Late Shortening and Extensional Structures and Veins in the Western Margin of the Taconic Orogen (New York to Vermont)

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ABSTRACT

Along the western margin of the Taconic orogen in New York and Vermont, undeformed quartz-calcite veins commonly occur in the belt of melange that formed beneath the westward-advancing Taconic Allochthon during the Middle-Late Ordovician Taconic orogeny. The veins have mineral slickenfibers recording either reverse or normal slip. In New York, the reverse-motion veins recording the latest phase of shortening are crosscut by the normal-motion veins and faults. The shortening indicated by the reverse-motion veins is correlated with the convergence along the Champlain thrust in Vermont, which is also crosscut by a significant, strike-parallel normal fault. Fluid inclusion data from the veins, complemented by stable isotope data, lead to a reconstruction of the sequence of events in the context of a cooling of the fluids, which is consistent with crosscutting relationships among the veins. Following cessation of the convergence, there was regional extension of the western margin of the Taconic orogen, analogous to modern arc-continent collisional orogens. Extension progressed to normal faulting without vein precipitation, and the normal faulting significantly modified the Allochthon-melange contact. The timing of extension is constrained to follow the late Taconic thrusting and predate the latest Silurian, based on the similar fluid temperature/salinity of the reverse- and the normal-motion veins, and contrast with veins nearby in Devonian rocks. Extensional, partial collapse of the orogen was accompanied or followed by rebound of the foreland basin, perhaps due to reduction of the thrust load and/or subducted slab breakoff. The systematic gradient of homogenization temperatures exhibited by the reverse-motion veins along the orogen margin is interpreted to be caused by real differences in temperature of the vein-forming fluids.

Online enhancements: appendix, figure, table.

Introduction

Thickened continental crusts, marked by large mountain ranges, commonly "collapse" at some point, principally driven by lateral variations of gravitational potential energy (England 1983; Jones et al. 1996; Rey et al. 2001). The extensional collapse has been documented in active convergent deformation zones such as the Himalaya/Tibet (Burchfiel and Royden 1985; Burchfiel et al. 1992) and the Andes (Liu et al. 2002; McNulty and Farber 2002). Such a paradoxical association of extension with shortening explains how the thick crusts of orogens return to normal thickness without much erosional denudation but with the widespread preservation of upper crustal sequences (Dewey 1988).

The northern Appalachians in eastern New York and western Vermont, along the Hudson and the Champlain Valleys, preserve the Ordovician foreland basin and thrust belt of the Taconic orogeny. The deformation in this region is a consequence of the stacking of thrust slices driven by a collision in which there was attempted subduction of the ancient North American passive continental margin beneath island arcs (Rowley and Kidd 1981; McKerrow et al. 1991). Major components of the Taconic orogen comprising allochthonous or parautochthonous sequences, tectonic melange, and folded/imbricated flysch, however, have been understood only in a framework of shortening, and possible modifications of the shortening structures by late extensional collapse have not been considered. Strike-parallel normal faults occur within the Champlain Valley, but their ages mostly have not

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been determined or remain speculative (Stanley 1980; Fisher 1985; McHone 1987).

This article presents a strain history of the western margin of the Taconic orogen in New York and Vermont based on field study of the vein system developed along the foreland zone, as well as fluid inclusion analyses of the veins. Quartz-calcite veins, occurring commonly in the melange, are not deformed in most cases and cut across the prevalent phacoidal cleavage, which was produced during the deformation of the melange. The veins, postdating the melange development (Plesch 1994), mostly have distinct mineral slickenfibers, providing an exceptional opportunity to understand a late strain history along the western margin of the Taconic orogen. Our study documents a late shortening and a subsequent significant extensional event associated with the Ordovician collision in the area investigated. We discuss correlations of those events along the Hudson and the Champlain Valleys and the timing of the extension as well as the effects of the extension on the evolution of the foreland basin. Our results also document a systematic and regional along-strike variation in the temperatures of vein-forming fluids in the foreland zone during late orogeny.

Geologic Setting and Field Relationships

General Geology. The area investigated extends from the border between United States and Canada to the New York Capital Region (Albany and vicinity) and covers the Taconic foreland basin and the westernmost Taconic orogen, which comprise, from east to west, the Taconic Allochthon and parautochthon, melange and thrust deformation belt, and autochthonous Late Ordovician flysch over Cambrian-Ordovician carbonate platform sequences on Precambrian basement (fig. 1).

The Taconic Allochthon consists of the remnants of a series of stacked thrust sheets that underwent more than 120 km of tectonic transport onto the Laurentian continental margin during the Taconic orogeny in the later Ordovician, mostly in the Caradocian in New York (Bradley 1989). The rocks of the Allochthon consist of late Precambrian and/or Early Cambrian through early Late Ordovician sedimentary sequences deposited in the synrift basin or on the former Laurentian continental rise (Rowley and Kidd 1981). The parautochthon consists of a Cambrian-Ordovician section, mostly shelf carbonates, tectonically transported about 60–80 km along the Champlain thrust system over the autochthonous platform (Rowley 1982; Stanley 1987; Hayman and Kidd 2002*b*).

The carbonate platform sequence in the New York foreland, consisting of Late Cambrian basal sandstone overlain by carbonate rocks of up to late Early Ordovician age, is interpreted to have been deposited on the continental shelf that developed along the passive margin of ancient North America under conditions much different from those for its deep-water equivalent in the allochthon (Rodgers 1968; Bird and Dewey 1970). The top of the carbonate platform sequence is disconformably to unconformably overlain by a Late Ordovician (Caradocian) foreland-basin fill consisting of the Black River–Trenton limestones at the base, followed by dark argillites (Utica Shales), followed in turn conformably upward by flysch graywackes and shales (Rowley and Kidd 1981; Cisne et al. 1982; Bradley and Kusky 1986).

In western Vermont and continuing into southern Quebec, the transition from the Taconic orogen to the foreland is marked by the Champlain thrust, which places the parautochthonous Cambrian-Ordovician shelf section onto the Late Ordovician basinal shales (Rowley 1982; Stanley 1987). To the south in eastern New York, the Champlain thrust system continues into the extensive Taconic flysch basin of the Hudson Valley (Hayman and Kidd 2002*b*), and the transition from the Taconic orogen to the foreland is marked there by a wide deformation belt that is the lateral equivalent to the Champlain thrust system (Plesch 1994; Kidd et al. 1995).

Deformation Belt. In New York, the Late Ordovician flysch extends many hundreds of kilometers west of the present margin of the Taconic Allochthon, and equivalents are preserved along the length of the Appalachians (Bradley 1989). In the Capital Region, a deformation belt 16–20 km wide and comprising several zones that are distinguished based on the structure and assemblage of fragmented blocks is developed between the undeformed, untransported flat-lying flysch to the west and the Taconic Allochthon to the east (Kidd et al. 1995). Contacts between the individual zones are suggested to be thrusts (Plesch 1994). The westernmost zone consists of folded and faulted graywacke and shale with local incipient tectonic melange, but these strata are not pervasively disrupted (Vollmer 1981; Bosworth 1989; Plesch 1994). The zones in the eastern two-thirds of the deformation belt dominantly consist of highly disrupted and sheared rocks, a melange (Plesch 1994). The melange zones as a whole are referred to as the Cohoes Melange (Kidd et al. 1995).

The deformation belt is generally interpreted as a consequence of intense ductile and brittle shear-

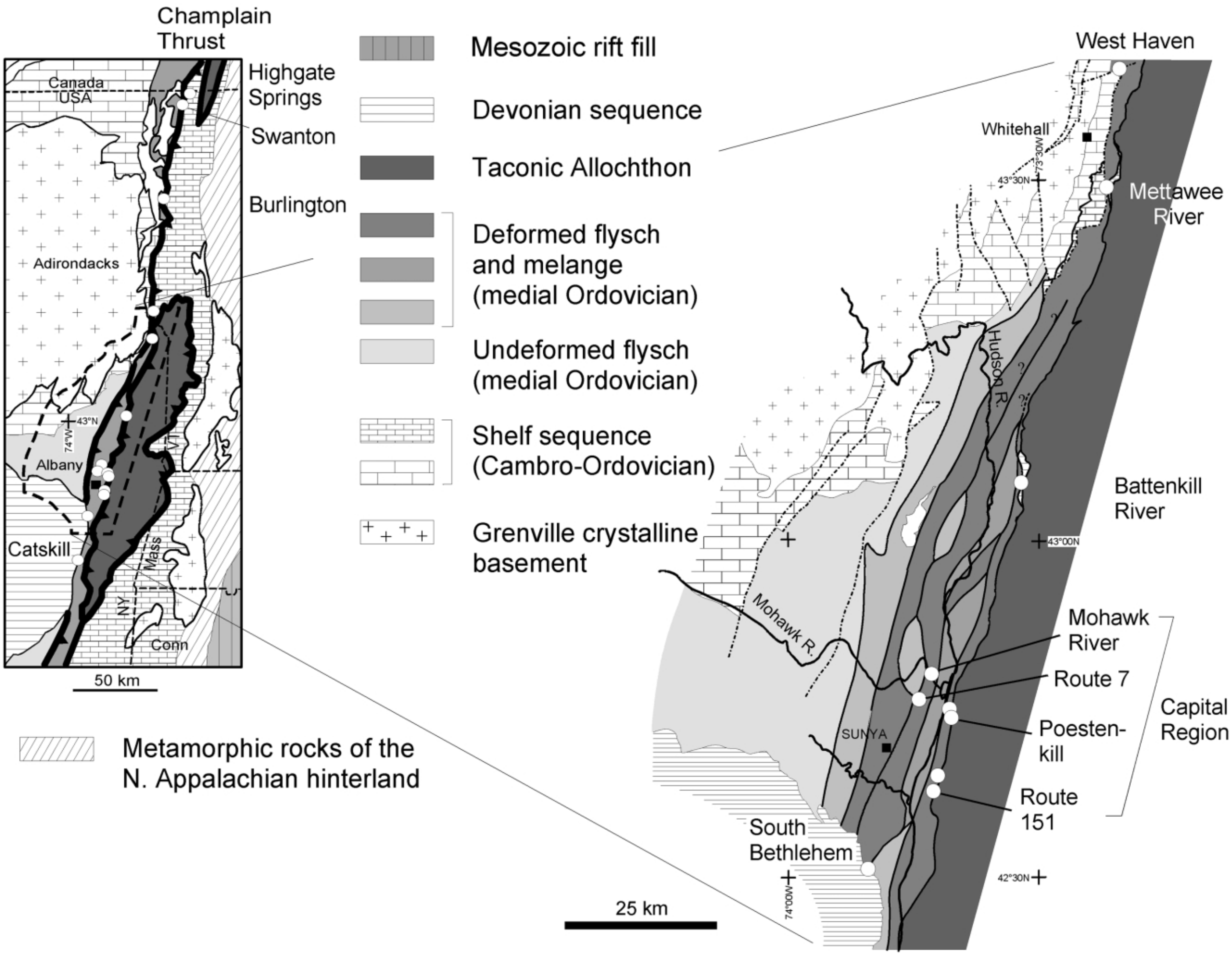


Figure 1. Geologic maps showing the general geology of the western margin of the Taconic orogen and the foreland and the locations of the outcrops for measurement and sampling (*circles*). The small brick pattern indicates transported shelf sequences (parautochthon), and the large brick pattern indicates shelf sequences without tectonic movement (autochthon). *SUNYA* = State University of New York at Albany.

ing of the flysch beneath the Taconic Allochthon as it overrode the foreland basin (Bosworth and Vollmer 1981; Rowley and Kidd 1981; Vollmer 1981; Bosworth 1989; Plesch 1994). South of Albany, New York, the deformation belt is overlain on an angular unconformity by latest Silurian/earliest Devonian carbonates of the Helderberg Group, demonstrating that it represents a Taconic (Ordovician) structure (Bosworth and Vollmer 1981).

Melange and Veins. The Cohoes Melange consists of dark gray shale/siltstone matrix and widely dispersed, fragmented blocks of graywacke, and less common "exotic" blocks (carbonate rocks, chert, and mudstone). Three principal mechanisms may have operated to produce the melange: (1) folding, boudinage, and disruption of graywacke-shale sequences due to viscosity and ductility contrasts; (2) imbrication and out-of-sequence thrusting that resulted in intercalation of sedimentary units; and (3) tectonization of olistostromes derived from exposed fault scarps (Vollmer and Bosworth 1984). The matrix characteristically shows phacoidal cleavage, which was probably formed by microfaulting, diffusive mass transfer, and phyllosilicate growth and rotation during shearing (Bosworth and Vollmer 1981).

The phacoidal cleavage is commonly crosscut by faults, with fault surfaces marked, in most cases, by well-developed veins. The veins consist of quartz and calcite and are essentially planar with continuity of <1 m to several meters or more. They

Figure 2. Reverse-motion vein with a multilayer mat of mineral slickenfibers. The orientations of fibers are slightly different in each layer, but all of them indicate essentially the same movement direction and shear sense, westward displacement of a hanging wall that was in contact with the vein when it precipitated. Route 7 roadcut, Latham, New York.

are generally a few millimeters to a few centimeters thick, but the thickness reaches nearly 10 cm in some places. In most cases, the veins have mats of elongate quartz or calcite crystals on their surfaces (fig. 2), forming the slickenside lineations (Means 1987). The mineral fibers grow parallel to the local displacement direction across slowly dilating voids on fault surfaces (Durney and Ramsay 1973). Veins with fiber intergrowths oriented perpendicular to the vein walls, formed by incremental cracking and sealing (Ramsay 1980), are rare, although some do occur in the Cohoes Melange, particularly above Cohoes Falls. However, most of the veins in the melange have mineral fibers arranged subparallel to the fracture walls. Except for rare cases where a few veins are folded, boudinaged, and rotated with the matrix, the veins almost everywhere crosscut phacoidal cleavage and are not folded. Thus, most veins precipitated after the cleavage-forming event.

Structural Measurements. Orientations of veins and slickenfibers have been measured in outcrops of the Cohoes Melange distributed from the Capital Region, New York, to West Haven, Vermont (fig. 1). A sense of displacement along the slickenfibers has been determined based on a stepping-down sense of mineral fiber mats, but where possible, the shear sense was determined based on as many criteria as available, such as an offset of markers and an oblique fabric inside the veins. In addition to field observation of the shear criteria, for key outcrops the shear sense was determined by microscopic examination of rotation of host-rock inclusions, offset of microcracks, and oblique fabrics within the veins.

Field observation and measurement show that most of the veins in the Cohoes Melange were subject to either reverse or normal displacement, and these two types of veins occur together without any distinct pattern in geographic distribution. Figure 3*a* summarizes the orientations of the veins and mineral fibers. In general, both the reverse- and normal-motion veins strike north to northeast and dip southeast, with the latter dipping more steeply. The phacoidal cleavage also shows similar strike, and it dips at an angle close to the reverse-motion veins (fig. 3*b*). The mineral fibers on the reverse-motion veins show remarkably consistent orientations plunging ESE at low to moderate angles, regardless of where they were measured (fig. 3*a*). By contrast, the orientations of mineral fibers on the normal-motion veins (fig. 3*a*) plunge at higher angles and are oriented more toward the south (Route 7) or toward the northeast (Mohawk River)

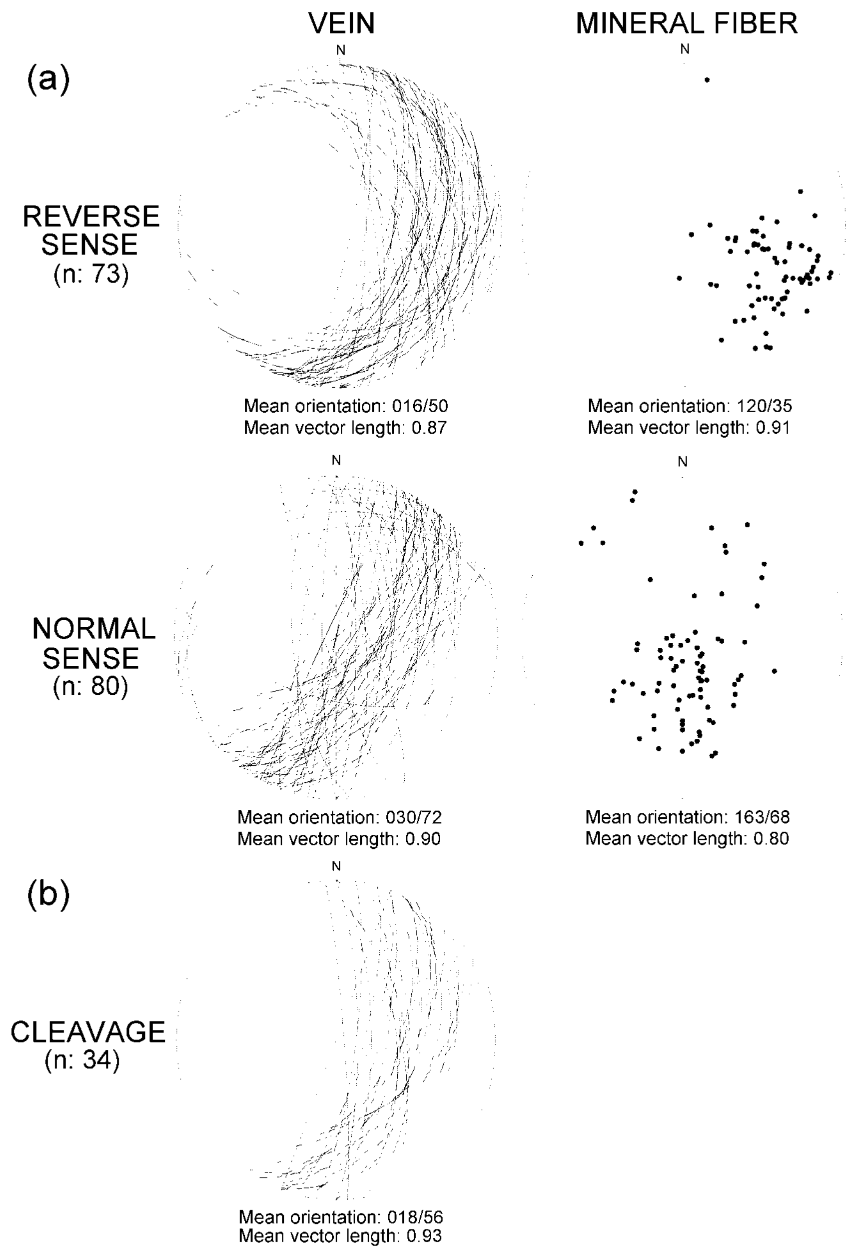


Figure 3. Results of field measurements. *a*, Equal-area stereonets of the vein and mineral fiber orientations. The mineral fibers with reverse motion plunge toward the southeast, and the displacement of the hanging wall was toward 300° in average orientation. The mineral fibers with normal motion plunge more toward the south and show large variation in orientation. Mean vector length refers to the length of the mean vector divided through the number of measurements (value 1 means all vectors are parallel and value 0 that vectors are randomly oriented). Lower hemisphere projection. *b*, Equal-area stereonet for orientation of phacoidal cleavage. Reverse- and normal-motion veins are close to the cleavage in strike, despite different slip directions. Lower hemisphere projection.

and show large variations within a given locality as well as among the localities (fig. 4). This indicates that the reverse motion occurred in an essentially single direction from ESE toward WNW, while the normal displacement was a multidirection event that occurred under a stress regime very different from the stress during the shortening. In addition, the directions in which the melange was broken up to produce the reverse- or normal-motion veins may have been controlled by the ori-

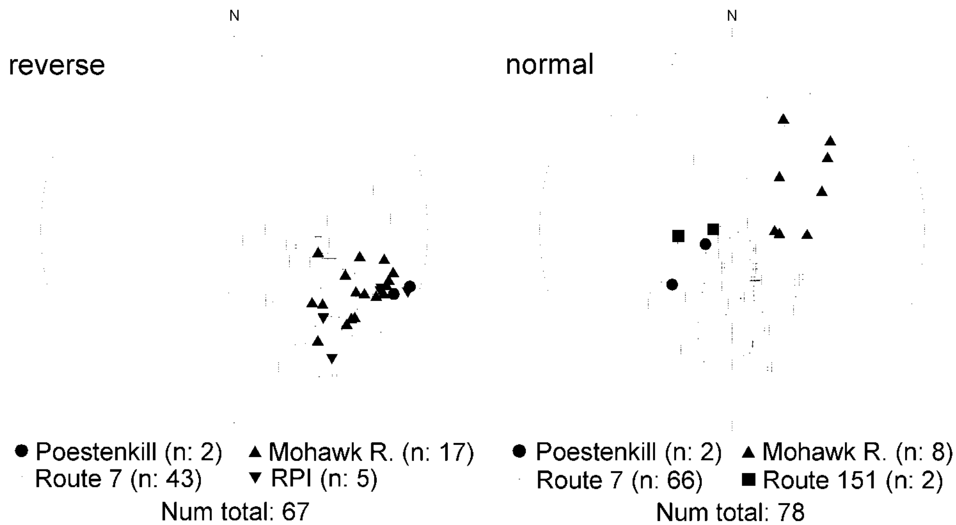


Figure 4. Equal-area stereonets showing orientation of mineral fibers in each locality. The reverse-motion fibers indicate a consistent ESE direction regardless of localities. By contrast, the orientation of normal-motion fibers shows large variations within each locality as well as among the localities. Localities with only one measurement for each type of vein are not included. Lower hemisphere projection. *RPI* = outcrops in the Rensselaer Polytechnic Institute, located 1.5 km N of the Poestenkill Falls outcrop.

entation of predominant phacoidal cleavage (fabric control) because both types of veins have strikes generally parallel to the strikes of the cleavage in spite of the different directions of movement (fig. 3).

Crosscutting Relationships. Crosscutting relationships of the veins are well exposed in roadcuts along the northern side of Route 7 located between overpasses of Route 9 and Old Loudon Road in Latham, New York. Here, a series of large, gently dipping veins occur within much of the section between the two overpasses. They show the following sense of shear: (1) left-lateral, if they strike northwest and dip to the northeast; (2) reverse, if they strike northeast and dip to the southeast; and (3) normal, if they strike northeast and dip to the northwest. These shear senses and orientations indicate a consistent northwestward movement of the melange blocks above the veins, that is, precipitation of those veins during a thrusting to the northwest. The slip and attitude of the veins in (2) and (3) may be comparable to the P- and R-fractures of Petit (1987) in larger scale, respectively, if the bulk shear is assumed to have been parallel to the subhorizontal displacement direction of the allochthon. This vein system is crosscut by many normal-motion veins and vein-free normal faults, all with east-side-down offset, which indicates that the normal-motion veins and faults postdate the slip along the reverse-motion veins (fig. 5). The normal-motion veins and normal faults strike northeast and dip moderately to steeply east or very steeply to the west. The amount of observed offset ranges between 2 and 60 cm but more commonly is about 10–20 cm.

Normal faults without veins are common in the Cohoes Melange, and their relationship with the reverse- or normal-motion veins is shown well in part of the Route 7 roadcut. Here, both the reverse- and the normal-motion veins are truncated by a large, vein-free normal fault dipping steeply to the east, indicating that the latter postdates both types of veins (fig. 6). This suggests that normal faulting without vein formation occurred after normal faulting involving vein precipitation (normal-motion veins), which, in turn, formed after thrusting (reverse-motion veins).

An outcrop at West Haven in Vermont shows that the Mettawee Fault (Fisher 1985), an undated normal fault juxtaposing parautochthonous flysch on the east against the Champlain thrust-transported carbonate platform sequence on the west (Hayman and Kidd 2002*a*), has no vein (but has a centimeter-wide gouge) along the fault, although a number of reverse-motion veins occur just above it in slates and are oblique to the fault. If the Mettawee Fault is of the same age as those veins, it is difficult to explain how the fault was not a place for mineral precipitation from migrating fluids de-

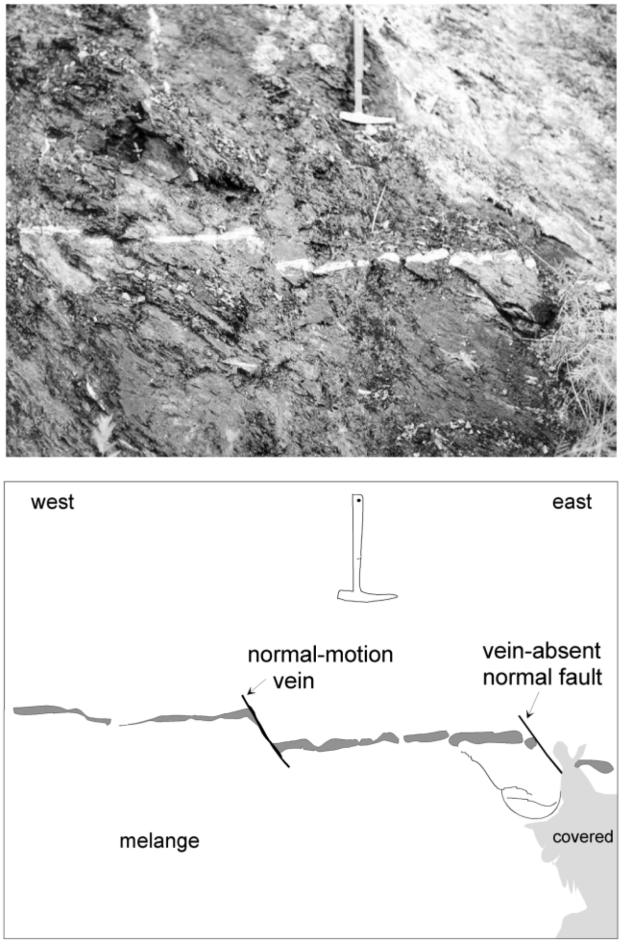


Figure 5. Normal-motion vein and a normal fault cut across a thick vein dipping at 18° to the northwest. The block above the thick vein moved toward northwest. Roadcut on northern side of Route 7, Latham, New York.

spite its proximity to other fractures that served as fluid pathways (veins); the closest vein is just 4 cm above the Mettawee Fault. This suggests that the Mettawee Fault and the reverse-motion veins above it cannot be of the same age but that the Mettawee Fault formed later than the reverse-motion veins. Thus, the reverse motion followed by the normal slip found here is consistent with the order of events determined from the Route 7 roadcut in New York, with veins representing normal faulting with accompanying mineralization absent from the West Haven outcrop.

Fluid Inclusion and Stable Isotope Data

The purpose of microthermometric analysis is to determine the relationships among veins on the basis of homogenization temperature and salinity of fluid inclusions within the veins. Most of the samples analyzed were taken from veins from which structural data were obtained. A total of 17 Taconic and three Acadian vein samples, processed into thin, doubly polished slices, were analyzed in a Fluid gas-flow heating/cooling stage that had been calibrated using synthetic fluid inclusions. The fluid inclusions analyzed are hosted in quartz or calcite and are generally 6–10 μm in long axis dimension. Because of their generally small size, nearly all inclusions were measured using a cyclic heating/cooling method to get homogenization temperatures and ice melting temperatures with optimum precision. The fluid inclusions analyzed are of aqueous, two-phase type (liquid + vapor) with vapor percentages of 3%–10% in most cases, and all of them homogenized into the liquid phase upon heating. The accuracy of measurement is ± 0.1°C, and the precision is ± 1°C (homogenization temperature) and ± 0.1°C (melting temperature). No carbonic fluid inclusions have been found. Eight samples were crushed in kerosene and four in glycerine, but the observed behavior of gas bubbles liberated from inclusions did not reveal the presence of significant quantities of methane, based on criteria of Roedder (1970) for kerosene and on criteria of A. Lacazette (pers. comm., 2005) for glycerine. Ideally, in order to characterize the primary and the secondary inclusions in one sample, many tens of both kinds of inclusions need to be analyzed, although no fixed numbers can be recommended (Shepherd et al. 1985). In this study, it was possible to analyze an average of only eight to nine inclusions in each sample because of stretching and decrepitation of the inclusions commonly encountered during heating to high temperature, a problem characteristic of this suite. We think that the number of inclusions analyzed is sufficient because, in most individual samples, the range of homogenization temperature is not large and there is virtually no difference between primary and secondary inclusions in homogenization temperature range for a given sample.

The fluid inclusions in the Taconic vein samples give homogenization temperatures ranging from 133° to 297°C, with more of the measurements from primary inclusions (fig. 7*a*). Ice melting temperatures range from −0.6° to −8.5°C, with a large peak between −1.0° and −2.0°C and with little difference between primary and secondary inclusions (fig. 7*b*). These results suggest that the fracturing that formed the secondary inclusions occurred soon after the precipitation of minerals, before the parent fluids had cooled much, introducing along the frac-

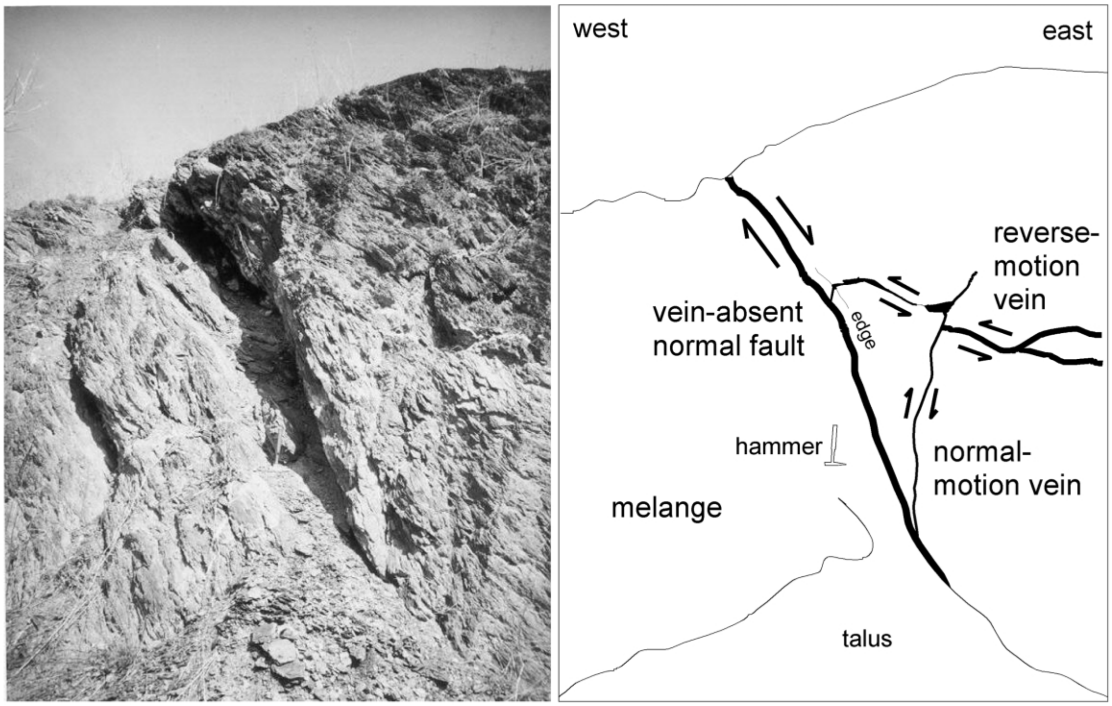


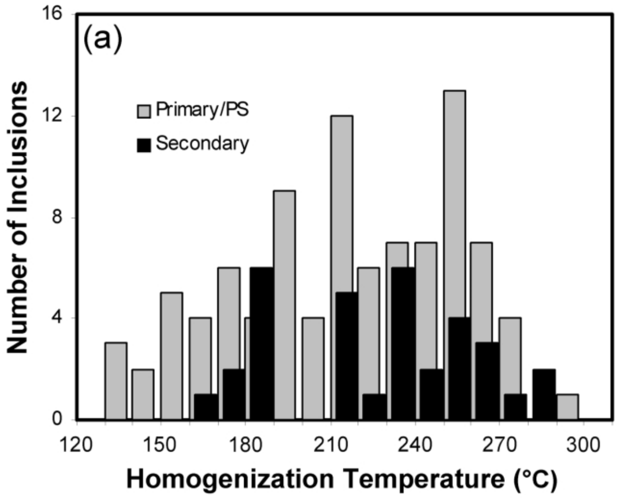
Figure 6. Reverse-motion vein (east side up) is crosscut by a nearly vertical normal-motion vein (east side down), and both veins are truncated by a vein-absent normal fault. Roadcut on northern side of Route 7, Latham, New York.

tures fluids of similar temperature and composition to those trapped in the primary inclusions.

Homogenization temperatures of the reverse-motion veins gradually decrease along strike to the north (fig. 8*a*) from the Capital Region (219°–297°C), Battenkill River (200°–249°C), Mettawee River (183°–220°C), and West Haven (153°–211°C) to Burlington (151°–199°C) and Swanton (168°–219°C). Westward across strike, there is a gradual but smaller decrease of homogenization temperatures from Poestenkill Falls (232°–297°C) through the Mohawk River (257°–276°C) to Route 7 (219°–264°C) . This regional trend reflects a variation of temperatures of vein-forming fluids during the last shortening in the area (see the "Discussion").

Ice melting temperatures were converted to salinity based on the table of Bodnar (1993). Most fluid inclusions in the Taconic samples give low salinity, ranging 1–5 wt% NaCl equivalent (fig. 9). Such low-salinity water probably originated from metamorphic reactions at depth (Larroque et al. 1996) or clay dehydration during cleavage formation (Goldstein et al. 2000). The salinity and its range increase only in veins very close to the carbonate platform sequences. The highest salinity range (5.4–12.3 wt% NaCl equivalent) was found in the West Haven locality, where the vein is located immediately above a fault contact with shelf-sequence dolomite. Fluids with very different salinities (densities) are sluggish to mix, which results in an array of salinities that lies between the original salinities when those fluids are trapped as inclusions (Shepherd et al. 1985; O'Reilly and Parnell 1999). The large variation of salinity observed at two places (West Haven and Mettawee River) was probably caused by the mixing of a low-salinity water with a saline formation water derived from the carbonate platform sequences. As discussed previously, the current fault contact between the melange and the dolomite (Mettawee Fault) did not exist when the veins formed at that place. We infer that there was a thrust between the two types of rock near the current fault that served as a major migration path and as a mixing zone for the fluids.

The one vein sample from the Lower Devonian limestone in South Bethlehem, analyzed to compare with the Taconic fluid inclusion data, yields homogenization temperatures of 98°–178°C and low salinity. The two vein samples from roadcuts in the same Helderberg Group near Catskill also give low homogenization temperatures of 98°–130°C and have low-salinity water. The Lower Devonian Helderberg limestones in the outermost



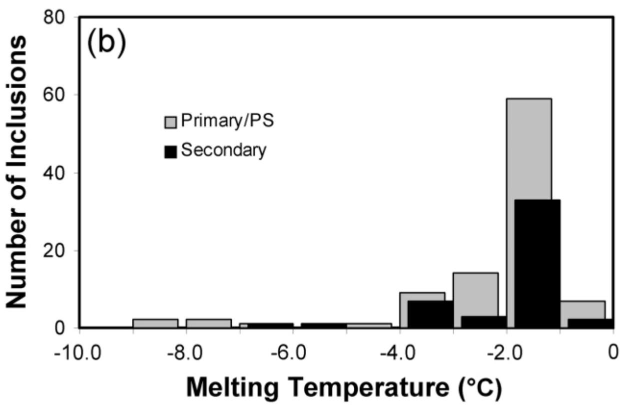


Figure 7. Histograms showing all measurement data from the Taconic vein samples. *a*, Homogenization temperatures of aqueous fluid inclusions show a wide variation among the veins. This variation is generally a function of location (reverse-motion veins) and strain type of the vein. *b*, Ice melting temperatures of fluid inclusions show little variation for most samples, with a strong peak in the −1.0° to −2.0°C interval. Some low melting temperatures are probably due to mixture of this fluid with other high-salinity fluids. *PS* = pseudosecondary.

part of the Acadian fold and thrust belt, from which the Catskill samples were taken, were at conditions ranging from 200°C, 1.8 kbar (7 km depth) to 250°C, 2 kbar (7.5 km depth) caused by tectonic-depositional burial during Acadian thrusting (Zadins 1989). If the veins had formed under those ambient conditions, fluid inclusions that had trapped low-salinity water would give homogenization temperatures between 110°–170°C (lithostatic pressure) and 140°–200°C (hydrostatic pressure). Although significant uncertainty does exist regarding the trapping conditions, the overlap between the measured and the simulated ranges of homogenization temperatures is consistent with the idea that the veins from the Helderberg limestones formed during the last stage of the Acadian orogeny (Late Devonian).

Stable isotope analysis was undertaken to compare the veins in terms of isotopic compositions and to understand the nature of the vein-forming fluid. Calcite from 15 Taconic and three Acadian samples that yielded fluid inclusion data was analyzed to determine its oxygen and carbon isotopic composition. Analytical procedures and precision/accuracy data are given in the appendix (available online and also from the *Journal*'s Data Depository on request). The $\delta^{18}O$ and $\delta^{13}C$ of the Taconic vein samples range from 16.1‰ to 19.7‰ (VSMOW) and from −6.4‰ to 0.7‰ (VPDB), respectively (fig. 10). If the precipitation temperature of each vein is determined based on the homogenization temperatures of fluid inclusions in the veins and a pressure correction corresponding to a temperature difference of approximately 100°C (Goldstein et al. 2000), the $\delta^{18}O$ of vein-forming fluids calculated from the calcite-water fractionation curve (O'Neil et al. 1969, as plotted in Friedman and O'Neil 1977) ranges from 9.0‰ to 13.5‰, regardless of shear sense or locality. These compositions are completely within the $\delta^{18}O$ range of metamorphic water and barely overlap the upper end for primary magmatic water (Sheppard 1986; Taylor 1997). However, pervasive stockwork veins, hornfels, and high sulfide contents that are typical of magma-related hydrothermal systems (Peters et al. 1990) are not present. There is no evidence of Ordovician plutonism within or near the area investigated (Drake et al. 1989), suggesting that magmatic water is not a likely source. We think that the vein-forming fluids originated from dehydration of minerals during metamorphism.

Discussion

Along-Strike Correlation of the Late Shortening and Extension. Most veins in the Cohoes Melange and in the melanges to the north cut across the phacoidal cleavage and are rarely deformed along with the matrix. Thus, the reverse motion along the veins occurred after the deformation in the matrix ceased. This indicates a shift of stress-accommodation mechanisms from partitioning of strain along the numerous surfaces of the phacoidal cleavage and through deformation of phacoids that resulted in distributed, individually small-scale displacements to a mechanism in which strain was partitioned along veins that accommodated more

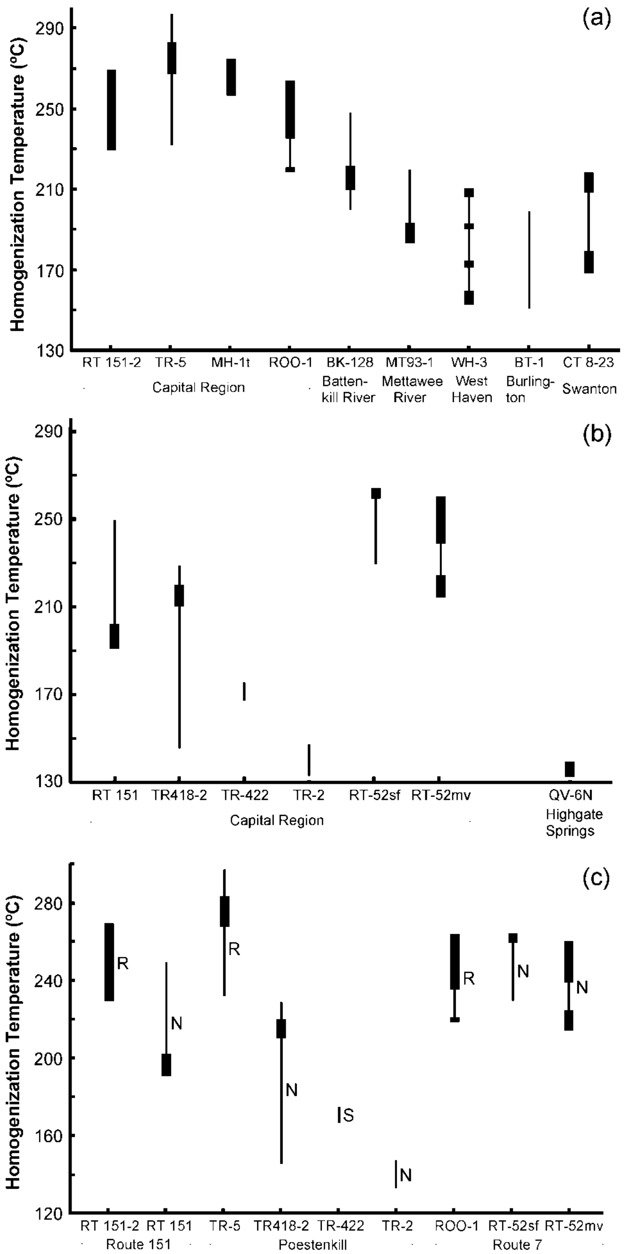


Figure 8. Ranges of homogenization temperatures. The thin lines are whole temperature ranges including outliers. The heavy lines mark a temperature range where two or more inclusions plot within 10°C of each other. *a*, Homogenization temperatures in the reverse-motion veins. Along-strike variation of temperatures refers to the variation from the Capital Region northward to Swanton. Across-strike variation is the variation to the west from TR-5 (Poestenkill) through MH-1t (Mohawk River) to ROO-1 (Route 7). Note that variation of homogenization temperatures is gradual. *b*, Homogenization temperatures in the normal-motion veins. Plot includes a strike-slip vein from Poestenkill Falls (TR-422). The variation of temperatures is irregular and abrupt. *c*, Comparison of homogenization temperatures of the reverse- and normal-motion veins occurring in the same outcrops. In Route 7, there is no difference in homogenization temperature between the reverse- and normal-motion veins. In Route 151 and Poestenkill, however, the normal-motion veins yield homogenization temperatures lower than the reverse-motion veins. In contrast to homogenization temperatures, there are virtually no differences in salinity among different vein types (see fig. 9). N = normal-motion vein, R = reverse-motion vein, S = strike-slip vein.

concentrated, individually larger displacements (Plesch 1994). From the Capital Region to West Haven, the orientations of reverse slip along the veins are remarkably consistent regardless of localities, with an average azimuth of 300° (fig. 3*a*). This average orientation exactly coincides with the displacement direction of the Champlain thrust in Vermont (also 300°) based on slickenlines, grooves, and fault mullions (Stanley 1987). Thus, the reverse motion along the veins from the Capital Region to West Haven is correlated with the reverse motion along the Champlain thrust and represents the last stage of shortening in New York and Vermont during the Taconic orogeny.

In the Capital Region, the normal-motion veins and normal faults, in conjunction with crosscutting relationships, suggest that a distinct extension occurred in the melange zones after cessation of the youngest shortening. The same strain history is suggested in western Vermont, where the Champlain thrust system is regionally truncated by the normal-sense Mettawee Fault (Hayman and Kidd 2002*b*). Absence of vein material along the Mettawee Fault in the West Haven outcrop is also consistent with the post-thrusting extension, as discussed above. We correlate the extension in the Capital Region with the normal displacement along the Mettawee Fault and suggest that there was a regional extension of the western margin of the Taconic orogen in New York and Vermont following soon after cessation of the youngest shortening.

Exposures of the western boundary fault of the Taconic Allochthon, known as the Taconic Frontal Fault (or Thrust), are scarce. One of the better exposures is at Route 9G near Mount Merino, west of Hudson, New York. The fault juxtaposes the Taconic Allochthon with the melange along a vein-absent, high-angle (almost vertical) fault. The Taconic Frontal Fault is also well exposed at Poestenkill Falls in Troy where the fault at the base

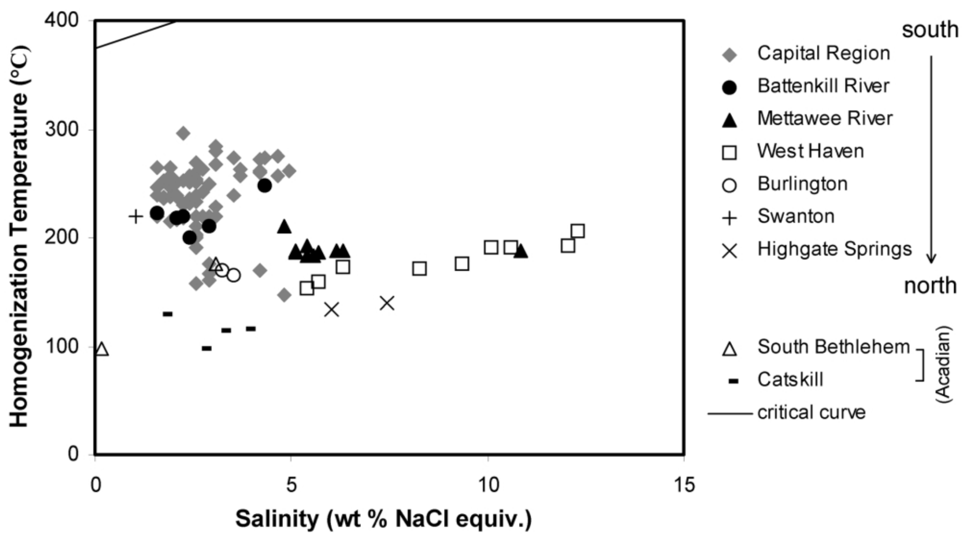


Figure 9. Homogenization temperature and salinity of fluid inclusions. Most inclusions have salinity <5%. Some samples with higher salinity are from outcrops that are close to carbonate shelf rocks. Only inclusions in which both homogenization temperature and melting temperature were determined are shown. The Devonian (Acadian) samples from South Bethlehem and Catskill are shown for comparison. Critical curve after Shepherd et al. (1985).

of the falls lacks vein material (fig. 11, available online and also from the *Journal's* Data Depository on request), but higher up the south slope in the same exposure, a vein up to 1 cm thick occurs along part of the fault, which dips 40°–54° to the east. The fabric of the vein under the microscope indicates a normal sense of shear. This suggests that the Taconic Frontal Fault, at least here and probably as far south as Mount Merino, is not a thrust but a late normal fault that repositioned the Taconic Allochthon against/into the melange. The normal displacements along at least parts of the Taconic Frontal Thrust and along the normal faults of the melange in New York and the normal slip along the Mettawee Fault in Vermont all occurred with minimal precipitation of minerals along the faults. We think that these normal displacements occurred during the later part of the overall extensional event, given the observed crosscutting relationships and the relative abundances of vein material associated with the different sets of structures. The reverse-motion veins are generally thicker and more continuous than the normal-motion veins, and the later normal faults mostly lack vein material. If volume of the mineral precipitates indicates the level of fluid activity, the activity may have waned from late thrusting to early normal faulting and decreased further during the rest of the normal faulting.

The strain history of late shortening followed by the extension is well supported by fluid inclusion data. At the Poestenkill and the Route 151 outcrops, the normal-motion veins show homogenization temperatures lower than those of nearby reverse-motion veins, and in the Route 7 roadcut, adjacent veins with reverse or normal sense of shear show overlapping temperature ranges (fig. 8*c*). These temperature relationships support the observed crosscutting relationships, if the fluids cooled after cessation of the thrusting.

In the Poestenkill Falls outcrop, reverse-motion veins in the melange have the highest homogenization temperatures, and the normal-motion vein on the Taconic Frontal Fault has somewhat lower homogenization temperatures. At this locality, the Fault is cut and offset by a strike-slip fault (Plesch 1994; Kidd et al. 1995), which contains a slickensided vein that gives yet lower homogenization temperatures. This suggests that the vein-forming fluids progressively cooled down through the strain events. It is also possible that the late thrust shortening was immediately followed by extension closer to the foreland (Route 7), where the reverse- and normal-motion veins show almost the same ranges of homogenization temperature, while closer to the hinterland (Poestenkill Falls, Route 151), there may have been a hiatus between the end of thrusting and the beginning of extension, as suggested by the more obvious temperature decrease. A reconstruction of the sequence of events and their along-strike correlations is outlined in figure 12.

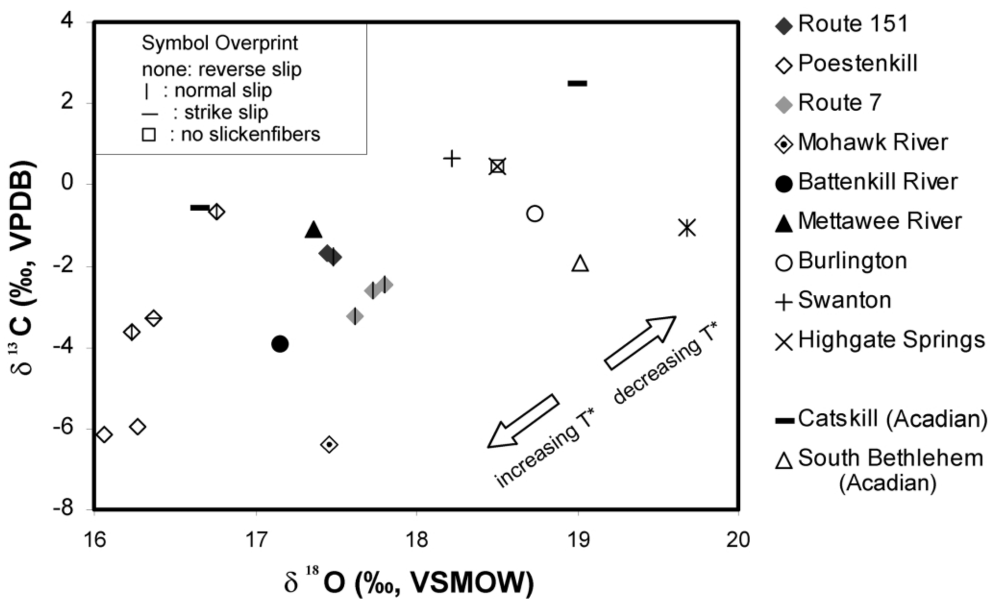


Figure 10. Isotopic compositions of vein calcite. In general, samples with higher homogenization temperatures are more depleted, and those with lower homogenization temperatures are more enriched. One problem could be that the majority of fluid inclusion data in this study is from quartz, while all isotope data are from calcite. However, when fluid inclusions in both minerals are analyzed in the same sample, they give essentially the same temperature range. Taconic vein calcite is suggested to have precipitated from metamorphic fluids; fluid-source type or types is not determined for the Acadian samples. T^* = homogenization temperatures of fluid inclusions. Samples from normal shear-sense veins and one sparry vein lacking shear-sense determination are indicated by overprints of the symbols.

Timing of the Extension. It is not straightforward to determine when the extension took place because the normal-motion veins, consisting solely of quartz and calcite, do not provide any materials suitable for routine isotopic dating. Lacking such data, we constrain the timing of extension in the following discussion.

Temperature Relationship to Acadian Orogeny. The extension might have occurred in association with the Devonian Acadian event that deformed the latest Silurian–Middle Devonian sequences unconformably overlying the Ordovician melange in the Hudson Valley fold-thrust belt, west and south of Albany (fig. 1). The South Bethlehem sample, collected from a slickensided vein in the folded Devonian limestone immediately above the Taconic unconformity, records homogenization temperatures of 98°–178°C. These temperatures are lower than those of the normal-motion veins (133°–265°C) and are significantly lower than the reverse-motion veins (219°–297°C) in the nearest sampled outcrops of the Taconic melange.

The folded and thrust late Silurian–Middle Devonian strata are ~0.5 km thick in the vicinity of Catskill and do not vary significantly along strike (Marshak 1986). This allows direct comparison of temperature data from the veins in the Catskill outcrops with the temperatures from the normal-motion veins in the Taconic melange because the effect of differential burial between the strata of the Catskill outcrops and the Taconic melange is insignificant. Two vein samples, collected from the deformed Lower Devonian Helderberg Group exposed in the Catskill roadcuts, yield homogenization temperatures of 98°–130°C, distinctly lower than those of the normal-motion veins in the melange near Albany (133°–265°C). According to Schimmrich and Marshak (1998), synkinematic veins in the deformed Silurian (?)–Devonian strata of the Hudson Valley fold-thrust belt precipitated from fluids reaching 180°–250°C, based on geochemical and fluid inclusion data. Thermal modeling for the Helderberg Group in the same locality also suggests a 200°–250°C range during the Acadian thrusting (Zadins 1989). Those temperatures overlap only the homogenization temperatures of fluid inclusions in the normal-motion veins whose trapping temperatures would be ~230°–370°C. Thus, the conditions under which the extension in the Taconic foreland zone took place near Albany were hotter than those for the Acadian deformation

Sequence of Events

Late Thrusting	Early Extension	Late Extension
* reverse-motion veins in the Capital Region	* normal-motion veins in the Capital Region	* normal faults in the Capital Region and Mount Merino
* thick and continuous vein precipitation	* thin and less continuous vein precipitation	* little or no vein precipitation
* high homogenization temperatures	* homogenization temperatures same as or lower than the reverse-motion veins	* lower homogenization temperatures, if any vein (Poestenkill)
* Champlain thrust in Vermont		* Mettawee Fault in Vermont

Figure 12. Reconstruction of the strain history based on crosscutting relationships, a fluid cooling trend, and scale and abundance of veining.

in the Hudson Valley fold-thrust belt, suggesting different ages.

Temperature Relationship to Alleghenian Orogeny. Deformation during the Carboniferous–Permian Alleghenian orogeny is known to have only a limited influence north of Pennsylvania, resulting in very low amplitude folding of the Upper Devonian sequences and a distinctive joint pattern in central New York (Engelder and Oertel 1985; Engelder 1989). The amount of post-Devonian overburden in New York is not certain. In contrast to the notion that the Devonian sediments in the Catskill Mountains were buried at only about 1 km (Zen 1981) and a suggestion of a comparable shallow burial in central New York (Karig 1987), studies based on vitrinite reflectance (Friedman and Sanders 1982) and fission track analyses (Lakatos and Miller 1983; Johnsson 1986; Miller and Duddy 1989) have suggested about 4–7 km of post-Devonian burial in New York. Recent $^{40}Ar/^{39}Ar$ analyses of K-feldspar also suggest a comparable deep burial of the basin in eastern New York during the Carboniferous (Heizler and Harrison 1998). If the rocks around the Middle-Upper Devonian boundary were heated up to near 200°C by burial in the Alleghenian foreland basin (Friedman and Sanders 1982; Lakatos and Miller 1983), then, assuming geothermal gradients of 25°–35°C/km and a maximum thickness near Catskill of sediments between the Ordovician melange and the base of the Upper Devonian strata of <1.2 km (Rickard 1975), the maximum temperatures that the Ordovician melange could have experienced during that period would have been about 230°–240°C. Those temperatures, at most, just overlap the upper range of homogenization temperatures of the normal-motion veins and cannot approach most of the trapping temperatures of the fluid inclusions in the veins, unless an anomalously high geothermal gradient (up to 140°C/km) or the former existence of thick upper Middle Devonian sequences in the Capital Region is assumed, neither of which is likely. Thick accumulation of sediments in the Alleghenian basin, if it did occur, is therefore not sufficient to explain the range of homogenization temperatures of the normal-motion veins.

An alternative hypothesis would be that the normal-motion veins precipitated sometime after the Ordovician from transient, hot orogenic fluids whose temperature was not controlled by burial depth. The thermal history based on $^{40}Ar/^{39}Ar$ analyses of K-feldspar, however, suggests progressive heating to a thermal maximum during the Late Carboniferous followed by slow cooling for the next 100 Ma, consistent with burial heating rather than a brief thermal event (Heizler and Harrison 1998). The transient, anomalously hot and younger fluid model, furthermore, requires a remarkable coincidence to explain the proximity of temperature ranges of the normal-motion veins to those of the reverse-motion veins: the homogenization temperatures of normal-motion veins overlap (Route 7) or are consistently just lower (Poestenkill, Route 151) than those of adjacent reverse-motion veins (fig. 8*c*; see "Along-Strike Correlation of the Late Shortening and Extension" for discussion). We think that such a temperature relationship indicates a close temporal linkage between precipitation of the reverse- and the normal-motion veins. The hot fluids that migrated to the Taconic foreland zone cooled down as the Ordovician convergence ceased, which is suggested by the correspondence between decreasing homogenization temperatures and crosscutting relationships, and we think that the cooling was probably rapid. This means that the regional extension, or the beginning of normal-motion vein formation, was concurrent with or occurred very soon after the end of reverse-motion vein precipitation. In this region, there are no other tectonic or thermal events known that could have produced such hot fluids in these rocks.

From the temperature relationships, therefore, we suggest that the regional extension started following the latest thrusting during the Caradocian and ended before the Silurian. This timing of the extension is consistent with the fact that no evidence of regional extension is found in the Silurian–Devonian sequences preserved in the vicinity of Catskill (Marshak 1986). Such timing of the extension is also consistent with the rapid cooling of the top of the basement to 75°C by ~430 Ma in eastern New York, inferred from the thickness of the pre-Silurian strata that remained after the Ordovician event (Heizler and Harrison 1998). All normal-motion veins record homogenization temperatures substantially higher than 75°C.

Extensional Collapse in Arc-Continent Collision Belts. Regional extension within orogens has been demonstrated in modern arc-continent collision belts such as Taiwan (Crespi et al. 1996; Lu et al. 1998) and Indonesia (Charlton 1991), where collision is still in progress. By comparison with the Taiwan and the Indonesia collision belts, we suggest that the regional extension that occurred along the western margin of Taconic orogen in New York and Vermont represents part of the gravitationally driven collapse associated with the collision between ancient North America and the Taconic island arcs. As the modern examples show, the extension may have started earlier in the high-elevation internal part of the Taconic orogen during ongoing convergence. The extension later occurred along the western margin of the orogen when the convergence ceased.

Regional Effect of the Extension. In eastern New York, synorogenic deep-water deposits in the Taconic foreland basin mostly consist of a westward-prograding facies of turbiditic sandstones and shales. Toward the west, the underlying deep-marine black shale and shelf-derived carbonate turbidites are exposed, along with overlying siliciclastic turbidites. Much farther west, beyond the deep-water basin, most of the synorogenic deposits are shallow-water carbonate rocks (Hay and Cisne 1989; Bradley and Kidd 1991). Isopach maps show that the synorogenic deposits are thickest adjacent to the hinterland in eastern New York and rapidly become much thinner toward the west (Rickard 1973; Zerrahn 1978). This suggests that a linear depocenter for the synorogenic sedimentation was located in eastern New York, close to the advancing allochthon. By contrast, Late Ordovician molasse deposits occur in central to western New York and are scarce or absent in eastern New York, suggesting that the depocenter for the molasse was located a significant distance west of the synorogenic deep-water basin (Lehmann 1993; Lehmann et al. 1994). The molasse deposits (Pulaski, Oswego, Queenston, and equivalent Formations extending to Ontario and Pennsylvania) are of very modest thickness (maximum 450 m; Fisher 1977) and of dominantly fine grain size. This suggests that the Taconic hinterland in eastern New York and Vermont changed into a significantly less active sediment source before deposition of the molasse (Zerrahn 1978; Lehmann 1993), although a modest increase in both thickness and grain size toward southeastern Pennsylvania implies the existence of a somewhat more persistent sediment source there. In other words, the Taconic orogen, which had produced a large volume of sediments during the Caradocian, quickly changed into a much lower-relief terrain at the time of the beginning of molasse deposition.

The absence of molasse in eastern New York might be attributed to postdepositional erosion, but complete removal of a thick package of molasse by erosion is unlikely because the eastward pinching out of the molasse deposits accompanies both a thinning of each formation and a decrease of grain size, suggesting that the eastern margin of the molasse basin was not located far from the present eastern terminus of the molasse in central New York (Lehmann 1993). The dramatic change from a synorogenic deep-water basin to a subaerial environment in eastern New York implies rapid filling and rebound of the basin, probably following

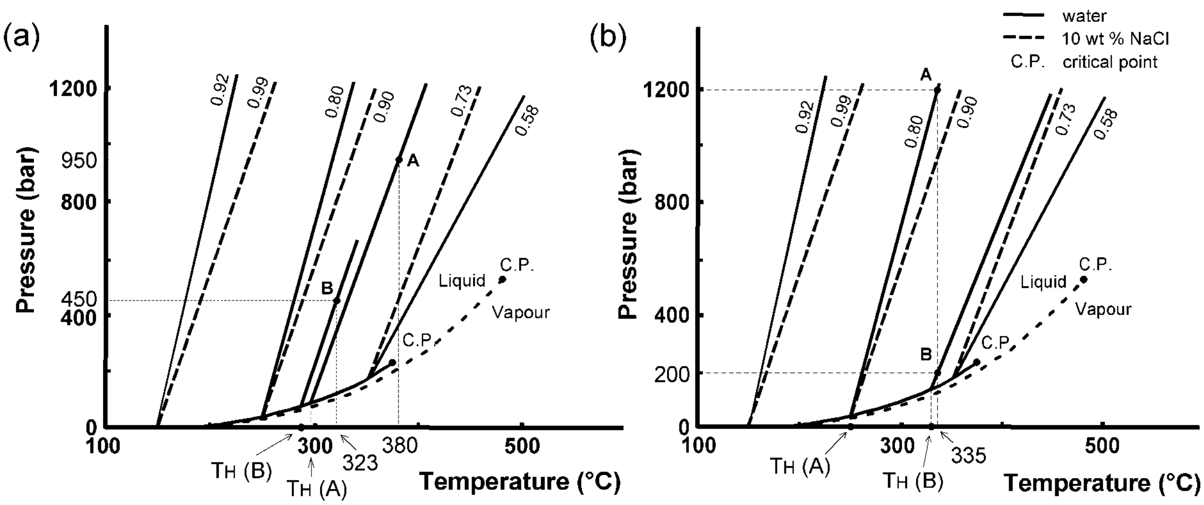


Figure 13. Simulations of the homogenization temperatures of fluid inclusions that initially trapped a single-phase water in a hypothetical crust that has only a vertical geothermal gradient and no horizontal gradient. Numbers on the upper part of curves are fluid density. Diagrams modified after Shepherd et al. (1985). *a*, Fluid inclusions A and B form at different temperatures and pressures, but there is only a small difference in homogenization temperature. *b*, Fluid inclusion A is trapped at a pressure 1 kbar higher than inclusion B and at the same temperature. However, the homogenization temperature of inclusion A is only 78°C lower than that of inclusion B.

the end of Taconic plate convergence (Bradley and Kidd 1991).

Foreland basins form by flexural downwarp of continental margins under thrust loads (Quinlan and Beaumont 1984; Stockmal et al. 1986; Tankard 1986). If a major thrust load is reduced by extension of an orogen, the foreland basin will rebound isostatically, resulting in deposition of shallowing-upward sequences overlying deep-water deposits or subaerial erosion of deep-water sequences. We suggest that, in eastern New York, enough rebound occurred to turn the synorogenic deep-water basin into a low-relief land or, at least, to change the deep-water basin into a basin shallower and more slowly subsiding than the molasse depocenter in central New York. The sedimentation pattern suggests that the rebound in central New York and Ontario was of smaller magnitude, declining from east to west. We propose that this rebound of the foreland basin was caused by the rapid extensional collapse of a segment of the Taconic orogen whose margin is now along the Hudson and Champlain Valleys, following immediately after the end of Taconic convergence. Additionally, or alternatively, breakoff of the last subducted oceanic lithosphere slab might have been involved.

Variation of Homogenization Temperature. Variations of homogenization temperatures of fluid inclusions among the normal-motion veins are independent of sampling localities, and they probably reflect how much fluids that passed through one locality had cooled when each vein was precipitated (fig. 8*b*). By contrast, the variations of homogenization temperatures among the reverse-motion veins have a distinct spatial trend: homogenization temperatures gradually decrease to the north, from the Capital Region to Swanton (fig. 8*a*). This along-strike variation might indicate (1) a diachronous formation of the veins accompanying cooling of the fluids, that is, diachronous thrusting; (2) a regional thermal gradient for contemporaneous veins developed during the same latest shortening event, as suggested by consistent slip orientations; or (3) artifacts of exposure of progressively different depth levels or of gradually varying trapping pressures. The third possibility is discussed below, and a study of the other two possibilities is under way.

Differential Erosion. The variations of homogenization temperatures might result from exposure of different erosion levels of the orogen that originally had little horizontal variation of temperatures. This argument can clearly be made for temperatures determined from geothermometers, illite crystallinity, or conodont alteration index if no local heat source is known (Johnsson 1986). Fluid inclusions are, however, different from those temperature proxies in that, except under special circumstances (e.g., boiling), homogenization temperature does not indicate trapping temperature of the fluid but is a temperature at which liquid and vapor phases separate as fluid follows a cooling path (isochore) determined by trapping *PT* conditions and fluid density. The homogenization temperature coincides with an intersection between the isochore and the liquid-vapor curve (Roedder 1984; Shepherd et al. 1985).

The following discussion tests the effects of differential erosion (burial) on the homogenization temperature of fluid inclusions, which are not necessarily straightforward. The test is about a hypothetical crust that has only a vertical temperature gradient, no horizontal temperature gradient, and a vertical pressure gradient of 3.8 km/kbar (Spear 1993). If the vertical thermal gradient is 30°C/km and inclusion A is, for example, trapped at 380°C and 950 bar, the inclusion will yield a homogenization temperature of 296°C (fig. 13*a*). If inclusion B is trapped at a depth 1.9 km shallower than A, trapping pressure and temperature based on the gradients are 450 bar and 323°C, respectively. Interestingly, such trapping condition is located close to the isochore for A, giving a homogenization temperature of 287°C, although B was trapped at a temperature 57°C lower than A. Table A1 (available online and also from the *Journal's* Data Depository on request) summarizes approximate homogenization temperatures determined from the same diagram for other thermal gradients. As the results show, the homogenization temperatures are not much different between inclusions trapped at different depths, especially for thermal gradients of 30°–40°C/km. The high thermal gradient of 70°C/km results in, at most, 71°C of differential homogenization temperature for 1.9 km of differential depth.

The extrapolations in table A1 are based on a lithostatic pressure model in which a fluid has the same pressure as the surrounding rocks. If a hydrostatic pressure model is assumed, however, the fluid pressure is 2.7 times less than the surrounding rocks (Shepherd et al. 1985). If inclusion B is trapped at a depth 1.9 km deeper than inclusion A under hydrostatic pressure and at a thermal gradient of 30°, 40°, or 70°C/km, the differences in homogenization temperature between the two inclusions are still <80°C. Moreover, hydrostatic pressure conditions are unlikely because veins are open only rarely.

This test shows that aqueous fluid inclusions yield homogenization temperatures that are largely unaffected by differential exhumation, if they were trapped under a thermal gradient typical of upper continental crust. Although rocks exposed deeper in a section could have been subject to higher temperatures, they also must have been under higher pressures whose effects on homogenization temperatures tend to counteract those of the higher temperatures. The gradual decrease of about 80°–90°C in homogenization temperature from the Capital Region to northwestern Vermont is hard to explain by differential erosion because it would require much larger differences of fluid-trapping temperature combined with different pressures and there are no grounds to assume an extraordinary vertical thermal gradient (>70°C/km). This suggests that there was a significant real variation in the trapping temperatures of the fluid inclusions among the reverse-motion veins.

Isothermal Pressure Variation. In special circumstances, such as rapid tectonic burial, fluid inclusions may form under different pressures but at similar temperatures. In figure 13*b*, inclusions A and B are trapped at the same temperature (335°C) but at pressures that differ by 1 kbar. Inclusion A gives a homogenization temperature of 250°C and inclusion B of 328°C, a difference of 78°C. In order to explain the observed 80°–90°C of along-strike variation, >1 kbar (3.8 km) of differential pressure (erosion) would be needed. Total thickness of the synorogenic Caradocian sequences (Black River and Trenton Groups, Utica and Schenectady Formations) ranges up to 1.2 km adjacent to the Taconic Allochthon/parautochthon along the Hudson and the Champlain Valleys and gradually decreases to the west (Rickard 1973). If there had been 3.8 km (1 kbar) of differential erosion (pressure), it would have been enough to expose the Grenville basement rocks at the present position of Champlain thrust in northwestern Vermont, which is not the case (fig. 1). Thus, variation of pressure along strike is unlikely.

Moreover, isotopic data do not support the isothermal condition. Oxygen and carbon isotopic compositions of vein calcite show a moderate variation that generally matches the variation of homogenization temperatures of fluid inclusions, with depleted samples characterized by high homogenization temperatures and enriched samples by low homogenization temperatures (fig. 10). Such a distinct trend, observed when carbonate rocks and a hydrothermal fluid equilibrate over a range of temperatures (Valley 1986; Bouzenoune and Lecolle 1997), suggests that there were real variations of temperature of vein-forming fluids. Equilibrium isotopic fractionation is a function of temperature, while the effect of pressure on the isotopic fractionation between calcite and water is negligible (O'Neil 1986; Chacko et al. 2001). If isothermal

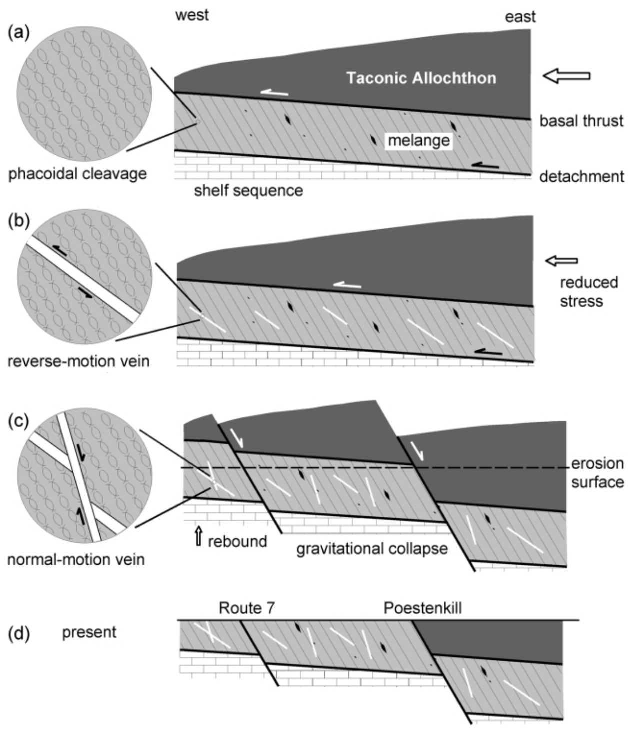


Figure 14. Schematic temporal cross sections through the western margin of the Taconic orogen in the Capital Region. *a*, Transport of the Allochthon over deforming melange is shown. The last shortening event of the thrust-sense veins in *b* is correlated with the shortening along the Champlain thrust in Vermont. The early extension in *c* is manifested as the normal-motion veins in the melange. The later extension in *c* occurred with little or no veining, and good examples are the vein-free normal faults in the Route 7 roadcuts and the western boundary fault of the allochthon in the Poestenkill Falls outcrop. To the north, this later extension is correlated with the normal faulting along the Mettawee Fault.

pressure differences caused the variation of homogenization temperatures, the isotopic compositions would not vary.

In this study, therefore, the variation of homogenization temperatures shown by the reverse-motion veins is regarded as being due to a variation of trapping temperatures. By extending the above considerations, particularly the variation of isotopic compositions and the insignificant role of differential erosion on homogenization temperature, we also regard the variations of homogenization temperatures in the other types of veins studied here as indicating real differences in trapping temperature. Thus, the veins in the melange precipitated from fluids of varying temperature that probably originated from dehydration reactions beneath the thick thrust stack and migrated to the western margin of the Taconic orogen.

Proposed Model. We propose a strain history of the western margin of the Taconic orogen of the New York and Vermont segment from the last period of shortening, illustrated in figure 14. During late overriding of the Taconic Allochthon in the Hudson Valley, the stress imposed on the melange was accommodated by distributed, numerous small-magnitude slips along the phacoidal cleavage as well as by intense folding and rotation (fig. 14*a*). Near the end of the convergence between the continental margin of ancient North America and the Taconic island arcs, this strain mechanism became inactive, and instead, the stress was absorbed along many fluid-rich thrusts, which resulted in the reverse-motion veins (fig. 14*b*). This last shortening in the melange in the Capital Region is correlated to the last thrusting along the Champlain thrust in Vermont. The termination of the convergence was followed by regional extension, driven by gravity as the thickened orogen lost compressional support coming from boundary forces. The early extension resulted in many small normal faults in the melange, with accompanying fluid activity, producing the normal-motion veins currently exposed in the Capital Region. The extension continued into a major phase of normal faulting, which was not accompanied by vein formation (fig. 14*c*). The current central-southern segment of the contact between the Taconic Allochthon and the melange was established during this period of extension. Along strike to the north in Vermont, the Mettawee Fault was also formed during this time, truncating the Champlain thrust stack. Because this extension reduced thrust loading, the Taconic foreland rebounded to change the deep-water basin to a shallow-water molasse basin and/or terrestrial environment. The regional extension and the rebound of the Taconic foreland zone might have been reinforced or initiated by breakoff of the last subducted oceanic lithosphere slab.

Conclusions

We conclude that there was a regional extension along the western margin of the Taconic orogen in what is now New York and Vermont, quickly following cessation of the thrusting during the late stages of the Taconic orogeny. Such a sequence of events is supported by fluid inclusion data from the veins which provide an independent base to suggest the relative timing of each event in the context of a cooling of the vein-forming fluids. The along-strike variation of homogenization temperatures determined from the reverse-motion veins reflects real differences in temperatures of the vein-forming fluids that reached the present exposures of the regional fault zone represented by the Cohoes Melange and the Champlain thrust during the later part of the Taconic orogeny.

Precipitation of both the reverse- and the normal-motion veins started from fluids significantly hotter than deformation/burial conditions during both the Acadian and the Alleghenian events in New York. Close or overlapping temperature conditions of the normal-motion vein precipitation with those of the reverse-motion vein development suggest that the start of extension was immediately or very soon after the end of the Taconic thrusting in the area investigated. These temperature relationships constrain the period of regional extension to follow the Late Ordovician Taconic thrusting and to be before the Silurian, along the foreland zone of the New York to Vermont segment of the northern Appalachians.

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