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A TRANSECT THROUGH THE FORELAND AND TRANSITIONAL ZONE OF 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 base 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; in press) and her students (Gregory, 1982; Myrow, 1983; Teetsel, 1985; Bulter, 1986; MacLean, 1986). 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. 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 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

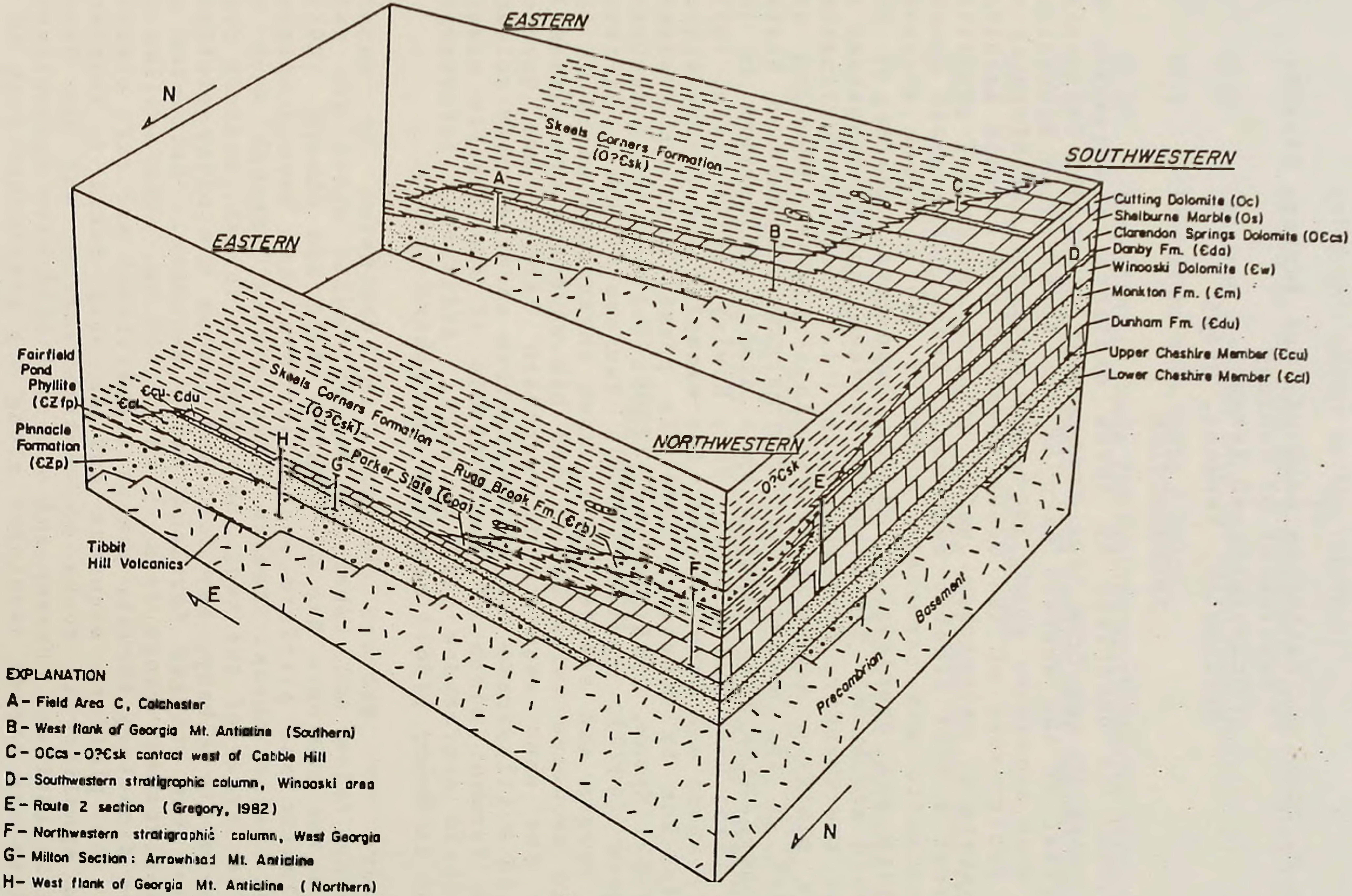
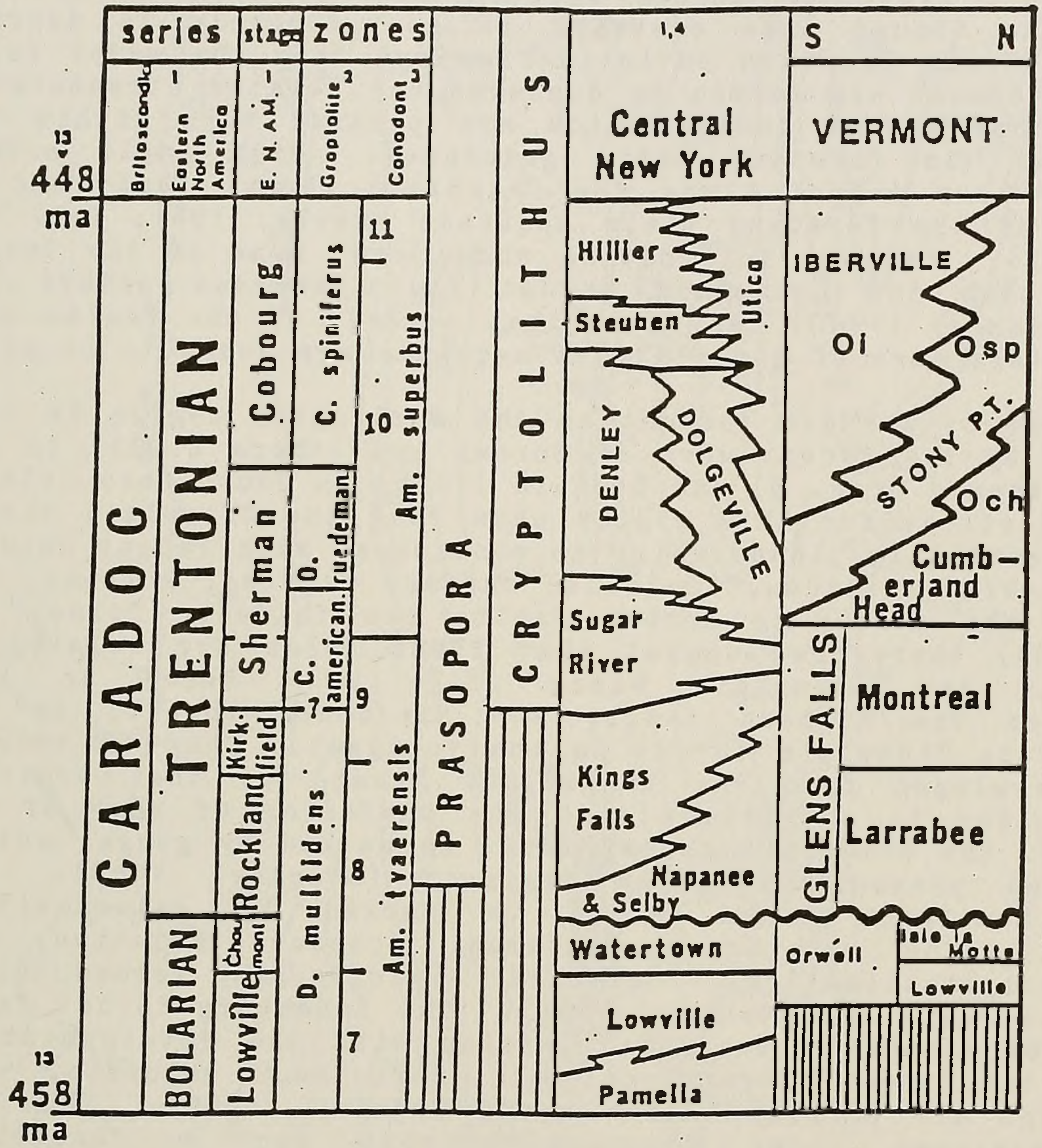


Figure 1 - Diagrammatic representation of the original depositional protolith for the Milton quadrangle reconstructed from stratigraphic information in the indicated areas.

Dorsey and others (1983)



MacLean 1986

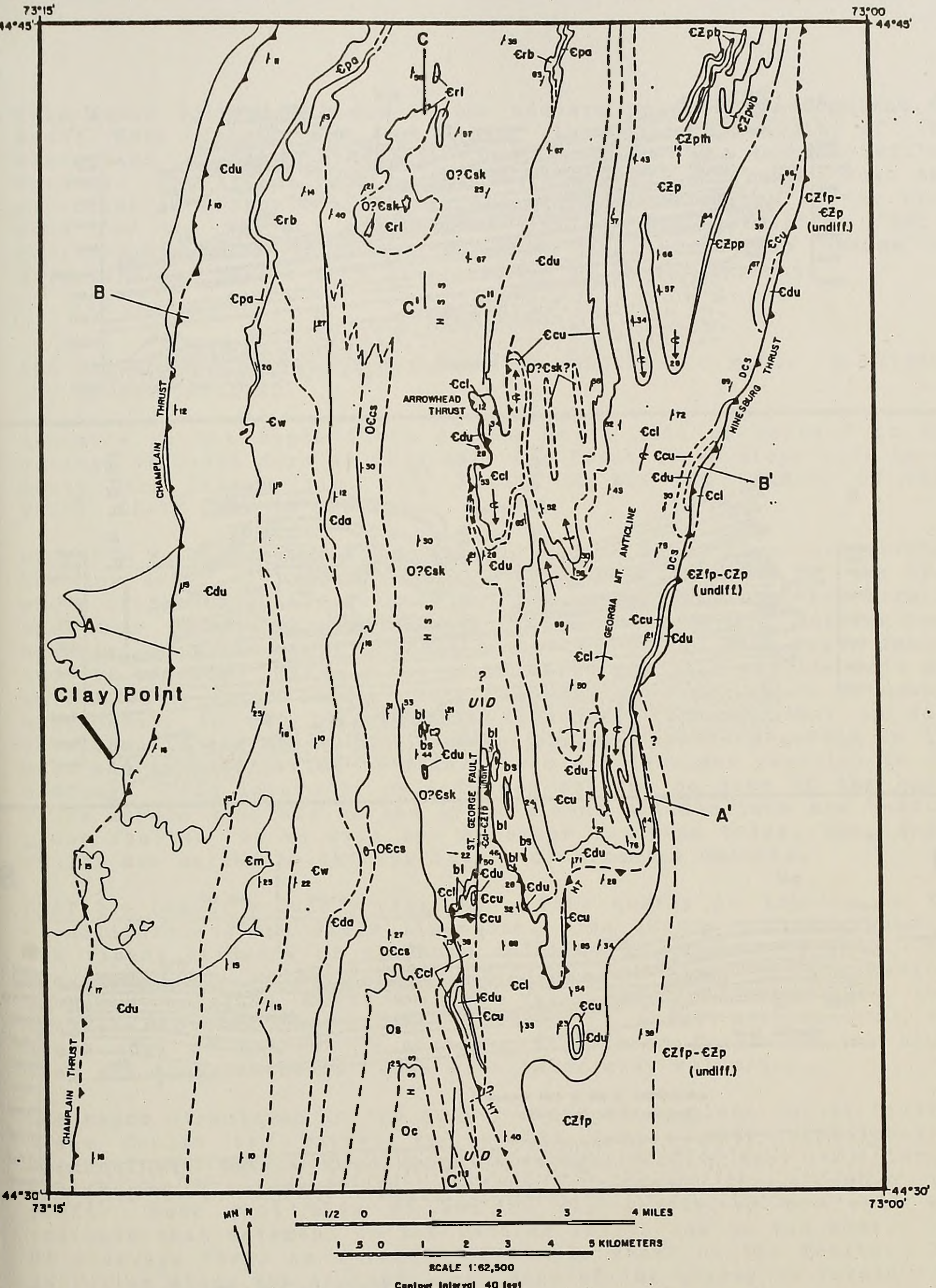
Figure 2 - Stratigraphic column for the western Vermont and Central New York. For further stratigraphic discussion read Erwin (1957) and Hawley (1957, 1972).

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. This conclusion is 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). The Arrowhead Mountain thrust fault is marked by cataclasites, dislocation glide and creep features, pressure solution, and limited recrystallization. Oriented sericite is formed along cleavages (Strehle and Stanley, 1986). The Hinesburg thrust fault is marked by extensive recrystallization with the development of protomylonites and ultramylonites. Sericite, chlorite, and stilpnomelane are present (Strehle and Stanley, 1986). Thus the fabrics developed under progressive more ductile conditions generated at higher temperatures and pressures. 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.

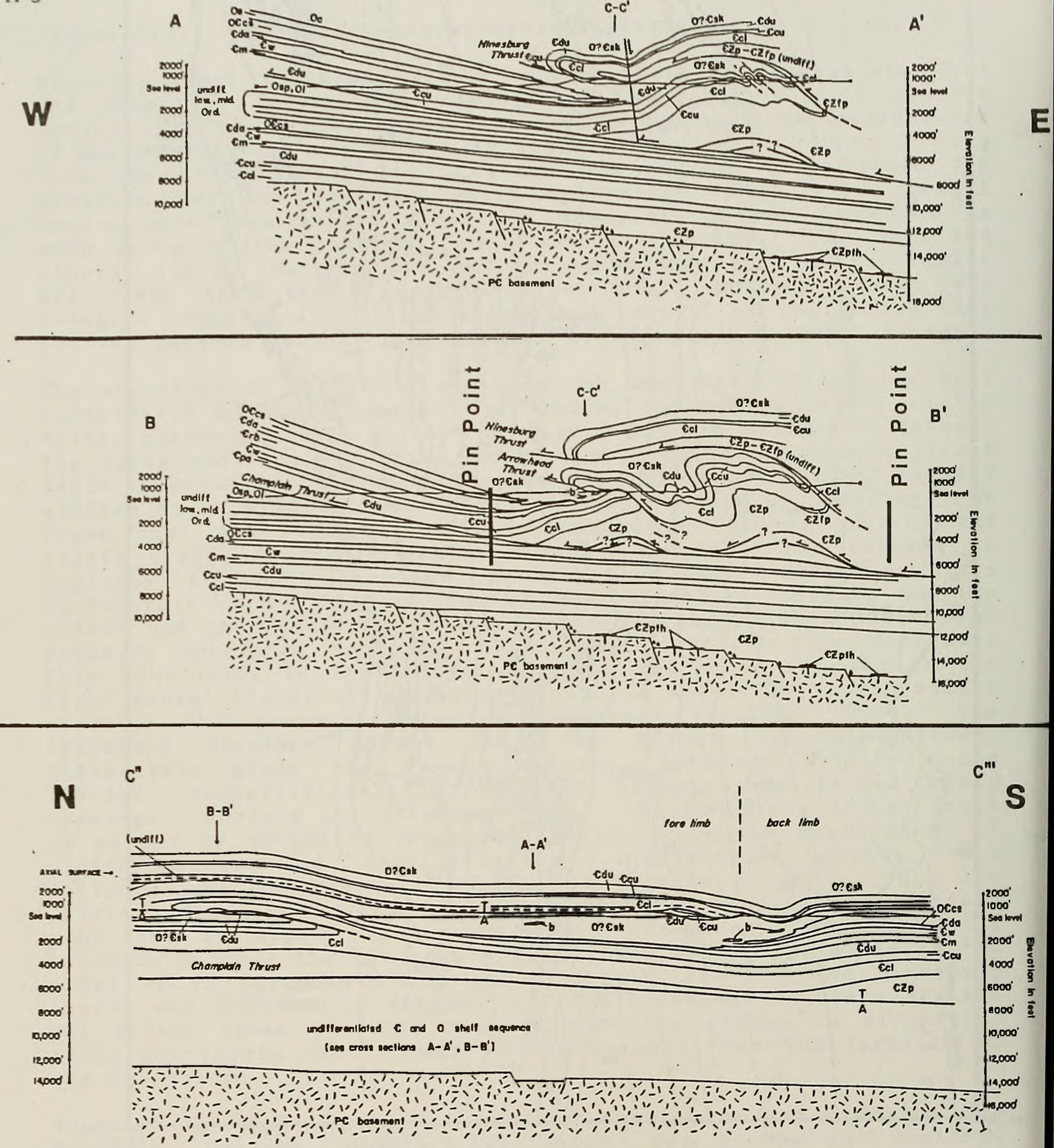
Most of the deformation in western Vermont occurred during the Taconian Orogeny. This conclusion is based on regional considerations (Stanley and Ratcliffe, 1985) and the analysis of available isotopic data (Sutter and others, 1985). Mesozoic deformation in the form of extensional faults and igneous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



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

This field trip will begin in the western part of the foreland in South Hero Island and end along the western margin of the hinterland at the Hinesburg thrust fault at Mechanicsville, Vermont. The trip across central Vermont will continue across the pre-Silurian hinterland through the northern extension of the root zone for the Taconic allochthons Stanley and Ratcliffe, 1985). Participants will therefore be able to observe the change in structure and fabric from the foreland to the hinterland.

ITINERARY

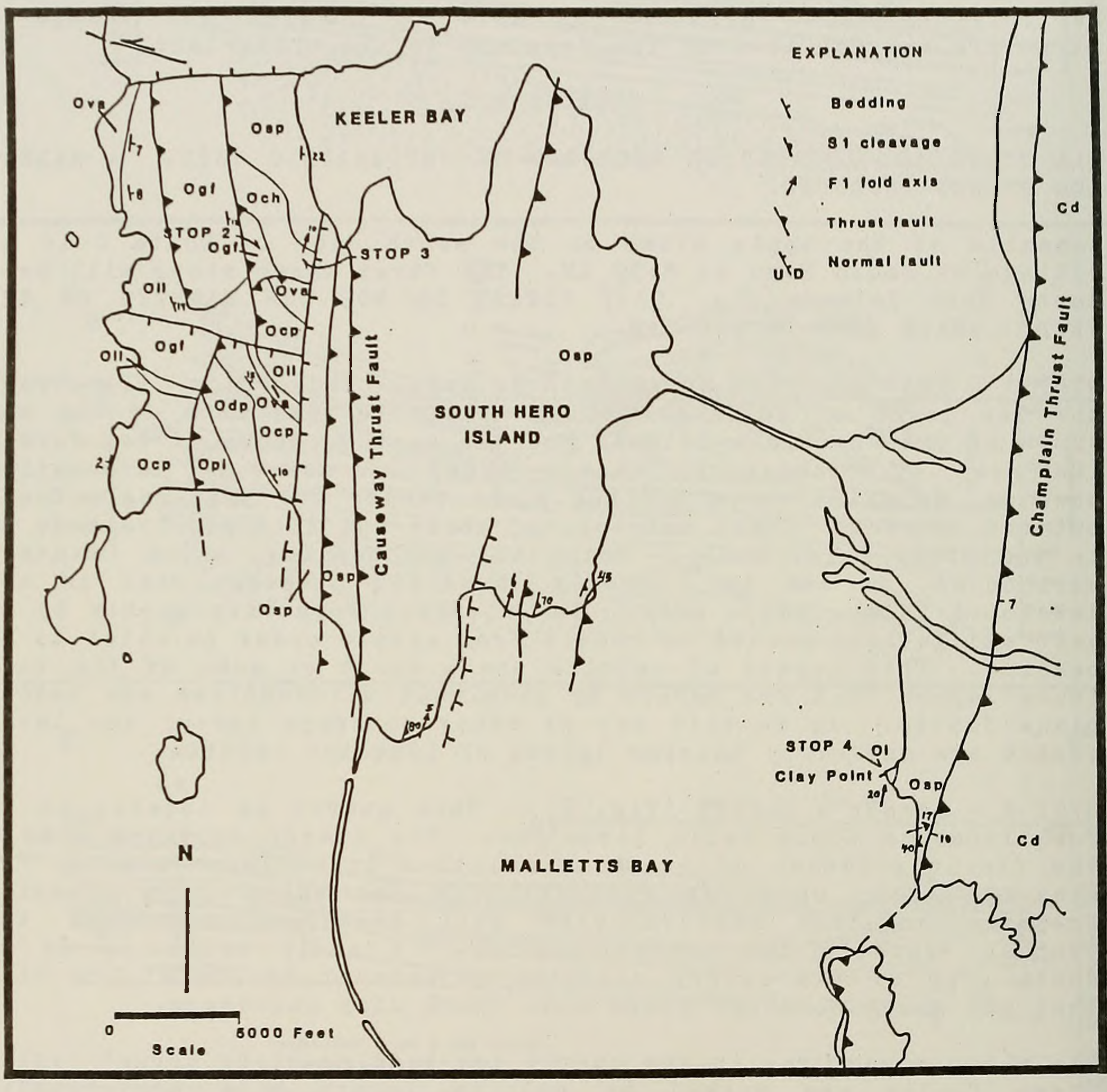
ALL STOPS ARE LOCATED ON GEOLOGIC OR TOPOGRAPHIC MAPS. A MILEAGE LOG IS NOT INCLUDED.

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.

<u>STOP 1 - West Shore of South Hero Island.</u> 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.

<u>STOP 2 - Lessor's Quarry (fig. 6)</u> - 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.

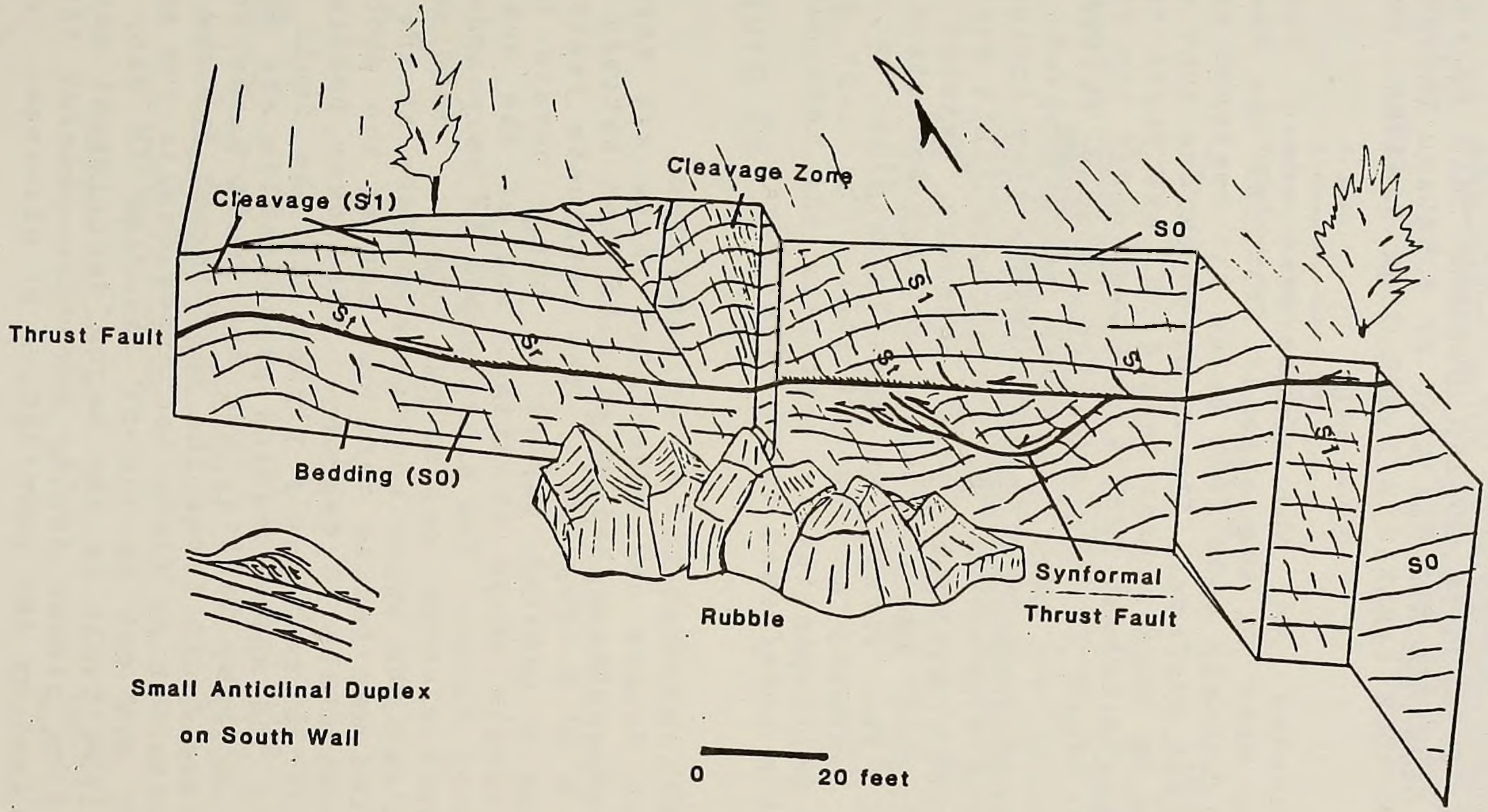
The major structures in the quarry are bedding-plane thrust faults. These faults are marked by calcite layers with west-trending slickenlines and a fault-zone cleavage (St). Near the larger faults the S1 cleavage is rotated (Sr) toward the plane of the fault. Note that both St and Sr dip gently to the 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.



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

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.

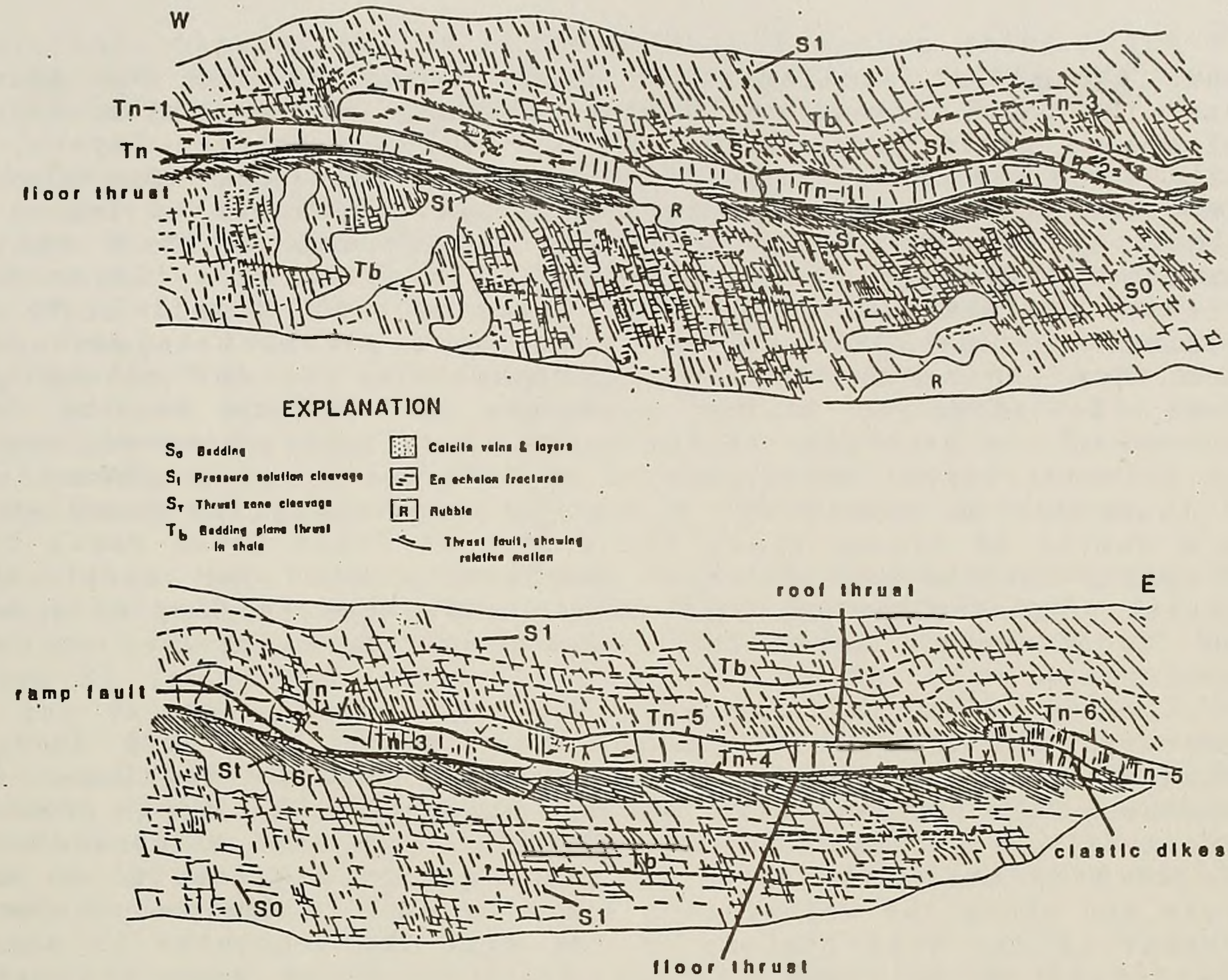
STOP 3 - "THE BEAM" - THIS IS A SUPERB OUTCROP THAT SERVES AS A FIELD LABORATORY FOR RESEARCH AND TEACHING OF FORELAND DEFORMATION. PLEASE STUDY IT. USE YOUR CAMERAS BUT NOT YOUR HAMMERS. REFER TO FIGURES 7 AND 8.

The outcrop is located in the Cumberland Head Formation approximately 5 miles west of the exposed front of the Champlain thrust fault or approximately 4600 feet 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. 8).

THE OUTCROP

Five imbricate thrust faults and associated ramps are exposed in profile section in a foot thick bed of micrite that extends 45 feet 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 8 feet of flats. The micrite bed is surrounded by at least 5 feet 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 the fault zone is thick. In comparsion the upper bedding plane faults 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 and are offset by the penetrative S1 cleavage.

All the fault surfaces are covered by layers of sparry calcite that vary in thickness from several mm to 4-6 cm. The thickest zone is

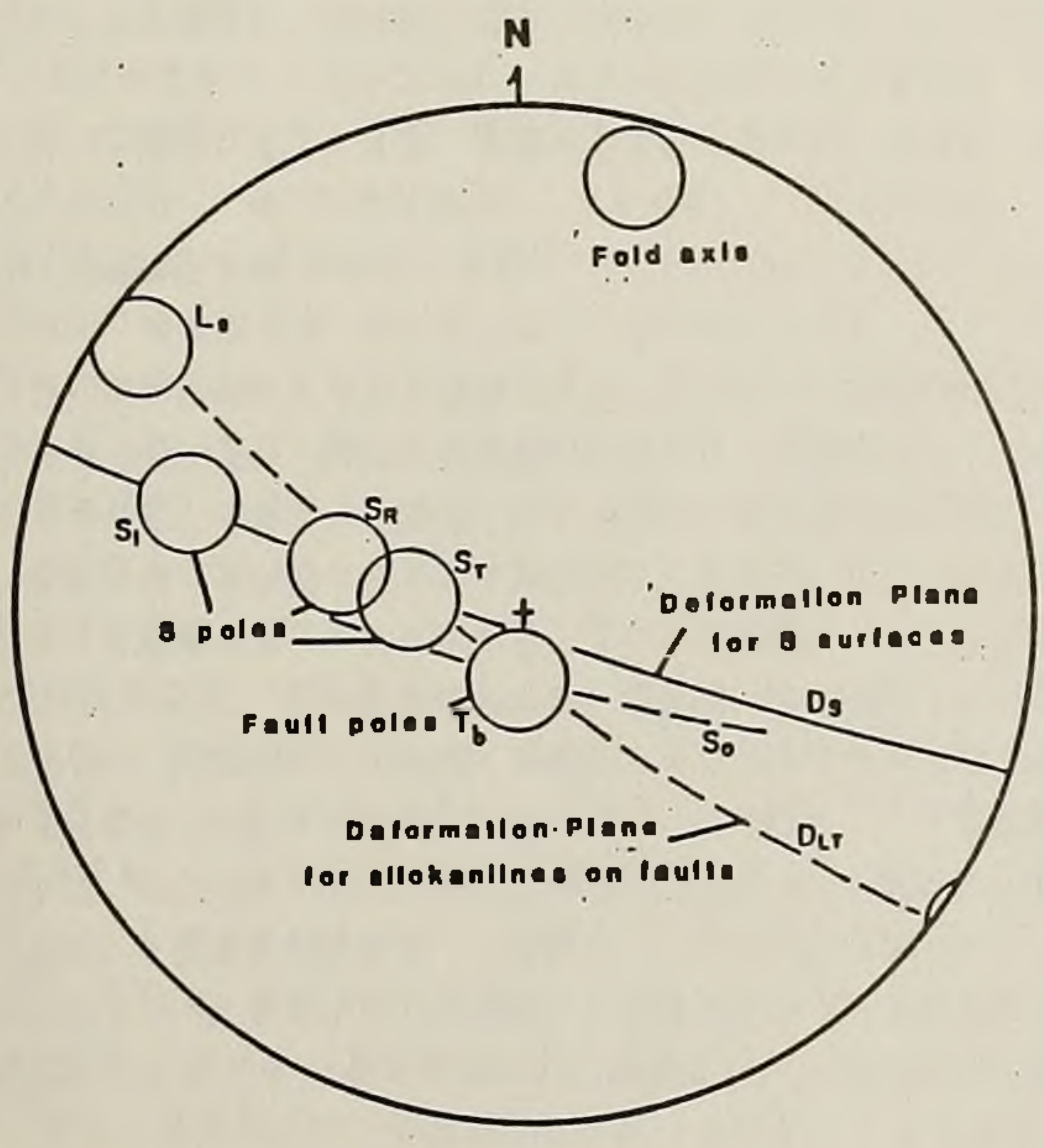


Stanley 1985

Figure 7a

DEFORMATION OF THE CUMBERLAND HEAD FORMATION

SOUTH HERO, VERMONT



Lower hemisphere equal area projection showing the dominant position of the major structural elements in the outcrop of the "beam". The structural elements are identified in figure 7a. Ls refers to the dominant orientation of slickenlines on the thrust faults. The slickenlines on the older thrust faults are rotated along the deformation plane for slickenlines.

Figure 7b

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 (fig. 7b). In sections oriented perpendicular to the layering and parallel to the slickenlines the shale selvedges are more parallel to each other than they are 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 5 feet 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 (figs 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. S1 surfaces are covered by a black, carbon-rich selvedge of illite and kaolinite which is less than two tenth of an inch 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 is gradually 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 2.2 inches 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 a mm thick), more closely spaced (about a 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.

OTHER IMPORTANT RELATIONS

Ramp Faults

All the east-dipping ramp faults form from arrays of west-climbing, en echelon fractures. These arrays appear to nucleate along the west-facing limbs of slightly asymmetrical buckle folds. The sequence continues with the rotation of the individual fractures by distributed shear strain along the west-climbing array. The individual fracture may continue to grow as the older central parts of the fractures are rotated counterclockwise to produce S shaped fractures. New arrays of planar fractures develop over the older arrays and produce a weakened zone along the trend of the array. As generations of fractures are superposed they coalesce and develop into ramp faults as demonstrated by deformed fracture arrays in the footwall and hangwall that 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 are abundant in the ramp regions.

Fault Chronology

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.

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. Second, the floor fault is continually reactivated during the evolution of the fault system. The average age of this fault zone therefore becomes younger to the west. Furthermore, it is the most continuous fault in the outcrop. This second conclusion explains why the fault zone for the floor fault is much 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.

Fault-Zone Deposits

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.

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 5 or 10 feet 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) The slickenlines on each of the vein layers are formed by grooves rather than calcite fibers.

6) 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 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. We suggest that preferrential 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 was 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

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 3 to 19 inches which adds up to a total displacement of 52.8 inches or 4.4 feet over a present horizontal distance of 35 feet. The five anticlines over ramps and the broad folds along the flats account for approximately 5 inches of additional shortening so that the total shortening equal 57.8 inches or 4.8 feet. These values correspond to a shortening of 13.7 percent.

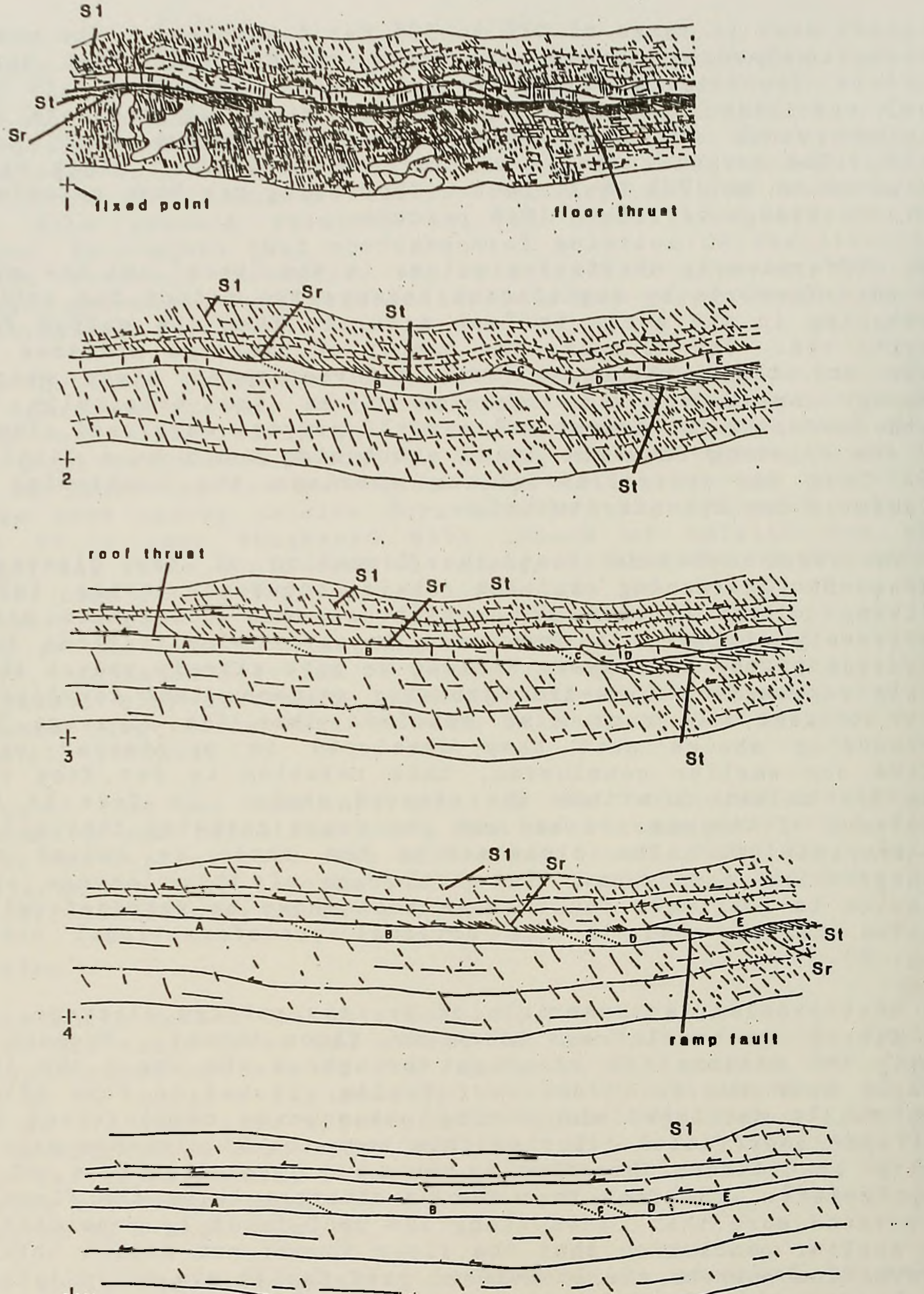
In the shale the corresponding shortening is provided by volume reduction across the cleavage surfaces. In order to see if the shortening determined from the micrite bed is comparable to the shortening in the shale, an independent estimate was made for the shale by determining the percentage of insoluble 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 across a present width of 26 feet was then counted. The total width of the selvedges (a range of 1.16 to 1.78 foot) was then multiplied by 2.8 to given an estimate of the original width now represented by the cleavages (3.25 to 4.98 feet). The original length of the present 26 foot width was then estimated to be 29.3 to 31 feet. Shortening was then calculated to be in the range of 7.3 to 10.6 percent.

This difference in shortening values in the "beam" and the shale is not considered to be significant because the method for estimating shortening in the shale is less accurate than the method for the micrite bed. For example, the number of cleavage surfaces in the shale and their thickness were underestimated since many thin cleavage selvedges were overlooked in the originally count. Furthermore, if the lowest value of 32 percent were used along with all the existing data the total shortening would be a little over 14%. Thus the range of values overlaps the shortening value calculated for the micrite bed.

We 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 we have already proven that the fsults represent a time-transgressive sequence that developed from east to west, we must also conclude that the cleavage in the surrounding shales must have developed in a similar mannner. Unlike our earlier conclusion, this relation is far from obvious from the relations within the cleaved shale. In fact it is the existence of the micrite bed and its fault geometry that allows us to conclude that the cleavage in the shale is indeed a time transgressive phenomena. The evolution of the cleavage and its relation to imbricate faulting is best shown by retrodeforming the faulted micrite bed to its original predeformational condition (fig. 8).

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 throughout the shale and is only present near the floor and roof faults, it had to form after the ramp faults developed and during subsequence displacement on the roof and floor faults. During this time the S1 cleavage near these faults is rotated in simple shear in the direction of fault displacement. The fact that the Sr cleavage below the floor fault is rotated more than it is along the roof fault is consistent with our earlier conclusion that the floor thrust was active throughout deformation whereas the individual roof faults are short lived.

The St cleavage, which is only present along the floor thrust, clearly formed after Sr because it cuts across Sr at a low angle and is not folded or rotated. Its absence along the roof faults is consistent with their short history of displacement. St is a true fault zone cleavage because it is restricted to a thin region below



RETRODEFORMED SECTION OF THE CUMBERLAND HEAD FORMATION

SOUTH HERO, VERMONT

Figure 8

Stanley, 1987

the floor fault. Furthermore, the St cleavage is in the orientation predicted by the Ramsay equation Tan 2θ'=2/gamma (Ramsay and Huber, 1984) where gamma of 36 degrees is the rotation of S1 (Sr) as it is traced into the fault zone. This relation proves that St is the result of simple shear along the floor fault.

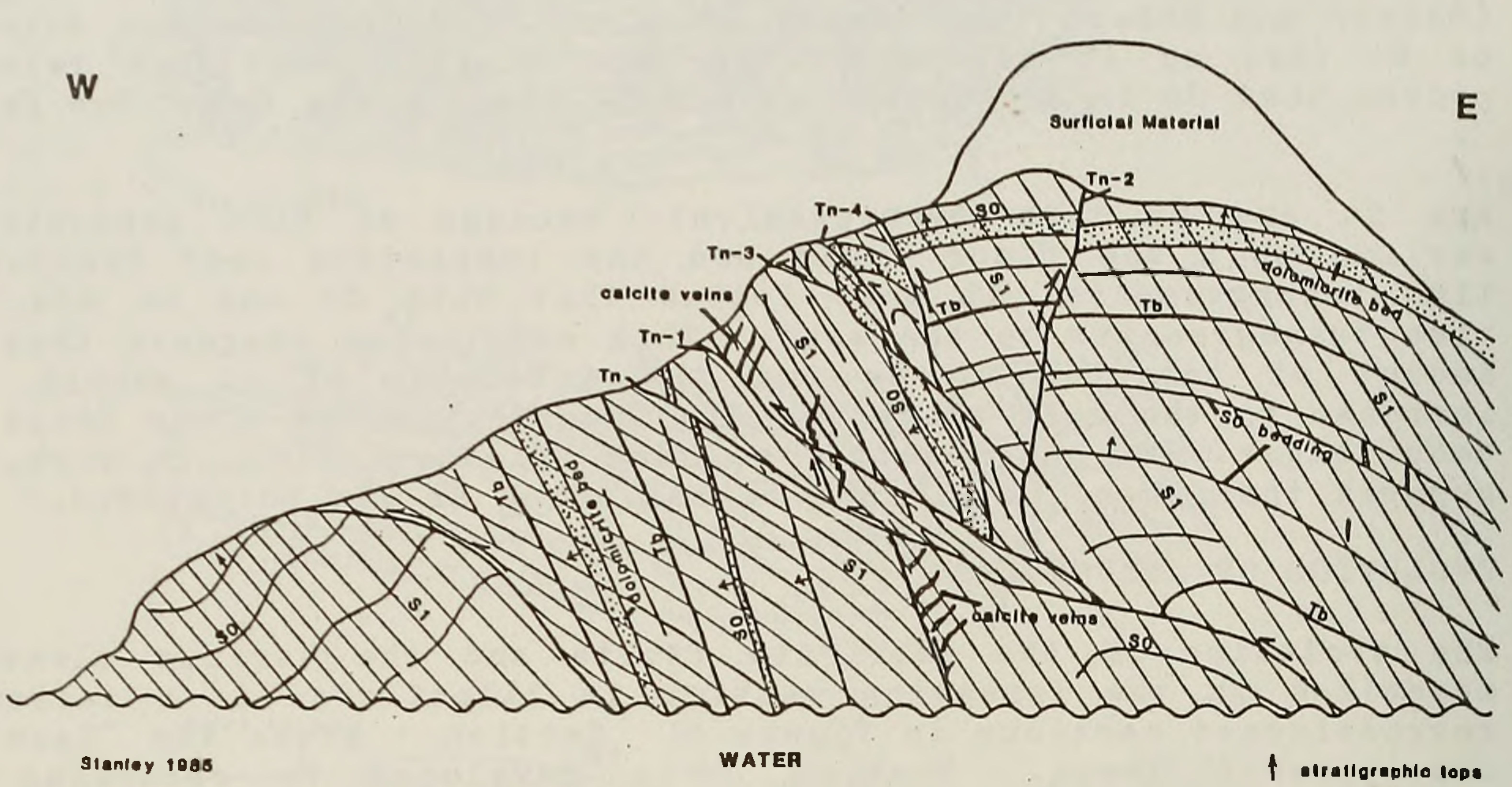
Are St and Sr time transgessive? Because we have demonstrated earlier that the floor fault and the respective roof faults are time transgressive, it must follow that both Sr and St are also time transgressive to the west. This conclusion suggests that the amount of rotation of Sr and the intensity of St should also increase to the east where the displacement on the floor fault has been longer. We could detect no such relation which, in turn, may suggest that there is a limit beyond which Sr can be rotated.

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 8. Section 1 shows the "beam" 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. For example, the rocks of the hanging wall block (B, section 2) are unfolded as they are returned to their original position east of the footwall block (A, section 2). During the time represented by section 2 the active floor fault in the eastern part of the diagram climbs section along the ramp fault below block C and continues along the top of block A and B. As a result the St and Sr cleavages below blocks A and B are absent. Furthermore, the S1 cleavage is shown to be more abundant in the upper plate than the lower plate because it is actively moving.

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.

STOP 4 - CLAY POINT (fig. 9) - Clay Point is located several 1000 feet west of the trace of the Champlain thrust fault (fig. 3 and 5). The rocks consists of medium to dark gray, noncalcareous shales, dolomitic siltstones and beds of brown-weathered dolomicrite of the Iberville Formation, a foreland basin flysch deposit. This sequence is rhythmically layered with the base of each cycle marked by yellowish-brown weathered, dark gray laminated siltstone which contains ripple laminations (Hawley, 1972). These beds grade upward into dark grey, well cleaved shale that are marked by thin layers of lineated calcite. These layers represent bedding plane faults (Tb, fig. 9a). which are identical to the bedding plane faults at the "beam". Like the bedding, these faults are folded into a west-facing, overturned anticline that is cut by a prominant, east-dipping, pressure solution cleavage (S1, fig. 9b). This cleavage has the same orientation as the S1 cleavage at the "beam" (STOP 3) and is part of the same generation although it



DEFORMED IBERVILLE FORMATION AT CLAY POINT, VERMONT

Figure 9a

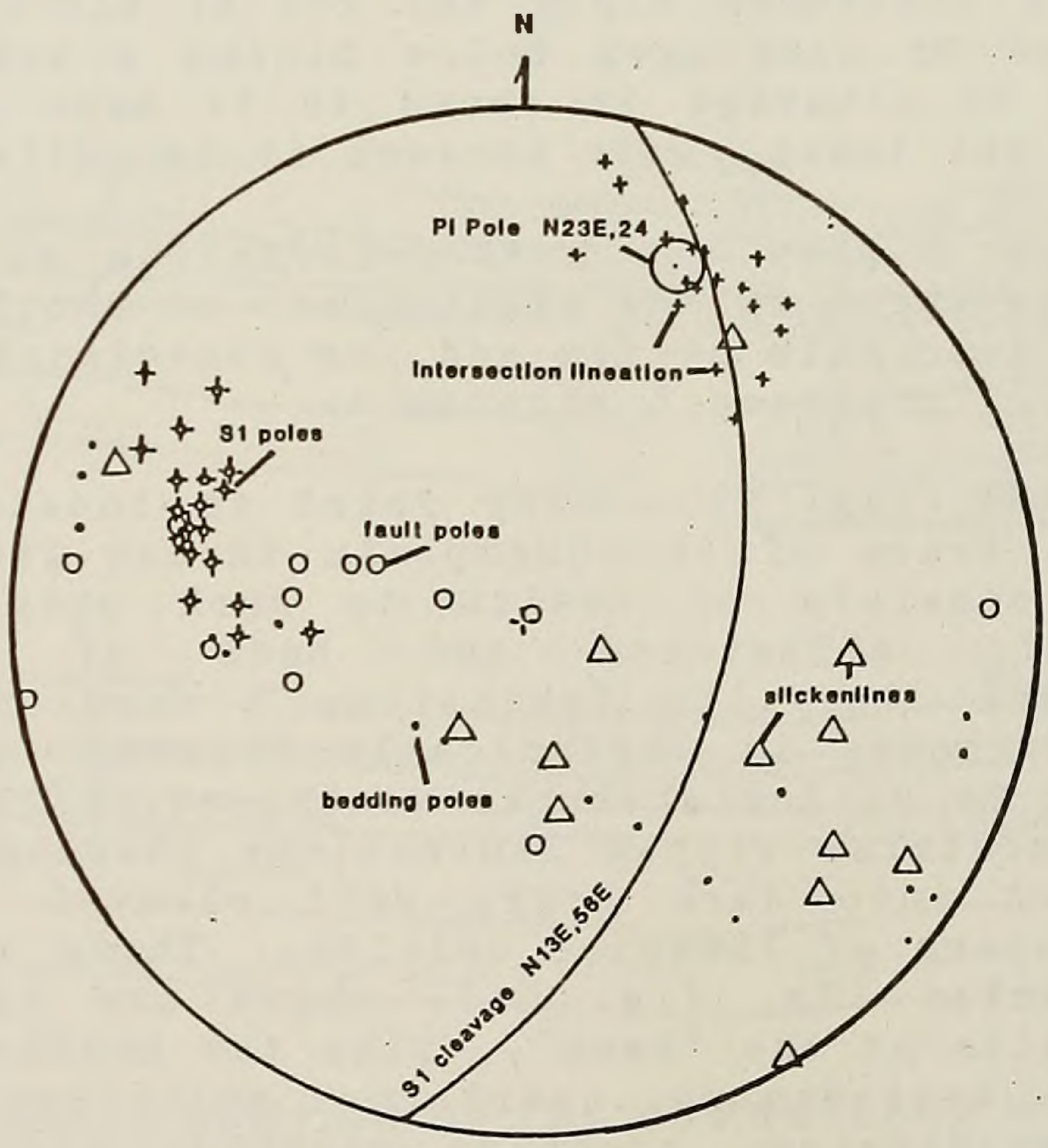


Figure 9b - Lower hemisphere equal area projection showing the orientation of structural elements at Clay Point.

formed earlier because it is situated farther to the east and closer to the Champlain thrust fault. The overturned limb of the anticline is cut by a series of thrust faults (Tn through Tn-4) that are clearly influenced more by the cleavage than they are by the bedding. The sense of displacement on these faults is provided by the rotated S1 cleavage adjacent to the fault surface. We believe that faults of this type that are controlled by the cleavage are younger than the faults seen at the "beam". This relation can be seen at the east end of the "beam" where a younger fault has developed from the older ramp and is climbing through the upper plate shale.

The other features that you should study are:

1) Calcite fractures (fig 9a). Many of these fractures contain cross fibers and indicate slow extension parallel to the fiber direction. Several of the veins are compound with an outer border of fiber calcite and an inner core of sparry calcite, which indicates rapid extension. These veins therefore opened more rapidly after an earlier period of slow extension.

2) Several of the faults on western side of the outcrop have two directions of slickenlines. Although this fact suggests a change in fault motion during the evolution of the structure at Clay Point, another explanation is suggested by the fact that many of these faults are bedding plane faults (Tb). During folding the older northwesterly slickenline direction was rotated out of the plane containing the transport direction (deformation plane). Subsequent movement along the same northwesterly direction produced the new slickenlines which cut across the older direction.

3) A west-trending Mesozoic dike cut the anticline just north of the cross section.

4) Study the laminate calcite along the faults. Look at the fabric perpendicualr and parallel to the slickenline direction.

<u>STOP 5 - 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.

<u>STOP 6 - THE HINESBURG THRUST FAULT AT HINESBURG, VERMONT</u> - This is the classic and best exposed locality for the Hinesburg thrust fault. 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 Vermont Geology No.3). This publication also contains analysis of other fault zones of western Vermont which will be seen during this NEIGC. 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.

The following features should be studied here:

1) The change in fabric as the fault surface is approached. The quartzite grades from a protomylonite to an ultramylonite along the fault surface.

2) The presence of east-over-west asymmetrical folds. These folds are related to simple shear along the fault.

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

4) "Z" shaped quartz veins that are associated with beds of quartzite.

5) Rare east-dipping shear bands.

6) Late fractures and associated en echelon fracture arrays.

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

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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).

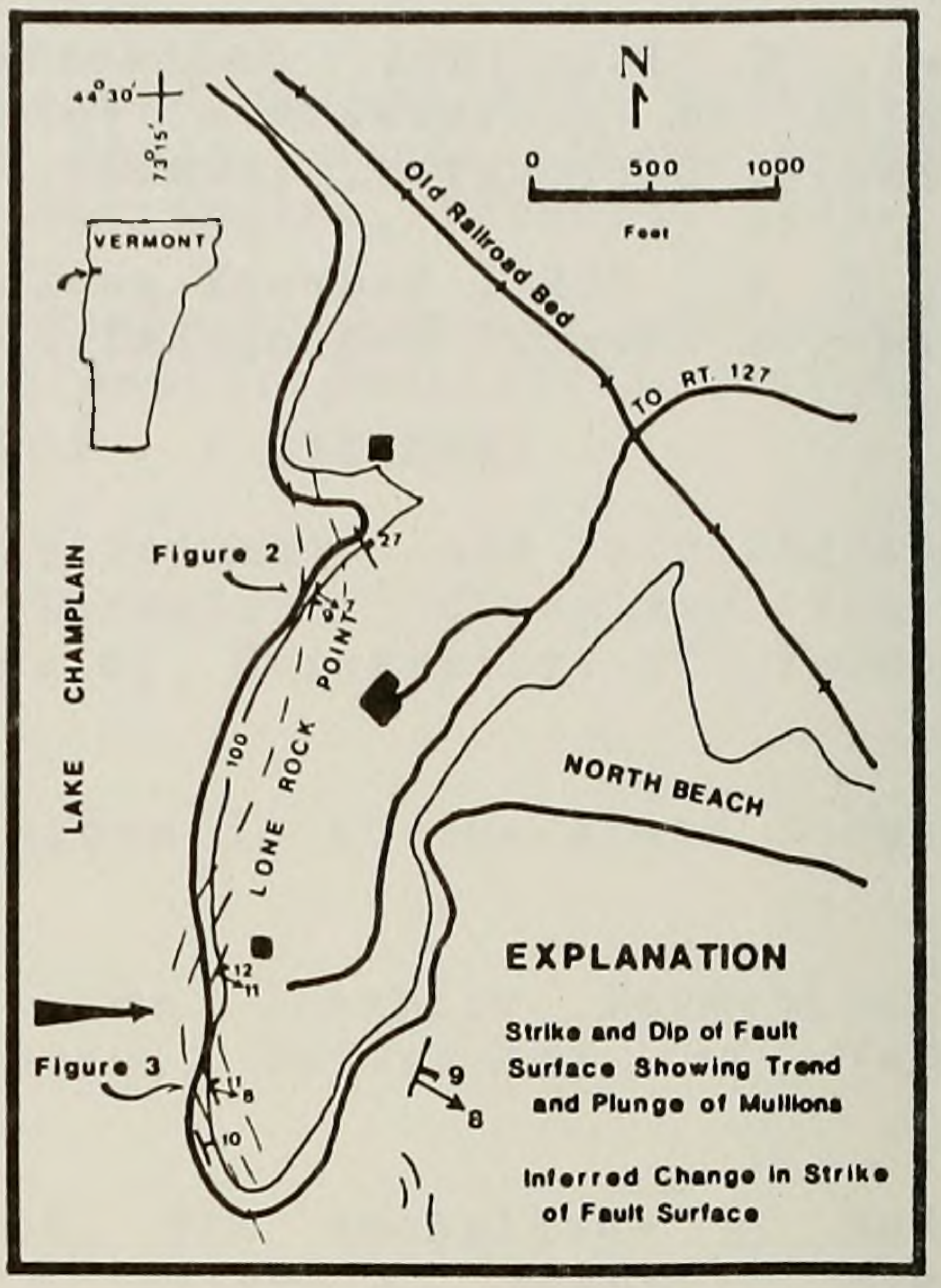

Figure 1. Location map of the Champlain thrust fault at Lone Rock Point, Burlington, Vermont. The buildings belong to the Episcopal Diocesan Center. The road leads to Institute Road and Vermont 127 (North Avenue). The inferred change in orientation of the fault surface is based on measured orientations 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

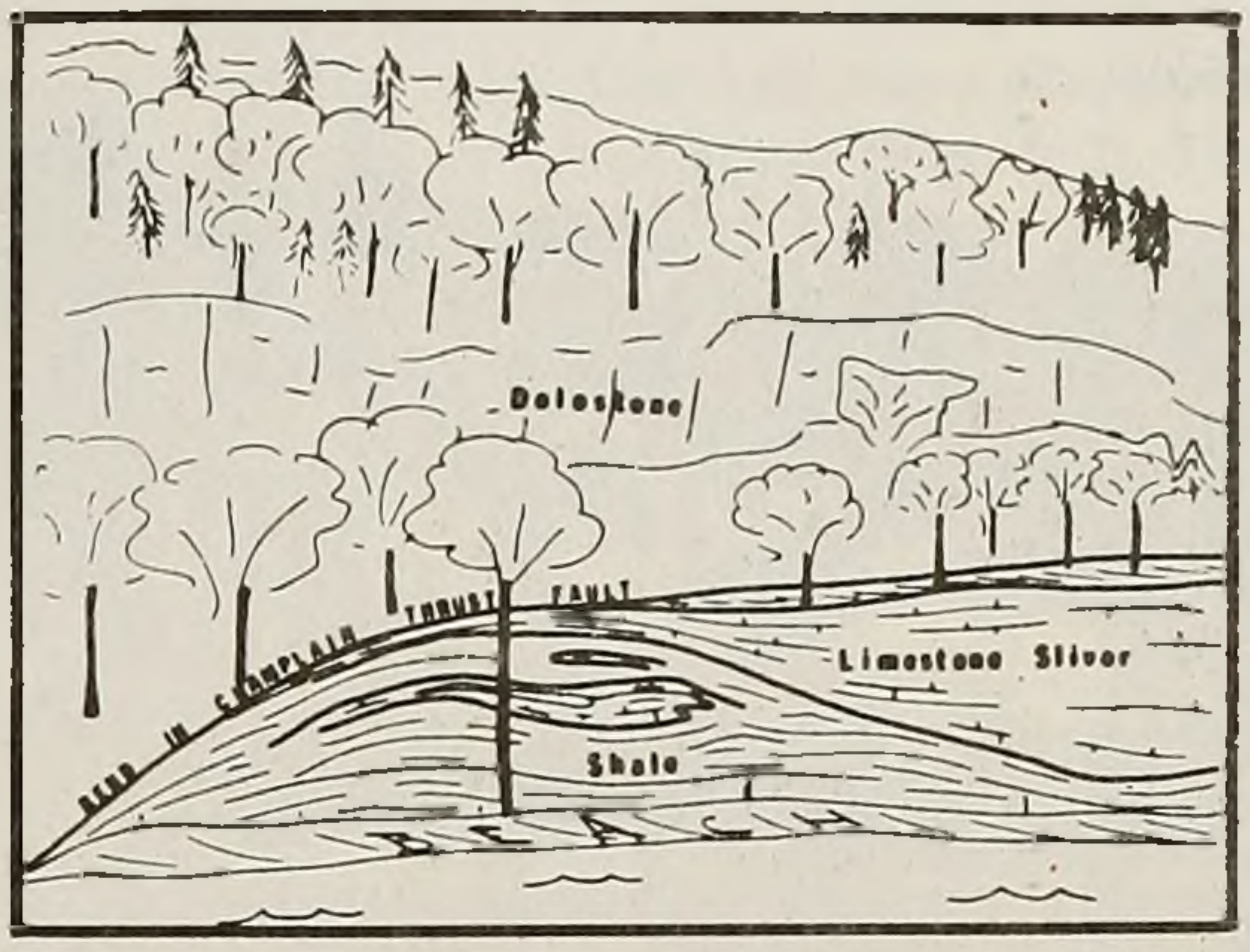

Figure 2. A sketch of the Champlain thrust fault at the north end of Lone Rock Point showing the large bend in the fault zone and the slivers of Lower Ordovician limestone. The layering in the shale is the S1 cleavage. It is folded by small folds and cut by many generations of calcite veins and faults.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

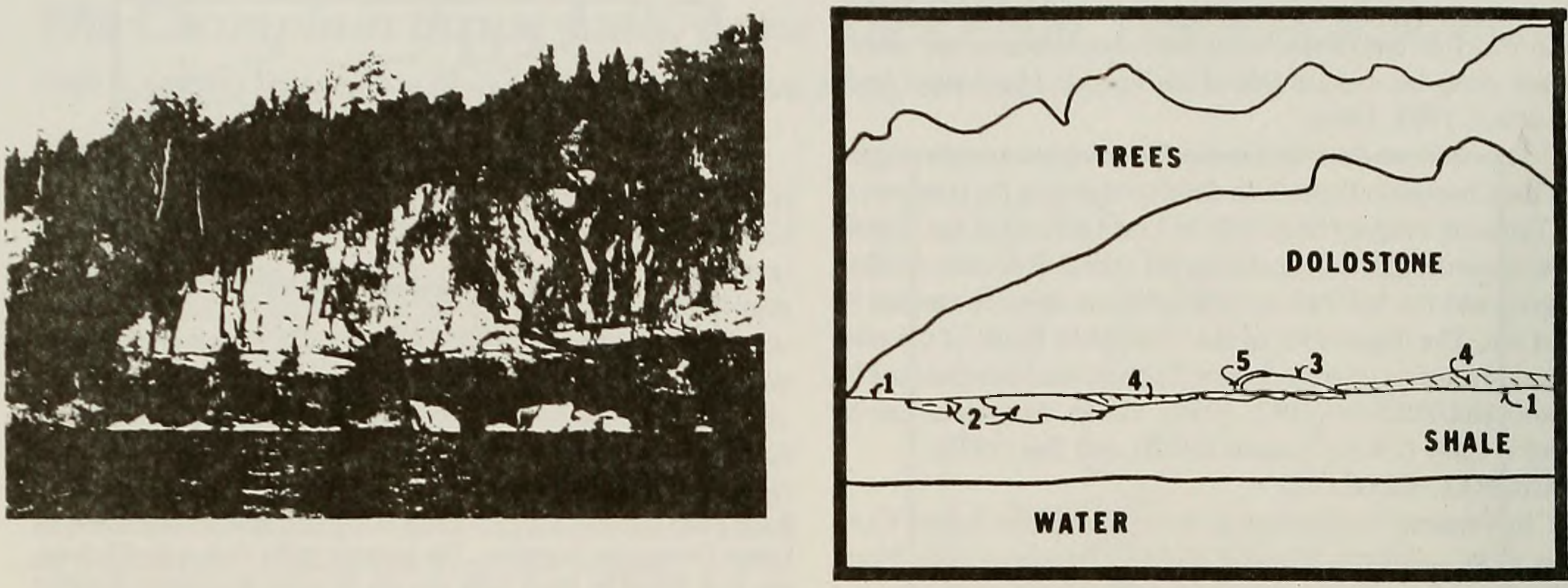

Figure 3. View of the Champlain thrust fault looking east at the southern end of Lone Rock Point (Fig. 1). The accompanying line drawing locates by number the important features discussed in the text: 1, the continuous planar slip surface; 2, limestone slivers; 3, A hollow in the base of the dolostone 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

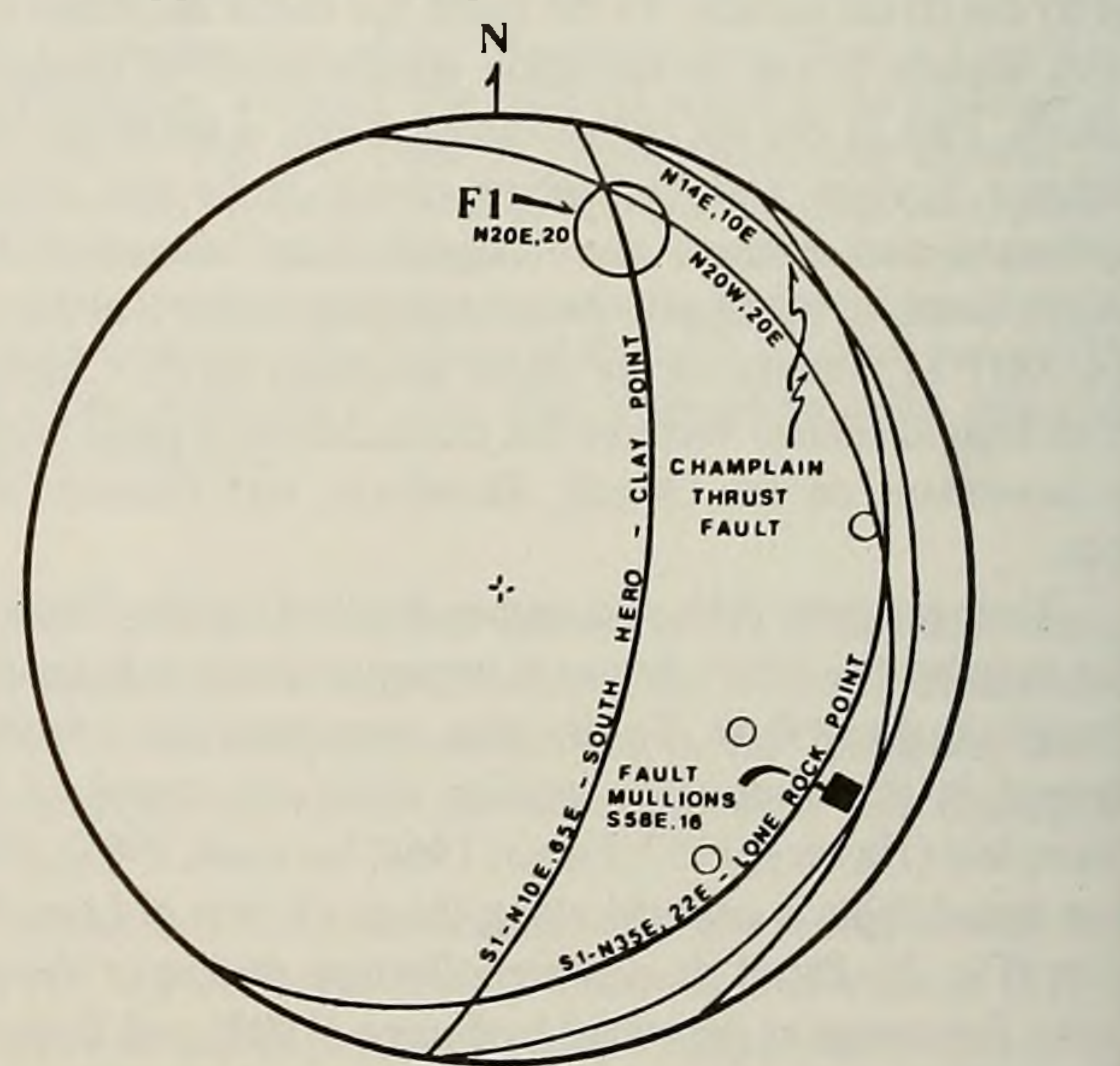

Figure 4. Lower hemisphere equal-area net showing structural elements associated with the Champlain thrust fault. The change in orientation of the thrust surface varies from approximately N20°W to N14°E at Lone Rock Point. The orientation of S1 cleavage directly below the thrust is the average of 40 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 Fl 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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