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**Notes**

# Tectonic synthesis of the Taconian orogeny in western New England

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## ABSTRACT

A tectonic synthesis based on stratigraphic and structural analysis of western New England is proposed for the Ordovician Taconian orogeny. It emphasizes arc-continental collision in which ocean-floor, continental-margin, and ensialic-rift rocks were imbricated westward in a repeatedly deformed accretionary wedge. Continued compression displaced segments of the North American sialic crust to the west and deformed the earlier emplaced slices of the Taconic allochthons which were derived from the continental margin.

Critical arguments for this synthesis are (1) the west-to-east stratigraphic relations among the basal rift clastic rocks of the Dalton, Pinnacle, and Hoosac Formations of late Precambrian to Early Cambrian age; (2) the stratigraphic and sedimentological similarities between the rocks of the lower Taconic sequence and rocks in the Pinney Hollow and Underhill slices to the east and north of the Green Mountain massif; (3) the environmental similarities between the Cambrian and Lower Ordovician section of the Giddings Brook slice and the age-equivalent section in the St. Albans synclinorium; (4) the presence of carbonate platform rocks as slivers between each of the successively higher and younger premetamorphic slices (groups 1 and 2) of the Taconic allochthons; (5) the presence of synmetamorphic, fault-related structures in the youngest and highest slices (group 3) of the Taconic allochthons; (6) the recognition of extensive thrust zones in the pre-Silurian eugeoclinal sequence east of the middle Proterozoic basement of the Housatonic, Berkshire, Green Mountain, and Lincoln massifs; (7) the location of the Taconic root zone within the pre-Silurian eugeoclinal sequence; (8) the recognition of numerous faults in the serpentinite belt; (9) the similarity between the rocks of the Moretown Formation and modern fore-arc basin sequences; (10) the recognition that the volcanic arc-continental complexes of the Ascot-Weedon and Bronson Hill have been displaced westward over the Moretown and/or Hawley Formations along such faults as the Bristol and Coburn Hill thrusts; (11) the allochthonous and internally imbricated nature of the North American basement in the Berkshire massif; (12) the proposition that the Housatonic, Green Mountain, and Lincoln massifs, as well as the middle Proterozoic cored domes of southeastern Vermont, are also thick sialic slices of North American basement; (13) the recognition of medium-high- to high-pressure metamorphic mineral assemblages in the pre-Silurian eugeoclinal rocks of Vermont; and (14) the recent synthesis of isotopic age data by Sutter and others (1985).

On the basis of an analysis of the foregoing arguments and relationships, a chronological sequence of seven structural sections between Albany, New York, and the Bronson Hill anticlinorium in central Massachusetts is used to depict the evolution of the Taconian orogeny. Retrodeformed distances are based on structural overlap and restoration of the Taconic slices to their depositional setting along the ancient North American continental margin. These easterly younging, diverticulated slices formed as a result of horizontal compression rather than gravity sliding. This palinspastic analysis implies the following. (1) Approximately 1,000 km of shortening has occurred during the emplacement of the Taconic allochthons and the subsequent imbrication of North American basement as thick sialic slices. Approximately 330 km of this shortening is attributed to multiple cleavage generations. (2) Repeated movement along such major surfaces as the Cameron's Line-Whitcomb Summit-Belvidere Mountain thrust zone has buried the Taconic root zone. We suggest that the northern extension of this root zone is exposed to the east of the Lincoln massif in Vermont where the Underhill, Pinney Hollow, and Hazens Notch Formations are exposed. These formations, here considered thrust slices, disappear along the Belvidere Mountain-Whitcomb Summit thrust zone as it is traced southward into western Massachusetts and western Connecticut. (3) Taconian metamorphic rocks, particularly the older medium-high-pressure rocks in northern Vermont, have been transported westward on such reactivated surfaces as the Belvidere Mountain thrust. (4) The anticlinorial form of the middle Proterozoic basement in the Green Mountain and Lincoln massifs may have resulted from fault-bend folding on deep mantle-involved thrusts that developed late in the Taconian orogeny.

## INTRODUCTION

Since 1945, the Taconic allochthons of western New England have been interpreted as gravity slides that originated from a depositional site on top of the present location of the Green Mountain and Berkshire massifs and that were emplaced into a deep basin to the west (Cady, 1945; Thompson, *in* Billings and others, 1952; Zen, 1961, 1967, 1972; Rodgers and Neal, 1963). This interpretation was based on the presence of wildflysch deposits and soft-sediment deformation along and beneath the westernmost slice (Giddings Brook) and the recognition that the youngest rocks in the allochthon are coeval with sediments beneath the thrusts. In this model, a root zone is not required. Furthermore, this model was consistent with the concept of a coherent depositional stratigraphy in the pre-Silurian

Plates 1 and 2, which include Figures 2 and 6, appear as folded inserts in this issue.

eugeoclinal belt east of the Green Mountain–Lincoln massifs. The gravity-slide hypothesis was proposed to explain the wildflysch deposits, the submarine emplacement, and the absence of a recognized root zone.

The gravity-slide model, however, had to be modified when Skehan (1961) and Thompson (1972) showed that the basal part of the autochthonous carbonate platform rests unconformably on middle Proterozoic rocks at the eastern edge of the Green Mountain massif, thus ruling out the Green Mountain basement as the original depositional site for the rocks of the Taconic allochthons. In addition, the palinspastic width of the Taconic allochthons far exceeds that of the Green Mountain or Berkshire basement (Ratcliffe and others, 1975). Zen (1967, p. 56–62; 1972, p. 44–45) concluded that gravity sliding from these basement areas might be a correct working model, although the surface available was minimal, if not inadequate. Studies of the Berkshire massif in Massachusetts showed that the middle Proterozoic basement consists of multiple overlapping fault slices reflecting extensive structural foreshortening (Ratcliffe, 1969a, 1975, 1984; Ratcliffe and Harwood, 1975; Norton, 1967, 1975; Harwood, 1975; summarized by Ratcliffe and Hatch, 1979). The cover rocks on the highest structural slices in the Berkshire massif are unconformable on the basement and consist of upper Proterozoic and Lower Cambrian rocks that interfinger with shelf rocks. These cover rocks are unlike the rocks in the Taconic sequence. These relationships rule out both the Green Mountain and Berkshire basements as potential depositional sites for the Taconic sedimentary basin and require a depositional site situated still farther to the east. This interpretation was proposed early in this century (Keith, 1913, 1932; Prindle and Knopf, 1932; Zen, 1967; Rodgers, 1970). Keith (1932) and Prindle and Knopf (1932) proposed that the Taconic allochthons were derived from a root zone in the pre-Silurian eastern Vermont sequence within the Hoosac-Rowe part of the section.

We believe that a root zone for the Taconic allochthons lies, in part, within the Hoosac Formation and below a prominent tectonic zone, the Whitcomb Summit thrust (WST, Fig. 2; this figure appears on a folded insert included in this issue) which carries serpentized ultramafic rocks in its upper plate and marks the locus of repeated displacement during a westward-progressing collisional episode in Middle and Late Ordovician time (Stanley and Ratcliffe, 1980, 1982, 1983). This zone is defined by the Hoosac-Rowe contact of western Massachusetts (Zen and others, 1983).

We have examined the gravity-slide hypothesis for the Taconic allochthons and conclude that hard-rock thrusting, with the deposition of forethrust olistostromes in front of an overriding accretionary wedge, better explains the available data. Rowley and Kidd (1981) also proposed an accretionary model for the development of the Taconian orogeny on the basis of an analysis of the Austin Glen graywacke, relations in Newfoundland and Quebec, and their deductive application of modern island-arc–continental boundaries in the southwestern Pacific to western New England. Earlier, Rowley and others (1979) suggested that an obducted ophiolite was subsequently carried “piggy back” by younger, more westerly situated slices of the Taconics during the evolution of the orogen. Rowley and Kidd (1981) modified this model to accommodate the emplacement of sialic slices in accordance with evidence worked out by Ratcliffe, Norton, and Harwood for the Berkshire massif (see earlier citations). Recently, Rowley (1983) and Bosworth and Rowley (1984b) recognized a thrust chronology in the northern part of the Taconic allochthons that confirms the observations of Zen (1967, 1968) and other workers and, thus, have put to rest their recent discussion on the relative ages of the frontal thrust faults and the development of slaty cleavage.

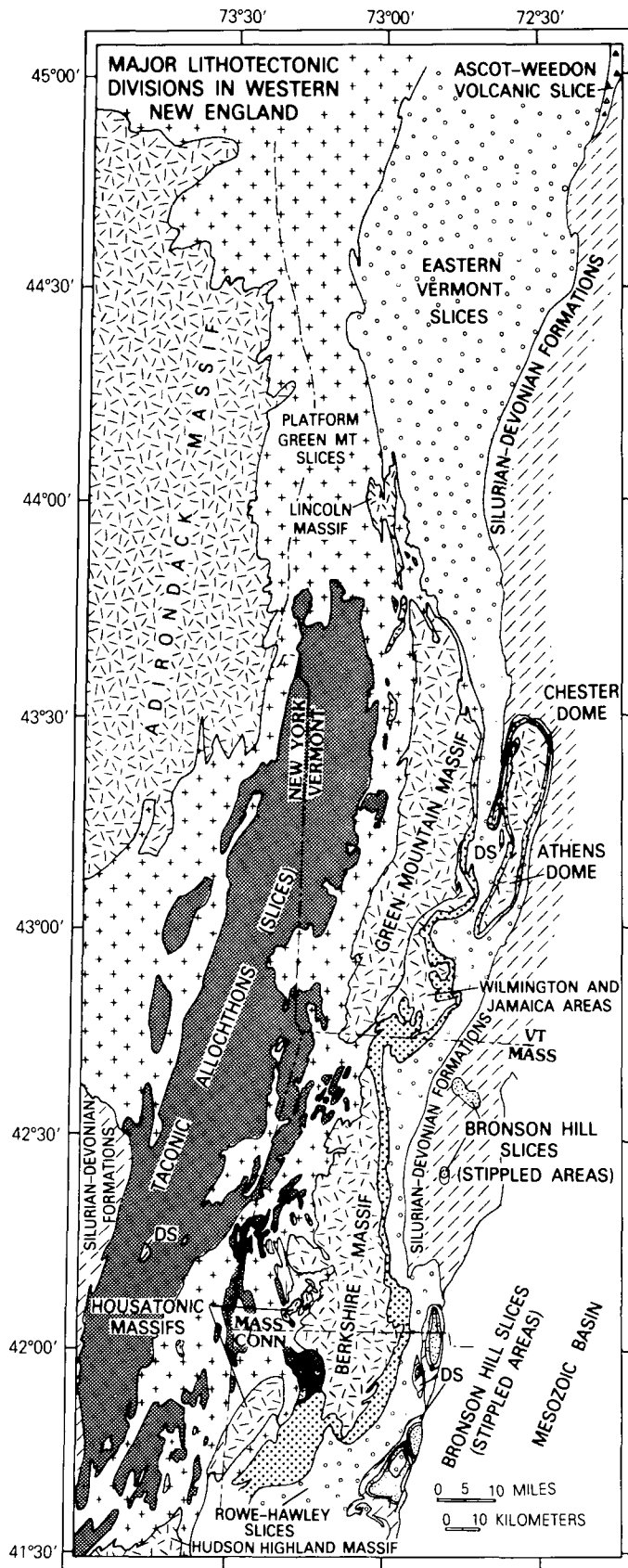
We contend that the model presented by Rowley and Kidd (1981) for the Taconian orogeny is too simple and does not adequately account

**Figure 1. Generalized geologic map of western New England north of 41°30', showing the major lithotectonic units. Unit 1: Parautochthonous middle Proterozoic Green Mountain (Y), Lincoln (YL), Housatonic (Yh), and Adirondack massifs (random dashed pattern) with their platform cover (plus symbols) labeled “Platform-Green Mountain slices.” Unit 2: Taconic allochthons. Unit 3: Allochthonous middle Proterozoic basement and late Proterozoic to Lower Cambrian cover rocks of the Berkshire massif and the domes of southeastern Vermont (Chester dome, Yc, Athens dome, Ya, and the Wilmington-Jamaica areas shown in more detail in Fig. 2). Unit 4: Allochthonous Hoosac Formation located east of the Housatonic, Berkshire, and Green Mountain massifs (dot pattern). Unit 5: The eastern Vermont eugeoclinal rocks (circle pattern) labeled “Eastern Vermont slices” to the north and “Rowe-Hawley slices” to the south. Unit 6: The Bronson Hill slices (stippled pattern) exposed in the domes west of the Mesozoic Basin. DS refers to Silurian and Devonian rocks at Becraft Mountain. Figures 1 and 2 have been modified from Doll and others (1961), Fisher and others (1971), Zen and others (1983), and Rodgers (1982a).**

for the data and arguments worked out by numerous geologists for the Taconic region and the pre-Silurian eugeoclinal belt. The Taconic allochthons, as they are preserved today, did not accrete in a simple foreland direction, but developed with the youngest slices near the hinterland, consistent with the diverticulated sequence of Zen (1967). Furthermore, with the possible exception of Belvidere Mountain in Vermont (Doolan and others, 1982), there is no evidence of an ophiolitic slab with the dimensions of the examples in Quebec and Newfoundland as shown by Rowley and others (1979) in the eugeoclinal belt of western New England, although fragments of serpentized dunite and peridotite are common. In addition, much of the reinterpreted structure in the eugeoclinal belt reported here and elsewhere (Stanley and Roy, 1982; Zen and others, 1983; Stanley and others, 1984) was either being worked out or written during 1978–1981 and was not available to Rowley and others. Finally, our plate-tectonic interpretation for western New England is inductive, although we have liberally reinterpreted relations, particularly in the pre-Silurian eugeoclinal belt in Vermont; it is based on a systematic retrodeformation of known geologic relations that begins with their present arrangement as seen in maps and cross sections. Thrust slices and allochthonous terranes are restored to their original deposition sites according to stratigraphic, metamorphic, structural, and chronological evidence or our interpretation of that evidence. As a result, many of our arguments may appear to be too detailed for those readers not familiar with New England geology, but we believe that this discussion is necessary for our synthesis. The arguments which we present here are based largely on our detailed mapping in the central Taconics region, the Berkshire massif, the eastern Vermont sequence from western Connecticut to northern Vermont, and in Taiwan between the west Philippine and Asian plates.

## MAJOR LITHOTECTONIC UNITS

The western part of central and northern New England can be divided into six major lithotectonic units that are marked by distinctive rock sequences (Fig. 1). Although the boundaries between many of the units are



shown as faults, this interpretation is largely speculative, except for the Taconic allochthons (unit 2; Fig. 1). Data bearing on these faults are presently available only in western Massachusetts (Zen and others, 1983), along the eastern border of the Green Mountain massif (Karabinos, 1984; Karabinos and Thompson, 1984), north and east of the Lincoln massif in central Vermont (Tauvers, 1982b, 1982c; DiPietro, 1983a, 1983b; Stanley and others, 1985), and the Jay area in northern Vermont (Stanley and others, 1984). In Vermont, the thrust slices and most of the basal thrust faults are named for the dominant formation as mapped by Doll and others (1961). For example, the Underhill slice is made up largely of the Underhill Formation and is floored by the Underhill thrust. A letter-number coordinate system in Figure 2 is used to locate areas referred to in the text. These locations are shown in parentheses without reference to Figure 2; for example, 3g, J4 refers to the Greylock slice in grid location J4.

#### Parautochthonous Green Mountain Massif (Yg), Lincoln Massif (YL), and Cover Unit 1 (Fig. 1)

Middle Proterozoic gneiss of the Green Mountain and Lincoln massifs forms a parautochthonous basement upon which three successive sequences were deposited: (1) an upper Proterozoic rift-clastic sequence, (2) a Lower Cambrian through Lower Ordovician carbonate-siliciclastic shelf sequence, and (3) a Middle Ordovician limestone-shale, back-arc basin sequence. Major unconformities separate sequences 1 and 3 from the underlying rocks. Basement rocks of the Housatonic massif (Yh) and the Hudson Highlands in New York and Connecticut are overlain by cover sequences 2 and 3 above. These basement blocks, along with the Cambrian and Ordovician cover, are correlative with the autochthonous Adirondack basement and its surrounding cover.

#### Taconic Allochthons (Unit 2; Fig. 1)

We have adapted the slice concept of Zen (1967) for the Taconic allochthons but modified it by revisions based on recent mapping. We have further combined his slices into three groups. Group 1 slices, which were emplaced on soft, unconsolidated rocks, as evidenced by wildflysch deposits, include the Giddings Brook-Sunset Lake slice [shown as one slice (1gb) in Fig. 2] and several areas of predominantly Austin Glen Graywacke directly west of the Giddings Brook slice, which Fisher (1977) interpreted as soft-rock slide masses (Fig. 2, 1Ls). Group 1 slices have the most complete stratigraphic range of any of the groups. They consist of upper Precambrian to Middle Ordovician rocks that presently occupy the greatest area. Group 2 slices consist of the Chatham (2c), Rensselaer Plateau (2rp), Bird Mountain (2bm), Berlin Mountain (2b), and Everett (2e) slices of Figure 2. These slices contain mainly the upper Precambrian through Cambrian part of the Taconic section, with distinctive basaltic volcanics and associated graywacke. They are in fault contact with group 1 slices. The fault zones commonly contain slivers of the carbonate platform in complex tectonic breccias (Fig. 2, symbol X). Group 1 and group 2 slices were emplaced before metamorphism.

Group 3 slices consist of the Dorset Mountain (3d), the Greylock (3g), the Canaan Mountain (3cm), and the June Mountain (3j) slices of Figure 2. The upper Precambrian through Lower Cambrian rocks within group 3 slices resemble the rocks in the Hoosac-Cavendish sequence that rest unconformably on 1-b.y.-old basement of lithotectonic units 3 and 4 (Fig. 1). Synmetamorphic reclined folds and mineral lineations mark the fault zones. They result from westward displacement of the slices, their final emplacement postdating earlier metamorphism in the autochthon.



### **Allochthonous Berkshire Massif (Yb) and Middle Proterozoic Domes (Yc, Ya, Yj, Yw) of Southeastern Vermont (Unit 3; Fig. 1)**

The allochthonous middle Proterozoic basement of the Berkshire massif is separated from its adjoining lithotectonic units 1 and 4 by major thrust faults: the Hoosic thrust (HoT) to the west and the Hoosac Summit thrust (HST)–Middlefield fault zone (MFZ) to the east. The massif is internally imbricated by ten overlapping thrust slices, locally separated by slivers of cover rocks. The western half of the massif is unconformably covered by basal clastic rocks which locally grade upward into shelf carbonate rocks. The uppermost and easternmost slice in the massif on Hoosac Mountain is unconformably overlain by upper Precambrian or earliest Cambrian rocks that interfinger with the clastic rocks of the western shelf sequence and are considered to be a transitional facies with the rocks of the allochthonous Hoosac Formation to the east. Basement gneiss in the Wilmington area (Yw, I5), the Jamaica area (Yj, H6), and in the Chester-Athens domes (Yc-Ya, G6, H6) contains a cover similar to that found on the highest basement slice in the Berkshire massif. These basement areas in southeastern Vermont are here considered to be allochthonous and part of the Berkshire massif tectonic unit.

### **Allochthonous Hoosac (Unit 4; Fig. 1)**

Lithotectonic unit 4 consists of a narrow belt of Hoosac Formation at the eastern margin of the Berkshire massif that extends northward as far as Middletown, Vermont (8, G6), and widens southward into Connecticut, where it merges with the Manhattan terrane (Hall, 1980) south of the area shown in Figure 2. The sole of this unit is a major thrust, the Middlefield fault zone (Fig. 2, MFZ), that Norton (1976) mapped along the eastern margin of the Berkshire massif. Near North Adams (location 1, J4), the fault leaves the Proterozoic Y-Hoosac contact and cuts upward into the Hoosac Formation as the Hoosac Summit thrust (Fig. 2, HST) (Ratcliffe, 1979a). Along its entire length, the Hoosac Summit–Middlefield fault zone exhibits mylonitic fabrics and is locally intruded by granite in thin, sill-like sheets (Ratcliffe, 1975; Ratcliffe and Harwood, 1975). We believe that this zone represents the root zone for the structurally youngest slices (group 3) of the Taconic allochthons.

### **Eastern Vermont and Rowe-Hawley Slices (Unit 5; Fig. 1)**

This complex is separated from underlying tectonic units to the west by a zone of thrust faults associated with overlapping thrust slices that bring westward more easterly situated rocks as the belt is traced southward from the Canadian border (Figs. 1 and 2). It is a cryptic suture for the Taconic allochthons, with as much as 660 km of displacement in western Massachusetts. In northern Vermont, the Hinesburg thrust (HT, B4, C4) has been shown to be a recumbent fold, the lower limb of which was sheared out during emplacement (Dorsey and others, 1983). This fold-thrust system places phyllite of the Fairfield Pond Formation, as used by Tauvers (1982c) and DiPietro (1983b), and shallow-water argillaceous quartzite of the Cheshire Quartzite on top of the carbonate platform-bank edge and slope sequence of the Hinesburg and St. Albans synclinorium (Gillespie, 1975; Dorsey and others, 1983). North of Burlington, Vermont, the lower part of the carbonate platform is in depositional contact in several places with the underlying argillaceous rocks of the Hinesburg slice (OCp on HS, B4) (Dorsey and others, 1983). These remnants are important in locating the northern root zone for the Taconic slices. Total displacement across the Hinesburg thrust zone at this latitude is estimated to be 10 km.

The Hinesburg thrust extends southward in the Lincoln area where Tauvers (1982c) and DiPietro (1983b) showed that it dies out in the overturned limb of the fold cored by the eastern part of the Lincoln massif

(YL, D5). More importantly, these geologists demonstrated that the eastern contact along the northern part of the massif is tectonic, with major thrust faults containing mylonitic zones extending northward along the western boundary of the Underhill Formation (Fig. 2, Jerusalem and Underhill slices). The location of these thrust faults to the north is speculative. We consider these faults to be part of the northern root zone for the Taconic slices and, thus, the major zones of movement compared to the Hinesburg thrust.

To the south, where the Underhill pinches out between the Hoosac (Fig. 2, CZh–CZt) and the Pinney Hollow Formations (Fig. 2, PHS, CZph) at the latitude of the north end of the Green Mountain massif (F5), the western margin of the eastern Vermont slices traces southward along the western contact of the Pinney Hollow, as mapped by Doll and others (1961), to the Massachusetts border. Here, it joins the Whitcomb Summit thrust (Fig. 2, WST), which separates the allochthonous Hoosac Formation (unit 4; Fig. 1) from the Rowe Schist to the east. The Pinney Hollow–Underhill contact in Vermont is interpreted as being a thrust fault on the basis of regional relations, although there is no direct local evidence to support this interpretation. The Whitcomb Summit thrust in Massachusetts, which is based on field evidence (Stanley, 1978; Ratcliffe, 1979a; Stanley and others, 1984), continues southward into Cameron's Line, a major tectonic contact in western Connecticut (Rodgers and others, 1959; Rodgers, 1970, p. 94, 1982a; Hatch and Stanley, 1973, p. 34; Hall, 1980).

### **Core Rocks of the Domes of Western Massachusetts and Connecticut (Unit 6; Fig. 1)**

The eastern edge of the Eastern Vermont and Rowe-Hawley slices is shown as the basal contact between the Cambrian and Ordovician rocks and the Silurian and Devonian cover. Lithotectonic unit 6 is represented by the core gneiss of the Bristol, Collinsville, Granville, Goshen, and Shelburne Falls domes (Figs. 1 and 2) and is interpreted as the thin western edge of the eastern volcanic arc–continent complex (Bronson Hill plate of Robinson and Hall, 1980), presently exposed along the Bronson Hill anticlinorium in central New England. The lower contact is not exposed in Vermont or Massachusetts. In the Bristol (BD, N5) and Waterbury (WD, N4) domes in western Connecticut, however, it is mapped as a discordant contact between the Bristol Member of the Collinsville Formation and the Taine Mountain Formation (Stanley, 1964, Pl. 1; Simpson, 1974). This contact is here interpreted as being a thrust fault and is named the "Bristol thrust."

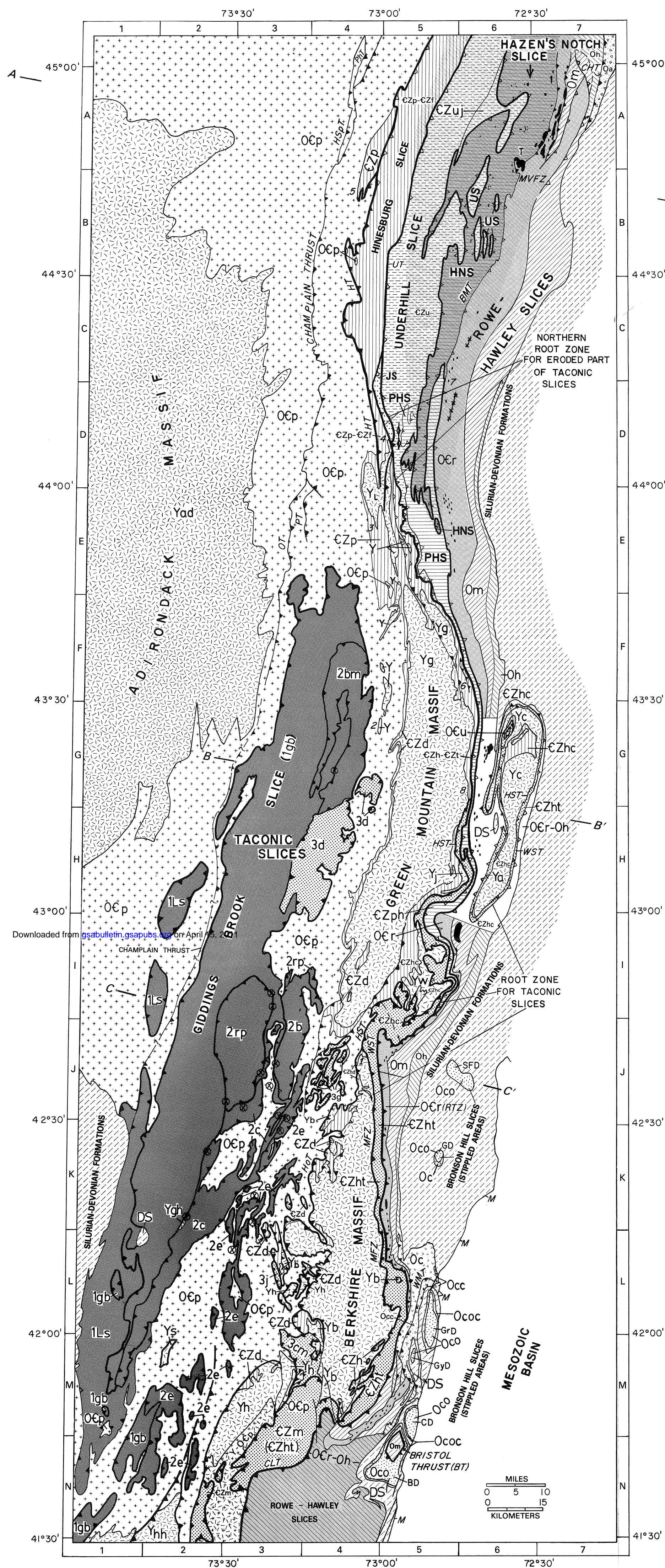
Although the Ascot-Weedon sequence in northern Vermont and southern Quebec (Oa, A7) has not been treated separately in our discussion, it does contain plutonic and volcanic rocks suggestive of an island arc (Gale, 1980; Hoar, 1981; Doolan and others, 1982). This sequence is located west of the Bronson Hill anticlinorium and disappears beneath the Silurian-Devonian unconformity ~30 km south of the Canadian border. Although the relation of the Ascot-Weedon sequence to the rocks of the Bronson Hill anticlinorium is unknown at present, it may have been overridden by the Bronson Hill arc complex with the triple junction now buried beneath the Silurian and Devonian rocks to the east, as suggested by Osberg (1978), for example. Gale (1980) and Hoar (1981) mapped the Coburn Hill thrust (Fig. 2, CHT) along the western border of the Ascot-Weedon sequence in northern Vermont.

### **IMPORTANT STRATIGRAPHIC RELATIONSHIPS AMONG LITHOTECTONIC UNITS**

Palinspastic reconstruction of the western margin of Iapetus and its offshore tectonic elements requires correct interpretation of not only the location of major boundaries but also the relative sedimentologic position



Figure 2—INTERPRETATIVE TECTONIC MAP OF WESTERN NEW ENGLAND



## PLATE 1—TECTONIC SYNTHESIS OF THE TACONIAN OROGENY IN WESTERN NEW ENGLAND

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Figure 2A— INTERPRETATIVE GEOLOGIC SECTION—NORTHERN VERMONT

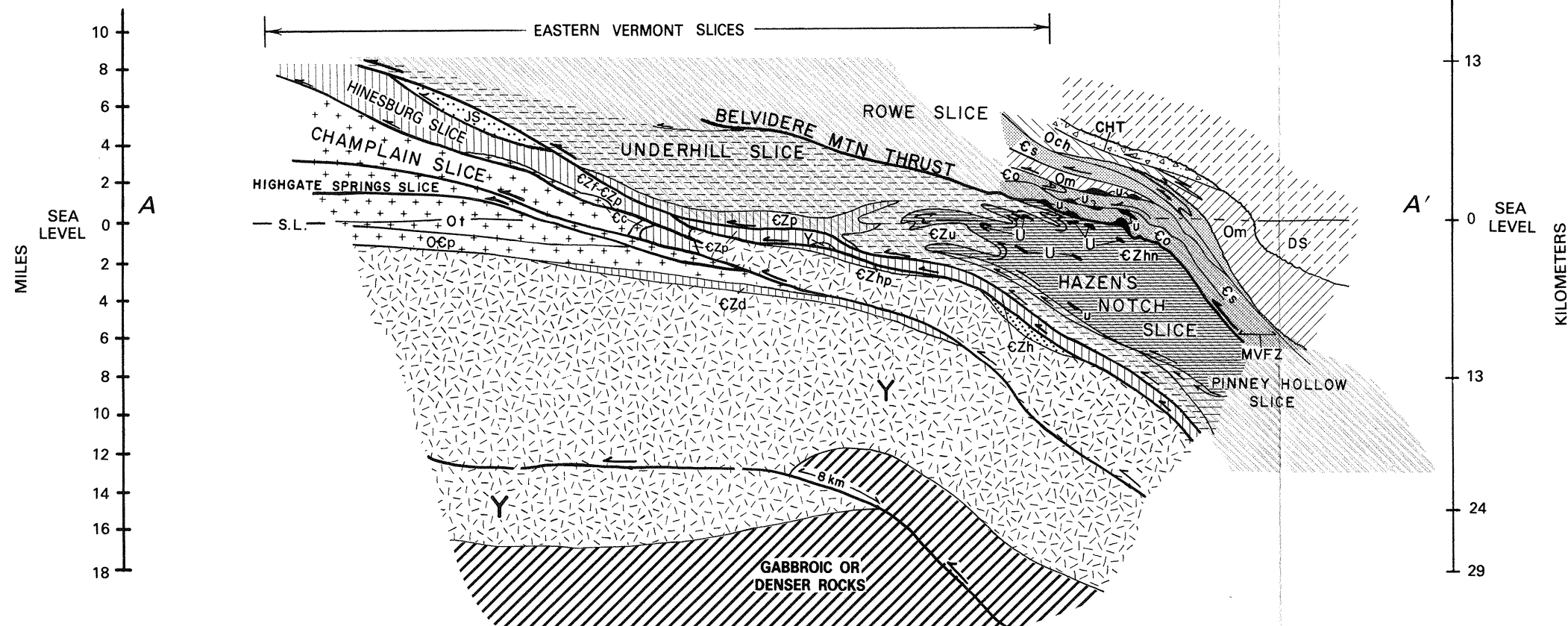


Figure 2B—INTERPRETATIVE GEOLOGIC SECTION—SOUTHERN VERMONT

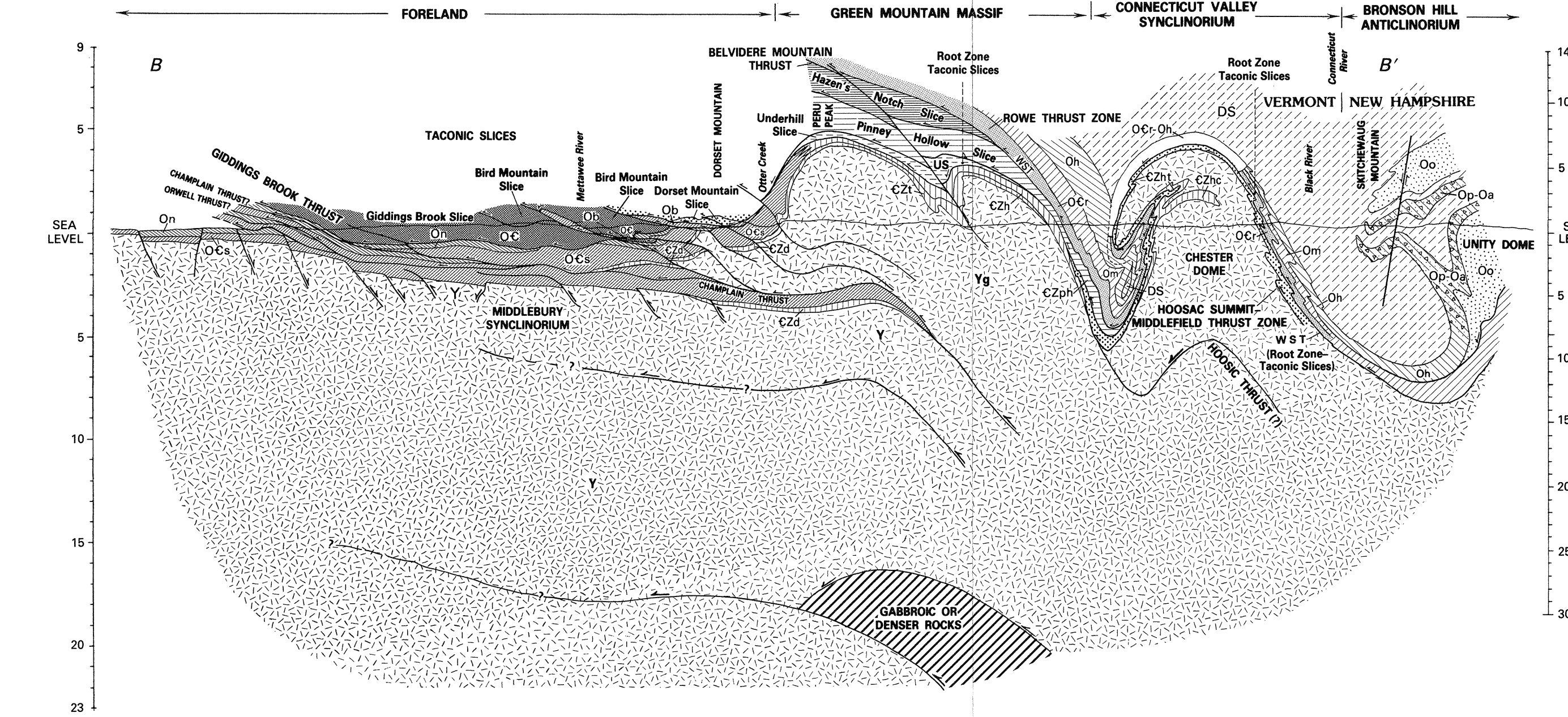
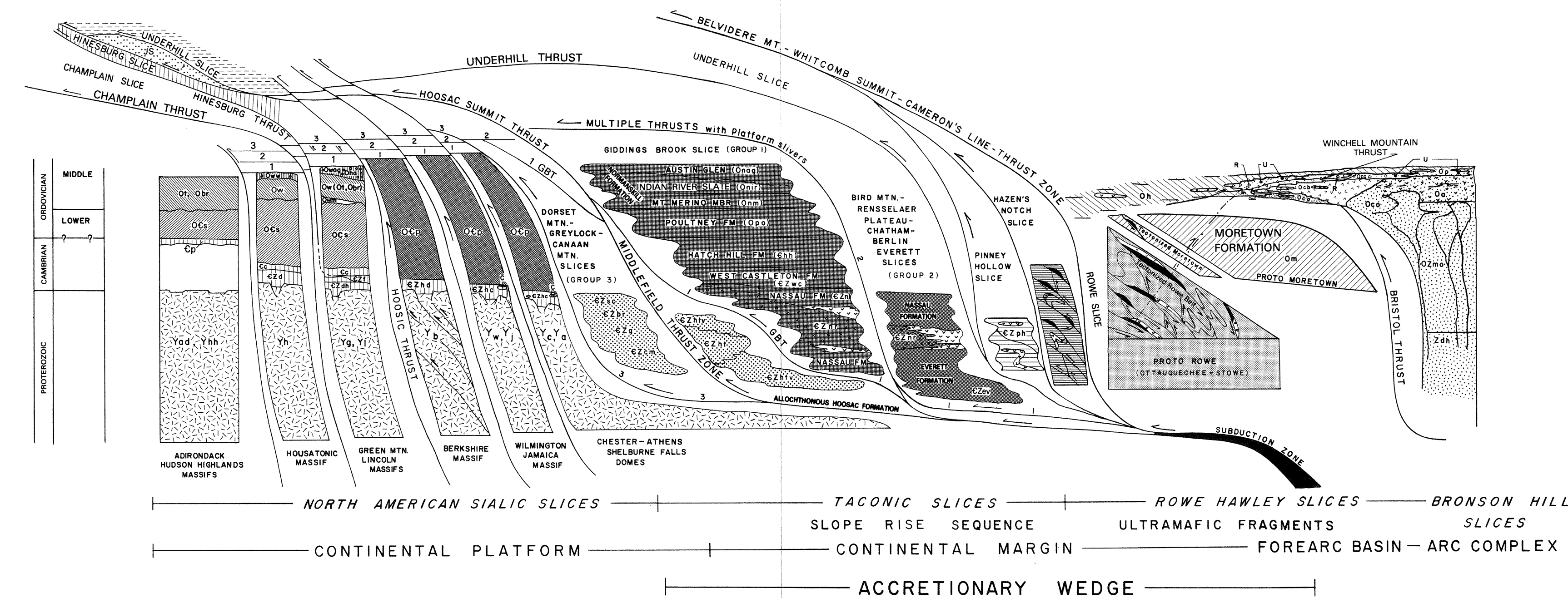


Figure 6— INTERPRETATIVE PALINSPASTIC DIAGRAM OF WESTERN NEW ENGLAND



**Figure 2**  
Interpretative Tectonic Map of Western New England north of latitude 41°30' showing in more detail the distribution and internal character of the lithotectonic units of figure 1. A letter (latitude) and number (longitude) grid system locates features or localities referred to in the text. These locations are in parenthesis without reference to figure 2. For example, (3g, J4) refers to the Greylock slice (3g) in grid location J4. Numbers 1-8 mark specific locations referred to in the text and are indicated by the number and the grid location. For example, (6, F6) refers to the Plymouth section in grid location F6. Symbol  $\Delta$  indicates important localities of carbonate platform slivers in the Taconic slices. Symbol T in AB is a glauconite locality at Tillston Peak. Short line with x's (Worcester Mountains) and line with rhombs (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones of Albee (1968). Widely-spaced diagonal lines in northwestern Vermont outline the region that contains medium-high pressure amphibolites described by Laird and Albee (1981b). Irregular black marks are ultramafic bodies. Dashed line in J5 separates allochthonous Hoosac with mafic volcanic rocks to the north from allochthonous Hoosac without mafic volcanic rocks. Section lines A-A', B-B', and C-C' refer to enlarged cross sections shown in figure 2a and 2b of plate 1 and section 1 of plate 2 respectively. Open teeth of thrust fault symbols mark speculative thrust zones.

**SYMBOLS FOR PLATES 1 AND 2 ARE LISTED UNDER THE APPROPRIATE LITHOTECTONIC UNIT SHOWN IN FIGURE 1. SYMBOLS FOR MAJOR STRUCTURES ALONG THE WESTERN BOUNDARY OF EACH LITHOTECTONIC UNIT ARE LISTED BETWEEN THE APPROPRIATE UNITS.**

**PARAUTOCHTHONOUS NORTH AMERICAN CRUST AND COVER—UNIT 1—Yad.** Middle Proterozoic of the Adirondack massif. Yg, Middle Proterozoic of the Green Mountain massif; YL, Middle Proterozoic of the Lincoln massif; Yh, Middle Proterozoic of the Housatonic massif; Y, Middle Proterozoic between the Green Mountain massif and the Taconic slices, Vermont; Ygh, Middle Proterozoic of Ghent block (K2); Ya, Middle Proterozoic at Stissing Mountain (M2); Ocp, Cambrian and Ordovician rocks of the carbonate-siliclastic platform; CZp, Pinnacle Formation; CZI, Fairfield Pond Formation; CZd, Dalton Formation; CZdh, interlayered Dalton and Hoosac Formations where the Dalton predominates; CZh-CZl, autochthonous Tyson-Hoosac sequence of the Plymouth section on the northeast side of Green Mountain massif (F5). Cm, Monkton Quartzite; Cp, Potsdam Formation; Ocs, Stockbridge Formation and equivalent rocks; Ob, Bascom Formation; Ow, Wallomac Formation and equivalent rocks; Owm, marble at the base of the Wallomac Formation; Oww, wildflysch in the Wallomac Formation; Owag, Austin Glen Graywacke in the Wallomac Formation; Oha, Hathaway Formation of northwestern Vermont; Odr, Black River Group; Ot, Trenton Group; PHT, Phillipsburg thrust; HSPt, Highgate Springs thrust; PT, Pinnacle thrust; OT, Orville thrust.

**TACONIC ALLOCHTHONOUS—UNIT 2—Group 1 slices:** 1Ls, Livingston slice (Austin Glen lithotectonic unit) of Fisher (1977); 1gb, (GB) Giddings Brook slice (including Sunset Lake slice); On, Normanskill Formation and equivalent rocks; Group 2 slices: 2bm, (BM) Bird Mountain slice; 2b, (B) Berlin slice; 2rp, (RP) Rensselaer Plateau slice; 2e, (C) Chatham slice; 2c, (E) Everett slice; Group 3 slices: 3d, Dorset Mountain slice; 3p, Greylock slice; 3j, June Mountain slice; 3cm, Canaan Mountain slice; CZsc, St. Catherine Formation of Vermont; CZbr, Breese Formation of Vermont; CZg, Greylock Formation of Massachusetts; CZcm, Canaan Mountain Formation of Connecticut.

**Boundary—HoT, Hoosic thrust.**

**ALLOCHTHONOUS BERKSHIRE MASSIF AND MIDDLE PROTEROZOIC DOMES OF SOUTHEASTERN VERMONT—UNIT 3—Yb.** Middle Proterozoic of Berkshire massif; Yw, Middle Proterozoic around Wilmington, southeastern Vermont; Ya, Middle Proterozoic of Athens dome; Yc, Middle Proterozoic of Chester dome; Yj, Middle Proterozoic at Jamaica; CZnd, autochthonous Hoosac Formation deposited on the Middle Proterozoic of Berkshire massif; CZhc, autochthonous Hoosac Formation deposited on the Middle Proterozoic of the Chester, Athens and Wilmington domes.

**Boundary—MFZ, Middlefield thrust zone; HST, Hoosac Summit thrust.**

**ALLOCHTHONOUS HOOSAC FORMATION—UNIT 4—Czht.** CZht, mafic volcanic rocks in the allochthonous Hoosac Formation.

**Boundary—CLT, Cameron's Line thrust; WST, Whitcomb Summit thrust; PHT, Pinney Hollow thrust; UT, Underhill thrust; HT, Hinesburg thrust.**

**EASTERN VERMONT SLICE AND ROWE-HAWLEY SLICES—UNIT 5—U.** ultramafic rocks; CZhp, Hoosac and Pinnacle Formations; CZh, autochthonous Hoosac Formation; CZph, Pinney Hollow Formation; CZu, Underhill Formation; CZuj, Jay Peak Member of Underhill Formation; Ocu, Undifferentiated Cambrian-Ordovician rocks around the Chester and Athens domes; Co, Ottauquechee Formation; Cs, Stowe Formation; Ocr, Rowe Schist; Ocr, gray to black carbonaceous schist member; Om, Moretown Formation; Oms, amphibolites; Oms, garnet schist; UH, Umbagog Hill Formation; Oh, Hawley Formation and its equivalents in Vermont; Ohp, plagioclase-hornblende schist; Ohf, felsic granolites; Ohv, mafic volcanics; Oc, Cobble Mountain Formation; Oca, Ocb, Ocs, Members A, B, and C of Cobble Mountain Formation; Ooc, Catcaule Member of the Collinsville Formation; JS, Jerusalem slice; US, Underhill slice; HNS, Hazens Notch slice; MVFZ, Missisquoi Valley fault zone; PHS, Pinney Hollow slice; RTZ, Rowe thrust zone; BMT, Belvidere Mountain thrust; WMT, Winchell Mountain thrust.

**Boundary—BT, Bristol thrust; CHT, Coburn Hill thrust.**

**CORE ROCKS OF DOMES OF WESTERN MASSACHUSETTS AND WESTERN CONNECTICUT—UNIT 6—Zdh.** Dry Hill Gneiss; OZmo, Monson Gneiss; Oco, Collinsville Formation; Op, Partridge Formation; Oa, Ammonoosuc Volcanics; Oo, gneisses of the Olverian domes; SFD, Shelburne Falls dome; GD, Goshen dome; GRD, Granville dome; GvD, Granby dome; CD, Collinsville dome; BD, Bristol dome; WD, Waterbury dome; although the Ascut-Wooden sequence (Oa) is included in this unit, we do not consider it to be part of the Bronson Hill arc complex which includes the dome rocks of unit 6.

**Unconformity—DS.** Silurian and Devonian rocks; Og, Goshen Formation; Dgl, Goshen Formation with calcareous beds; Dw, Waits River Formation; Dwt, Waits River Formation with thick calcareous beds; Dgm, Gile Mountain Formation with abundant quartzite; DI, Littleton Formation.

**Unconformity—M.** Mesozoic border faults. TRS, Triassic rocks.

(Wallomac and its Formation in Massachusetts and Vermont), is present beneath the Dorset Mountain slice. This basement thrust is interpreted to be a reactivated older normal fault along which the upper plate has moved over a ramp to produce the anticline shown in the section. Some of these late Taconian faults are shown cutting the older slices of the Taconic allochthons. This interpretation may explain the presence of other Proterozoic blocks between the Green Mountain massif and the Taconic allochthons (see Doll and others, 1961). 3) our analysis indicates that the eastern surface exposure of the root zone for the Taconic slices is located along the east side of the Chester and Athens domes at this latitude. Symbols are explained in figure 2 and 6. Heavy lines with arrows are faults. Arrow indicates the displacement of the hanging wall or upper plate. Lighter lines represent depositional contacts. The cross section is enlarged by 105 percent compared to the map scale of figure 2. The letters B and B' are in the same geological location as they are in figure 2. The correlation of the sialic slices between cross sections A-A' and B-B' of figure 2a and 2b and cross section C-C' (section 1, plate 2) is the following: a. The slices of the Berkshire massif (Yb) are correlated to the slices of the Chester and Athens domes (Yc,Ya) as well as to the Middle Proterozoic rocks in the Wilmington and Jamaica area of southeastern Vermont (Yw, Yj, fig. 2). Although this correlation is proposed on the basis of structural and stratigraphic reasons, faults may separate many of these masses. Figure 6 shows these faults schematically although no surface evidence for them is known at present; b. The Green Mountain massif is present in all sections and is floored by the Champlain thrust fault; either the Housatonic slice or Reading Prone may correlate with the sialic slices that contains denser rocks in the structural section to the north. Thus, the lower two slices of cross section C-C' (section 1, plate 2), the Housatonic (Yb) and the Green Mountain (Yg) slices, are thick plates that may be regionally continuous. The Berkshire slice (Yb) and those in southeastern Vermont (Yw, Yj, Yc, Ya) are missing in the northern section (fig. 2a). These slices clearly increase in number in the latitude of the Berkshires (Ratcliffe and Hatch, 1979) and to the south (Ratcliffe, 1982, 1984).

**Figure 6**  
Composite palinspastic diagram for western New England showing relative depositional sites for Late Proterozoic to Middle Ordovician rocks deposited on or between the North American continent and the Bronson Hill arc complex. Autochthonous carbonate-siliclastic-shale sequence (Ocp and equivalents) with its clastic rift basal section (CZd, CZdh, CZnd, CZht) rest on imbricated slices of Middle Proterozoic North American crust (random dash pattern). To the east along the continental margin are the rocks of the Taconic slices. The irregular vertical contacts are shown to emphasize the gradational stratigraphy of the slices and the fact that they represent a continuous continental margin sequence. CZnr represents the Rensselaer Graywacke Member of the Nassau Formation; the v pattern refers to mafic volcanic rocks. The Pinney Hollow, Hazens Notch, and Rowe slices represent eastward or basal sequences deposited on transitional or ocean crust. The folds and faults in each of these slices show that these rocks were deformed repeatedly in the earlier accretionary history of the margin before they were displaced as a whole to the west during the election of the Taconic slices. The black lenses represent "sheared-in" fragments of ultramafic rocks that originated from oceanic crust. The lenses with "V" symbols are mafic volcanic rocks that were either part of ophiolites or interstratified with ocean floor or continental margin sediments. The proto-Rowe refers to the original sediments that have undergone a long history of repeated deformation resulting in the Rowe Schist that was present at the end of the Taconian orogeny along the eastern flanks of the Middle Proterozoic masses. The tectonized part of the Moretown Formation contains ultramafic fragments and represents the basal part of the forearc basin that was tectonically involved with the accretionary wedge. Proto-Moretown represents earlier undeformed sediments in the forearc basin. The present "pinstriped" fabric typical of the Moretown rocks is considered to be a tectonic fabric resulting from compression between the accretionary wedge and the arc-complex to the east. The Bristol thrust, which places the arc-complex of the Bronson Hill over the forearc sequence of the Moretown, is unconformably overlain by the Cobble Mountain Formation and its equivalents. This unit represents a late forearc basin fill derived largely from the arc itself. Fragments of ultramafic rocks (u) and Rowe Schist (R) are olistoliths in member C (Ooc) of the Cobble Mountain Formation (Oo) derived from the emerged accretionary wedge to the west. The Winchell Mountain thrust is interpreted as a west-over-east or backthrust that places distal rocks over proximal rocks of the lower two members of the Cobble Mountain Formation.

Schematic thrust zones are represented by curved lines with arrows marking the direction of movement. Their relative age is shown by crosscutting relations. For example, the Belvidere Mountain-Whitcomb-Summit-Cameron's Line thrust zone is a reactivated zone which has continued to move late in the Taconian orogeny, structurally overlapping and burying beneath it more westerly-situated slices. Thus, it is shown cutting the Hazens Notch and Pinney Hollow slices. In the Taconic slices the thrust zones flooring the three groups are numbered 1, 2, and 3. The cross cutting relations in their depositional site show when they formed. Group 2 slices, therefore, are shown cutting Group 1 slices. The resulting stacking order of Group 1, 2, and 3 in ascending order are shown to the west on the autochthonous rocks of the continental platform. The sialic slices, which formed late in the sequence, are shown cutting all the easterly derived slices. Except for the eastern slices of the Taconic allochthons, much of the surface evidence for these age relations has been eroded and is based on our interpretation of the chronology. The slice relations in northwestern Vermont are in the upper part of the diagram. Thrusts flooring the eastern two sialic slices are queried because they may be the same as the Hoosac thrust. The Hinesburg thrust is shown to die out in the basal part of the Cheshire Quartzite on the west side of the Lincoln massif. Appropriate depositional environments and plate tectonic settings are shown at the bottom of the diagram. Symbols are explained in figure 2.



of sedimentary-volcanic assemblages contained in the lithotectonic units mentioned above.

### Age of the Basement Rocks

Mineral and whole-rock dates in excess of 900 m.y. (Naylor, 1976) indicate that the core rocks of the Chester and Athens domes are age equivalents of the middle Proterozoic rocks of the Green Mountain and Berkshire massifs. Dates have not been obtained from the Wilmington area (Yw, I5). Rosenfeld (1972, p. 174) noted that the calc-silicate gneiss, rusty graphitic schist, and granitic gneiss similar to those in the Green Mountain and Berkshire massifs are unconformably overlain by cover rocks in the Chester dome. Granitic gneisses of probable intrusive origin (Tyringham Gneiss or Stamford Granite Gneiss) are common to all areas and are probably 1 b.y. old, based on zircon and Rb/Sr whole-rock studies in the Berkshire massif (Ratcliffe and Zartman, 1976). East-west structural trends, which developed in association with a pre-Dalton dynamothermal metamorphic event, are present in the Lincoln (DelloRusso, 1984, personal commun.), Green Mountain (Doll and others, 1961), and the Berkshire massifs (Zen and other, 1983). Unconformable sedimentary contacts with upper Proterozoic or Lower Cambrian cover rocks are preserved in each dome and massif.

Rocks exposed in the core of the Wilmington area (Yw, I5) were regarded by Skehan and Hepburn (1972) as middle Proterozoic because they are similar to the rocks of the Green Mountain massif. Ratcliffe (1979b) concluded that the complex assemblage of amphibolite, biotite-plagioclase gneiss, calc-silicate rock, calcite marble (Sherman Marble), and potash feldspar and megacrystic granitic gneiss (Wilmington Gneiss of Skehan, 1961), which Skehan (1961) previously assigned to the cover-

rock sequence, is equivalent to rocks of the Berkshire massif of proven middle Proterozoic, or older, age. All of these gneisses are interpreted as being exposures of North American continental crust that was consolidated prior to late Proterozoic rifting. Gneisses of the southeastern Vermont domes and the Berkshire massif belong to the same structural level. They are separated from the basement of the Green Mountain, Housatonic, and Hudson Highland massifs by the Hoosac thrust. This interpretation is required by the evidence of at least 21 km of composite foreshortening on the Berkshire massif on the basal detachment surface (Beartown Mountain thrust; Ratcliffe, 1975).

### Cover Rocks of Middle Proterozoic Massifs and Domes

The Dalton and Pinnacle Formations, Forestdale Marble, and Moosalamoo Phyllite west of the Green Mountain, and Berkshire massifs and the Hoosac Formation and Cavendish Formation (of Doll and others, 1961) directly to the east form a coherent clastic blanket that immediately overlies the basement unconformably. Although the cover is broken by faults, the marked similarity between each section of cover rocks and critical exposures of their interfingering relations demonstrate their original stratigraphic relations despite the lack of fossil control in the eastern sections. These relations require that the root zone for the Taconic slices be located east of the massifs rather than on top of them, as argued by Zen (1967, 1972). Furthermore, the facies relations in the clastic cover suggest that the Berkshire massif was originally situated east of the Green Mountain massif and was part of an eastern North American basement that is now exposed in the Wilmington area (Yw, I5) and in the Chester and Athens domes (Yc, Ya, G6, H6). Figures 3, 4, and 5 summarize the critical stratigraphic relations and form the basis of our discussion.

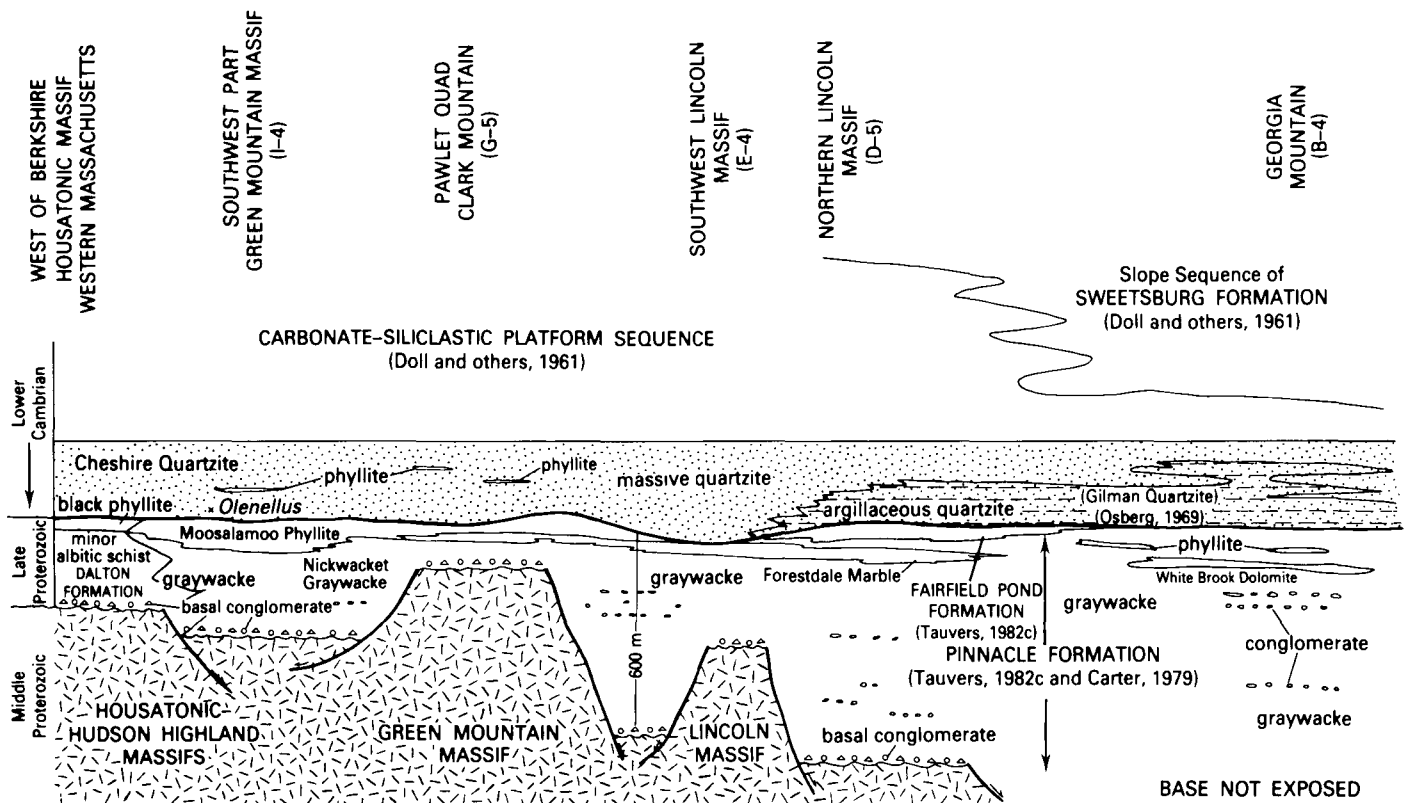


Figure 3. South-to-north generalized stratigraphic relations of the late Proterozoic to Lower Cambrian section along the western side of the parautochthonous middle Proterozoic massifs of lithotectonic unit 1, showing suggested correlation of units that have been used in various ways by different authors cited in the text. Letter-number designations in parentheses refer to geographic location on Figure 2 of Plate 1.

The clastic sequence that overlies basement rocks along the western boundary of the Lincoln Mountain (YL, D5, E5) and Green Mountain massifs is shown in a south-to-north section (Fig. 3). Included here is the clastic sequence overlying the Housatonic and Hudson Highland massifs to the south. Four important relations are:

1. The unsystematic northward increase in thickness of the pre-Cheshire graywackes, arkoses, and conglomerates of the Pinnacle Formation below the carbonate-black-shale section of the Forestdale and Moosalamoo suggests a horst and graben configuration (Fig. 3). The faults shown here probably trend northward parallel to the mountain belt and thus cut the section at a small angle.

2. The sedimentary rocks of the Pinnacle Formation indicate a rift environment (Tauvers, 1982a, 1982b, 1982c; DiPietro, 1983a, 1983b). The basal conglomerates are clast-supported fluvial deposits indicative of rapid, subaerial flash floods. The clasts in the conglomerate are identical to the local middle Proterozoic plagioclase gneiss of the Lincoln massif. Blue quartz and magnetite-bearing graywacke and schistose graywacke overlie the basal conglomerates. Intraformational conglomerates higher in the section are matrix-supported, contain slate and chert, and are associated

with graded sequences of graywacke, indicating subaqueous deposition. Plagioclase persists through the section up to the top of the Forestdale Marble, whereas potash feldspar is present in the argillaceous part of the Cheshire Quartzite. Tauvers (1982b) suggested that the change in feldspar marks a change in source from local horst-derived clastic sediments of the sub-Forestdale section to a more cratonal, western source for the overlying sequence that is transitional to the shallow-water marine sediments of the Cheshire.

3. Rusty schistose dolostone, brown- to buff-weathering massive dolostone, and sandy dolostone of the Forestdale Marble appear in the section between the Pinnacle graywackes and the black shale of the Moosalamoo Phyllite north of Pawlet (G5) but are absent to the south. At the north end of the Lincoln massif, the carbonate rocks are found in both the upper and lower parts of the Pinnacle. Further to the north at Georgia Mountain (CZp, B4), buff- to brown-weathering sandy dolostone is present in the upper Pinnacle (Carter, 1979; Dorsey and others, 1983) and is correlated with the White Brook Dolomite mapped between the Pinnacle and the Fairfield Pond Formations (as used by Tauvers, 1982c) near the Canadian border (Dennis, 1964). Although these carbonate units have been correlated with each other, their discontinuous occurrence indicates that they may not be time equivalents. In the Lincoln area, Tauvers (1982b and 1982c) found graded beds and debris flows with carbonate and shale clasts in the dolostone, suggesting a redeposited doloarenite shed from a

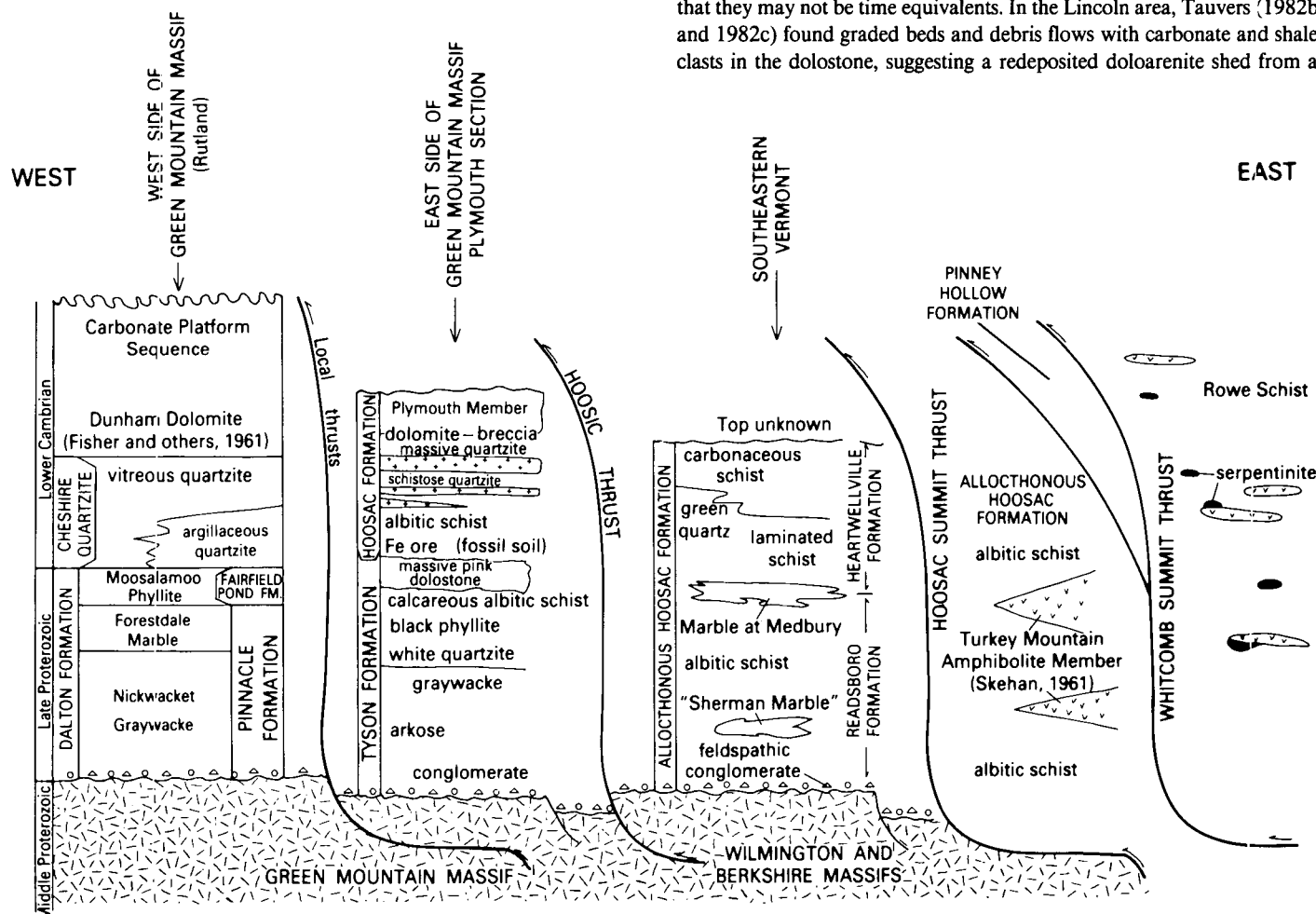


Figure 4. West-to-east relations of the Dalton Formation, Cheshire Quartzite, and Tyson and Hoosac Formations in central and southern Vermont between latitudes  $43^{\circ}00'$  and  $43^{\circ}30'$ . Stratigraphic names for the Plymouth section are adapted from Doll and others, 1961. Stratigraphic names for the autochthonous and allochthonous Hoosac Formation are adapted from Skehan, 1961. The Whitcomb Summit thrust (Fig. 2, WST) and Rowe Schist (Pl. 1, Fig. 2, OCr) are extended northward along the west boundary of the Rowe-Hawley slices in Vermont so as to show their relations to the western sequences.



local carbonate bank. These relations suggest that the carbonate rocks in the upper Pinnacle or along the Pinnacle–Fairfield Pond contact represent a series of local, nontime-equivalent carbonate banks capping the filled or partially uplifted rift sequence.

4. Minor, but important, albitic schist is in the Pinnacle on the west side of the Lincoln massif (Osberg, 1952) and in the middle two members of the Dalton Formation in western Massachusetts (Ratcliffe, *in* Zen and others, 1983). These rocks are interpreted as representing the western facies of the albitic schist of the allochthonous Hoosac Formation on the east side of the Berkshire massif.

West-to-east relations in the vicinity of the Green Mountain massif and those at the north end of the Berkshire massif are important in showing the gradual change in facies eastward (Figs. 4 and 5). The section at Plymouth, Vermont (location 6, F6), is assumed to be lithologically similar and correlative with cover rocks along the west limb of the Green Mountain massif. Thompson (1972, Fig. 1) showed that the basal upper Proterozoic Tyson Formation (of Doll and others, 1961) in the Plymouth section is physically continuous with the Dalton Formation across the northern end of the Green Mountain massif. Karabinos (1984) and Karabinos and Thompson (1984) confirmed these facies relations from the Jamacia area (F6) to the northern end of the Green Mountain massif, although the basement and cover are cut by numerous thrust faults throughout this region. The massive pink dolostones in the upper part of the Tyson may well represent the same type of local carbonate bank deposits as does the Forestdale along the western limb of the massif, although they may possibly be older. The lower calcareous albitic schist of the Tyson could be doloarenite flows similar to those in the Forestdale. The concentration of iron oxide at the top of the Tyson led Thompson (*in* Chang and others, 1965) to suggest an erosional unconformity. The basal part of the Tyson is

magnetite-bearing graywacke, arkose, and conglomerate, lithically similar to the Pinnacle to the west. Albitic schist is prevalent throughout the upper part of the Tyson and represents, according to us, an eastward facies change from graywacke to albitic rocks of the late Proterozoic section around the domes of southeastern Vermont.

The overlying Hoosac Formation in the Plymouth section has many features that are similar to rocks mapped as part of the Cheshire–Dunham sequence along the west side of the Green Mountains. Although albitic schist is characteristic of the Hoosac, the middle part of the formation consists of schistose quartzite and interlayered carbonaceous schist that passes upward into massive quartzite (Chang and others, 1965). This sequence is very similar to the Cheshire to the west, which is argillaceous quartzite at the base and massive quartzite at the top (Myrow, 1983). Gray to buff marble with local dolomitic breccia makes up the Plymouth Member, as used by Chang and others (1965), which is the uppermost member of the Hoosac. Thompson (*in* Chang and others, 1965) correlated this with the Dunham Dolomite of the western Vermont sequence. The total sequence at Plymouth is ~720 m thick, compared to 650 m on the west side of the massif.

To the south in the Wilmington area (Yw, I5), Skehan (1961) defined a sequence (Cavendish Formation of Doll and others, 1961) that he interpreted as having been deposited earlier than was the Tyson Formation in the eastern Vermont sequence. Zen (1967, p. 25) suggested that this sequence is, indeed, a western facies of the lower part (Middle and Upper Cambrian Ottauquechee Formation and older) of the more easterly situated eastern Vermont eugeoclinal sequence. We interpret the stratigraphic range of the Cavendish, however, to be restricted to the eastern Hoosac Formation of late Proterozoic and Early Cambrian age. The Cavendish was divided into the Readsboro and Heartwellville Formations by Skehan

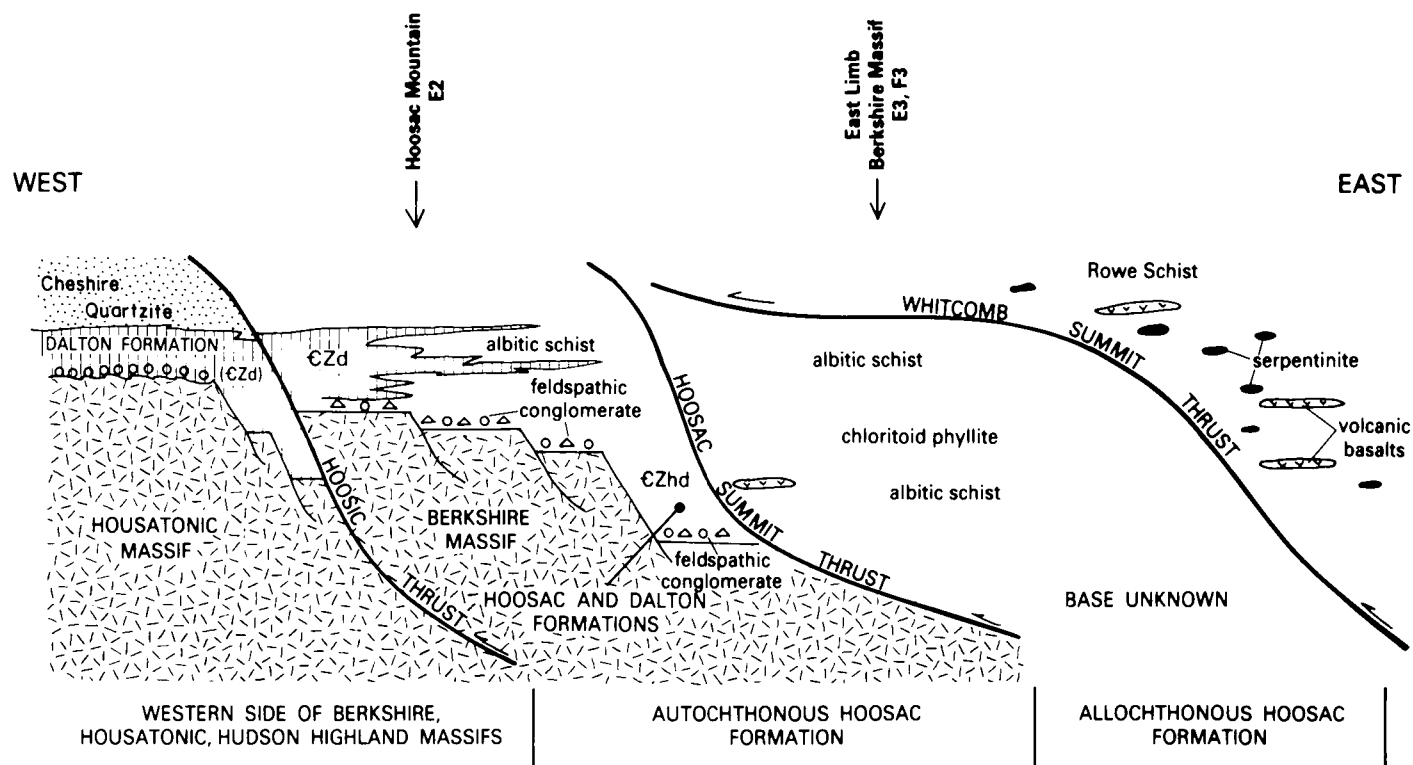


Figure 5. West-to-east relations of the Dalton and Hoosac Formations at the north end of the Berkshire massif along an approximate latitude of 42°42' in grid location J4. Stratigraphic names are adapted from Zen and others, 1983. Middle Proterozoic rocks are marked by a random dash pattern.



(1961). The Readsboro contains albitic schist with discontinuous lenses of marble (including some at Medbury) which may be similar in origin to the marbles of the Tyson Formation and Forestdale Marble to the west (Ratcliffe, 1979b). The basal conglomerates are feldspathic, similar to those found unconformably on the middle Proterozoic gneisses of the Berkshires. The graywacke and arkose of the Tyson are not present, and the albitic rocks in the lower sequence represent a more easterly, albite-rich pelitic facies of the clastic rocks to the west. We therefore interpret the basal albitic rocks of the Cavendish Formation (Readsboro Member of Doll and others, 1961) as being an easterly, and perhaps older, facies of the more westerly clastic rocks of the Tyson and Pinnacle Formations.

Although both the Hazens Notch Formation in northern Vermont (Fig. 2) and the allochthonous Hoosac to the south contain albitic carbonaceous and noncarbonaceous schist and mafic volcanic rocks, the Hazens Notch is distinctly different in that it contains serpentinite bodies, which are mapped along faults in the eastern part of the belt (Stanley and Roy, 1982; Stanley and others, 1984). We believe that these ultramafic rocks were tectonically incorporated into the basal albite-rich sediments that originally were deposited on transitional ensimatic crust. They thus represent a still more easterly basal facies of the allochthonous Hoosac Formation of southern Vermont and western Massachusetts.

From the above discussion, we conclude that at least the lower part of the carbonate platform exists to the east of the Green Mountain massif (Cheshire-Dunham part of the section at Plymouth, Vermont). This conclusion is important in locating the Taconic root zone.

The pre-Cheshire section in western Massachusetts (Fig. 5) displays critical relations at the northern end of the Berkshire massif on Hoosac Mountain where the autochthonous Hoosac Formation (location 1, J4) has been studied by Ratcliffe (1979a). In accordance with the earlier conclusions of Pumpelly and others (1894), Prindle and Knopf (1932), and Herz (1958), the type Hoosac is known to rest unconformably on middle Proterozoic basement. A basal boulder conglomerate consisting of clasts of gneiss in an albitic matrix appears to fill uneven depressions in the underlying basement complex. The basal conglomerate is overlain by albitic granofels with very distinctive disarticulated beds that produce pseudoconglomerate. Quartz pebble and cobble conglomerate decreases upward and is overlain discordantly by a regionally mappable, dark colored, carbonaceous, biotite-muscovite-albite-garnet-quartz schist that locally rest directly on basement and may be lithically similar to the black phyllite of the Moosalamoo Phyllite along the west side of the Green Mountain massif (Fig. 3). Local relief on top of the basement exceeds 50 m and may be as much as 100 m. This relationship, coupled with the extremely coarse-grained, unsorted character of the basal conglomerate, suggests local infilling of depressions that were possibly controlled by faults. Infilling was completed before deposition of the carbonaceous quartzose schist. Distinctive light yellow-gray feldspathic quartzite typical of the Dalton Formation is interbedded with this dark schist and is present sporadically upward in the dark albitic schists of the Hoosac (CZd, Fig. 5). This dark schist passes upward into feldspathic quartzite or flaggy tourmaline-rich quartzite typical of the Dalton Formation. This sequence of rocks closely resembles the mapped sequence of flaggy quartzite, identified as Tyson, beneath the Plymouth Member of the Hoosac Formation northeast of Plymouth, Vermont (Thompson, 1972). Eastward, the autochthonous sequence on Hoosac Mountain is more schistose but still contains beds of Dalton or Cheshire-like quartzite (Norton, 1967).

We propose that the "autochthonous" Hoosac sequence in depositional contact with Berkshire middle Proterozoic gneisses on Hoosac Mountain is equivalent, in part, to the Hoosac (Cavendish Formation of Doll and others, 1961) that overlies the basement of Wilmington Gneiss of Skehan (1961). The autochthonous Hoosac sequence grades upward into

light green albite schist and chloritoid phyllite containing minor beds of vitreous quartzite. This sequence is interbedded with feldspathic quartzite of the Dalton Formation on Hoosac Mountain (Fig. 5; Ratcliffe, 1979a, Fig. 5).

Albitic and lustrous chlorite-garnet schists similar to those described above are present on Hoosac Mountain east of the Hoosac Summit thrust, which extends southward into the Middlefield fault zone. The allochthonous Hoosac Formation also contains 0.5- to 1-m-thick beds of greenish vitreous quartzite similar to the Curtis Mountain Quartzite (of Fisher, 1962) of the Taconic slices. Feldspathic granofels of the Dalton is absent. Locally, epidote layers or lenses in amphibolite are recognized (Turkey Mountain Amphibolite Member of the Hoosac Formation of Skehan, 1961). This sequence of allochthonous Hoosac resembles very closely the Greylock Schist of the Greylock slice (3g, J4; Ratcliffe, 1979a), whereas the autochthonous Hoosac does not resemble any of the Taconic allochthonous rocks. Lustrous green to gray garnetiferous schist, lithically similar to the Gassetts Schist of the Cavendish Formation in the Chester dome, has been mapped in the allochthonous Hoosac by Norton (1976) and Ratcliffe (1979a). Herz (1958) and Norton (1976, p. 364, unit 7) also reported thin calcite-bearing schist south of Hoosac Mountain but no dolostone or calcareous marbles similar to the marble of the Plymouth Member. The allochthonous Hoosac Formation has few calcareous rocks and is, therefore, a more easterly facies of the autochthonous Hoosac Formation (formerly Cavendish) of southeastern Vermont.

From Hoosac Mountain (location 1, J4), the allochthonous Hoosac is traced northward into Vermont where it is interbedded with mappable belts of amygdaloidal amphibolite and actinolitic greenstone (15, Turkey Mountain Amphibolite Member; Skehan, 1961). This section continues northward to Middletown (location 8, G6), Vermont, where it terminates against the western contact of the Pinney Hollow Formation. Directly north of Middletown in the section at Plymouth, Vermont (location 6, F6), greenstone and amphibolite are absent from the autochthonous Hoosac. South of Sherman Reservoir (15, dashed line) in southern Vermont, the mafic volcanic rocks are missing from the allochthonous Hoosac Formation except for a few small outcrops near latitude 42°23' (Hatch, 1969). The plagioclase schist, so typical of the allochthonous Hoosac to the north, disappears near latitude 42°15'. It is replaced by gray to gray-brown, rusty and nonrusty quartz-plagioclase-mica schist and gneiss, kyanite-garnet-quartz-mica schist with distinctive knots and ribs of quartz, and plagioclase and minor calc-silicate granofels (Hatch and Stanley, 1976). These schists and a rusty, muscovite-spangled schist continue southward where they are mapped as Hoosac Formation (Schnabel, 1976, Fig. 1). Amphibolite is again important in the allochthonous Hoosac (Martin, 1970, Pl. 1) and becomes very abundant in the rocks east of the Housatonic massif (Gates, 1961, Pl. 1), which consist of feldspathic mica granofels and "hobnailed" garnet-sillimanite schist lithically identical to the schist near latitude 42°15' and the schist making up the Canaan Mountain slice (3cm, M3; Harwood, 1975). The following conclusions are important.

1. The pre-Cheshire clastic rocks form a coherent sedimentary sequence that was originally deposited in ensialic rift basins. The western part of this sequence consists of locally derived graywacke, arkosic sandstones, conglomerate, and minor albitic schist of the Dalton Formation. To the east, this sequence grades into albite-rich, tourmaline-bearing, sandy and silty shales and basal clastic rocks of the autochthonous Hoosac Formation.

2. The allochthonous Hoosac represents a still more eastern facies in which aluminous and iron-rich schist (green, albitic, and lustrous, chloritoid-bearing Hoosac) are interlayered with the more westerly autochthonous Hoosac. Basalt is important to the north and south and probably indicates local rift volcanic rocks.



PLATE 2—PALINSPASTIC RECONSTRUCTION OF THE  
TACONIAN OROGENY IN WESTERN NEW ENGLAND

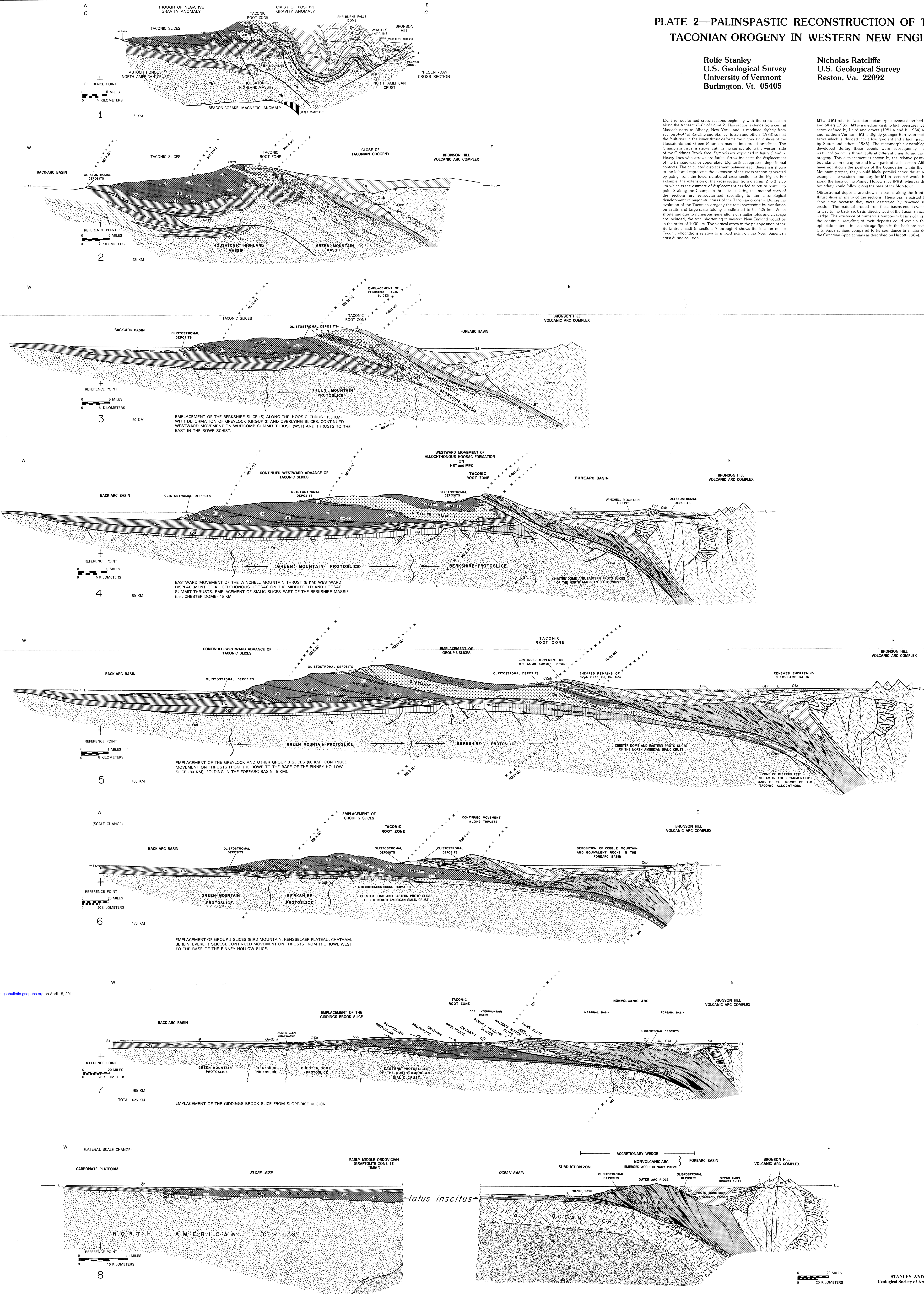
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Eight retrodeformed cross sections beginning with the cross section along the transect C-C' of figure 2. This section extends from central Massachusetts to Albany, New York, and is modified slightly from section A-A' of Ratcliffe and Stanley, in Zen and others (1983) so that the fault rise in the lower thrust deforms the higher sialic slices of the Housatonic and Green Mountain massifs into broad anticlines. The Champlain thrust is shown cutting the surface along the western side of the Giddings Brook slice. Symbols are explained in figure 2 and 6. Heavy lines with arrows are faults. Arrow indicates the displacement of the hanging wall or upper plate. Lighter lines represent depositional contacts. The calculated displacement between each diagram is shown to the left and represents the extension of the cross section generated by going from the lower numbered cross section to the higher. For example, the extension of the cross section from diagram 2 to 3 is 35 km which is the estimate of displacement needed to return point 1 to point 2 along the Champlain thrust fault. Using this method each of the sections are retrodeformed according to the chronological development of major structures of the Taconian orogeny. During the evolution of the Taconian orogeny the total shortening by translation on faults and large scale folding is estimated to be 625 km. When shortening due to numerous generations of smaller folds and cleavage are included, the total shortening in western New England would be in the order of 1000 km. The vertical arrow in the paleogeographic of the Berkshire massif in sections 7 through 8 shows the location of the Taconic allochthons relative to a fixed point on the North American crust during collision.

M1 and M2 refer to Taconian metamorphic events described by Sutter and others (1985). M1 is a medium-high to high pressure metamorphic series defined by Land and others (1981 a and b, 1984) for central and northern Vermont. M2 is slightly younger Barrovian metamorphic series which is divided into a low gradient and a high gradient series by Sutter and others (1985). The metamorphic assemblages which developed during these events were subsequently transported westward on active thrust faults at different times during the Taconian orogeny. This displacement is shown by the relative position of the boundaries on the upper and lower parts of each section. Although we have not shown the position of the boundaries within the Taconian Mountain proper, they would likely parallel active thrust zones. For example, the western boundary for M1 in section 6 would be located along the base of the Pinney Hollow slice (PHS) whereas the eastern boundary would follow along the base of the Moretown.

Olistostromal deposits are shown in basins along the front of active thrust slices in many of the sections. These basins existed for only a short time because they were destroyed by renewed uplift and erosion. The material eroded from these basins could eventually find its way to the back-arc basin directly west of the Taconian accretionary wedge. The existence of numerous temporary basins of this type and the continual recycling of their deposits could explain the lack of ophiolitic material in Taconic-age flysch in the back-arc basins of the U.S. Appalachians compared to its abundance in similar deposits in the Canadian Appalachians as described by Haxel (1984).





3. We consider the Hazens Notch Formation to be an even more easterly derived facies of the Hoosac Formation that may well have been deposited on transitional or ensimatic crust, as evidenced by the ultramafic rocks.

4. Group 3 slices of the Taconic allochthons must root east of, rather than above, the middle Proterozoic rocks of the domes and massifs of southeastern Vermont and western Massachusetts for the following reasons. (a) The autochthonous Hoosac Formation was deposited unconformably on the Proterozoic gneisses, although faults now locally offset the unconformity. (b) The autochthonous Hoosac Formation is most reasonably interpreted as being an eastern facies of the Plymouth section. (c) The quartzites and carbonate rocks of the Hoosac Formation in the Plymouth section probably represent the lower part of the carbonate platform on the east side of the Green Mountain massif. For these reasons, the root zone for the slope-rise sequence of the Taconic slices (group 3) must be located within, or beneath, the Pinney Hollow Formation east of the Green Mountain massif rather than on top of it, as suggested by Zen (1967, 1972) and others.

5. The clastic cover of the Berkshire massif is more similar to the late Proterozoic cover of the dome gneisses in southeastern Vermont than it is to the cover of the Green Mountain–Lincoln massifs. This relationship indicates that the Berkshire massif represents a more easterly part of the North American crust compared to the Green Mountain and other massifs to the west.

#### Stratigraphic Parameters in the Taconic Slices

The stratigraphy of the Taconic region has been extensively discussed and reviewed by Zen (1967, 1968, 1972), Rodgers (1970, p. 79–82), Ratcliffe (1974a, 1974b, 1974c, 1979a), Ratcliffe and others (1975), Ratcliffe and Hatch (1979), Potter (1963, 1972, 1979), and Harwood (1975). The following discussion begins with group 3 slices and continues downward through the stacking sequence of group 2 and group 1 slices which lie successively westward.

**Group 3 Slices.** From north to south, group 3 slices include the Dorset Mountain slice (3d, H4), the Greylock slice (3g, J4), the June Mountain slice (3j, L3), and the Canaan Mountain slices (3cm, M3). The previous discussion of the autochthonous and allochthonous Hoosac Formation and its equivalents provides the sedimentological framework for locating the root zone of group 3 slices. There is a strong latitudinal correspondence between facies in the allochthonous Hoosac and in the individual slices of group 3. In southern Massachusetts and western Connecticut, the Canaan Mountain slice (3cm, M3) and, presumably, the June Mountain slice (3j, L3) root along the western thrust fault of the allochthonous Hoosac (€Zht, M3, M4) (Harwood, 1975, Fig. 2; 1979). Harwood (1975, p. 122) interpreted the Canaan Mountain Formation as being a transitional facies between the Dalton Formation to the west and the Hoosac Formation to the east. For example, micaceous granofels in the Canaan Mountain slice is very similar to granofels in the Dalton Formation. Associated with these rocks in the Canaan Mountain Formation, however, are “hobnail” garnet-sillimanite schist, muscovite-spangled gneiss, and amphibolite, typical of the allochthonous Hoosac to the east.

Ratcliffe (1979a, p. 399–401) showed that the Greylock slice (3g, J4) is lithically very similar to the allochthonous Hoosac east of the Hoosac Summit thrust and, therefore, suggested that its root zone is within the allochthonous Hoosac or buried beneath the Whitcomb Summit thrust. Prindle and Knopf (1932) suggested this earlier but included the Rowe Schist to the east of the Whitcomb Summit thrust in the root zone. A root zone along the Hoosac Summit thrust, however, cannot be ruled out. Minor, salmon-pink-weathered dolostone and quartz pebble conglomer-

ate described by Ratcliffe (1979a, p. 397) in the Greylock Schist (J4) may well be equivalent to the orange-weathered dolostones described by Skehan (1961) at Medburyville and Mount Pisgah in the autochthonous Hoosac (€Zhc, I5). Pink dolostone and quartz conglomerate are not present in the allochthonous Hoosac. Furthermore, the Greylock Schist contains none of the Dalton-like quartzite and dark schist of the autochthonous Hoosac on the Berkshire massif (€Zhd, J4).

Similarity between rocks of the Dorset Mountain slice (3d, G4) on Dorset Mountain (Thompson, 1967) and rocks of the Plymouth section (Chang and others, 1965; location 6, F6) led to the conclusion that the original depositional site of the Dorset Mountain slice could have been on the east side of the Green Mountains. As for the Greylock slice, the match with the eastern sequence is not perfect. The albitic rocks of the Netop Formation (of Thompson, 1967) and the greenish, chloritoid-bearing St. Catherine Formation (of Doll and others, 1961) closely resemble rocks of the autochthonous Hoosac and Pinney Hollow Formation near Plymouth. Sporadic, gray dolomite beds in the Dorset Mountain slice are associated with gray albitic phyllite and gray quartzite and are similar to the Plymouth Marble. Notable differences between the Pinney Hollow and the St. Catherine Formations are the absence of green quartzite and graywacke in the Pinney Hollow and the absence of basaltic volcanic rocks in the St. Catherine. In addition, feldspathic, Dalton-like quartzite interbedded in the Tyson Formation–autochthonous Hoosac Formation at Plymouth are absent from the Dorset Mountain slice. These systematic differences suggest that the Dorset Mountain slice can be reconstructed to a stratigraphic position transitional between the autochthonous Tyson–Hoosac section at Plymouth and the Pinney Hollow Formation directly east beneath the surface defined by the western boundary of the allochthonous Hoosac (€Zht) and the Pinney Hollow (PHS) north of Middletown, Vermont (location 8, G6).

In summary, all of the slices of group 3 are rooted either along the Middlefield–Hoosac Summit thrust (Fig. 2, MFZ-HST) or directly to the east within the allochthonous Hoosac (€Zht) between western Connecticut and Middletown, Vermont. To the north, the root zone would lie along the western contact of the Pinney Hollow or Underhill slices of Figure 2.

**Group 2 Slices.** Group 2 slices include the Bird Mountain (2bm, F4, G4), the Rensselaer Plateau (2rp, J3), the Chatham (2c, K2, K3, L2, M1), and the Everett (2e, L3, M2, N2) slices. The aerial extent of each slice is here modified from Zen's 1967 compilation on the basis of extensive new mapping. Assignment of all areas to individual slices is still in doubt, and along-strike correlations are especially uncertain. As noted by Zen (1967), the Chatham, Rensselaer Plateau, and Bird Mountain slices are all marked by the presence of coarse-grained graywacke (Rensselaer Graywacke Member of the Nassau Formation, Bird Mountain Grit as used by Zen, 1964) and associated basaltic volcanic rocks. Gray-green and purple slate or phyllite, some of which are well laminated, are associated with thick-bedded, medium- to coarse-grained feldspathic or lithic graywacke. Distinctive, but sporadically developed, diabasic basalts, pillow lavas, and pyroclastic volcanic rocks are spatially associated with the base of the graywacke facies which reaches its greatest distribution in the Rensselaer Plateau slice (2rp). These rocks interfinger with purple-green slate and phyllite of the Metawee Member of the Nassau Formation. Graywacke, volcanic rocks, slate, and phyllite, along with quartzite, are present throughout the Chatham slice (2c) where they comprise the lower part of the Nassau Formation. Light-green to greenish gray, chloritoid-bearing phyllite with minor amounts of quartz-pebble graywacke and quartzite characterize the Everett slice (2e) and may be equivalent to the basal part of the Nassau (Ratcliffe and others, 1975, Fig. 2) and the Biddie Knob Formation (Zen, 1961). The facies relations within, and between, these slices have been summarized by Ratcliffe and others (1975, Fig. 2, Table



1), Ratcliffe (1979a, Fig. 2), and Potter (1979, Fig. 3). The composition of the units within all of the slices was also discussed by Zen (1967, Fig. 4), who showed that the rocks in group 2 slices are probably restricted to the upper Proterozoic and Lower Cambrian section. Limestone conglomerate similar to the *Eliptocephala asphoides*-bearing Ashley Hill Limestone Conglomerate Member of the Nassau Formation (Bird and Rasetti, 1968) or North Britton Conglomerate Member of the Bull Formation (Zen, 1961) of the Giddings Brook slice is present in the Bird Mountain (2bm) and Rensselaer Plateau (2rp) slices. The Bird Mountain and Rensselaer Plateau slices contain the Middle Ordovician strata (Indian River Slate), but rocks in these slices are largely the equivalent of the pre-*Olenellus* Nassau Formation of Massachusetts and eastern New York (Ratcliffe, 1979a, p. 392). Zen (1967) and many other workers noted that the Nassau Formation in these slices is distinctly more feldspathic and coarser grained than are the Nassau or Bull Formations of the Giddings Brook slice.

Stratigraphic sequences equivalent to the rocks of group 2 slices are not preserved anywhere in the Eastern Vermont slices (Fig. 1) from the north end of the Green Mountains southward into Massachusetts. Graywacke and associated mafic volcanic rocks, important in group 2 slices, are present only in the Pinnacle and Underhill Formations to the north. Volcanic rocks are absent from the autochthonous Hoosac and Cavendish Formations on the east side of the Lincoln and Green Mountain massifs. The varicolored slates and phyllites, some of which contain chloritoid, are not known in the Dalton and autochthonous Hoosac Formations but are present in the allochthonous Hoosac as the pale green, lustrous chloritoid phyllites on Hoosac Mountain. The lack of coarse graywacke and basaltic volcanic rocks in the autochthonous Hoosac Formation on the middle Proterozoic massifs, therefore, indicates that the Rensselaer Graywacke Member and lower parts of the Nassau Formation in the group 2 slices were deposited east of the restored position of the rocks in the Dorset Mountain (3d), Greylock (3g), and Canaan Mountain (3cm) slices of group 3. The rocks in group 2 therefore are more distal and possibly older than those preserved in the Tyson-Hoosac Formations on the east flank of the Berkshire and Green Mountain massifs and mantling the domes in southeastern Vermont. The facies relations between the allochthonous Hoosac (EZht) and the autochthonous Hoosac (EZhd-EZhc) require that the root zone for group 2 slices be located east of the allochthonous Hoosac belt.

In western Massachusetts, the Everett Formation in the Everett slice (2e) resembles the light green to light bluish gray, quartz-laminated schist (OCr, Zen and others, 1983) of the Rowe Schist east of the Whitcomb Summit thrust (Zen, 1972). There are, however, some very important differences between the phyllites in the Taconic slices and those in the Rowe Schist. For example, the Rowe lacks the distinctive graywackes and associated volcanic and volcanoclastic rocks of group 2 slices. It does contain basaltic amphibolites, but many of these are in contact with lenses of serpentinite, which are totally missing from group 2 slices. Other amphibolites are in sharp contact with the pelitic units of the Rowe. The gray to black, fine-grained, slightly rusty weathering, carbonaceous schist interlayered with dark gray to white quartzite typical of the Rowe (Ottawaquechee Formation in Vermont) is not known in group 2, although their lithic correlative, the Hatch Hill Formation, is present in the group 1 Giddings Brook slice. The lithic difference between group 2 rocks and those in the Rowe Schist therefore makes the Rowe an unlikely root zone. The similarities between these two belts of rocks, however, suggest that the Rowe is an eastern, perhaps deeper-water facies of the rocks in the group 2 slices. The foregoing stratigraphic evidence indicates that the root zone for group 2 slices is buried beneath the Whitcomb Summit thrust in western Massachusetts and the Pinney Hollow thrust in Vermont.

Facies relationships of rocks now preserved east of the Hinesburg thrust zone (HT, D4, C4, B4) and north of the exposed middle Proterozoic

rocks of the Lincoln massif (YL) support the foregoing palinspastic interpretation. In the Jerusalem and Underhill slices of northern Vermont (A5, B5, C5), the Pinnacle Formation consists of greenish gray to gray-blue, quartzose graywacke and minor beds of greenish quartz-sericite-chlorite phyllite with quartz veins and laminae much like the rocks of the overlying Underhill Formation (Christman, 1959; Christman and Secor, 1961; Dennis, 1964). Boulder and pebble conglomerates are common in the graywacke. In Quebec, polymict conglomerates associated with the Tibbit Hill Volcanic Member of the Pinnacle contain granitic, gneissic boulders as well as autoclastic fragments of quartzite and siltstone (Booth, 1950). Cross-bedding, graded bedding, and turbidite features have been reported at a number of localities by these geologists. Throughout this region, the basaltic volcanic rocks of the Tibbit Hill are interbedded with the graywacke and consist of massive, fine-grained to coarse-grained, dark green vesicular greenstone containing amygdules. Pillow basalt is known from a few localities. Schistose greenstones, tuff and related pyroclastic rocks, feldspathic greenstone, actinolite porphyroblastic greenstone, and calcareous greenstone, which grade into graywacke, are reported from many localities by these workers. In the Lincoln area (YL, D5), the western belt of Pinnacle (EZp, 4), which was deposited on middle Proterozoic basement, contains carbonate rocks (Forestdale Marble, Fig. 3) but lacks the distinctive rocks of the Tibbit Hill (Tauvers, 1982c; DiPietro, 1983b). It therefore represents a more westerly facies of the Nassau Formation graywackes and associated basalts of group 2 slices.

North of the Lincoln massif in the Jerusalem and Hinesburg slices, the Pinnacle Formation is overlain to the west by the Fairfield Pond Formation (as used by Tauvers, 1982c) which contains uniformly greenish gray phyllite with thin, quartz-rich, sandy and silty laminae. Mafic volcanic rocks are absent. Near the base, red and purple phyllites are locally present. This sequence is lithically similar to phyllites of the Nassau Formation. Although Doll and others (1961) mapped the Fairfield Pond as a member of the Underhill Formation, it has recently been remapped by Tauvers (1982b and 1982c) and DiPietro (1983b) as a separate formation because it is lithically very distinct from the rocks in the Underhill slice. More importantly, both of these geologists showed that the Underhill north and east of the Lincoln massif (YL, D5) is in fault contact with the rocks to the west (EZh-EZt with EZp-EZf, Fig. 2). North of the Lincoln massif, the Jerusalem slice (JS, D4) contains a very distinctive quartz-laminated, greenish gray schist. Middle Proterozoic slivers or mylonitic rocks are found along both thrust faults.

The rocks of the Underhill slice are quite varied, and they consist of (1) silvery gray to greenish gray sericite-chlorite-quartz-magnetite schist containing conspicuous quartz veins and pods much like the rocks of the Jerusalem slice, (2) amphibolite, (3) greenstones, (4) carbonaceous schist and dark quartzite, (5) albitic schist and dark quartzite, (6) albitic schist and gneiss, and (7) minor amounts of graywacke (Christman and Secor, 1961; Thompson, 1974; Eiben, 1976; Aubrey, 1977). The lower part of the Underhill Formation in the Underhill slice is considered a facies of the Pinnacle and Fairfield Pond Formations (EZp-EZf), which was deposited on the middle Proterozoic rocks of the Lincoln massif (4, D5; Doll and others, 1961; DiPietro, 1983a, 1983b). The albitic rocks and the carbonaceous schists and quartzites bear strong resemblances to the Hoosac and the Ottawaquechee Formations, respectively, and thus suggest that the Underhill slice contains a stratigraphic section that extends through the Upper Cambrian Hatch Hill Formation of the Taconic slices, although it may be cut up internally by faults. Serpentinites and associated mafic rocks are notably absent from the Underhill but are present in the Hazens Notch Formation to the east. Recent mapping east of the Lincoln massif has delineated synmetamorphic thrust faults within, and between, these two formations (Stanley and others, 1985). For this reason, the two formations are separated by a thrust fault in Figure 2.



The resemblance of the rocks in the Taconic slices (group 2) to the pre-*Olenellus* rocks on the Hinesburg, Jerusalem, and Underhill slices is striking. Shallow-water quartzite, basaltic lava flows, tuffaceous sedimentary rocks, black terrigenous shale, iron-rich dolostone, turbiditic ensialic graywacke rich in blue quartz and fragments of middle Proterozoic gneiss all suggest a depositional environment proximal to North American sialic crust. The rocks of the Pinnacle, Tibbit Hill, and much of the Underhill are interpreted as being a pre-shelf rift facies punctuated by disconformities and intervals of nondeposition. These rocks represent an eastern facies of the basal clastic rocks of the Dalton and Hoosac Formations and equivalent rocks of lithotectonic units 1 and 4 (Fig. 1 caption).

Group 2 slices are interpreted as being the remains of a regionally extensive slice that was originally continuous with the Jerusalem-Underhill slices (Fig. 2). This original slice was, then, fragmented into a series of imbricate slices to produce the general west-to-east stacking sequence of the Bird Mountain, Rensselaer Plateau, Chatham, and Everett slices. Subsequent erosion has removed these slices north of latitude 43°45'. The root zone for these slices would now be located along the western part of the Jerusalem and Underhill slices in northern Vermont, the Underhill-Pinney Hollow thrust faults in central and southern Vermont, the Whitcomb Summit thrust in western Massachusetts, and Camerons Line in western Connecticut (Pl. 1, Fig. 2).

**Group 1 Slices.** The Giddings Brook slice, including the Sunset Lake slice, is the largest exposed slice in the Taconic allochthon (Figs. 1 and 2). It contains a complete late Proterozoic to Middle Ordovician (zone 12 graptolites) section (Zen, 1967, Fig. 4) and has suffered less metamorphism than have the slices of groups 2 and 3. Fossils preserved in many parts of the Giddings Brook slice provide age correlation with the carbonate platform (Zen, 1967, Pl. 2).

The stratigraphy of the Giddings Brook slice was summarized by Zen (1967, p. 18, Pl. 2), and Potter (1963, 1972, 1979) discussed the relations for the eastern part of the slice near Hoosick Falls. Briefly stated, the varicolored slates and minor graywackes (Bird, 1963; Zen, 1967; Potter, 1972) of the Nassau Formation are overlain by black or dark gray carbonaceous slates or phyllites interlayered with thin quartzite and some limestone of the West Castleton and Hatch Hill Formations. The overlying gray to black siliceous slates and argillites of the Poultney Formation are, in turn, overlain by red, green, and maroon slate of the Indian River Slate and by sooty black slate and interbedded black chert of the Mount Merino Member of the Normanskill Formation (Fig. 6; this figure appears on a folded insert included in this issue). Easterly derived graywackes (Bird, 1963, p. 19) and slates of the Austin Glen Graywacke unconformably overlie the lower part of the Normanskill and are the first rocks to mark the advance of thrust slices from the east. The whole section thus consists largely of different colored slates with distinctive quartzites and graywackes of turbidite origin in the lower and upper parts of the section. Total postdeformational thickness was estimated at 600–1,800 m on the eastern side (Potter, 1972, Fig. 2; 1979, p. 229) and a minimum of 1,000 m for the western side near Granville (Rowley and others, 1979, Fig. 2). The comparatively thin sequence (380 m–840 m) deposited from late Early Cambrian to late Middle Ordovician time (top of Bomoseen Graywacke Member of the Nassau Formation to the base of Pawlet Formations;  $\epsilon$ Zn-Onag, Pl. 1, Fig. 6) suggests a slow rate of deposition compatible with a continental-rise environment (Rowley and others, 1979, p. 206–207) and supports earlier suggestions by Zen (1967, p. 46–47, p. 65) that the Taconic sequence represents a relatively starved section. Shelf-derived carbonate debris flows support a continental-rise environment (Keith and Friedman, 1977).

Several features of the Giddings Brook slice led to the continental slope-rise model for their site of deposition. (1) Carbonate rocks are more abundant in the western slices (group 1) than in the eastern slices (groups

2 and 3) (Zen, 1968, Pl. 1). (2) Carbonate breccias and conglomerates are present in the Hatch Hill Formation and in the upper part of the Poultney Formation (Bird, 1963; Potter, 1972; Rowley and others, 1979). The carbonate breccias, conglomerates, and westerly derived turbidites that thin eastward leave little doubt that the rocks of the Giddings Brook and Sunset Lake slices were deposited seaward of the carbonate platform where they received slump deposits, debris flows, and turbidites. The lack, or paucity, of carbonate and carbonate breccia in the Nassau Formation and the eastward thinning of the graywackes, as seen in group 2 slices, suggest that they were formed prior to development of the carbonate shelf. Rocks of the Giddings Brook slice may have been deposited, in part, on top of this proximal pre-shelf facies as the late Proterozoic to Early Cambrian transgression led to the establishment of deeper-water conditions typical of a slope-rise environment.

As pointed out by Zen (1968, p. 131, Fig. 9-1), the sequence consisting of the Parker Slate and the Rugg Brook and Sweetsburg Formations (as used by Doll and others, 1961) in northern Vermont can be viewed as a transitional facies between shelf rocks and the allochthonous rocks in Taconic groups 1 and 2. Specifically, the Cambrian rocks of the St. Albans synclinorium (A4, B4) have many sedimentary characteristics that are common to the equivalent section of the Giddings Brook slice. Here, the boundary between the Cambrian and Lower Ordovician carbonate bank to the south and the slope deposits trends northwesterly just north of Burlington, Vermont (C4) (Rodgers, 1968, Figs. 10-1 and 10-3). The contact between the two is gradational (Stone and Dennis, 1964, Pl. 1; Dorsey and others, 1983). The section to the north is largely dark gray to black shale with thin beds and laminae of calcareous or dolomitic siltstone. Limestone breccias, limestone or dolostone conglomerates, and sandy dolostones occur throughout the section as discontinuous beds, lenses, or single blocks. The lithologies of some of the blocks can be readily matched to the rocks of the carbonate platform. Blocks of limestone breccia or conglomerate are found within the shale and suggest a compound history wherein limestone and dolostone were broken up and recemented on the carbonate platform and then slumped as larger blocks into the muds seaward of the platform.

Recent work by Carter (1979) and Dorsey and others (1983) along the eastern side of the St. Albans-Hinesburg synclinoria (A4, B4, C4) showed that Nassau-like graywackes of the Pinnacle Formation grade upward into argillaceous quartzites and clean quartzites of the Cheshire Quartzite which, in turn, are conformably overlain by the Dunham Dolomite. These carbonate rocks grade into the Cambrian and Lower Ordovician shales of the St. Albans region (Gregory, 1982; Dorsey and others, 1983). Although minor faults occur along the Cheshire-Dunham contact, the Hinesburg thrust is mapped to the east of the western belt of Pinnacle ( $\epsilon$ Zp, A4, B4). The St. Albans section therefore contains not only the slope deposits of the post-Dunham section, but, more importantly, it preserves part of the older rift clastic rocks that were overlapped by the eastern edge of the lower part of the carbonate platform.

The St. Albans section, then, provides a clue to the depositional position of the rocks of the Giddings Brook slice relative to the rocks of the other Taconic slices and to the partially preserved section east of the Hinesburg thrust in northern Vermont. Although the Giddings Brook slice lacks carbonate platform deposits, it does contain many of the shelf-derived turbidite deposits that are represented in the St. Albans section. For this reason, the rocks in the Giddings Brook slice must have been deposited in an equivalent position or slightly east of the section represented by the St. Albans sequence of rocks. Moreover, the bulk of the Giddings Brook slice represents material deposited on top of the older rift clastics represented by the Pinnacle and Fairfield Pond Formations directly east of St. Albans ( $\epsilon$ Zp, A4, B4) and in the Lincoln area (D4). Zen (1967) originally proposed that rocks of the Giddings Brook slice do not



interfinger extensively with the oldest part of the Rensselaer Plateau and Chatham slices and could, conceivably, have been deposited on top of the rift clastics. The rocks in the St. Albans region bear this out. If the coarse graywackes and associated volcanic rocks of group 2 slices and the equivalents of the lower part of the Nassau Formation do represent a pre-shelf rift assemblage, then slope-rise rocks of the Giddings Brook have few, if any, contemporaneous counterparts in group 2 slices. This stratigraphic interpretation was one of Zen's (1967) reasons for proposing soft-rock emplacement of the Giddings Brook slice by a process of unpeeling or diverticulation driven by gravity sliding. In this interpretation, the Giddings Brook slices would not require a conventional root zone.

In the present interpretation, the Giddings Brook slice was emplaced as a result of horizontal compression from a root zone directly east of the Proterozoic massifs. Although there is, at present, no recognized section in the Underhill slice that is convincingly similar to the section in the Giddings Brook slice, there are similarities and differences which indicate that part of the Underhill Formation may be an eastern facies of the Giddings Brook. For example, rocks lithically similar to the Hatch Hill Formation of the Giddings Brook slice are largely undifferentiated in the Underhill slice except for isolated belts of carbonaceous schist and interlayered dark quartzite and thin carbonaceous limestone (€Zuc, as used by Doll and others, 1961, but recognized as the Ottauquechee Formation by Christman, 1959). To the east in the Rowe-Hawley slices (Fig. 2), carbonaceous schist is more abundant, and calcareous rocks are minor in the Ottauquechee Formation. Metamorphosed mafic volcanic rocks and associated volcanogenic rocks are abundant, whereas they are scarce in the equivalent rocks in the Underhill slice and absent in the Hatch Hill Formation in the Giddings Brook slice. This carbonaceous shale-quartzite sequence must then represent a west-to-east transition from a quartz-carbonate-rich section to a shale-volcanic-rich section and suggests that the metamorphosed rocks of the Underhill slice are possible facies equivalents to, at least, the Cambrian and older parts of the Giddings Brook slice.

Given these findings, the root zone for the northern extension of the Giddings Brook slice in northern Vermont is located east of the Hinesburg thrust, by virtue of the occurrence of the basal part of the carbonate platform, (Cheshire Quartzite and Dunham Dolomite) as isolated patches on the Hinesburg slice north of latitude 44°30' (O€p, B4). The eastern boundary of the root zone must be located west of the Hazens Notch slice because this slice contains serpentinites. The presence of the medium-high-pressure metamorphic series preserved in many of the mafic volcanic rocks of the Underhill slice (Laird and Albee, 1981a and 1981b) further argues that the Giddings Brook slice roots along the western edge of the Underhill slice. We suggest that much of the Upper Cambrian and Ordovician section of the Giddings Brook slices has been overridden by later westward movement of the Underhill and more easterly situated slices containing medium- to medium-high-pressure metamorphic rocks. Part of the section may exist as presently unrecognized slices with the Underhill slice. The root zone for group 1 slices therefore is the same as it is for group 2. It follows the Underhill-Pinney Hollow-Whitcomb Summit-Cameron's Line thrust faults southward and reappears at the surface around the Chester and Athens domes in southeastern Vermont.

#### **Serpentinites, Ultramafic Rocks, and Gabbros East of the Proterozoic Massifs**

Serpentinite and other ultramafic rocks in the eugeoclinal rocks of eastern New England have been thought to be Ordovician intrusive bodies in the Cambrian and Ordovician section (Cady and others, 1963; Doll and others, 1961; Chidester and others, 1967; Hatch and Stanley, 1976; Chidester, 1978, for example). They are elongate bodies parallel to the general

trend of lithic units and range in length from a metre to several kilometres, with an average length of <200 m. Their width rarely exceeds 100 m. Their contacts with the surrounding metasedimentary rocks are highly sheared and lack obvious contact metamorphic aureoles. Slickensided fault surfaces and phacoidal shear zones are common within the serpentinites. Many of the ultramafic rocks consist of serpentine, talc, and minor amounts of magnesite, tremolite, magnetite, and chromite. The larger ultramafic bodies commonly have an inner core of dunite and/or peridotite, particularly in Vermont where the metamorphic grade is lower than it is to the south. Gabbro is found in some of the bodies. Internal contacts in these plutonic rocks are discordant with the surrounding country rocks. Detailed mapping near North Troy, Vermont (A7), has shown that several bodies consist of serpentinitized dunite, peridotite, and gabbro in a highly sheared matrix of serpentinite without any slivers of the enclosing sedimentary rocks (Winner, 1981; Stanley and Roy, 1982; Stanley and others, 1984). Throughout the belt, many of the ultramafic rocks are thoroughly altered to steatite or talc-carbonate rocks.

Serpentinized ultramafic rocks are found in the Hazens Notch, Ottauquechee, Stowe, and Missisquoi Formations of Vermont (as shown by Doll and others, 1961) and are significantly absent from the Pinnacle and Underhill Formations. In Massachusetts and western Connecticut, they are restricted to the Rowe Schist, western part of the Moretown Formation, Member C of the Cobble Mountain Formation, and the Sweetheart Mountain Member of the Collinsville Formation (Stanley, 1964; Zen and others, 1983). They are absent from the local Silurian and Devonian rocks. A few bodies are found in the middle Proterozoic of the Chester and Athens domes in southeastern Vermont (Doll and others, 1961) and the Berkshire massif in western Massachusetts (Zen and others, 1983) where the evidence indicates that they are intrusive plugs.

Recent mapping in northern Vermont has demonstrated that all of the ultramafic rocks occur along fault zones as delineated by 1:10,000 mapping in the surrounding metasedimentary rocks (Stanley and Roy, 1982; Stanley and others, 1984). The stratigraphic interval of the Ottauquechee and Stowe and the western part of the Moretown contain numerous fault slices with distinctive, but gradational, depositional sequences interpreted as being of continental-margin and ocean-floor affinities. A significant conclusion of this work is the realization that many, if not all, of the formations of the pre-Silurian eugeoclinal zone in eastern Vermont are lithotectonic units of regional scope that contain fault slices from neighboring units.

It has been suggested that the same pattern holds true for western Massachusetts where Acadian deformation and metamorphism are more severe than they are in northern Vermont (Stanley and others, 1984; Stanley and Hatch, in press). In Massachusetts, 1:24,000-scale mapping has shown that >50% of the ultramafic bodies are found along lithic contacts in the Rowe Schist, in the western part of the Moretown, and in the upper part of the Cobble Mountain Formation. The lack of lithic continuity in the Rowe Schist (Zen and others, 1983) suggests that distribution of the ultramafic rocks is fault-controlled in a manner similar to that in northern Vermont (Stanley and others, 1984). Although the remaining 50% occur within boundaries of units in western Massachusetts, we believe that these may also mark faults.

Most of the ultramafic rocks, therefore, along the east side of the Proterozoic massifs are here interpreted as fault slivers rather than orostromal deposits or intrusive plugs. Correlation with ultramafic rocks in the fragmented ophiolitic rocks of Quebec (Laurent 1975, 1977; St. Julien and others, 1976) and geochemical analysis in northern Vermont (Doclan and others, 1982) suggest that these rocks are fragments of ocean crust that were imbricated with the continental-rise and ocean-floor sediments during the early development of the accretionary wedge. Continued move-



ment on some of these older faults and the development of younger faults in the older slices of the accretionary wedge extended the tectonic emplacement of ultramafic and related rocks into sediments near the continental margin. Mélanges of serpentinite with older ultramafic and mafic fragments, as found in North Troy, Vermont, may represent older deformed material from a ridge or transform fault.

#### **Sediment Source and the Paleoenvironment of the Moretown Formation and Lithically Equivalent Rocks**

The Moretown Formation and similar rocks form a continuous belt extending from the Canadian border southward to Long Island Sound. They consist of distinctive quartz-plagioclase granofels with thin, mica-rich laminae, light colored pelitic rocks, mica-bearing quartzites, and mappable belts of carbonaceous pelitic rocks and thin cherts. Mafic volcanic and cogenetic rocks are also present but constitute <30% of the formation. The Moretown and equivalent rocks are bordered to the west by light green, green to bluish green pelite containing thin quartz veins and lamellae, greenstone, black carbonaceous pelite and interbedded black, gray, or white quartzite of the Rowe Schist (Pinney Hollow, Ottauquechee, and Stowe Formations of Vermont). To the east, the Moretown is bordered by black carbonaceous pelite and volcanic rocks of the Hawley belt (Fig. 2).

The rocks of the Moretown Formation originated from two quite different sources. Metamorphosed bedded and nonbedded greenstones, pillowed greenstones, amphibolites, minor rhyolitic tuff, and minor volcanogenic material are found throughout the Moretown Formation (Cady, 1956; Cady and others, 1963; König, 1961; Chang and others, 1965; Skehan, 1961, for example) and indicate a source to the east in the island-arc setting of the Bronson Hill slice (Fig. 2) and Ascot-Weedon slice (Oa, A7). Some of the bedded greenstones contain large, angular quartz grains, suggesting contamination of arc volcanics by quartz sands from another source (König, 1961, p. 24–25). Minor limestone is present in the volcanic rocks in some places (Cady, 1956; König and Dennis, 1964), which indicates deposition above the carbonate compensation depth. Dikes and sills of intermediate to mafic composition are found in many places in the Moretown Formation (Cady and others, 1963; König, 1961; Osberg and others, 1971). In southern Vermont, the Barnard Volcanic Member of the Missisquoi Formation (of Doll and others, 1961) contains light and dark volcanic rocks (Chang and others, 1965) and is interpreted as being an arc-related volcanic sequence below the black pelitic rocks and volcanic rocks of the Cram Hill and Hawley Formations.

The quartz-rich granofels, interlayered pelites, and quartzites, which make up a large percent of the Moretown Formation, appear to originate from a relatively nearby source rich in quartz-bearing rocks and shales. In northern Vermont, where the metamorphic grade is low, the fabric consists of “fine angular to rounded clastic grains that are unsorted as to size and are set in a finer grained matrix” (Cady and others, 1963, p. 27). The quartz grains are more rounded and sorted in the quartzites. The ratio of albite to quartz is 1:3, and potash feldspar is rare. These same rocks are reported to the south in Vermont and Massachusetts despite higher metamorphic grades. These rocks were not derived from a volcanic arc because they are rich in quartz and relatively poor in feldspar (Schwab, 1981, Table 1). A western source from the North American continent also seems unlikely because the equivalent rocks of the Poultney Formation and the lower part of the Normanskill Formation in the Taconic allochthons are shales and thin cherts (Ratcliffe and Hatch, 1979, Fig. 3) which were deposited between the North American continent and the depositional site of the Moretown Formation.

A possible source area is suggested by conglomerates of the Umbrella Hill Formation which is in depositional contact with the base of the

Moretown Formation in northern Vermont (Cady and others, 1963; Badger, 1979). Badger recognized three facies in the conglomerates. The first is a coarse quartz-pebble, nonbedded framework conglomerate (20%) with rounded white pebbles of vein quartz, poorly sorted recrystallized sandstone, and gray or black quartz clasts which are similar to the quartzites in the Ottauquechee Formation to the west. Feldspathic clasts are rare. The second is a tan to gray Moretown phyllite with rounded to subangular quartz clasts and phyllitic breccia (40%). The phyllite clasts are white, red, and black and include fragments of the Moretown, chloritoid-bearing phyllite. Volcanic fragments are rare. The third facies consists of beds of Moretown phyllite (40%) interlayered with facies 1 and 2. Badger (1979) suggested a debris-flow mechanism for the Umbrella Hill based principally on the bimodal association of rounded quartz pebbles and angular phyllitic clasts. Although he suggested a source to the northeast in northern Maine and southern Quebec (Badger, 1977), Doolan and others (1982) concluded that a western source is likely. Veins and pods of quartz are abundant in the Stowe and the Underhill Formation directly west of the Umbrella Hill. White quartzites are present in the Hazens Notch Formation and parts of the Ottauquechee, although black quartzite is more characteristic of the Ottauquechee. On the basis of this evidence, we suggest that the Umbrella Hill and the quartz-rich rocks of the Moretown Formation originated from emerged parts of an accretionary wedge made up of older rocks now exposed in the Rowe Schist, Underhill, and equivalent formations. The wedge was situated between the North American continent and the Bronson Hill arc complex. The Moretown Formation and equivalent rocks were, therefore, deposited in a broad forearc basin that received sediment from islands or high submarine ridges of the accretionary wedge to the west and the volcanic arc to the east. This basin may have been wide, and the relief in the source areas, low during most of Moretown Formation time because the rocks are well bedded, fairly well sorted, and olistostromal deposits have not been recognized, except for the Umbrella Hill Formation in northern Vermont.

#### **Middle Ordovician Sequence within All Six Lithotectonic Units**

Black, carbonaceous shales and graywackes of Middle Ordovician age form a sedimentary cap on each of the major lithotectonic assemblages. The change to Middle Ordovician black shale deposition and formation of unconformities herald important tectonic changes that are associated with the early stages of collision leading to the events of the classical Taconian orogeny.

In the miogeoclinal belt, a principal Middle Ordovician unconformity is widely recognized beneath the Snake Hill Formation in southern New York (member A of the Manhattan Schist of Hall, 1976, in Westchester, New York), beneath the Walloomsac in eastern New York, western Massachusetts, and southern Vermont, and beneath the Ira Formation and Hortonville Slate in the northern Taconic area. A thin, basal limestone, which is locally conglomeratic, rests on miogeoclinal shelf rocks throughout the belt from Vermont to New Jersey. This basal limestone (Balmville Limestone of Fisher, 1962, Walloomsac Formation limestone, Whipple Marble Member of the Ira Formation) contains zone 8 conodonts (Sweet and others, 1970) and other late Middle Ordovician fossils. Graptolites within the overlying Walloomsac include *Diplograptus foliaceus* [sic] (Prindle and Knopf, 1932, p. 274). *Climacograptus bicornis* (zone 12 of Berry, 1960) was found in the Walloomsac, by Potter (1972), beneath wildflysch conglomerate on Whipstock Hill at the east margin of the Giddings Brook slice (I4). Fisher and Warthin (1976, p. B6–8) reported graptolites identified by John Riva as indicative of his *Diplograptus multidentis* zone in Snake Hill Formation immediately overlying Balmville Limestone in Dutchess County, New York. These and other data (Zen,



1967, p. 43) indicate that the Balmville, Snake Hill, Walloomsac sequence is medial Mohawkian (late Middle Ordovician) and is equivalent to subtidal limestones of the Trenton Group of the Black River area of New York (Fisher, 1977, Pl. 2).

Eastward toward the middle Proterozoic massifs, the hiatus at the unconformity increases, and basal Walloomsac or Ira rests locally on Proterozoic basement (Doll and others, 1961; Norton, 1967; Thompson, 1967; Ratcliffe, 1969b; Ratcliffe, *in* Zen and others, 1983). Facies changes are also apparent as the basal marble becomes thicker eastward and becomes more impure, containing abundant dolomite and quartz sandstones, thinly laminated quartzite, and locally feldspathic conglomerate. Zen (1961) and Ratcliffe (1969b) suggested that this facies change reflects a greater degree of erosion onto the lower dolomitic units of the shelf sequence and a contribution of detritus from areas of exposed Proterozoic basement east of the Walloomsac basin. In southeastern New York and adjacent New Jersey, Fisher and Warthin (1976) noted evidence for pre-Balmville karst features, clearly indicating subaerial exposures of the shelf. Savoy and others (1981) found that the Jacksonburg and Balmville Limestones contain progressively younger conodonts northeastward from Pennsylvania to Newburg, New York, and that consistently older rocks of the Beekmantown Group are overstepped northeastward. From these relationships and from study of the range of cobbles of shelf rocks in the basal conglomeratic deposits, Savoy and others concluded that regional tilting was sufficient to expose 300 to 430 m of shelf rocks to erosion in southern New York prior to Balmville transgression. Pre-Middle Ordovician block faulting has been invoked in Vermont, New York, and Massachusetts (Zen, 1967, 1968, p. 134), in part to account for the very rapid local variation in age of shelf rocks found beneath the black shale sequence. In western Vermont, igneous activity (Baker Brook Volcanics of Thompson, 1967) may have accompanied block faulting.

The younger age and structural assignment of the rocks of the Austin Glen Graywacke found above the Snake Hill Formation is more controversial. The upper part of this section contains easterly derived turbidites (Bird, 1963, 1969) that onlap westward into the Hudson Valley (I1) (Zen, 1967, p. 29) and are found in both Taconic allochthons (Pawlet and Austin Glen of group 1 slices) and the autochthon (Austin Glen and upper members of the Normanskill and Snake Hill). Nevertheless, Fisher (1977) considered the Austin Glen as older than some of the Snake Hill and included all Austin Glen as allochthonous slices (1Ls, H2, I2, K1, L1). In the autochthon, however, wildfisch deposits are present along the western and eastern edges of the Giddings Brook slice (Zen, 1967; Potter, 1963, 1972). The deposits west of the Giddings Brook slice contain clasts of the allochthon in a matrix of black shale containing zone 13 graptolites (Berry, 1960) and represent material eroded from the advancing Taconic slices and deposited in the submarine muds proximal to the thrust (Zen, 1967, p. 35–40).

The Middle Ordovician section is found only in group 1 slices where it is largely represented by the Normanskill Formation. The red and green slates of the Indian River Slate and the sooty black slate and chert of the Mount Merino Member in the lower part of the Normanskill are distinctly different from the transgressive black shale of the autochthon. Graywackes and turbidites in the upper part of the Middle Ordovician section (Pawlet and Austin Glen) may be present in both the allochthonous and autochthonous sequences. These black slates and cherts of the allochthon, however, are lithically similar to carbonaceous schist and metamorphosed chert in the Hawley Formation in the Rowe-Hawley slices (Fig. 2). A Middle Ordovician unconformity equivalent to that in the autochthon is not recognized in the allochthon. A younger unconformity is recognized at the base of the Pawlet Formation, but it represents only a relatively short period of time (Zen, 1967, p. 29).

The Middle Ordovician section in the Eastern Vermont and Rowe-Hawley slices (Fig. 1) and the Bronson Hill slice are remarkably similar. Both contain carbonaceous, sulphidic schist and interlayered mafic volcanic rocks (Hawley and Partridge Formations). Felsic volcanic rocks are subordinate in the eastern Vermont slices but are plentiful in the Bronson Hill where the Ammonoosuc Volcanics underlies the Partridge Formation.

The Sweetheart Mountain Member and an unnamed hornblende gneiss member of the Collinsville Formation (Oco, L5, M5, N5) in the domes of western Connecticut are part of the Middle Ordovician sequence. The feldspathic gneiss and schist of the Cobble Mountain Formation (Oc and Occ, L5) of southern Massachusetts and northern Connecticut are lateral facies of the Hawley Formation. Although no regional Middle Ordovician unconformity has been recognized in the Eastern Vermont slices, Robinson (Zen and others, 1983) proposed one just beneath the Partridge Formation which locally cuts out the Ammonoosuc and rests on the Monson Gneiss. Stanley (Zen and others, 1983) proposed an unconformity at the base of the Hawley-Cobble Mountain sequence on the basis of regional relations (Stanley and Hatch, *in press*).

Despite the fact that the Middle Ordovician black shales and related rocks at higher grades from the different lithotectonic units bear a resemblance to each other, the differences as seen in the presence or absence of carbonate rocks, volcanics, and volcanogenic rocks indicate quite different settings. For example, the black shale in the Eastern Vermont slices probably formed in a forearc basin, whereas the black shales in lithotectonic unit 2 to the west formed on top of the carbonate platform in a gradually deepening basin between the North American craton and the western side of the slowly rising accretionary wedge. There is no evidence that these basins were once continuous and, indeed, they may well have been separated during much of the time of their formation. Because age control is available only for the western basin, the possibility remains that lithically equivalent rocks formed at slightly different times to the east in the forearc region in, and around, the Bronson Hill arc complex.

#### **The Relation of the Cobble Mountain Formation to the Middle Ordovician Black Shale and the Core Gneiss of the Domes of Unit 6 (Fig. 1)**

The Cobble Mountain Formation (Oc, L5) forms the southern continuation of the Hawley Formation. Lithically equivalent rocks are found in the Goshen (GD, K5), Granville (GrD, L5), Collinsville (CD, M5), Bristol (BD, N5), and Waterbury (WD, N4) domes (Stanley and others, 1980). The Cobble Mountain Formation consists of feldspathic gneiss, interlayered feldspathic schist, granofels, aluminous schist, amphibolite, minor felsic gneiss, and lenses of serpentinite. This formation is pertinent to the plate-tectonic reconstruction of southern New England because it contains serpentinite and rocks that are lithically distinct from the black shales and volcanics of the Hawley with which it interfingers (L5).

The lower two members of the Cobble Mountain Formation (Oc, L5) are in depositional sequence and consist of interbedded granular rocks and pelites that resemble metamorphosed turbidites. The granofels of the lowest member is finer grained, less feldspathic, and occurs in thinner beds compared to the next younger member where the gneisses and granofels are coarser, more feldspathic, and the beds are thicker. Amphibolite is more abundant in the younger unit. Graded beds are common in the older unit. The lowest member (member A) is considered a distal deposit of the younger member (member B) (Stanley and others, 1980, 1982). The abundance of feldspar and the presence of mafic and felsic volcanic rocks indicate a source to the east in the Bronson Hill arc complex. Both of these members are considered Middle Ordovician because member A grades laterally into the Hawley. Member C of the Cobble Mountain Formation



(Occ, L5) is in fault contact with member B along the Winchell Mountain thrust (WMT). It is lithically important because it contains lenses of kyanite-sillimanite schist and schistose amphibolite that are very similar to the schist and amphibolite in the Rowe Schist (Hatch and Stanley, 1973, p. 12; Knapp, 1977, 1978). Furthermore, it contains lenses of serpentinite which are absent from members A and B. Because the bulk of member C is feldspathic schist, which is identical to the schist in member B, it is interpreted to be a distal (westward) facies of B. Stanley and others (1982) suggested an olistostromal origin for the Rowe-like fragments and serpentinites in member C, the source being an emerged part of the accretionary wedge to the west. They further suggested that member C was thrust eastward over member B.

Rocks (Ococ) equivalent to member C are found in the domes directly east of the Cobble Mountain Formation. Here, they overlie amphibolites and gneisses that are correlated with the Ammonoosuc Volcanics of the Bronson Hill anticlinorium (Zen and others, 1983). Member C, therefore, rests upon two quite different sections: the amphibolites and gneisses of the domes and the feldspathic turbidites of the Cobble Mountain Formation directly to the west. On the basis of the lithic similarities between member C and the schists of member B, a Middle Ordovician age is suggested for member C (Stanley and others, 1982).

#### **Correlation of the Core Rocks of the Domes of Western Massachusetts and Western Connecticut with the Ammonoosuc Volcanics and Older Gneisses of the Bronson Hill Slice**

The rocks in the domes in Massachusetts and Connecticut west of the Mesozoic Basin consist of gneiss, granofels, amphibolite, subordinate schist, and minor manganiferous chert grouped in three formations: the Waterbury Gneiss, the Taine Mountain Formation (Om in the Bristol dome, Fig. 2), and the Collinsville Formation (Oco). The first two, which are structurally lower, appear only in the Waterbury (WD) and Bristol (BD) domes, whereas the Collinsville is present in all of the domes and has the closest resemblance of all three to the gneisses and volcanics in the Bronson Hill anticlinorium.

The Bristol Member of the Collinsville Formation is largely absent west of the domes in the Hawley Formation and equivalent rocks with which it has been correlated (Hatch and Stanley, 1973). In contrast, the upper feldspathic schist of the Collinsville (Ococ) is present to the west as member C of the Cobble Mountain Formation which overlies the Moretown (Om). The Bristol Member, therefore, forms a westward-thinning wedge of gneiss and amphibolite that is exposed in the domes but is largely absent to the west, where its place is taken by schist and granofels of the lower part of the Cobble Mountain Formation.

The significance of these relations is that the Bristol Member and equivalent rocks in the domes of Figure 2 closely resemble the Ammonoosuc Volcanics and the core gneisses in the Bronson Hill anticlinorium directly east of the Mesozoic basin. These plagioclase-quartz gneisses (Monson Gneiss and Fourmile Gneiss of Ashenden, 1973) and associated amphibolites are considered to be of mixed volcanic and plutonic origin (Robinson and Hall, 1980, p. 74). They also overlie gneisses and metasedimentary rocks of late Proterozoic age exposed in the Pelham dome (Zen and others, 1983). The similarity between the plagioclase gneiss of the Collinsville Formation and the plutonic gneisses of the Bronson Hill slice is particularly significant because similar plutonic rocks are absent from the underlying Moretown Formation (Om, Taine Mountain Formation) in the Bristol and Waterbury domes. The basal contact of the Collinsville Formation therefore is interpreted as a thrust, the Bristol thrust, of regional extent, and it is shown as such in Figure 2 and Plate 2. (Plate 2 appears on a folded insert included in this issue.) Recent mapping by Dietsch (1983) in the Waterbury dome shows that mylonites occur along this boundary.

## **STRUCTURAL CONSIDERATIONS**

Westward-directed thrust slices dominate the Taconian geology of western New England (Pl. 1). Eastward, toward the Bronson Hill anticlinorium, these thrust faults are progressively deformed by regional folds, westward-directed thrusts, and metamorphism of the Acadian orogeny, which severely overprints the earlier Taconian fabric. The younger events mask the older fabrics east of the Green Mountain massif but diminish along this belt to the north, revealing older Taconian fabrics.

### **Development of the Taconic Slices (Unit 2; Fig. 1)**

The Taconic allochthon and the Berkshire massif consist of imbricate slices, those of which are situated to the east rest on the trailing edge of the next most westerly slice (Fig. 2). The west-to-east stacking order for the Taconic slices was originally proposed by Zen (1967) and has since been confirmed and elaborated on by Potter (1972, 1979), Ratcliffe (1969b, 1974a, 1974b, 1974c, 1979a, 1984a), and Harwood (1975). A fundamental disagreement, however, exists on the mechanism and order of assembly of the slices. Ratcliffe (1979a) proposed that the easternmost slices (group 3) are youngest and have traveled the least distance, compared to the western slices (group 1). This sequence in the Taconics is markedly different from the "piggyback" sequence described for the Quebec and Newfoundland Appalachians where the easternmost, higher slices have traveled farther and were emplaced earlier than the western slices (Williams, 1975, for example). The critical evidence is best preserved in the Quebec Appalachians where St. Julien and Hubert (1975, p. 352) described wildflysch deposits that are younger, based on fossils, for each successive premetamorphic thrust slice in the westward progression. As each of the major thrust slices broke the surface, erosion along the leading edge produced olistostromal deposits with fossils preserved in the matrix between the eroded clasts. Significantly, carbonate slivers, which are abundant in group 2 slices in the Taconics, have not been reported in the Quebec sequence. In a similar sequence in western Newfoundland, an ophiolite forms the highest slice and is systematically underlain by thrust-bounded lithic packages that were originally deposited closer and closer to the foreland (Williams, 1975). The westernmost, lowest thrust slice therefore was the last to move, traveled the least distance, and brought on its back a stack of preassembled thrust slices capped by the ophiolite.

This piggyback model has been reported for many mountain belts of the world and appears to be an important process in the development of the accretionary wedge along subduction zones. Foreland fold and thrust belts, as in the Canadian Rockies and the Appalachians, have major thrusts that developed progressively from the hinterland to the foreland (Bally and others, 1966; Elliot, 1976; Boyer and Elliot, 1982). Definitive evidence for the sequence is the migration and younging of forethrust depositional basins onto the foreland. In the accretionary-wedge model, ocean sediments and crust are sheared off the oceanic plate along a subduction zone and are underplated beneath older thrust slices (Karig and Sharman, 1975). Most of the active deformation is confined to the lower part of the trenches, deformation decreasing upward through the accretionary wedge. The accretionary mechanism is supported by a number of geophysical and sedimentological studies throughout the Pacific (Seely and others, 1974; Kulm and Fowler, 1974; Hamilton, 1979; Moore and others, 1980; Moore and Karig, 1980). It may well have operated during the early stages of such mountain belts as the Taconide zone of the Appalachians, as suggested by Rowley and others (1979, Fig. 4) and Rowley and Kidd (1981, Fig. 4).

Despite the fact that the piggyback, or accretionary sequence, is common, there is mounting field and theoretical evidence that the actual

sequence is complicated by the reactivation of major thrusts to the rear of the accretionary wedge or foreland of an active mountain belt (Burchfiel and others, 1974, p. 1016–1017; Burchfiel, 1982; Davis and others, 1983). Major displacement of these thrusts overlaps or buries thrusts along the leading edge of the belt. A modern example of this process is seen in Taiwan where the steady-state volume of the mountain chain is balanced by the rate of convergence between the West Philippine and Asian plates (Suppe, 1981). In order to maintain the constant cross-sectional wedge shape, faulting must continue in the hinterland, either as reactivated older thrusts or as backthrusts or backfolds (Page and Suppe, 1981; Stanley and others, 1981). In fact, several major thrusts in the hinterland are planar and, therefore, are capable of renewed movement (Suppe, 1980). Continued deformation, in the hinterland, of a mountain belt is actually an inherent requirement for thin-skinned tectonics, whether one favors the strictly gravitational spreading process (Elliot, 1976) or a combined process of horizontal compression and gravitational spreading (Chapple, 1978; Davis and others, 1983). We take this view for western New England where the evidence supports the more complicated emplacement sequence rather than a simple westward-younging piggyback sequence suggested by Rowley and Kidd (1981).

The evidence in western New England that is critical to the west-to-east (toward the hinterland) younging sequence for the Taconic slices is as follows.

**Carbonate Slivers (Fig. 2, X Symbol).** Zen and Ratcliffe (1966) described carbonate slivers at a number of localities along the thrust surface of the Everett slice (2e, K3, L3, M2). Zen (1967, p. 31–34, Fig. 8) reported similar occurrences to the north in Vermont where they are found between the Giddings Brook (1gb), Bird Mountain (2bm), and Dorset Mountain (3d) slices. Subsequent work by Potter (1972, 1979) and Ratcliffe (1974a, 1974b, 1974c, 1979a) in the central and southern Taconics has revealed carbonate slivers along all of the thrust surfaces in group 2 slices, regardless of whether or not the upper slice is resting on the Middle Ordovician black shale of the autochthon or on a slice of one of the westerly situated allochthons. The rocks of the slivers can be matched to all of the units of the Stockbridge Formation down to, but not including, the Cheshire Quartzite (Ratcliffe, 1979a, p. 392). Fragments of the Walloomsac Formation and slivers from adjacent slices are also present. In many of these slivers, the bedding was recumbently folded prior to, or possibly during, their incorporation into the fault zone (Ratcliffe, 1979a, p. 407–413, Figs. 8 and 9). Subsequent deformation of the fault zones has refolded the older fabric of the slivers and produced a cleavage recognized as the regional Taconic cleavage (Zen and Ratcliffe, 1966, location 3 on Figs. 2 and 4). The carbonate slivers, therefore, require that the eastern slices be younger than those situated directly to their west. If the allochthons had been assembled by the piggyback mechanism, only the basal, more westerly situated thrust of the total Taconic allochthon would contain slivers of the platform. Carbonate slivers would be absent from surfaces higher in the stack because these slices developed earlier and were isolated from the platform by the underplating process. The carbonate slivers, then, preclude a simple east-to-west accretionary sequence for assembling the Taconic slices of groups 2 and 3.

The carbonate slivers along the leading edge of the Giddings Brook slice, for example, at Bald Mountain, are not clearly of the same origin as are the carbonate slivers found at the soles of group 2 slices. For a recent discussion of this, see Rodgers (1982b) and Bosworth and Rowley (1984b). The precise origin of these carbonate slivers is uncertain because late faults cut the slaty cleavage along the frontal zone of the Taconic allochthons (Bosworth and Rowley, 1984b, and labeled Champlain thrust on Pl. 1, Fig. 2), and some of the carbonate blocks of the Bald Mountain area may, indeed, be olistoliths. It is most significant that carbonate slivers

are never found on the trailing edge of the Giddings Brook slice, whereas they are widely present along the sole and trailing edges of group 2 slices. In addition, at many places, group 2 slices rest directly on the platform sequence and, therefore, had to have been emplaced onto the shelf.

In view of these considerations, two distinctly different fault-zone or fault-related structures are found: (1) the earlier wildflysch or olistostromal deposits containing zone 13 graptolites and coeval fault-zone mélange which developed when the first Taconic slices (now represented by group 1 slices) broke the Earth's surface and overrode its own debris; and (2) the younger, folded slivers of the carbonate-shale platform that occur along the soles of group 2 hardrock slices. These formed when the eastern part of the Giddings Brook slice was imbricated (Pl. 2, sections 7 and 6). The regional Taconic cleavage parallels the axial planes of folds that deform the fault zones. A sample of fine-grained muscovite from this cleavage in the purple phyllite of the Chatham slice has yielded a K-Ar age of  $442 \pm 16$  m.y. (Ratcliffe, 1979a, p. 406). We further suggest that late Taconian movement along the Taconic frontal thrust faults reworked the older mélanges and produced the fabrics reported by Bosworth and Rowley (1984b).

**Group 3 Metamorphic Fabrics.** The fault zones of group 3 slices differ from those of the other two groups in that they contain small-scale reclined folds that deform an older schistosity with new chlorite drawn out parallel to their axes. Ratcliffe (1979a, p. 395–398) showed that the highly deformed Greylock slice contains older, pre-thrust, reclined to recumbent mappable folds that, in turn, are refolded by syn-thrust recumbent folds of Taconic age. The oldest schistosity deformed in the fault zone is axial planar to the oldest mappable folds within the slice, and the younger schistosity is parallel to the thrust surface and the axial surfaces of the younger recumbent folds. It is important to note that structures identical to these are found in the Hoosac Formation between the Middlefield and Whitcomb Summit thrusts in the proposed root zone for group 3 slices.

The main point here is that the slices of group 3 contain older, pre-thrust, metamorphic fold structures that are transected and deformed in the fault zone. These characteristics are notably absent in group 1 and group 2 slices to the west. Although pre-thrust, pre-slaty cleavage folcs are suggested by the map pattern in some of these western slices in the northern Taconics (Zen, 1961) and in the Chatham slices in the central and southern Taconics (Ratcliffe and others, 1975), the sequence generally is right-side up. Most importantly, the regional Taconian cleavage cuts across the stacked western slices and is associated with their subsequent folding into the allochthon. This foliation is considered to be coeval with the older deformed schistosity in the Greylock slice. On the basis of these observations, group 3 slices were emplaced after groups 1 and 2.

In conclusion, the eastward-younging sequence originally proposed by Zen (1967) is definitively supported by slivers of the carbonate platform along the basal thrust fault of each slice, the presence of pre-thrust folds, and the existence of metamorphic emplacement fabrics. The New England region, more importantly, differs from Quebec and Newfoundland, where the stacking order youngs toward the foreland. Furthermore, unlike the Canadian Appalachians, the North American crust in western New England has been transported westward in thrust slices of different thickness. This is best displayed in the Berkshire massif where Ratcliffe (1975, 1979a), Ratcliffe and Harwood (1975), Harwood (1975), and Norton (1975) conclusively demonstrated an internal stack of ten slices that are allochthonous or parautochthonous with respect to the carbonate platform. The Berkshire massif and its sedimentary cover (Dalton Formation and autochthonous Hoosac Formation) were emplaced on the carbonate platform and group 3 slices along synmetamorphic imbricate thrust faults in Late Ordovician time (Pl. 2, section 1; Ratcliffe and Hatch, 1979, p. 183; Sutter and others, 1985). We propose a similar tectonic style for



the basement massifs in southeastern Vermont, the Green Mountain and Lincoln massifs, the Housatonic massif, and the Hudson Highlands (Pl. 1; Ratcliffe, 1982). This interpretation is supported by seismic-reflection information across southern Vermont (Ando and others, 1983, 1984).

We propose that the differences between western New England and the Canadian Appalachians are primarily due to the irregular geometry of the eastern border of the North American plate compared to the relative linear form of the Bronson Hill arc complex. This concept has been described for the Appalachians by a number of workers (Rankin, 1975; Thomas, 1977; Williams and Doolan, 1979, for example). In western New England, the area from central Vermont south to New York City is dominated by an increased number of basement thrust slices and high-grade syntectonic metamorphism of Taconian age (Sutter and others, 1985; Ratcliffe, 1984a, 1984b). The conclusion that this southern area was under a very thick tectonic overburden is inescapable. The area from central Vermont north to Quebec, however, lacks evidence for a thick stack of imbricated sialic slices and high-grade dynamothermal metamorphism. We believe that the structural complications involving late tectonic imbrication of earlier formed thrust-and-fold systems in Vermont and Massachusetts is the result of greater structural overlap and convergence of North American crust from central Vermont south. Ratcliffe (1975) suggested that the large eastward-directed promontory of the North American crust in western New England accounted for the unusually severe collisional tectonics in the Berkshire massif. During collision, this promontory formed a pressure point with the relatively straight western edge of the Bronson Hill plate. Deformation persisted in this latitude, and many sialic slices were imbricated without changing the polarity of the subduction zone because the area of the promontory was small compared to the Bronson Hill plate. To the south, off the south edge of the promontory, the sialic slices pass into steeply dipping, right oblique thrust faults in the northern end of the Hudson Highlands and near the Cortlandt complex. These faults are marked by mylonitic zones of late synmetamorphic Taconian age based on  $^{40}\text{Ar}/^{39}\text{Ar}$  data (Sutter and others, 1985; Ratcliffe, 1984b). Slip lines from these faults in the Berkshire massif and Hudson Highlands trend  $\text{S}70^\circ\text{--}75^\circ\text{W}$  (Ratcliffe, 1969a, 1984b). To the north of central Vermont, an increasing number of cover slices appear from beneath the western border of the Rowe-Hawley slices (Fig. 2) as they are traced into the Quebec re-entrant. Slip lines from the Hinesburg and Champlain thrust faults in western Vermont trend  $\text{N}50^\circ\text{--}60^\circ\text{W}$  (Stanley and Sarkesian, 1972). We believe that the change from left oblique thrust faulting in northern Vermont to right oblique thrust faulting in Massachusetts and New York during late Taconian time supports our contention that the Massachusetts area was located near the apex of the promontory and consequently suffered much greater deformation and tectonic overburden. The differences in the tectonic evolution of the Taconian orogeny in western New England and Quebec, as portrayed by St. Julien and Hubert (1975) and St. Julien and others (1983), thus are real and reflect important variations that are to be expected in any orogen that closes along irregular margins.

### The Eastern Vermont Slices and the Geometry of the Taconic Root Zone

Until the later part of the 1970s, the metamorphic stratigraphy along the east limb of the Proterozoic massifs had been considered depositional (Hatch and Stanley, 1973), although some faults had been recognized (Thompson, 1972, and *in* Chang and others, 1965; Stanley, 1968). Norton (1971, 1975) was the first to suggest that the Middlefield fault zone (MFZ) separated the Proterozoic rocks of the Berkshire massif from the Hoosac Formation. Berkshire gneisses interlayered with Hoosac rocks are found

throughout this zone, and the thickness of the Hoosac drastically changes from Vermont to western Connecticut. Subsequent work by Ratcliffe (1979a) showed that the Middlefield fault zone (MFZ) merges northward with the Hoosac Summit thrust (HST, J4) and forms the western boundary of the allochthonous Hoosac in southeastern Vermont. This thrust zone is interpreted as being Taconian in age on the basis of insecure Rb/Sr geochronology of intrusive granitic rocks that cut the fault zone (Ratcliffe and Mose, 1978) and the correlation of the granites with similar alaskite and trondhjemites that intrude fault zones where  $^{40}\text{Ar}/^{39}\text{Ar}$  data are available (Sutter and others, 1985).

Major rethinking of the metamorphic stratigraphy of the Rowe-Moretown interval began in the late 1970s, during compilation of the Bedrock Map of Massachusetts (Zen and others, 1983). Work along the Rowe-Hoosac contact showed that mappable units, which define an older structure in both the Hoosac and the Rowe, are truncated along this boundary throughout its length in Massachusetts (Fig. 2; Stanley, 1978; Ratcliffe, 1979a, Fig. 5; Stanley and others, 1984, Fig. 13). This boundary was interpreted as a major thrust zone, the Whitcomb Summit thrust (WST), which severely flattened and smeared out the older structure during movement and subsequent deformation. Sheared and truncated beds of quartz-rich schist and blue-green schist are preserved in many outcrops of the Rowe near the thrust and are interpreted to be related to Taconian deformation, although Acadian events have overprinted the fabric. Well-developed mineral lineation and quartz rods plunge down the dip of the dominant schistosity in the thrust zone and are similar to lineation described by Ratcliffe (1979a) as being associated with the Hoosac Summit thrust, the soles of imbricate faults being in the Berkshire massif. Lenses of Rowe Schist are present in the eastern part of the allochthonous Hoosac, and large lenses of Hoosac have been mapped locally in the Rowe along the Deerfield River (J5) (Chidester and others, 1967; Zen and others, 1983). A continuous exposure through one of these Hoosac lenses shows that the different rock types in the Hoosac are not symmetrically repeated across the lenses, indicating that the lenses are slivers rather than isoclinal folds. The lenses in the Rowe and Hoosac indicate that the Whitcomb Summit thrust is a zone of distributed thrust slices of varied widths and lengths.

Further evidence for the Whitcomb Summit thrust is based on (1) drastic changes in thickness of the Rowe Schist from ~2,000 to 100 m northward across Massachusetts (Fig. 2); (2) the fact that ultramafic bodies are restricted to the Rowe-western Moretown belt and are absent from the allochthonous Hoosac Formation, as well as from the late Proterozoic and Cambrian sedimentary cover on the North American crust; and (3) the finding that the facies line separating volcanic from nonvolcanic parts of the allochthonous Hoosac to the west is discordant with the thrust zone (Fig. 2; dashed line, J5). This suggests that considerable displacement has occurred across the Whitcomb Summit thrust and that it postdated the tectonic emplacement of the ultramafic bodies into the Rowe Schist and Moretown Formation.

The Whitcomb Summit thrust continues southward into western Connecticut where it joins Cameron's Line (Fig. 2, CLT, N3). In northern Connecticut, the Moretown Formation, or its equivalent, is in contact with the allochthonous Hoosac. The intervening Rowe is missing, although it is present with serpentinite lenses directly to the east (M5; Hatch and Stanley, 1973, Pl. 1; Stanley, 1968, Fig. 3). Cameron's Line can be traced southward into New York where it juxtaposes schist and metavolcanic rock of the Hartland Formation (Fig. 2, Rowe-Hawley slices) against Proterozoic basement and its carbonate platform cover (Fordham Gneiss-Lower Quartzite-Inwood Marble-Manhattan Schist sequence) (Hall, 1976, Fig. 2; Hall, 1980, Fig. 2). This line has long been recognized in Connecticut as a fundamentally important tectonic surface (Rodgers,

1970, p. 94). Hall (1980) showed it to be the root zone for the Taconic allochthons.

The Whitcomb Summit thrust has not been traced north into Vermont, and its existence there may be questioned. If, however, we consider the thrust as the surface that separates rocks on the east with ultramafic lenses from rocks to the west without ultramafic lenses, then the trace of the Whitcomb Summit thrust in Vermont could be drawn approximately along the western contact of the Ottauquechee Formation with the Pinney Hollow Formation, at least as far north as the village of Moretown, Vermont (location 7, D5). At Moretown, the contact between the Hazens Notch slice (HNS) and the Pinney Hollow slice (PHS) terminates against the western contact of the Ottauquechee (OEc). Thompson (*in* Chang and others, 1965, p. 20–21) described the contact between the Pinney Hollow and Ottauquechee Formations, ~60 km south of Moretown (G6), as a “transition zone a third of a mile wide” in which “at least three distinct bands of black phyllitic quartzite, each about 200 feet wide, are intercalated” (with) “pale green chlorite-sericite-quartz schist of the Pinney Hollow (that) is here locally albitic and biotitic. Similar conditions are found both farther north and south along the strike. The width of the transition zone, however, is not constant, and it may be wholly a zone of infolding of Pinney Hollow and Ottauquechee types.” The relations are identical to those that we observe in equivalent rocks along, and east of, the Whitcomb Summit thrust in Massachusetts, and they lead us to propose that the Whitcomb Summit thrust continues northward, at least to Moretown, Vermont, approximately along the Pinney Hollow–Ottauquechee contact. A different way of thinking of these relations is that the Pinney Hollow slice is structurally overlapped by the thrust (Whitcomb Summit thrust) along the western border of the Ottauquechee Formation as the slices are traced southward into western Massachusetts.

North of Moretown, Vermont (7, D5), the Whitcomb Summit thrust is thought to follow the contact between the Rowe–Hawley slices and the Hazens Notch slice (HNS) and has been named the “Belvidere Mountain thrust” (BMT) for exposures at Belvidere Mountain (Hollis-Gale, 1980). Evidence for this correlation is found between Lowell and North Troy (A7), where a complex fault zone has been mapped along this contact (Stanley and others, 1984). The imbricated slices of Ottauquechee and Stowe lithologies to the east are part of the Missisquoi Valley fault zone (MVFZ). To the west, mappable folds and imbricated slices of albitic schist and gneiss in the Hazens Notch slice are tectonically interlayered with carbonaceous schist and quartzite similar to the Ottauquechee Formation. The albitic rocks are lithically similar to the Hoosac Formation to the south and differ only by the presence of serpentinite bodies. The Belvidere Mountain Amphibolite forms discontinuous slivers along this zone, which is traceable southward to Belvidere Mountain where Laird and Albee (1981b) reported medium-high-pressure, sodium-rich amphibole. Hollis-Gale (1980) and Doolan and others (1982) showed that the ultramafic complex and garnet amphibolite in this area form a series of southeastward-inclined thrust slices with faults marked by slivers of adjacent metasedimentary rocks and mylonitic and cataclastic fabrics, suggesting repeated movement along this zone. Vestiges of epidote-amphibolite facies (garnet-barroisitic hornblende) are preserved in the coarse- and fine-grained amphibolite where they have been partially altered to greenschist-facies assemblages (Hollis-Gale, 1980; Laird and Albee, 1981b). Directly to the north at Tilliston Peak, a high-pressure/low-temperature assemblage of glaucophane omphacite, garnet, and phengite is reported from the Belvidere Mountain Amphibolite (Laird and Albee, 1981a). We propose that the Whitcomb Summit thrust is a regionally extensive tectonic zone extending from Cameron’s Line (CLT) in western Connecticut northward to the Canadian border, a distance of 500 km.

The Hazens Notch, Pinney Hollow, and Underhill Formations to the west also form major thrust slices, although our evidence is far less com-

elling, compared to the Whitcomb Summit thrust and its along-strike equivalents. The major arguments for the Hazens Notch slice are (1) the presence of ultramafic lenses in the Hazens Notch Formation and their absence in the Underhill Formation and (2) the truncation of the Jay Peak Member (€Zuj, A6) of the Underhill Formation (Doll and others, 1961) by the contact between the Hazens Notch and Underhill Formations. The arguments for the Underhill slice are that (1) it is sedimentologically distinct from the Pinnacle and Fairfield Pond Formations on the west limb of the Lincoln anticline; (2) it contains medium-high-pressure amphiboles (barroisite and winchite) (Laird and Albee, 1981b) which are absent to the west; and (3) part of the western boundary is marked by slivers of middle Proterozoic rocks, mylonitic quartzite, graywacke, and metabasite (Towers, 1982b, 1982c; DiPietro, 1983b). Mylonitic fabrics are also typical of the Jerusalem slice (DiPietro, 1983a, 1983b). The Pinney Hollow slice is based solely on its map pattern with the other slices and its contact relations, described above, with the Ottauquechee Formation west of the Chester dome (G6). The existing evidence does not preclude the presence of smaller slices in each of these major slices. Current mapping east of the Lincoln massif shows that the eastern contact of the Underhill slice with the Hazens Notch and Pinney Hollow slices is, indeed, a zone of synmetamorphic thrust faults (Stanley and others, 1985).

The relationships in the Rowe Schist and the lower part of the Moretown Formation, which are poorly preserved in western Massachusetts, are more clearly displayed in the equivalent rocks in northern Vermont (A7, Stanley and Roy, 1982; Stanley and others, 1984). The critical findings are these. (1) The present configuration of the Ottauquechee, Stowe, and Moretown Formations is largely the result of tectonic juxtaposition of depositional sequences along thrust faults of undetermined westward displacement. (2) Deformed metadiabase dikes and sills are restricted to fault slices distinguished by mafic volcanic rocks, fine-grained phyllites, and metasiltstones. (3) Serpentinized ultramafic and associated igneous rocks occur in fault zones. (4) Mapped faults have been folded and, in some cases, refolded by one of four fold generations. Most of the thrust faults are F2 in relative age and synmetamorphic, although faults coeval with F1 and F3 are present. These data suggest a complicated evolution involving early mafic intrusion, westward imbrication, and associated folding followed by refolding and fault reactivation in which easterly rooted thrusts were cut by westerly rooted thrusts, resulting in a more complex chronology than the simpler westward accreting sequence of the piggyback model. On the basis of structural evidence, the major deformation is considered to be Ordovician (Taconian orogeny); regional tilting and minor folding occurring during the Devonian (Acadian orogeny) (Stanley and others, 1984). The available isotopic age evidence indicates an older Taconian metamorphism, but disagreement still exists as to the importance of the younger Acadian metamorphism in the Cambrian and Ordovician rocks (Laird and others, 1984; Sutter and others, 1985).

The northern Vermont geology in the Missisquoi Valley fault zone (MVFZ) is traceable southward into the Rowe Schist in western Massachusetts where its members are discontinuous. Many of the ultramafic rocks in this belt occur along the contacts of these members (Zen and others, 1983; Stanley and Hatch, *in press*). Although most of the quadrangle mapping was done before the tectonic origin of the Rowe Schist was recognized, the details of the belt are probably similar to the Prospect Hill area (L5; Stanley and others, 1984) which was mapped after the tectonic origin was suspected.

The structural fabric and metamorphic history that has been mapped in northern Vermont, and suggested by earlier work in Massachusetts, is similar to the mapped geology in the exposed parts of such accretionary wedges as Taiwan (Suppe, 1980; Stanley and others, 1981; Liou, 1981) and Nias (Moore and others, 1980; Moore and Karig, 1980). This relationship justifies the interpretation that the eugeoclinal rocks from the



allochthonous Hoosac Formation eastward to the lower part of the Moretown Formation and, in particular, the serpentinite belt, originated within an accretionary wedge of Taconian age.

### Geometry of the Taconic Root Zone

The previous stratigraphic discussion has shown that the Taconic slices are rooted along the western part of the Eastern Vermont–Rowe–Hawley slices east of the middle Proterozoic massifs (Fig. 1). As the root zone is traced southward from northern Vermont, it is structurally overlapped by more easterly situated thrust slices. For example, the Underhill slice is overlapped by the Pinney Hollow slice (PHS at E5) which is, in turn, overlapped by the Rowe–Hawley slices at the latitude of the Vermont and Massachusetts border (I5). Vestiges of the Taconic slices along the root zone therefore become severely overlapped from the southern end of the Quebec re-entrant toward the promontory in southern New England where the Whitcomb Summit thrust and Cameron's Line indeed form a cryptic suture for not only the Taconic slices but for other such easterly situated slices as the Hazens Notch (Fig. 2). We believe that there has been ~660 km of displacement across this zone at the latitude of section C–C' in western Massachusetts (Pl. 1, Fig. 2; Pl. 2, section 1). This overlapped geometry suggests that the Whitcomb Summit thrust and its northern continuation in Vermont, the Belvidere Mountain thrust (BMT), has been reactivated repeatedly during the Taconian orogeny and, consequently, has ridden across slices situated to the west.

### ANALYSIS OF THE GRAVITY-SLIDE HYPOTHESIS

The gravity-slide hypothesis raises two questions. (1) Were the Taconic allochthons emplaced by gravity sliding? (2) Were the rocks soft when transported? Gravity emplacement may involve either soft sediment or lithified rock. Push-from-the-rear tectonics does require indurated, if not lithified, material to sustain such large thrust slices as we see in the Taconics. Zen (1961, 1967) concluded, from a series of independent observations, that a gravity-sliding model involving unconsolidated sediments best satisfied many important observations. These were the following.

1. Wildflysch conglomerate is present at the leading and trailing edges of the Giddings Brook slice, suggesting that the allochthon overrode unconsolidated sediments.

2. The age of the wildflysch and the age of the youngest sediments in the allochthon overlap. Some rocks in the allochthon therefore could be unlithified as well.

3. No difference in metamorphic grade exists between autochthon and allochthon; therefore, the allochthon was not emplaced as metamorphic rocks.

4. Foliation in the wildflysch crosscuts clasts and matrix alike, as well as overlying allochthon; therefore, foliation and metamorphism did not precede emplacement.

5. Gravity sliding was an attractive idea because a deepening, black mud sea (exogeosynclinal Walloomsac Formation) formed immediately prior to emplacement and provided a regional slope toward the foreland.

6. There is an absence of a recognized root zone east of the middle Proterozoic massifs.

Zen concluded that gravity sliding was a plausible mechanism not in conflict with the available data which was drawn largely from the Giddings Brook slice. Although the gravity-slide analysis is logical and consistent, there is little evidence preserved in the allochthonous rocks that uniquely demonstrates the mechanism. At present, there does not seem to be much data for soft-sediment deformation in the Giddings Brook slice. In addition, wildflysch-like rocks in themselves do not require a gravity-slide mechanism, only that the brows of the slices break surface and produce

erosional debris that is subsequently overridden. These facts by themselves do not preclude hard-rock gravity sliding, although the dimensions of the Giddings Brook slice and its spatial mismatch to the supposed source area on top of the external Proterozoic massifs generate serious reservations on the gravity-sliding interpretation.

According to the analysis presented in the foregoing pages, the gravity-slide mechanism does not seem likely. The critical arguments for this position are:

1. The stratigraphic-sedimentological relations that palinspastically restore the Taconic slices to a continental slope environment east of the easternmost occurrence of North American sialic basement.

2. Recognition of the lower part of the carbonate platform southeast of the Green Mountain massif.

3. Recognition of major thrust zones in the pre-Silurian eugeoclinal rocks of the Eastern Vermont and Rowe–Hawley slices.

4. Palinspastic inequality between the Taconic slices and the supposed depositional site on the middle Proterozoic massifs as proposed by Zen, 1967.

5. Synmetamorphic fabrics in group 3 slices and the presence of carbonate slivers between the slices of group 2 indicate hard-rock emplacement of these slices.

6. The regionally coherent stratigraphy of the Giddings Brook slice and its large size suggest that it was lithified prior to emplacement and thus could support horizontally directed stress.

7. There is a lack of definitive evidence for a westward-inclined paleoslope that would allow sliding of continental margin sediments over the carbonate platform with its black shale cover.

8. Recognition that deformation increases to the east across the Taconic slices rather than to the west, as one would expect if they were emplaced by gravity sliding.

We therefore suggest that the Taconic slices were emplaced as a coherent slice that failed to the east, as it moved across the carbonate platform, to produce the stacking order that is observed today (Stanley and Ratcliffe, 1980, 1982, 1983). Westward imbrication occurred along the west margins, as suggested by the recent work of Rowley and others (1979) and Bosworth (1982). Emplacement of the slices resulted from horizontal compression generated by eastward subduction of the North American plate beneath the Bronson Hill plate. Although gravity spreading (Elliot, 1976) probably contributed to the westward movement of the slices, we believe that horizontal compression was the dominant mechanism and prefer to interpret Taconian history using the mechanisms described by Chapple (1978), Suppe (1981), and Davis and others (1983).

### EVOLUTION OF THE WESTERN MARGIN OF NEW ENGLAND DURING THE TACONIAN OROGENY AS CONSTRUCTED FROM RETRODEFORMED SECTIONS

Although the following discussion emphasizes the Taconian orogeny, severe Acadian deformation, which pervades much of New England, has overprinted this older geology. For example, cross section B–B' (Pl. 1, Fig. 2B) shows that the Precambrian through Devonian section is recumbently folded over the Chester dome. These folds and the accompanying metamorphic events are the western front of a series of regional west-facing fold nappes that dominate the Acadian geology of the Bronson Hill anticlinorium (Thompson and others, 1968; Robinson, 1979; Robinson and others, 1979; Robinson, *in* Zen and others, 1983) and become more extensive in southern Massachusetts and Connecticut. In Vermont, the intensity of the Acadian orogeny diminishes to the north. Although we suggest that much of the large-scale tectonic fabric of western New England is Taconian, the direct evidence for this age assignment is uncertain at present. Subsequent extension during the Mesozoic produced normal faults throughout the

region, but their influence on our reconstruction is very minor, with the exception of the large listric faults in central New England which extend northward along the Vermont–New Hampshire border (section 1, Pl. 2).

The evolution of the western margin of New England is described by studying the sequence of retrodeformed sections on Plate 1. Section 1 shows the geology as it is interpreted to exist today from the Bronson Hill anticlinorium in central Massachusetts to Albany, New York. Sections 2–8 successively retrodeform that geology to produce the pre-Taconian configuration. The displacements and the major surfaces that were active during each of the stages are described below each section. The total displacement beginning with the emplacement of the Giddings Brook slice, is on the order of 655 km. This value would be reduced to 620 km if the Green Mountain slice is the same as the Housatonic slice. These values do not include the shortening associated with multiple generations of cleavage reported for these rocks. If 50% shortening is assumed for this process, a conservative estimate to be sure, then the total shortening would be on the order of 930 to 980 km, which is rounded off to 1,000 km. This value is reasonable if one considers the terminal velocity of plate subduction to be 60 to 90 km/m.y. (Forsyth and Uyeda, 1975, p. 178). According to Pfiffner and Ramsay (1982, Fig. 1B), the time span for deformation in mountain chains is between 1 and 30 m.y., a common rate being 1 to 3 m.y. In the latitude of Taiwan, the collision between the western Philippine plate and the Eurasian plate has covered a time span of 15 m.y. The processes and plate geometry that have led to the formation of Taiwan are analogous to those that took place in western New England during the Taconian orogeny. We therefore use a figure of 15 m.y. for the Taconic orogeny, beginning just before the emplacement of the allochthons and ending with the final displacement of the sialic slices. The total shortening would be 1,050 km, using Taiwan rates (70 km/m.y., Seno, 1977), which basically agrees with the displacements arrived at by the retrodeformational process.

### Construction and Significant Features of Sections 1–8 (Pl. 2)

Eight important features of these sections are emphasized here.

1. Reversing the deformation from sections 1 through 8 requires considerable restoration of material to each of the sections. This restored material was originally eroded from the cross sections during each stage in the evolution of the western margin. The restored material has been assigned to appropriate slices dictated by the scheme of emplacement. For example, restoring the eroded material in sections 3 and 4 means that parts of group 2 slices appear on top of group 3 slices (Greylock slice). Evidence for this, however, is not present today because of extensive post-Taconian erosion.

2. The topographic profiles that are assigned for the “Taconian Mountains” during their evolution are realistic if we consider topographic profiles for active converging margins. For example, the maximum elevation for Taiwan today is 13,000 ft (3,940 m), with an annual uplift rate of 5 mm/yr (Peng and others, 1977). Recent analysis of the topography in Taiwan relative to its converging rate of 70 km/m.y. has shown that the topography is balanced by the rate of erosion, which is, in turn, influenced by the tropical climate in Taiwan (Suppe, 1981). The same type of configuration is shown in the sections for the western margin of New England where the climate was comparable to Taiwan in the Middle Ordovician (Bambach and others, 1980, Fig. 6). The sections are dynamic in the sense that they not only attempt to show the evolution of the margin by retrodeformation but also to depict the topographic configuration and, hence, aspects of the evolution that would not otherwise be apparent (points 1 and 3, for example).

3. An important aspect of the inferred topography is the morphological expression of active thrust faults. These features are shown as steep

slopes where they intersect the Earth's surface. Some of these slopes may have been subaerial (Whitcomb Summit thrust), whereas others were clearly subaqueous (Giddings Brook thrust). Mass wasting from the upper plate along the advancing front would form olistostromal deposits in basins on the lower plate. These deposits were largely ephemeral, but some were preserved along the leading edge of, and beneath, the Giddings Brook slice. Elsewhere, they soon disappeared as the higher basins were uplifted and eroded to produce more mature deposits. Continual recycling of olistostromal material may well explain the lack of ultramafic debris in the Austin Glen Graywacke along the front of the Giddings Brook slice. The sections certainly emphasize the possibility that the material in the Austin Glen could have been recycled many times from its original source and, hence, have nothing to do with the slices as they are found today.

4. One of the more subtle aspects of the present cross sections and the events that have led to their formation is the degree to which the North American crust has responded to loading during westward imbrication. This becomes apparent as the younger sections are retrodeformed along eastward-dipping thrust zones. For example, simply moving the upper plate of the Middlefield–Hoosac Summit thrust zone eastward in section 4 results in the eastern part of the section ending up far below sea level. To correct for this error, the dip of the thrust zone is reduced in the retrodeformation process so that the Bronson Hill arc complex and associated accretionary wedge in section 5 are in a reasonable position near sea level. In short, reversing the collisional process unloads the eastern part of the North American crust, causing it to rise toward sea level (compare sections 1 and 8).

The thickness of the imbricated slices on the North American crust in each of the sections (4–7) may be, in fact, too thin compared to many modern accretionary margins. Independent depth estimates, however, based on pressure-sensitive mineral assemblages, are generally lacking for western New England because the available isotopic age data are unclear as to the age of recrystallization of many of these assemblages. Our interpretation assumes that most of the movement on the faults discussed in the paper are Taconian, although we cannot rule out Acadian motion. The solution of this problem must await future isotopic work.

5. The emplacement of the Taconic slices (sections 5, 6, 7) is shown to result from horizontal compression and gravity spreading during collision between the North American and the Bronson Hill plates. Gravity sliding is not employed for any of the slices. During compression, the wedge-shaped volume of continental margin sediments thickened, which raised the center of gravity of the mountain range and provided an added lateral force to the existing horizontal compression of plate collision. The dominant movement is east-over-west, although important motion in the opposite direction may have occurred in the forearc basin when member C of the Cobble Mountain Formation was deposited and the Winchell Mountain thrust subsequently formed (sections 4 and 7). The sections show that the Taconic slices developed from a single, large, coherent slice (section 7) in which the eastern part broke up into the smaller slices of group 2 that moved over the western part of the original Giddings Brook slice (section 6). Slivers of the carbonate platform were dragged up along these thrust faults during this time.

During emplacement, the Taconic slices are shown as an internally deforming package of slices that moved over, and deformed, the carbonate platform (compare sections 5, 6, 7, 8). As a result, the frontal parts of all of the active thrust faults were continually accreting material between adjacent plates, and olistostromal (or wildflysch) deposits formed along the steepened fronts. This process was particularly active along the leading edge of the Giddings Brook slice and led to a complex history of repeated imbrication and tectonic mixing of autochthonous Middle Ordovician shales, Giddings Brook rocks, and recycled olistostromal deposits.

The Rowe Schist and the Moretown Formation contain the frag-



mented remains of an older stage of the accretionary wedge—a stage that originally developed from oceanic sediments and gradually incorporated more and more of the slope-rise section as the North American crust moved eastward into the subduction zone. During this time, slices and fragments of oceanic crust were sheared off and incorporated into the accretionary wedge. The highly fragmented and tectonized rocks were then thrust, as a more or less coherent unit, westward with the Taconic slices to produce the relations that we see today in the Massachusetts cross section (section 1).

7. We have also attempted to incorporate the metamorphic Taconian history into the retrodeformed sections. Evidence of the polymetamorphic events has been reported by a number of writers during the past 15 yr, and, recently, much of this information has been synthesized by Sutter and others (1985). They suggested that three metamorphic domains can be recognized in western New England: (1) an older, high- to medium-high-pressure/low-temperature metamorphism; and (2) two Barrovian sequences—a western, low-gradient metamorphism and an eastern, high-gradient metamorphism. An interpretation of these events is shown in the sections. The older event (M1, Pl. 2) must have occurred in the subduction zone before the slope-rise sequence was emplaced onto the continent as the Taconic slices. These higher-pressure rocks were then displaced westward as a series of thrust slices so that they now rest tectonically on lower-pressure and lower-temperature rocks, as, for example, along the Underhill thrust in Vermont. A subsequent, low-gradient Barrovian metamorphism (M2, L.G., Pl. 2) extended farther to the west (section 6, for example) and overprinted the older metamorphic event, as described by Laird and Albee (1981b), but they do not agree on a Taconian age for this event. The high-gradient metamorphism (M2, H.G., Pl. 2) developed with the emplacement of group 3 slices and culminated with the westward transport of the sialic slices. The position of the metamorphic isograds on the upper plates of the sections after peak metamorphism is then controlled by subsequent erosion and the relative movement of the respective slices.

8. In sections 3–7, the black shales of the Middle Ordovician are shown in as many as four separate basins. Two of these, one between the accretionary wedge and the volcanic arc, and the other between the continent and the wedge, persist through all of the diagrams. Others are more temporal and are caused by irregularities in the accretionary wedge itself. The deposition of the black shales, therefore, was not strictly contemporaneous, although they are commonly considered to be the same age. For example, the black shales in the basin between the accretionary wedge and the continent are probably older than those to the east or west, if our interpretation is correct. Shales accumulated here first and are now represented by the Normanskill Formation in the Taconic slices. With time, these shales transgressed westward over the carbonate platform in Middle Ordovician time. To the east, a smaller basin developed in the forearc region as the accretionary wedge grew in size. Sections 4–8 clearly show that the Middle Ordovician shales of the Walloomsac, Normanskill, Hawley, and Partridge Formations were probably not deposited in one continuous basin and, therefore, may not be equivalent in a strict stratigraphic sense.

### Tectonic Summary

The evolution of the western margin from the Middle Ordovician to the present is depicted in Plate 2, beginning with section 8 and ending with section 1. The earliest compressional event for which we find evidence is the intense imbrication of ocean-crust-rise material represented by the Rowe Schist and its northern equivalents in the serpentinite belt in northern Vermont. This thrust zone developed in an early accretionary wedge offshore of the continental margin of North America (section 8). How much displacement had occurred before this time is unknown. Although

section 8 represents the plate configuration some time in the early Middle Ordovician (perhaps graptolite zone 11 time) prior to the first stage in the emplacement of the Taconic slices, the beginning of subduction along the western edge of the Bronson Hill plate is uncertain but probably started in the Early Ordovician, if not in the Late Cambrian. This estimate is based on the generally accepted Early Ordovician age of the unfossiliferous Moretown Formation, which is correlated with the Lower Ordovician Poultney and lower part of the Normanskill Formation of the Giddings Brook slice (Zen, 1967; Ratcliffe and Hatch, 1979). The Moretown is here interpreted as being a forearc deposit receiving material from the eastern volcanic arcs and the emerged parts of the western accretionary wedge, which produced such debris-flow deposits as the Umbrella Hill Conglomerate. The deformed terrane situated seaward of the North American continent in section 8 may, in fact, correspond to the region presently exposed in Quebec, northern New England, and parts of the eastern piedmont where the Penobscotian orogeny is recognized, or thought to exist (Pavlidis and others, 1964; Newman, 1967; Pavlidis and others, 1968; Drake, in press; L. M. Hall, 1984, personal commun.).

Prior to the time depicted in section 8, coarse clastic rocks of the Dalton Formation and its equivalents accumulated in fault-bounded basins to the west while seaward, albite-rich clastic rocks of the Hoosac and equivalent formations were deposited in similar basins. These rocks were covered by carbonate rocks and quartz-rich clastic rocks of the carbonate platform which, in turn, graded eastward into the shales, siltstones, and sandstones of the Taconic sequence in the slope-rise region. Later, Middle Ordovician black shales were deposited in a large basin between the accretionary wedge and the carbonate bank and in a smaller one to the east in the forearc region. We speculate that deposition of the shales began in the slope-rise region (Taconic sequence) and gradually prograded westward over the carbonate platform to form the configuration of section 8. In this view, the Middle Ordovician pre-shale unconformity likely formed over the outer swell as the eastern margin of the North American plate approached the subduction zone. How far the trench of the subduction zone was situated to the east is unknown. We cannot specify the total width of the section.

Subsequent movement of the continental crust into the subduction zone incorporated slope-rise material of the Taconic sequence into the accretionary wedge and displaced it westward onto the eastern edge of the carbonate platform in the form of the Giddings Brook slice (section 7). At this time, the more highly deformed and metamorphosed high- to medium-high-pressure/low temperature (M1) rocks of the older accretionary wedge moved over the eastern part of the Giddings Brook slice as the slope-rise sequence was underplated beneath it in the subduction zone. The Giddings Brook slice is shown as a relatively thin slice compared to its length. Although the thickness is based on known stratigraphic estimates of the Taconic sequence, the inferred dimensions do raise the question of whether the ancient Giddings Brook slice was strong enough to move as a largely coherent unit. We suggest that abnormal fluid pressures were generated along the base of the ancient Giddings Brook slice. Pressures of this type are common in many slope-rise deposits along modern accretionary margins (Suppe and others, 1981; Namson, 1982; Davis and others, 1983). These pressures reduced frictional resistance and facilitated the movement of the Giddings Brook slice onto the carbonate platform. Subsequent dewatering may have contributed to the later failure of the eastern part of this slice (section 6).

To the east in section 7, the younger part of the accretionary wedge, made up largely of tectonized Rowe Schist, is shown as a nonvolcanic arc with slide deposits forming to the east and the west. Along the eastern slope, olistoliths of serpentinite, mafic volcanic rocks, and older deformed and metamorphosed pelites of the Rowe were deposited into volcanogenic flysch of the Cobble Mountain Formation in the forearc basin. Middle

Ordovician shales were deposited, therefore, in three separate, subaqueous basins, the first in the forearc region, the second between the emerged accretionary wedge and the continent, and the third as an exogeosyncline west of the Taconic slices. A smaller intermontane basin with olistostromal deposits is shown in front of the Pinney Hollow and Hazens Notch slices because these slices were probably active at this time.

With continued compression, the deeper and eastern parts of the slope-rise terrane failed and overrode the westerly part of the Giddings Brook slice. Underthrusting of the platform sequence during failure of the Giddings Brook slice deformed the carbonate platform and interleaved slivers of black shale (Ow, Pl. 1; Pl. 2, sections 5 and 6) and the carbonate rocks (OCs) between the slices of group 2. This process continued during westward movement of the Taconic slices (sections 6 and 5). Secondary accretion was active during all of these stages as the Taconic slices moved onto the continent. To the east, the black shale basin was continuous with the forearc basin. There was no intervening arc over the accretionary wedge. Nonvolcanic arcs, however, undoubtedly existed elsewhere along the accretionary ridge, just as gaps existed between the arcs during the time represented in section 6. Modern nonvolcanic arcs are generally less continuous than are their volcanic counterparts (for example, the Indonesian arc; Hamilton, 1979), and, hence, we show them as such in our sections.

After the emplacement of group 2 slices, the eastern margin of North America was tectonically thickened, which contributed to renewed heating and the development of low-grade Barrovian metamorphism (M2, L.G.). This thermal event overprinted the older, higher-pressure subduction-zone event and strain-softened the eastern margin so that subsequent deformation folded the older thrust faults and the underlying platform.

Continued overlapping of the North American and Bronson Hill plates displaced the basal, albite-rich, metamorphosed eastern facies (allochthonous Hoosac Formation) of the basal platform clastic rocks westward as group 3 slices (Greylock slice, section 5). With them were carried the basal parts of group 2 slices (Everett slice, section 5). Fragments of the carbonate platform were undoubtedly incorporated along group 3 thrust faults at this time. Syntectonic Barrovian metamorphism (M3, H.G.) accompanied deformation of the older (M2, L.G.) metamorphic fabric, producing elongated recrystallized micas in the new thrust-related foliation. The Greylock slice is shown in section 5 as the western continuation of the volcanic bearing, allochthonous Hoosac whose root zone now forms the eastern border of the Berkshire massif as the Middlefield-Hoosac Summit thrust zone. As described in previous sections, movement continued along the Taconic root zone where the upper plate consisted of the eastern remains of the Pinney Hollow and Hazens Notch slices. These are preserved today in central and northern Vermont. As seen in sections 7 through 4, these slices are gradually overridden in the latitude of Massachusetts as the Rowe-Hawley slices moved westward on the Whitcomb Summit thrust.

Continued entry of the North American plate into the subduction zone was accompanied by increasing temperature which steepened the Barrovian isograds (sections 4, 3, 2). Strain-softening of the sialic crust resulted in failure of the North American crust along many thrust faults. These are represented by the ten slices of the Berkshire massif and the more-coherent and less-ductile slices of the Housatonic and Green Mountain massifs to the west. The enlarged thickness of the eastern margin of North America at this time may have resulted in partial melting of the sialic and subsialic crust and subsequent intrusion of granite, diorite, and gabbro. These rocks are found today along the eastern side of the Berkshire massif and to the south in western Connecticut and eastern New York. The large size and lesser density of the sialic slices compared to the oceanic crust to the east were probably the main factors stopping the collision

process (Chapple, 1973). The emplacement of the Green Mountain and Housatonic slices produced the Green Mountain anticlinorium and its western counterpart, the Middlebury synclinorium, as the basal thrust faults stepped up over large risers (section 2). Subsequently, uplift, erosion, deposition of Silurian and Devonian rocks, Acadian deformation, and Mesozoic rifting resulted in the configuration of section 1. The degree to which Alleghenian deformation contributed to this cross section is unknown, although it is presently considered to be very minor.

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