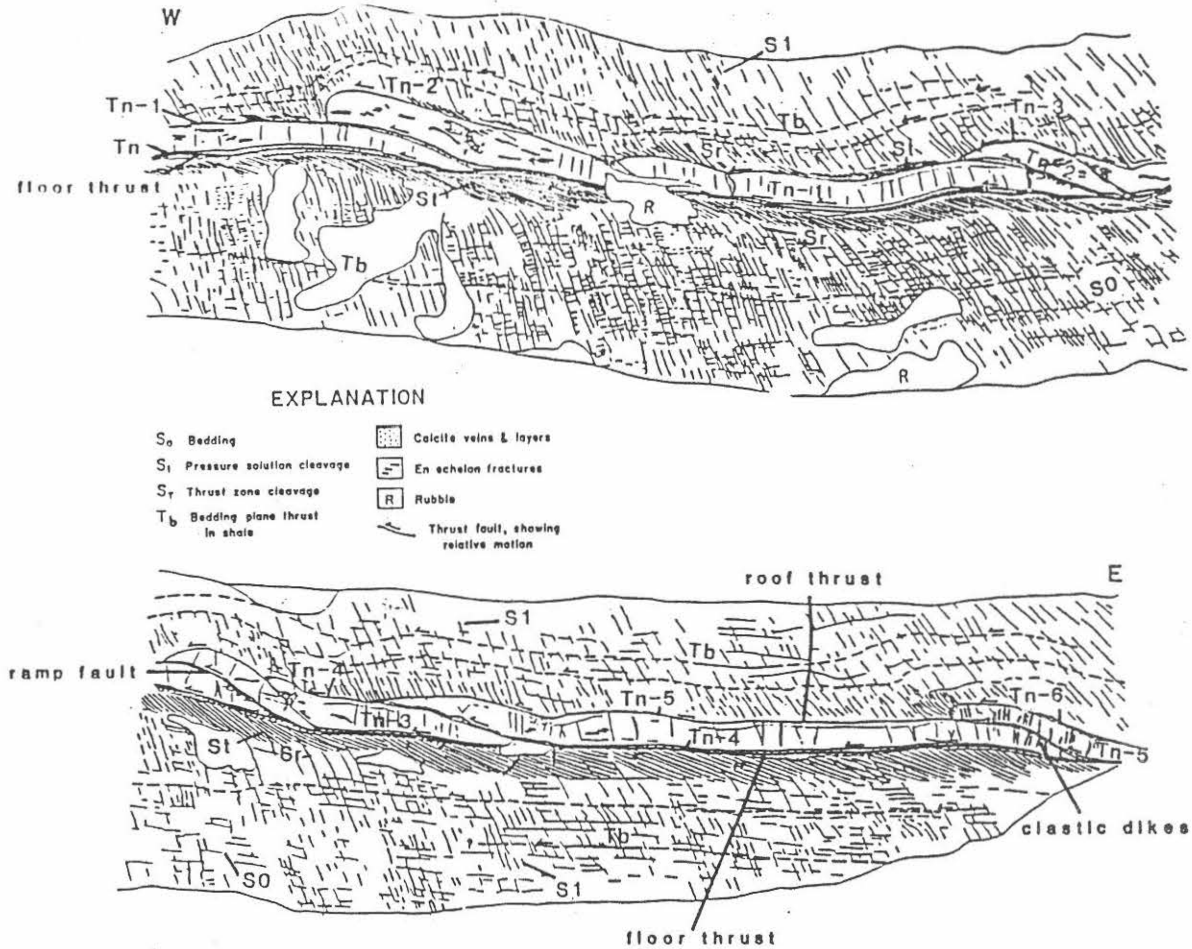
On the northeast side of the quarry (fig. 6) a syncline and an associated blind , synformal thrust fault are truncated by the major thrust fault that is continuous across the north wall of the quarry. The origin of this structure is not clear, but it is thought to be associated with a duplex or ramp below the level of the quarry floor.

STOP 3 - "THE BEAM" - THIS IS A SUPERB OUTCROP THAT SERVES AS A FIELD LABORATORY FOR RESEARCH AND TEACHING OF FORELAND DEFORMATION. PLEASE STUDY IT. USE YOUR CAMERAS BUT NOT YOUR HAMMERS. REFER TO FIGURES 7, 8, AND 9.

The outcrop is located in the Cumberland Head Formation approximately 5 miles (8 km.) west of the exposed front of the Champlain thrust fault or approximately 4600 ft. (1400 m.) below the restored westward projection of the thrust surface. The major questions that will be discussed are: 1. How do ramp faults form ?, 2. Are there criteria to determine if imbricate thrust faults develop toward the foreland or hinterland ?, 3. What is the relation between faulting and cleavage development ?, 4. What processes are involved in the formation of fault zones ?, 5. Are there criteria that indicate the relative importance and duration of motion along a fault zone ?, 6. Is there evidence that abnormal pore pressure existed during faulting ?, and finally 7. What is the structural evolution of the imbricate faults ? The first six questions will be largely addressed by direct evidence at the outcrop. The last question will be answered by palinspastically restoring the imbricated and cleaved sequence to its undeformed state (fig. 9).

THE OUTCROP

Five imbricate thrust faults and associated ramps are exposed in profile section in a foot thick bed of micrite that extends 14 m along an azimuth of N 80 E (fig. 7a). The imbricate fault can be further classified as a central duplex of three horsts that are separated from two simple ramps at either end of the outcrop by approximately 2.5 m of flats. The micrite bed is surrounded by at least 1.5 m of well-cleaved calcareous shale. Bedding plane faults are present along the upper and lower surface of the micrite where they merge with ramp faults that cut across the micrite bed at an angle of approximately 30 degrees. The lower bedding plane fault or floor thrust is relatively planar and



Stanley 1985

Figure 7a

DEFORMATION OF THE CUMBERLAND HEAD FORMATION

SOUTH HERO, VERMONT

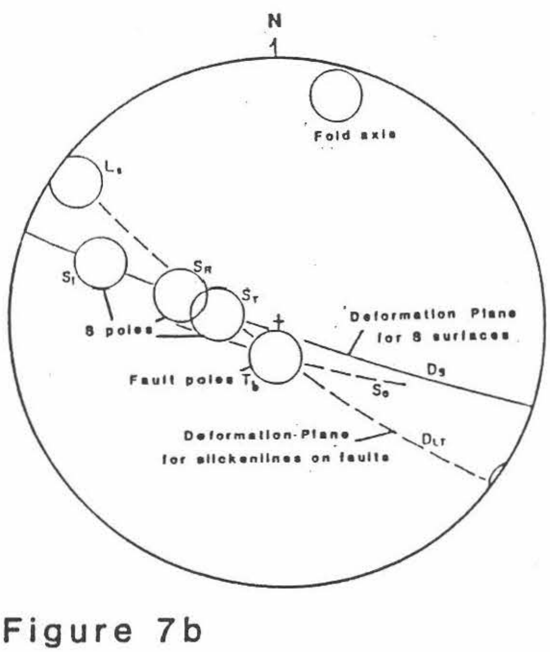


Figure 7b

the fault zone is thick compared to the upper bedding plane faults which are folded in the ramp areas, cut by ramp faults, and the fault zones are thin. The upper bedding plane fault forms the roof thrust for the central duplex. Along the intervening flats the upper faults are generally planar although they are cut by the S1 cleavage in many places. The lower bedding plane fault is the major decollement across the outcrop. Older bedding planes faults are also present throughout the shale where they are offset by the penetrative S1 cleavage.

All the fault surfaces are covered by layers of sparry calcite that vary in thickness from several millimeters to 4-6 cm. The thickest zone is found along the decollement or floor thrust whereas the thinest zone is found along the older bedding plane faults in the calcareous shale (fig. 7a). In all but the thinest layers, the calcite is arranged in distinct layers that are separated by discontinuous selvedge of dark gray shale. Each of the layers are marked by grooves or slickenlines that trends N 56 W and are essentially parallel from layer to layer (7b). In sections oriented perpendicular to the layering and parallel to the slickenlines the shale selvedges are more parallel to each other than in sections cut perpendicular to the slickenlines where the selvedges either anastomse or conform to the cross section of the grooves. In the parallel sections, however, some of the selvedge layers are truncated by more continuous surfaces. One of these can be traced for 1.5 m or more along the decollement. At a number of places along the different fault zones small dikes of sparry calcite have intruded the lower part of the calcite-shale layers. The fault-zone fabrics are most clearly displayed along the decollement where the calcite-selvedge layers are more abundance.

Fractures are common throughout the micrite bed where they are oriented at either a high angle or low angle to the bedding. Most fractures are filled with sparry calcite. The most prominent fractures are arranged in en echelon arrays that climb either to the west or the east. Most of these arrays are located in ramps areas and along the west-facing limbs or small flexures. A few are present in the flat regions of the bed. The fractures in many of the arrays in the ramp regions are folded and some are cut by younger generations of en echelon fractures. In the eastern ramp the hangingwall and footwall are cut by near-vertical fractures that are filled with fragments of the surrounding micrite embedded in sparry calcite so as to form clastic dikes.

Two, well developed, pressure-solution cleavages are present throughout the calcareous shale. The first and most conspicuous one, S1, strikes N 20 E and dips 60 to the east (fig. 7b). This orientation is an average based on measurements taken across the vertical face of the outcrop because the individual S1 surfaces are quite wavy along strike. As a result they form a distinct diamond-shape pattern on the bedding surface. The acute angle of the diamond pattern is approximately 30 degrees in the shale and 50 degrees in the micrite bed. This geometry indicates a moderate level of cleavage development according to the scheme of Alvarez and others (1976). S1 surfaces are covered by a black, carbon-rich selvedge of illite and kaolinite which is less than 0.5 cm thick. Although many of the cleavage surfaces are vertically continuous through the shale, some of them are short and discontinuous with tapered ends. The thickest selvedge occurs on the most continuous surfaces. The surfaces of the selvedge are not lineated although some are polished. The S1 cleavage offsets bedding and the older bedding-plane thrust faults with a down-to-the-east sense throughout much of the outcrop. This displacement is greatest where a selvedge is the thickest and it gradually is reduced to zero as a selvedge thins toward the tapered ends of the shorter cleavage surfaces. The average width of the microlithons between the S1 surfaces is 5.6 cm thick. The S1 cleavage not only cuts the older bedding-plane faults but cuts across most of the roof faults on top of the micrite bed.

A second, well developed cleavage, St, is resticted to a foot-thick zone directly below the decollement (fig. 7a). The individual cleavage surfaces are thinner (about 1 mm thick), more closely spaced (about 1 cm or more), and are covered with a thin selvedge (less than 1 mm). Furthermore, St dips to the east at only 6 degrees compared to the steeper dip of S1. St is also developed along the roof thrust of the micrite bed, but the zone is thinner and it is more difficult to recognize because the roof faults have been folded in the ramp areas and deformed by S1. The St cleavage is definitely related to movement of the thrust faults because it is only found near the faults and it is absent away from them.

In the zone near the decollement and the roof faults the S1 cleavage is rotated eastward so that the steeper 60 degree dip in the shale is reduced to 25 degrees (fig. 7b). The strike of the rotated S1 (hereafter referred to as Sr) is the same as S1 away from the faults. The counterclockwise rotation of S1 through an angle of shear of 35 degrees indicates that movement on the decollement was east-over-west along a direction of N 56 W as indicated by the slickenlines along the decollement. This sense of displacement is consistent with the orientation of St since the normal to St would correspond to the direction of maximum finite compressive strain.

Cleavage, similar to S1 and St, is totally absent from the micrite bed. In a few places, however, a very thin (less 1 mm), styolitic to very irregular cleavage is oriented perpendicular to bedding in the micrite. This cleavage is not developed uniformly throughout the micrite. Where it is formed the cleavage surfaces are separated from each other by at least 6-10 cm. The form, orientation and limited distribution of this cleavage indicates that it formed very early in the deformation sequence while compression was essentially parallel to the planar micrite bed.

RAMP FAULTS

All the east-dipping ramp faults develop from arrays of west-climbing, en echelon fractures. Although a few of the fractures are cavities covered by terminated calcite crystals, most of them are completely filled with sparry calcite. The sequence by which these faults develop is well displayed in a few gently folded parts of the micrite bed and in the ramp regions where the en echelon arrays are folded or cut by continuous ramp faults. These arrays appear to nucleate along the west-facing limbs of slightly asymmetrical buckle folds. The individual extension fractures within the arrays dip more steeply to the west by values of 7 to 11 degrees than the nearby bedding at the boundary of the micrite bed. The sequence continues with the rotation of the extension fractures by distributed shear strain along the west-climbing array. The individual fracture continued to grow as the older central parts of the fractures were rotated counterclockwise to produce S shaped fractures. Shear strains along these arrays were calculated using equation 2.3 of Ramsay and Huber (1983) and are in the order of 1.4 (54 degrees). These strains indicate that at least 5.6 cm of displacement occurs across the arrays before they fail along a ramp fault. During this time new arrays of planar fractures were developed over the older arrays thus producing a weakened zone along the trend of the array. As generations of fractures were superposed they eventually coalesced and developed into ramp faults as demonstrated by deformed fracture arrays in the footwall and hangwall which are truncated by the ramp faults. The presence of sparry calcite and cavities in the fractures indicates that the fractures opened rapidly and at a shallow enough level in the crust so that the micrite bed was strong enough to support the shape of the cavities.

East-climbing, en echelon fracture arrays are also present in the micrite where they tend to concentrate in the ramp regions. Unlike the fractures in the west-climbing arrays, the individual extension fractures in the east-climbing arrays are either parallel to or dip more steeply to the east than bedding. These angular relations are important because they indicate that the east-climbing arrays were not formed at the same time as the west-climbing arrays. Although some of the fractures in these arrays are deformed into Z shapes, none of the arrays developed into west-over-east faults or backthrusts. Calculated shear strains along these arrays give a maximum value of 1.1 (47 degrees). I suggest that they formed as a result of continued compression after the ramp faults were locally locked and before failure developed in another ramp zone farther to the west.

The orientation of the extension, or perhaps true tension, fractures in the micrite bed are important because they indicate the relative position of the bedding relative to the direction of maximum compressive stress (Sigma 1) which was essentially horizontal during deformation. During the early stage of deformation bedding was essentially parallel to Sigma 1 because the styolitic cleavage, which forms perpendicular to sigma 1, is perpendicular to bedding. The bedding then rotated eastward so that the extension fractures in the west-climbing arrays formed at a more-westerly inclined angle than bedding. East-over-west displacement in bedding plane faults could occur at this time because the finite shear stress was now high enough to overcome frictional resistence. Subsequent formation of ramp faults and their evolution led to the development of the east climbing arrays as a result continued compression. The orientation of these extension fractures is then controlled by the local orientation of sigma 1 in the ramp regions. Single, bed-parallel fractures in the flat regions of the micrite bed could have formed any time during the deformation sequence.

FAULT CHRONOLOGY

The discussion here will concentrate on the chronology of faults that are present within and in direct contact with the micrite bed. The bedding plane faults in the surrounding shales are clearly older than these faults because they are cut and offset by the S1 cleavage. Although the bedding plane faults that are in direct contact with the micrite bed may have formed originally at this time there is abundant evidence that they certainly were active long after those in the shale. Their average age therefore is younger. Furthermore, they play an essential role in the development of the ramp faults.

The evidence for the relative age of each of the faults is found in the regions where the ramp faults merge with the roof and floor faults. For example, in the western part of figure 7a the ramp fault (Tn-1) cuts across the upper-bedding plane fault (Tn-2) of the hangingwall block but merges asymtopically with the floor fault (Tn) of the footwall block. Furthermore, the roof fault of the hangingwall block is folded and cut by S1 cleavage. At the junction of the ramp fault and the floor fault, the quasiplanar slip surfaces in the calcite-shale fault zone cut across the fault zone of the ramp fault. The relative age relations therefore are clear - the oldest fault of the three is the roof thrust and is designated Tn-2. The youngest fault is the floor thrust

(Tn). This same relative chronlogy applies to all the ramps and duplexes to the east.

Several important conclusions result from this analysis. First, the imbricate faults which comprise the roof, ramp, and floor fault system become younger to the west or the foreland for the Northern Appalachians. This conclusion is based on direct evidence and not an assumption on the fault sequence. Second, the floor fault is continually reactivated during the evolution of the fault system. The average age of the fault zone therefore becomes younger to the west. It is a time transgressive structure in very sense of the word. Furthermore, it is the most continuous fault in the outcrop. This second conclusion explains why the fault zone for the floor fault is thicker than the fault zones along the ramp or roof faults. The floor fault has been active for a longer period of time than any of the other faults. The thickness and continuity of a fault zone therefore are directly related to the duration of motion of a given fault. How far this relationship can be carried to faults in other sequence is uncertain. It does seems to hold true for the shale section in western Vermont.

The geometry of the westward-younging faults in the micrite bed is distinctive - a flat, continuous floor fault with folded roof faults that are cut by ramp faults along their steeper west-facing limbs (fig. 7a). This is the standard "piggyback" sequence described for many mountain chains in which the thrust faults young toward the foreland (Boyer and Elliot, 1982; for example). Would the geometry, however, be different for an eastward-younging sequence or, in this case of the Northern Appalachians, a sequence that would young toward the hinterland? As shown in figure 8, the answer is yes. Here the roof fault can take on various configurations as irregularities caused by earlier ramp anticlines are eliminated by continued movement on the roof fault zone. Note that each of the ramp faults are truncated by the younger roof thrust. The floor faults are irregular and are cut successively by each of the ramp faults to the east. Ramps and duplexes with this geometry have not been recognized in the shales of western Vermont.

FAULT-ZONE DEPOSITS

With our understanding of fault chronology and importance, I now consider the processes by which the sparry calcite and shale are formed along the fault zones. Because the floor fault has been active throughout the development of the imbricate system one can find the most complete history there compared to the other faults where the history has been shorter and less complex. The fact that the average age of the fault zone along the floor fault is older to the east than the west, further suggests that more of its history will be recorded at the eastern part of the outcrop than the

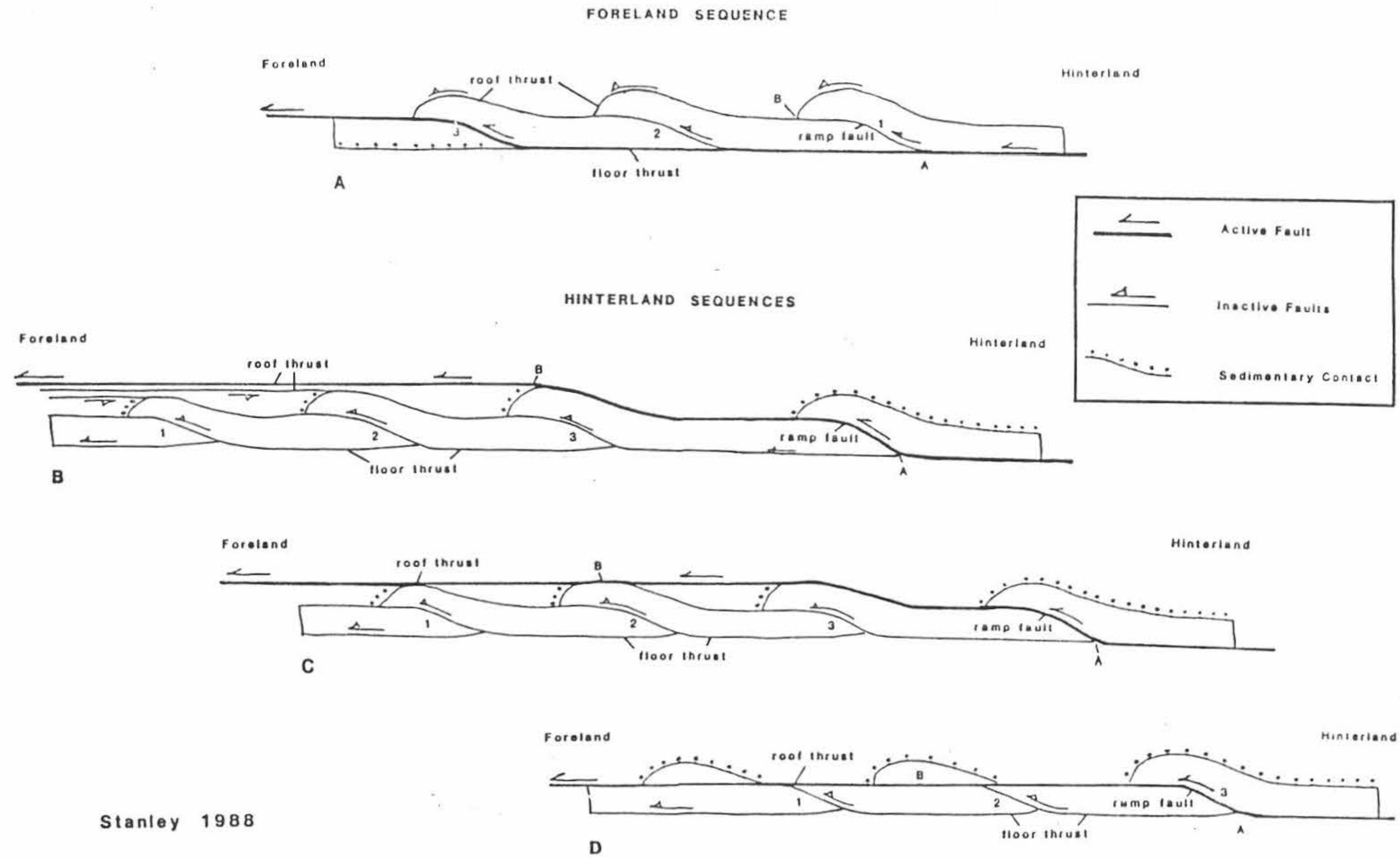


Figure 8 Comparison between imbricate thrust faults that become younger toward the foreland (foreland sequence) and those that become younger toward the hinterland (hinterland sequences). In all models the ramp faults are numbered 1, 2, and 3 in the direction of the ramp with the youngest age. In A, the foreland sequence, the older roof thrusts are cut by the younger ramp faults (loc. B) whereas the ramp faults are cut by the floor thrust nearest the foreland (loc. A). The age and the fault-zone fabric along the floor thrust become older and more complex toward the hinterland. In the hinterland sequences each floor thrust toward the foreland is cut by the ramp fault with the next higher number toward the hinterland (loc. A, for example). The roof thrusts must cut across folded bedding (not shown) along the limb facing the foreland (loc. B) in B and C. D is considered an unlikely possibility in nature because the roof thrust must cut across the strong rocks of the "beam" rather than follow the weak layer between the "beam" and the overlying shale. In all hinterland sequences the age and fault-zone fabrics along the roof thrust become older and more complex toward the foreland. In C the number of roof thrusts increases toward the foreland. Their individual fault fabrics may be relatively simple since they may not be involved with repeated movement during each imbrication. The position of the sedimentary contacts, here defined as a contact that is not involved with shortening of the "beam", is a very important criterion that separates foreland from hinterland sequences. In the hinterland sequences either the anticlinal limb facing the foreland (B and C) or the full ramp anticline (D) are sedimentary contacts. In the foreland sequence the ramp thrust caps all the anticlines. A basic assumption in all these models is that faults that have undergone repeated movement during their evolution will be planar or geometrically simple. These fault are the labelled the "active faults" in the models.

western part. The fabric along the floor fault, however, appears to be just as complex to the west as it is to the east. I have not been able to identified any feature that can be correlated directly with age. Other such factors as fault junctions in the ramp or duplex areas and broad flexures along the intervening flats control the thickness and complexity of the fault zone and, hence, overshadow any age-related fabric that might be preserved along the length of the fault. Leonard (1985) studied the fault zones in the field and in oriented thin sections cut parallel and perpendicular to the slickenlines.

The important facts that bare on the evolution of the fault-zones are the following:

1) All the faults zones are filled with veins of sparry calcite and minor quartz which are generally oriented either parallel or at a low angle to the fault surface. Vertical veins, which commonly join sills higher in the fault zone, are more common in the lowest layers of the fault zone and in the shale directly beneath the fault zone. Here in the shale some of the thinnest veins are filled with fibrous calcite and shale chips which are alined parallel to the vein walls in the typical "crack-seal" pattern of Ramsay (1980).

2) Discontinuous, dark, styolitic clay laminae with concentrations of quartz adjacent to or within the laminae are interlayered with the calcite in all but thinnest zones. The laminae are identical in appearance to the selvedges on the S1 cleavage surfaces. Some of the clay laminae are continuous with chips of shale which are either completely enclosed within a calcite layer or occur at the boundary between two calcite veins. These shale chips preserve varying degrees of pressure solution. For example, some of the chips are similar to the undeformed shale in microlithons away from fault zones whereas others have dark, thin selvedges within the chip and along their edges.

3) A relative planar surface decorated with shale laminae cuts older surfaces in the fault-zone deposit and is continuous for 1.5 or 3 m along the floor fault. Such surfaces as this are called <u>slip surfaces</u>.

4) The size of the sparry calcite is directly proportional to vein width. Most grains are bladed in form, but their long axes is not prefferentially alined. The calcite in all the layers is twinned with the greatest density occuring in the thinner layers between shale laminae where the grains are turbid and small. The larger grains in thicker layers nearest the planar slip surface are generally more twinnned than are those grains in veins further away.

5) Quartz occurs in a variety of forms. Large euhedral quartz crystals are found here and there in the middle of wide calcite veins. The quartz is deformed by fracture

whereas the calcite is twinned. Small crystals occur along the edges of selvedges. Very small grains with oriented C axes and indistinct borders are oriented parallel to some vein boundaries. In other areas calcite and quartz appear to be randomly intergrown with indistinct and irregular boundaries. The quartz in the microlithons in the shale away from faults is equant whereas it is elliptical in the well cleavage zone within a foot of the floor fault.

6) The slickenlines on each of the vein layers are formed by grooves rather than calcite fibers.

7) The calcite-shale layers tend to be more parallel in sections cut parallel to the slickenlines rather than they are in sections cut perpendicular to the slickenlines where the layers are irregular or anastomose.

It is clear from the foregoing information that the calcite has been intruded along the faults after the initial cohesion had been broken along the shale-shale or shale-micrite contacts where the strength contrast is the greatest and the cohesive strength the weakest. Once such a zone has developed and is filled with calcite, it becomes a zone of weakness. Subsequent failure likely occurred along the calcite-shale interface and resulted in scabs and chips of the shale being incorporated into the fault zone. Other shale fragments may have been sheared in along faults or carried in along thick veins. During renewed movement the calcite and quartz were dissolved from the shale to form the black selvedges which are interlayered with the calcite and decorate the slip surfaces. These boundaries then formed weak planes along which subsequent movement occurred within the fault zones. As the fault zone thickened movement could occur along planar surfaces (slip surfaces) which smoothed out the irregular geometry formed by ramp zones and broad folds in the intervening flat regions of the micrite bed. Movement therefore was no longer restricted to the calcite-shale boundaries of thin fault zones, but could occur along any favorably situated calcite-selvedge boundary. The resulting clay selvedge then acted as a catalyst that facilitated solution of calcite from the selvedge-calcite boundary of the surrounding veins. This process resulted in the stylolitic form of the selvedge. Leonard (1985) suggested that preferential solution in the direction of fault movement produced the slickenline grooves and the stylolitic selvedges best seen in sections cut perpendicular to the slickenlines.

Because the floor fault was continually active during the evolution of the imbricate system, it is not surprising to find evidence for repeated vein injection in the form of numerous crosscutting veins in the thick fault zone deposit. During each of these events the influx of fluid and the subsequent crystallization were relatively rapid so that sparry calcite formed rather than fibered calcite. As the fault zone thickened with layers of calcite and clay selvedge, new veins could form along any surface of weakness within the fault zone rather than being confined to the outer borders with the country rock. As movement continued across the fault zone, the calcite in the older veins became heavily twinned and severely strained. Repeated solution of calcite along their boundaries with the adjacent clay selvedges reduced their thickness and produced the common observation that calcite in many of the thinner veins are heavily twinned.

The senario that has been inferred from the fault zone fabrics and the relative age relations among the faults suggests that fault movement was intermittent with each event occuring rapidly. During the intervening time deformation in the fault zone may have been restricted to twinning in the calcite.

SHORTENING AND CLEAVAGE DEVELOPMENT

The shortening across the outcross is conveniently recorded by the folds in the ramps areas and structural overlap across the faults. The displacement on each of the faults in the micrite bed ranges from 7.6 to 48.3 cm which adds up to a total displacement of 134 cm (1.34 m) over a present horizontal distance of 10.7 m. The five anticlines over ramps and the broad folds along the flats account for approximately 12.7 cm of additional shortening so that the total shortening equal 146.7 cm (1.47 m). These values correspond to a shortening of 13.7 percent of which folding only accounts for 8.7 percent of the total reduction in length.

In the shale the corresponding shortening is provided by volume reduction across the cleavage surfaces, S1 and to a far lesser degree St. It is clear that these cleavages formed by pressure solution because they are marked by insoluble residue. Furthermore the abundance of calcite and minor quartz in fractures and along the faults indicates that these two minerals were dissolved from the calcareous shale and deposited in nearby openings. In order to see if the shortening determined from the micrite bed is comparable to the shortening in the shale, an independent estimate therefore was made for the shale by determining the percentage of insoluble material. Samples of suitable material from four different microlithons where immersed in hydrochloric acid until all the soluble material was eliminated. The final average residue was 36 percent of the original mass (a range of 32% to 39% for 4 samples). The number of cleavage selvedges was then counted across a present outcrop width of 7.9 m measured perpendicular to the cleavage. The total width of the selvedges (a range of 35.4 to 54 cm) was then multiplied by 2.8 to given an estimate of the original width now represented by the cleavages (99 to 152 cm). The original length of the present 7.9 m width was then estimated to be 8.8 to 9.5 cm. Shortening was then calculated to be in the range of 11 to 16 percent. Although this range overlaps the shortening value determined from the "beam", the shale value is less accurate because there is more error in estimating the thickness and number of the cleavage selvedges.

I therefore conclude that the formation of the cleavage and consequent shortening in the shale occurred during imbricate faulting in the micrite bed. Although this conclusion may seem intuitively obvious, it does have important implications for the evolution of the cleavage. Because I have already proven that the faults represent a time-transgressive sequence that developed from east to west, I must also conclude that the cleavage in the surrounding shales developed in a similar mannner. Unlike my earlier conclusion, this relation is far from obvious if my observations were restricted just to the cleaved shale. In fact it is the existence of the micrite bed and its fault geometry that allows me to conclude that the cleavage in the shale is indeed a time transgressive phenomena.

The next problem is the origin of Sr, the rotated cleavage, and St the finely spaced cleavage below the floor thrust. Because Sr is simply the dominant S1 cleavage that has been rotated in simple shear near the floor and roof faults, it had to form during subsequent displacement on these faults. Measurements of cleavage rotation taken at 8 locations below the micrite bed averaged 36 degrees whereas similar measurements taken at 7 locations above the micrite bed averaged 27 degrees (Table 1). These line rotations correspond to angular shear strains of 51 degrees for the floor thrust and 33 degrees for the various roof thrusts using equation 2.3 of Ramsay and Huber (1983) in which Gamma, the shear strain = cot alpha - cot alpha'. In this equation alpha and alpha' are the angles of S1 and St respectively from the fault surface. The fact that the Sr cleavage below the thrust has been rotated in most cases through a greater angle of shear than Sr above the thrust is consistent with my earlier conclusion that the floor thrust was active throughout deformation whereas the individual roof faults are short lived and are only active during the time interval between the formation of each of the imbricate faults.

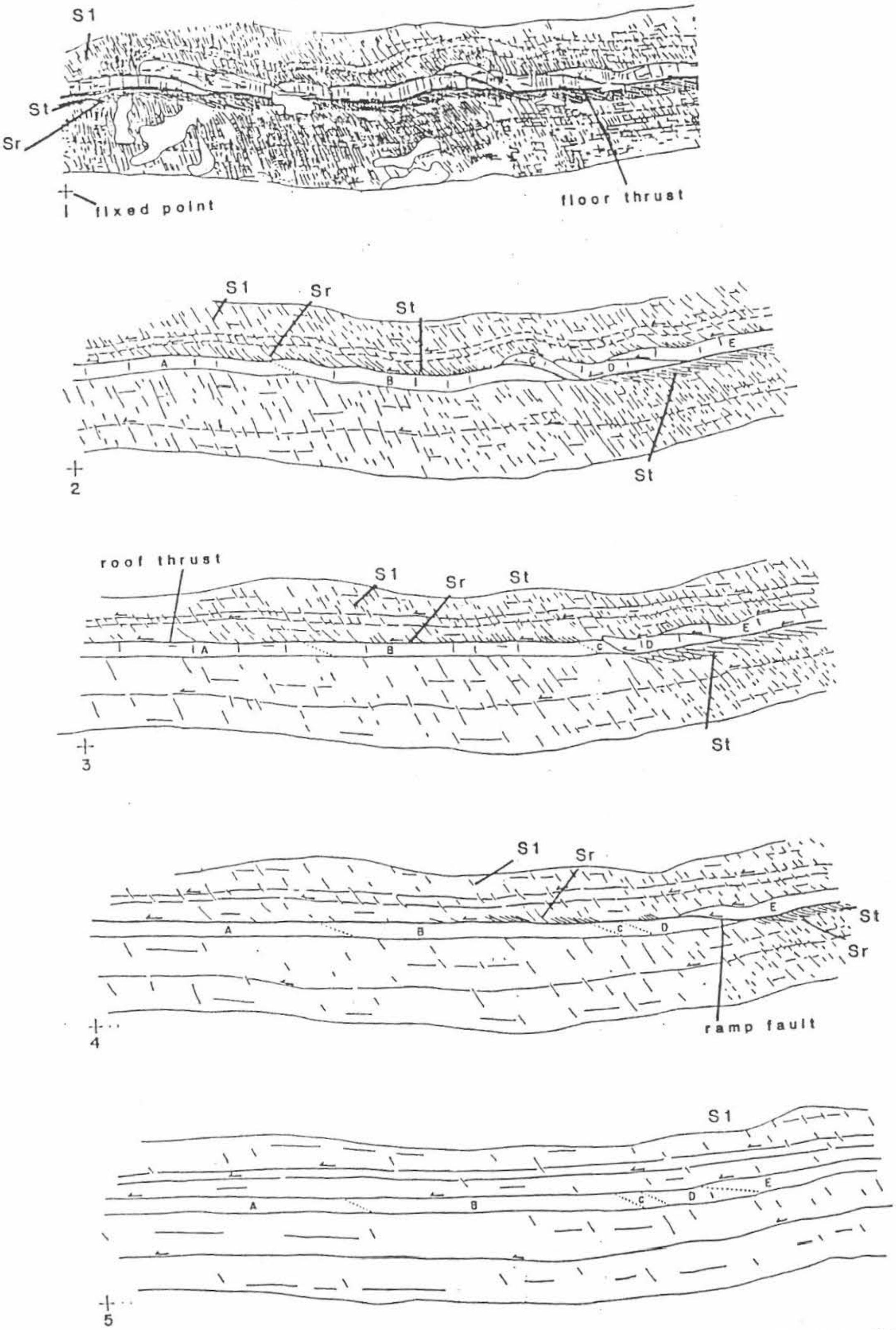
The St cleavage clearly formed after Sr because it cuts across Sr at a low angle and is not folded or rotated near any of the faults. Although St is best developed below the floor thrust, it is also present just above the roof thrusts where the cleavage is less distinct and occupies a thinner zone compared to the comparable cleavage along the floor thrust. Because the St cleavage is confined to a narrow zone near the faults, it most likely developed as a result of simple shear which rotated the S1 cleavage. In order to test this hypothesis the angle between St and the nearby fault, O', was measured and compared with a calculated value predicted by the Ramsay equation Tan 2O'=2/gamma (Ramsay and Huber, 1983, equation 2.4;Table 1, this paper). The values of gamma were calculated from the rotation of the cleavage using the equation described in the foregoing paragraph (Ramsay and Huber, 1983, eq. 2.3). This equation predicts the acute angle, O', between the plane defined by Lamda 1 and Lamda 2 of the strain ellipsoid (long axis of the strain ellipse in 2 dimensions) and the plane of simple shear which in this case is the fault surface. In the ideal case of simple shear the St cleavage will be parallel to the Lamda 1-Lamba 2 principal plane of the strain ellipsoid. Thus the calculated and observed values should be the same within the margin of error providing no mechanism other than simple shear was involved. Along the floor fault the calculated angle between St and the fault is an average of 29 degrees for 8 separate locations. The observed value, however, was an average of 21 degrees. Along the higher roof thrust the calculated angle between St and the nearby fault was an average of 37 degrees for 6 locations whereas the observed value was 32 degrees. Although the sample population was small and the corresponding standard deviations large for both the calculated and observed values, the differences do suggest two important relations. First, the angle O' between the St cleavage and the fault is larger along the roof thrust than it is for the floor thrust. This relation simply reflects the smaller shear strain along the roof thrust compared to the floor thrust. Second, and more importantly, the calculated angle O' between St and the nearby faults is consistently larger than the observed angle by values that range from 8 degrees for the floor thrust to 5 degrees for the roof thrust. This relation is important because it suggests either that the St cleavage has been subsequently flattened by some sort of vertical loading after simple shear or that the volume reduction, which has occurred during the formation of the St cleavage, has indeed reduced the O' angle. The volume reduction associated with the formation of the St cleavage was probably less than the 10 percent calculated for the S1 because the St surfaces are much thinner than S1 although they are more numerous.

Are Sr and St time transgressive? Because I have demonstrated earlier that the floor fault and the respective roof faults are time transgressive, it must follow that both Sr and St are also time transgressive to the west. This conclusion suggests that the amount of rotation of Sr should also increase to the east particularly along the floor thrust where the displacement is larger and has taken place over a longer period of time. Although this feature was evaluated across the outcrop, I could not detect within the error of my measurements any such relation for Sr. This observation suggests that the process by which the S1 cleavage is bent essentially work hardens and another deformational mechanism progressively takes over as the angle of line rotation increases during simple shear. This process appears to be the formation of the pressure-solution St cleavage because it is confined to the fault zone and is not bent like the older Sr cleavage. My measurements suggest that the St cleavage probably developed when the S1 cleavage was rotated somewhere in the neighborhood of 35 degrees.

EVOLUTION OF STRUCTURES

The evolution of the imbricate faults and the various cleavages described in the foregoing section is illustrated in a series of retrodeformed sections in figure 9. Section 1 shows the outcrop at South Hero in its present state. Section 2 is developed by reversing the deformation associated with the youngest ramp fault at the western part of the outcrop. The remaining four sections are formed by systematically unfaulting the micrite bed from west to east in the reverse order of their formation. For example, in section 2 the hanging wall block (B, section 2) is unfolded as it is returned to its original position east of the footwall block (A, section 2). During this time the active fault which carried the eastern sequence westward was the floor thrust below segments D and E, the ramp fault just below segment C, and the roof thrust above segments B and A, which were then undeformed. In the shale above and below this active fault the S1 cleavage is shown in a flatter position (Sr) as a result of rotation generated by east-over-west simple shear. Note that the rotated cleavage, Sr, is absent below blocks A and B since, at this time, the micrite bed is attached to the underlying shale. The fault-zone cleavage, St, is also shown in the shale along the active fault. I believe that this cleavage developed during the evolution of the imbricate structures rather than after all the imbricate faults had formed because St is present, although poorly developed, along the roof thrust above each segment of the micrite bed. If the fault-zone cleavage had formed after all the imbricate faults had moved into place, then it (St) would only be found along the floor thrust. The St cleavage therefore must have developed during the formation of each imbricate fault.

As a consequence of reversing the displacement on the fault and unfolding the ramp fold, the volume in the adjacent shale must be increased by eliminating much of the cleavage below segments A and B of the micrite bed. Note, however, that the density of S1 cleavage in the thrust plate above segments A and B is essentially the same across section 2. This diagram is drawn in this way because I believe that the upper plate of the active fault was probably being shortened as a result of ramps or irregularities along the fault surface farther to the west. In actual fact, however, the density of S1 in the upper plate must decrease to the west because, first, this is



RETRODEFORMED SECTION OF THE CUMBERLAND HEAD FORMATION

Figure 9 SOUTH HERO, VERMONT Stanley, 1987 what is observed to the west across South Hero Island and, second, the overall stress intensity during deformation in theoretical models diminishes from the hinterland to the foreland (for example, Hubbert, 1951; Chapple, 1979; Davis and others, 1982). Thus at any one time the cleavage density not only dimishes toward the foreland but it also changes in the same way from the upper plate to the lower plate (compared sections 2 through 5). As deformation progresses older cleavage surfaces continue to grow in thickness and lateral extent and younger surfaces nucleate in older microlithons and grow essentially parallel to the older S1 surfaces. In the end the S1 cleavage is uniformly developed across the outcrop and appears to the casual observer to be coeval across the outcrop because its style and orientation are the same. As can be seen in these series of sections, however, that the S1 cleavage is clearly a time transgressive structure.

Sections 3, 4 and 5 show the retrodeformation continuing to the east and are constructed in the same manner as section 2. Thus the evolution of the imbricate system and its associated structures can be seen by studying diagrams 5 through 1. Clearly all the structures are time transgressive from east to west as they are traced through sections 5 to section 1. During deformation the floor thrust or basal decollement undergoes repeated west-directed movement. As a result the Sr cleavage is rotated to a larger angle along the floor thrust than it is along the roof thrusts. When this angle reaches values in the range of 30 to 35 degrees, it appears that pressure solution takes over as the dominant shortening mechanism and the fault-zone cleavage, St, begins to be developed. Because the floor thrust continually moves the St cleavage is better developed there than along the higher roof faults.

<u>STOP 4 - THE CHAMPLAIN THRUST FAULT AT LONE ROCK POINT, BURLINGTON, VERMONT</u> - The following discussion is reprinted from The Centennial Field Guide, Volume 5, of the Geological Society of America in 1986. All the figure numbers for this stop refer to those figures in the reprint. The reprinted discussion appears in Appendix 1. Figure 10 is a regional map showing the location of the Champlain thrust fault in western Vermont. The Long Rock Point locality is identified as "LRP". The Arrowhead Mountain thrust fault is located east of the Champlain thrust fault and is identified by the letters "AMTF".

<u>STOP 5 - THE HINESBURG THRUST FAULT AT HINESBURG, VERMONT</u> - This is the classic and best exposed locality for the Hinesburg thrust fault (fig. 10, "HTFM"). It contains many fault related fabrics that have recently been studied by Strehle (1985) and published by Strehle and Stanley (1986) in a bulletin of the Vermont Geological Survey (Studies in

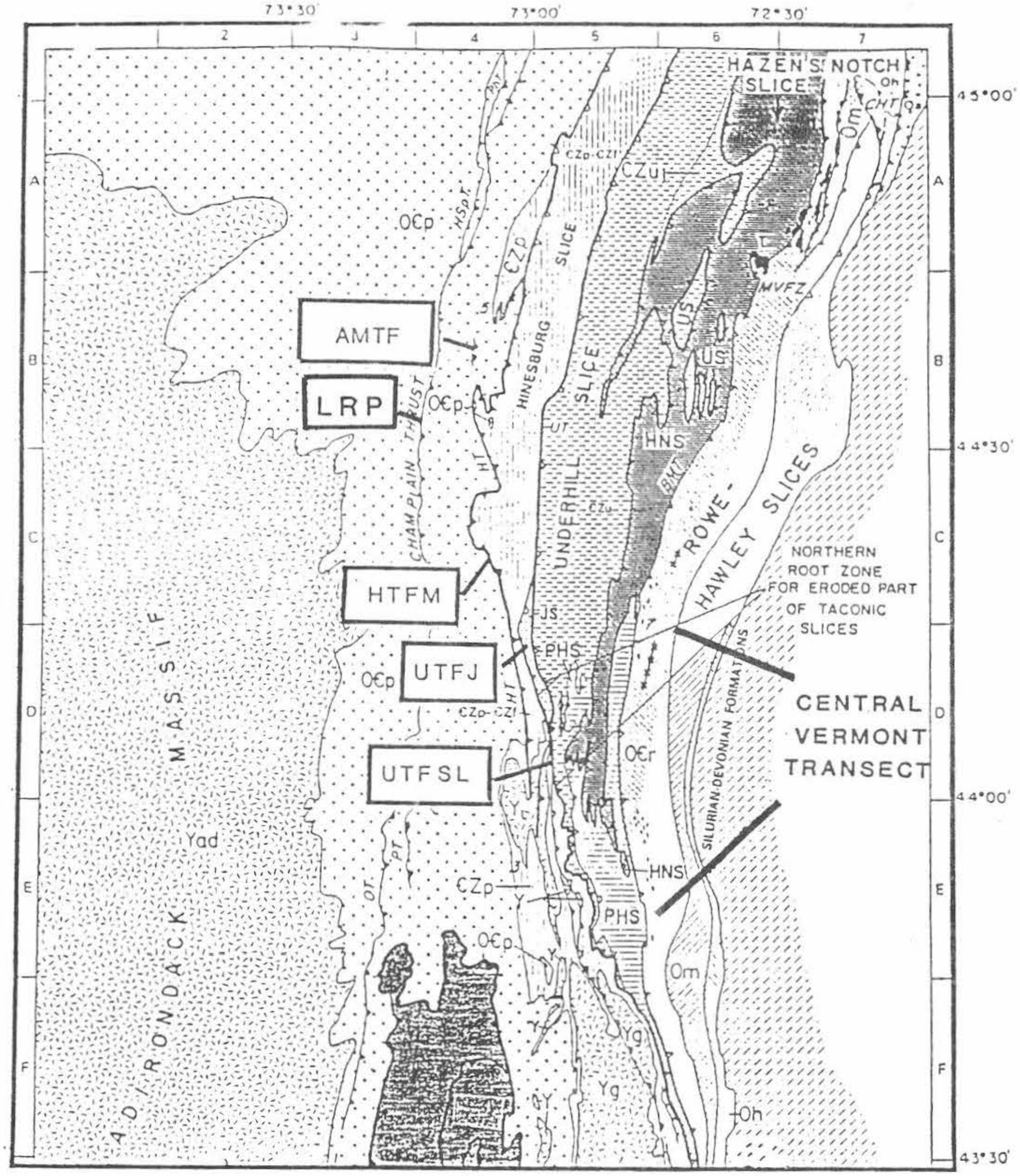


Figure 10 Interpretative Tectonic Map of Vermont and eastern New York showing the general location of the Arrowhead Mountain thrust fault (AMTF), the Hinesburg thrust fault at Mechanicsville (HTFM), and the Underhill thrust fault at Jerusalem (UTFJ), and South Lincoln (UTFSL). The geological map is taken from Stanley and Ratcliffe (1985, Pl. 1, figure 2a). Symbol T in A6 is a glaucophane locality at Tilliston Peak. Short line with x's (Worcester Mountains) and line with rhombs (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones of Albee (1968). Widely-spaced diagonal lines in northcentral Vermont outline the region that contains medium-high pressure amphibolites described by Laird and Albee (1981b). Irregular black marks are ultramafic bodies. Open teeth of thrust fault symbols mark speculative thrust zones. The following symbols are generally listed from west to east. Yad, Middle Proterozoic of the Adirondack massif; Yg, Middle Proterozoic of the Green Mountain massif; YL, Middle Proterozoic of the Lincoln massif; Y, Middle Proterozoic between the Green Mountain massif and the Taconic slices, Vermont; OCp, Cambrian and Ordovician rocks of the carbonate-siliciclastic platform; rift-clastic sequence of the Pinnacle (CZp) and Fairfield Pond Formations (CZf) and their equivalents on the east side of the Lincoln and Green Mountain massifs, PhT, Philipsburg thrust; HSpT, Highgate Springs thrust; PT, Pinnacle thrust; OT, Orwell thrust, UT, Underhill thrust ; HT, Hinesburg thrust; U, ultramafic rocks; CZu, Underhill Formation; CZuj, Jay Peak Member of Underhill Formation; OCr, Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation and its equivalents in Vermont; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens Notch slice; MVFZ, Missisquoi Valley fault zone; PHS, Pinney Hollow slice; BMT, Belvidere Mountain thrust; CHT, Coburn Hill thrust; Oa, Ascot-Weedon sequence in grid location 7A.

Vermont Geology No.3). This publication also contains analyses of other fault zones of western Vermont which will not be seen during this conference. The reader is referred to this paper or an earlier NEIGC trip by Gillespie and others (1972).

The Hinesburg thrust fault separates the Cambrian-Ordovician rocks of the platform sequence from the older, highly deformed metamorphic rocks of the eastern hinterland. As shown in figure 4, the Hinesburg thrust fault developed along the overturned, sheared limb of a large recumbent fold. This fault probably broke out from the overturned limb of a fault-propagation fold (Suppe, 1985) and therefore is similar in origin to the Arrowhead Mountain thrust fault. To the south the Hinesburg thrust fault dies out somewhere in the overturned limb of the Lincoln massif (Tauvers, 1982; DiPietro, 1983; DelloRusso and Stanley, 1986).

At the Mechanicsville locality the lower 40 m. of the Cheshire Quartzite is structurally overturned along the base of the upper plate of the Hinesburg thrust fault. Higher up the cliff the quartzite grades into the Fairfield Pond Formation of Tauvers (1982). The lower plate rocks, which are poorly exposed, consist of carbonates of the Lower Ordovician Bascon Formation. Slivers of dark gray phyllite of the Brownell Mountain Phyllite are found at several localities along the fault trace. Chlorite, muscovite, and stilpnomelane are present in the quartzite. Muscovite and chlorite are present in the schist. Quartz is thoroughly recrystallized, but feldspar grains are fractured and bent.

The following features should be studied here:

1) The change in fabric as the fault surface is approached. The quartzite grades from a protomylonite away from the fault to an ultramylonite near the fault. In thin section quartz becomes finer grained and quartz porphyroclasts decrease in number toward to fault.

2) The presence of east-over-west or "S" shaped asymmetrical folds. These folds are related to simple shear along the fault and are not related to the overturned limb of the older Hinesburg nappe. Parasitic minor folds related to this older structure would show a west-over-east or "Z" shaped asymmetry.

3) The prominent compositional layering. This layering is not bedding but represents the axial surface schistosity of the older parasitic folds related to the Hinesburg nappe. It is overprinted by a younger schistosity that is associated with the east-over-west folds. These two schistosities become a composite mylonitic schistosity as the fault surface is approached.

4) The prominent mineral lineation consisting of elongate quartz and quartz clusters.

5) "Z" shaped quartz veins that are associated with beds of quartzite. These structures are particularly interesting because the record progressive shear strain and mylonitization along the fault. The veins are confined to the metasandstone and metasiltstone layers and occur in at least 3 orientations. The least defomed veins are oriented either perpendicular to the layering or dip steeply to the west. They are filled with straight quartz fibers oriented perpendicular to the vein walls and show no signs of recrystallization except where they cross older "Z" shaped veins (Warren, 1988, pers. comm.). Here the fibers are reduced to smaller, nearly equant grains with undulose extinction. The most deformed veins are "Z" shaped and dip to the east. The older quartz fibers are completely replaced by fine grained, recrystallized quartz that forms a mylonite with distinct whitish layers oriented essentially parallel to the vein boundary. The third set is intermediate in orientation and deformation between the other two. The "Z" pattern develops because the shear strain (8.5 or an angle of shear of 84°) is higher in the pelitic units that surround the metasandstones where the shear strain is 1.48 (angle of shear of 56°). Strehle and Stanley (1986) suggested that these veins developed as shear fractures during east-over-west shear. This interpretation is not consistent with the fact that the youngest veins are vertical and consist of quartz fibers orientated essentially parallel to the layering (Warren, 1988, pers. comm.). It therefore appears that the veins formed as extension fractures during periods of vertical or near vertical loading. These periods of flattening (pure shear) were then followed by longer periods of east-over-west simple shear during which the veins were rotated westward and the characteristic "Z" pattern developed. The quartz, which was originally fibered, became progressively mylonitized as the extension veins were rotated into the favorable, east-dipping, shear position. Movement on these east-dipping veins continued even after they were cut by younger, unrotated veins because the fibered segments of the younger veins are deformed and recrystallized.

6) Rare west-dipping shear bands. These structures deform the younger schistosity which is parallel to the axial surfaces of the "S" shaped folds.

7) Westward displacement (N 75 W) of the upper plate of the Hinesburg thrust is documented by "S" shaped folds, "Z" shaped-quartz veins, quartz porphyroclasts, and late shear bands.

8) Late fractures and associated en echelon fracture arrays. These structures thought to be related to Mesozoic normal faults which cut the Champlain and Hinesburg slices.

One of these faults is located about 1000 m. directly west of this locality.

The interpretation of these structures and the thin sections fabrics are discussed in Strehle and Stanley (1986).

ACKNOWLEDGEMENTS

Many students have helped in the collection of data and the discussions that have resulted in my interpretations at the "beam", the Hinesburg thrust fault, and western Vermont. Although each has contributed in their own way, I would like to mention a few who have contributed above and beyond the "call to duty". At the "beam" John Humphrey provided data and calculations for shortening in the shale. John Delphia was one of the first to retrodeform the beam and demonstrate the time-transgressive nature of the cleavages. Vinnie DelloRusso supplied much of the data for figure 7b. Steve Taylor analyzed the quartz-fiber geometry associated with late pyrite at the Hinesburg thrust. Tom Armstrong stimulated revision of my earlier interpretations on the composition layering and the "S" shaped folds. Marian Warren provided important information on the "Z" shaped quartz veins at Hinesburg. Her work is still continuing. The work from South Hero was part of Kitty Leonard's Master of Science thesis (1985) while the Hinesburg thrust was part of Dick Gillespie's thesis in 1972. Barbara Strehle's work on the Arrowhead Mountain and Hinesburg thrust faults was very important to our present understanding of the transition from foreland to hinterland fault-zone fabrics. Last, but far from least, was the critical work and synthesis of Becky Dorsey. Her success was in part a result of earlier "pioneer" work by Eric Rosencranz, Paul Agnew, and Craig Carter. To all of these students and those that I have failed to mention - "thanks for the memories".

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APPENDIX 1 - LONG ROCK POINT - REPRINT.

The Champlain thrust fault, Lone Rock Point, Burlington, Vermont

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LOCATION

The 0.6 mi (1 km) exposure of the Champlain thrust fault is located on the eastern shore of Lake Champlain at the north end of Burlington Harbor. The property is owned by the Episcopal Diocesan Center. Drive several miles (km) north along North Avenue (Vermont 127) from the center of Burlington until you reach the traffic light at Institute Road, which leads to Burlington High School, The Episcopal Diocesan Center, and North Beach. Turn west toward the lake and take the first right (north) beyond Burlington High School. The road is marked by a stone archway. Stop at the second building on the west side of the road, which is the Administration Building (low rectangular building), for written permission to visit the field site.

Continue north from the Administration Building, cross the bridge over the old railroad bed, and keep to the left as you drive over a small rise beyond the bridge. Go to the end of this lower road. Park your vehicle so that it does not interfere with the people living at the end of the road (Fig. 1). Walk west from the parking area to the iron fence at the edge of the cliff past the outdoor altar where you will see a fine view of Lake Champlain and the Adirondack Mountains. From here walk south along a footpath for about 600 ft (200 m) until you reach a depression in the cliff that leads to the shore (Fig. 1).

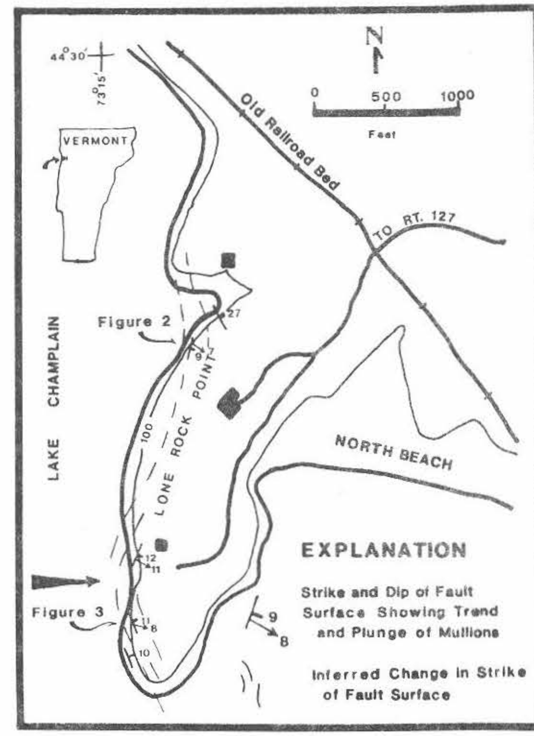

Figure 1. Location map of the Champlain thrust fault at Lone Rock Point, Burlington, Vermont. The buildings belong to the Episcopal Diocesan Center. The road leads to Institute Road and Vermont 127 (North Avenue). The inferred change in orientation of the fault surface is based on measured orientations shown by the dip and strike symbols. The large eastward-directed arrow marks the axis of a broad, late syncline in the fault zone. The location of Figures 2 and 3 are shown to the left of "Lone Rock Point." The large arrow points to the depression referred to in the text.

SIGNIFICANCE

This locality is one of the finest exposures of a thrust fault in the Appalachians because it shows many of the fault zone features characteristic of thrust faults throughout the world. Early studies considered the fault to be an unconformity between the strongly-tilted Ordovician shales of the "Hudson River Group" and the overlying, gently-inclined dolostones and sandstones of the "Red Sandrock Formation" (Dunham, Monkton, and Winooski formations of Cady, 1945), which was thought to be Silurian because it was lithically similar to the Medina Sandstone of New York. Between 1847 and 1861, fossils of pre-Medina age were found in the "Red Sandrock Formation" and its equivalent "Quebec Group" in Canada. Based on this information, Hitchcock and others (1861, p. 340) concluded that the contact was a major fault of regional extent. We now know that it is one of several very important faults that floor major slices of Middle Proterozoic continental crust exposed in western New England.

Our current understanding of the Champlain thrust fault and its associated faults (Champlain thrust zone) is primarily the result of field studies by Keith (1923, 1932), Clark (1934), Cady (1945), Welby (1961), Doll and others (1961), Coney and others (1972), Stanley and Sarkisian (1972), Dorsey and others (1983), and Leonard (1985). Recent seismic reflection studies by Ando and others (1983, 1984) and private industry have shown that the Champlain thrust fault dips eastward beneath the metamorphosed rocks of the Green Mountains. This geometry agrees with earlier interpretations shown in cross sections across central and northern Vermont (Doll and others, 1961; Coney and others, 1972). Leonard's work has shown that the earliest folds and faults in the Ordovician sequence to the west in the Champlain Islands are genetically related to the development of the Champlain thrust fault.

In southern Vermont and eastern New York, Rowley and others (1979), Bosworth (1980), Bosworth and Vollmer (1981), and Bosworth and Rowley (1984), have recognized a zone of late post-cleavage faults (Taconic Frontal Thrust of Bosworth and Rowley, 1984) along the western side of the Taconic Mountains. Rowley (1983), Stanley and Ratcliffe (1983, 1985), and Ratcliffe (in Zen and others, 1983) have correlated this zone with the Champlain thrust fault. If this correlation is correct then the Champlain thrust zone would extend from Rosenberg, Canada, to the Catskill Plateau in east-central New York, a distance of 199 mi (320 km), where it appears to be overlain by Silurian and Devonian rocks. The COCORP line through southern Vermont shows an east-dipping reflection that roots within Middle Proterozoic rocks of the Green Mountains and intersects the earth's surface along the western side of the Taconic Mountains (Ando and others, 1983, 1984).

The relations described in the foregoing paragraphs suggest that the Champlain thrust fault developed during the later part of the Taconian orogeny of Middle to Late Ordovician age. Subsequent movement, however, during the middle Paleozoic Acadian orogeny and the late Paleozoic Alleghenian orogeny can not be ruled out. The importance of the Champlain thrust in the plate tectonic evolution of western New England has been discussed by Stanley and Ratcliffe (1983, 1985). Earlier discussions can be found in Cady (1969), Rodgers (1970), and Zen (1972).

REGIONAL GEOLOGY

In Vermont the Champlain thrust fault places Lower Cambrian rocks on highly-deformed Middle Ordovician shale. North of Burlington the thrust surface is confined to the lower part of the Dunham Dolomite. At Burlington, the thrust surface cuts upward through 2,275 ft (700 m) of the Dunham into the thick-bedded quartzites and dolostones in the very lower part of the Monkton Quartzite. Throughout its extent, the thrust fault is located within the lowest, thick dolostone of the carbonate-siliciclastic platform sequence that was deposited upon Late Proterozoic rift-clastic rocks and Middle Proterozoic, continental crust of ancient North America.

At Lone Rock Point in Burlington the stratigraphic throw is about 8,850 ft (2,700 m), which represents the thickness of rock cut by the thrust surface. To the north the throw decreases as the thrust surface is lost in the shale terrain north of Rosenberg, Canada. Part, if not all, of this displacement is taken up by the Highgate Springs and Philipsburg thrust faults that continue northward and become the "Logan's Line" thrust of Cady (1969). South of Burlington the stratigraphic throw is in the order of 6,000 ft (1,800 m). As the throw decreases on the Champlain thrust fault in central Vermont the displacement is again taken up by movement on the Orwell, Shoreham, and Pinnacle thrust faults.

Younger open folds and arches that deform the Champlain slice may be due either duplexes or ramps along or beneath the Champlain thrust fault. To the west, numerous thrust faults are exposed in the Ordovician section along the shores of Lake Champlain (Hawley, 1957; Fisher, 1968; Leonard, 1985). One of these broad folds is exposed along the north part of Lone Rock Point (Fig. 2). Based on seismic reflection studies in Vermont, duplex formation as described by Suppe (1982) and Boyer and Elliot (1982) indeed appears to be the mechanism by which major folds have developed in the Champlain slice.

North of Burlington the trace of the Champlain thrust fault is relatively straight and the surface strikes north and dips at about 15° to the east. South of Burlington the trace is irregular because the thrust has been more deformed by high-angle faults and broad folds. Slivers of dolostone (Lower Cambrian Dunham Dolomite) and limestone (Lower Ordovician Beekmantown Group) can be found all along the trace of the thrust. The limestone represents

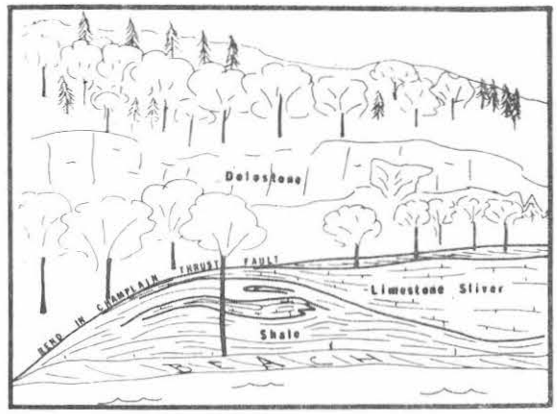

Figure 2. A sketch of the Champlain thrust fault at the north end of Lone Rock Point showing the large bend in the fault zone and the slivers of Lower Ordovician limestone. The layering in the shale is the S1 cleavage. It is folded by small folds and cut by many generations of calcite veins and faults. The sketch is located in Figure 1.

fragments from the Highgate Springs slice exposed directly west and beneath the Champlain thrust fault north of Burlington (Doll and others, 1961). In a 3.3 to 10 ft (1 to 3 m) zone along the thrust surface, fractured clasts of these slivers are found in a matrix of ground and rewelded shale.

Estimates of displacement along the Champlain thrust fault have increased substantially as a result of regional considerations (Palmer, 1969; Zen and others, 1983; Stanley and Ratcliffe, 1983, 1985) and seismic reflection studies (Ando and others, 1983, 1984). The earlier estimates were less than 9 mi (15 km) and were either based on cross sections accompanying the Geologic Map of Vermont (Doll and others, 1961) or simply trigonometric calculations using the average dip of the fault and its stratigraphic throw. Current estimates are in the order of 35 to 50 mi (60 to 80 km). Using plate tectonic considerations, Rowley (1982) has suggested an even higher value of 62 mi (100 km). These larger estimates are more realistic than earlier ones considering the regional extent of the Champlain thrust fault.

Lone Rock Point

At Lone Rock Point the basal part of the Lower Cambrian Dunham Dolostone overlies the Middle Ordovician Iberville Formation. Because the upper plate dolostone is more resistent than the lower plate shale, the fault zone is well exposed from the northern part of Burlington Bay northward for approximately 0.9 mi (1.5 km; Fig. 1). The features are typical of the Champlain thrust fault where it has been observed elsewhere.

The Champlain fault zone can be divided into an inner and outer part. The inner zone is 1.6 to 20 ft (0.5 to 6 m) thick and consists of dolostone and limestone breccia encased in welded, but highly contorted shale (Fig. 3). Calcite veins are abundant. One of the most prominent and important features of the inner fault zone is the slip surface, which is very planar and continuous throughout the exposed fault zone (Fig. 3). This surface is marked by very fine-grained gouge and, in some places, calcite slickenlines. Where the inner fault zone is thin, the slip surface is located

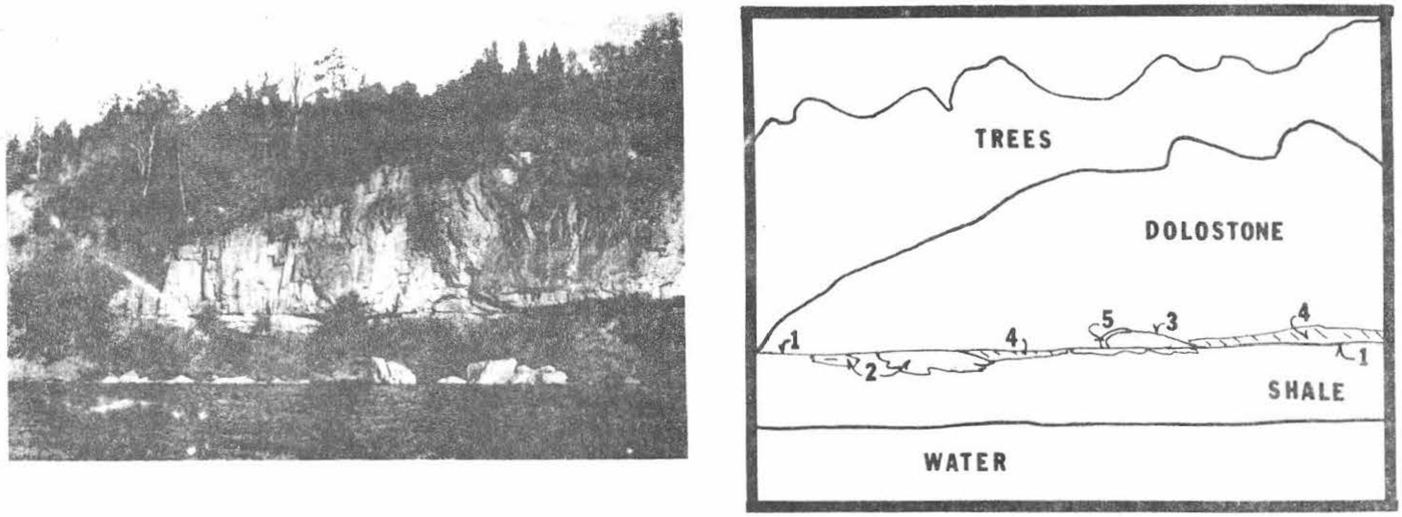

Figure 3. View of the Champlain thrust fault looking east at the southern end of Lone Rock Point (Fig. 1). The accompanying line drawing locates by number the important features discussed in the text: 1, the continuous planar slip surface; 2, limestone slivers; 3, A hollow in the base of the dolostone is filled in with limestone and dolostone breccia; 4, Fault mullions decorate the slip surface at the base of the dolostone; 5, a small dike of shale has been injected between the breccia and the dolostone.

along the interface between the Dunham Dolomite and the Iberville Shale. Where the inner fault is wider by virtue of slivers and irregularities along the basal surface of the Dunham Dolomite, the slip surface is located in the shale, where it forms the chord between these irregularities (Fig. 3). The slip surface represents the surface along which most of the recent motion in the fault zone has occurred. As a consequence, it cuts across all the irregularities in the harder dolostone of the upper plate with the exception of long wave-length corrugations (fault mullions) that parallel the transport direction. As a result, irregular hollows along the base of the Dunham Dolomite are filled in by highly contorted shales and welded breccia (Fig. 3).

The deformation in the shale beneath the fault provides a basis for interpreting the movement and evolution along the Champlain thrust fault. The compositional layering in the shale of the lower plate represents the well-developed S1 pressure-solution cleavage that is essentially parallel to the axial planes of the first-generation of folds in the Ordovician shale exposed below and to the west of the Champlain thrust fault (Fig. 4). As the trace of the thrust fault is approached from the west this cleavage is rotated eastward to shallow dips as a result of westward movement of the upper plate (Fig. 4). Slickenlines, grooves, and prominent fault mullions on the lower surface of the dolostone and in the adjacent shales, where they are not badly deformed by younger events, indicate displacement was along an azimuth of approximately N60°W (Fig. 4; Hawley, 1957; Stanley and Sarkesian, 1972; Leonard, 1985). The S1 cleavage at Lone Rock Point is so well developed in the fault zone that folds in the original bedding are largely destroyed. In a few places, however, isolated hinges are preserved and are seen to plunge eastward or southeastward at low angles (Fig. 4). As these F1 folds are traced westward from the fault zone, their hinges change orientation to the northeast. A similar geometric pattern is seen along smaller faults, which deform S1 cleavage in the Ordovician rocks west of the Champlain thrust fault. These relations suggest that F1 hinges are rotated towards the transport direction as the Champlain thrust fault is approached. The process involved fragmentation of

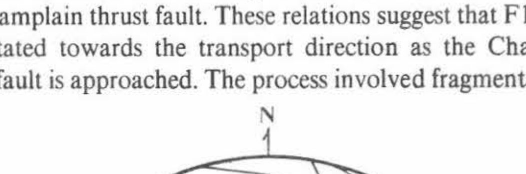

Figure 4. Lower hemisphere equal-area net showing structural elements associated with the Champlain thrust fault. The change in orientation of the thrust surface varies from approximately N20°W to N14°E at Lone Rock Point. The orientation of S1 cleavage directly below the thrust is the average of 40 measurements collected along the length of the exposure. S1, however, dips steeply eastward in the Ordovician rocks to the west of the Champlain thrust fault as seen at South Hero and Clay Point where F1 hinges plunge gently to the northeast. Near the Champlain thrust fault F1 hinges (small circles) plunge to the east. Most slickenlines in the adjacent shale are approximately parallel to the fault mullions shown in the figure.

the FI folds since continuous fold trains are absent near the thrust. Much of this deformation and rotation occurs, however, within 300 ft (100 m) of the thrust surface. Within this same zone the S1 cleavage is folded by a second generation of folds that rarely developed a new cleavage. These hinges also plunge to the east or southeast like the earlier F1 hinges. The direction of transport inferred from the analysis of F2 data is parallel or nearly parallel to the fault mullions along the Champlain thrust fault. Stanley and Sarkesian (1972) suggested that these folds developed during late translation on the thrust with major displacement during and after the development of generation 1 folds. New information, however, suggests that the F2 folds are simply the result of internal adjustment in the shale as the fault zone is deformed by lower duplexes and frontal or lateral ramps (Figs. 1, 2). The critical evidence for this new interpretation is the sense of shear inferred from F2 folds and their relation to the broad undulations mapped in the fault zone as it is traced northward along Lone Rock Point (Fig. 1). South of the position of the thick arrow in Figure 1, the inferred shear is west-over-east whereas north of the arrow it is east-over-west. The shear direction therefore changes across the axis of the undulation (marked by the arrow) as it should for a synclinal fold.

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