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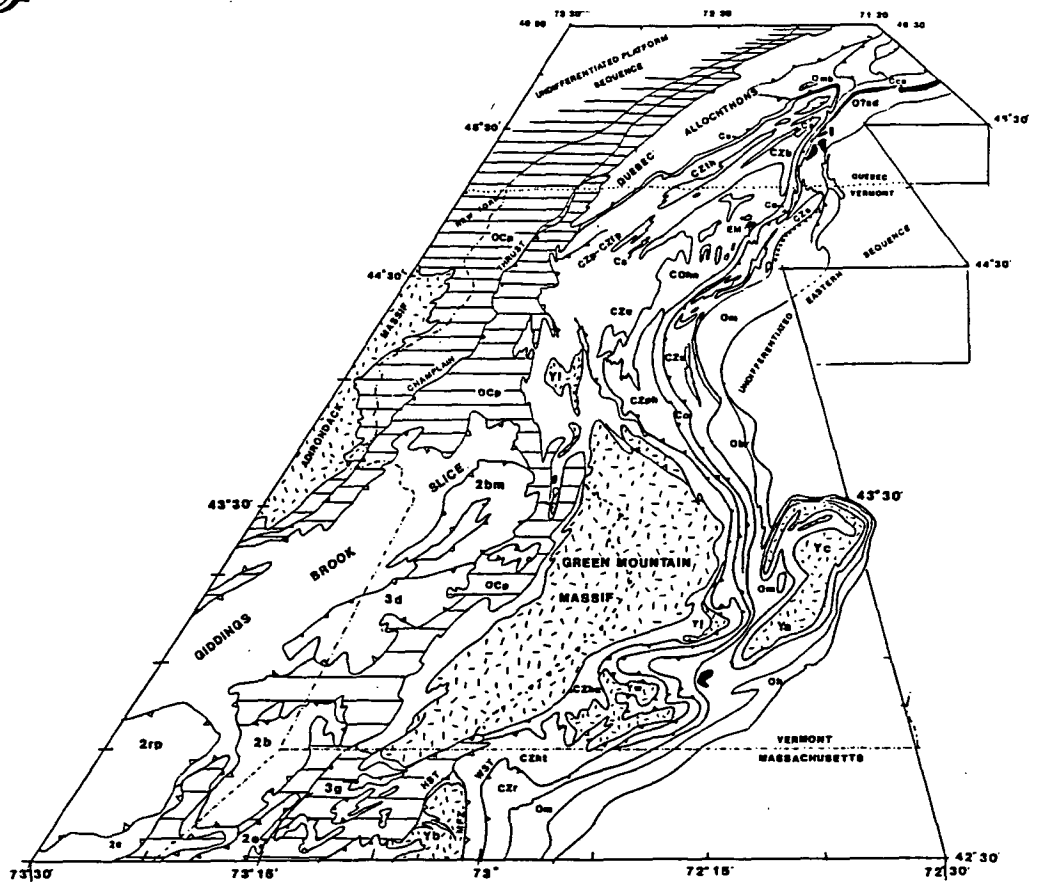

APPALACHIAN WORKSHOP

1989

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APPALACHIAN WORKSHOP

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Cover: *Perspective block diagram of the Québec-Vermont Orogen (from Doolan, 1989; see Doolan, this volume).*

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PROGRAM AND CONTENTS

SATURDAY, APRIL 15, 1989

EVOLUTION OF THE ANCIENT NORTH AMERICAN MARGIN IN THE QUEBEC REENTRANT

Chairperson:

Charles A. Ratté

GEOCHRONOLOGIC CONSTRAINTS ON THE TIMING OF RIFTING AND CONTINENTAL BREAKUP AT THE SUTTON MOUNTAINS TRIPLE JUNCTION

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The geological evolution of a segment of the Appalachian orogen in southeastern Québec and northern Vermont has been influenced by an Iapetan RRR triple junction - the Sutton Mountains triple junction - and its vestiges (Kumarapeli, 1985). The prominent bend of the orogen, commonly referred to as the Sutton Mountains salient or the Québec reentrant, is the most striking feature inherited from the triple junction.

The biostratigraphic and geochronologic evidence from the general region does not provide tight time constraints on the problems of rift initiation and rift-drift transition. No fossils have yet been discovered from the lower part of the Oak Hill Group (or its equivalent in Vermont, the Camels Hump Group) which appears to be made up of synrift sedimentary and volcanic assemblages. However, the Gilman and Dunham Formations, probable syn- and/or post-breakup deposits higher up in the Oak Hill sequence (see Dowling and others, 1987) contain fragments of Early Cambrian brachiopods and trilobites (Clark, 1936). Thus, the synrift part of the Oak Hill Group (and hence the initiation of rifting) was considered to have been formed in the Early Cambrian or earlier in the latest Proterozoic.

An important geochronologic constraint on the problems is provided by the 565 Ma ages of alkalic-carbonatitic complexes (Gittins and others, 1967) emplaced at the distal end of the Ottawa graben which is the failed arm of the Sutton Mountains triple junction. A similar age of 573±32 Ma (K-Ar whole-rock) has been obtained from a group of small outliers (in Grenvillian rocks) of trachyandesitic lava flows, also of probable synrift origin, located close to the northern boundary of the Ottawa graben in Buckingham area, Québec (Lafleur and Hogarth, 1981). A reliable age for the tholeiitic Grenville dike swarm, which appears to radiate westwards from the Sutton Mountains triple junction into the craton, also could provide an important time constraint on the problems. However, its age is not known with any precision. The age of 575 Ma, quoted in the most recent literature (Fahrig and West, 1986), seems to be an estimate based on a wide range (741±71 to 415±70 Ma; Kretz and others, 1985) of available K-Ar dates.

Another important geological constraint is provided by the recently acquired age of the metavolcanic rocks of the Tibbit Hill Formation which constitutes the lowest exposed part of the Oak Hill Group (in northern Vermont, the Tibbit Hill metavolcanics are interbedded with greywackes). The Tibbit Hill Formation consists predominantly of greenstones derived from Within-Plate transitional basalts (Coish and others, 1985; Pintson and others, 1985) which erupted at the triple junction in a setting probably similar to that of the Afar triangle. Although the outcrop pattern of the Tibbit Hill Formation forms a relatively narrow (<10 km), discontinuous zone of metavolcanic rocks its excess mass gives rise to the most positive (+54 mGal) gravity anomaly in the northern Appalachians. Analysis of this gravity anomaly (Kumarapeli and others, 1981) shows that the Tibbit Hill Formation represents one of the most voluminous remnants of Iapetan synrift volcanics known. In Waterloo area, Québec, the Tibbit Hill Formation contains a relatively minor unit of metafelsic rocks whose volcanic parent appears to

have been comenditic in composition. The rocks from this unit have yielded a reliable U-Pb zircon age of 554 +4/-2 Ma (Kumarapeli and others, 1989). This date gives the age of a chemostratigraphic unit of the Tibbit Hill Formation. However, in Québec where the volcanic sequence is thickest, there is no hint of thick interbeds of sediments. Therefore, we prefer the view that the Tibbit Hill volcanics formed during a relatively short span of time (≈5 Ma), and that the age of the comenditic unit is essentially the age of the Tibbit Hill Formation.

Pre-breakup (but post-Grenvillian), extension-related magmatic products from the continental margin of Laurentia show a wide spread of ages. The segment extending from southern Pennsylvania through Virginia to North Carolina has yielded ages ranging from 570 to 734 Ma (Goldberg and others, 1986; Lukert and Banks, 1984; Mose and Nagel, 1984; Odom and Fullagar, 1984). Reliable dates from the Newfoundland segment are approximately 600 Ma (Williams and others, 1985). The presence of extension-related magmatic products dating back from over 700 Ma has been interpreted by some (for example, Williams and others, 1985; Goldberg and others, 1986) to mean protracted rifting in the late Proterozoic prior to the onset of sea-floor spreading in the Early Cambrian. A somewhat different view, based on the age dates from the central Appalachians, is that extension-related magmatism along the nascent continental margin of Laurentia was related to two distinct events (Badger and others, 1988). The earlier event (≈730-650 Ma) did not lead to successful rifting but the later one (≈570 Ma) was successful and led to the eventual opening of Iapetus. No evidence has yet come to light from the Sutton Mountains region to suggest a protracted period of rifting, or a failed rifting event, prior to continental breakup and sea-floor spreading unless the Grenville dike swarm is older than present estimates of its age. The only known rift-related igneous event older than the Tibbit Hill volcanism is the emplacement of the alkalic-carbonatitic complexes in the Ottawa graben. Their emplacement in the very early part of Early Cambrian may have taken place at or shortly after the beginning of rifting at the Sutton Mountains triple junction. The Tibbit Hill volcanism, being the youngest major episode of volcanism known from the ancient continental margin of Laurentia, may have taken place shortly before and possibly as a precursor to sea-floor spreading at the Sutton Mountains triple junction.

PALEOGEOGRAPHIC ANALYSIS OF THE PRESHELF FACIES IN THE OAK HILL AND CAMELS HUMP GROUPS: CONSTRAINTS ON SYN-RIFT AND POST-BREAK-UP SUBSIDENCE IN THE QUÉBEC REENTRANT

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The preshelf units of the Oak Hill Group in Québec and the western facies of the Camels Hump Group in northern Vermont comprise the late Precambrian to early Cambrian syn-rift and post-break-up rocks in the Québec Reentrant. The subsequent deformation and metamorphism of these rocks during Taconian and Acadian orogenesis does not preclude a first-order restoration of the depositional environments of the majority of the syn-rift facies, especially in Québec. Recent field work in the hinterland of the Sutton and Green Mountains has defined several spatial and lithologic associations that are instrumental in developing a model for the paleogeographic setting of the syn-rift facies, and suggests that the distribution of lithologies in the preshelf sequence is dependent upon their spatial relation with respect to progressive amounts of lithospheric thinning during the earliest stage of Iapean rifting.

The western facies of the Camels Hump Group and the preshelf rocks of the Oak Hill Group are generally divisible into two rock sequences, which likely represent two lithotectonic assemblages: 1) syn-rift facies, and 2) a post-break-up facies. The rift facies is characterized by volcanic, volcanoclastic, and coarse clastic rocks (Tibbit Hill Formation, Call Mill Slate/Pinnacle Formation, and White Brook Dolomite). The post-break-up facies is composed of non-volcanic fine-grained clastic rocks which include the West Sutton Slate, Frelighsburg Formation, and the Fairfield Pond Formation.

The area of interest encompasses a region extending from central Vermont to Sutton, Québec within several kilometers of the Lincoln anticline, Enosburg Falls anticline, and the Pinnacle Mountain anticline. The importance of this region to the study of Iapean rifting in the Québec Reentrant is highlighted by the regional along-strike facies changes exemplified by the transition from Grenville-age basement exposures at the Lincoln massif to a thick sequence of rift-related volcanic rocks in northern Vermont and southern Québec. Clearly, this transect defines an across-strike view of the rift basin, where the Lincoln massif represents part of the margin of the rift basin, which passes northward (basinward) into progressively thinned continental crust as indicated by intrusions of thick rift-related volcanic rocks.

The preshelf clastic rocks of the Oak Hill Group and the western facies of the Camels Hump Group are comprised of a suite of rocks ranging in thickness from 2000 m to 2500 m, although the relative thicknesses of the syn-rift and post-break-up lithologies varies with the degree of lithospheric attenuation. The clastic rocks are underlain in northern Vermont and southern Québec by volcanic and volcanoclastic rocks called the Tibbit Hill Formation. Two major belts of the Tibbit Hill Formation comprise the majority of the rift-related volcanic rocks in the region. Preliminary mapping of the clastic cover rocks at the southern end of the northern volcanic belt show a marked contrast in mineralogy and stratigraphy when compared with lithologic descriptions (Doolan, 1988) of the clastic rocks overlying the southern volcanic belt. Further, rift clastics south of the southern volcanic belt suggest varied and different depositional settings from those prevalent to the north. For these reasons, the discussion of the preshelf clastic stratigraphy of the Oak Hill and Camels Hump Groups will be

addressed as they occur in 3 regions. The northern region (Region I) is defined by clastic rocks that overly the 2 belts of Tibbit Hill volcanic rocks. Clastic rocks that overly the northern volcanic belt have been mapped in detail by Colpron (1988) and Dowling (1988). These rocks maintain their stratigraphic integrity to the southern terminus of the volcanic belt near Enosburg, Vermont. Region II consists of those clastic rocks that occupy an intermediate, and largely unmapped, region between the exposed Tibbit Hill belts and the Lincoln massif. Region III includes clastic rocks that rest directly on basement of the Lincoln massif. These rocks have been mapped by Tauvers (1982a).

The northern belt (Region I) of Tibbit Hill volcanic rocks is composed of a monolithic sequence of chlorite-epidote greenstones. Felsic volcanic rocks have been reported by Rankin (1976) north of Brome Lake, but this is not a common lithology in the Tibbit Hill Formation. Discontinuous slate beds from 2 to 10 cm thick are the only non-volcanic clastic units in the Tibbit Hill Formation. Kumarapeli et al. (1981) suggest a maximum thickness of 8 km for the northern belt of Tibbit Hill volcanic rocks based on a profound positive gravity anomaly. Based on the stratigraphic proximity to the Pinnacle Formation, and the presence of slate beds separating volcanic rocks, the Tibbit Hill Formation is interpreted as a series of basaltic flows, punctuated by periods of quiescence in a shallow water environment.

The Tibbit Hill Formation is overlain by a thin, but continuous, black slate called the Call Mill Slate. It is identical to the slate beds present in the Tibbit Hill Formation, with exception of minor coarse clastic beds, and was deposited on a submerged plateau of an inactive volcanic terrane.

The Pinnacle Formation overlies the Call Mill Slate and is interpreted as a beach deposit, based on the maturity of the detrital component (greater than 95% quartz and heavy minerals), swash zone bedforms, and the prevalence of magnetite bedding and placer deposits in excess of 6 m thick.

A semi-continuous quartz-rich dolostone called the White Brook Dolomite overlies the Pinnacle Formation. In Region I, the White Brook Dolomite generally marks the end of coarse clastic sedimentation. An abundance of dolomitic pods and carbonate cement in the uppermost Pinnacle Formation suggest that the White Brook Dolomite was deposited as carbonate mud in a slowly transgressive environment.

In northern Vermont, the syn-rift rocks overlying the southern volcanic belt contain significant differences in both sedimentology and stratigraphy from their equivalents to the north. Extensive mapping by Doolan (1988) indicates that a major characteristic of the Pinnacle Formation in northern Vermont is the presence of graded beds. Indeed, Dennis (1964), has documented turbidite sequences in the region of Bakersfield, Vermont. This evidence, supported by the absence of magnetite bedding and the presence of a significantly more immature framework component indicates that the majority of the coarse grained, syn-rift rocks were deposited in a deeper, but more proximal setting in the rift basin than those to the north. This is consistent with the observation that the White Brook Dolomite, as observed in Québec, is largely absent, except where it overlies an anomalous fluvial facies of the Pinnacle Formation exposed in the Georgia Mountain anticline.

The stratigraphy and sedimentology of the syn-rift units in Region II are poorly understood, largely due to the absence of detailed mapping. However, excellent exposures of the Pinnacle Formation in this region indicate that the stratigraphy contains a component of graded beds, typically capped with slate. Where observed, these beds are nearly a meter thick, and contain numerous intraformational rip-up clasts. This part of the stratigraphy represents turbidite deposits, and suggests a deep-water rift basin setting.

The syn-rift clastic rocks that overly the Grenville-age basement in Region III were mapped and interpreted by Tauvers (1982b) as a sequence of subaerial and subaqueous alluvial fans derived from proximal rift margins. The great thickness of these deposits (2000 m) is in contrast to thicknesses ranging from 200 to 300 m for coarse clastics in Region I and probably Region II.

Volumetrically, the fine-grained post-break-up lithologies (Frelighsburg Formation, Fairfield Pond Formation, and West Sutton Slate) comprise the majority of the preshelf sequence, except along the rift margins. Distal facies of these rocks, particularly the Fairfield Pond Formation, are interbedded with coarse clastic rocks and are, therefore, time correlative with proximal coarse grained syn-rift rocks. However, the vertical transition from the Pinnacle Formation/White Brook Dolomite to the fine-grained post-break-up rocks is rapid on a regional scale.

The analysis of the stratigraphy, sedimentology, and depositional environments of the preshelf lithologies is a viable means for the study of the crustal response during syn-rift and post-break-up stages to the early rift-related volcanic intrusion of the Tibbit Hill Formation. An examination of the relative stratigraphic thicknesses of the rift volcanic rocks, syn-rift clastics, and the post-break-up lithologies is summarized by the observation that the volume of volcanic rocks is inversely proportional to the thickness of the coarse-grained rift clastic rocks and proportional to the thickness of the post-break-up sequence. The Pinnacle Formation comprises essentially all of the syn-rift deposits; however, this unit is extremely thin where it overlies the volcanic terranes. Where the Pinnacle Formation overlies the Grenville-age basement, it is in excess of 2000 m thick. Conversely, the post-break-up rocks are thickest over the northern volcanic belt and thinnest over the basement.

Sedimentological and paleogeographical evidence indicates that variations in the thicknesses of the clastic overburden are controlled by differential heat flow during the early crustal rifting. Beach facies deposits in the syn-rift Pinnacle Formation overlying the northern volcanic belt clearly indicate that the Tibbit Hill volcanic rocks created a shallow water plateau, which remained topographically elevated through the duration of syn-rift sedimentation. The mechanism for sustaining this plateau is the residual heat contained within the Tibbit Hill magma body. Traced southward into Regions II and III, syn-rift rocks rest on progressively less thinned crust, and apparently contain an increasing proportion of turbidite sequences and immature sediments, indicating deeper water environments more proximal to the rift margin during the rift stage.

A scenario for the early evolution of the rift basin requires that initial rift subsidence is greatest along the relatively cool margins of the rift basin (Region III). Syn-rift subsidence decreases progressively toward the axis of the basin where recently active volcanism produces a thermally buoyant terrane. Cooling of the magma bodies results in the development of a break-up unconformity and produces a second stage of basin subsidence where the amount of subsidence is proportional to the amount of crustal replacement by dense rift-volcanic rocks. This is reflected by the thick sequence of fine-grained rocks (Fairfield Pond and Frelighsburg Formations) overlying Region I, especially in Québec.

EVOLUTION OF THE CAMBRIAN PLATFORM AND PLATFORM MARGIN IN NORTHWESTERN VERMONT

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The Cambro-Ordovician platform sequence in northwestern Vermont can be divided into two sequences: a Western Shelf and Eastern Basinal Sequence (Figure 1). The Western Shelf Sequence is comprised of alternating units of siliciclastic and carbonate composition with shallow water platform affinities. The Eastern Basinal Sequence is composed of conglomerate and sandstone units in a shale matrix representing platform margin and basin deposits.

Rocks of the Western Shelf Sequence were deposited on a tectonically stable shelf following late Precambrian rifting and formation of the Iapetus Ocean. Eocambrian rift related sediments have been described by Dowling (1988) and Tauvers (1982a). They are overlain by Early Cambrian arkoses, subarkoses and quartz arenites of the Cheshire Quartzite, recording the transition from the rift basin fill sediments to the stable carbonate platform (rift-drift transition). The 400 meter thick Cheshire can be divided into a lower, finer-grained more feldspathic unit and an upper, coarser-grained quartz-rich unit (Myrow, 1983). The Cheshire contains a suite of sedimentary structures including shale-draped ripples, *Monocraterion* burrows, and herringbone and planar cross stratification. The unit was interpreted by Myrow to represent tidally-influenced shallow marine sandstone deposition.

The Dunham Dolomite is a thick (300 meter) arenaceous dolomite which can be subdivided into 4 lithofacies: peritidal, tidal channel, open platform and proximal platform margin (Gregory, 1982). The bulk of the unit exhibits 10 meter thick shallowing up cycles of open shelf muds capped by peritidal horizons. Open shelf muds have yielded specimens of *Salterella conulata* (Mehrtens and Gregory, 1984). The Dunham is capped by horizons of polymictic breccia interpreted as proximal platform margin deposits. They are interbedded with shale horizons of the Parker Slate. Recognition of the platform margin deposits is important because it documents the existence of a sharp gradient on the Early Cambrian shelf and mapping the distribution of the platform margin facies enables us to recognize the margin of the Dunham platform over a wide area in northwestern Vermont (Mehrtens and Dorsey, 1987, Mehrtens and Borre, 1988).

The Dunham Dolomite is overlain by the Monkton Quartzite, a 300 meter thick red sandstone, siltstone shale and dolomite unit (Rahmanian, 1981). Lithofacies of the Monkton include tidal flat, subtidal, tidal channel and shelf margin deposits. Sedimentary structures such as shale-draped symmetrical ripples, interference ripples, mudcracks, and vertical and horizontal burrows are common. Shallowing-up cycles composed of subtidal sandstones, tidal flat sands and shales, and supratidal dolomites characterize the unit. The uppermost horizons of the unit consist of dolomite beds with oolite ghosts and polymictic breccia horizons. The distribution of the latter facies was mapped and the position of the Monkton shelf margin delineated.

The Winooski Dolomite has been the least-studied unit in the Vermont sequence, owing to a general lack of macroscopic sedimentary structures. The 400 meter thick dolomite unit lies in gradational contact with the underlying Monkton and overlying Danby Quartzites. Basal dolomite horizons contain cryptalgalaminite structures resembling LLH stromatolites, suggesting that the base of the unit represents a shallow marine deposit. Uppermost horizons of the Winooski are also polymictic breccias interpreted as platform margin deposits.

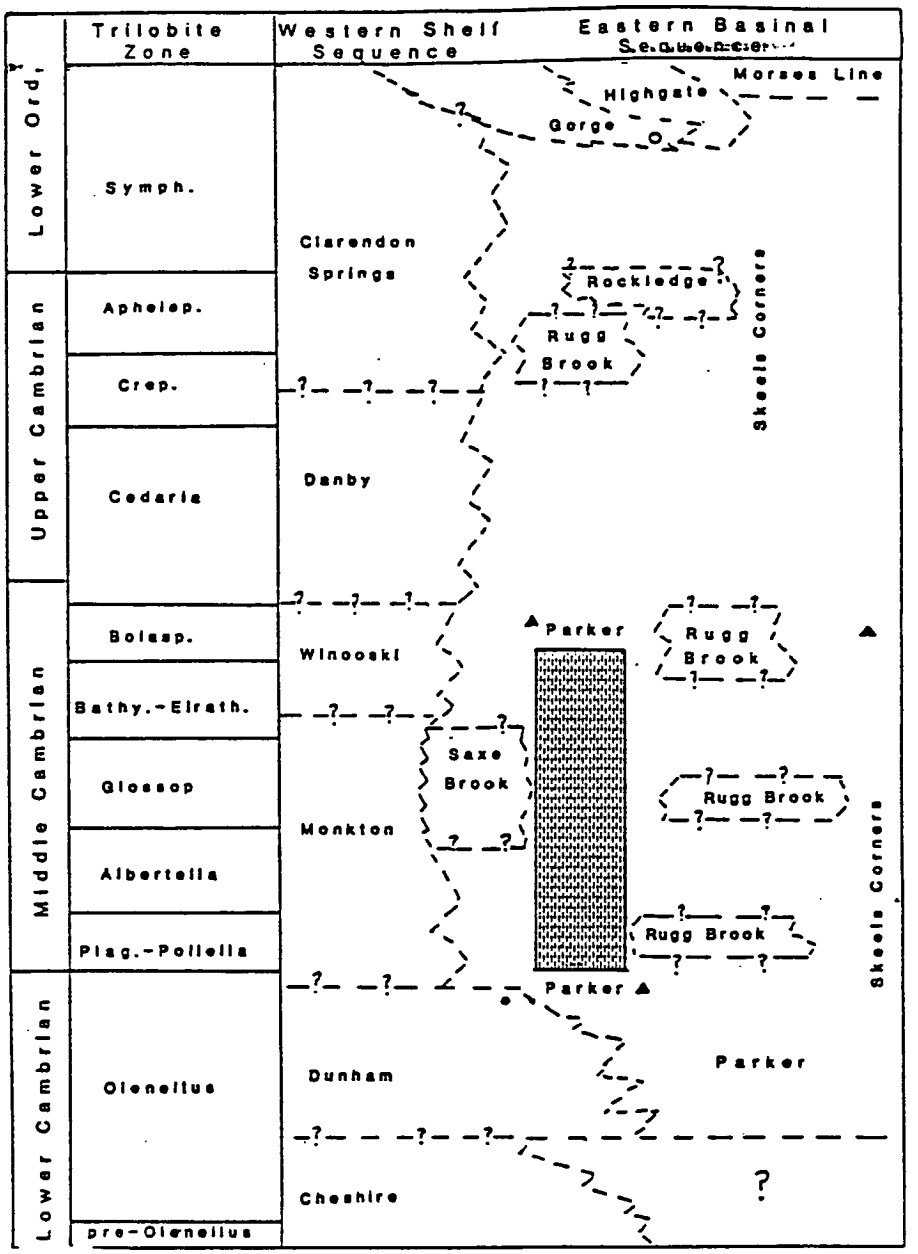

Figure 1. Correlation chart for Cambro-Ordovician rocks in northwestern Vermont. Ages of the units comprising the Western Shelf Sequence are approximate, and are, with the exception of the Dunham Dolomite, based on intertonguing relationships with the basinal shales. Ages of the units within the Eastern Basinal Sequence are from Shaw (1958) and Palmer (1970), with recent revisions: solid triangle-Dorsey, et al., 1983; solid circle- Mehrtens and Gregory, 1983; open circle-Landing, 1984. Shaded area corresponds to the duration of an unconformity within the Parker Slate.

The Upper Cambrian Danby Quartzite consists of mixed siliciclastic and carbonate sediments recording storm-influenced deposition on tidal flat, shallow subtidal, open platform and platform margin environments (Butler and Mehrtens, in press). Tidal flat sediments are represented by shale-draped symmetrical ripples, wave-generated interference ripples and herringbone cross beds. The shallow subtidal sediments include oncolites and planar cross-stratified and hummocky cross-stratified sandstones. The unit is capped by platform margin dolomitic and quartzite breccia horizons and cross bedded sandstones interpreted as platform margin sand shoals.

The Danby Quartzite is overlain by various dolomite units of Lower Ordovician age, in northwestern Vermont the Clarendon Springs Dolomite and to the south, the Beekmantown Group, and the shallow water affinities of these units is well documented (Mazzullo and Friedman, 1975).

A depositional model for the Cambrian of northwestern Vermont was presented by Dorsey, and others (1983) and their interpretations have been confirmed by continued field mapping and sedimentological studies focusing on the nature of the platform to basin transition. Several key points in the model include: 1) platform margin facies are recognized in every Cambrian unit except the Cheshire; 2) platform margin facies can be mapped out and the shape of the platform margin reconstructed; 3) the platform vertically aggraded through the Cambrian and did not build out into the basin; 4) vertical aggradation on the platform was accomplished through accumulation of shallowing-up cycles that prograded towards, but never reached, the platform margin; 5) regardless of whether sedimentation was dominantly siliciclastic or carbonate the same environments are present on the platform: inter- (or peri-) tidal, shallow subtidal, open shelf, and platform margin.

The Eastern Basinal Sequence is composed of the Parker and Skeels Corners Slates and various breccia, conglomerate and sandstone horizons of the Rugg Brook, Rockledge, Highgate and Gorge Formations (Figure 1). Mehrtens and Dorsey (1987), Mehrtens and Borre (1988), and Mehrtens and Hillman (1988) mapped out the distribution of the sandstones, breccias and conglomerates and recognized that they represent carbonate apron deposits of debris flow and high density turbidity current origin, derived off the adjacent shallow water platform. In the case of the Rockledge Formation petrography of the clasts indicates that the platform source was rich in pellets, oolites and *Epiphyton* algae (Mehrtens and Hillman, 1988).

Several key points can be made about the evolution of the Cambrian platform in northwestern Vermont: 1) An intra-shelf basin, termed the St. Albans Reentrant, existed in the platform. This basin is recognized by the distribution of platform margin and basinal deposits, and also on the abrupt eastward pinch out of platform units; 2) The age of the St. Albans Reentrant is established on the basis of the first occurrence of platform margin breccias in the Dunham Dolomite. The basin was destroyed by Upper Cambrian time when the adjacent shallow water platform foundered, recorded by the Highgate and Gorge Formations; 3) The earliest formation of the St. Albans Reentrant can be narrowly confined to mid-to late Dunham time (late Lower Cambrian) because both the Cheshire Quartzite and Dunham Dolomite extend considerable distances eastward (down dip) and platform margin facies are not recognized in these units in eastern Vermont. During Dunham time the platform margin was established in western Vermont in a position which remained stationary throughout the remainder of the Cambrian; 4) The platform margin was stationary, suggesting that its position might reflect structural control, perhaps by a reactivated rift-related fault.

EVOLUTION OF THE CAMBRO-ORDOVICIAN SHELF IN THE CHAMPLAIN VALLEY: INTERACTION OF AN EVOLVING CARBONATE BANK WITH TECTONIC SUBSIDENCE AND SEA-LEVEL OSCILLATIONS

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The Cambro-Ordovician carbonate shelf sequence in the Appalachians was deposited on an actively subsiding margin of the Laurentian shield. Major low-stands of sea-level intermittently interrupted the growth of the carbonate bank. The interaction of these two factors has produced the stratigraphic sequence seen today in the Champlain Valley.

Major submarine canyons developed during the low-stands and segmented the shelf, isolating the developing Champlain Valley carbonate sequence from similar sequences in the Mohawk Valley and in Canada. The lack of sedimentologic connection between adjacent segments prevents lithologic correlation of Champlain Valley strata with coeval strata in the Mohawk Valley and in Canada.

Major low-stands of sea-level are marked by major dis/paraconformities, paleokarst features, dolomitization and dedolomitization events, and breaks in paleontologic continuity. Recent detailed stratigraphic and paleontologic studies of the Beekmantown group in the Champlain Valley clearly identify major low-stands of sea-level between Beekmantown B and C, within Beekmantown D, between Beekmantown D and E, and after Beekmantown E. Other major low-stands have been tentatively identified in the mid-Early Cambrian and in the Middle Cambrian.

Following each major low-stand of sea-level, the carbonate depositional system was reestablished. In many instances, the tectonic subsidence during the depositional hiatus resulted in excessive post-low-stand water depths over the outer shelf. As a result, the low-stands generally coincide with major cratonward steps of the shelf margin.

Each cratonward step of the shelf margin necessitated the reestablishment of the accompanying slope and rise system. Thus, a mini-rise was built atop the foundered edge of the pre-hiatus shelf until a continuous slope redeveloped. Intermittently, carbonate shelf strata managed to build out onto the mini-rise strata, so the mini-rise sequences include stringers of coeval shelf-edge carbonates. This is especially evident in the Rysedorph Hill sequence, the mini-rise developed following a major low-stand during Beekmantown D. Other recognizable mini-rise sequences include the Middlebury limestone, developed during the post-Beekmantown E low-stand, and the New Haven sequence, probably developed following the mid-Early Cambrian low-stand.

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THE OCEANIC DOMAIN FROM EASTERN TOWNSHIPS APPALACHIANS, QUÉBEC: TACONIAN VERSUS ACADIAN TECTONICS

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In the southern Québec Appalachians, cambro-ordovician rocks of the oceanic domain (the Dunnage zone of Williams, 1979) belong to three distinctive tectonostratigraphic units. These are, in stratigraphic order, the St-Daniel Mélange, the Ascot Complex and the Magog Group (figure 1).

The St-Daniel Mélange is defined as an ophiolitic mélange with a pebbly mudstone matrix in which are enclosed blocks and slices of sandstones, mafic and felsic volcanics, granitoïds and mafic-ultramafic ophiolitic rocks (Slivitzky and St-Julien, 1987). It is interpreted as the remnants of an accretionary prism (St-Julien and Hubert, 1975; Cousineau, 1988). The **Ascot Complex** is a new stratigraphic appellation to designate a volcano-sedimentary unit known as the **Ascot Formation**. Recent work (Tremblay, 1987; 1988) indicates that this unit is made up of distinctive assemblages of volcanic and volcaniclastic rocks tectonically separated by mélange-type sediments correlative with the St-Daniel pelitic matrix. Volcanic rocks have geochemical signatures typical of subduction-related volcanism (Tremblay and Bergeron, 1989). The Magog Group is essentially a turbidite sequence unconformably overlying the St-Daniel Mélange and forming the bedrock of the St-Victor Synclinorium (SVS). It is interpreted as an accretionary forearc basin (Cousineau, 1988). These cambro-ordovician rocks are unconformably overlain by silurian units belonging to the Glenbrooke Group and to the Lake Aylmer Formation.

To the northwest, the Baie Verte-Brompton Line (Williams and St-Julien, 1982) separates the oceanic domain from rocks belonging to the passive continental margin (the Humber zone). To the southeast, the La Guadeloupe fault (St-Julien et al., 1983) marks the eastern limit of the oceanic domain. This fault is a major acadian thrust along which silurian and devonian rocks of the St-Francis Group, which form the Gaspé-Connecticut Valley Synclinorium (GCVS), are overthrusted on cambro-ordovician units of the Dunnage zone (Tremblay and St-Julien, 1989).

Recently, it has been stressed that regional folding of rocks from the oceanic domain of southern Québec is related to the Acadian orogeny (Cousineau, 1988; Cousineau and Tremblay, 1989). Structural analysis of rocks from the Sherbrooke area (figure 1), where the Ascot Complex, the Magog and St-Francis Groups are well-exposed, agrees with this statement and indicates that the SVS and GCVS are contemporaneous acadian-related structures. While the Magog and St-Francis Groups form relatively simple folded sequences, the Ascot Complex is a strongly polydeformed terrane. It shows at least three phases of deformation. D_2 and D_3 phases are acadian-related as they are also present in adjacent units. Regional folding and faulting are related to the D_2 phase. D_3 is associated with a crenulation cleavage and open folds of variable intensity. The earlier phase of deformation (D_1) of the Ascot Complex seems to be related to the tectonic accretion of volcanic rocks within the accretionary prism and is attributed to the Taconian orogeny. The acadian tectonic evolution of the area is schematically depicted in three steps (figure 2). The first one (figure 2A) shows a profile of the oceanic domain in pre-acadian times. The St-Daniel Mélange is seen as the basement of the Magog Group, and the Ascot Complex, as partially accreted and thrusted volcanic material within and against the St-Daniel. The Magog Group shows no, or very few, effects of a taconian origin. Post-ordovician

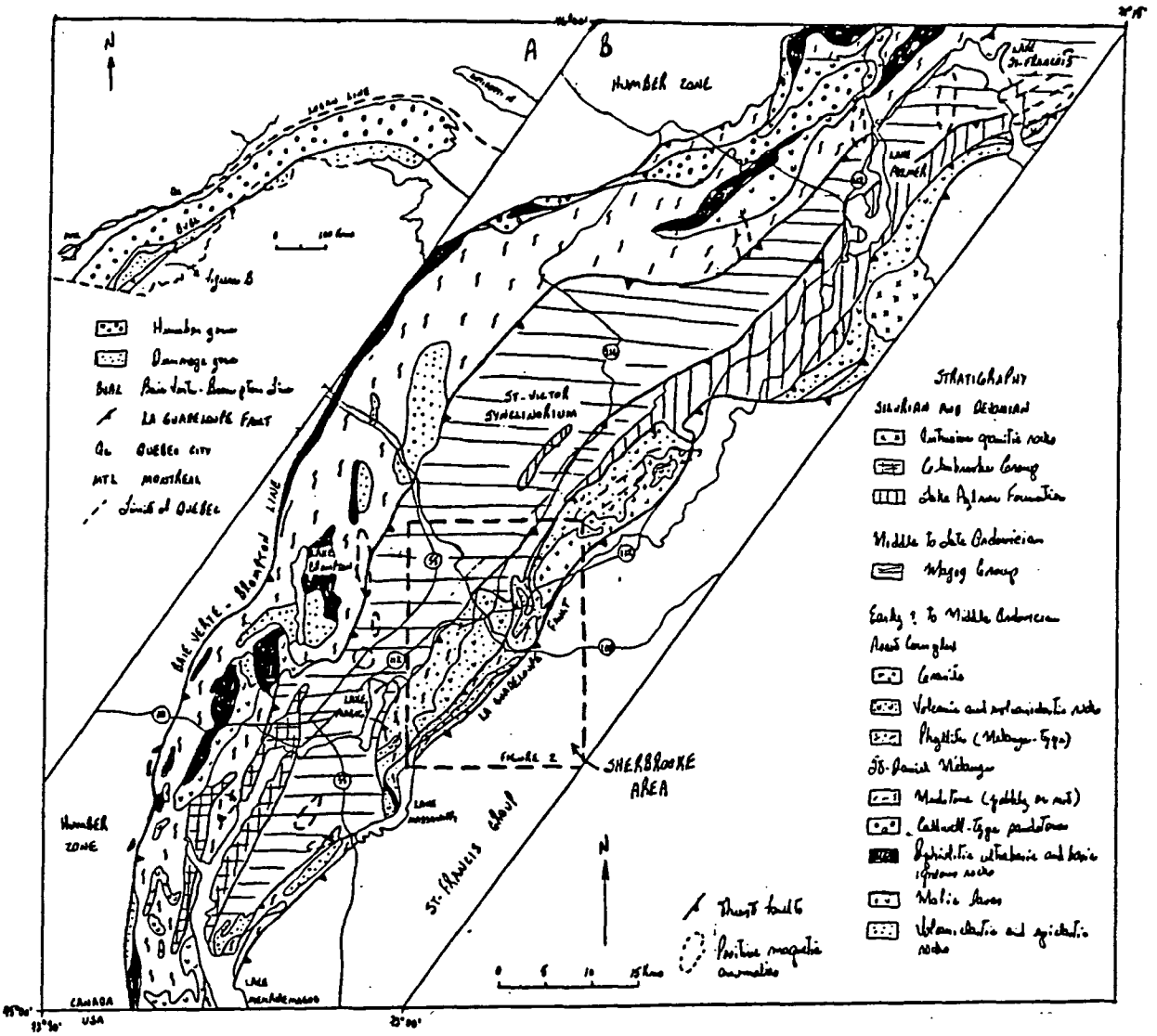

Figure 1: Geology of southern Québec oceanic domain between the international border and Lake St-François. Location map in the upper left corner. Modified after Slivitzky and St-Julien (1987).

rocks are viewed as belonging to distinctive sedimentary basins on both sides of the volcanic belt. First pulses of the Acadian orogeny lead to the formation of the SVS and GCVS (figure 2B) and of an antiformal culmination along the Ascot Complex. At a more advanced stage (figure 2C), a system of high-angle reverse faults was initiated. The La Guadeloupe fault belongs to this system and represents the main décollement surface. Due to a pre-existing anisotropy, the Ascot Complex reacted more ductilly than adjacent units during the Acadian orogeny.

In the Appalachians, the Taconian orogeny is attributed to a collision between an east-facing passive margin and an island-arc terrane over an east-dipping subduction zone (Osberg, 1978; and others). In southern Québec, the Ascot Complex and correlatives are interpreted as relics of this volcanic arc (St-Julien and Hubert, 1975; Cousineau, 1988). Systematic absence of taconian deformations in the Magog Group is a fact hardly compatible with this tectonic interpretation, considering that both the continental margin and the Ascot Complex were strongly deformed during the Taconian orogeny. Another problematic feature to be considered is that, in western New England, cambro-ordovician rocks, correlative with the oceanic domain of southern Québec, which means those east, or above, the Whitecomb Summit Thrust, show numerous examples of Taconian-related deformations (Stanley and Ratcliffe, 1985). On the other hand, the St-Daniel Mélange seems to be the only unit which possesses correlative rocks south of the international border (Doolan et al., 1982) and this could be the key to explain apparent lithologic and structural discrepancies between Québec and New England Appalachians. As a matter of fact, the St-Daniel Mélange is an orogen-scale unit with high internal complexities and characterized, like modern accretionary prism, by various styles of early structural disturbances produced by interdependent sedimentary and tectonic processes.

The role of promontories and reentrants along the border of the ancient continental margin has already been pointed out by various authors (e.g. Thomas, 1977), as an important structural control of the Taconian orogeny. In addition to diachrony in sedimentation and deformation along the continental margin, this particular original geometry could also be related to early tectonic disturbances in the oceanic domain without necessarily implicating a major tectonic collision. In fact, the Taconian orogeny could have resulted solely from an attempt to subduct a non-linear continental margin, consequently bordered by variable oceanic facies. An actual similar tectonic environment could be exemplified by the Sumatra-Banda subduction zone (Hamilton, 1979), where the nature of subducted material vary from oceanic to continental along strike.

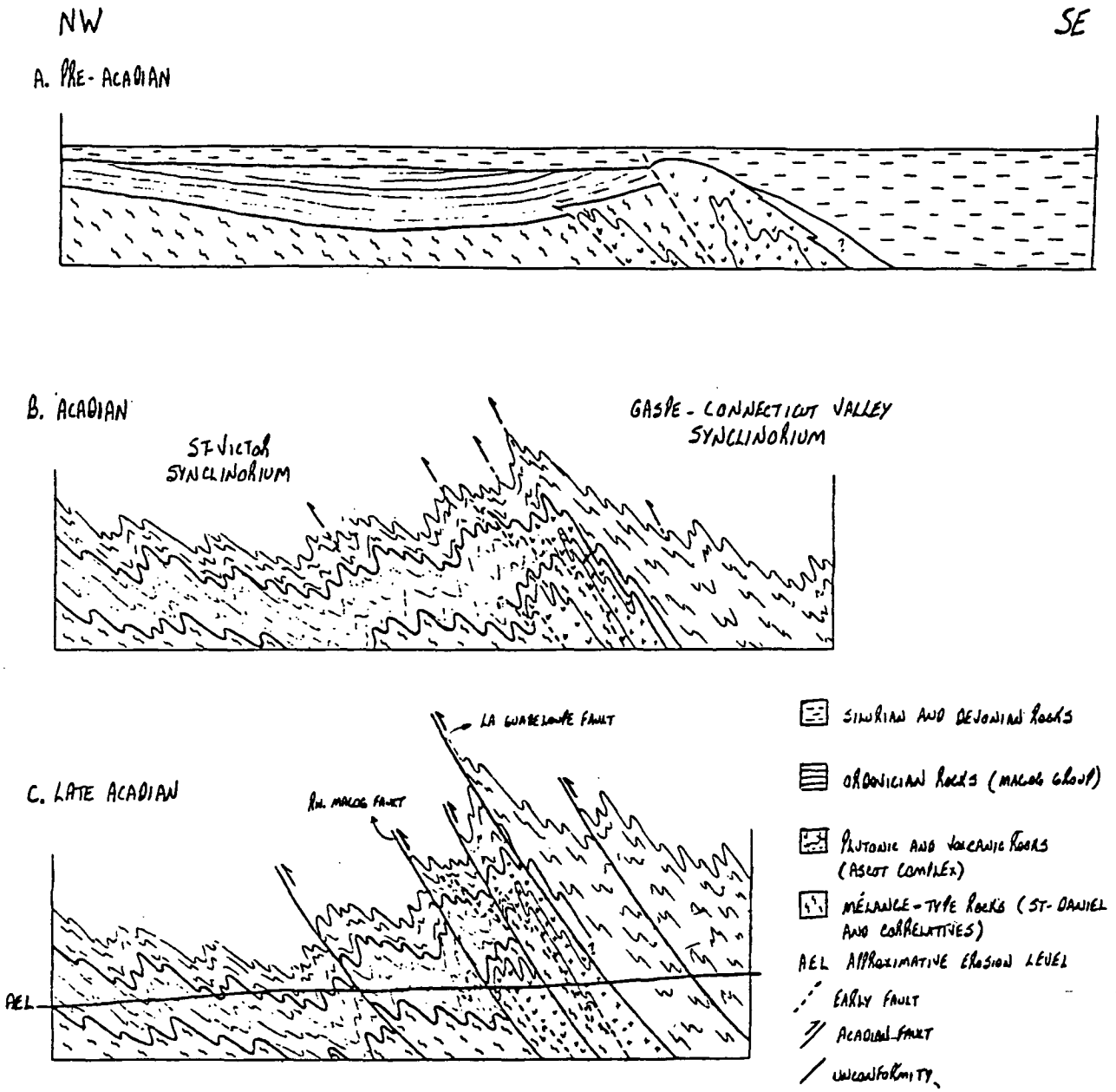

Figure 2: Schematic structural evolution of the oceanic domain during the Acadian orogeny. See text for discussion.

EVOLUTION AND COLLAGE OF THE SAINT-DANIEL MÉLANGE IN THE DUNNAGE ZONE OF THE QUÉBEC APPALACHIANS

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The Saint-Daniel Mélange is one of the four assemblages making up the Dunnage Zone in the Québec Appalachians where it extends continuously for 300 km. The mélange is either in contact to the NW with the Eastern Townships ophiolites or with the Caldwell Group; in the absence of ophiolites its NW limit is thus part of the Baie Verte-Brompton Line. To the SW, the Saint-Daniel Mélange is in contact with the Magog Group.

The Saint-Daniel Mélange is interpreted as a relict of an accretionary prism. This view is based of its structural position, the mélange lying unconformably on the ophiolites and being unconformably overlain and/or in fault contact with the fore-arc sediments of the Magog Group. This interpretation is also supported by sedimentological and structural data.

The Saint-Daniel Mélange is made up of numerous sub-units of varying dimensions. Most of these are sedimentary rocks which are called lithofacies; east of the Chaudière River, these lithofacies have been grouped together into 4 sedimentary assemblages on the basis of lithological and structural similarities. Others are tectonic slivers of intrusive and/or extrusive rocks; east of the Chaudière River, two tectonic slivers have been documented. No fossils have ever been found in the Saint-Daniel Mélange. However, one sedimentary assemblage can be correlated with rocks of the Magog Group and another sedimentary assemblage can be shown to be derived from Caldwell-type rocks. Thus, by combining these considerations and field relationships, it is possible to establish specific sedimentary and structural episodes and to place these in a stratigraphic order.

The presence of blocks of the Chain Lakes Massif in the Saint-Daniel is now well established. These blocks, together with blocks of gabbros and ultramafic rocks, comprise the Rivière des Plante ophiolitic Mélange. Within the Chain Lakes blocks, granofels of the upper amphibolite to lower granulite facies are present in the southwestern extremity of the ophiolitic mélange. But in the northeast, mylonitized arkosic wackes are found instead and these have similar relationships with nearby lithologies as do the granofels to the SW. This relationship and the general composition of the Chain Lakes Massif suggest that arkosic wackes could be the protolith to the granofels.

Blocks of Chain Lakes Massif are in contact with serpentinized harzburgite and dunite which have been carbonatized, occasionally pervasively, and transformed into ophicalcite. This ophiolitic mélange is also in contact with an extensive, thin, sheared serpentinite sheet and with a polygenic conglomerate belonging to the uppermost part of the Thetford Mines ophiolite. Ophicalcite and polygenic volcanogenic breccias are common along modern oceanic (transform) faults. Thus, the environment proposed here to account for the formation of this ophiolitic mélange is that of basin located near a continent-ocean transform-type fault.

The Ware Volcanics are intermediate to felsic volcanic/pyroclastic rocks which form a series of slivers near the village of St-Luc, to the NE of the Rivière des Plante ophiolitic Mélange. Lead isotope studies of galenas from these volcanic rocks and from the Ascot Complex by Gariepy (pers. comm., 1988) show similar ratios suggesting possible correlations between these two units. Studies of the Ascot Complex by Tremblay (1988), on the deep seismic profile underlying the Connecticut

Valley-Gaspé Synclinorium in the Eastern Townships by Bernard (1987), of the sedimentology of the Magog Group by Cousineau (1988), and of the sedimentology of the Mictaw Group by De Broucker (1986) all suggest that the Chain Lakes Massif could be the basement on which part of the Ascot (and Ware Volcanics) formed a magmatic arc. The Chain Lakes Massif, together with a slice of oceanic crust, would therefore be the basement of the Saint-Daniel Mélange.

Assemblage 1 of the Saint-Daniel is the assemblage most commonly encountered in the mélange. Three lithofacies are present in this assemblage: green and black shales, calcareous siltstone and shale, and black sandstone and shale. In general, bedding is seldom well preserved. All three lithofacies contain interbeds of black shale. Interstratification of the green and black shales lithofacies with the calcareous siltstone and shale lithofacies is frequent. Dismembered beds with layer-parallel extension structures are common. Centimeter to decameter scale debris flows with calcareous siltstone fragments occur regularly; debris flows with sandstone fragments can also be seen but are unusual. Several outcrops show a strong but local transposition of the sedimentary structures by the regional cleavage. Sandstone composition is quite variable from one locality to an other. Overall structure and composition suggest that this assemblage represents a pelagic to hemipelagic sedimentation scraped off the subducting plate onto the accretionary prism. Subsequently, as in modern accretionary prisms, these offscraped sediments are dismembered, partly resedimented in small accretionary basins, and repeatedly deformed.

Assemblage 2, even though less abundant than assemblage 1, is the assemblage the most characteristic of mélanges in general. A pebbly shale lithofacies is the only lithofacies of this assemblage. This pebbly shale is a totally chaotic mixture of fragments of different size and shape set in a pervasively deformed black and/or black and green shale matrix. A phacoidal cleavage is typically present in the matrix. Fragments are almost entirely derived from lithologies of the previous assemblage. Hence, assemblage 2 postdates or is partly contemporaneous with assemblage 1.

Assemblage 2 forms lenses of different dimensions. Smaller lenses (0.5 km in length) occur along contacts between two other assemblages. Larger lenses (2-4 km in length) may occupy the entire width of the mélange. The pebbly shale does not crop out extensively and contacts between it and adjacent assemblages are rarely observed. However, where this pebbly shale can be studied in detail, no ghost or graded bedding were seen or could be deduced since no changes in either clast composition or ratios were observed. The absence of interbeds of undeformed well-stratified sequences or transitions to dismembered sequences militate for a simple olistostromal origin for this assemblage. Instead, we favor a mud diapirism/volcanism origin for this pebbly shale. Such phenomena are common in accretionary prism today.

Assemblage 3 is made up of three lithofacies which can be correlated with lithologies of a different formation within the adjacent Magog Group. The black siliceous mudstone lithofacies is equivalent to similar rocks found in the Beauceville Formation, the green siliceous mudstone lithofacies to the Etchemin Formation equivalents, and blue-green sandstone lithofacies to the Frontière Formation. The two siliceous lithofacies are poorly exposed east of the Chaudière River but are the dominant lithologies in small basins in the Saint-Daniel near the Saint-François River. The blue-green sandstone lithofacies occupies a basin more than 5 km in length. Sandstone composition is similar to that of the Frontière sandstone but bedding is markedly thinner. This would indicate a more distal character for the blue-green sandstone lithofacies compared to a more proximal character for the Frontière. Deformation is less intense than in previous assemblages but is consistent with the general deformation style of the Saint-Daniel Mélange. No dismembered beds have been seen but tectonic transposition by the regional cleavage is locally intense in the blue-green sandstone,

especially near its limits with other assemblages. In this same lithofacies, folding is tight and plunges are steeper than in the Magog Group. Considering that no fragments of this assemblage are present in the pebbly shale assemblage, the overall less-deformed state and better defined basin limits of assemblage 3, and its correlation with part of the Magog Group, it is proposed that assemblage 3 represents a more recent sedimentation episode than assemblages 1 and 2. This "Magog-type" assemblage would thus represent sediment overspills from the fore-arc basin onto basins lying on the accretionary prism. Strata in these accretionary basins would subsequently be deformed together with the underlying offscraped sediments.

Assemblage 4 consists of 3 lithofacies which all contain predominantly clasts of lithologies of the adjacent Caldwell Group. A series of aligned kilometer-size blocks make up the first "lithofacies". Bedding and sedimentary structures in these blocks are identical to those found in the adjacent Caldwell Group. Rocks within southeastern blocks are more deformed (recrystallized) than those within the northeastern blocks. The second lithofacies is a sandstone-matrix conglomerate. This conglomerate lies only along the edges of the larger blocks where it forms a discontinuous band. Clast composition is identical to lithologies found in the adjacent blocks and the matrix is a green sandstone. The last lithofacies is a shale-matrix conglomerate which lies farther away from the big blocks and/or in areas lacking big blocks.

In the second conglomerate, clast and matrix composition are variable. Conglomerates may be either clast- or matrix-supported; the matrix is generally a black shale but locally it is a black sandy shale. Clasts are generally sub-rounded but angular or rounded pebbles and boulders can be seen. Most of the clasts are green Caldwell-type sandstone and mudstone, but Rosaire-type sandstone and volcanics (Caldwell-type ?) are found uncommonly. In both conglomerates, rafts of meter-size sandstone beds can be seen. Regional deformation in the shale matrix sandstone is not regular. In the less deformed area, the matrix is itself weakly deformed and does not resemble that of the pebbly shale (assemblage 2). In the Saint-Pamphile area, rocks of the shale matrix conglomerate broadly define an anticlinal structure in the core of which rocks of assemblage 1 are present. Transposition by cleavage is intense only along the edges of this "Caldwell-type" assemblage where it is in contact with the "Magog-type" assemblage 3.

Structures and textures of assemblage 4 differ from those of previous assemblages and are more similar to those expected in an olistostromal deposit. Sliding and breaking down of big olistoliths with formation of numerous debris flows, as is envisaged here for formation of this assemblage, is also well documented in modern accretionary prisms. Caldwell-type sediments in Québec Appalachians have been shown to be derived from the Laurentian craton and to have been deposited early in a continental slope environment. They thus belong to the Humber Zone and not to the Dunnage Zone like the "Magog-type" or other previous assemblages. Sedimentation of the "Caldwell-type" assemblage therefore could not have been possible before the tectonic collage of the Humber and Dunnage Zones during the Taconian Orogeny. Then, passive margin units were incorporated into the accretionary prism, brought up to surface along thrust faults, and partly eroded to give olistostromal deposits in accretionary basins. Assemblage 4 is thus the youngest sedimentary episode recorded in the Saint-Daniel Mélange.

Finally deformation of all assemblages continued until the end of the Taconian Orogeny. The actual collage of the various units of the Saint-Daniel Mélange and the development of the regional cross-cutting cleavage was an Acadian event. This can be demonstrated by the fact that the Silurian Cranbourne Formation, which lies unconformably on the Baie Vert-Brompton Line is folded and cut by the same regional cleavage as rocks of the underlying Saint-Daniel Mélange.

ORDOVICIAN AND SILURIAN BASIN EVOLUTION AND TECTONICS ALONG THE BAIE VERTE-BROMPTON LINE IN NORTHWESTERN MAINE.

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General

Northwestern Maine is located along the western margin of the Connecticut Valley-Gaspé synclinorium (CVGS) where a thick low greenschist, Early-through-Middle Devonian sequence is in fault contact with units of Cambrian to Late Silurian age (Fig. 1). The fault, called the La Guadeloupe Fault in the Eastern Townships of Québec, extends through the northwestern "tip" of Maine as the Rocky Mountain Fault and then into, and well beyond, the Lake Temiscouata region of Québec where it is unnamed. The pre-Devonian units record a depositional and tectonic history that spans and follows the Taconian Orogeny in the Québec Reentrant along the Baie Verte-Brompton Line. The Ordovician and Early Silurian (?) Depot Mountain Formation in Maine is a continuation of the Magog Group of the St. Victor Synclinorium (SVS) from the southwest and correlates with Cabano Formation in Québec again to the northeast. Recent Taconian plate tectonic reconstructions for western New England by Stanley and Ratcliffe (1985), for eastern New York by Bradley and Kusky (1986), and for eastern Québec by Cousineau and St. Julien (1985, 1986) suggest that the synclinorial succession was deposited in a forearc basin within an accretionary wedge. Of particular interest here is the Depot Mountain Formation which contains Caradocian graptolites in its Lower Member while its, so far unfossiliferous,

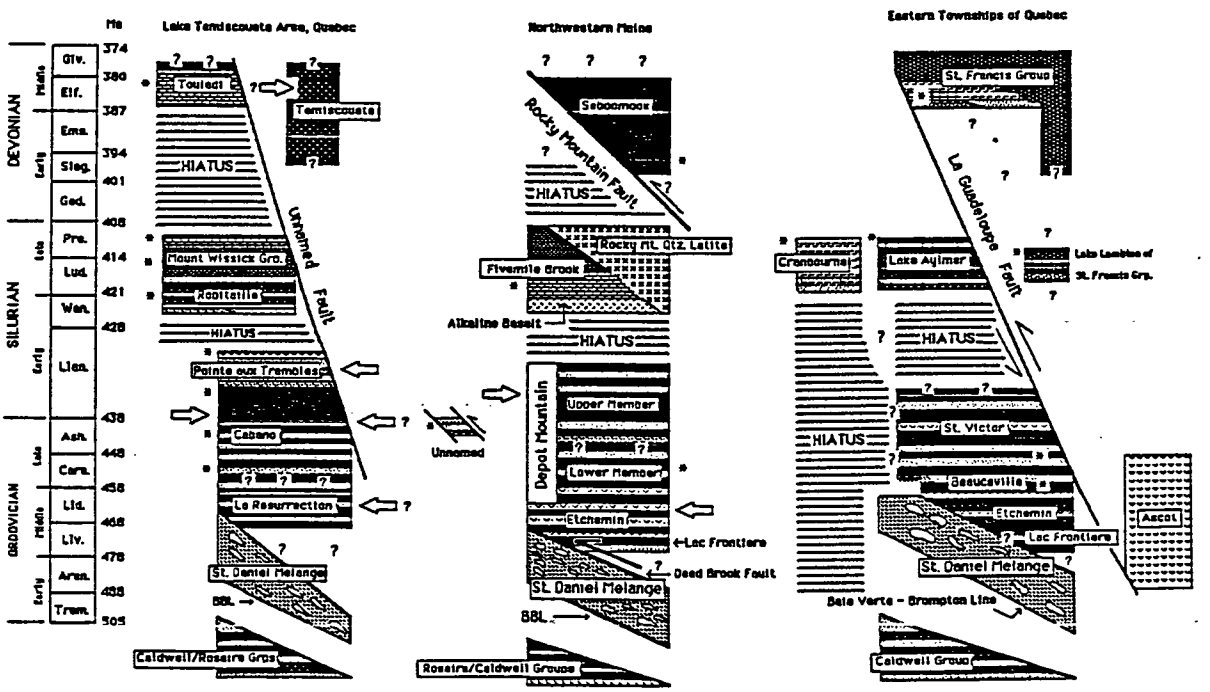

Figure 1. Comparison of the stratigraphic succession in the St. Victor Synclinorium of northwestern Maine with that of the Eastern Townships and Lake Temiscouta regions of Quebec. Asterisks indicate general locations of fossil localities. Open arrows show inferred sediment transport directions based on sandstone provenance; left-pointing arrows indicate transport from the southeast (present-day direction), right-pointing arrows indicate transport from the northwest.

Upper Member is a good lithologic correlative of the Caradocian-early Llandoverian Cabano Formation to the northeast. The Cabano is the oldest unit in what has usually been regarded as a post-Taconian successor basin that extends from the Lake Temiscouata region to the Gaspé Peninsula. Recently Caradocian and Ashgillian graptolites have been found in the lower part of the Cabano and it is now likely that the Depot Mountain and Cabano formations, together with the Magog Group, were deposited in a basin that developed penecontemporaneously with the Taconian Orogeny and extended into the more tectonically quiet Early Silurian. Some Taconian deformation of these units, especially the Late Ordovician phases, appears likely but has been difficult to separate from the affects of the Acadian. The units of the SVS are also of low-greenschist grade and share a single cleavage with the Devonian sequence in the CVGS. However, faults mapped in the Eastern Townships between some of the units are shown as folded on recent maps by Slivitzky and St. Julien (1987) and Cousineau and others (1984).

St. Victor Synclinorium Stratigraphy in Maine

Sub-synclinorium Mélange: Polydeformed gray, red, and green phyllite interbedded with quartzite and generally containing abundant blocks of quartzite, graywacke, and chert forms a broad belt along the Maine-Québec border northwest of the SVS. Divided into the "Estcourt" and "Lac Landry" sequences by Boudette and others (1976), this unit was designated as the Estcourt Road Formation by Roy (1980; 1982) and shown as the St. Daniel Formation on the recent Maine bedrock map (Osberg and others, 1985). **The St. Daniel Formation** has, in recent years, been carried as a narrow fault-bounded belt to the Maine border near St. Pamphile Québec by Cousineau (1987, personal communication). However, the width of the belt as presently mapped in Maine is larger than in Québec and is currently being subdivided by S. G. Pollock. In Maine the St. Daniel is made up in large part of "disrupted" or "broken" stratigraphy of the Caldwell and Rosaire groups. It contains blocks in a wide size-range of vitreous quartzite , red/green feldspathic quartzite, limestone, and very calcareous siltstone. Portions of the belt look very much like disrupted stratigraphy of the Trinity Group as mapped in, and northeast of, the Lake Temiscouata region. As in the Eastern Townships, the St. Daniel Formation abuts the Depot Mountain Formation of the SVS along a fault, the Dead Brook Fault. Where this fault is observed or closely approached, blocks of volcaniclastic graywacke similar to the Lower Member of the Depot Mountain Formation or the Etchemin Formation are seen in the mélange. This suggests that deformation of the mélange extended into the later part of the Ordovician and may even have been concurrent with deposition of the younger parts of the of the SVS sequence. Thus in Figure 1 the lithodemic St. Daniel is shown as a diachronous mélange which is, at least in part, temporally equivalent to parts of the SVS stratigraphy while being generally younger than the Caldwell and Rosaire groups.

Etchemin Formation: The Etchemin Formation of Cousineau (1987) and Cousineau and St. Julien (1985) extends into Maine between Lac Frontière and St. Pamphile where it appears to underlie the Depot Mountain Formation but may be in large part equivalent to the Lower Member of that formation. The Etchemin consists of interbedded gray and green volcaniclastic sandstones, water-laid crystal-lithic tuffs, and silicious slate. The sandstones are typically graded and show the sedimentary features of turbidites.

Depot Mountain Formation: Originally designated as the "Depot Mountain Sequence" by Boudette and others (1976) the unit was slightly redefined and designated as the Depot Mountain Formation by Roy (1980; 1982). On the state bedrock map (Osberg and others, 1985) the unit is divided into two phases, Odms and Odmp, that are designated as the Lower and Upper members respectively by Roy (in press). A conformable contact between the two members has not been observed. In the two places where the contact zone has so far been seen it has proven to be a fault. The **Lower Member** consists of thick-bedded, dark-gray, fine-to-coarse grained feldspatholithic graywacke

interbedded with gray and black slate. The graywacke has a volcanic provenance. Locally, as at the type locality at Depot Mountain, lithic and crystal water-laid tuffs and volcaniclastic sandstone and angular-clast conglomerate form sequences that are at least tens of meters thick. The volcaniclastic sequences are likely perched at multiple levels within the member but cannot be traced along strike very far due to insufficient exposure. Roundstone conglomerate with felsic volcanic clasts are also present in a few places. The Lower Member is largely correlative with the Beauceville Formation to the southwest and lithologically very similar to the La Resurrection Formation to the northeast. The **Upper Member** is composed of dark-gray, usually well-foliated, lithic graywacke interbedded with an equal or greater proportion of laminated gray slate. Pebble and cobble conglomerate are commonly seen within thicker packages of sandstone. The Upper Member is lithologically similar to the St. Victor Formation to the southwest and Cabano in the Lake Temiscouata region. The age of the Depot Mountain Formation is understood to be Caradocian to Early Llandovery base on an Ordovician graptolite locality at Depot Mountain (Boudette and others, 1976) and lithologic correlation of the Upper Member with the well-dated Cabano Formation.

Unnamed Unit: An unnamed succession of quartzite, quartz-clast conglomerate, calcareous siltstone and slate is sparsely exposed west of Rocky Mountain. This unit has produced a rich fauna of brachiopods, pelecypods, and trilobites that have been assigned to the Ashgillian (Roy, in press). Although a probable temporal equivalent of the Depot Mountain Formation to which it was previously assigned by Roy (1982) and Roy and Lowell (1983), rock-types prominent in the unit are not seen in the Depot Mountain. The unnamed unit is apparently in a fault sliver that is surrounded by St. Daniel mélange. S. G. Pollock (personal communication, 1988) has found similar rocks elsewhere in St. Daniel terrane last summer and suggests that a much broader distribution for the unnamed unit is likely.

Fivemile Brook Formation and Rocky Mountain Quartz Latite: The Depot Mountain Formation is overlain disconformably by the both the Fivemile Brook and Rocky Mountain Quartz Latite. The hiatus represented by the disconformity is probably roughly the same as the well documented Wenlockian hiatus between the Pointe-aux-Trembles Formation and younger units of the Late Silurian described by Lajoie and others (1968) in the Lake Temiscouata area (Fig. 1). The **Fivemile Brook Formation** is a sequence of light greenish gray, variably calcareous phyllite interlayered with thin beds of biomicrite. Locally, as at the type section at Fivemile Brook, massive alkali basalt flows are present and at other places both basalt and felsic tuff are within the unit (Schwartz and Hon, 1983; Schwartz and others, 1984; Dubois and Roy, 1985; Dubois, 1986). The **Rocky Mountain Quartz Latite** consists of as much as 1000 m of fragmental siliceous crystal and lithic tuffs and lesser alkali basalt (Boudette and others, 1976; Schwartz and Hon, 1983). The main mass of the unit at Rocky Mountain does not have interbedded sedimentary rocks but the presence along strike of felsic volcanic rocks in the Fivemile Brook Formation suggests widespread distribution of eruptive materials and the penecontemporaneous nature of the two rock units.

Conclusions

Figure 2 illustrates the interpretation of the stratigraphy in the SVS and relates that interpretation to the volcanic arc history of the BHBM Terrane and the basins of the KCMS. The Ordovician stratigraphy of the SVS formed in a forearc basin and was deposited on an "active" mélange. The mélange developed over much of the Early and Middle Ordovician, possibly overlapping with the Penobscottian Orogeny in the BHBM belt. The forearc basin sediments, especially the early deposits, were probably mildly deformed during the Middle and possibly Late Ordovician during accretionary-wedge thrusting in the Internal Domain of the Notre Dame Mountains. Early basin fill was largely derived from the volcanic units of the BHBM and the Ascot Formation. The upper part of the fill was

mostly derived from metasedimentary terranes in the accretionary wedge to the northwest or from the pre-arc stratigraphy of the, then dormant and uplifted, arc to the southwest. Uplift of the forearc basin in the late Llandoverian and early Wenlockian was associated with and extensional regime that is present on the eastern side of the CVGS in the BHBM terrane. The extensional history which produced alkaline and bimodal volcanism is associated with the development of a deepening and widening Devonian flysch basin complex in the CVGS. The Devonian flysch basins evolved as the youngest axis of deposition in a prograding series of basins that "evolved out of" the "rear of arc" basin in the KCMS. The Acadian Orogeny deformed the CVGS stratigraphy in the Middle Devonian as the last phase of a regionally diachronous orogenic event.

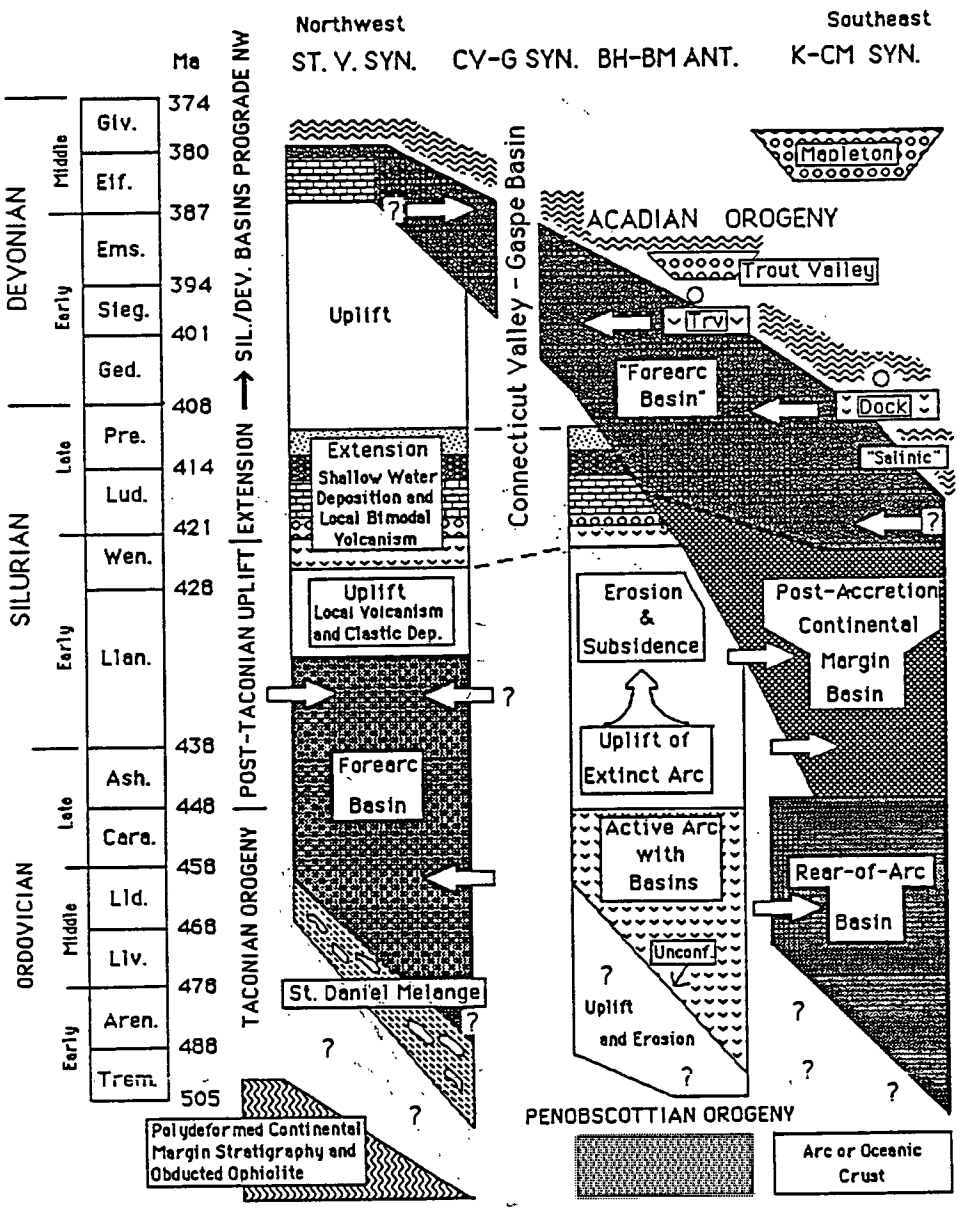

Figure 2. Tectonostratigraphic interpretation of the evolution of the St. Victor Synclinorium in northwestern Maine and correlations of that history with the evolution of the Bonson Hill-Boundary Mountain belt and the Kearsarge-Central Maine Synclinorium of north-central and eastern Maine.

THE NATURE OF THE BRONSON HILL ZONE DURING THE TACONIAN OROGENY

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To the student of the Taconian orogeny in western New England, the Bronson Hill volcanic arc was the Never-neverland in the misty east related to an east-dipping subduction zone believed to have existed through early and medial Ordovician time. To the student of the Bronson Hill rocks themselves this relationship was far less obvious, and the pre-Silurian stratigraphy and plutonism of the region have been interpreted in two distinctly different ways. The sequence of rock units in one interpretation is: I) Late Proterozoic microcline gneisses, schists and quartzites exposed only in the core of the Pelham gneiss dome and in southeastern Connecticut, 2) Late Proterozoic through early Ordovician plagioclase gneisses and amphibolites of plutonic and volcanic aspect (Monson, Fourmile, Swanzey, Pauchaug) unconformably overlain by 3) medial Ordovician Ammonoosuc Volcanics of probable arc affinities and Partridge Formation black shales with volcanics, also of probable island arc affinities, in turn overlain unconformably by 4) Lower Silurian (Llandoverian, 438-428 m.y.) Clough Quartzite. In the second interpretation, part of the plagioclase gneisses of 2) are considered to be intrusive into the Ammonoosuc and younger than medial Ordovician. Common to both interpretations has been a correlation of the Partridge to a medial Ordovician (Caradocian; 458-448 m.y.) graptolite locality in the Boundary Mountain anticlinorium, northwestern Maine; and an unconformity at the base of the Clough Quartzite.

New radiometric dating of carefully selected zircon separates from central Massachusetts and southwestern New Hampshire gives the following results :

Microcline gneiss of the Dry Hill Gneiss in the Pelham dome, interpreted as metamorphosed alkali rhyolite, is confirmed as late Proterozoic, yielding an age of 614 +9/-7. Another sample shows strong evidence of inheritance with a minimum mean age of 1400.

Plagioclase gneisses (Monson, Fourmile, Swanzey, Pauchaug) and related amphibolites, now believed to represent a complex assemblage of largely intrusive calc-alkaline igneous rocks, yield ages of 454 +4/-2, 454 +9/-3, 452 +3/-2, 447 +3/-2, close to 443 and close to 443, thus ranging from medial Ordovician through late Ordovician, consistent with other dates on Oliverian granitoid rocks in the region. A gneissic gabbro within the Swanzey Gneiss yields 454 +3/-2. No sample of plagioclase gneiss yet obtained in this study has yielded a Late Precambrian or Cambrian age like those obtained from lithically similar gneisses in coastal Connecticut.

A quartz-phyric rhyolite in the Upper Member of the Ammonoosuc Volcanics, about 275 meters above the top of the Lower Member and 115 meters below the base of the Partridge Formation, that was collected from a new artificial exposure in November 1988, yields an age of 453 ±2. A pyroclastic quartz-phyric rhyolite bed with clear primary textures from the Partridge, 20 meters above the top of the Ammonoosuc Volcanics, yields a refined zircon age of 449 +3/-2, significantly different from a previously reported age of 432 +16/-5 based on three discordant analyses. The Ammonoosuc and Partridge localities are 3 km from the Bernardston fossil locality where the Clough contains Silurian fossils and the Fitch Formation contains lowest Devonian conodonts.

These results show that there is almost complete temporal overlap between the Ammonoosuc-Partridge cover sequence and the physically underlying plutonic gneisses of the Monson, Fourmile, Swanzey, and Pauchaug. These results rule out the unconformity hypothesis, previously espoused by us on the basis of field relationships in southern New England, in which the Ammonoosuc-Partridge was unconformably above the plagioclase gneisses. These results also appear to rule out a hypothesis proposed by others that the plagioclase gneisses are intrusive into the Ammonoosuc, unless the Lower Member is substantially older than the Upper Member dated here. The hypothesis that the plagioclase gneisses are intrusive into the Ammonoosuc is also not supported by field data in southern New England.

The Monson, Fourmile and other plagioclase gneisses of the Bronson Hill zone appear to represent the plutonic roots of a calc-alkaline volcanic suite that was formed in the Bronson Hill arc between approximately 454 - 443 m.y. The Upper Member of the Ammonoosuc Volcanics and the Partridge Formation belong to a cover sequence that spanned from at least 453 to 449 m. y. Despite their apparent contemporaneity, the igneous rocks in the two sequences have very little geochemical similarity. What then was the process by which they were brought to their present close juxtapositon, probably before erosion and deposition of the Early Silurian Clough Quartzite?

In studies of modern island arcs, it has commonly been suggested that the arc lithosphere is undergoing extension. Extensional detachment faults are features that are commonly ascribed to such environments and that are also a plausible geometrical way to bring deep-seated plutonic rocks into contact with volcanic-sedimentary cover of the same age. While we have not yet worked out the necessary geometry and geometrical evidence that such a fault was operative in the Bronson Hill arc during the late Ordovician, this hypothesis seems worthy of further exploration. In its favor is the striking localization of a zone of pre-metamorphic sulfidic hydrothermal alteration in Monson Gneiss of the Orange area. This zone, 10 km long and up to 1 km wide lies adjacent to the basal contact of the Ammonoosuc Volcanics.

The new zircon ages confirm that the Ammonoosuc Volcanics and Partridge Formation were deposited at essentially the same time as the Caradocian emplacement of the Giddings Brook thrust sheet in westernmost New England and eastern New York. This suggests that tectonic features known to post-date the Giddings Brook slice may have formed in the late Ordovician, which may be the true time of final closure of Iapetus Ocean in New England. This can also be equated conveniently with the notion that the early Silurian extension of the Merrimack trough might have been related to late phases of back-arc extension east of the Bronson Hill arc. However, the new zircon ages give no hint of a stratigraphic or plutonic record of the period from late Cambrian through early middle Ordovician time, when one might imagine that an early version of the Bronson Hill volcanic arc was approaching North America.

A second enigmatic contact is between the apparently plutonic plagioclase gneisses (Fourmile) of the Pelham dome and the underlying the late Proterozoic volcanic and sedimentary strata with their possible Avalon affinities and traces of granulite facies metamorphism. Geologic mapping has shown that this contact was already established and recumbently folded during early phases of the Acadian orogeny. If the Bronson Hill volcanic arc was built on a fragment or margin of Avalon as we have proposed previously, then an Ordovician closure of Iapetus as suggested here is inconsistent with paleomagnetic data that shows Avalon in high southern latitudes in middle Ordovician time. An alternative currently being explored is that this contact and the apparent metamorphic discontinuity is a major fault, that brought Avalon crust against crust of the Bronson Hill volcanic arc early in the Acadian.

RECENT DEVELOPMENTS IN THE CONNECTICUT VALLEY TROUGH, VERMONT

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Reconnaissance remapping and restudy of the rocks in the Connecticut Valley trough in eastern Vermont north of 43° 30' north latitude has resulted in a number of proposed changes in the stratigraphy, structure, and age of the stratified units.

The changes proposed here relate to rocks mapped for decades as the Gile Mountain (characterized by light-gray. micaceous quartzites interbedded with dark-gray metapelites) and Waits River (characterized by punky-brown- weathering carbonate-quartz granofels interbedded with dark-gray metapelites) Formations. The Gile Mountain Formation has been subdivided into nine previously recognized and unrecognized lithologic members. In northeasternmost Vermont, the Halls Stream Grits and the thick-bedded graded-bedded metagraywacke were both mapped by Myers (1964). One of the two most widespread units in the Gile Mountain is the quartzite and schist member (Dgqs) which consists of fine-grained light-gray micaceous quartzite interbedded with dark-gray, graphitic pelitic schist. The other most widespread member of the Gile Mountain is the previously unrecognized, thick-bedded gray-brown micaceous quartzite (Dgq). This unit is distinguished by its distinctly brownish color (due to biotite) and the 1-3 m thick beds (see, for example figure 5 of Hatch, 1988b). The final, and previously unrecognized member of the Gile Mountain is the relatively massive, fine-grained quartzite member (Dgmq) which is continuous along strike to the north with the rhythmically graded member.

One new local subunit of the Waits River Formation, Dwq, the quartzite-bearing member, has also been recognized. It is distinguished from the main body of the Waits River (Dw) solely by the presence in Dwq of beds as much as 1 meter or more thick of light-gray, fine- to medium-grained micaceous quartzite typical of unit Dgqs of the overlying Gile Mountain Formation.

Finally, to avoid questions of how many of the lenses of metavolcanic greenstones, amphibolites, and intermediate metavolcanic rocks within and between the Waits River and Gile Mountain Formations are really Standing Pond Volcanics Member of the Waits River Formation, all of the metavolcanic rocks in the trough are here mapped simply as unit Dv--Devonian metavolcanic rocks.

Fisher and Karabinos (1980) were the first to describe and use the wellpreserved rhythmically graded beds (unit Dgr) in the western (Townshend-Brownington syncline) belt of the Gile Mountain Formation near Royalton, Vermont. By careful mapping across this belt they demonstrated that at least this western belt of Gile Mountain Formation is stratigraphically above the bounding belts of Waits River Formation(Dw). Reconnaissance mapping of the rest of the western belt of Gile Mountain shows the same stratigraphic relations--a stratigraphic as well as a structural syncline.

Not previously reported are many graded beds in the predominantly dark-gray slates/phyllites in the Northfield Member of the Gile Mountain Formation (Northfield Formation of Doll and others, 1961) along the west margin of the Connecticut Valley trough (see for example figure 7 of Hatch, 1988b). These graded beds in the Northfield not only look just like the beds in the western belt (unit Dgr) of Gile Mountain to the east, but the great majority of them face to the west, strongly suggesting that the Northfield is stratigraphically above the adjoining belt of Waits River Formation, in the same

stratigraphic position as the western belt of Gile Mountain. For these reasons, I propose that the Northfield is a western, more distal facies (Member) of the Gile Mountain Formation stratigraphically above the Waits River rather than the basal unit of the trough sequence below the Waits River as shown by Doll and others (1961). Furthermore, if the Northfield is indeed distal Gile Mountain above the Waits River, then its western contact, the"R.M.C.", has to be a fault, as recently proposed on the basis of field evidence by Westerman (1985, 1987), rather than an unconformity as shown by Doll and others (1961).

In addition to the long-recognized belt of Northfield along the west margin of the trough, I have mapped as Northfield a belt of dark-gray crinkly phyllites that extends from about five km southwest of South Barre, Vt., south to a point about seven km northwest of Woodstock, Vt. This belt consists of about 90 percent dark-gray phyllites identical to the phyllites of the Northfield, a few percent well-graded, light- to dark-gray quartzite/ metapelite beds, and the remainder being beds as much as a few meters thick of punky-brown-weathering carbonate-quartz granofels characteristic of the Waits River Formation. This belt has not been distinguished before, although it corresponds in part to a belt shown by R. H. Jahns on an unpublished manuscript map (without an explanation) of the Barre 15' quadrangle. This belt contains an abundance of graded beds intermediate between that of the Northfield belt and that of the western belt of Gile Mountain Formation. Furthermore, the tops sense of the graded beds in this belt suggest that the belt is a syncline of Gile Mountain/Northfield, which I propose is intermediate in facies between the western Gile Mountain (Dgr) and the Northfield (Dn).

Along its east margin, the trough is bounded by what has come to be known over recent decades as the "Monroe line". Originally (Eric and others, 1941) described as a fault, it went through about 20 years of controversy until Doll and others (1961) settled on it being an unconformity, with the trough rocks (Meetinghouse Slate) resting on the "New Hampshire sequence" rocks to the east. Recent restudy, aided greatly by new cuts on Route I-91 and at Union Village Dam, show two distinct periods of faulting along the Monroe line (Hatch, 1987, 1988a). Ductile faulting, probably thrusting, believed to have occurred during the Acadian orogeny, is evidenced by local intense shearing, local truncation of beds, and intercalation of "New Hampshire sequence" rocks with Connecticut Valley trough rocks. Superposed on the ductile Acadian structures are kink bands, gouge zones, crushed-rock zones and slickensides interpreted as indicative of postmetamorphic, probably Mesozoic, brittle faulting. I thus concluded (Hatch, 1988a) that the "Monroe line" should revert to its original (Eric and others, 1941) name, the "Monroe fault."

Along most, but not all, of the length of the Monroe fault, on the west, or trough, side of the fault, is a narrow (generally less than 1 km), discontinuous band of dark-gray slate known (at least since Doll, 1944) as the Meetinghouse Slate Member of the Gile Mountain Formation. All of the 15' quadrangle maps from which Doll and others (1961) compiled their State Map show the Meetinghouse as the top member of the Gile Mountain facing east into the Monroe fault. Yet Doll and others (1961), for unexplained reasons, mapped the Meetinghouse as the basal member of the Gile Mountain, unconformably deposited on the rocks of the "New Hampshire sequence" to the east. The original very fine grain size of the Meetinghouse (presumably clay) makes it a rather unusual basal unit. Furthermore, the strong evidence for faulting along the Monroe line re-opens the possibilities of the Meetinghouse facing either east or west. And finally, tops can be read at many localities in the Meetinghouse in delicate thin (a few mm) metasiltstone beds, and, where best exposed, they indicate that the Meetinghouse stratigraphically overlies the main mass of Gile Mountain to its west. I thus conclude that I agree with the pre-State Map quadrangle mappers (Johansson, 1963; Eric and Dennis, 1958; Hall, 1959; White and Billings, 1951; Hadley,1950; Doll, 1944; and Lyons, 1955) that the Meetinghouse is at the top, rather than at the bottom, of the Gile Mountain, in a stratigraphic position comparable to that of the Northfield.

Although early recognized as a closely interrelated package, the age of the rocks of the Connecticut Valley trough has long been uncertain and a subject of intermittent controversy. Over the years the trough rocks have been assigned ages ranging from Middle Ordovician to Early Devonian. In the summer of 1988 a group of concerned geologists and paleontologists led by Wallace A. Bothner went to southern Québec, northernmost New Hampshire, and north-central Vermont in an effort to resolve the question of the age of the trough rocks. The principal results were two. First, all of the "graptolites" reported by Richardson (1918 and earlier references) and by Bothner (Bothner and Berry, 1985; Bothner and Finney, 1986) were determined to be inorganic mineral streaks, "pseudograptolites", or plants. Whatever they are, they were determined by Stanley C. Finney not to be Ordovician graptolites, or indeed graptolites of any age. The second result of the trip was the discovery and rediscovery of about five localities of plant fossils in southern Québec and northernmost New Hampshire identified as Emsian (late Early Devonian) by Francis M. Hueber, paleobotanist with the Department of Paleobiology, National Museum of Natural History, Smithsonian Institution, Washington, D.C.

GEOCHIMIE DES CUMULATS ULTRAMAFIQUES DU COMPLEXE OPHIOLITIQUE DE THETORD MINES, QUÉBEC: MODELE MAGMATIQUE DE LA DISTRIBUTION DES PLATINOIDES

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Nous présentons les résultats des travaux de géochimie détaillée dans le complexe ophiolitique de Thetford Mines (figure 1; Laurent et Kacira, 1987) afin de comprendre la polycyclicité de la séquence à cumulats stratigraphiquement superposée à la base dunitique où l'on connaît des indices en platinoïdes (Gisement Hall). L'étude vérifie les points suivants:

1- La consanguinité de la séquence (2 blocs tectoniquement distincts)
2- Les relations pétrologiques des platinoïdes avec les types de roches.
3- Le potentiel en platinoïdes de la séquence à cumulats où les pyroxènes deviennent une phase de fractionnement importante.

Quarante-cinq échantillons proviennent de deux affleurements montrant bien la séquence litée (figure 2). Trente-quatre de ces échantillons appartiennent à l'affleurement A et se composent en majorité de cumulats riches en olivine (C.R.O.; dunite et wehrlite). Les litages magmatiques sont d'ordre décimétrique. Les onze autres échantillons appartiennent à l'affleurement C et se composent en majorité de cumulats riches en pyroxène (C.R.P.) d'épaisseur métrique. L'échantillonnage de ces deux affleurements s'est fait de deux façons. Pour le bloc A, les 34 échantillons ont été répartis sur les 20 m d'affleurement dans le but de vérifier la polycyclicité de la séquence par les variations géochimiques des éléments majeurs, mineurs et en traces. Pour le bloc C, l'échantillonnage est plus espacé et a pour but de vérifier la consanguinité de cette séquence avec celle du bloc A et les échantillons plus près du gisement. Dans la figure 2 les lignes d'attache reliant les échantillons mettent en évidence les échantillons pris dans des lits adjacents les uns aux autres.

Les analyses des éléments majeurs et mineurs ont été exécutées par fluorescence-X et celles des éléments en traces par activation neutronique avec préconcentration au sulfure de nickel pour les platinoïdes (Bergeron, 1988).

Les résultats préliminaires montrent que les éléments compatibles (Cr, Ni, Co, Sc) reflètent les proportions des phases minérales qui fractionnent pour former les cumulats. Les phases sont dans l'ordre: olivine, chromite, clinopyroxène, orthopyroxène. Il ne semble pas que les platinoïdes soient liés au degré de différenciation de la roche (sauf Ir) mais plutôt au type pétrographique. Dans le bloc supérieur, toutes les anomalies en platinoïdes mesurées (ppb) sont dans les C.R.O.: Ir: 5.5; Pt: 132; Ru: 35. Dans le bloc inférieur (près du gisement Hall) toutes les anomalies sont mesurées dans l'une ou l'autre des lithologies; C.R.O., Pd: 111, Rh: 17.4; C.R.P.: Pd: 234; Ir: 4.2; Au: 109.

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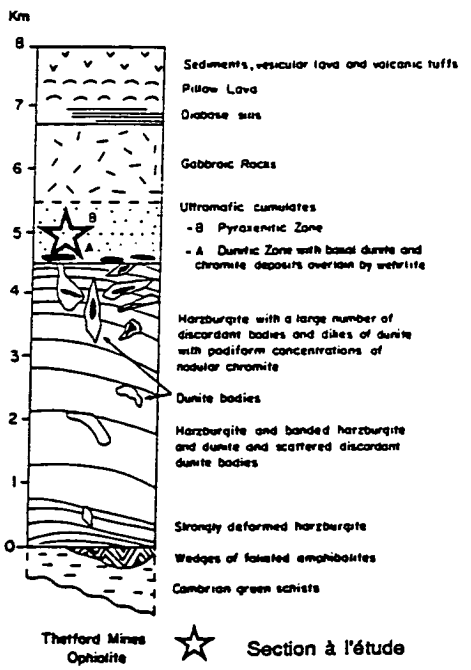
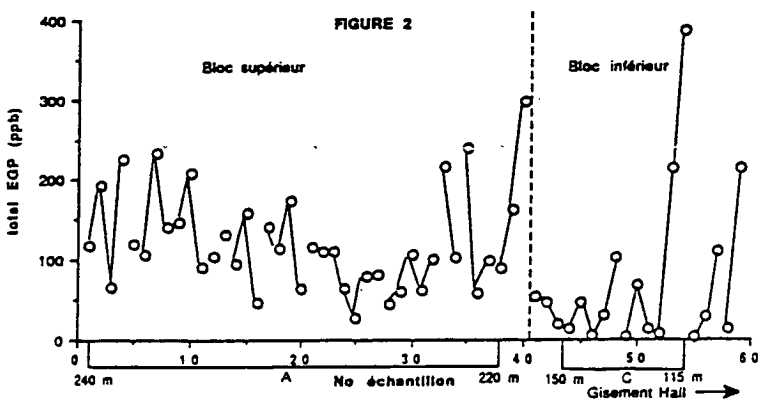
Les teneurs en soufre des échantillons sont très faibles. La plus haute teneur analysée (0.18%) appartient à une unité de C.R.O. située à environ 200 m du gisement Hall, près de l'affleurement A. On doit remarquer que l'augmentation de la teneur en soufre des échantillons ne se corrèle pas nécessairement avec une augmentation de la somme des platinoïdes. Ce dernier point suggère que les platinoïdes sont présents à la fois sous forme d'alliage et de sulfures. La figure 2 représentant la distribution des éléments du groupe du platine versus leur position stratigraphique montre une augmentation de la teneur de fond lorsqu'on s'éloigne du gisement. Toutefois les plus hautes teneurs anomaliques se situent à droite du graphique en direction du gisement Hall. Ces anomalies ponctuelles semblent se corréler à certaines anomalies observées en Na, La, Al_2O_3 suggérant que les teneurs actuelles sont des concentrations de deuxième génération d'origine métamorphique (?).

Un modèle magmatique et métamorphique de la distribution des éléments compatibles et des platinoïdes sera présenté.

Figure 1





DESTRUCTION OF A PASSIVE MARGIN I:

EVOLUTION OF THE

QUEBEC / NORTHERN VERMONT OROGEN

Chairperson:

Philip H. Osberg

THE SUTTON MOUNTAINS ANTICLINORIUM: ITS RELATIONS WITH ADJACENT UNITS

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The termination of a major structure of the Appalachian Humber zone, the Sutton Mountain Anticlinorium (SMA), was mapped by the author at the scale of 1: 20 000 from 1984 to 1987 (Marquis, 1984, 1985a, 1985b, 1985c, 1986, and 1987). This mapping of the Richmond area, in the Québec Eastern Townships (figure 1) was financed by the Ministère de l'Énergie et des Ressources du Québec.

The SMA is made of two Cambro-Ordovician units, the Oak Hill Group and the Sutton Suite (Table I). In the Richmond area, the Oak Hill Group consists of 8 formations situated in the chlorite zone of metamorphism. The Tibbit Hill (TH), Call Mill (CM) and Pinnacle (PI) Formations define the lower Oak Hill Group, while the White Brook (WB), Gilman (GI), Dunham (DU), Sweetsburg (SW) and Melbourne (ME) Formations define the upper Oak Hill Group. The lower Oak Hill Group represents an intracratonic rift accumulation of alkaline volcanics (TH) overlain by chloritoïd bearing schist (CM) derived from laterisation of the volcanics, and quartzite (PI) representing a braided delta deposit. The upper Oak Hill Group, a marine nearshore and platform sequence consists of dolomitic sandstone and phyllite (WB, DU), green phyllitic siltstone (GI), and black phyllite interbedded either with quartz sandstone (SW) or graphitic limestone (ME).

The Sutton Suite consists of intensely deformed metamorphic schists of a slightly higher metamorphic grade then the Oak Hill Group rocks. In the Richmond area, we have mapped three different schists; a green chlorite + epidote + calcite schist of volcanic origin interlayered with a quartz + chlorite + muscovite metasedimentary schist, and a layered quartz + chlorite + muscovite + albite schist. The tholeiitic composition of the chlorite + epidote + calcite schist (Table 2) and its stratigraphic relationships with the metasedimentary schists suggest that it is younger than the Tibbit Hill Formation of the lower Oak Hill Group. We also propose that the metasedimentary schists we have mapped in the Richmond area could represent deeper marine equivalents of the Gilman and Sweetsburg Formations of the Upper Oak Hill Group.

The contact between the Sutton Suite and the Oak Hill Group displays complex imbrication of Oak Hill's Sweetsburg lithology with chlorite + epidote + calcite schist, and quartz + chlorite + muscovite schist of the Sutton Suite. The chlorite + epidote + calcite schist is locally completely altered to a talc + tremolite schist suggesting intense fluid circulation and a tectonic contact, interpreted as the northern extension of the Mansville phase of the Sutton area.

The contact between upper Oak Hill Group lithologies (GI, DU, SW) and thinly bedded phyllite and limestone of the Ordovician Stanbridge Group along the west flank of the SMA is defined by a tectonic melange. Northward, the Stanbridge Group disappears and the SMA is in contact with the Bulstrode Formation, a middle to late Ordovician graphitic limestone similar and contemporaneous with the Melbourne Formation, the uppermost unit of the Oak Hill Group. These two units represent contemporaneous and similar lithologies, distinguished mainly on their different tectonic settings. The Bulstrode forms a large moderately folded nappe of the Appalachian foreland, while the Melbourne, restricted to a few square km in the SMA, is more intensely deformed.

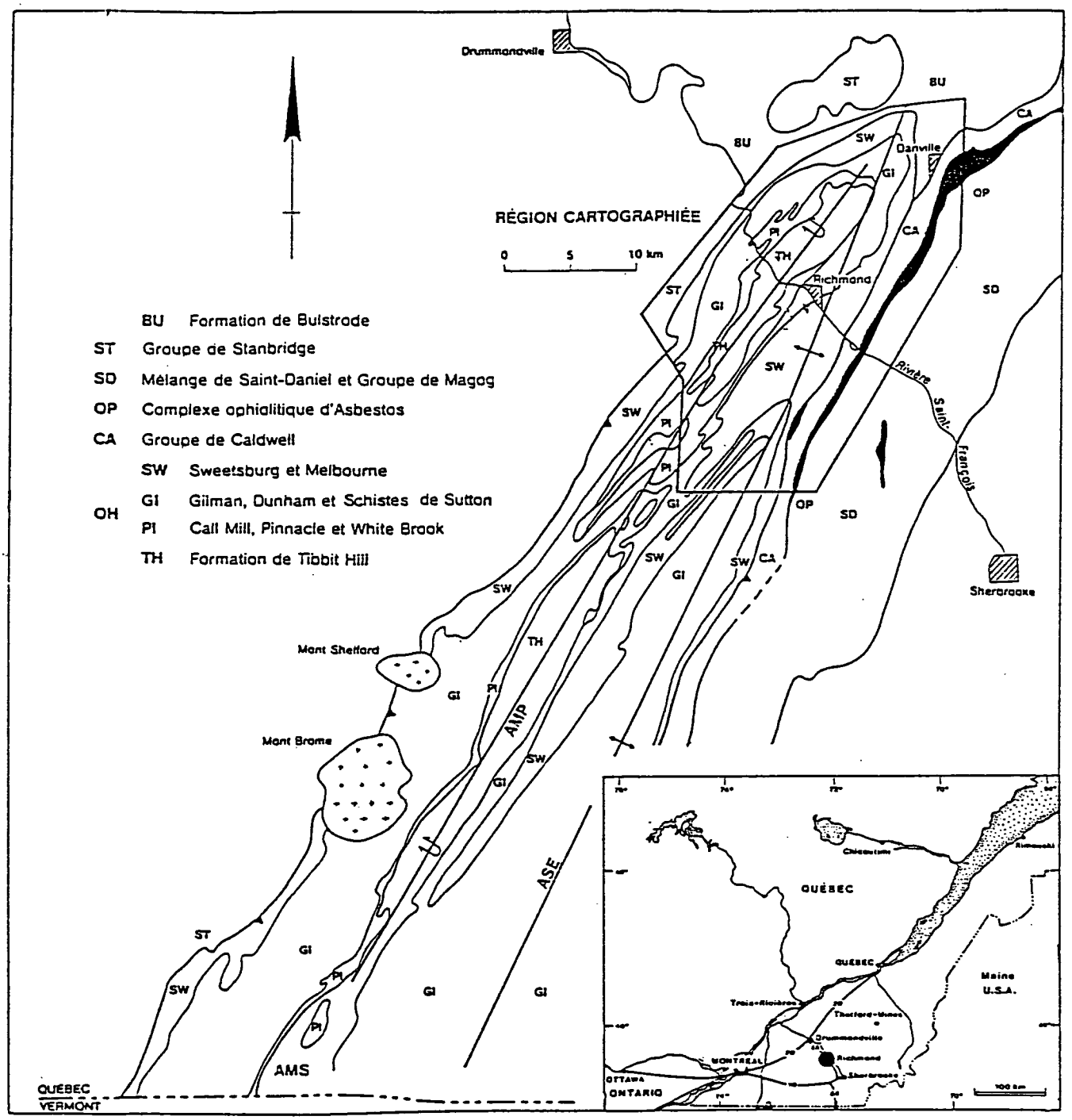

Figure 1: Regional geology showing the area mapped by the author between 1984 and 1987. Geology after Osberg (1965) and Hubert et al. (1977). List of symbols: Oak Hill Group (OH), Sutton Mountains Anticlinorium (AMS), St-Etienne Antiform (ASE), Pinnacle Mountain Anticline (AMP). Ultramafic rocks are shown in black.

The fault traced on the west flank of the SMA has to wrap around the nose of this shallow NE plunging anticlinorium in the Danville area, considering that the SMA is thrusted over the Bulstrode, so this fault is a major décollement cutting through the Bulstrode-Melbourne assemblage. The hanging wall of this décollement, the Bulstrode, moved further to the northwest while the footwall, the Melbourne, remained attached to the Upper Oak Hill. Absence of the Melbourne Formation in the Oak Hill Group type locality, near Sutton, may suggest that this décollement is cutting at a deeper level in the Sutton then in the Richmond area.

The relationship between the Bulstrode and the Stanbridge is best exposed outside the map area near Kingsey Falls, where the Stanbridge is thrusted on top of the Bulstrode. This fault probably postdates emplacement of the SMA over the Bulstrode.

On the eastern flank of the SMA, from north to south, the upper Oak Hill Group (GI, SW, ME) and the Sutton Suite schists are in contact with turbiditic sandstone, siltstone and phyllite of the Caldwell Group. This contact is marked by a major steeply dipping fault zone, along which pegmatitic gabbro, ophiolitic mélange, pyroxenite and serpentinized ultramafic rocks occur sporadically. Poor exposure of the fault zone, precludes determination of its relative motion. Nevertheless, it seems reasonable to consider this fault as a major Taconic thrust. Of course, we cannot dismiss the hypothesis that it could have been reactivated during the Acadian Orogeny, to accommodate backthrusting and backfolding known to have occurred in the adjacent SMA, a delamination process associated with emplacement of crustal wedges at depth.

TABLE I: Stratigraphy of the rock units in the Richmond area.

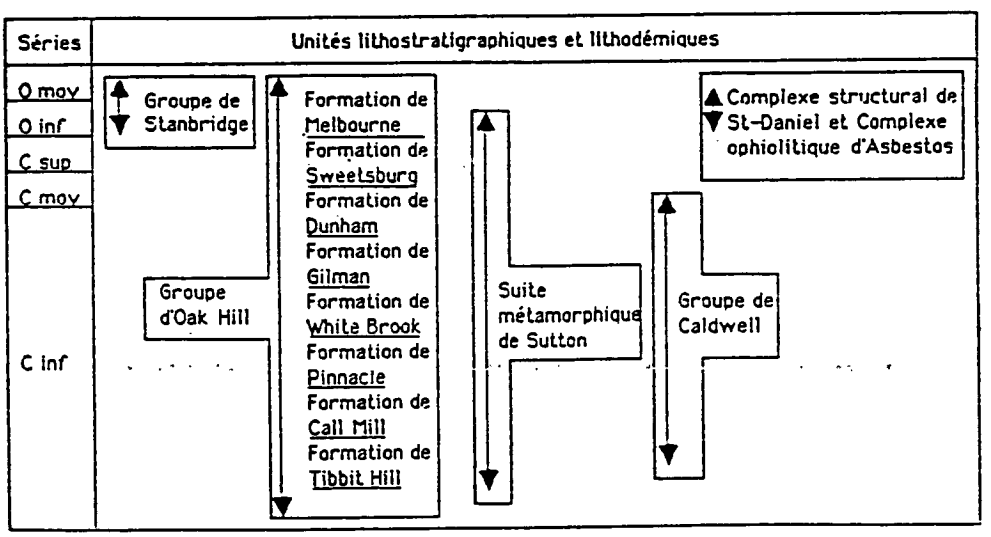

TABLE II: Lithogeochemistry of the chloritic schists from the Tibbit Hill Formation and Sutton Suite.

Unit	TIBBIT HILL		SUTTON			
Lab number	85-016884	85-016887	87-018379	87-018383	87-023186	87-023187
field number	85-BG-114	85-BG-196	87-CC-156	87-RM-268	87-CC-230	87-RM-385
SiO2	44.10	49.80	47.10	47.80	46.70	48.50
Al2O3	16.90	13.40	15.70	16.30	13.70	12.80
Fe2O3t	14.70	12.90	14.90	14.60	15.40	16.60
MgO	2.14	2.59	8.18	8.01	6.15	7.59
CaO	4.57	11.40	5.08	5.34	10.90	5.02
Na2O	0.73	3.01	2.11	0.83	0.56	4.22
K2O	7.79	0.29	0.01	0.01	0.07	0.05
TiO2	3.66	2.97	1.88	1.59	1.62	2.02
MnO	0.07	0.27	0.20	0.25	0.21	0.23
P2O5	1.68	0.41	0.11	0.16	0.14	0.17
PAF	3.09	2.05	5.10	5.35	3.15	2.47
La	112	35	5	10	7	9
Ce	217	73	48	15	12	14
Pr	<2	<2	<10	<10	<2	<2
Nd	395	265	80	80	105	120
Sm	8	<2	<2	<2	<2	<2
Eu	10	5	2	4	5	5
Dy	21	4	8	8	8	10

TACONIAN STRUCTURES FROM THE HINTERLAND OF THE EASTERN TOWNSHIPS OF QUÉBEC: THE RESULT OF CONTINENTAL (A-TYPE) SUBDUCTION ?

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The peculiar style of Taconian deformation encountered in the Québec Appalachians has long been a puzzling problem in reconstructing the tectonic evolution of the hinterland of the Eastern Townships. Early workers all report cleavage fans and/or backfolds (Rickard, 1965; Osberg, 1965) but did not have mechanical models that can explain such a structural style. The advent of plate tectonics provided the theoretical background that lead St-Julien & Hubert (1975) to relate the deformational style of the Taconian hinterland with a complex subduction history that culminated in a continent-island arc collision. However, there is growing evidence that the Cambro-Ordovician island arc complex (Ascot/Weedon Complex ?) did not collide with the continental margin during the Taconian orogeny (Tremblay, this volume). Moreover, this model does not explain the mechanical processes resulting in the development of both synthetic and antithetic folds/thrusts in the Taconian hinterland of the Eastern Townships of Québec.

Recent detailed mapping by the author in the Brome Lake area (figure 1a), coupled with seismic reflection data, provides new insights on the overall tectonic evolution of the Eastern Townships hinterland. The Humber zone of Québec (or "continental domain") was subdivided into external and internal domains by St-Julien & Hubert (1975). The study area is mostly underlain by metamorphosed sequences of the internal domain. These rocks are interpreted to have been deposited during rifting and subsequent passive margin development along the western margin of a Cambrian Iapetus ocean. They structurally overlie more proximal shelf and slope deposits to the west. To the east, they are juxtaposed against assemblages of the Dunnage zone along the Baie Verte-Brompton Line (Williams & St-Julien, 1982; see also Tremblay, and Cousineau, this volume).

Lithotectonic Assemblages

As a result of this study, the internal part of the Humber zone has been subdivided into three lithotectonic assemblages which are briefly summarized below (figure 2):

I- The westernmost assemblage comprises the volcanic/sedimentary facies of the **Oak Hill Group** (Clark, 1936). This assemblage includes a basal sequence of (bimodal) alkalic flood basalts which is overlain by rift-clastics that are interpreted to record two discrete phases of crustal extension (Colpron, 1989a). The rift sequence is in turn overlain by "clean" arkosic sandstones, dolomitic marble and a turbiditic siltstone/shale assemblage. This post-rift sequence is interpreted to record sedimentation in the shelf-edge/slope part of a thermally subsiding, transgressive passive margin.

3 project funded by the Ministère de l'Énergie et des Ressources, Québec.

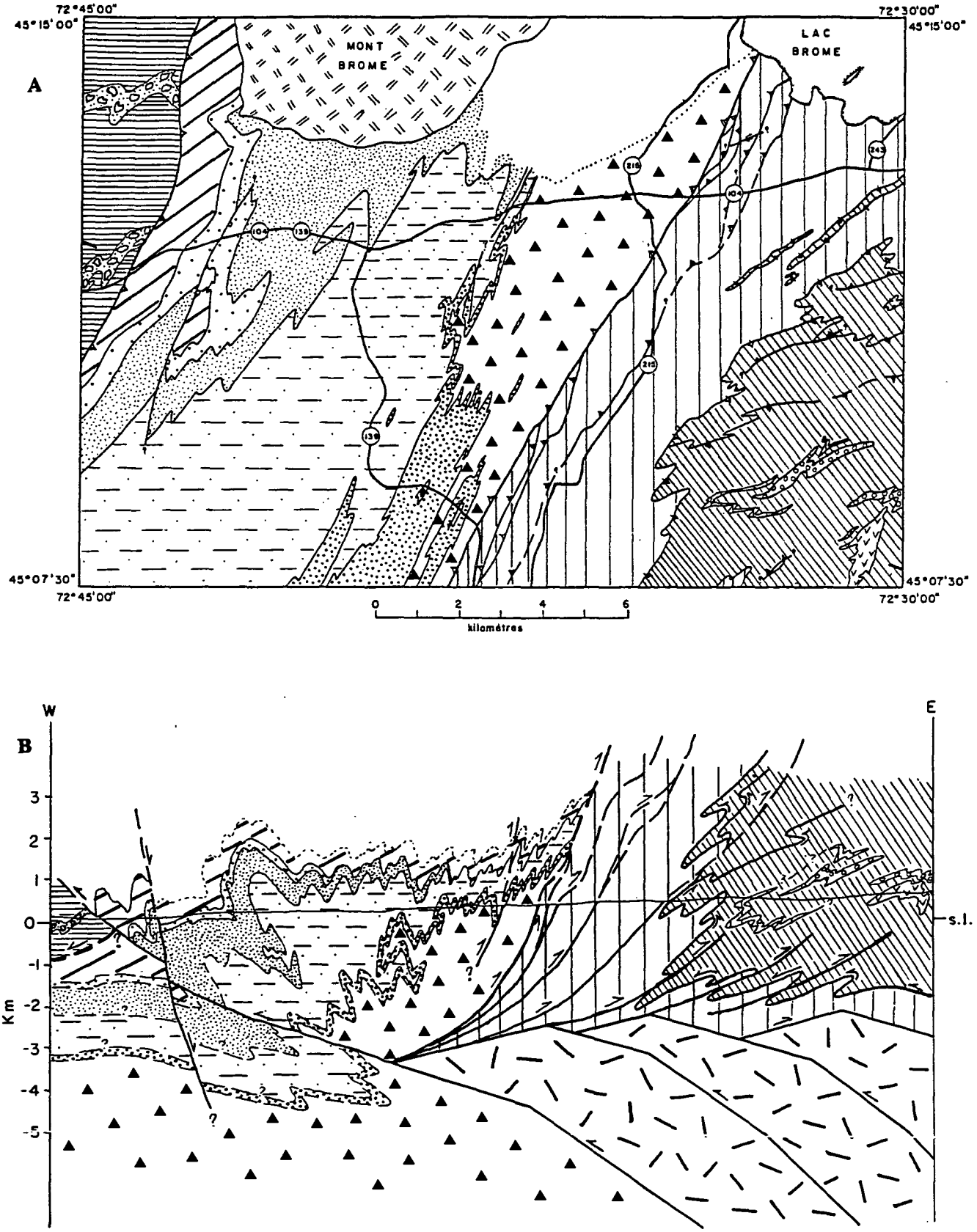

Figure 1: Schematic geologic map (a) and cross section (b) of the Brome Lake area. *Patterns:* **Oak Hill Gp.:** triangles = Tibbit Hill Fm.; coarse stippled = Pinnacle-West Sutton Fms.; dashed lines = lower Frelighsburg Fm.; fine stippled = upper Frelighsburg Fm.; dots = Cheshire-Dunham Fms.; wide diagonal lines = Sweetsburg Fm.; **Mansville Suite:** vertical lines; **Sutton Suite:** diagonal lines; and **Stanbridge Gp.:** horizontal lines; **Mont Brome (monterigian):** double dashed; **Quaternary:** blank.

II- To the east, the **Mansville Suite** (Colpron, 1989b, in progress) is characterized by a post-rift cover in many ways similar to that of the Oak Hill Group. This post-rift section overlies various rift assemblages confined to discrete fault slices. The rift sequences differ from the Oak Hill Group in their sedimentary facies and scarcity of volcanic rocks.

III- The easternmost assemblage corresponds to the highly deformed/metamorphosed rocks of the **Sutton Suite** (Marquis, 1989, in progress; this volume). Although a rift "stratigraphy" similar to that of the Mansville Suite and the Oak Hill Group can be interpreted in the metasedimentary rocks of the Sutton Suite, several lines of evidence, in particular the tholeiitic signature of the metabasalts and the presence of ultramafic rocks, suggest that they were outboard (i.e. east) of their present day position, resting on thinned Laurentia basement. The observed relationships between the different lithic packages are believed to result from tectonic processes. Part of the Sutton Suite is actually interpreted as a tectonic mélange involving both rift and post-rift lithologies, developed at (or near) the toe of an advancing accretionary prism (Colpron & Armstrong, 1988).

East of the Sutton Suite, but west of the Baie Verte-Brompton Line, lie two belts which are excluded from the present discussion: the Caldwell Group and the Ottauquechee Formation (Slivitzky & St-Julien, 1987; Lamothe, 1979).

Structural Style

The deformational history of the Eastern Townships hinterland, as recognized in the Brome Lake area, is marked by three generations of structures. The third phase structures (D_3), which are common to all lithotectonic zones, are characterized by a steeply east-dipping spaced cleavage axial planar to open folds and undulations. The "intensity" of D_3 structures rapidly decreases westward, D_3 folds being ubiquitous in the Sutton Suite and practically lacking in the Oak Hill Group.

The second generation structures (D_2) dominate the map pattern in all lithotectonic zones. In the western part of the Oak Hill Group, D_2 is characterized by a moderately to steeply east-dipping slaty-cleavage, or anastomosing pressure-solution cleavage, axial planar to upright to west-verging, open to tight, cylindrical folds, plunging shallowly to the north. These structures gradually pass eastward into a "syn-metamorphic", vertical to steeply west-dipping, crenulation cleavage, axial planar to tight to isoclinal, non-cylindrical, east-verging folds. Along the contact with the Mansville Suite, namely the Brome Thrust, the foliation is locally mylonitic and the D_2 folds plunge moderately to the north or to the south.

In the Mansville Suite, the D_2 structures are characterized by a closely-spaced cleavage, marked by more intense mica crystallization. It is sub-parallel to mylonitic zones marking the boundaries between discrete fault slices. Although the mylonitic foliation is well preserved in outcrops, several lines of evidence indicate that it was overprinted by late-D_2 flattening of the shear zones. The D_2 folds are strongly non-cylindrical and are best described as sheath-like structures. The progressive syn-D_2 rotation of fold axes within both the Oak Hill Group and the Mansville Suite result in an apparent "basin and dome" map pattern.

In the Sutton Suite, the D_2 cleavage is only weakly developed and is sub-parallel to the strong D_1 schistosity defining a composite foliation. D_2 sheath folds appears to have a SSE vergence and shallow plunges to the west. Minor D_2 shear zones indicate a NNW over SSE relative transport.

The D_1 structures are best developed in the Sutton Suite where they are defined by a strong composite foliation, marked by numerous quartz veins and a compositional layering. A number of pre- to early-metamorphic faults are interpreted to be sub-parallel to the D_1 foliation. These faults are best marked by serpentine bodies and unit truncations. In both the Mansville Suite and the Oak Hill Group the D_1 structures are characterized by a bedding-parallel schistosity axial planar to rarely observed isoclinal folds. Hook-and-crescent map patterns are interpreted to result from the interference of D_1 and D_2 folds. This type of interference pattern therefore requires that D_1 folds had axial planes dipping moderately to shallowly to the east (presumably) and axes sub-parallel to the superposed D_2 axes. When projected into a cross-section (figure 1b), the D_1 folds of the Oak Hill Group all appear to "climb" toward the western closure of a large west-facing recumbent fold, further suggesting that the Oak Hill experienced an early nappe-stage.

Timing of Deformation

Field relations suggest that the different phases of deformation were propagating from east to west. Available K/Ar dating (Rickard, 1965) supports this view, as muscovites along the D_2 cleavage in the Oak Hill Group appear to be contemporaneous with those along the D_3 cleavage in the Sutton Suite (440±20 Ma and 420±20 Ma, respectively). These cooling ages are interpreted to record a Taconian metamorphism.

Tectonics of a Taconian A-type subduction zone

Hodges et al. (1982) describe an A-type subduction zone as one in which continental crust is actively involved in subduction processes. In such cases, the continental crust is usually sliced up, possibly as a mean to reduce the basement buoyancy.

In the Québec reentrant, the presence of basement imbrications at depth is now a well known feature (St-Julien et al., 1983; Bardoux & Marquis, this volume; and unpub. seismic surveys, University of Vermont). It is proposed here that these imbrications are responsible for most of the supracrustal deformation observed in the Taconian hinterland of the Eastern Townships. The introduction of "rigid" basement slices is interpreted to result in the delamination of the supracrustal cover rock sequence (figure 1b).

The structural evolution of the Eastern Townships Taconian hinterland, as documented in the Brome Lake area, can therefore be summarized as follows:

1) The overriding of the Laurentian passive margin by a Taconian accretionary prism (part of the Sutton Suite ?) induced the development of large recumbent folds in the Mansville Suite and the Oak Hill Group, possibly in late-Arenig/early-Llanvirn.

2) Continued westward compression resulted in the introduction of basement wedges that created antithetic structures in the supracrustal cover sequence, as the cover sequence reacted passively relative to basement imbrications.

3) Late compression and rising of a basement duplex resulted in the development of the arch-like anticlinorial (D_3) structure of the Sutton Mountains, while antithetic thrusting/folding was still going on in the Oak Hill Group.

4) Finally, break-through of a basal décollement (floor thrust of lowermost basement slice ?) resulted in westward transport of the hinterland over shelf and shelf-edge sequences to the west.

In this model, only the earliest deformation observed in the Sutton Suite (part of which is interpreted as a tectonic mélange) appears to be related to "true" accretionary processes thought to have occurred in a B-type subduction setting.

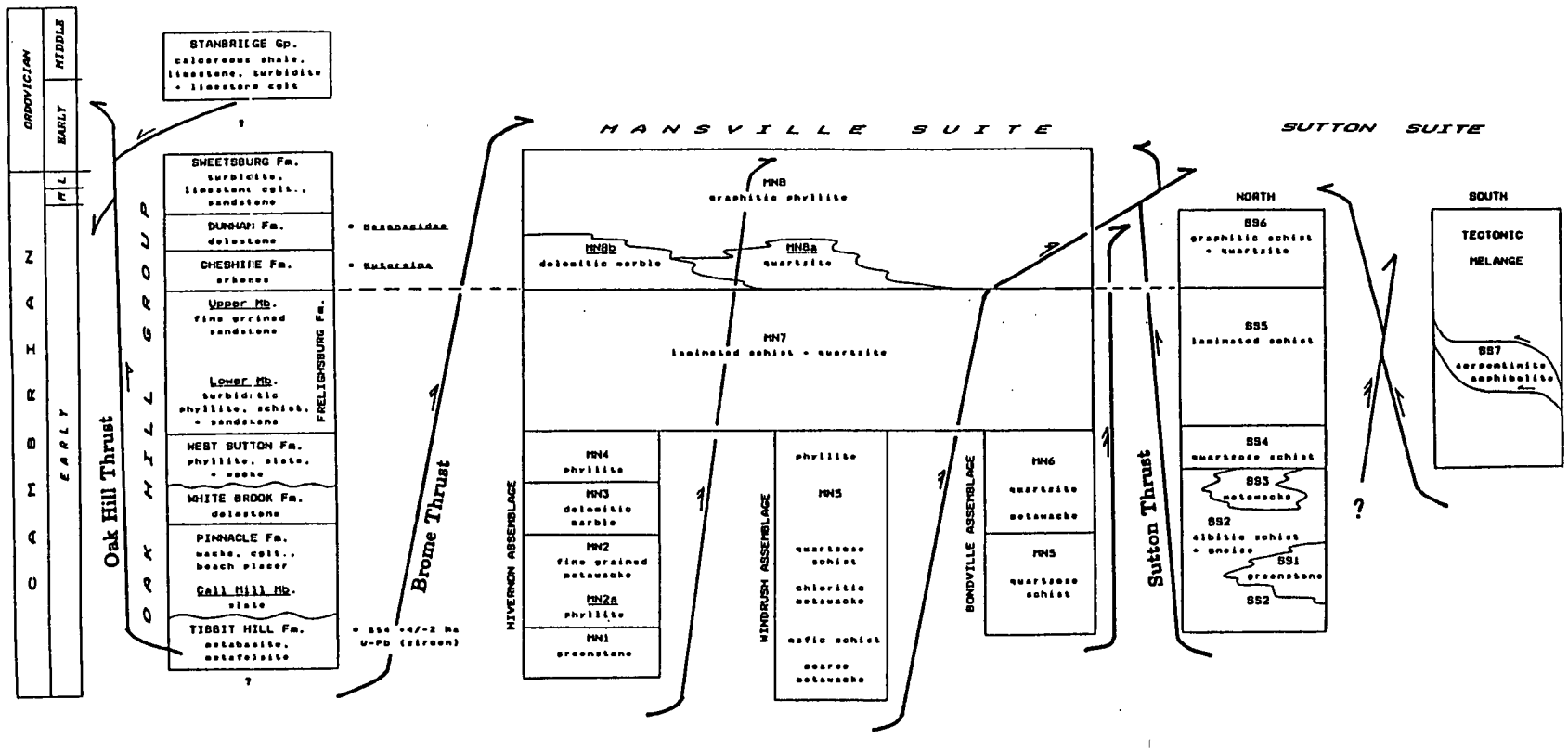

Figure 2: Lithotectonic Assemblages of the internal part of the Humber zone in southern Québec. The zircon age for the Tibbit Hill Fm. is from Kumarapeli & Pintson (this volume); the fossil ages from Clark (1936). Relative age of the thrusts as follow: D_1 = single arrow; D_2 = double arrow; D_3 = full arrow.

HINTERLAND TECTONICS OF THE VERMONT - QUÉBEC APPALACHIAN OROGEN

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STRATIGRAPHIC BIAS

Reconstruction of the polyphase deformation in the Vermont-Québec hinterland is hindered by accurate palinspastic restoration of lithostratigraphic units which comprise the orogen. This analysis of hinterland tectonics necessitates that certain assumptions of the stratigraphic succession be made since reconstructions not based on a restorable passive margin encompassing both rift and drift stage sequences are likely to be incorrect. The following five controversial assumptions, are made in restoring the lithostratigraphy:

1) the Sweetsburg Group of NW Vermont and the upper part of the Oak Hill Group of Québec correlate at least in part, with the Ottauquechee Fm. of the Eastern Townships ultramafic belt and central Vermont (Osberg, 1965; Marquis, this volume);

2) the Cambrian sandstone shale facies of St. Julien and Hubert (1975; includes the Caldwell Fm., Armagh Group; Granby Group, etc) are considered to represent late stage rift sequences of probable age, if not lithic equivalents of the Frelighsburg (Charbonneau, 1980) and lower Gilman (P. St. Julien, personal communication, 1988);

3) the entire Camels Hump Group of Doll et al. (1961) with the exception of certain carbonaceous phyllite horizons and the Hazens Notch Formation correlates either directly with the lower part of the Oak Hill Group of Clark (1936; Tibbit Hill, Pinnacle, White Brook, Fairfield Pond [=West Sutton and Frelighsburg]) and equivalents or indirectly as their more deformed and outboard counterparts (Underhill, Mt Abraham, Pinney Hollow);

4) the Stowe Formation is included with the units listed in (3) on the basis of lithic similarity and possible structural continuity with the Pinney Hollow (Armstrong, this volume). The geochemical signature of basalts in the Stowe argue further that the Stowe be restored outboard from the craton relative to the Pinney Hollow (Coish, et al, 1986);

5) the Hazens Notch formation is not an original restorable stratigraphic unit of the passive margin; rather, it represents a mix of lithologies including rift facies, drift facies and ultramafic material presumably derived from Iapetus during closure (Doolan, 1988). The Rosaire formation of the Notre Dame mountains of Québec includes lithologies identical to those referred to as Hazens Notch in Vermont.

MAJOR STRUCTURAL FEATURES

The reentrant has had profound control on the distribution of the lithostratigraphic succession of the original passive margin as well as the distribution of the present day structural features of the orogen derived during Paleozoic deformation. The Vermont - Québec hinterland stretching from the Massachusetts border to Richmond, Québec can be subdivided on the basis of differences in structural trends, allochthon development, basement involvement, metamorphism and vergence into two parts of approximately equal size (see cover map). The two halves are here simply referred to as the "Québec arm" and the "Vermont arm" of the orogen to emphasis the importance of the original reentrant shape in controlling the effects of closure of the system. In reality, the Québec arm extends well into Vermont to approximately the Winooski River transect (Thompson and Thompson, this volume).

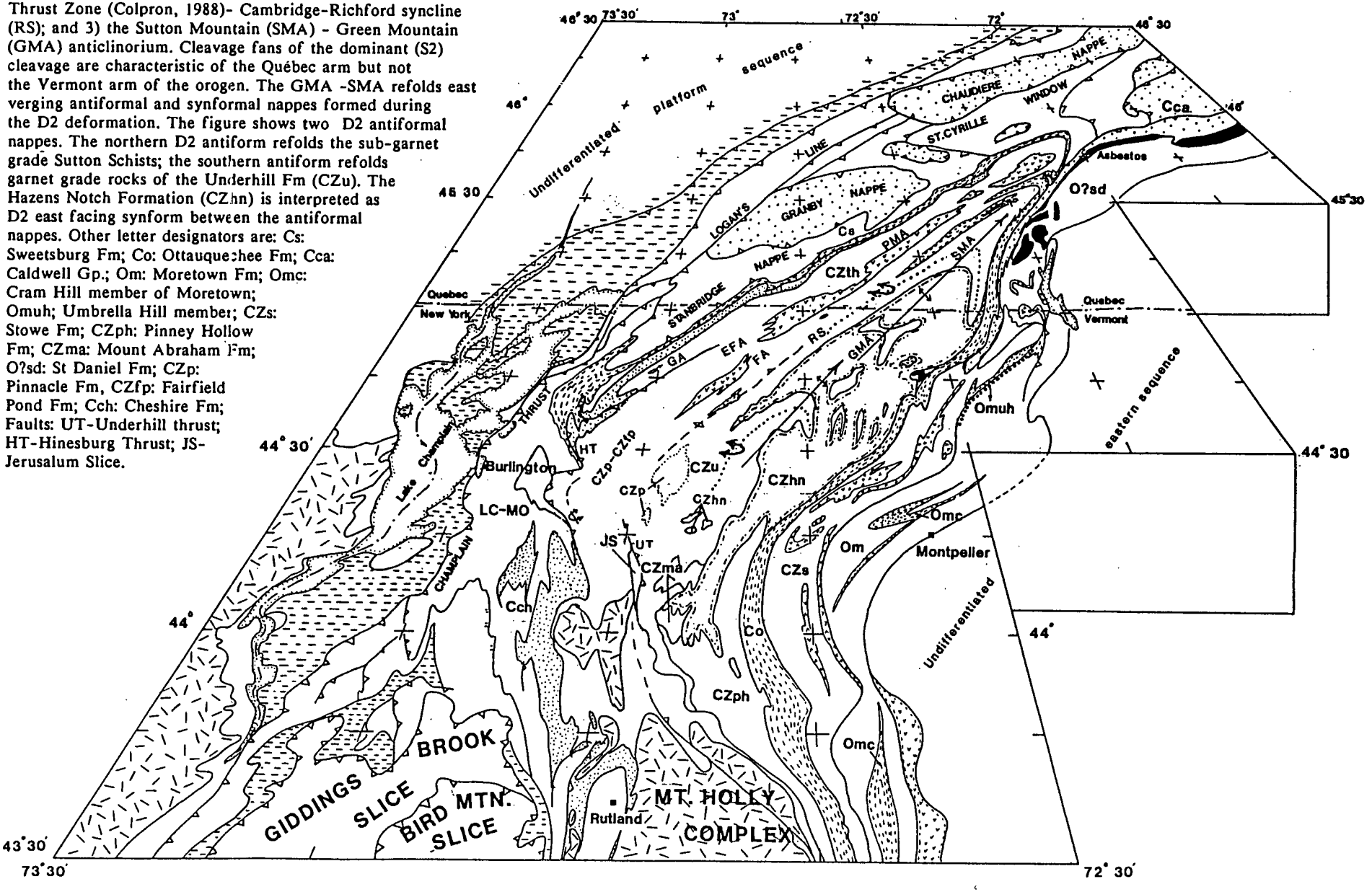

Figure 1. Geology of the Québec arm of the Vermont-Québec Appalachian orogen. The major structural features of the Québec arm hinterland include: 1) the Pinnacle Mountain (PMA) - Enosburg Falls (EFA) - Fletcher (FA) "anticlinorium"; 2) the Mansville Thrust Zone (Colpron, 1988)- Cambridge-Richford syncline (RS); and 3) the Sutton Mountain (SMA) - Green Mountain (GMA) anticlinorium. Cleavage fans of the dominant (S2) cleavage are characteristic of the Québec arm but not the Vermont arm of the orogen. The GMA -SMA refolds east verging antiformal and synformal nappes formed during the D2 deformation. The figure shows two D2 antiformal nappes. The northern D2 antiform refolds the sub-garnet grade Sutton Schists; the southern antiform refolds garnet grade rocks of the Underhill Fm (CZu). The Hazens Notch Formation (CZhn) is interpreted as D2 east facing synform between the antiformal nappes. Other letter designators are: Cs: Sweetsburg Fm; Co: Ottauquechee Fm; Cca: Caldwell Gp.; Om: Moretown Fm; Omc: Cram Hill member of Moretown; Omuh; Umbrella Hill member; CZs: Stowe Fm; CZph: Pinney Hollow Fm; CZma: Mount Abraham Fm; O?sd: St Daniel Fm; CZp: Pinnacle Fm, CZfp: Fairfield Pond Fm; Cch: Cheshire Fm; Faults: UT-Underhill thrust; HT-Hinesburg Thrust; JS-Jerusalum Slice.

THE QUÉBEC ARM

The major structural features of the Québec arm of the hinterland are from west to east the Enosburg Falls-Pinnacle Mountain anticlinorium (EPMA), the Mansville Thrust Zone (or Cambridge-Richford syncline; CRS) and the Green Mountain/Sutton Mountains anticlinorium (GMA). Folds and faults in the northern end of the Québec arm verge to the **southeast** across the entire hinterland and are refolded along the GMA to downward facing structures just west of the Baie Verte Brompton line (Marquis, this volume). Further south (e.g. Brome Lake; Colpron, this volume), a well developed cleavage fan displaying west verging, upright, and east verging structures is found. The transition from west to east vergence occurs just west of the EPMA. In northern Vermont, along the Lamoille River traverse, the transition is located near the CRS (Doolan, 1988; Mock, in progress). Further south along the Winooski River traverse, the transition lies east of the CRS but west of the Hazens Notch Formation (Thompson and Thompson, this volume). Throughout the Québec arm the SE verging structures and cleavage fans are D2 in relative age of deformational events. Metamorphism increases progressively to the south along strike of the major anticlinoria. This, in combination with the predominant northerly plunge of fold axes of D2 and D3 age, and the continuity of major structural features (EPMA, CRS, and GMA) justifies using a down plunge view of small scale compilation maps to reconstruct cross sections of the Québec arm. Using this technique, the Hazens Notch Formation structurally underlies an east facing anticlinal nappe composed of Bonsecours schist in the Sutton Mountains. To the south the Hazens Notch structurally overlies garnet grade east facing anticline of the Underhill (Thompson and Thompson, this volume; see cover map). The Hazens Notch Formation is interpreted as an east facing D2 recumbent syncline (P. Thompson, 1975) which is cored by ophiolite sole, aureole and blueschist assemblages near Eden Mills and Tillotson Peak (Hollis Gale, 1980; Laird and Bothner , 1986; Laird et al., this volume). If this simplistic view of the orogen is correct the Québec arm of the orogen is dominated by large scale backfolds which predate the D3 structures associated with the GMA (Doolan, 1988).

THE VERMONT ARM

The Vermont arm of the hinterland by contrast is dominated by syn-metamorphic east over west faults, significant basement involvement and comparatively higher temperatures of recrystallization than the rocks exposed in the Québec arm (Stanley and Ratcliffe, 1985; Stanley, this volume). The first order structural features of the Québec arm (EPMA, CRS and GMA) are poorly defined and difficult to recognize. While these differences make correlations in stratigraphy and relative structural chronologies difficult between the two arms of the hinterland, the fact is that the differences are real. It remains a challenge to incorporate the features of both arms into a consistent retrodeformable model which reconstructs the differences in the original reentrant configuration and the contrasts in deformational history.

PROBLEMS

Many problems remain which will hopefully be addressed at this conference. The dilemna of the palinspastic restoration of the Ottauquechee Formation with respect to the hinterland to the west appears to be especially problematic. In the northern limit of the Québec arm the evidence is strong that the Ottauquechee is rooted from the west by backfolding. Stanley and Ratcliffe's (1985) model of the Ottauquechee/Stowe (=Rowe) representing an accretionary prism outboard of the North American passive margin has since been modified (e.g. Stanley and others, 1988a) to include both the Stowe and the Ottauquechee as distal components of the reconstructed passive margin. While this configuration is in accordance with the palinspastic restoration for northern Vermont, the problem shifts to the relationship between the present Ottauquechee belt and the suture between North American craton and Iapetus.

RECOGNITION OF THE IAPETUS-NORTH AMERICAN SUTURE

The suture between Iapetus and the North American craton is only rarely preserved along the orogen. In Vermont the suture is defined where the ophiolite tectonite sole exposed at Eden Mills is in tectonic contact with the Hazens Notch Formation. A well preserved aureole beneath the tectonite grades downward from garnet amphibolite to amphibolite to mica schist to greenstone (Belvidere Mountain Complex; Hollis Gale, 1980; Doolan et al., 1982). The serpentinized sole, aureole and Hazens Notch rocks at Eden Mills have been backfolded during the development of the strong amphibole lineation in the rock. The correlatives of the Belvidere amphibolite at Tillotson Peak to the northwest contain blueschist assemblages (Laird and Albee, 1981) and are interpreted as a detached sliver of the aureole in the Hazens Notch. It appears likely that the aureole, the blueschists, and the serpentine contacts with the Hazens Notch are the preserved fragments of the Iapetan suture. The backfolding in the aureole and the entire Hazens Notch Formation into an east facing recumbent syncline between non ophiolite bearing rocks of the Bonsecours and Underhill schists supports the view that the Hazens Notch formed as an accretionary melange below obducted Iapetan lithosphere and above the rift clastic section of the ancient North American passive margin. If the source rocks of the graphitic parts of the Hazens Notch/Rosaire lithologies are Ottauquechee correlatives, then the Ottauquechee rocks could be palinspastically restored to positions cratonward of the Hazens Notch. The present position of the Ottauquechee belt east of the Hazens Notch is difficult to reconcile without backfolding if the Hazens Notch indeed formed in the suture zone.

DISCUSSION

The northern, Québec arm of the hinterland formed at higher structural positions in comparison to the Vermont arm as evidenced by the present distribution of metamorphic isograds, the differences in basement involvement and the differences in the allochthon emplacement history . The large scale backfolding in the Québec arm of the hinterland is a diachronous event involving the rapid rise of blueschist assemblages during and subsequent to their formation at 468 Ma. (Laird and others, this volume). Continued D2 deformation involving backfolding affected rocks as young as the Melbourne formation (ca. 450 Ma.; Marguis, this volume) deformed at higher and more cratonward positions in the orogen. The complex deformational history of the Iapetus suture records the effects of both A-type and B-type subduction. This complex history contrasts sharply with rocks east of the BVB line.

The contrasts between the tectonic development of the Québec - Vermont hinterland may thus reflect the same fundamental process of subduction and collision of an irregular margin. Taken together the hinterland preserves the intermediate and deep crustal record of subduction tectonics unavailable to those working in present day orogenic belts.

L 2001: A NEW INTERPRETATION OF NORTHWESTERLY-DIPPING REFLECTORS ABOVE THE BASEMENT-CORED SUTTON-NOTRE-DAME ANTICLINORIUM NEAR THETFORD-MINES, QUÉBEC

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South of Québec City, the Appalachians consist of Cambro-Ordovician sediments and volcanics originally deposited on a rifted Grenvillian basement. Atlantic-type sedimentation started with rift clastics overlying calc-alkaline volcanics. Rift sediments are laterally and vertically grading into carbonates to the northwest and into turbiditic flyshoid sequences to the southeast. The whole stratigraphic sequence is gradually thickening to the southeast.

Destruction of this early Paleozoic extensional basin (Iapetus) started during the Taconic Orogeny in middle Late Ordovician time. Thin-skinned detachment of the supracrustals produced: northwesterly vergent recumbent nappes; tectonic melanges; and thrusts forming a foreland fold and thrust belt. Certain southeasterly verging nappes are superposed on these structures in the hinterland and may have formed as the result of the interaction in a southeast-dipping subduction zone context.

Further compression forced the imbrication and abduction of oceanic crust fragments mixed with accretionary prism melange and abyssal plain sediments. Though the Brompton-Baie Verte Line (BBL) represents the suture zone between the accreted oceanic terrane and miogeoclinal sediments it is still poorly defined in this part of the Québec Appalachians.

Closure of Iapetus during the Taconic Orogeny created considerable shortening of the miogeoclinal and oceanic assemblages causing major crustal thickening. Compression continued in late Devonian time during the Acadian Orogeny. Under greater lithostatic pressure and higher temperature due to the stacking of Taconic nappes, preexistent and new ductile shear zones were activated in the Grenvillian basement. Basement wedges thus formed were bounded above and below by thrust faults with opposite vergence, producing delamination at high structural levels along southeasterly verging backthrusts and forming large southeasterly vergent backfolds in the supracrustals of the hinterland.

Several basement wedges underneath the Québec Appalachians form distinct reflectors on seismic Line 2001. These appear to be stacked in a duplex style at about 6 km depth. It is proposed that these basement wedges created the large open synclinoria and anticlinoria mapped at the surface in the hinterland. This interpretation implies that: northwesterly dipping foliations; backfolds; and backthrusts of this region should be Acadian structures formed in response to deep crustal wedging.

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Grenvillian basement is known to occur at the surface underneath Rift assemblages in Vermont (Doolan et al., 1987). In Québec, the only well known basement sliver occurs in a major detachment surface 40 km to the northeast of the present study area near St-Malachie (Vallières et al., 1978).

Several interpretations of line 2001 have been made in the past. St-Julien et al. (1983) suggested the presence of a basement uplift underneath the Notre Dame Anticlinorium (NDA) and Bernard (1987) followed-up with a similar proposition underneath the St-Victor Synclinorium and Connecticut Valley - Gaspé Synclinorium. Each of these large basement wedges is bounded by northwesterly dipping reflectors at the top and southeasterly dipping reflectors at the base. This geometry suggests a conjugate shear system typical of a "rigid-plastic" to brittle behavior. Several of the northwesterly dipping faults occur at the surface without clear interpretation. The present paper proposes a new interpretation to these features, focussing particularly on the reflectors occurring on the northwest flank of the NDA.

Rocks underlying the NDA consist of complexly folded and imbricated sequences of Chlorite-garnet schists; chloritized metavolcanics; amphibolites; and serpentinites. The rocks are at biotite greenschist grade and were probably underneath 10-15 km of Ordovician overburden. Structural patterns suggest large scale northwesterly verging nappes superposed by younger southeasterly verging backfolds and northwesterly dipping foliations.

Flanks of the NDA are outlined by the presence of a strained serpentinite belt to the southeast and by highly altered and highly tectonized assemblage of predominantly Tibbit Hill volcanics (THV) to the northwest. The THV are the oldest of the Oak Hill Group units which normally rests unconformably over Grenvillian basement. These are bounded on either side by faults with unknown stratigraphic separation. This one should be in the order of several hundred meters as the THV are juxtaposed against younger Oak Hill-Rosaire Group rocks. Serpentinite fragments also occur in the strained zone along with THV.

At least three serpentinite bands and several aligned serpentinite pods straddle across the NE-SW axis of the NDA. We propose that those intensely deformed and highly altered (steatite) serpentinites carrying amphibolite fragments (Harvey Hill), from deeper structural levels, represent individual thrust surfaces. Displacement along these surfaces in not known but should be relatively large if we assume that serpentinites root into the oceanic domain, the suture of which (BBV) is currently located approximately 10 km to the southeast. The parallelism between serpentinite layers may be due to tight refolding or to thin repetitive imbricate slices. These layers are clearly merging with the serpentinite band along the southeastern flank of the NDA and they appear to line up with the serpentinite fragments along the northwestern flank of the NDA.

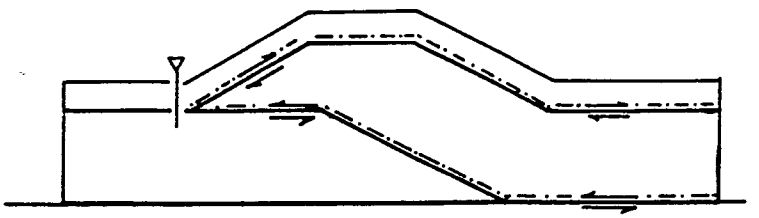
At the surface, the structural corridor bounding the THV on the northwestern flank of the NDA appears to be a late feature as it cuts older structural features and is parallel with the axis of the NDA. These later faults can be described by the following hypothetical geometries:

A - They may be northwest-dipping, northwest verging, thrusts that have been folded over the NDA which would be thereby a late structural window. None of the maps currently available are detailed enough to show that the northwest-dipping faults are closing around this metamorphic culmination (NDA).

B - The faults may be northwest-dipping southeast-vergent backthrusts.

We favor the second case and relate those features of the tectonic wedging of deep basement duplexes during the late tectonic compression affecting this part of the Appalachian Orogen in Acadian time.

Delamination



DESTRUCTION OF A PASSIVE MARGIN II:

NEW DEVELOPMENTS IN THE VERMONT APPALACHIANS

Chairperson:

Nicholas M. Ratcliffe

COMPARATIVE TECTONICS OF BASEMENT MASSIFS IN THE NORTHERN APPALACHIANS WITH SPECIAL REFERENCE TO THE GREEN MOUNTAIN MASSIF OF VERMONT

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Approximately 1.3 to 1.0 by-old gneisses of the Blue Ridge, Reading Prong, Hudson Highlands, Housatonic, Berkshire, Green Mountain and Lincoln Mountain massifs form a series of external massifs, marking cores of west to northwest verging fold-nappes and basement wedges. A second belt of similar-age rock forms the core of internal massifs in the Baltimore Gneiss domes, Manhattan Prong, Sadawaga and Chester-Athens domes. The structural geology, age of deformation and intensity of remobilization varies greatly along strike of the orogen reflecting both the extent and degree of deformation in the Taconic, Acadian and Alleghanian orogenies. Reuse of geologic structures, principally of faults, by reactivation tectonics may be widespread, and is especially prevalent in early Mesozoic extension.

Taconian collisional tectonics culminated in large-scale westward overthrusting and metamorphism-induced ductile deformation of North American sialic crust, following ejection and emplacement of the Taconic allochthons. Strongly lineated tectonites, regionally important fold-thrust tectonics and Barrovian dynamothermal metamorphism culminated this deformation that is well-expressed from Pennsylvania to Vermont. Late Taconian structures, folding of thrust faults, imbrication of allochthons and active uplift may have continued into the latest Ordovician in the foreland and along faults at the eastern margins of the massifs. In Late Ordovician through Early Silurian(?) a series of late to post tectonic granites and alkali-gabbros intruded the composite Taconian terrane, possibly related in part to thermal-lag following Taconian crustal build up.

Acadian dynamothermal metamorphism in the Green Mountains, the internal massifs, the eastern 1/3 of the Berkshire massif and Manhattan Prong produced thrust-faulted antiformal-domal features perhaps resulting from delamination and consequent disharmonic folding above reactivated Taconian thrusts.

Alleghanian deformation, principally transport of imbricate rigid slices, retransported basement massifs from New York southward. Finally Mesozoic extension, controlled by reactivation of Taconian and younger faults chiefly in the central Appalachians produced the form and distribution of early Mesozoic basins such as the Newark basin.

The Green Mountain massif and more easterly exposures of Proterozoic basement in the Chester-Athens and Rayponda domes contain many of the complexities present in areas to the south. However, the allochthony of the northern massifs and internal deformation by thrust faulting appears less than in exposures to the south. Through-going thrust faults of major significance do not appear to be present in the Green Mountains although a zone of concentrated thrust faulting probably does separate the basement of the internal domes from the massif proper. Variations of cover rock facies resting on the Green Mountains and on the internal massifs suggest that the ancient Lower Cambrian shelf and older pre-shelf deposits extended eastward at least to the eastern margin of the Rayponda dome and its cover wherever that paleodepositional site was. Important south to north facies variations in cover rocks leading into the Québec re-entrant should be considered in reconstructions of paleodepositional environments of the subsequent collisional tectonics. The general mismatch between stratigraphy and

volume of the Taconic allochthons in comparison to the "eastern" cover sequence suggest that the sedimentary rocks comprising the Taconic allochthons were deposited east of the exposed basement. Detailed mapping in the eastern cover from Connecticut north to Jamaica, Vermont, reveals numerous disruption of stratigraphic continuity and abundant evidence for thrust faults within the classical Tyson-Hoosac-Pinney Hollow sequence, or Hoosac-Rowe sequence of Massachusetts.

The oft-cited suggestion that younger metamorphism and deformation has destroyed or eliminated older structures in reworked basement rarely is true. Rather, the gloriously old and magnificently complicated Middle Proterozoic basement of the Appalachians is analogous to a palimpsest manuscript, that yields, with proper scrutiny and detailed geologic mapping, much of the ancient history of the orogen.

SHARED DEFORMATIONAL FABRICS IN BASEMENT AND COVER ALONG THE WEST FLANK OF THE SOUTHERN GREEN MOUNTAINS, VERMONT

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A persistent stratigraphy marked by lateral continuity of thin, distinct metasedimentary units is present in the Late Proterozoic-Early Cambrian Dalton Cheshire cover-sequence rocks along the west flank of the Green Mountain massif in southern Vermont. This stratigraphy has permitted the detailed mapping of Paleozoic structures and determination of their relative ages and deformational effects on the cover sequence. Two major episodes of deformation have been found. The first event produced a locally well-developed schistosity (S1) which is subparallel to bedding and is axial planar to minor recumbent folds (F1). Initial orientation of these folds is problematic due to later folding. The second event produced north-south-trending folds with moderately- to steeply east-dipping axial planes which deform the earlier fabric and are responsible for the conspicuous digitations in the basement-cover contact along the west flank of the Green Mountains near Arlington and Manchester, VT.

Post-Grenville structures in the basement rocks of the massif can be linked to the structures in the cover sequence. A pervasive shallow- to moderately dipping retrograde foliation in the gneisses north of Stratton Mountain is axial planar to minor recumbent folds in gneissic layering. Near the basement-cover contact these structures are subhorizontal to west-dipping and are believed to be contemporaneous with the early (S1) schistosity in the cover rocks. Upright F2 folds in cover rocks are well-developed near small high-angle reverse faults in basement near Bennington, VT. In addition, the axial surfaces of large-amplitude F2 folds in cover rocks parallel steeply-dipping, second-generation mylonite zones in underlying basement near Manchester, VT.

Muscovite from a phyllitic layer in the basal Dalton has yielded a $^{40}Ar/^{39}Ar$ age of 390-400 Ma, which has been interpreted as a Taconian cooling age (Ratcliffe and others, 1988). Textures in the dated sample consist of an early micaceous schistosity (S1) that is folded and transposed into a later crenulation cleavage (S2). Unstrained, recrystallized muscovite defines both the S1 schistosity along the noses of the crenulation folds and the S2 cleavage on the limbs, and is all considered to be of S2 generation. The second-generation structures in both the cover and basement are thus likely to be Taconian in age, and probably the first generation as well. Post-S2 upright folds produce broad dome and basin structures along the basement-cover contact and in the gentlydipping retrograde fabric in the basement; these are probably Acadian in age.

Coextensive fold sets and deformational fabrics in the cover and basement show that large-scale faulting and decoupling of cover from basement is not required to explain the structural geology along the southwest flank of the Green Mountain massif. Rather, the basement responded by semi-ductile shearing and small-scale faulting to accommodate strain accompanying folding of the cover sequence. The overall form of the Green Mountain massif in this area is a product of these deformations.

TECTONIC EVOLUTION OF THE SOUTHERN LINCOLN MASSIF AND RIFT-CLASTIC COVER SEQUENCE: CENTRAL VERMONT

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Introduction

The Lincoln massif, located in west-central Vermont (Fig.1, inset), marks the northernmost exposure of para-autochthonous basement within the New England-Québec Humber Zone. Such "external massifs" of the northern Appalachians (see also Ratcliffe, this volume) are located at the western edge of the most intensely deformed terrain of the Taconian hinterland (internal domain of the Humber Zone). Basement exposures are characterized by rocks of Grenville age metamorphism, exposed in anticlinoria and/or thrust sheets of primarily Taconian age. The presence of such massifs presents an opportunity 1) to observe rift-clastic rocks in contact with the basement on which they rest, and thereby investigate west-to-east facies relationships; 2) to study the mechanical role of "rigid" basement in Taconian deformation, where crustal levels were high enough to produce a contrast in behavior between basement, eastern cover, and western cover; 3) to investigate the link between style of Taconian deformation and earlier rift-related structures which may be recorded in the basement and lowermost rift-clastics.

The present study is located immediately west of the topographic crest of the Green Mountains, approximately 80 km (50 miles) southeast of Burlington. Middle Proterozoic gneissic rocks of the Lincoln massif are exposed in two north-south trending doubly-plunging anticlines; basement rocks are unconformably and, locally, tectonically overlain by eastern and western cover sequences. A syncline of western cover rocks separates the two basement anticlines in the southern part of the area (Fig. 1). Three, perhaps four, orogenic events which have contributed to the overall tectonic evolution are recorded in the Lincoln massif and its cover: the Grenville orogeny; the opening of Iapetus, and the Taconian orogeny. Although the effects of the Acadian orogeny are believed to be minimal, Acadian deformation may have contributed to some of the late structures observed in this area.

Middle Proterozoic rocks

The basement of the southern Lincoln massif is composed largely of felsic gneisses of varying texture and composition. Basement lithologies are more difficult to identify and compare toward the east, as greater Taconian deformation has progressively destroyed older textures and mineral assemblages. The **western Lincoln massif** (WLM) is characterized by 1) abundant fine- to medium-grained "granular" or weakly foliated qtz-plag-(bio)-(kspar) granitic gneiss; 2) well-foliated qtz-plag-bio-(kspar) gneiss; 3) distinct microcline augen gneiss characterized by 1-5 cm kspar megacrysts within a fine-grained quartz-biotite-plag-(kspar) matrix; 4) minor amounts of chlorite-(biotite)-epidote-quartz-(actinolite)-(plagioclase)-calcite schistose amphibolite, which is observed in sharp contact with basement lithologies 2 and 3. The contact between basement lithologies 1 and 2 appears to be gradational. Finally, blue quartzite is observed at a few localities.

The **eastern Lincoln massif** (ELM) is characterized primarily by lithology 1 of the WLM, which is in the ELM most commonly a coarse-grained white quartz-plagioclase-(biotite) gneiss, often intruded by tourmaline-bearing pegmatite. A sequence of plagioclase augen gneiss, foliated qtz-plag-bio gneiss and gneissic plagioclase-hornblende amphibolite is observed within the eastern part of the ELM. Contacts between these lithologies are sharp. Finally, sheared (Paleozoic deformation) mafic schist or amphibolite, perhaps similar to lithology 4 of the WLM, is observed along the eastern boundary of the

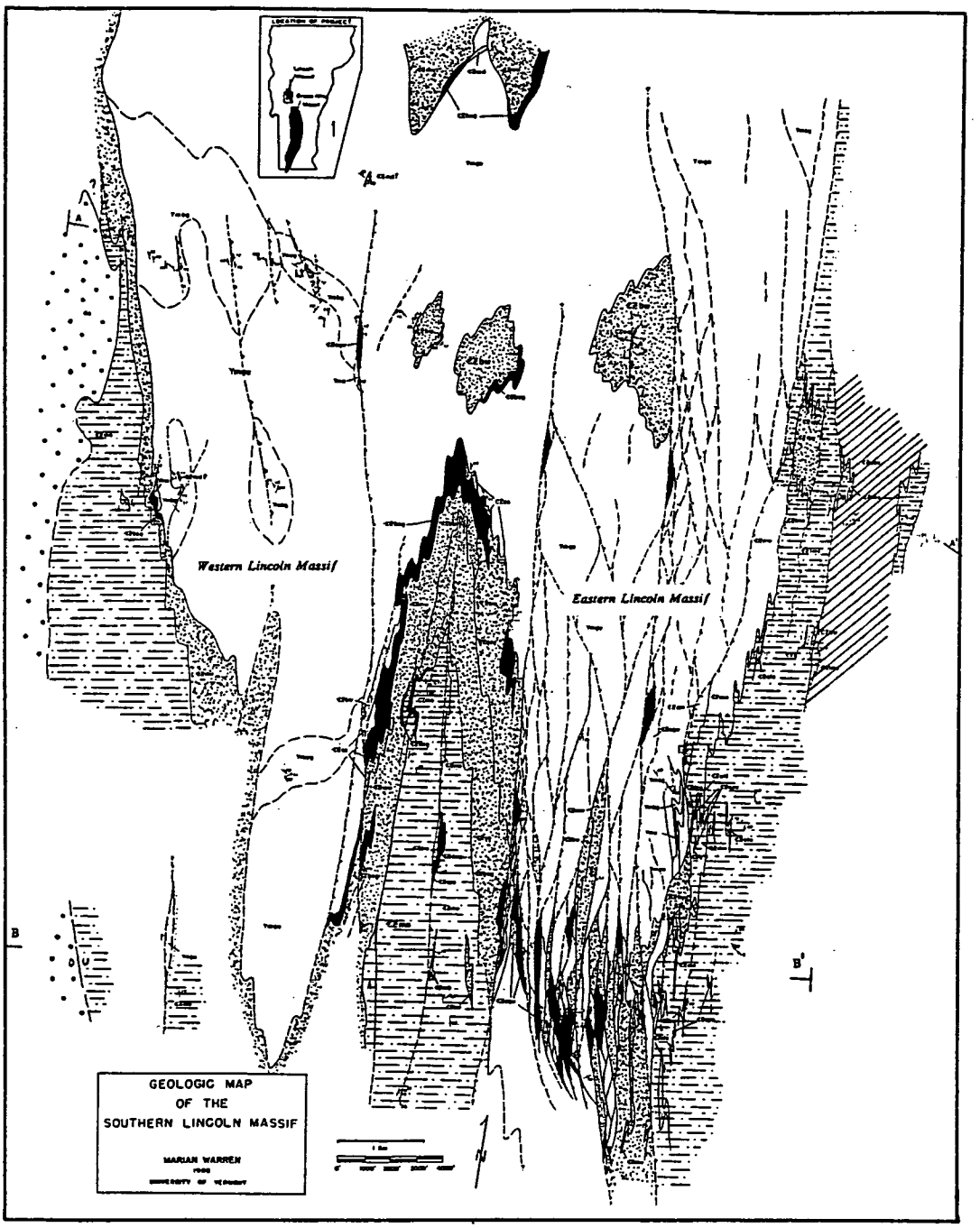

Figure 1: Generalized geologic map of the southern Lincoln massif, central Vermont. White = undifferentiated Grenville basement: solid = undifferentiated conglomerates: small dots = arkosic metawackes: dashed = quartz-muscovite metawacke, schist, or phyllite, locally graphitic: large dots = Cheshire Quartzite: diagonal rule = albitic schist. Line of cross-section (Fig.2) is B - B'.

ELM. Similar suites of basement rocks, with a similar distribution across the massif, are also reported in the northern Lincoln massif, where they are better exposed (DelloRusso and Stanley, 1986).

These gneissic rocks experienced at least upper-amphibolite facies metamorphism and accompanying deformation during the Grenville orogeny. To date, only one Grenville fabric has been recognized at any one locality in the basement; it is not known if there are multiple Grenville fabrics. The Grenville fabric is everywhere best identified by the presence of biotite, either as anastomosing seams, or as a compositional layering within quartzo-feldspathic gneisses; the Paleozoic fabric is characterized by muscovite and quartz in the retrograded felsic basement. At two localities in this study area, and also within the northern Lincoln massif (DelloRusso and Stanley, 1986), the biotite Grenville fabric is clearly truncated by the basement-cover unconformity, while the Paleozoic muscovite fabric cuts the unconformity. The Grenville fabric typically strikes NW-SE, in contrast to the more N-S striking Paleozoic fabrics.

Several-meter-wide mafic schists are observed within the basement which appear to cut the Grenville fabric, but are truncated by the basement-cover unconformity. These features are similar to mafic dikes interpreted by DelloRusso and Stanley (1986) as possible rift-volcanic feeder dikes. In addition, several mafic rocks previously mapped as retrograded Grenville amphibolite are located along newly identified Paleozoic fault zones which also involve boulder conglomerates; these rocks should also be considered as igneous material of possible younger age.

Rift-clastic cover

The rift-clastic cover sequences, although deformed by folds and faults, represent coherent sequences deposited on the basement of the Lincoln massif. To the west, this sequence includes all lithologies beneath the Cheshire Quartzite, whereas to the east, a similar rift-clastic sequence has been mapped east to the topographic crest of the Green Mountains. The syncline of cover rocks separating the two basement anticlines contains lithologies of the western sequence.

The eastern and western sequences are both characterized by 1) a lower section of coarse arkosic wackes and pebble conglomerates, which commonly contain cobble or boulder conglomerates at or near the base. The "basal" conglomerates are of two distinct types: 1) a more common, typically clast-supported cobble conglomerate with well-rounded, well-sorted clasts of quartzite, several gneissic lithologies, and metasediments, and 2) a less common, matrix-supported conglomerate containing a small percentage of poorly sorted, subrounded boulders of the immediately underlying basement lithology. The type 2) matrix is variable, consisting of interfingering arkosic wackes, mafic schist interpreted as possible volcaniclastic sediment, and layered dolostone or dolomitic marble. Conglomerate 1 is most commonly found along undisrupted unconformable basement-cover contacts, while conglomerate 2 is uniquely found along zones of basement-cover imbrication.

The wackes appear to become poorer in biotite and feldspar, and richer in quartz and muscovite upwards in the section, although the local heterogeneity of these deposits should be stressed. The wackes grade upwards into A) finer-grained meta-sandstone, meta-siltstone, or quartzite (west) or laminated quartz-muscovite schist or micaceous quartzite (east). This fine-grained quartz-rich material grades upward into B) phyllite with silty laminae or locally graphitic dark phyllite (west) and aluminous schist commonly characterized by quartz laminae, also locally graphitic (east). An important feature within both eastern and western sequences is the presence of lenses of dolostone and/or overlying dolostone-clast, rounded quartz pebble to cobble conglomerates at several stratigraphic horizons, with a major horizon between lithologic packages A and B. It is believed that the differences in the western and eastern sequences primarily reflect differences in metamorphic grade and intensity of deformation,

rather than dramatic differences in protolith. Relative amounts of feldspars, quartz, and micas change vertically, but do not appear to change significantly from west to east.

Significance of the basement-cover unconformity

In several locations, the "basal" boulder conglomerate does not rest directly on basement, but instead overlies discontinuous lenses of relatively well-sorted sandy quartzite or wacke meters to tens of meters thick. At one locality, bedding in the sandstone is defined by concentrations of magnetite, perhaps similar to that observed more extensively in the Pinnacle Formation of southern Québec (Dowling, 1988; Colpron, 1988). Furthermore, clasts of fine-grained sandy material are found within the overlying conglomerates and wackes, indicating recycling of cover lithologies. Mafic dikes in the basement are truncated by the unconformity, and there is apparent lack of metavolcanics, or even of widespread convincing metavolcaniclastics, in the rocks immediately overlying the Lincoln massif. These observations suggest that rift-related erosion of the massif and its initial rift-clastic cover took place before deposition of the currently preserved sequence, which is itself characterized by evidence for periodic erosion.

Paleozoic deformation

The field area is characterized by a metamorphic and strain gradient which increases from west to east. Detailed mapping at 1:12,000 has shown that the ELM is in fact primarily a ductile shear zone of imbricated wackes, schists, and boulder conglomerates. This conclusion is in marked contrast to the interpretation of the geologic map of Vermont (Doll and others, 1961). Imbrication has occurred on a scale of meters to hundreds of meters within a zone one to two kilometers wide. Mylonitic fabric is pervasive within the basement slivers, and is observed progressively to overprint the earlier Grenville fabric toward fault contacts.

The ductile faults of the eastern "massif" are interpreted to have developed initially as ductile shears along the limbs of minor folds associated with the eastern basement anticline. This deformation was probably preceded and accompanied by multiple "phases" of continuous deformation within the more ductile and earlier-affected cover further to the east. Continued strain is interpreted to have partitioned along the basement-cover contact in response to mechanical "buffering" by the developing rigid basement structure. Severe flattening of the massif in the hinge region produced an anastomosing system of ductile shear zones and minor imbrication of basement and cover along faults of both synthetic and antithetic senses of motion. Along the eastern limb of the massif, simple shear was a greater component of strain, due to a steeply east-dipping zone of rheologic contrast. Strain in the eastern cover was locally concentrated along this contact, resulting in local detachment of cover from basement.

The eastern "massif" and its immediate eastern cover, of biotite to garnet grade, was subsequently juxtaposed along its overturned western limb against the less ductilely deformed, primarily chlorite grade syncline of cover rocks which separates the ELM from WLM. Material west of this boundary, including basement of the WLM, is characterized by a much simpler strain style, documented by 1) the development of a bedding-parallel schistosity; 2) a pervasive schistosity axial planar to the large-scale folds; and 3) sporadically developed late cleavage(s). Tight folds, rather than faults, characterize the western rocks, although late- to post-metamorphic faults are more commonly observed than previously documented (DelloRusso and Stanley, 1986).

The late stages of deformation were characterized by out-of-sequence thrusting immediately east of the ELM when flattening in the eastern massif and westward-younging thrusts in the western massif (Warren, 1989), were not able to efficiently accommodate strain. The westward-younging structures are probably associated with motion on the Champlain thrust (Stanley, 1989).

Preliminary conclusions of this study, germane to the overall geologic history of this region, are 1) although poorly exposed, basement rocks reveal relationships similar to those in the northern Lincoln massif; Grenville lithologic packages exist in this region at or beyond the scale of the entire Lincoln massif; 2) the rift-clastic cover rocks represent sedimentation which post-dates initial rifting in this region and which was not characterized by dramatic subsidence; 3) the ELM behaved as a rigid mechanical buttress between the highly deformed eastern cover and the "protected" rocks of the WLM and western cover sequence during collision-stage Taconian deformation, but was forced to accommodate westward-propagating strain via ductile deformation; 4) despite the contrast in structural style, however, comparison of eastern and western rift-clastic rocks, coupled with structural analysis, suggest that tectonic shortening across the Lincoln massif as a whole is limited to less than 20 km; 5) occurrence of coarsest rift-clastic rocks and possible rift-volcanic dikes along the major Taconian fault zones suggests that some of the Taconian structures may result from reactivation of rift-related structures.

I wish to thank Rolfe Stanley for helpful comments. This research is funded by NSF Grant #EAR 8516879, awarded to R. S. Stanley.

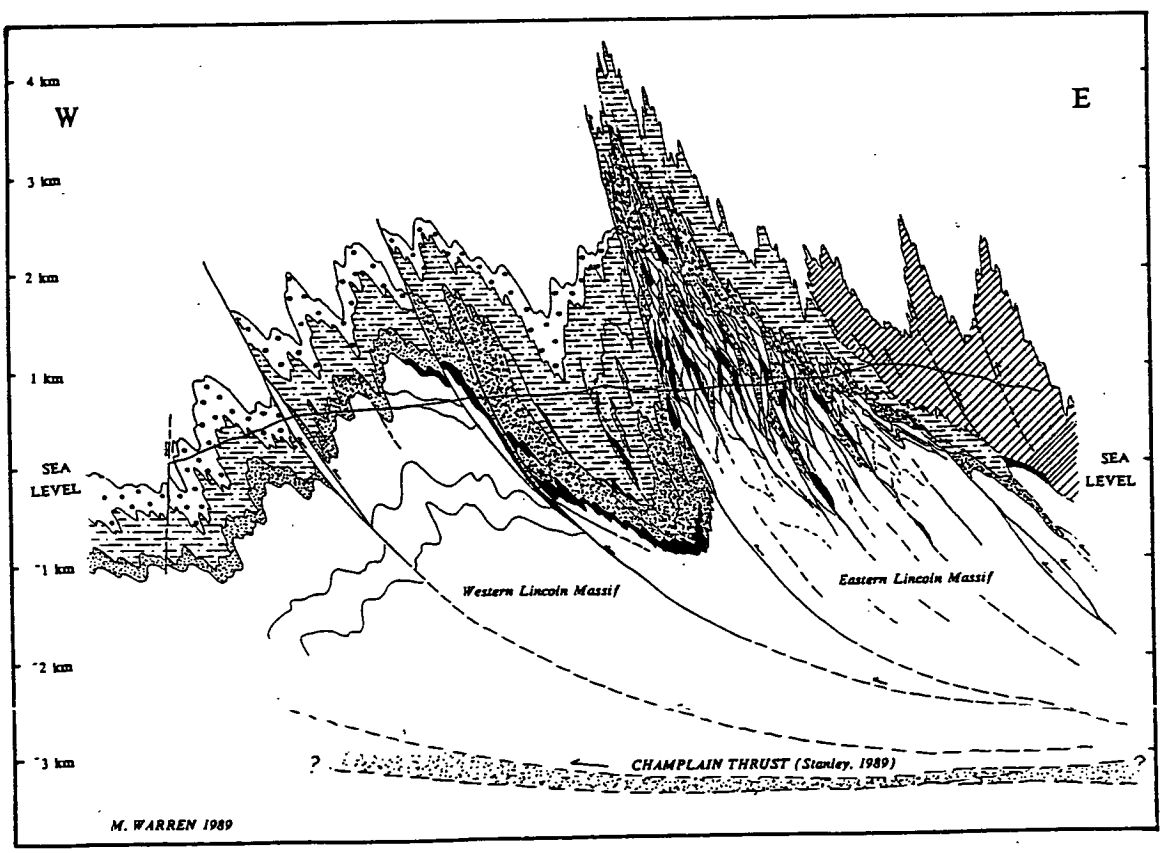

Figure 2: Interpretive cross-section (B - B')through the southern Lincoln massif, central Vermont. White = undifferentiated Grenville basement: solid = undifferentiated conglomerates: small dots = arkosic metawackes: dashed = quartz-muscovite metawacke, schist, or phyllite, locally graphitic: large dots = Cheshire Quartzite: diagonal rule = albitic schist and other lithologies of the Monastery Fm (Armstrong, this volume). Champlain thrust from Stanley (1989). No vertical exaggeration.

THE VERMONT PRE-SILURIAN COVER SEQUENCE: REGIONAL TECTONOSTRATIGRAPHIC RELATIONSHIPS BASED ON DETAILED STRUCTURAL AND LITHOLOGICAL ANALYSIS.

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Recognition of regional thrust faults bounding discrete lithologic sequences in the Taconide Zone of central Vermont has led to the reinterpretation of stratigraphic correlations, clarifying of structural geometry, and development of orogenic models based upon the theories of plate tectonic processes (Stanley and Ratcliffe, 1985; Karabinos, 1988; Ratcliffe et al. 1988; Stanley et al., 1988a).

Fault bounded lithotectonic packages contain stratigraphic sequences deposited in various depositional environments. Sedimentological and stratigraphic characteristics of these sequences allow palinspastic restoration to relative proximal or distal positions; A - **pebble conglomerate, arkosic conglomerate, and quartzite of (Dalton facies)** depositionally overlie Middle Proterozoic gneiss on the western side of the Green Mountain massif (GMM; fig. 1: Ratcliffe et al., 1988). This proximal rift sequence varies in thickness and grades upwards into arkosic and clean quartzite of the Cheshire Formation, the base of the pre-Silurian carbonate platform. North of Rutland, Dalton proximal rocks appear to grade into; B - **interlayered pebble/boulder conglomerate, fine grained wacke, quartzite, albite schist, aluminous schist, and massive dolarenite (Pinnacle facies).** This sequence progressively thickens north toward the Lincoln massif (LM) and becomes more albitic, less coarse grained toward the eastern side of the GMM (Karabinos, 1987). **B** appears to be transitional between **A** and; C - **thick fine grained albite-muscovite-chlorite-biotite schist with interlayered aluminous schist (Monastery facies).** White quartzite, similar to the Cheshire Fm., forms the upper part of this distal basement cover.

The **A - C** section is characterized by progressive westward thinning and coarsening. Iapean feeder dikes within the Grenville gneiss of the Lincoln massif, similar in chemistry to metabasalt of the Pinnacle Formation in northern Vermont (Coish, 1987), are unconformably overlain by **B**. These characteristics support the interpretation depicting the **A - C** facies as late rift sedimentation, subsequent to deposition of metabasalt - bearing clastics of the Group 2 Taconic slices and the central Vermont Underhill Formation. These older sequences are interpreted as the stratigraphically lowest part of **C**. The rapid thickening from **B** to **C** suggests that the position of the rift margin hingeline is in close proximity to the trace of the Underhill Thrust Zone, along the east side of the Lincoln massif (figure 1). This relative age delineation differs from previous ones depicting the Dalton Formation as the oldest pre-Silurian lithology.

C is conformably overlain by; D - **a thin zone of interlayered non graphitic and graphitic schist, and thick, either massive or bedded, dolomite and marble (Plymouth Member-Hoosac Fm., Battell Formation of Armstrong, 1989).** This carbonate is the interpreted distal correlative of the Lower Cambrian Dunham Dolomite (Thompson, 1972). If this correlation is correct, **D** must represent the distal part of the carbonate platform (with underlying clastic cover, **C**). Locally, the carbonate is interlayered and overlain by graphitic schist with dolomitic laminations that may represent turbidites deposited immediately down slope of the shelf edge. Similar sequences are seen in the Group 1 slices of the Taconic allochthon and the Oak Hill Group of southern Québec where the Dunham Dolomite grades laterally into slope-rise graphitic turbidites of the Sweetsburg Formation (Clarke, 1934; Colpron, 1988).

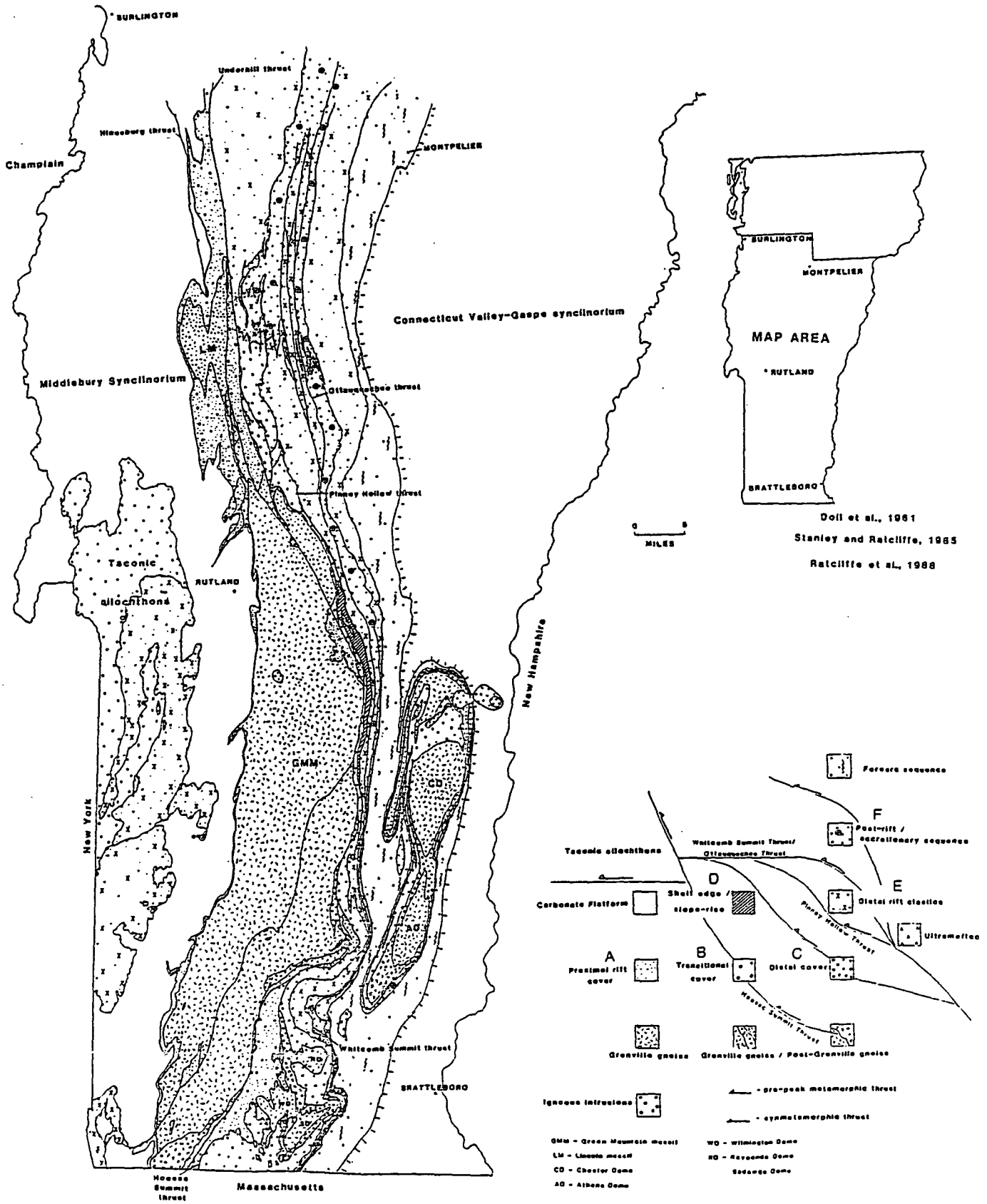

Figure 1. Regional tectonic lithofacies map with insert showing tectonic relationships amongst the various lithofacies-types (letters refer to lithofacies, described in abstract).

E - **muscovite-chlorite-chloritoid phyllites, albite-chlorite-muscovite schist, minor fine grained wacke, quartzite, and locally abundant mafic volcanics** represent distal rift clastics of the central Vermont eastern cover sequence and the Group 3 Taconic slices (Armstrong et al., 1988b; Stanley et al., 1988b). The initial depositional relationship between **E** and **C** is demonstrable in central Vermont due to the tectonic character of the **C / E** boundary (the Pinney Hollow thrust). The lack of a transition between these two sequences means they either do not correlate and were deposited within isolated parts of the basin, or tectonic overlap of **C** and **E** was sufficiently large enough to eradicate the transition. Ratcliffe (1979, 1987) described the facies change between coarse clastics and basalt of the Group 2 Taconic slices (similar to **B** and **C**) with Group 2 fine grained chloritoid phyllite and basalt (similar to **E** and the Group 3 Taconic slices). This section is interpreted as the **C / E** transition with Group 3 slices representing the western facies of **E**. Comparison of the mafic volcanic trace element signatures of Group 2 basalts with those found in central Vermont indicates that they erupted from a mantle source intermediate in depletion character between **C** and **E** (Coish et al., 1985, 1986; Ratcliffe, 1987; Stanley, et al., 1988b); geochemical and lithologic, and structural analyses independently agree on the palinspastic position of the Taconic slices between **C** and **E**.

E also includes a significant amount of chlorite-rich phyllite and mafic volcanics (Stowe Fm.), separated (predominantly) from the western part of **E** by the Ottauquechee Formation. Prior to the recognition of major thrust faults, the Stowe was believed to be the easternmost, and youngest, member of the pre-Missisquoi Group cover sequence (Osberg, 1952; Doll et al., 1961). Recent detailed mapping within the Granville-Hancock area of central Vermont has shown that the Stowe Formation, and more western parts of **E** (Pinney Hollow Fm.), are depositionally continuous over a narrow (≈ 100m) transition zone identifiable in windows through the tectonically overlying Ottauquechee Formation (Armstrong et al., 1988a). Geochemical analyses of volcanics within **D** show an eastward progression in depletion trends with the Stowe rocks displaying MORB tholeiite signatures (Coish et al., 1986). This trend most likely coincides with progressive eastward thinning of the rifted continental crust.

E - **non carbonate - bearing graphitic albitic schist (Granville Fm.) and graphitic phyllite (Ottauquechee Fm.)** are correlated with graphitic rocks of **C**. The lack of carbonate and dolomitic laminations suggests that **F** was formed below the calcium compensation depth in the distal part of the slope-rise or even on the abyssal plain, presumably during thermal subsidence of the lapean margin following continental breakup. A transition between Granville and Ottauquechee rocks is recognized in central Vermont and may exist to the north where the intervening Pinney Hollow Formation has been mapped terminating in a fold closure (Doll et al., 1961). The presence of serpentinite within the Ottauquechee Formation, along thrust faults, including the tectonic contact with **E**, suggests that this formation was in close association with oceanic crust and may partially represent a tectonic mélange / accretionary complex formed during Taconian orogenesis (Armstrong et al., 1988a; Colpron and Armstrong, 1988).

Lithological analysis and stratigraphic correlation of the Taconide Zone in Vermont allow development of a palinspastically restorable lapean passive margin. This restoration is further constrained by modern day basin models and theories on rift and post - rift subsidence mechanisms, which provide comparison to structural - based retrodeformation. An internally consistent passive margin model for the central Vermont lithofacies predicts the presence of basement highs as part of a promontory extending southeastward into Massachusetts. A basement high associated with the Lincoln massif is demonstrated by coarse fluvial deposits (Tauvers, 1982b) and the erosion of lapean feeder dikes with subsequent unconformable deposition of the Pinnacle facies, **B**. A basement high associated

with the Green Mountain massif is characterized by thin, coarse cover and relatively thick shelf edge carbonate, extending outboard of similar sequences to the north; this shows that the early shelf edge was oblique to the north - south Taconian orogenic trend. This initial, rift-related geometry may have had a profound influence on subsequent orogenic style and differential uplift of progressively deeper crustal levels to the south. The shelf edge geometry, in conjunction with detailed structural and lithologic analysis, constrains the palinspastic position of the post - rift proximal turbidites of Group 1 Taconic slices, and associated Groups 2 & 3 rift facies, intermediate to **C** and **E**, above.

Present lithotectonic relationships in central Vermont suggest that initial pre/early metamorphic emplacement of the Taconic allochthons could have been in piggy-back fashion during attempted subduction of the North American margin. Later out-of-sequence thrusting, during continent-arc collision, could account for the present structural arrangement of Taconic slices and diverticulation of lithofacies **C** - **E** along the Hoosac Summit / Underhill Thrust Zones.

ASPECTS OF THE TACONIAN OROGEN IN CENTRAL VERMONT[5]

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Synthesis of recent 1:12,000 mapping, seismic reflection data, and geochemical and mineralogical studies of mafic and pelitic schists, has shown that Middle Proterozoic North American crust and its adjacent pre-Silurian rift-drift cover consist of numerous pre- to post metamorphic thrust faults of varying age and complexity (figs. 1 and 2). The central Vermont transect contains evidence that records the progressive collapse of the North American margin during its collision with an eastern island arc complex during the Taconic Orogeny of Middle Ordovician age. Although deformation generally progressed toward the foreland, it was marked by numerous out-of-sequence thrust faults particularly in the metamorphosed hinterland east of the Middle Proterozoic slices.

Deformation of the Middle Proterozoic crust and its western rift-platform sequence involved detachment along the east-dipping Champlain thrust zone and the younger, more westerly-situated Vergennes thrust (fig. 2). Continued shortening and detachment of the Champlain slice along the Champlain thrust zone deformed the Middle Proterozoic crust of the Lincoln massif and its cover into large west-verging asymmetric fault-propagation folds which became progressively flattened and fragmented toward the hinterland by cleavage, shear zones, and high-angle ductile faults (DelloRusso and Stanley, 1986). This deformation occurred within a temperature range of 250°C to 450°C at depths in the order of 8 to 20 km assuming geothermal gradients in the order of 20°C to 30°C/km.

The pre-Silurian metamorphosed hinterland directly east of the Lincoln massif consists of at least 7 pre/early metamorphic thrust slices that contain in varying amounts metawacke, garnetiferous schist, albitic-quartz schist, quartzite, aluminous chloritoid schist, and carbonaceous schist and quartzite (fig. 3). Mafic schist of basaltic composition is characteristic of the metawacke and albitic-quartz schist sequences whereas serpentinite and talc schist are restricted to the eastern part of the carbonaceous rocks. All of these rocks have been previously assigned by Doll and others (1961) to the Hoosac, Underhill, Hazens Notch, Mt. Abraham, Pinney Hollow, Ottauquechee, and Stowe Formations. The more easterly-situated Missisquoi Formation which has been interpreted by Stanley and Ratcliffe (1985) as a forearc basin deposit is excluded from this analysis.

The pre/early metamorphic thrusts have been deformed into folds of isoclinal, reclined, or sheath geometry. A pervasive east- to southeast-plunging stretching lineation is oriented either parallel or at low angles to the hinges of reclined and sheath folds in the Pinney Hollow and Ottauquechee belts where the shear strain is high. These folds are cut by numerous out-of-sequence syn- and late-metamorphic

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5 all figures included in pocket at the end of the volume.

thrusts with east-over-west asymmetric displacement fabrics which are complicated by intermittent flattening. All fault contacts are marked by synmetamorphic fabrics and/or truncations in the footwall and/or the hangingwall.

Barroisitic hornblende, indicative of the medium-high to high pressure metamorphic facies series, is preserved in mafic schists of the Hazens Notch, Pinney Hollow, Ottauquechee, and Stowe Formations in the eastern part of the hinterland where they are overprinted by a lower pressure, greenschist assemblage of actinolite-chlorite-epidote-albite-quartz (Laird and Albee, 1981; Laird and others, 1984; Laird, 1987; fig. 4). To the west near the Lincoln massif tschermakitic hornblende, indicative of medium-pressure metamorphic facies series is common in mafic schists. The assemblages in the surrounding pelitic rocks indicate upper greenschist to lower amphibolite facies metamorphism followed by synkinematic retrogression. These relations suggest that the older pre- or early metamorphic thrust slices occurred before peak medium-high pressure series metamorphism whereas the younger thrusts largely developed during greenschist to lower amphibolite facies metamorphism and subsequent retrogression. Isotopic age analyses suggest that deformation was largely Taconian with possibly some Acadian overprint.

Comparison of the pre-Ordovician sequence in the Taconic allochthons with the hinterland sequence in central Vermont suggests that the northern extension of the Taconic root zone is located to the east of the metawacke-rich sequence and to the west of the albitic-rich sequence (figs. 2 and 3). This position is further supported by the geochemistry of the mafic rocks as described by Coish (1987), Coish and Sinton (1988) and Ratcliffe (1987). As shown in figure 5 the major and REE of the Taconic metabasalts are more similar to the mafic rocks of Zone 2 than they are to either Zone 3 or 4. The relative positions of these zones are shown in figures 2 and 6. O'Loughlin and Stanley (1986) and Lapp and Stanley (1986) have mapped a major syn- to late metamorphic fault zone along this boundary (loc.5; fig. 2). Because the Taconic allochthons were emplaced prior to any substantial metamorphism, they would correlate either with the early thrust faults recognized in the Vermont hinterland or perhaps with still earlier faults whose record is basically destroyed or yet to be recognized. Thus the fault zone marked as the "Root Zone for the Taconic Allochthons" on figure 2 has had a very long and complex history.

Palinspastic analysis based on structural sequence and stratigraphic relations among the thrust slices and coeval rocks in the Taconic allochthons and the Oak Hill Group in southern Quebec suggests that the rocks represent the eastern North American passive margin sequence which developed during Late Proterozoic through the Cambrian time (Armstrong and others, 1988b; fig. 6). The restored arrangement is supported by mafic schist geochemistry in which the average concentrations of TiO_2, Zr, and La/Yb (a measure of LREE enrichment) progressively change from with-plate signatures in the Middle Proterozoic crust of the Adirondack dome and Lincoln massif in the west to MORB signatures in the Ottauquechee and Stowe Formations to the east (fig. 5). These relations suggest that the mafic rocks formed from progressively depleted magmas during eastward thinning of the ancient sialic crust (Coish, 1987, Coish and Sinton, 1988). Evolution of the Taconian hinterland included westward transport of the thrust slices, ejection of the Taconic allochthons from a zone west of the Hazens Notch and Mt Abraham Formations, and overall shortening of several hundred kilometers by foreland-younging and out-of-sequence thrusts. The serpentinite and talc schist in the carbonaceous rocks of the Ottauquechee and correlative units may represent altered derivatives of oceanic crust incorporated into the post-rift sediments during initial subduction of oceanic crust and associated accretion of post-rift sediments.

STRUCTURAL DEVELOPMENT OF THE CENTRAL CHAMPLAIN VALLEY: LITHOTECTONIC CONTROLS IMPOSED BY A COMPLEX SHELF MARGIN

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Recent mapping in the central Champlain Valley of Vermont has revealed that the structure of the area is a complex thrust belt containing several distinct lithotectonic packages. The prior view of the area as a simple thrust system underlying a major synclinorium was based on miscorrelation of stratigraphic units and lack of recognition of key structural features.

The presently recognized lithotectonic packages are: 1) a stable shelf sequence including the Orwell and Weybridge thrust sheets and the foreland; 2) a series of mini-rise sequences, developed on foundered shelf edges, including the Middlebury limestone, the Rysedorph Hill sequence, and the New Haven sequence; 3) a major rise sequence - the Taconics; 4) foundered shelf edge pieces, including the Sudbury thrust sheet, the `marble belt', and the Champlain thrust sheet; and 5) basement-cored pieces, specifically the Green Mountains. The complex areal relations among these packages are caused by a broad promontory of the carbonate shelf in the central Champlain Valley and by the generally greater slope of the thrust systems than of the shelf margin.

Taconic deformation is generally limited to the outer slope and rise strata and foundered portions of the shelf, with the majority of the shelf strata remaining undeformed even where overridden by the shelf-rise tectonic wedge. The protruding shelf portion deflects the regional structural trends in the overriding packages, resulting in a window to the outer shelf. The outer parts of the shelf promontory are deformed.

Because the thrust surfaces generally dip hindeward more steeply than the original shelf margin, older portions of the outer shelf are tectonically superimposed on younger shelf and rise strata. Likewise, basement is tectonically stacked on shelf and rise strata.

Synthesis of the structural relations within and among the lithotectonic packages indicates that they were accreted to the advancing thrust wedge in the following order: 1) Taconics, 2) Rysedorph Hill sequence, 3) Green Mountains, 4) New Haven sequence, 5) Champlain thrust sheet, 6) marble belt, 7) Sudbury thrust sheet, 8) Middlebury limestone, and 9) Weybridge and Orwell thrust sheets. Events 1, 2, 4, and 8 involve detachment of portions of the rise from the shelf, whereas events 3, 5, 6, 7, and 9 involved detachment of pieces of the shelf. To the south of Brandon, VT, the absence of the protruding shelf eliminates events 7 and 9, and similar relations north of Vergennes, VT, eliminate events 6, 7, 8, and 9.

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Massachusetts 01104

GEOLOGY OF THE WINOOSKI RIVER TRANSECT, NORTH CENTRAL VERMONT

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Introduction

In north central Vermont the Winooski River cuts across the Green Mountain anticlinorium, exposing rocks of the Camels Hump Group. Christman and Secor (1961) mapped areas of Pinnacle, Underhill, and Hazens Notch Formations in the Camels Hump quadrangle. We have remapped central portions of the quadrangle and have found their formation descriptions workable and valuable, but detailed mapping snows that the map pattern is more complicated than Christman and Secor's interpretation (Fig. 1). The area is part of the "internal nappes domain" (St-Julien et al., 1983. Isoclinal east-verging (Fn) folds and faults deform obscure, pre-Fn (west-verging?) structures. Post-Fn folds related to the Green Mountain anticlinorium deform all the above and are dominant in most outcrops.

Stratigraphy

Our mapping is based on the lithologic descriptions outlined in Christman and Secor (1961), a summary of which follows. "The Pinnacle Formation... contains principally thick sections of [meta]graywacke with some interlayered [phyllite and schist]." (p.14).

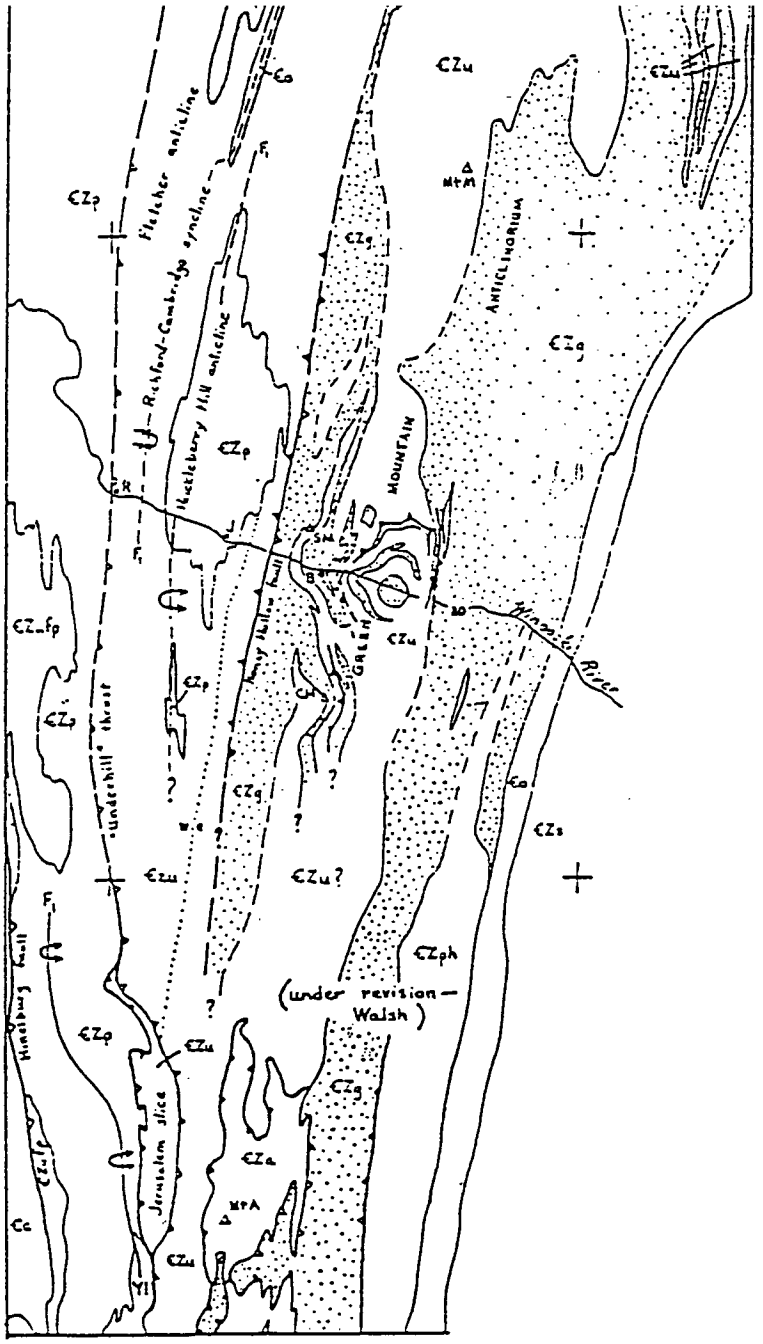
"...in the western half of the quadrangle, the rocks of the Underhill formation are typically phyllite, but some interlayered metagraywacke is common...The rocks contain chlorite, sericite, quartz and some albite, with traces of graphite locally...As the Underhill formation is traced eastward...the rocks are more metamorphosed and are typically quartz-albite-muscovite schist with variable amounts of biotite, chlorite, calcite and garnet... In places...albite may comprise as much as 75% of the rock..." (p.19-21). Most samples in the table of modes for the Underhill (p.19) contain one to two per cent magnetite.

We have substituted the name Granville Formation for Hazens Notch Formation for reasons explained in a later section. "the Hazens Notch formation consists of many different rock types, with the most diagnostic being graphitic schist and phyllite. These contain variable amounts of graphite and some pyrite and the rock, when weathered, has a characteristic rusty color. [Albite is commonly dark gray to black due to graphite inclusions.] Locally, thin interbeds of blue to dark gray quartzite occur...Schist which might be called greenstone occurs locally in the formation." (p.36-37)

Map pattern and structures

The deepest level in the center of the anticlinorium between Bolton village and Bolton Dam is poorly exposed, but consists of Granville Formation and a locally thick sequence of greenstones. The greenstones grade up into the Underhill Formation. The Underhill, occupying the next higher level and up to 1000 m thick is cut by numerous faults and contains numerous slices of Granville. Compositional layering, the dominant foliation (Sn) and the faults are all deformed by the anticlinorium. Minor post-Sn folds related to the anticlinorium have opposite rotation senses from one side of the anticlinorium to the other, and in the more pelitic rocks a spaced cleavage dipping steeply east is parallel to their axial planes. The older Fn minor folds in this central region have a west-over-east rotation sense.

On the east limb of the anticlinorium, there is a mixed zone marking the eastern edge of the main mass of Underhill Formation. Thin layers of greenstone and lithologies from both Underhill and Granville can be followed along strike for up to hundreds of meters. Mapping at 1:10,000 reveals a few



KEY

Geographic place names:

MtM	Mt. Mansfield
SM	Stimson Mountain
CH	Camels Hump
MtA	Mt. Abraham
R	Richmond
J	Jonesville
B	Bolton
BD	Bolton Dam

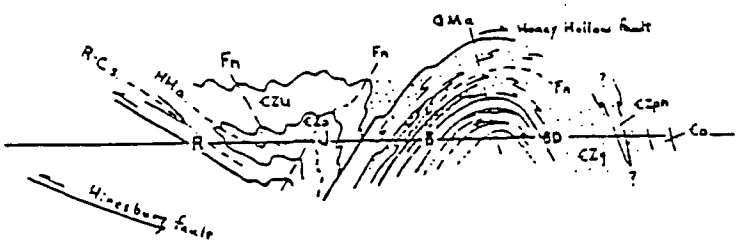
Lithologic units:

€c	Cheshire Quartzite
€o	Ottauquechee Fm.
€Zg	Granville Fm.
€Zs	Stowe Fm.
€Zu	Underhill Fm.
€Zufp	Fairfield Pond Mbr.
€Zph	Pinney Hollow Fm.
€Za	Mt. Abraham Schist
€Zp	Pinnacle Fm.
Yl	Lincoln Mountain Massif

Other symbols:

+	corners of Camels Hump quadrangle
	pre-Fn fault
F_1	pre-Fn fold
	syn- or post-Fn fault
w/e	"western" and "eastern" Underhill Formation

Schematic Cross Section along Winooski River:



HHa. Huckleberry Hill anticline
RCs. Richford-Cambridge syncline
GMa. Green Mountain anticlinorium

Figure 1: Generalized geologic map of the Camels Hump quadrangle and surroundings (adapted from Doll et al., 1961; Eiben, 1976; Aubrey, 1977; DiPietro, 1983; and O'Loughlin and Stanley, 1986). Scale: 1:340,000.

isoclinal Fn folds, and in some cases faults cut the Fn folds. Mylonitic fabrics range from incipient to pervasive. Some fault zones are marked by papery phyllites from a few centimeters to a meter thick. East of Bolton Dam there are no Underhill horizons, with the exception of a 100 m thick layer on Crossett Hill.

The lack of obvious symmetry across the anticlinorial crest toward the west is one of the intriguing puzzles of the Camels Hump area. Older folds and faults are responsible, but the relationships are not entirely clear. About one km west of the crest, near Joiner Brook Valley and Bolton village, Fn minor isoclinal folds are either neutral, or have both rotation sense in the same outcrop. Father west Fn folds are east-over-west. THerefore it would seem that there must be a synclinal Fn fold hinge near Joiner Brook. If so, the western overturned limb is cut by faults, obscuring the symmetry across the hinge. Many outcrops show evidence for sheared-out folds, and a mylonitic zone cuts across a large isoclinal fold east of Camels Hump. Still farther west, west of Stimson Mountain, greenstones, Underhill, Granville and Pinnacle occur in a mixed zone. The deformation style is similar to that in the mixed zone on the eastern side of the anticlinorium, but so far, mapping to the north and south has not discovered a place where the zones connect across the anticlinorium.

There is an important north-south fault along Honey Hollow south of the Winooski, and along Duck Brook to the north. Greenstone horizons and layers of Underhill east of the fault cannot be followed across the fault. The area to the west is dominated by Underhill, with minor Granville layers. Minor Fn folds are east-over-west on both sides of the fault. The proportion of graywacke increases westward, and on the hills west of Duck Brook, Pinnacle is infolded with Underhill. Toward the south this Pinnacle/Underhill contact diverges gradually from the fault. To the west, toward Richmond, the intensity of deformation and metamorphism decreases, and Fn structures are more nearly upright.

Regional Context of Structures

The east-verging Fn folds are analogous to those mapped in the Bonsecours and Sweetsburg Formations by Osberg (1965) in the Sutton Mountain anticlinorium, Québec. We suspect the Granville Formation may extend from the Stimson Mountain area northward to tie in with the Granville south of Jeffersonville, Vermont (Hazens Notch of Doll et al., 1961), to form a continuous belt west of the anticlinorium. The greenstones at Stimson Mountain would thus occupy a similar stratigraphic within he Granville as the Peaked Mountain Greenstone (Dennis, 1964; Thompson, 1975)

The Honey Hollow fault is proposed as a west-dipping fault, much like those mapped along the east side of the Fletcher Mountain anticline (Doolan et al., 1987) and Enosburg Falls anticline in Québec (Osberg, 1965). Reverse movement on the fault brought less-metamorphosed Pinnacle and Underhill west of the fault upwards into juxtaposition with the more intensely deformed terrane to the east. There may have been pre-Fn west-directed motion on the same fault. The mass of Pinnacle Formation between Jonesville and Richmond apparently forms the core of a west-verging pre-Fn anticlinal nappe, for which we propose the name Hucklebury Hill anticline. It is deformed by nearly upright Fn folds with axial planes which fan across the area. Post-Fn structures are less well developed (Thresher, 1972). The Richford-Cambridge syncline (pre-Fn overturned to the west?) strikes toward the area west of the Hucklebury Hill anticline. The extension of all these structures to the south is uncertain; they lie east of the so-called "Underhill thrust" (Tauvers, 1982b).

Stanley and Ratcliffe (1985) suggested that the contact between Underhill and Hazens Notch extending from the Winooski River past Mt. Mansfield (Doll et al., 1961) may be a major boundary between pre-Fn thrust slices. Our mapping suggests that no special significance should be attached to this one contact. Some Underhill /Granville contacts appear to be depositional, some are Fn faults, and

some may be pre-Fn faults. There is local evidence for pre-Fn faults, in one instance folded by an Fn fold, but control of the map pattern by pre-Fn faults is obscured by younger structures.

Stratigraphic problems

The Underhill Formation east of the Honey Hollow fault is a chlorite schist with variable amounts of albite and magnetite, and in many respects is similar to the Pinney Hollow Formation. In some places, most notably on the open ledges east of Camels Hump, chlorite-paragonite schist similar to the Mount Abraham Schist is interlayered with albite-bearing schist. Magnetite porphyroblasts are distinctive but not ubiquitous in these rocks. We conclude that the Mount Abraham, Pinney Hollow, and Underhill were depositionally related, but not necessarily exactly coeval. Toward the west the Underhill changes character, largely due to lower metamorphic grade. Albite porphyroblasts are smaller, magnetite is sparse to absent, and the platy minerals become finer so that the rock is more phyllitic than schistose. Foliation surfaces weather tan rather than silver-green. The contact between western and eastern Underhill nearly coincides with the garnet isograd mapped by Christman and Secor (1961). At the latitudes of the Winooski River the western Underhill contains discontinuous greenstones and more abundant graywacke horizons. These differences are due to primary facies changes. To the south the isograd is farther west so that at the latitude of Jerusalem the higher-grade eastern Underhill also contains greenstones and more graywacke. The relationship between the eastern and western Underhill is similar to that between the West Sutton and the Bonsecours Formations in Québec; the Bonsecours is the deformed and metamorphosed equivalent of the West Sutton (Slivitsky and St-Julien, 1987).

The name Hazens Notch Formation is currently controversial. Doolan (in review) suggests that the Hazens Notch is a mélange of other stratigraphic units, ranging from the Pinnacle to the Ottauquechee Formations. We propose that the graphitic rocks defined as Hazens Notch by Christman and Secor (1961) be called Granville Formation, a name used for similar rocks to the south (Osberg, 1952). Doolan's term "Hazens Notch Terrane" could then be applied to the broad area between the Honey Hollow fault and the Baie Verte-Brompton line in which several stratigraphic units, especially the Underhill and Granville (Underhill and Hazens Notch on the state map) are tectonically mixed at scales from lenses smaller than the outcrop to slices several kilometers thick. The location of the boundary between Stanley and Ratcliffe's (1985) "Hazens Notch slice" and "Underhill slice" remains uncertain.

GEOCHEMISTRY OF THE METABASIC ROCKS FROM THE PRE-SILURIAN RIFT-CLASTIC SEQUENCE OF CENTRAL VERMONT.

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Major and trace element geochemical signatures of metabasic rocks from the pre-Silurian rift-clastic sequence are separated into the zone classification of Coish (1989b; this volume). Geochemical signatures of greenstones and amphibolites from the Underhill Formation (Zone 2) are distinct from those of the Fayston Schist (formerly non-carbonaceous white albitic schist of the Hazens Notch Formation), Mount Abraham Schist and Pinney Hollow Formation (Zone 3), and Stowe Formation (Zone 4; Fig. 1). The metabasic rocks of Zone 2 have very high TiO_2, P_2O_5, Zr, and Y and plot as within-plate and alkali basalts (Figs. 2a.& b.). The geochemistry of Zone 2 is similar to Group A of Coish et al. (1985) which includes Huntington (Underhill Formation) and Tibbit Hill (Pinnacle Formation) greenstones and amphibolites (Fig. 2c.). Zone 3 rocks have intermediate TiO_2, P_2O_5, Zr, and Y and plot as transitional from within-plate basalts to tholeiitic basalts (Figs. 2a.& b.). Zone 3 geochemistry is

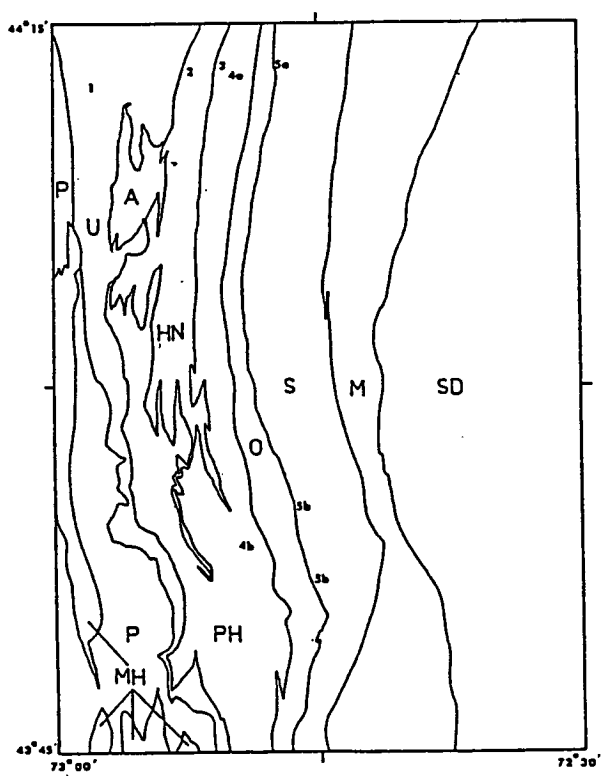

Figure 1. General locations of greenstones from which samples were obtained in central Vermont. ZONE 2: 1. Underhill greenstone and amphibolite, Stave Brook, Buels Gore. ZONE 3: 2. Fayston greenstone, Fayston. 3. Mount Abraham greenstone, Fayston. 4. Pinney Hollow greenstone, Fayston (4a.), and Rochester (4b.). ZONE 4: 5. Stowe greenstone, Waitsfield (5a.), and Rochester (5b). Zone classification of Coish (1989). Geology is based on the Vermont Centennial Map (Doll et al., 1961). Detailed sample descriptions,locations, and geologic maps are found in University of Vermont Master of Science theses by the authors (in progress). P = Pinnacle Formation, MH = Mount Holly Complex, U = Underhill Formation, A = Mount Abraham Schist, HN = Hazens Notch Formation, PH = Pinney Hollow Formation, O = Ottauquechee Formation, S = Stowe Formation, M = Mississquoi Formation, SD = Siluro-Devonian sequence of eastern Vermont.

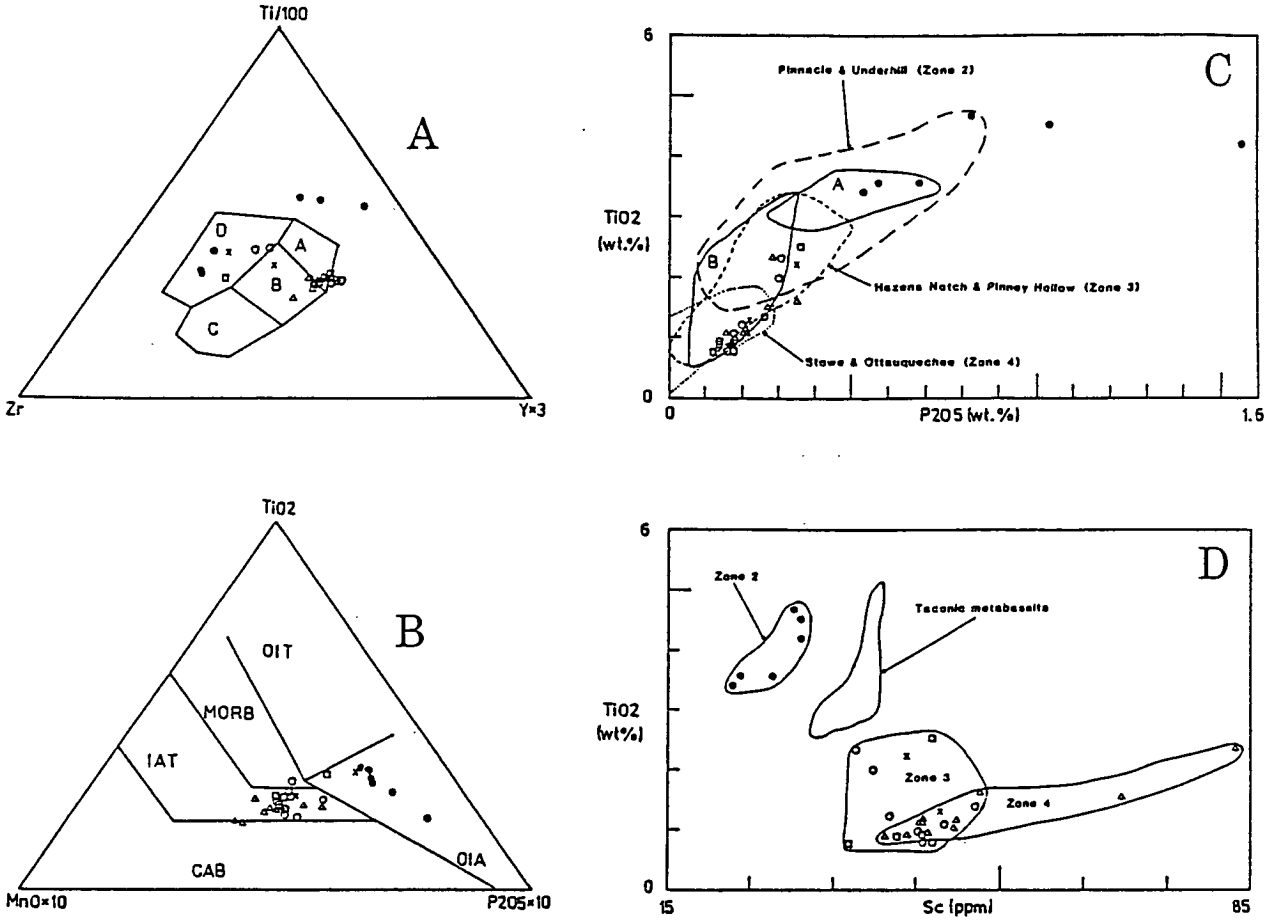

Figure 2. Geochemical diagrams: Closed circle = Underhill (6 analyses), open square = Mount Abraham (4 analyses), x = Fayston (2 analyses), open circle = Pinney Hollow (8 analyses), open triangle = Stowe (11 analyses). A. Zr-Ti-Y diagram. Tectonic fields from Pearce and Cann (1973): A and B = ocean floor basalts and low K tholeiites, C and B = calc-alkaline basalts, and D = within-plate basalts. B. TiO_2-MnO-P_2O_5 diagram. Tectonic fields from Mullen (1983): OIT = ocean island tholeiites, MORB = mid-ocean ridge basalts, IAT = island arc tholeiites, CAB = calc-alkaline basalts, OIA = ocean island alkali basalts. C. TiO_2-P_2O_5 diagram. Fields A and B from Coish et al. (1985). Pinnacle and Underhill (Zone 2), Hazens Notch and Pinney Hollow (Zone 3), and Stowe and Ottauquechee (Zone 4) fields from Coish (1987, 1989). D. TiO_2-Sc diagram. Fields for Zones 2, 3, and 4 are from this study. Field of Taconic metabasalts is from Ratcliffe (1987).

similar to Group B greenstones of Coish et al. (1985) from the Underhill and Pinney Hollow Formations, and to Type 1 greenstones from the Stowe Formation (Coish et al., 1986). Zone 4 rocks have intermediate to low TiO_2, P_2O_5, Zr, and Y and plot as tholeiitic basalts (Figs. 2a. and b.). In the absence of REE data, the TiO_2-Sc diagram shows the distinction between the three zones better than any other trace element diagram (Fig. 2d.). Zone 2's high TiO_2 versus low Sc values are similar to Hawaiian alkalic basalts and transitional basalts from the Kenyan rift (Basaltic Volcanism Project, 1981), a similarity first recognized for Group A by Coish et al. (1985). Zone 3's intermediate values are again similar to Group B of Coish et al. (1985) which are in the MORB range. Zone 4's lower TiO_2 and higher Sc also plot within the MORB range and overlap with the Zone 3 field. This overlap coincides with detailed mapping that indicates the Pinney Hollow and Stowe Formations are depositionally related and are part of an east-west rift-clastic facies change (Stanley et al., 1987a. and 1987b.; and Kraus et al., 1988). The geochemical overlap between the Fayston, Mount Abraham, and Pinney Hollow greenstones within Zone 3 supports evidence that all three units are part of the east-west rift-clastic facies change (Walsh and Stanley, 1988), and that their current position west of the Stowe Formation is a reflection of their previous position in the rift-clastic sequence. The Underhill Formation includes the western most equivalent of the predominant rock type in the rift-clastic sequence of central Vermont--a non-carbonaceous, quartz-muscovite-white albite-chlorite schist. A depositional transition between the Underhill Formation and the Fayston Schist has not been recognized in central Vermont, and recent mapping indicates a tectonic boundary with a complex history between the two units. With the correlation of the lithologies from the Pre-Silurian sequence of central Vermont with those of the Taconic allochthons well established, it is logical that a geochemical correlation should follow. Metabasalts from the Rensselaer Plateau and Chatham slices (Ratcliffe, 1987) occur in rock types that are similar to those of the central Vermont rift-clastic sequence. All characteristics of the Taconic metabasalts indicate that they are tholeiitic to transitional alkalic basalts and basaltic tuffs. The geochemistry of the metabasalts is similar to that of the Tibbit Hill volcanics. The Taconic metabasalts have a high TiO_2 content similar to Group A of Coish et al. (1985), but the Sc values are higher. The geochemical comparison is most striking in the TiO_2-Sc diagram where Taconic metabasalts plot between Zones 2 and 3, thus filling in the geochemical void in the central Vermont rift-related metabasic rocks (Fig. 2d.). The agreement of the geochemical data with the recognition of a significant tectonic boundary between the Underhill Formation and Fayston Schist suggests that the Taconic root zone of Stanley and Ratcliffe (1985) may be located between the two units in central Vermont. If the geochemistry stands the test of time the TiO_2-Sc diagram may be an important tool for deciphering the Taconian history of central Vermont.

GEOCHEMISTRY AND GEOCHRONOLOGY

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LA CRÉATION DE L'OCÉAN IAPETUS; DU RIFT CONTINENTAL A LA FORMATION DU BASSIN OCÉANIQUE: CONTRAINTES PÉTROLOGIQUES ET GÉOCHIMIQUES

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Les unités structurales de la Nappe de la Rivière Chaudière et de l'Olistostrome de Drummondville font partie du domaine externe de la zone de Humber (Williams, 1979) des Appalaches du Québec. Cette dernière zone montre le développement et la destruction de la bordure continentale de l'océan Iapétus au Paléozoïque inférieur. Cette bordure, de type atlantique, repose sur un socle grenvillien (St-Julien et al., 1983). Les lithologies associées à la Nappe de la Rivière Chaudière correspondent à un assemblage de schistes argileux verts, rouges et gris à interlits de siltstones, d'arénites feldspathiques, de rudites et de grès quartzeux. Ces unités lithologiques font parties du Groupe de Sillery dont l'âge est évalué au Cambrien inférieur (Slivitzky et St-Julien, 1987). Localement, des roches volcaniques basiques et des rhyolites (Vermette, en prép.) forment une mince unité à la base de ce groupe (St-Julien et Hubert, 1975). L'Olistostrome de Drummondville est composé d'une matrice de shale contenant des blocs de calcaire cristallin, d'agglomérats volcaniques, des laves à amygdules et vésicules, et des trachy-andésites (Vermette, en prép.) dont l'âge de la matrice varie de Cambrien à Ordovicien moyen (Globensky, 1978).

1) MORPHOLOGIES ET MINERALOGIE DES VULCANITES

Les vulcanites basaltiques associées à la Nappe de la Chaudière et à l'Olistostrome de Drummondville ont été subdivisées, sur une base minéralogique et géochimique, en quatre groupes, soit: les basaltes tholéiitiques, les basaltes transitionnels à affinité tholéiitique et alcaline, et les basaltes alcalins. Les basaltes tholéiitiques (*groupe 1*) et transitionnels à affinité tholéiitique (*groupe 2*) se présentent sous forme de coulées massives et coussinées ainsi que sous forme de brèches et de tufs hyaloclastiques. Ces basaltes sont généralement aphanitiques et leur minéralogie primaire est composée de phénocristaux et de microphénocristaux d'olivine et de plagioclase dans une matrice contenant des microlites de plagioclase, du clinopyroxène, des minéraux opaques, et du verre chloritisé et palagonitisé. De la pumpellyite et de la préhnite, indicatrices d'un degré de métamorphisme très faible, ont également été

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pumpellyite et de la préhnite, indicatrices d'un degré de métamorphisme très faible, ont également été observées dans quelques échantillons.

Les basaltes transitionnels à affinité alcaline (*groupe 3*) se présentent sous forme de coulées massives vésiculaires et amygdalitiques, de coulées brèchiques et de laves à blocs. Leur granulométrie varie d'aphanitique à finement grenue. La minéralogie primaire des basaltes du groupe 3 comprend des phénocristaux et des microphénocristaux d'olivine et de plagioclase dans une matrice constituée de microlites de plagioclase, d'augite, d'olivine, d'apatite, de sphène, d'ilménite, et de verre volcanique chloritisé ou palagonitisé. Ces basaltes sont en contact net avec des calcaires cristallins à Acton Vale. Le principal faciès associé au basaltes alcalins (*groupe 4*) correspond à des coulées massives de basaltes finement à moyennement grenus et très peu vésiculaires. Ces basaltes sont observés en contact net avec des grès ou des schistes gris vert. La minéralogie primaire comprend des phénocristaux et des microphénocristaux de plagioclase, d'olivine et de Ti-augite. La matrice est composée de microcristaux de plagioclase, d'apatite, de biotite, de zircon, de spinelle et de minéraux opaques. On y observe également de la pumpellyite et de la préhnite comme minéraux de métamorphisme.

Les trachyandésites se présentent sous forme de coulées massives vésiculaires et amygdalitiques, et sont composées de phénocristaux et de microphénocristaux de plagioclase et d'augite. Des microlites de plagioclase constituent la mésostase des trachyandésites. Les rhyolites sont caractérisées par un passage graduel d'un faciès brèchifié à massif. Ces dernières sont composées de microphénocristaux de quartz dans une matrice de quartz et feldspath se présentant sous forme sphérulitique et des traces de biotite.

2) GEOCHIMIE DES VULCANITES

La classification géochimique, effectuée à l'aide du diagramme de la figure 1, nous indique que les basaltes du groupe 1 sont purement de type sub-alcalin alors que ceux du groupe 2 présentent un caractère transitionnel entre les champs sub-alcalin et alcalin. Les basaltes des groupes 3 et 4 possèdent un rapport Nb/Y supérieur à 0.8 qui est caractéristique des basaltes alcalins. Le métamorphisme le plus important affectant chacun des groupes de basaltes consiste en une albitisation des plagioclases qui se traduit par un enrichissement en Na_2O, K_2O, Rb et Cs par rapport à un appauvrissement en CaO, Sr et Ba. Quelques échantillons de basaltes présentent également des évidences de désodification et de désilicification (Vermette, en prép.). Avec l'évolution de la différenciation, chaque groupe de basaltes est caractérisé géochimiquement par le fractionnement de l'olivine, du clinopyroxène et de la chromite. La remobilisation des éléments alcalins rend difficile l'étude du fractionnement du plagioclase. Cependant, pour chacun des groupes de basalte, le comportement de l'Eu ne suggère pas le fractionnement de cette dernière phase (Vermette, en prép.). Les rhyolites constitueraient le produit ultime du fractionnement des basaltes transitionnels à affinité tholéiitique alors que les trachyandésites présentent un caractère consanguin avec les basaltes du groupe 3. Les données des travaux en géochimie isotopique suggèrent une contamination crustale pour certains de ces faciès.

Le patron des éléments des terres rares (figure 2) nous indique que les basaltes du groupe 1 montrent un appauvrissement important en LREE et un enrichissement en HREE. Un tel profil, de même que des rapports La/Sm et La/Yb dont les valeurs varient respectivement de 0.4 à 1.6 et de 0.4 à 1.0, sont typiques des N-MORB de l'Atlantique Nord (Schilling et al., 1983). Les basaltes du groupe 2 montrent un profil caractérisé par un faible enrichissement en LREE par rapport aux HREE et possèdent des rapports La/Yb (1.8 à 4.2) et La/Sm (0.4 à 1.0) semblables aux basaltes de type T-MORB de l'Atlantique Nord (Schilling et al., 1983). Les basaltes des groupes 3 et 4 présentent un fort enrichissement en LREE dont les valeurs varies entre 60 et 200 fois les valeurs chondritiques. De plus, pour les basaltes du groupe 3, les rapports La/Yb et La/Sm varient respectivement de 5.5 à 14.5

et de 2.6 à 4.8 indiquant un caractère transitionnel alcalin semblable à ceux de Boina en Afar (Barberi et al., 1975). Les basaltes du groupe 4 possèdent des rapports La/Yb et La/Sm variant respectivement de 17.1 à 29.3 et de 3.0 à 4.3. Bien que ces rapports ne présentent pas de similitudes avec ceux de la région de Boina ou du Gregory Rift du Kenya (Baker et al., 1977), leurs valeurs élevées, de même que le patron des REE, confèrent à ces basaltes un caractère alcalin intraplaque continental. Ces vulcanites seraient reliées à différents époques de la création et de l'ouverture de l'océan Iapétus. Les types alcalins sont générés au stade du rift initial en milieu continental, alors que les types tholéiitiques seraient associés à un environnement de rift continental mature et de rift océanique. Les résultats de géochimie isotopique préliminaire sur les couples (Sm/Nd) suggèrent une distribution des âges à la limite Précambrien-Cambrien.

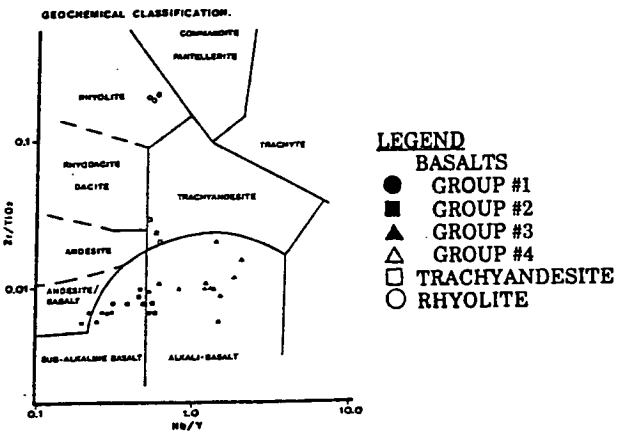

Figure 1: Classification géochimique des roches basaltiques de la Nappe de la Rivière Chaudière et de l'Olistostrome de Drummondville. Diagramme d'après Winchester and Floyd (1977).

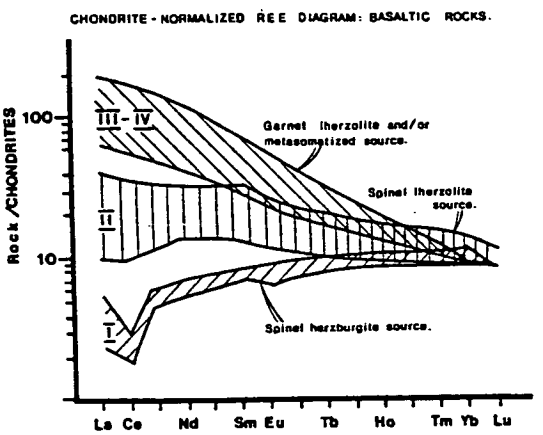

Figure 2: Patrons des éléments des Terres Rares pour les roches basaltiques de la Nappe de la Rivière Chaudière et de l'Olistostrome de Drummondville.

THE SIGNIFICANCE OF GEOCHEMICAL TRENDS IN VERMONT GREENSTONES

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Greenstones from western and central Vermont show systematic geographic trends that can be correlated with changes in chemistry accompanying the rifting and opening of Iapetus ocean. Here I outline those trends and compare the chemistry of some of the central Vermont greenstones to basalts from the Thetford Mines ophiolite, Québec. To facilitate discussion, I have divided the mafic rocks into 4 zones, representing a west to east sequence.

Zone 1 - Mafic dikes cut the Adirondack basement in eastern New York and the Lincoln massif, just north of Ripton, Vermont (Isachsen et al., 1988; Coish and Sinton, 1988; Sinton, 1988; DelloRusso and Stanley, 1986; Stanley et al., 1987a) . The dikes are believed to be Precambrian in age. Adirondack dikes are undeformed and only slightly metamorphosed but the Vermont samples are metamorphosed and somewhat deformed.

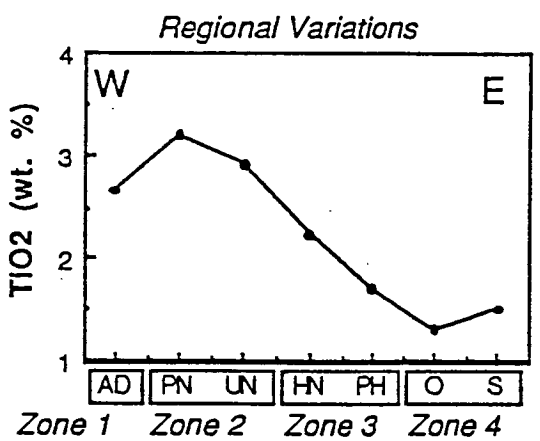

Figure 1. Average concentration of Ti in all geochemical zones. The trends probably indicate a changing mantle source area for the magmas that were erupted during progressive splitting of the ancient Adirondack continent, eventually leading to formation of the Iapetus ocean. (AD - Adirondack Dikes; PN - Pinnacle Fm; UN - Underhill Fm.; HN - Hazens Notch Fm.; PH - Pinney Hollow Fm; O - Ottauquechee Fm; S - Stowe Fm.)

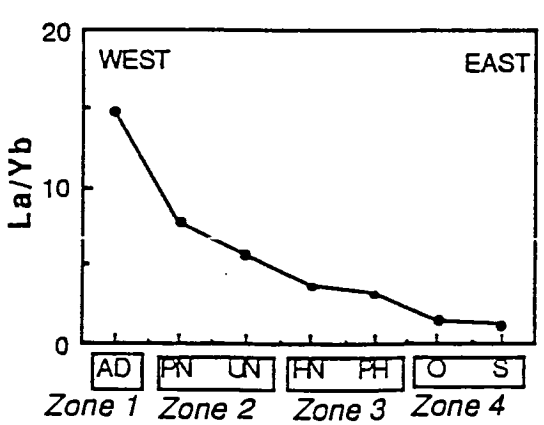

Figure 2. Variation in La/Yb ratio (a measure of the light rare-earth enrichment) for Vermont greenstones.

Geochemically, the dikes cutting the basement are transitional basalts. They have high Ti, Zr, Y, P and rare earth element (REE) concentrations (Figs. 1 & 2). They have enriched LREE patterns (La/Yb >> 1.0, Fig. 2). The dikes are similar in chemistry to within-plate basalts, including rift regions. Compared to greenstones in overlying formations in Vermont, they have many chemical similarities; however, some dikes (especially those from Vermont) have higher K, and lower Ti/P and Ti/Y ratios. These differences may reflect a slightly different mantle source for Zone 1 mafic dikes or continental crust contamination.

Zone 2 *- Pinnacle and Underhill Formations:* Greenstones from the Pinnacle and Underhill Formations are grouped together in Zone 2 because they are geochemically very similar, and thus, probably have the same origin. Their chemistry indicates the samples were transitional basalts with very high concentrations of Ti, Zr, Y, and the rare earth elements (Figs. 1 & 2). These geochemical features are typical of basalts produced in the early stages of continental rifting. Although there is variation in chemical composition among all the samples, there is no clear difference between the Pinnacle & Underhill greenstones. This indicates that the Pinnacle and Underhill greenstones were basalts generated from the same type of mantle source rock. Different amounts of melting and fractionation along with some post-formation alteration probably resulted in the chemical variation seen in Zone 2 greenstones.

Zone 3 *- Hazens Notch and Pinney Hollow Formations:* Greenstones from these two formations were grouped in Zone 3 because of their overall geochemical similarity, and their differences from greenstones to the west (Zone 2) and east (Zone 4). Geochemically, the greenstones from Zone 3 are high Ti, Zr, Y and REE basalts; concentrations of these elements overlap with the low end of the envelope for Zone 2 greenstones and extend to lower values typical of greenstones from Zone 4. Samples from one particular body exhibit nearly the entire chemical range shown by the whole group. This suggests either 1) magmas erupted in a single place were generated at different depths in the mantle or from different parts of a single magma chamber, or 2) the different basalts were generated in disparate regions and intimately mixed later by faulting. Whatever the reason, it is clear that the Hazens Notch and Pinney Hollow formations contain basaltic greenstones with a wide range of chemical compositions.

Zone 4 *- Ottauquechee and Stowe Formations:* Greenstones from the Ottauquechee Formation are indistinguishable chemically from greenstones from the Stowe Formation. All samples from Zone 4 are basaltic in composition and have much lower concentrations of Ti, Zr, Y and REE than most greenstones from more westerly zones (Fig. 1 & 2). Also, the rare earth element patterns show a slight to moderate depletion in the LREE relative to the HREE (La/Yb < 1, Fig 2), contrasting with the LREE-enriched patterns of the western formations. In almost all geochemical features, Zone 4 greenstones are similar to modern mid-ocean ridge basalts. The depleted LREE pattern indicates that the greenstones must have been derived from a different mantle source than the western greenstones; in particular, the mantle source must have been depleted mantle. On all counts, Zone 4 greenstones are different from all greenstones in Zone 2 and many greenstones in Zone 3.

The Thetford Mines ophiolite in the Québec Appalachians has been interpreted as a section of Ordovician ocean crust, so it is natural to compare the Zone 4 greenstones to basalts in this ophiolite (Fig. 3). It is clear from the variations in Cr & Y that the Zone 4 greenstones of this study are **not** like the Thetford Mines ophiolite. In fact, the Zone 4 greenstones are like mid-ocean ridge basalts whereas volcanics from the Thetford Mines ophiolite are similar to island arc rocks (Coish, 1989a; Oshin and Crockett, 1986). The significance of this non-correlation is not fully understood, but it is interesting to note that greenstones from the eastern part of the Stowe Formation and western Moretown Formation show island arc affinities (Cua, 1988; Dick, work in progress). Preliminary interpretations suggest

that all volcanics west of the eastern part of the Stowe are rift or ridge-related with island arc rocks to the east.

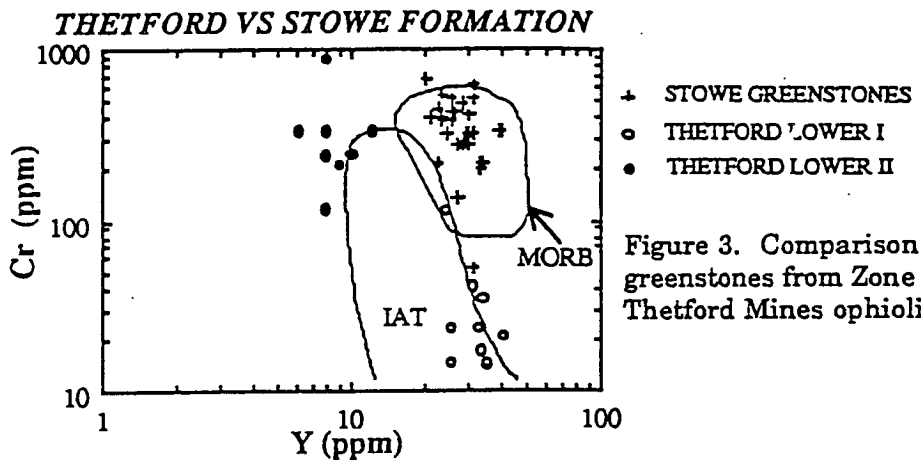

Figure 3. Comparison of Cr-Y values in greenstones from Zone 4 with basalts in the Thetford Mines ophiolite, Quebec.

TECTONIC IMPLICATIONS: Geochemical trends in modern basaltic rocks have been used to place volcanic rocks in a particular tectonic environment, i.e, mid-ocean ridge, subduction zone, or within-plate environment. Applying the same technique to ancient rocks, it can be shown that greenstones from Zones 1 and 2 formed in a within-plate environment. Furthermore, these greenstones appear to have been formed within a **continental** plate. Most modern within-continental plate volcanism occurs in regions where a plate is being stretched and pulled apart to form a rift valley. The tectonic environment of formation of Zone 4 greenstones appears to be oceanic. Greenstones from Zone 3 show chemical characteristics of both within-plate **and** oceanic environments.

The regional geochemical trends and implied tectonic environments of greenstones in central Vermont (e.g. Figs 1 & 2) can be rationalized by a fairly simple model of splitting of a continent and formation of an ocean basin. During the late Precambrian (~ 650? m.y.a), the Adirondack continent began to pull apart presumably due to a rising thermal plume in the mantle below. Volcanism accompanied this stretching resulting in the formation of basalts in Zones 1 and 2. The continental crust at this time was distended but still fairly thick; this can be called an early stage of rifting. Magmas produced at this stage were primarily from melting an enriched sub-continental lithosphere. Later (~600? m.y.a.), volcanism continued to produce basalts further east as the zone of magmatism migrated. The continental crust was now more stretched and an ocean basin had started to develop. This was a transitional stage in the development of ocean crust; the basalts produced here had transitional chemical features between true within-plate basalts and ocean ridge basalts. The basalts erupted at this time later became part of Zone 3. They were probably produced by melting both continental lithosphere and convecting mantle asthenosphere. Still later (~550? m.y.a.), the continental crust completely split and melting of depleted mantle asthenosphere produced ocean ridge type basalts in the region that later became part of Zone 4. The ocean basin at this time could have been very small. Greenstones farther east in Vermont (and the Thetford Mines ophiolite in Québec) appear to have been formed later when subduction resulted in island arc formation. Continued subduction led to the eventual closure of the Iapetus ocean basin during the mid-Ordovician resulting in metamorphism of all the basalts in Vermont to greenstones.

GEOCHEMISTRY OF ORDOVICIAN ISLAND-ARC AND OCEAN-FLOOR ROCK ASSEMBLAGES FROM THE QUÉBEC OPHIOLITES, CANADIAN APPALACHIANS

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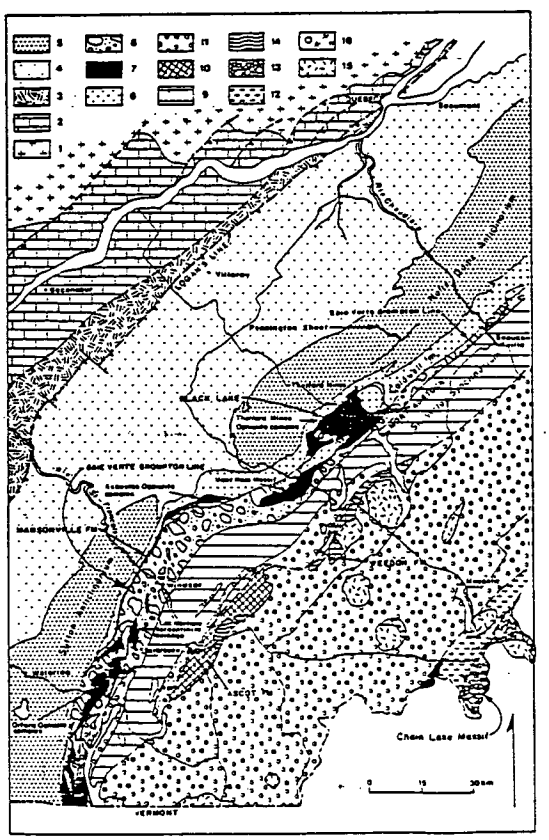
This work provides new geochemical and petrological data on ophiolitic slices which are tectonically incorporated into the Dunnage Zone of the Quebec Appalachians (Figure 1; St-Julien et al., (1983). These ophiolites are the remnants of former Iapetus basins of Early Ordovician age that were accreted to the North American craton during the Middle and Late Ordovician. By defining the various assemblages of lavas and diabases closely associated within the ophiolitic complexes of Thetford Mines, Asbestos, Orford and Bolton, we hope to unravel their original relationships, strongly disturbed during the Taconian and Acadian orogenies, and to restore the geodynamic environment of formation. The ophiolites contain several assemblages of extrusive and intrusive rocks that can be discriminated using field, petrographic and geochemical criteria.

In the Thetford Mines and Asbestos ophiolites occur two stratigraphically distinct groups of volcanic rocks (Laurent and Hébert, in prep.). The Ophiolitic Lower Volcanic Group (O.L.V.G.) is bounded by gabbroic rocks at the base and by red clays, cherts or polygenic breccias at the top (Laurent and Hébert, 1977; Laurent et al., 1979) O.L.V.G. is composed of aphyric, poorly vesicular dark green basaltic lavas, and of porphyric, vesicular light green boninitic lavas showing well preserved quench textures. This group is metamorphosed into greenschist facies with chlorite predominating in basalts and actinolite in boninites. Relict clinopyroxene is augite in basaltic flows, whereas it is Cr-diopside in boninites. In addition, Cr-rich (60-64% Cr_2O_3) spinels are preserved in mafic boninites. The Ophiolitic Upper Volcanic Group (O.U.V.G.), which is overlying O.L.V.G. is metamorphosed into greenschist facies with extensive development of fibrous actinolite. Both O.L.V.G. and O.U.V.G. occur in the volcanic sequence of the Orford ophiolite. This ophiolite is remarkable for having a unique 1000

Figure 1: Geological map of the southwestern part of the Québec Appalachians compiled from various sources by P. St-Julien (after St-Julien et al., 1983).

Explanation
1. Grenville basement (pC)
2. St.Lawrence platform (CO)
3. Foreland thrust belt (O)
4. Allochthons of the external domain (CO)
5. Allochthons of the internal (mainly Bennett Schist) (CO)
6. Caldwell Group and Mansonville Formation (C)
7. Ophiolites (C)
8. St.Daniel and brompton Formations (mélange) (LO)
9. St.Victor synclinorium (Magog Group)(MO)
10. Ascot-Weedon Formation (L-MO)
11. Connecticut Valley-Gaspé synclinorium (SD)
12. Frontenac Formation (O?)
13. Chain Lakes Massif (Helikian)
14. Ordovician granites
15. Devonian granites
16. Mesozoic alkaline intrusive rocks.



9 also: GIRGAB: Groupe Interuniversitaire de Recherches Géologique en Analyse de Bassins.

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meters thick Sheeted Sill Complex consisting of about 30% of boninitic sills and 70% of TiO_2-rich basaltic sills. The TiO_2-rich rocks are relatively fresh augite-plagioclase-phyric diabases. South of Orford, Ordovician black pelagic shales (similar to the shales overlying the O.U.V.G.) host pillowed basalts known locally as the Bolton lavas. These greenish basalts, which have accumulated in hundreds of meters thick piles, are sparsely phyric and poorly vesicular. They have abundant microphenocrysts of plagioclase and lesser amounts of altered olivine.

Major and trace elements allow to classify these rocks into three distinctive groups that are: (I) normal and transitional M.O.R.B. (Bolton lavas and Orford TiO_2-rich diabases respectively):, (2) arc tholeiites (basaltic rocks of O.L.V.G.), and (3) boninites (boninites of O.L.V.G. and the O.U.V.G.). These groups derive from at least two magmatic suites of distinct origin. Group 1 represents ocean floor and groups 2 and 3 island-arc assemblages. In the island-arc assemblages, SiO_2 decreases with Mg# while Al_2O_3 and P_2O_5 increase suggesting fractionation of clinopyroxene (and orthopyroxene ?) and no extensive fractionation of olivine, spinel, plagioclase and apatite. Larger variations of Ni and Cr contents in some boninites imply local fractionation of small amounts of orthopyroxene or olivine + spinel. Regular increases of V, Ti and Fe contents with decreases in Mg# suggest no significant fractionation of Fe-Ti-V oxides. Boninites have high Mg# values, high Ni and Cr contents and flat chondrite-normalized patterns and some samples show typical U-shape patterns; they are very depleted in REE (LREE 1.5 to 4 times chondrite). Arc tholeiites also are depleted in LREE (2 to 10 times chondrite) and their Ce/La, Zr/Y and La/Sm ratios are similar to those of the boninites indicating close magmatic consanguinity. Compared to the island-arc assemblages, normal and transitional MORB are respectively characterized by moderate to strong enrichments in incompatible elements (Zr, Y, Th, Hf, etc.) and by moderate to low concentrations of compatible elements (Ni, Cr, etc). They show LREE fractionation and flat chondrite-normalized patterns. The Orford TiO_2-rich diabases appear to be distinct from the Bolton lavas having LREE 15 to 30 times chondrite and much higher incompatible element contents.

These specific geochemical signatures call for sources ranging from normal chondritic mantle to strongly depleted mantle sources. Bolton lavas and Orford TiO_2-rich diabases are likely to have been emplaced along spreading ridge and plume-type spreading ridge segments respectively. These segments were probably part of an active back-arc basin system. Arc tholeiites formed in the early stages of intra-oceanic arc development while boninites were extruded during a fore-arc rifting event. Features of the ophiolitic assemblages suggest that the Ordovician ophiolites of the Quebec Appalachians originate from a suprasubduction zone environment. The Thetford Mines ophiolite is considered to be typical of ocean crust created by fore-arc rifting over a subduction zone. The Bolton lavas and Orford ophiolite are interpreted to have originated through back-arc spreading volcanism and plume-type ridge segments related to the major intra-oceanic arc system.

INTERPRETATION OF 40AR/39AR AMPHIBOLE AGE SPECTRA, VERMONT: EASIER SAID THAN DONE.

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Introduction

Total fusion $^{40}Ar/^{39}Ar$ amphibole ages from greenschist, epidote-amphibolite, and blueschist facies mafic rocks may seem to give a fairly straight-forward picture of metamorphic ages in Vermont west of the Connecticut-Valley Trough (CVT). Taconian metamorphism with total fusion amphibole ages between 471 and 439 Ma is recorded in north-central Vermont, and Acadian metamorphism with a total fusion amphibole age of 376 Ma is recorded in southeastern Vermont (Fig. 1; Laird et al. 1984). However, petrologic data indicate polymetamorphism, and therefore total fusion ages may "average" the timing of metamorphic events.

Devonian total fusion amphibole, biotite, and muscovite ages between 376 and 386 Ma from garnet to kyanite grade rocks, west-central Vermont (MG, Fig. 1), are interpreted as Taconian cooling ages by Sutter et al. (1985) and as representing Acadian metamorphism by Laird et al. (1984). At Elmore Mountain (EM, Fig. 1) zoned amphibole (magnesio-hornblende cores and actinolite rims) gives a total fusion age of 446 Ma and a disturbed 40Ar/39Ar age spectrum. These data are interpreted by Laird et al. (1984) to record Taconian and Acadian metamorphism at 470 and 380 (?) Ma, consistent with amphibole and muscovite ages in the Worcester Mountain (WM, Fig. 1). However, these data do not mean that all zoned amphibole in north-central Vermont record both Taconian and Acadian metamorphism. In spite of extensive petrologic and isotopic studies, the distribution and relative physical conditions of Taconian and Acadian metamorphism in Vermont are not completely understood.

Further and more detailed age spectrum studies on zoned and unzoned amphibole were undertaken to respond to these controversies/questions. Figure 1 summarizes previous $^{40}Ar/^{39}Ar$ amphibole age data and gives the localities discussed below. Names of units are from Doll et al. (1961). All amphibole names follow the nomenclature given by Leake (1978).

Southeastern Vermont, west of the Athens Dome (V118, Ottauquechee Fm., TD, Fig. 1; V119, Pinney Hollow Fm., sample directly north of TD, Fig. 1)

All samples are garnet amphibolites. Amphibole defines the foliation and is variably lineated. It is medium-grained in samples V118BB and V119. Both medium- and coarse-grained amphibole occur in V118CC. Amphibole is tschermakite and is only weakly and not systematically zoned. Medium- and coarse-grained amphibole are not compositionally distinct in V118CC. Garnet-biotite geothermometry using the experimental calibration of Ferry and Spear (1978) gives temperatures of 479°C at locality V119 and 496 to 555 °C at locality V118 (Laird and Albee, 1981, Table 2).

The majority of gas released from amphibole from sample V118CC gives an average age of 380 ± 2 Ma (Fig. 2a). A saddle-shaped spectrum is given by amphibole from V118BB (Fig. 2b). The age minimum is at 389 ± 2 Ma. A saddle is shown by the spectrum from V119 amphibole (Fig. 3) with a

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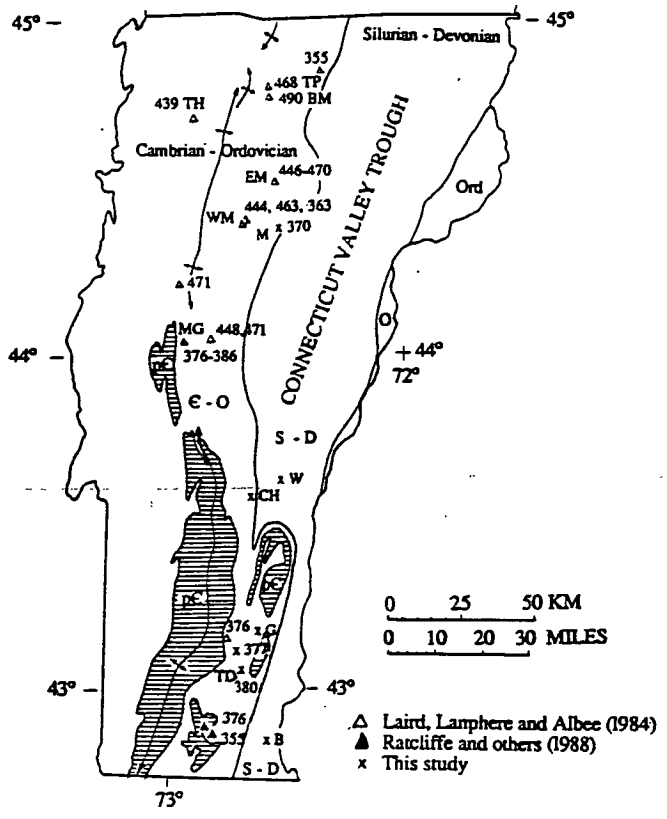

Fig. 1: Location of 40Ar/39Ar amphibole age data from Vermont.

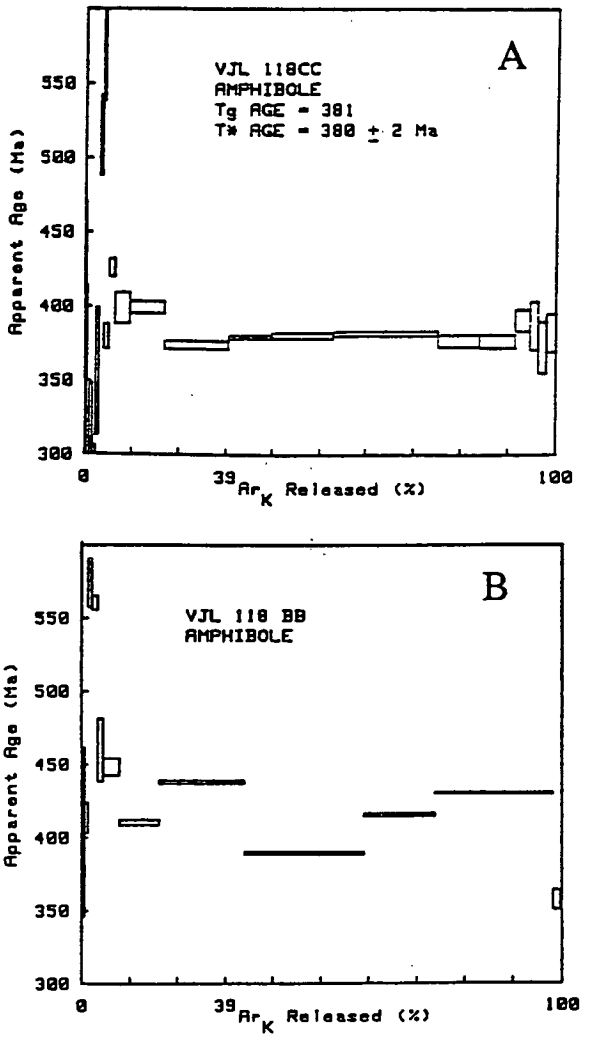

Fig. 2: 40Ar/39Ar age spectra from TD, Fig. 1. a) Tg is the total gas age and T* the plateau age. b) The spectrum for this sample leaves room for interpretation.

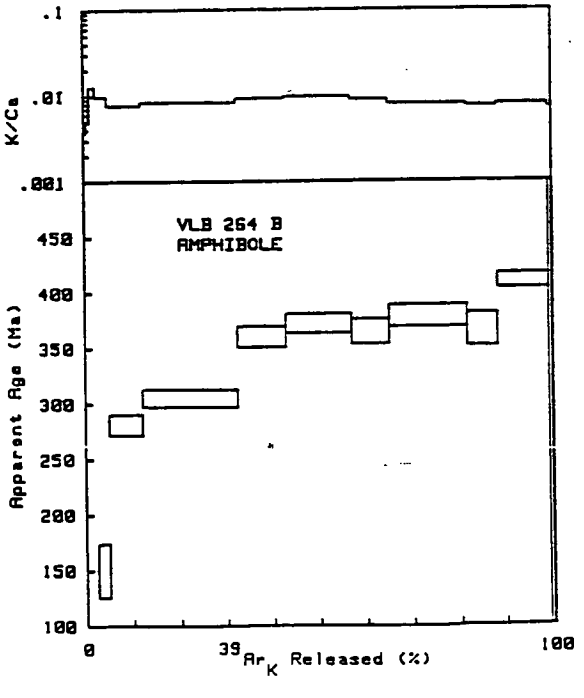

Fig. 4: 40Ar/39Ar age spectrum and K/Ca for zoned amphibole at M, Fig. 1.

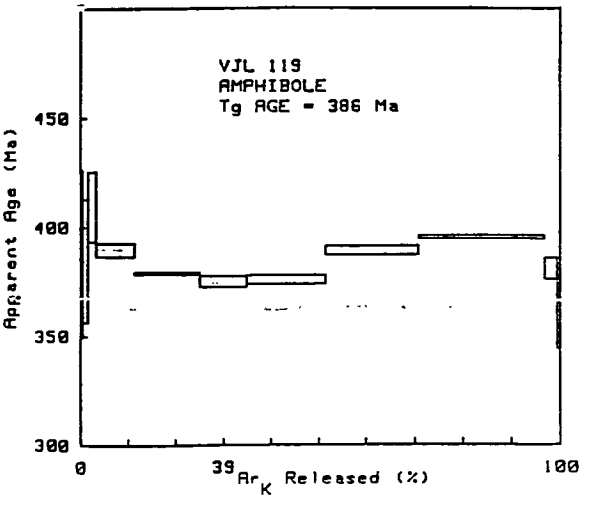

Fig. 3: 40Ar/39Ar age spectrum for amphibole north of TD, Fig. 1 Tg is total gas age.

minimum age of 377 ± 2 Ma and a total gas age of 386 Ma. These data and those obtained earlier (Fig. 1) support Acadian metamorphism at about 380 Ma. The saddle-shaped spectra leave room for interpretation. A "generic" explanation is excess Ar, possibly greater in finer grained samples.

North-central Vermont, east of the Worcester Mountains (V264B, Missisquoi Fm., M, Fig. 1)

Amphibole from metadiabase is primarily actinolite with minor, interior, magnesio-hornblende zones. The amphibole grains are poikiloblastic with epidote and titanite inclusions; they may be pseudomorphs of pyroxene.

An age spectrum shows diffusive argon loss (Fig. 4). The 370 Ma age at which the spectrum "levels off" may be the age of magnesio-hornblende cores because K/Ca increases where the apparent age jumps to about 370 Ma; K/Ca = 0.03 for the magnesio-hornblende and 0.01 for the actinolite (catatom units). The actinolite would thus be younger. An alternative explanation is that the gas fraction at about 410 Ma represents the magnesio-hornblende. Regardless, this last age fraction is experimentally real.

North-central Vermont, Tillotson Peak (VLB319E, Belvidere Mountain Amphibolite Member of the Hazens Notch Fm., TP, Fig. 1)

Glaucophane from this area gives a total fusion age of 468 ± 6.4 Ma (sample V337C, Laird et al. 1984). Amphibole in coarse-grained garnet amphibolite from a folded shear zone is zoned from magnesio-hornblende cores to actinolite rims and gives an uninterpretable scatter of ages during step-wise heating. Apparently, there is too little K to produce enough gas.

Conclusion

While it would be nice to have found evidence for Taconian metamorphism, it's not obvious in these spectra. Data support Acadian metamorphism in the Ottauquechee and Pinney Hollow Formations, southeastern Vermont at about 380 Ma and suggest metamorphism at 370 Ma in the Missisquoi Formation, north-central Vermont. Even though we may argue as to the meaning of the age spectra, examination of Figures 2b and 3 shows that total fusion ages may be misleading even if samples are not polymetamorphosed.

For the purpose of discussion J.L. and W.A.B. suggest that east of the Worcester Mountains a fault zone separates the Stowe and Missisquoi Formations, explaining the difference in structural style between the two units and the lack of medium-high-pressure facies series amphibole and Taconian metamorphism (?) in the Missisquoi Formation compared to the Stowe Formation. A northwest-southeast-trending thrust in central Vermont such as the southern extension of the Hinesburg thrust shown by Sutter et al. (1985, Fig.4A) may separate mafic rocks giving Ordovician ages to the north (Fig. 1) and showing evidence of medium-high-pressure facies series metamorphism from mafic rocks giving Devonian ages to the south and showing medium-pressure facies series metamorphism only.

Samples are being "processed" to determine the age of metamorphism of the Barnard Volcanic Member, Missisquoi Formation, central Vermont (CH, Fig. 1) and to address the following questions:

1) Is the total fusion age of 439 Ma at TH (Fig. 1) a "mixed age"? The amphibole is zoned compositionally.
2) Does apparent age change with metamorphic grade and grain size within the same unit?

3) How do age spectra of amphibole west of and within the CVTcompare in central and in southeastern Vermont?
4) What will laser studies on chemically zoned amphibole show?

Isotopic data to the south are discussed by Sutter et al. (1985) and Ratcliffe et al. (1988). It is hoped that isotopic data to the north and east will also be covered formally or informally during this conference.

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ASPECTS OF THE TACONIAN OROGEN IN CENTRAL VERMONT

R.S. Stanley, R.A. Coish, and Jo Laird

FIGURES

Québec - Vermont Appalachian Workshop

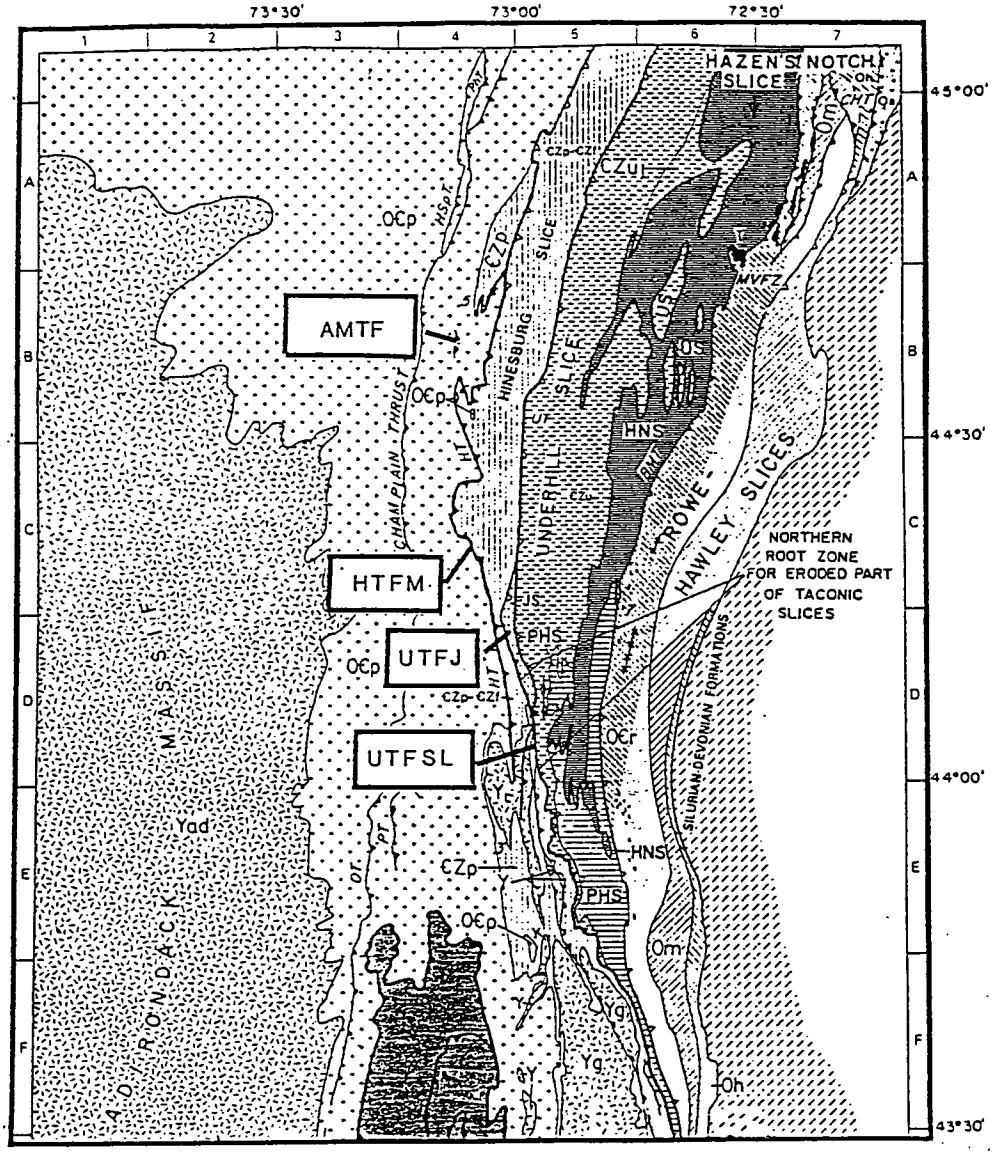


Figure 1 Interpretative Tectonic Map of Vermont and eastern New York showing the general location of the Arrowhead Mountain thrust fault (AMTF), the Hinesburg thrust fault at Mechanicsville (HTFM), and the Underhill thrust fault at Jerusalem (UTFJ), and South Lincoln (UTSL). The geological map is taken from Stanley and Ratcliffe (1985, pl. 1, fig. 2A). Symbol T in A6 is the 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 if Albee (1968). Widely-spaced diagonal lines in north central Vermont outline the region that contains medium-high pressure amphiboles described by Laird and Albee (1981). Irregular black marks are ultramafic bodies. 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 the Underhill Formation; OCr, Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation and its equivalents in Vermont; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens North 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.

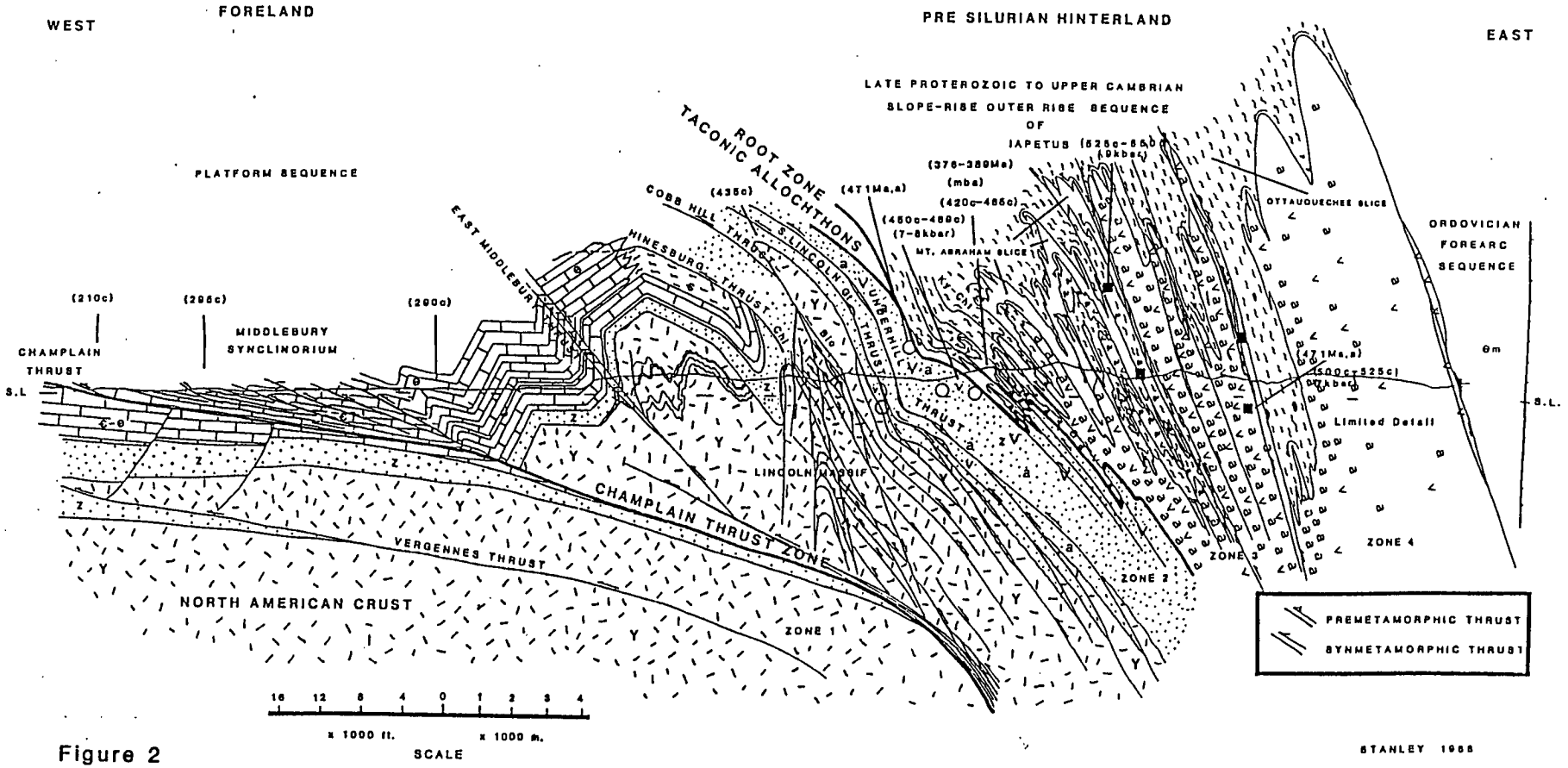
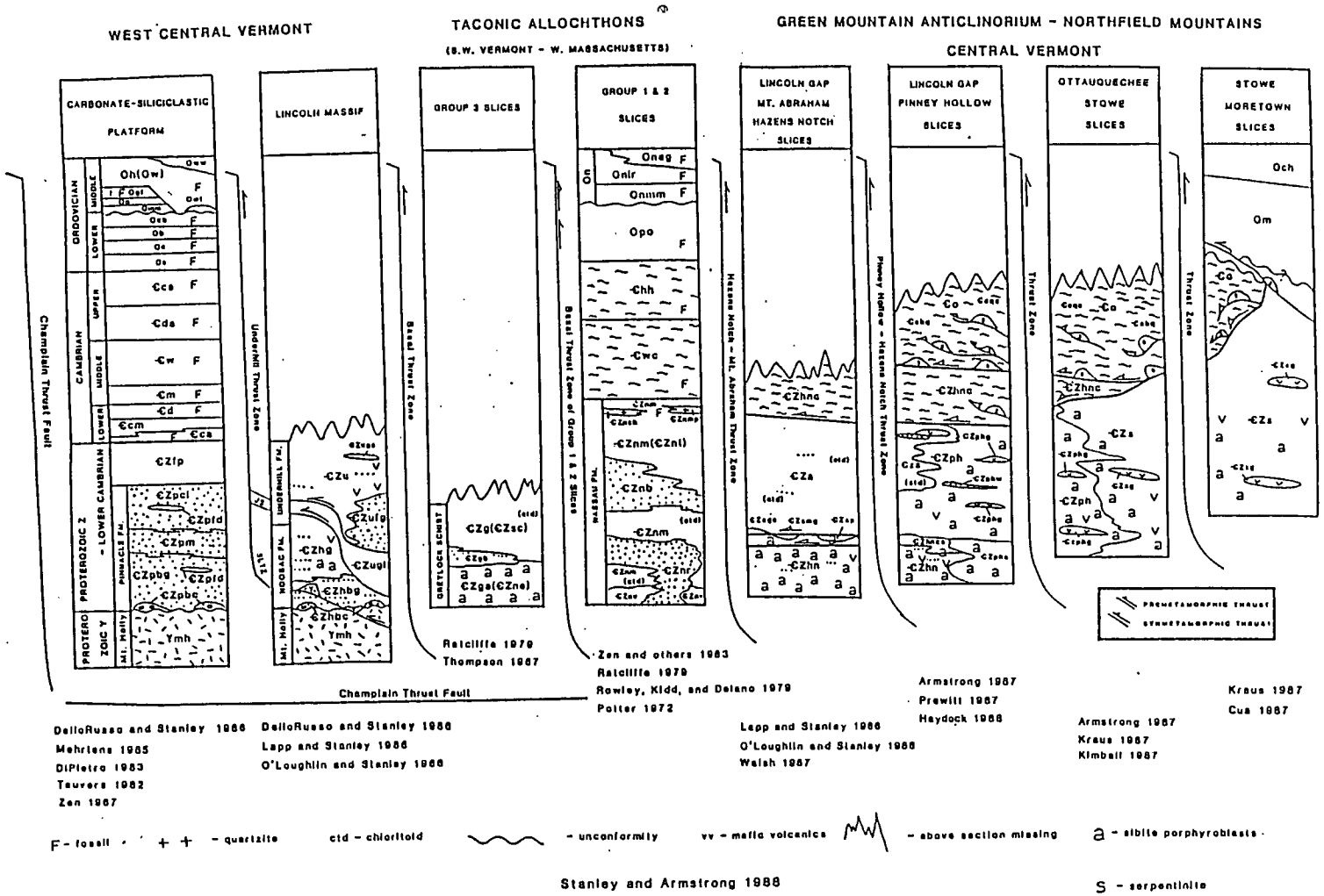


Figure 2

Geological cross section through the pre-Silurian foreland and hinterland along a latitude of 44° N. The section is based on mapping by Cady (1945), Washington (1987), and Harding and Hartz (1987) in the foreland, by Tauvers (1982) and DelloRusso and Stanley (1986) in the Lincoln massif, and by Lapp and Stanley (1986), Prewitt (1986), Stanley (1986, 1987), Kraus (1987), and Cady and others (1962) in the hinterland. North American Middle Proterozoic (Y) crust is shown by the random dash pattern. The overlying foreland rocks consist of Late Proterozoic (Z) rift clastic rocks (stippled pattern) and Cambrian and Ordovician platform rocks (carbonate and siliciclastic rocks overlain by shales) shown by the stacked rectangles. Symbols for the deformed and metamorphosed slope-rise and outer rise sequence correspond to those in figure 6. The sequence designations are those used in Stanley and Ratcliffe (1985) and figure 6. The black lenses in the hinterland sequence represent serpentinites. Zones 1 through 4 are based on mafic rock geochemistry (Colah, 1987, 1988). Temperatures of coexisting mineral assemblages are given as (i.e.290°c) and are based on carbon and oxygen isotope analyses by Sheppard and Schwarcz (1970) for the foreland, oxygen isotope analyses by Garlick and Epstein (1967) and calcite-dolomite and amphibole-plagioclase temperatures (Laird and others, 1984) for the western part of the hinterland. The 471 Ma ages are based on 40Ar/39Ar analyses of hornblende by Laird and others (1984). The 376-389 Ma ages are based on 40Ar/39Ar total fusion ages of muscovite and biotite (Laird and others, 1984; Lamphere and others, 1983). The symbol a refers to amphibole, b refers to biotite, and m refers to muscovite. The metamorphic series assignment is based on amphibole analyses by Laird (1987). The solid squares represent hornblendes from the high-pressure metamorphic series and the open circles represent hornblendes from the medium-pressure metamorphic series (see fig. 4). The hinterland rocks have been displaced westward along multiple generations of thrust faults and deformed by numerous generations of folds. The earliest recognized faults probably developed before or during the early high-pressure metamorphism (solid squares; fig. 4). These faults are deformed by reclined or sheath-like folds with axes oriented parallel or nearly parallel to the pervasive southeast-plunging grain lineation. These folds and the older faults are in turn cut by synmetamorphic thrust faults with well-developed mylonitic fabrics. Based on stratigraphic and sedimentologic information, it is suggested that the root zone for the northern extension of the Taconic allochthons is located along a complex zone of pre- to late metamorphic faults that forms the western boundary of the albite-rich and aluminous rocks of the Hazens Notch, eastern Underhill and Mt. Abraham formations. This location is supported by the fact that the metabasalts in the Taconic allochthons are chemically more similar to the mafic rocks of the western Underhill and Pinnacle formations (Zone 2) than [illegible] are to the mafic rocks to the east in Zones 3 and 4. The [illegible] Middle Proterozoic rocks of the Lincoln massif consist [illegible] major anticlines in which the eastern one has been severely flattened by numerous thrust faults which contain Paleozoic mylonites with east-over-west fabrics. A third anticline is buried beneath the platform sequence to the west of the East Middlebury thrust (Harding and Hartz, 1987). The breakthrough faults (East Middlebury thrust) and geometry of the western anticline suggests that the three anticlines of Middle Proterozoic rocks began their development as fault-propagation folds (Suppe, 1985) and evolved to flattened massifs and isolated slivers by the development of axial-surface cleavage and extensive high-angle faults. The available PT information suggests that this deformation occurs over temperatures that range from 290°C to the west to 435° C to the east. The temperatures would correspond to pressures in the range of 3 kbars (10 km) to the west and 4.5 to 5.0 kbars (16 to 18 km) to the east assuming a standard geothermal gradient of 20°C to 30°C per kilometer (Strehle and Stanley, 1986, fig. 5). This information again suggests that there was a significant tectonic load over the Lincoln massif when these structures formed. The minor imbricate faults of the Middlebury synclinorium are based on work by Washington (1987). The slice of North American crust floored by the Vergennes thrust is an interpretation based on some unpublished seismic reflection work. Shortening across this section is estimated to be between 400-500 km, of which approximately 70 km involved the folding and faulting of North American crust.



Stanley and Armstrong 1988

Figure 3

EXPLANATION FOR LITHOTECTONIC CORRELATION CHART FOR CENTRAL VERMONT- The lithic symbols shown in each column are explained by geographic locality starting with the oldest unit.

West Central Vermont - Ymh- granitic gneiss of the Mt. Holly Complex, Y - mylonitic gneiss at South Lincoln, CZpbc - basal conglomerate in the Pinnacle Formation, CZpbg - biotite wacke, CZpfd - Forestdale Marble, CZpm - muscovite wacke, CZpcl - chlorite wacke and chlorite schist, CZhbc - basal conglomerate in the Hoosac Formation, CZhbg - biotite wacke, CZhg - schistose wacke, CZugl - quartzite laminated schist of the Underhill Formation, CZufg, well foliated wacke, CZu - mixed unit consisting of garnet, chlorite, muscovite schists, thin quartzite, CZugs - garnetiferous schist, CZfp - gray, finely laminated phyllite of the Fairfield Pond Formation, Cca - argillaceous quartzite of the Cheshire Formation, Ccm - massive, bedding quartzite of the Cheshire Formation, Cd - Dunham Dolomite, Cm - Monkton Quartzite, Cw - Winooski Dolomite, Cda - quartzite and dolostone of the Danby Formation, Ccs - Clarendon Springs Dolomite, Lower Ordovician Beekmantown Group consists of limestone, dolostone and minor quartzite and shale of the Shelburne Formation (Os), Cutting Dolomite (Oc), Bascom Formation (Ob), and Chipman Formation (Ocb), Omm - Middlebury Limestone, Oo - Orwell Limestone, Ogf - Glens Falls Formation, Owl - fossiliferous limestone lenses near the base of the Walloomsac Formation, Oh - black carbonaceous slate and phyllite of the Hortonville Formation, Ow - dark gray, graphitic schist and phyllite of the Walloomsac Formation, Oww - Whipstock Breccia Member of the Walloomsac Formation.

Taconic Allochthons - Group 1 and 2 - CZnr, Rensselaer Graywacke Member of the Nassau Formation, CZnv - metabasalt and basaltic tuff, CZnm - lustrous, yellowish-green, purple laminated chloritoid - chlorite phyllite of the Mettawee Member, CZnt - Truthville Slate, CZnb - Bomoseen Graywacke Member, CZnzh - Zion Hill Quartzite, CZnmp Mud Pond Quartzite, Cwc - dark gray to black, pyritiferous and calcareous slate with thin sandy laminae of the West Castleton Formation, Chh - bluish-gray weathering black sulphidic slate and chert, Opo - white weathering, well laminated gray slate and chert, On - Normanskill Formation, Onmm - black slate and thin bedded cherts of the Mount Merino Member, Onir - red and green slate of the Indian River Member, Onag - Austin Glen Graywacke Member

Group 3- CZga, Light green to gray, white albitic schist with some magnetite, chlorite granofels of the Greylock Schist, CZne - Netop Formation in the Dorset Mt. slice, CZgb - black of dark gray chloritoid or stilpnomelane albite, quartz knotted schist, lenses of feldspathic quartzite, conglomerate, and pink dolostone, CZg - light green, lustrous chloritoid phyllite and minor beds of white albitic schist, CZsc - St. Catherine Formation in the Dorset Mt. slice.

Green Mountain anticlinorium - Northfield Mountains (Many of the symbols are repeated from column to column. These symbols are explained in the first column that they appear in as the columns are read from left to right)

Lincoln Gap-Mt. Abraham-Hazens Notch slices - CZhn silvery white to dark green to black schist spotted with white albite, minor white to gray laminated quartzite of the Hazens Notch Formation. Mafic schist abundant in the eastern part,CZhnc - rusty weathering, black albitic schist with widespread graphite and minor thin black or gray quartzite. CZhnc has been called the Battell Member of the Underhill (Doll and others, 1961) and the Granville Formation (Osberg, 1952). CZa - Mt. Abraham Schist. Silvery colored paragonite - muscovite - chloritoid - chlorite (garnet) schist with a distinctive pearly sheen on the schistosity, CZamg - similar to the main body of Mt. Abraham Schist but with abundant magnetite and chloritized garnet, CZags - Muscovite - chlorite schist of the Mt. Abraham Schist with large porphyroblasts of garnet and minor chloritoid,

Lincoln Gap-Pinney Hollow slices - Additional symbols are: CZhnca - rusty weathering, dark gray to black albite schist with discontinuous patches of graphite. Traceable into the Granville Formation. CZph - silvery green muscovite - chlorite - quartz schist of the Pinney Hollow Formation, CZpha light gray muscovite - chlorite -quartz schist with albite porphyroblasts, CZphg - mafic schists of the Pinney Hollow, CZphw - gray wacke with minor blue quartz, Co - black, pyritiferous and graphitc schist of the Ottauquechee Formation, Cobq - black and gray quartzite of the Ottauquechee Formation, Coqs - sandy quartzose schist of the Ottauquechee Formation, s - serpentinite and talc-carbonate rock.

Ottauquechee - Stowe slices - Additional symbols are: CZs - Stowe Formation - silvery green, muscovite - chlorite - quartz schist identical to CZph, some schists are richer in chlorite, CZsg - mafic schist in the Stowe Formation.

Stowe-Moretown slices - Om, pinstriped schist and mafic schist of the Moretown Formation, Och - black, graphitic schist and thin quartzite of the Cram Hill Formation.

The names below each column refer to the principal sources of information.

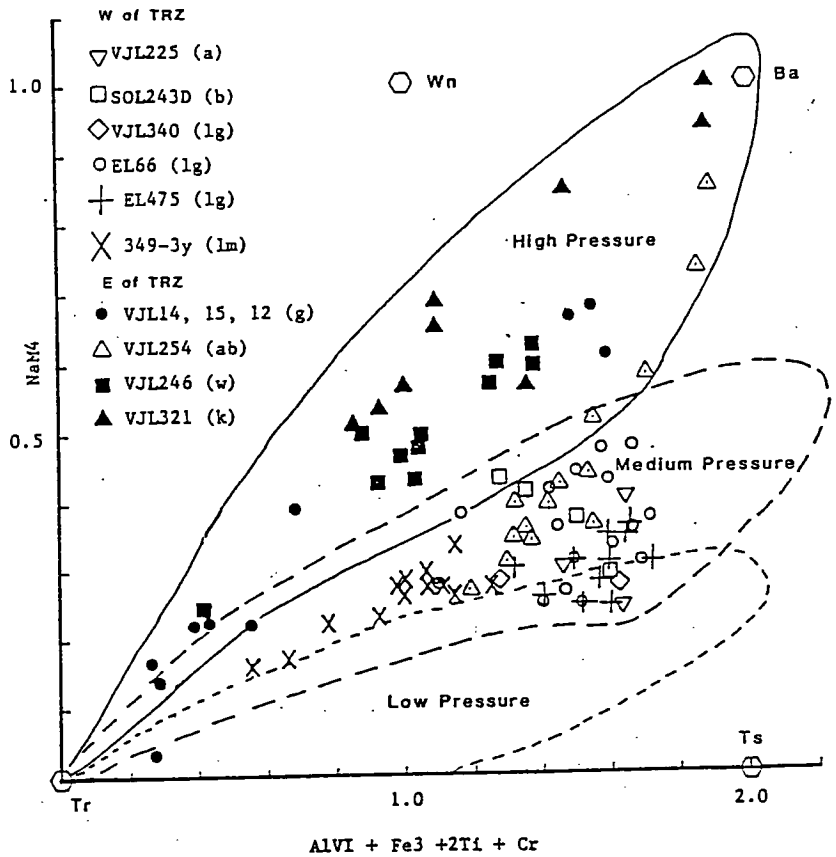


Figure 4

Formula proportion NaM^4 versus ($AlVI + Fe^3 + 2Ti + Cr$) for amphibole - chlorite - epidote - plagioclase - quartz schist from central Vermont. Sample localities identified by VJL numbers are from Laird and others (1984). Localities in ( ) are from areas west (W) of in the inferred Taconic Root Zone (TRZ) or from areas east (E) of the Taconic Root Zone as shown in figure 2. Areas of high-, medium-, low-pressure facies series metamorphism are from Laird and others (1984). Increasing temperatures of metamorphism results in increasing advancement along the X and Y axes. Tr (tremolite), Ts (tschermakite), Wn (winchite), BA (barroisite). Analyses are electron microprobe data and normalized to total cations in which (Na + K) = 13. The mafic volcanic rocks from east of the Taconic Root Zone have core compositions indicating higher metamorphic grade than rim compositions. For example, at Granville Gulf (VJL 12, 14, 15), where the 471 Ma age has been obtained (fig. 2), actinolite overgrew barroisite. Because the actinolite rims add to the total amphibole in the rock, they record a second, but lower temperature and pressure metamorphism rather than an simple alteration of an earlier barroisite during cooling. In some of the samples west of the TRZ (VJL 225 and 340) the zoning is continuous and indicates progressive metamorphism whereas samples from the Lincoln massif (349-3y) farther to the west record decreasing temperature with time with the original temperature being less than the overlying cover to the east. The zoning in the other samples west of the TRZ is more difficult to interpret because it is irregular and may be related to miscibility. This figure is taken from Laird (1987).

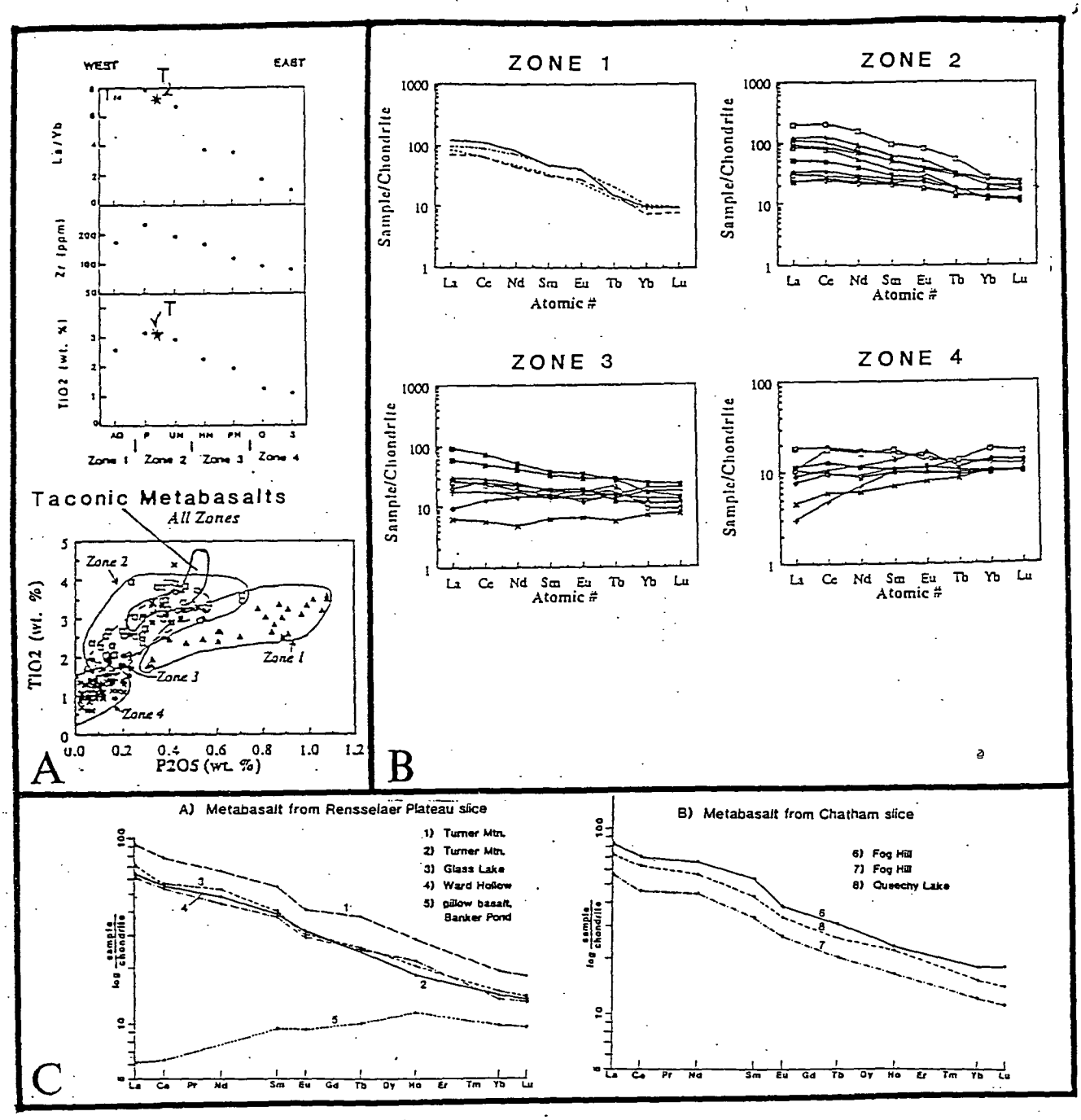


Figure 5

Geochemistry of mafic rocks in central Vermont (Coish, 1987; Coish and Sinton, 1988) and in the Rensselaer Plateau and Chatham slices of the Taconic allochthons (Ratcliffe, 1987). Diagrams in A show the average values of TiO_2, Zr, and La/Yb (a measure of LREE enrichment) and the variation in TiO_2 vs P_2O_5 (lower graph) for mafic rocks of basaltic composition from Zone 1 through Zone 4 from Vermont. The symbol "T" in the upper graph represents the average values for TiO_2 and La/Yb as given by Ratciffe (1987, table 1 and fig. 6B). The values for TiO2 and P_2O_5 are also plotted for the Taconic metabasalts. They show a distribution that overlaps with Zone 2 and Zone 3 rocks. Graphs in B show the distribution of LREE from mafic rocks from Zone 1 through Zone 4 in Vermont. Zone 1 is characterized by enriched LREE. Samples from Zone 2 show a general pattern that is similar to Zone 1. Several samples, however, are less enriched in LREE compared to the HREE. Zone 3 samples are similar to Zone 1 and 2 in that they are high in REE but their patterns are generally much flatter than the other two. The group as a whole shows patterns that are similar to several samples in Zone 2 and 4. Zone 4 samples have lower concentrations of REE than the other zones and either show flat patterns or are slightly or moderately depleted in LREE compared to the HREE. They are therefore quite different form all the mafic rocks in Zone 2 and many of the mafic rocks in Zone 3. The graphs in C show the REE distribution for 8 metabasalts from the Nassau Formation in the Taconic allochthons (Ratcliffe, 1987). These patterns are similar to Zone 2 mafic rocks in central and northern Vermont. Graphs in A, B, and C indicate that the Taconic metabasalts are quite similar to Zone 2 mafic rocks although they do show similarities to some samples from Zone 3.

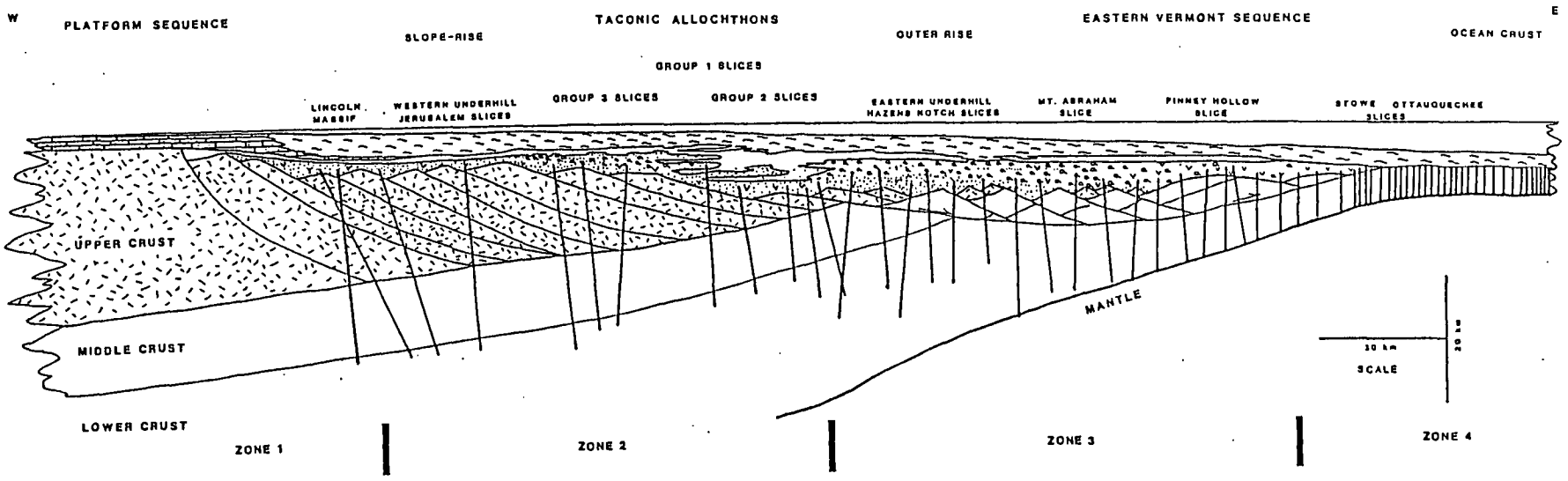


Figure 6

Simplified interpretative diagram of the ancient passive margin of North American with Iapetus during the Late Proterozoic through the Upper Cambrian. The diagram was modified from the model of a lower-plate margin as described by Lister and others (1986). This particular model is preferred over other models because it allows for a wide passive margin with relatively thin but somewhat different stratigraphic sequences deposited in various sedimentary depressions or basins which can be separated by relatively thin stratigraphic sequences. Restoration of the Taconic allochthons to their original depositional site east of the Green Mountain and Lincoln massifs and retrodeformation of the resulting Taconian hinterland suggests that the margin was at least 350 to 400 km wide, of which the carbonate-siliciclastic platform comprised at least 60 km of the total figure (see also Stanley and Ratcliffe, 1985, plate 2). This range is comparable to the Australian passive margin southeast of the Timor trough in the Indonesian convergent arc system. The present stratigraphic thickness of the Taconic allochthons has been estimated to be in the order of .6 km to 1.8 km. (Potter, 1972; Rowley and others, 1979). The original thickness of the slope-rise sequence was probably in the order of 4 km considering the fact that the preserved section is incomplete and has been dewatered and compacted. This value, although small, is comparable to the stratigraphic thickness along modern passive margins where the margin is wide and continental crust is being subducted beneath oceanic crust. For example, the slope-rise sediments along the Asian continental margin west of Taiwan are at least 7 km thick (Suppe, 1981) whereas those in the Browse Basin on the Australian passive margin southeast of Timor are at least 5 km thick (Hamilton, 1978). Although many of the rocks in the Taconic allochthons are lithically similar to rocks in the Taconian hinterland, the differences, particularly in the lower rift-clastic sequence, are significant and suggest deposition in somewhat isolated basins (Ratcliffe, 1987). For example, the rift clastic sequence in the Taconics allochthons is rich in metawacke, metasiltstone, and metasandstone whereas many of the correlative rocks in the Taconian hinterland are rich in albitic schist and relative poor in metawacke and metasandstone. During subsequent deformation, which is considered to be largely Taconian (Middle Ordovician), the passive margin sequence was shortened by at least 80 percent to produce the geological section shown in the Central Vermont Transect (fig. 2).

Lithic symbols for the sedimentary cover are the following: <u>the platform sequence</u>, shown by the rectangle and dot pattern, consists of carbonate and siliciclastic rocks of the Cheshire, Dunham, Monkton, Winooski, Danby, Clarendon Springs and equivalent formations. <u>The slope-rise and outer rise sequence</u> consists of four groups of rocks each of which is characterized by a dominant set of rocks which is considered distinctive for that group. Contacts between the groups are gradational. The basal sequence consists of two groups. <u>The stippled pattern</u> represents metawacke, metaconglomerate, metasandstone, metasiltstone and quartz-rich pelitic rocks of the Pinnacle and Underhill (western part) Formations in central Vermont, the Rensselaer Graywacke Member and Bomoseen Graywacke Member of the Nassau Formation in the Taconic allochthons (group 1 and 2) and minor, but mappable, units in the Hoosac, Underhill (eastern part), Hazen Notch and Pinney Hollow Formations in central Vermont. <u>The pattern with random "a" symbols</u> are largely quartz-bearing pelitic rocks with distinctive porphyroblasts of albite typical of such formations as the Greylock Schist and the Natop Formation in the Taconic allochthons (group 3), the Hoosac Formation of western Massachusetts and southern Vermont, the Underhill (eastern part), Hazens Notch, Pinney Hollow, and Stowe in central Vermont. Albitic schists are also present in the Hoosac and western part of the Underhill Formations in central Vermont. The basal two groups are characterized by mafic volcanics of basaltic composition (v symbol). These rocks are commonly absent from the upper two groups. <u>The blank pattern</u> of the central group consists of aluminous pelitic rocks which generally contain chloritoid at the appropriate metamorphic grade. Metasiltstone and metasandstone are minor. Rocks of this group are found in the Nassau Formation (Mettawee Member), the Everett Schist, the Greylock Schist, and the St. Catherine Formation in the Taconic allochthons, the Hoosac Formation on the east limb of the Berkshire massif, and the Mt. Abraham Schist in central Vermont. The Fairfield Pond Formation of central and northern Vermont is included in this group although it does not contain chloritoid nor does it appear to be as aluminous as the others. <u>The wavy dashed pattern</u> of the upper group consists of dark colored, rusty or brown weathering, carbonaceous pelitic rocks that contain interlayered grey, black, or white metasandstone and metasiltstone. Thin carbonate beds and limestone conglomerates are present in the Taconic allochthons (group 1) and the western part of the group. Rocks of this group are found in the Skeels Corner and Sweetsburg Formations of Vermont and Quebec, the West Castleton and Hatch Hill Formations of the Taconic allochthons (Group 1), the Hazen Notch, Battell, Granville, and Ottauquechee Formations of central Vermont. Mafic volcanic rocks of basaltic composition are rare in the upper two groups.

The rocks in each of the aforementioned groups provide a basis for interpreting the rift-drift sequence for the Late Proterozoic to Upper Cambrian margin. The rocks of the basal groups and their associated basaltic volcanics represent the clastic rocks deposited in basins and depressions during the rifting of the ancient margin. The presence of granitic clasts, blue quartz, and abundant secondary albite in many of these rocks supports this interpretation. The minor and trace element geochemistry of the associated mafic volcanics systematically vary from west to east (Coish, 1987; Coish and Sinton, 1988) and also support the rift interpretation. All the rocks have basaltic compositions. Those to the west (Zone 1) are transitional to mildly alkaline basalts with high concentrations of Ti, P, Zr, Y and LREE-enriched patterns and are similar to continental rift basalts whereas the volcanics to the east (Zone 4) are tholeiitic basalts with low concentrations of Ti, Zr, P and LREE-depleted patterns and are similar to ocean ridge basalts. The intervening two zones are transitional between Zone 1 and 4. For example, Zone 2 mafic rocks in the Pinnacle and western Underhill formations are similar to Zone 1 except they have lower P and slightly higher Ti contents. The mafic volcanic rocks in the late Proterozoic to Early Cambrian Nassau Formation of the Taconic allochthons have many of the chemical characteristics of Zone 2 volcanics of central and northern Vermont (Ratcliffe, 1987). Support for this comparison is discussed in figure 8. These chemical similarities therefore support the relative paleospastic position of the Taconic allochthons within or slightly east of Zone 2 . Zone 3 mafic volcanics have a mixed chemistry. Some of the volcanics are like those of Zone 1 and 2, whereas other bodies have lower concentrations of Ti, Zr, P, and LREE. The dark colored, carbonaceous rocks in the upper group are generally devoid of volcanic rocks and were deposited seaward of the carbonate-siliciclastic platform during thermal subsidence. The presence of graphite throughout many of these rocks indicates deposition under reducing conditions which were extensive when the polar icecaps were greatly reduced during the lower Paleozoic (Berry and Wilde, 1978). The aluminous rocks of the middle group are interpreted to be the transition from rift to drift conditions.

GEOLOGY AND GEOCHEMISTRY OF METABASALTIC ROCKS FROM THE ROXBURY AREA, CENTRAL VERMONT

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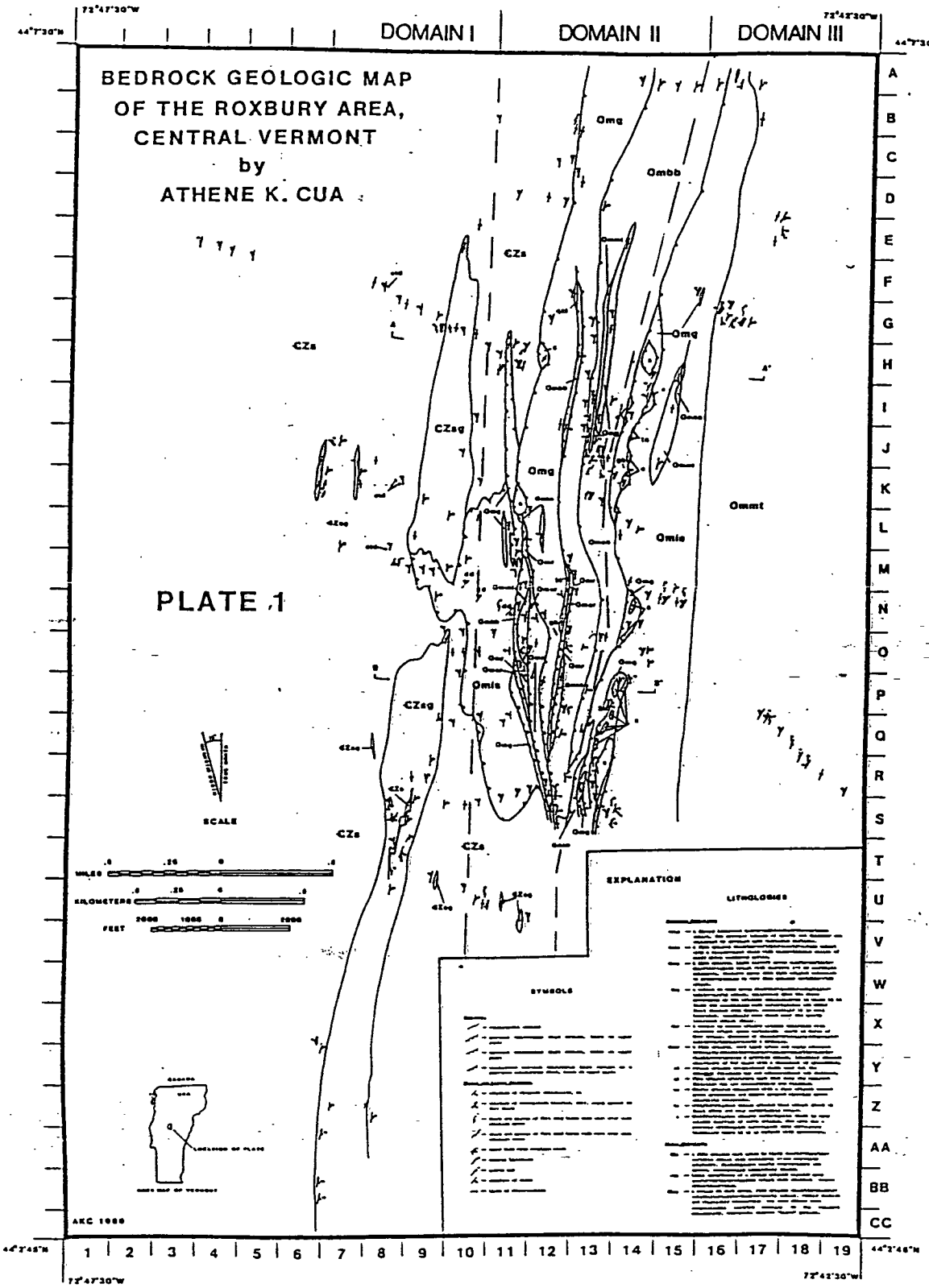
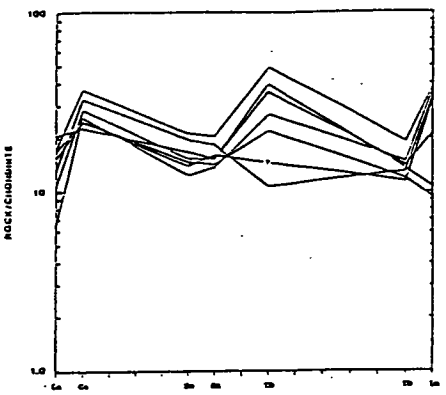
The lithologic units identified in the Roxbury area are metasedimentary and metaigneous rock types of the Stowe and Moretown Formations. The Stowe Formation to the west is divided into two units: a greenish black quartz-sericite-chlorite-albite schist with local occurrences of sericite-chlorite-quartz-chloritoid schist, and an epidote-chlorite-actinolite schist (greenstone). The Moretown Formation to the east contains three metasedimentary rock types: a finely laminated quartz-albite-sericite-chlorite schist, a bluish black sericite-chlorite-quartz schist, and a light green magnetite-bearing schist that ranges in texture from a fine grained metagraywacke to a very fine grained mica-rich schist. The metaigneous rock types of the Moretown include abundant greenstone units, serpentinites and associated ultramafic rocks. Minor amounts of metagabbro are present as well as significant amounts of mariposite-bearing rodingitized gabbros. At least three Paleozoic dikes intrude the quartz-laminated schist, two metadiabasic and one meta-basaltic, all of which are metamorphosed and deformed. Detailed geologic mapping at 1:12000 lead to the recognition of three local greenstone domains. From west to east, Domain I contains Stowe greenstones, Domain II contains Moretown greenstones associated with fault slivers of mariposite-bearing rodingite and two small serpentinite pods, and Domain III contains Moretown greenstones with large internal serpentinite bodies.

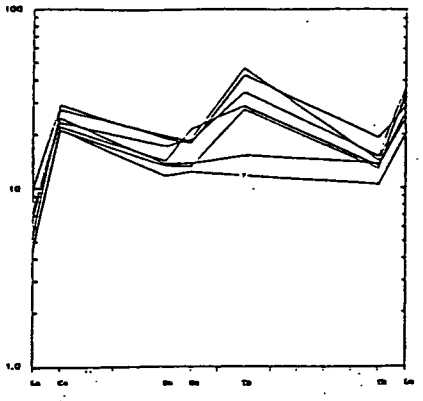
The metaigneous rocks were analyzed for major, trace and rare earth elements to determine their magmatic and tectonic histories. The greenstone domains defined by field criteria cannot be separated geochemically and therefore all the metabasalts are likely to have originated from the same source magma. REE patterns for the metabasalts are flat to slightly LREE-depleted, similar to N-type MORBs or MORB-like ophiolitic basalts. The greenstones plot in MORB fields in various trace element discrimination diagrams. The presence of voluminous ultramafic rocks, rodingitized gabbros and metagabbros strongly suggests that the meta-igneous sequence belongs to a highly dismembered ophiolite suite. The geo-chemistry of the greenstones is consistent with the regional geochemical trend of metabasalts from central Vermont.

The structural elements in the metasedimentary rocks record at least three episodes of deformation. The more competent metaigneous rocks are more resistant to deformation. The oldest fold and foliation generation, Fn-1 and Sn-1, is only found in the Stowe schist, and is not observed in any Moretown samples. The near vertical dominant schistosity, Sn, is pervasive and consistent throughout the field area, strongly influencing the map pattern. A younger foliation, Sn+1, is a spaced cleavage to the west and grades into a crenulation cleavage to the east. The majority of contacts between lithologic units are pre-, syn-, and/or post metamorphic faults. Mineral assemblages indicate the peak metamorphic event was at least at biotite grade. The assemblages were subsequently retrograded at chlorite grade conditions. The peak metamorphic event is believed to be Taconian, and the age of retrograde metamorphism is either late Taconian or Acadian.

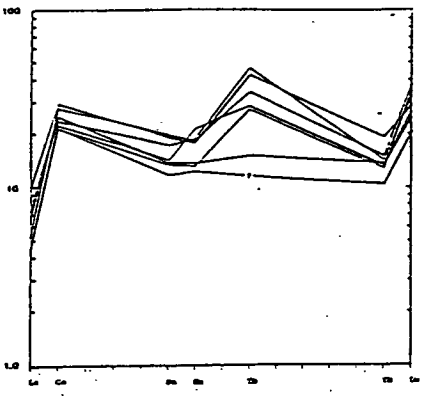

BEDROCK GEOLOGIC MAP
OF THE ROXBURY AREA,
CENTRAL VERMONT
by
ATHENE K. CUA
PLATE 1
DOMAIN I
DOMAIN II
DOMAIN III
Omg
Ombb
CZs
CZsg
Omg
Ommt
Omis
Omis
CZsg
CZs
CZs
CZs
SCALE
EXPLANATION
LITHOLOGIES
SYMBOLS
AKC 1988



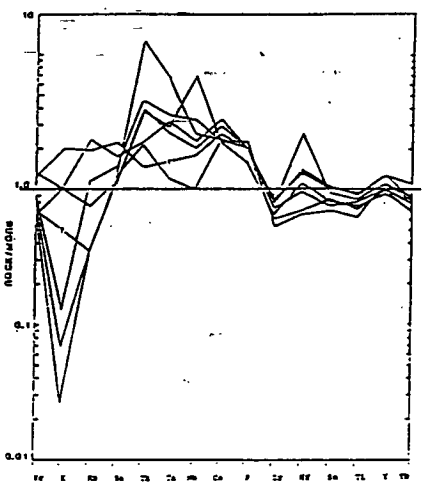
Rare earth element abundances in the Domain I greenstones from the Roxbury area normalized to chondritic values defined by Haskin and others (1968). The pattern is generally flat to slightly LREE depleted, similar to N-type MORB.



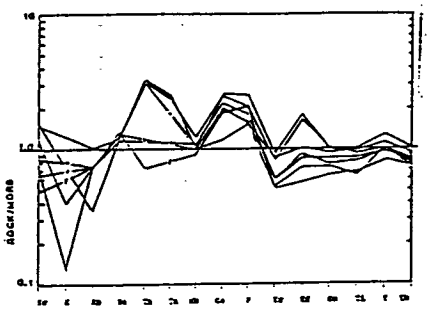
Rare earth element abundances in the Domain II greenstones from the Roxbury area normalized to chondritic values defined by Haskin and others (1968). The pattern is generally flat to slightly LREE depleted, similar to N-type MORB.



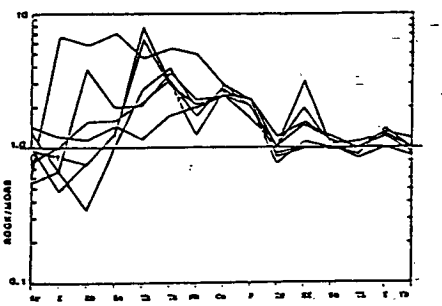
Rare earth element abundances in the Domain III greenstones from the Roxbury area normalized to chondritic values defined by Haskin and others (1968). The pattern is generally flat to slightly LREE depleted, similar to N-type MORB.



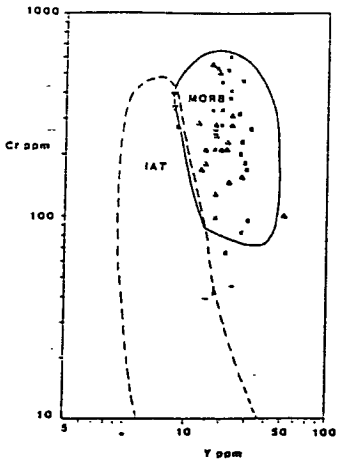
Trace and minor element pattern ("spidergram") in Domain I greenstones from the Roxbury area normalized to a typical tholeiitic (N-type) MORB. Normalization values are from Pearce (1983) and are given in the text. The elements Sr, K, Rb, and Ba are mobile during metamorphism and show an erratic pattern. Elements to the right of Ba are similar to N-type MORB.



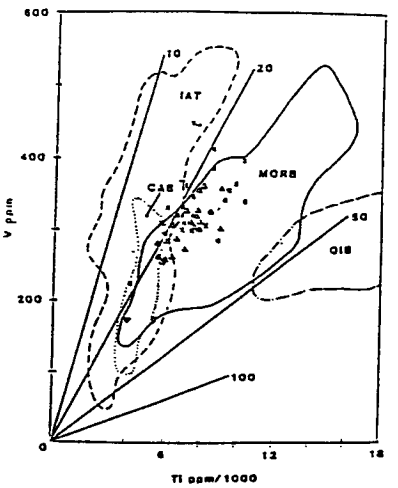
Trace and minor element pattern ("spidergram") in Domain II greenstones from the Roxbury area normalized to a typical tholeiitic (N-type) MORB. Sr, K, Rb, and Ba show an erratic pattern and elements to the right of Ba are similar to N-type MORB.



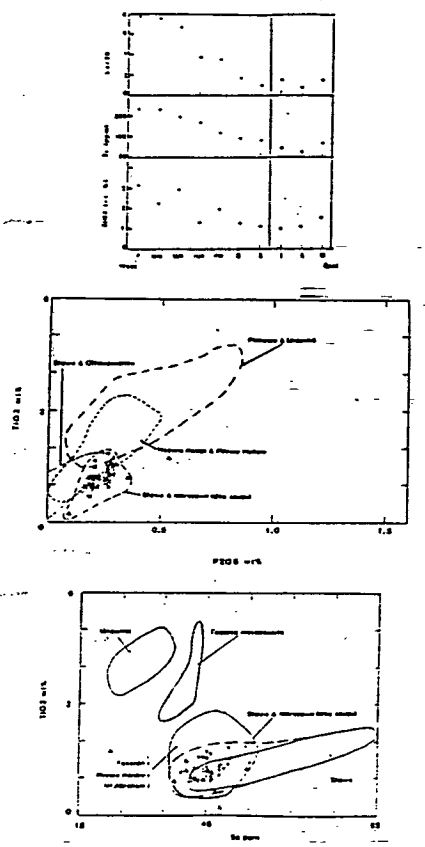
Trace and minor element pattern ("spidergram") in Domain III greenstones from the Roxbury area normalized to a typical tholeiitic (N-type) MORB. Sr, K, Rb, and Ba show an erratic pattern and elements to the right of Ba are similar to N-type MORB.



Discrimination diagram using Cr and Y from Pearce (1984). Abbreviations for fields: IAT=island arc tholeiites; MORB=mid-ocean ridge basalts. This is probably the best diagram for the separation of IAT and MORB since the area of overlap is small. The Y content of IAT is less than the Y content of MORBs. Nearly all greenstones from Roxbury plot in the MORB field. Interpretation of the fields in which metagabbro rocks and Paleozoic dikes plot is more equivocal.

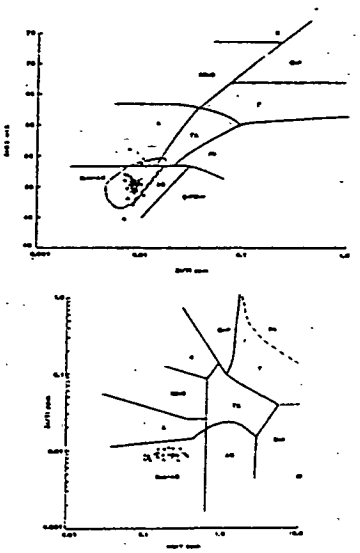


Discrimination diagram using Ti and V from Shervais (1982). Abbreviations for fields: CAB=calc-alkaline basalts; IAT=island arc tholeiites; MORB=mid-ocean ridge basalts; OIB=ocean island basalts. This is a good diagram to separate MORB (Ti/V ~20-40) and IAT (Ti/V ~<20). Even though there is an overlap between fields, the trend of Ti/V ratios in a set of samples is diagnostic. The Roxbury greenstones plot in the MORB and IAT overlap area but the trend of Ti/V is more similar to MORB. Metagabbro and dike samples are more difficult to interpret because of their small quantity.



Comparative diagrams of metabasaltic rocks from central Vermont.
A) Domains I, II, and III show similar geochemical trends to those of the Ottauquechee and Stowe Formations (Coish, this volume).
B) Domains I, II, and III all plot near the Stowe and Ottauquechee field of Coish (this volume).
C) Domains I, II, and III plot in an expanded field within the Stowe of Walsh and Kimball (this volume).

X - Domain I
Δ - Domain II
□ - Domain III
∇ - metagabbro
— - metadiabase dike
| - metabasaltic dike



Discrimination diagrams using SiO_2 and Zr/Ti, and Zr/Ti and Nb/Y from Floyd and Winchester (1978). Abbreviations for fields: AB=alkali basalts, hawaiites, mugearites, trachybasalts; sub-AB=sub-alkaline basalts (tholeiitic and high alumina); B+TB=+basanites, trachybasanites, nephelinites; A=andesites; D=RD=dacites and rhyodacites; R=rhyolites; TA=trachyandesites; T=trachytes; Ph=phonolites; C+P=comendites and pantellerites. Dashed field is that of modern ocean floor metabasalts, in which the majority of metabasic rocks from Roxbury plot. In 1.7b all of the metabasalts from the Roxbury area plot in the sub-AB field.

TECTONIC EVOLUTION OF THE SOUTHERN LINCOLN MASSIF, CENTRAL VERMONT: A PRELIMINARY MODEL FOR THE ROLE OF "RIGID" BASEMENT IN TACONIAN OROGENESIS

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Middle Proterozoic gneisses of the Lincoln massif and their overlying rift-clastic cover are exposed within two doubly-plunging anticlines which display markedly different structural styles. The eastern Lincoln massif (ELM) experienced a greater amount of strain at more ductile conditions during Taconian deformation. The western Lincoln massif (WLM) is characterized by folding and by semi-ductile to brittle faults; the eastern "massif" is reinterpreted as a ductile shear zone, characterized by an anastomosing system of mylonitic ductile faults which imbricate basement and immediately overlying wackes and boulder conglomerates. Imbrication has occurred on a scale of meters to a few hundreds of meters across a zone one to two kilometers wide.

The ductile faults of the eastern "massif" are interpreted to have developed initially as ductile shears during early development of an anticline. Along the eastern limb of this structure, simple shear resulting from concentration of strain along the steeply-dipping basement-cover contact caused detachment of isoclinally folded cover from more rigid basement. In contrast, the core and hinge region of the ELM were characterized by pure shear; flattening enhanced the entire structure vertically with continued compression. The entire deformed eastern "massif" and its eastern cover, of biotite to garnet grade, was subsequently faulted along its overturned limb against the less ductiley deformed, primarily chlorite grade syncline of cover rocks which separates the two anticlines. Some of the ductile faults to the east were progressively overprinted by more brittle fabrics, as the imbricated eastern "massif" was brought to higher crustal levels. Continued east-over-west deformation produced similar less ductile structures progressively to the west within the syncline and WLM, while out-of-sequence thrusting proceeded to the east of the ELM within the less competent eastern cover. Acadian deformation must be considered as a possible influence on these latest structures, although structural data suggest that the observed deformation occurred under a single regional stress field.

A retrodeformed cross-section suggests an overall tectonic shortening across the massif of 58%, or about 21 km. The marked similarities, however, between the eastern and western rift-clastic cover sequences, coupled with structural evidence, argue that the strain represents primarily flattening and internal imbrication of a single lithologic package, rather than juxtaposition of two very discrete packages.

The coarsest rift-clastic material consistently overlies basement within the large-scale fault zones, suggesting that the development of the Taconian structures may be largely controlled by earlier rift-related structures. The retrodeformed section supports other sedimentological evidence for an inherited change in crustal geometry across the eastern edge of the ELM, possibly representing a transition to much more thinned continental crust. This boundary subsequently provided a rigid "buttress" to westward-verging Taconian deformation that caused concentrated strain to the east, and "protected" the rocks to the west, thus producing the observed contrast in structural style.

TECTONIC EVOLUTION OF THE SOUTHERN LINCOLN MASSIF, CENTRAL VERMONT: A PRELIMINARY MODEL FOR THE ROLE OF "RIGID" BASEMENT IN TACONIAN OROGENESIS

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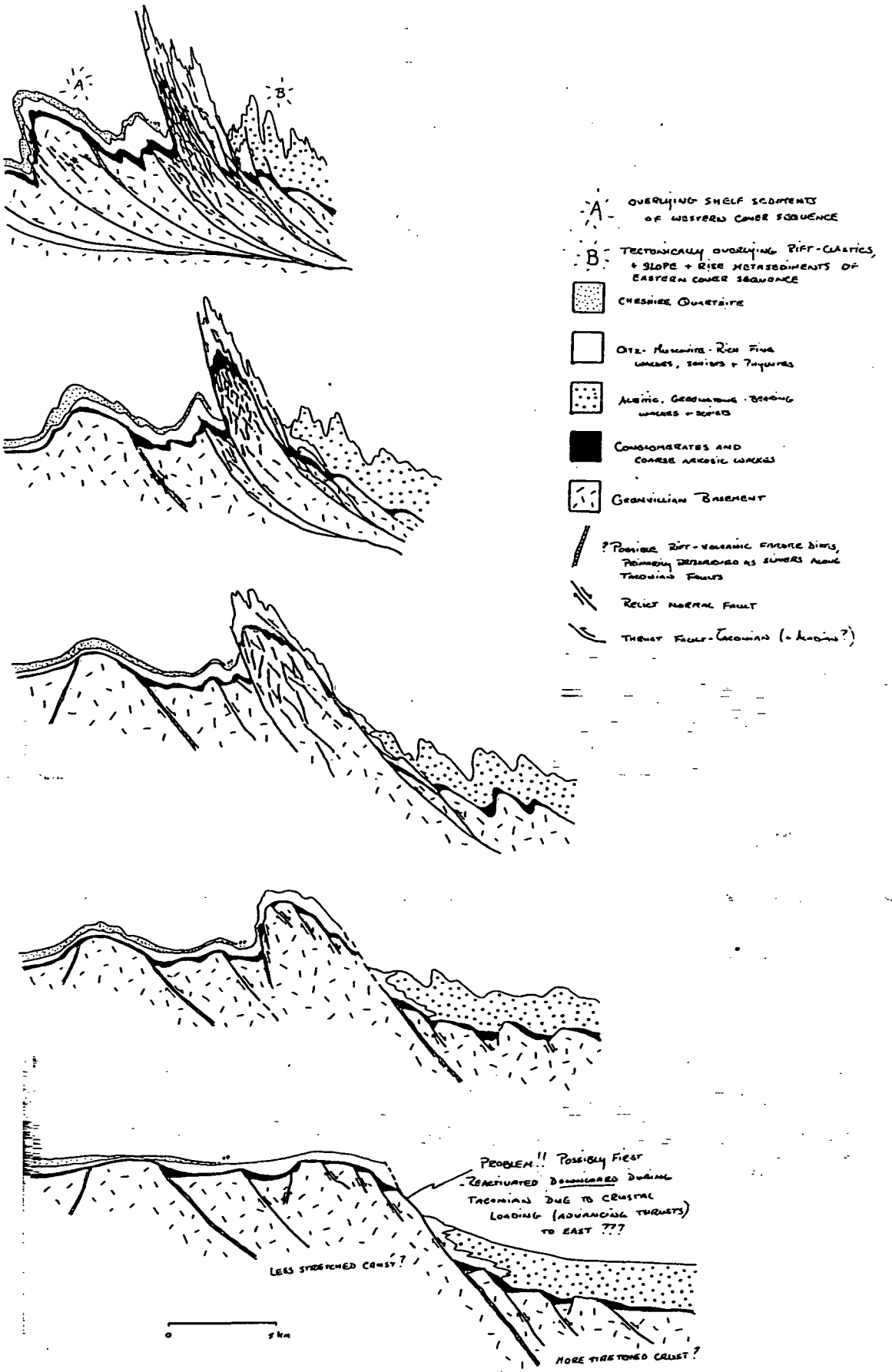

THE BRONSON HILL VOLCANIC ARC AND ITS COLLISON WITH NORTH AMERICA: THE RADICAL NEW INTERPRETATION IS WRONG, AN ENIGMA RETURNS

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The radical new interpretation proposed in our abstract submitted in fall 1988 is no longer tenable based on new zircon data produced during the period December 1988 - February 1989. Critical to this new position are the following: 1) Data on new zircon separates from Partridge rhyolite tuff (previous abstract) indicating a refined age of 449 +3/-2 (late Caradocian) as contrasted with the previously reported age of 432 +16/-5 based on three discordant analyses; 2) An age of 453 ±2 on separates from quartz-phyric rhyolite in the Upper Member of the Ammonoosuc Volcanics collected in November 1988 from a new artificial exposure.

These results show that there is almost complete temporal overlap between the Ammonoosuc-Partridge cover sequence and the physically underlying plutonic gneisses of the Monson, Fourmile, Swanzey, and Pauchaug. These results rule out the unconformity hypothesis, previously espoused by us, in which the Ammonoosuc-Partridge was unconformably above the plagioclase gneisses. They also appear to rule out the hypothesis proposed by others that the plagioclase gneisses are intrusive into the Ammonoosuc, unless the Lower Member is substantially older than the Upper Member dated here.

The Monson, Fourmile and other plagioclase gneisses of the Bronson Hill zone appear to represent the plutonic roots of a calc-alkaline volcanic suite that was formed in the Bronson Hill arc between approximately 454 - 443 m.y. The Upper Member of the Ammonoosuc Volcanics and the Partridge Formation belong to a cover sequence that spanned from at least 453 to 449 m. y. Despite their apparent contemporaneity, the igneous rocks in the two sequences have very little geochemical similarity. What then was the process by which they were brought to their present close juxtapositon, probably before erosion and deposition of the Early Silurian Clough Quartzite?

In studies of modern island arcs, it has commonly been suggested that the arc lithosphere is undergoing extension. Extensional detachment faults are features that are commonly ascribed to such environments and that are also a plausible geometrical way to bring deep-seated plutonic rocks into contact with volcanic-sedimentary cover of the same age. While we have not yet worked out the necessary geometry and geometrical evidence that such a fault was operative in the Bronson Hill arc during the late Ordovician, this hypothesis seems worthy of further exploration. In its favor is the striking localization of a zone of pre-metamorphic sulfidic hydrothermal alteration in Monson Gneiss of the Orange area. This zone, 10 km long and up to 1 km wide lies adjacent to the basal contact of the Ammonoosuc Volcanics.

The new zircon ages confirm that the Ammonoosuc Volcanics and Partridge Formation were deposited at essentially the same time as the Caradocian emplacement of the Giddings Brook thrust sheet in westernmost New England and eastern New York. This suggests that tectonic features known to post-date the Giddings Brook slice may have formed in the late Ordovician, which may be the true time of final closure of Iapetus Ocean in New England. This can be equated with the notion that the early Silurian extension of the Merrimack trough might have been related to late phases of back-arc extension east of the Bronson Hill arc.

A second enigmatic contact is between the plagioclase gneisses (Fourmile) of the Pelham dome and the underlying the late Proterozoic strata with their possible Avalon affinities and traces of granulite facies metamorphism. Geologic mapping has shown that this contact was already established and recumbently folded during early phases of the Acadian orogeny. If the Bronson Hill volcanic arc was built on a fragment or margin of Avalon, then an Ordovician closure of Iapetus as suggested here is inconsistent with paleomagnetic data that shows Avalon in high southern latitudes in middle Ordovician time. An alternative currently being explored is that this contact and the apparent metamorphic discontinuity is a major fault contact of early Acadian age.

THE BRONSON HILL VOLCANIC ARC: OLD INTERPRETATIONS AND NEW RADIOMETRIC AGES

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To the student of the Taconian orogeny in western New England, the Bronson Hill volcanic arc was the Never-neverland in the misty east related to an east-dipping subduction zone believed to have existed through early and medial Ordovician time. To the student of the Bronson Hill rocks themselves this relationship was far less obvious; and pre-Silurian stratigraphy and plutonism have been interpreted in two distinctly different ways. The sequence of rock units in one interpretation is: 1) Late Proterozoic microcline gneisses, schists and quartzites, 2) Late Proterozoic through early Ordovician plagioclase gneisses and amphibolites of plutonic and volcanic aspect (Monson, Fourmile, Swanzey, Pauchaug) unconformably overlain by 3) medial Ordovician Ammonoosuc Volcanics of probable arc affinities and Partridge Formation black shales with volcanics, also of probable island arc affinities, in turn overlain unconformably by 4) Lower Silurian (Llandoverian, 438-428 m.y.) Clough Quartzite. In the second interpretation, part of the plagioclase gneisses of 2) are considered to be intrusive into the Ammonoosuc and younger than medial Ordovician. Common to both interpretations has been a correlation of the Partridge to a medial Ordovician (Caradocian; 458-448 m.y.) graptolite locality in the Boundary Mountain anticlinorium, northwestern Maine; and an unconformity at the base of the Clough Quartzite.

New radiometric dating of carefully selected zircon separates from central Massachusetts and southwestern New Hampshire gives the following results:

Microcline gneiss of the Dry Hill Gneiss in the Pelham dome, interpreted as metamorphosed alkali rhyolite, is confirmed as late Proterozoic, yielding an age of 614 +9/-7. Another sample shows strong evidence of inheritance with a minimum mean age of 1400.

Plagioclase gneisses (Monson, Fourmile, Swanzey, Pauchaug), now believed to represent largely intrusive igneous rocks, yield ages of 454 +4/-2, 454 +9/-3, 452 +3/-2, 447 +3/-2, close to 443 and close to 443, thus ranging from medial Ordovician through late Ordovician, consistent with other dates on Oliverian granitoid rocks in the region. A gneissic gabbro within the Swanzey Gneiss yields 454 +3/-2. No sample of plagioclase gneiss yet obtained in this study has yielded a Late Precambrian or Cambrian age like those obtained from lithically similar gneisses in coastal Connecticut.

A quartz-phyric rhyolite in the Upper Member of the Ammonoosuc Volcanics, about 275 meters above the top of the Lower Member and 115 meters below the base of the Partridge Formation, yields an age of 453 ±2. A pyroclastic quartz-phyric rhyolite bed with clear primary textures from the Partridge, 20 meters above the top of the Ammonoosuc Volcanics, yields a zircon age of 449 +3/-2. The Ammonoosuc and Partridge localities are 3 km from the Bernardston fossil locality where the Clough contains Silurian fossils and the Fitch Formation contains lowest Devonian conodonts.