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NEW EVIDENCE FOR FAULTING ALONG THE "MONROE LINE", EASTERN VERMONT AND WESTERNMOST NEW HAMPSHIRE

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ABSTRACT. In 1941 the Monroe fault was defined as the boundary between two metasedimentary sequences, called the Vermont and New Hampshire sequences. By 1961 the boundary was reinterpreted as an unconformity and renamed the Monroe line. Recent mapping once again demonstrates the existence of a fault zone that has a protracted and complex history of movement. Kink bands, gouge zones, crushed-rock zones, and slickensides noted in this study between Plainfield, N.H. and Concord, Vt. indicate postmetamorphic brittle faulting. Concentrations of quartz veins, localized zones of intense flattening and truncation, and intercalation of Vermont sequence rocks with New Hampshire sequence rocks are interpreted as evidence for synmetamorphic ductile faulting. The brittle structures are believed to have formed during Mesozoic extension and to represent a northward continuation of the system of normal faults that bound the Connecticut Valley Mesozoic basins in Massachusetts and Connecticut. The ductile structures are thought to have resulted from thrusting during early Acadian (Middle Devonian) compression.

INTRODUCTION

In 1941, Eric, White, and Hadley proposed that the gray metasedimentary rocks of the eastern part of the "Vermont sequence" (White and Jahns, 1950, p. 182) in eastern Vermont were bounded on the east by the Monroe fault, named for the village of Monroe, N.H. (fig 1). East of the Monroe fault were Ordovician to Lower Devonian metasedimentary and metavolcanic rocks of the "New Hampshire sequence" (White and Jahns, 1950, p. 182; Billings, 1937, 1956). The age of the rocks of the eastern Vermont sequence has been, and still is, a matter of some controversy. Richardson (1908) reported Ordovician graptolites from these rocks, but the fossils were later discredited and then were lost. Doll (1943a,b) reported crinoids, gastropods, and a brachiopod of mid Silurian to Early Devonian age. More recently Bothner and Berry (1985) and Bothner and Finney (1986) described Ordovician graptolites from these rocks from southern Quebec and from southeast of Montpelier, Vt. Francis Hueber (written commun., 11/5/85) described plant fossils from these rocks from southern Quebec as suggesting a Lower Devonian, possibly Emsian, age. As cited in the original abstract (Eric, White,

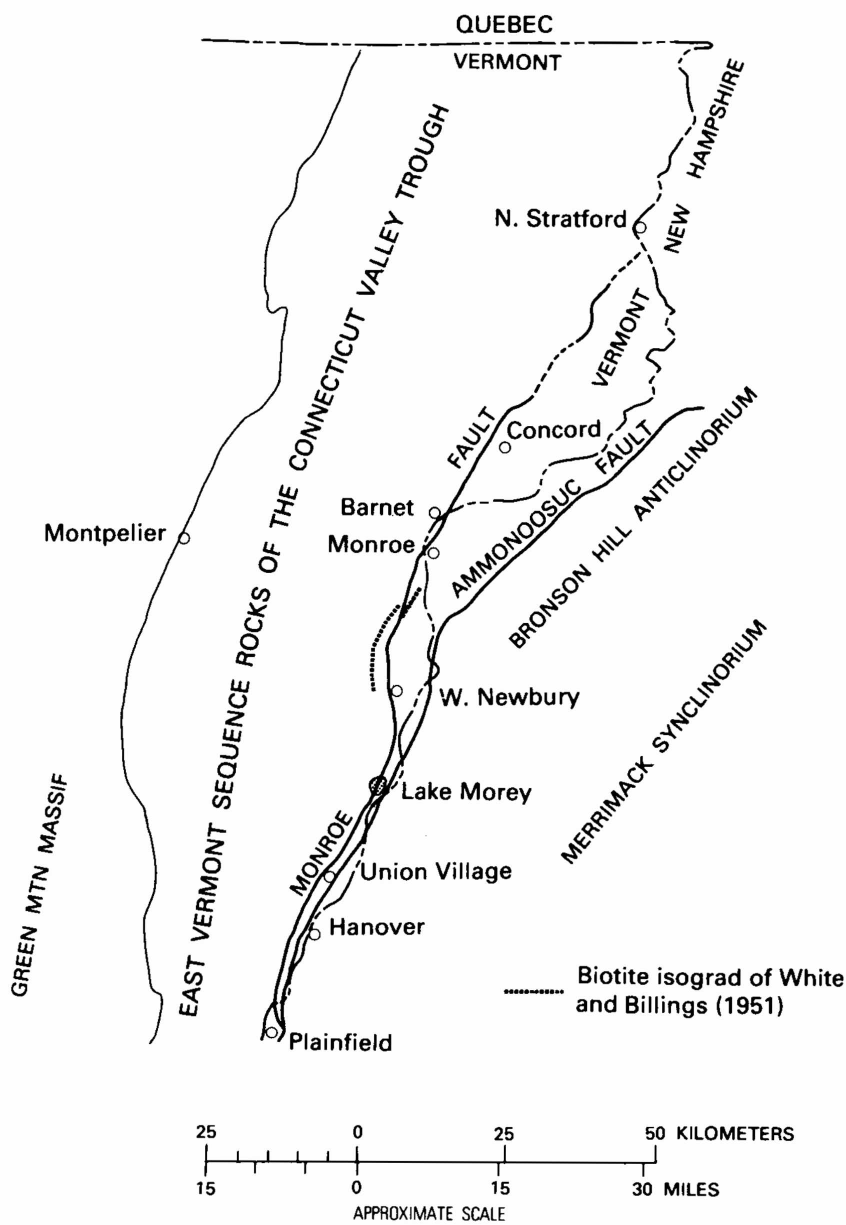

Fig. 1. Sketch map showing relation of the Monroe fault to surrounding geologic features and cited towns.

and Hadley, 1941), the specific evidence for the Monroe fault was "a distinct change in lithology" across the fault and "truncation of formations along the fault." They reported a dip for the fault of 65 degrees to the northwest at exposures 1.2 km north of the village of Monroe but indicated that "The average dip throughout the area mapped, however, is believed to be about vertical." The movement sense was indicated as northwest-side-up if the rocks of the Vermont sequence were older than the rocks of the New Hampshire sequence but southeast-side-up if the Vermont rocks were younger than the New Hampshire rocks. In other words, they advocated, or required, an older-over-younger movement sense on the fault.

All the authors of quadrangle reports and maps that include parts of the boundary between the New Hampshire and Vermont sequences between North Stratford, N.H., and Plainfield, N.H., (fig. 1) (Johansson, 1963; Eric and Dennis, 1958; Hall, 1959; White and Billings, 1951; Hadley, 1950; Doll, 1944; and Lyons, 1955), interpreted that boundary as a fault, the Monroe fault. The evidence cited by these authors included truncation of units of the New Hampshire sequence by the easternmost unit of the Vermont sequence, the Meetinghouse Slate; a slight offset of the biotite isograd (fig. 1; White and Billings, 1951, p. 670, pl. 1); very local discordance between the trends of bedding and cleavage in the rocks on both sides of the fault and the trend of the fault itself (Hadley, 1950, p. 27); a very localized breccia as much as 20 cm thick (Eric and Dennis, 1958, p. 56); and local exposures of faults in outcrops near, but not exactly at, the actual Monroe fault (Hall, 1959, p. 87). All these authors also assigned the easternmost unit of the Vermont sequence, the Meetinghouse Slate, to the *top* of the Vermont sequence, which thus required that this unit face or top to the east into the Monroe fault. None of them, however, presented any primary stratigraphic, or other, evidence to support this assignment. Hall (1959), Hadley (1950), Eric and Dennis (1958), Johansson (1963), and Doll (1944) all assigned an Early Devonian age to the eastern Vermont sequence rocks which, in combination with placing the Meetinghouse at the stratigraphic top of the Vermont sequence, *required* that the Monroe contact be a fault. None of these authors made it clear whether belief in the fault led to the stratigraphic order of the Vermont sequence, or whether belief in the order led to the fault, or neither, or both. White and Billings (1951), Lyons (1955), and Billings (1956) all assigned a pre-New Hampshire sequence Ordovician age to the eastern Vermont sequence rocks and retained the Meetinghouse at the top of the Vermont sequence.

By 1961 a variety of regional stratigraphic considerations had persuaded Doll and others (1961) to assign a Silurian to Early Devonian age to the eastern Vermont sequence rocks and to show them as resting unconformably upon the New Hampshire sequence to the southeast along what had become known as the "Monroe line." This choice was probably influenced at least in part by Doll's firm belief in his earlier reported (1943a,b) Silurian-Devonian fossils and by the strong similarity between the Meetinghouse Slate and the fossiliferous (Boucot and Arndt, 1960) Lower Devonian Littleton Formation across the Monroe "line" in Littleton, N.H. Coincident with, and perhaps as a direct consequence of, this decision, Doll and others also revised the stratigraphic order of the eastern Vermont sequence to place the Meetinghouse Slate at the bottom, rather than the top, of that sequence. To the best of my knowledge, no explanation or discussion was ever published of this reversal of interpretation of both the stratigraphy and the structure at and immediately west of the Monroe "line."

The purpose of this paper is to present evidence for post-metamorphic, brittle, probably Mesozoic normal faulting, and other evidence for synmetamorphic, ductile, probably Acadian thrust (?) faulting along the Monroe "line" in easternmost Vermont and adjacent New Hampshire. For consistency with this evidence, it will be referred to as the Monroe *fault* through the remainder of this paper. Preliminary results of this study were presented by Hatch (1985).

BRITTLE STRUCTURES

Structures indicative of post-metamorphic brittle movement along the Monroe fault include kink bands, gouge zones, crushed-rock zones, and slickensides.

Kink bands.—Lyons (1955, p. 132) described a set of structures from the Hanover, N.H., area (fig. 1) as follows: "The other type of chevron folding is a crinkling of steeply dipping schistosity along horizontal or gently westward-dipping slip planes. Along many of these slip planes the upper block has slid westerly, but this is not consistently so. These chevron folds are the youngest structures in the rocks, have a wave length seldom exceeding a few inches, and are found chiefly in the rocks immediately west of the Ammonoosuc and Monroe thrust faults." On the following page (133), in a discussion of the Monroe fault, Lyons says "The only structural indication of the Monroe fault in the Hanover quadrangle is the somewhat questionable evidence of chevron folding in some of the rocks west of the inferred position of the fault. . . ." My principal differences with Lyons' interpretations are that the movement that caused the "chevron" (kink) folds was normal rather than thrust (Lyons, 1955, p. 122) and that it was Mesozoic rather than Paleozoic (Acadian) (Lyons, 1955, p. 143) in age.

The kink folds are common in the immediate vicinity of the Monroe fault from Plainfield, N.H., to as far north as Concord, Vt. (figs. 1 and 2). My observation is that the kink folds, rather than being largely restricted to rocks just west of the Ammonoosuc and Monroe faults as Lyons suggested, are instead concentrated in more schistose rocks, including greenschists of the New Hampshire sequence and pelitic slates of the Vermont sequence, on *both* sides of but within about half a kilometer of the Monroe fault (and presumably also the Ammonoosuc fault). Most occur as kink bands striking north to northeast and dipping gently to moderately either northwest or southeast (fig. 2). The kink

A.

B.

Fig. 2. Photographs of kink folds and kink bands present along the Monroe fault. (A) in Meetinghouse Slate about 6.4 km south-southwest of Lake Morey, Vt., on US Route 5, looking southwest. (B) in Meetinghouse Slate, conjugate kinks parallel to the two 18-cm map scales, Union Village dam, Union Village, Vt., looking northeast.

folds have amplitudes ranging from about a centimeter to as much as a meter (fig. 2A). At a few localities, conjugate sets of gently dipping kink bands are present and indicate maximum compression in the verticle direction (fig. 2B). The kink folds deform all other structures in the rocks including bedding, schistosity, and quartz veins.

Gouge zones.—At scattered localities along the Monroe fault, but most conspicuously exposed in cuts on Interstate 93 about 3 km west of Concord, Vt. (fig. 1), are zones of gouge (fig. 3). At least one of the

Fig. 3. Gouge zone in gray slates of the eastern Vermont sequence. Looking southwest at cut in Interstate I-93 about 4 km west-southwest of Concord, Vt.

"several exposures of faults in the dark gray slate that crops out in the stream 0.1 mile south of Stevens Island" (3.6 km north of Monroe, N.H.) described by Hall (1959, p. 87) is a zone of gouge similar to those shown in figure 3. These zones strike north to northeast and dip moderately southeast to vertically; they are generally parallel or subparallel to the dominant schistosity in the host rocks. The gouge zones range from about 15 to 40 cm in width. In the fresh, unweathered cuts on Interstate I-93, the rock in many of the gouge zones has been ground to such an extent that it can be scooped out with bare hands. The gouge material ranges in size from flour-sized particles, which form a paste, to angular fragments a few centimeters across. In one 300-m-long interval of cut across strike on Interstate 93 west of Concord, Vt., five separate gouge zones are present. The rock immediately adjacent to some gouge zones in pelitic material is strongly kinked (fig. 3). Because the gouge zones essentially parallel both bedding and schistosity in the host rocks, it was not possible to ascertain either the sense or amount of movement along them. The open character of the zones, however, strongly suggests relatively near-surface fault movement.

Intensely deformed zones.—In many of the large cuts along Interstates I-91 and I-93 along or near the trace of the Monroe fault are zones of complexly and locally chaotically deformed rock (fig. 4). Some of the zones are entirely within one rock type, but others, such as that shown in figure 4B, form the boundary between rock types. The zones are a meter to tens of meters wide and, where an attitude of a bounding surface can be determined, are roughly parallel to the schistosity and bedding in the host rocks. In contrast to the narrow zones of gouge described above, the rock in most of these intensely deformed zones is not strongly disaggregated or finely ground. Rather it appears to have been intensely but unsystematically folded (fig. 4A). Many of the individual folds in these zones are angular or chevron and resemble the kink folds described above (note for example the top of fig. 4A and the left side of fig. 4B), but their orientations are complexly varied to random. A planar fault surface bounding a zone of intensely deformed rock is illustrated in figure 4C. It strikes about N 35 E, parallel to the regional S_1 schistosity. Although most of the rock in these zones retains the earlier metamorphic schistosity and, locally, bedding, isolated small volumes of rock in these zones are mechanically crushed and approach the degree of deformation achieved in the gouge zones described above. Because demonstrably large zones of chaotically deformed rock could only be documented in large exposures such as highway cuts, their distribution is difficult to define. They are, however, relatively common in the cuts of Interstate 91 where it roughly follows the Monroe fault between the latitudes of Hanover, N.H. and Concord, Vt. (fig. 1), and they were not seen in the cuts on Interstate 89 which angles across the eastern Vermont sequence from about 8 km south of Hanover, N.H. north-northwest to Montpelier, Vt. (fig. 1). This restricted occurrence strongly suggests a genetic relationship to the Monroe fault. The drag on the foliation in the rocks across the fault surface in figure 4B indicates west-side-down movement sense. How representative this is of the movement along other zones is unknown.

Slickensides.—Slickensides were seen at scattered localities along the trace of the Monroe fault between Plainfield, N.H., and Barnet, Vt. They were seen in both metavolcanic rocks assigned to the New Hampshire sequence and phyllites assigned to the Vermont sequence. In all cases the slickensided surfaces parallel the north-northeast-trending dominant schistosity in the host rock, and in all but one case the slickenlines on these surfaces plunge directly down the dip. The one exception is in an intensely deformed zone where broken fragments of slickensided rock were diversely oriented, presumably reflecting later breaking and rotation as a result of continuing movement. The movement that produced the slickensides therefore was essentially directly up or down the dip of northeast-striking fault planes parallel to the dominant Acadian schistosity.

Discussion of brittle structures.—A number of summary points or conclusions can be reached from the information presented in the preceding sections. First, all the structures described are restricted to a narrow belt probably no more than 2 km wide centered about the trace of the Monroe fault. This geographic restriction strongly suggests a genetic relationship between the structures and the fault. Second, the gouge zones, which can only be interpreted as faults, all trend north to

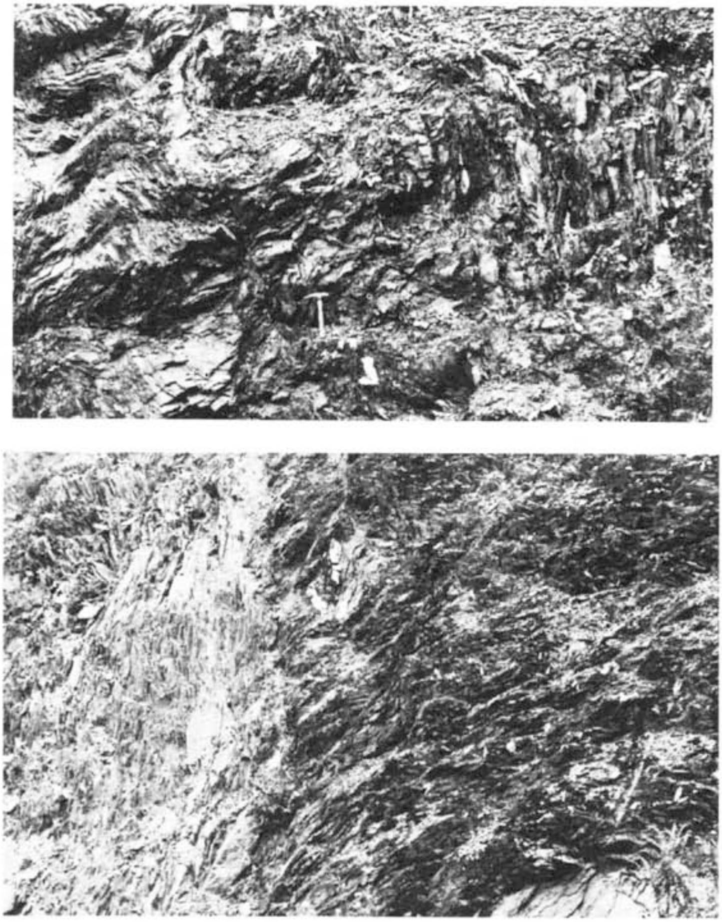

Fig. 4. Intensely brittlely deformed zones. All three views are looking northeast along Interstate I-91 about 1.5 to 2.5 km north of Lake Morey. (A) Chaotically folded S_1 schistosity in phyllites of the Meetinghouse Slate. In (B) note the sharp boundary in the middle of the photograph between gently west dipping Meetinghouse Slate on the right and vertically dipping schistose greenstone on the left (west) suggesting west-side-down movement.

northeast and dip steeply, parallel to the dominant metamorphic fabric (schistosity) in the rocks. Third, the observed slickensides indicate subvertical movement on surfaces parallel to that same schistosity. Fourth, the orientation of the kink bands, and particularly of the conjugate pairs of kink bands (fig. 2B), indicates a subvertical orientation of maximum compression entirely compatible with the movement inferred by the slickenlines and the gouge zones. Fifth, as exemplified by the five separate gouge zones in the cut west of Concord, Vt. noted above, movement at least locally was distributed along many individual

C.

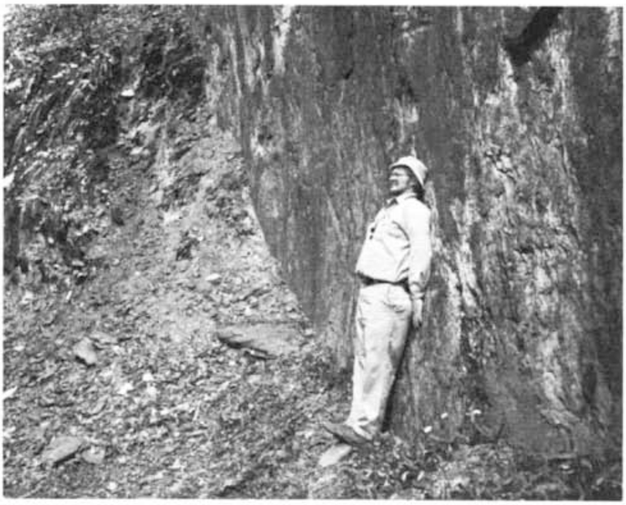

Fig. 4. (continued)
(C) Polished fault surface separating relatively undisturbed rock on the right (east) from chaotically folded rock on the west. The fault surface cuts both quartz veins and schistosity. Grooves on fault surface plunge within 2° of directly down the dip of the fault surface.

faults; there is no reason to suggest that this locality is atypical of the Monroe fault zone as a whole. The restriction of the kink bands (kink folds) and other structures to a narrow belt centered about the trace of the Monroe fault strongly implies that the faulting was restricted to a comparably narrow zone and that movement was distributed over many fault surfaces within this narrow zone.

The only clue to the net post-metamorphic offset along the aggregate fault zone is the report of White and Billings (1951, p. 670) that "the biotite isograd is slightly [about 4 km in plan view] offset in the vicinity of Boltonville" (about 12 km south of Monroe, fig. 1) (see also White and Billings, 1951, pl. 1). Assuming gently west-dipping isogradic surfaces, the indicated offset could be explained by a kilometer or less of subvertical fault movement; with more steeply dipping iosgradic surfaces, a greater offset could be required. But considering the differences in composition between the rocks of the Vermont and New Hampshire sequences in this area and the potential effect of those differences on the first appearance of biotite in the rocks, all that can be concluded from this one apparently offset isograd is that the post-metamorphic vertical movement along the Monroe fault was probably no more than a few kilometers. The west-side-down sense of offset of the biotite isograd is in agreement with the drag along the zone of intense deformation described above in indicating a west-side-down sense to the post-metamorphic movement on the Monroe fault.

DUCTILE STRUCTURES

The brittle structures described and discussed above are superposed on a variety of older structures and fabrics that I interpret as indicating earlier, more ductile, approximately synmetamorphic faulting along the same zone. These older structures include localized concentrations of quartz veins, localized intense shearing, truncation of fabrics along sharp surfaces, and intercalation of gray metasedimentary rocks typical of the Vermont sequence with metavolcanic rocks here believed to be part of the New Hampshire sequence.

Concentration of quartz veins.—Although quartz veins are locally common in the rocks of the New Hampshire and, particularly, the Vermont sequences, unusually high concentrations are present at many localities along or near the trace of the Monroe fault (fig. 5). Individual veins are generally 0.5 to 5 cm wide and 10 to 50 cm long, and collectively they may form as much as 25 percent of the local volume of the rock.

The origin of these veins is uncertain, but the following observations may be pertinent. The high concentrations of veins are in zones 1 to 10 m wide. These zones are roughly parallel to the dominant (S_1) schistosity in the host rocks and are particularly common near the trace of the Monroe fault. They are locally concentrated along contacts between Meetinghouse-type slate and greenstone, which, for reasons discussed in the section below on "Intercalation of Vermont sequence and New Hampshire sequence (?) rocks," I believe are faults. Most, but not all, individual veins in the zones of concentrated veins are parallel to the S_1 schistosity and have been folded and locally cut by Acadian F_2

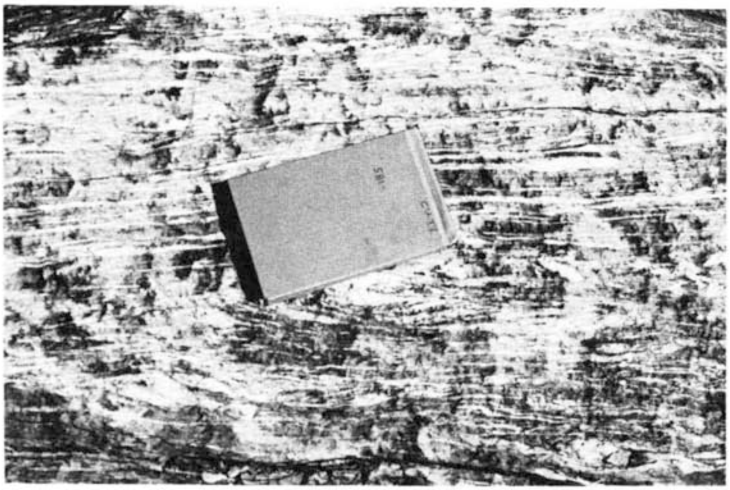

Fig. 5. Concentrations of quartz veins, looking down at subhorizontal surface of Meetinghouse Slate, U.S. Route 5 about 1 km south of Barnet, Vt. Notebook is 20 cm long.

folds and cleavage. Some are cut by planes subparallel to S_1 schistosity described below. Finally, in many places the matrix rock of these quartz veins, as well as the veins themselves, is strongly flattened and attenuated (fig. 5), suggesting that development of the quartz veins accompanied or preceded intense shearing.

These observations suggest that the veins formed early in the F_1 stage of Acadian deformation, which produced isoclinal folds throughout the area. Because the same isoclinal folds with the same axial planar (S_1) schisotosity are very common elsewhere in areas where the metamorphic grade is comparable but quartz veins are absent (Hatch, 1969), some process in addition to the folding and metamorphism must be responsible for the development of the quartz veins.

Localized intense flattening or shearing and truncations.—Some of the quartz veins in figure 5 are strongly attentuated and flattened. In these and other zones, normally equidimensional minerals such as garnet and biotite porphyroblasts are highly flattened and drawn out. In these zones, structures such as quartz veins and quartzose beds are truncated at very low angles by surfaces that locally can be traced subparallel to S_1 schistosity for few meters (fig. 6). These surfaces are interpreted as faults because they appear to have slip along them, but the fault surfaces are completely healed and are believed to be part of a ductile process. In the example shown in figure 6, at least some of the movement must have

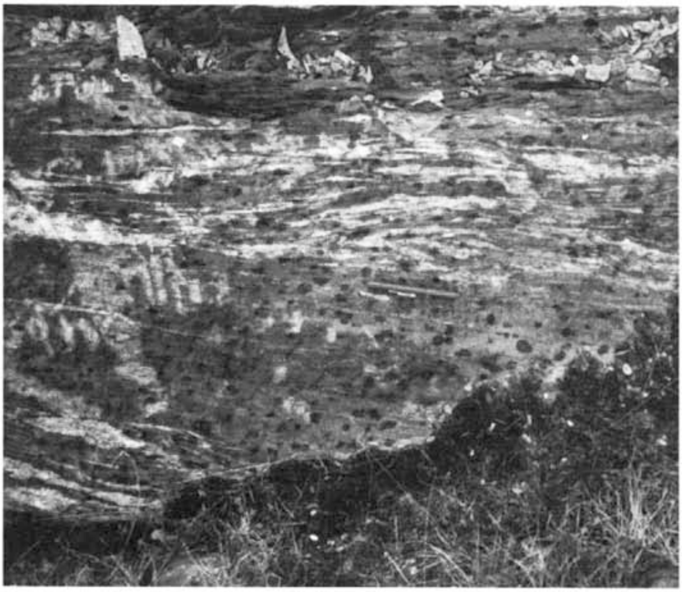

Fig. 6. Low-angle truncations. Looking down at subhorizontal surface showing truncation of quartz veins along fault surface parallel to 15 cm pencil. East Vermont sequence rocks, US Route 5, about 1 km south of Barnet, Vt.

post-dated formation of the quartz veins, which are truncated, but may have been approximately synchronous with growth of the flattened metamorphic porphyroblasts. As can be seen in figure 6, the fault movement also produced drag folds in the quartz veins and the schistosity adjacent to the faults. In this example, the dominant metamorphic fabric is subvertical and the axes of the drag folds are subhorizontal, similar in orientation to the Acadian F_1 folds. Without marker horizons it is not possible to calculate the amount of slip on any one of these ductile faults, but at every exposure where they were recognized, more than one fault was seen. Finally, because they are subparallel to the dominant fabrics in the rocks and are thus difficult to identify, it is possible that these synmetamorphic faults are very abundant, at least in certain zones. Total displacement of such a group of faults cannot be meaningfully estimated but could presumably be large.

Intercalation of Vermont and New Hampshire (?) rocks.—The third class of evidence suggestive of synmetamorphic ductile faulting consists of localities along the Monroe fault where rocks typical of the eastern Vermont sequence are intercalated with metavolcanic rocks similar to those of the New Hampshire sequence. Interpretation of these relations is somewhat speculative and is completely dependent upon assumed age or sequence assignments for both sets of rocks involved. These assumptions seem sufficiently supported, however, that the evidence is worth presenting.

At many localities, particularly in the vicinity of Lake Morey (fig. 1), dark-gray slates typical of the Meetinghouse Slate of the Vermont sequence are intercalated with greenstones and intermediate metavolcanic rocks and green quartzose schists. Previous workers, principally Hadley (1950) and White and Billings (1951), assigned both the metavolcanic rocks and the Meetinghouse-like slates east of them to the Orfordville Formation of the New Hampshire sequence. Neither their written descriptions nor my field observations point up any significant differences between the slates they assigned to the Orfordville and those they assigned to the Meetinghouse. Although it certainly is not impossible that the same lithology could have been deposited in this area during two different time periods, it seems at least equally plausible that all of the Meetinghouse-like slates are indeed Meetinghouse and that their present relations to metavolcanic rocks are a result of later folding or faulting.

Pertinent to this discussion are the following observations. Although metavolcanic rocks are present in the east Vermont sequence, they are concentrated at or near the contact between the Waits River and Gile Mountain Formations farther west in the trough, where they are known as the Standing Pond Amphibolite. They are rare in the eastern part of the Gile Mountain Formation and the western part of the Meetinghouse Slate. If the metavolcanic rocks discussed in the previous paragraph are stratigraphically interbedded, rather than structurally intercalated, with the Meetinghouse-like slates, one of two unusual situations exists. First, if the dark-gray slates are indeed Meetinghouse, the presence of interbedded metavolcanic rocks here is atypical of the Meetinghouse. Alternatively, if the metavolcanic rocks are indeed New Hampshire sequence, the presence of interbedded Meetinghouse-like slates in the New Hampshire sequence is unusual in its own right and is even more unusual in that they are restricted to an interval only about a kilometer wide at the extreme western edge of the sequence.

Without extensive paleontologic, isotopic, or geochemical data to tie down ages and geochemical characteristics, neither of the "unusual situations" described above can be ruled out. Until then, the alternative interpretation that the intercalation of Meetinghouse-like rocks and New Hampshire-like metavolcanic rocks is structural, probably fault-generated, in origin remains at least viable, and I propose it here.

About 2.6 km north of Lake Morey on the paved road just west of Interstate 91 is an exposure in which at least three intervals of dark-gray Meetinghouse-like slate are separated by at least three intervals of schistose greenstone or green schist typical of the New Hampshire sequence rocks to the east. These intervals parallel the dominant schistosity and range in width from about a meter to approx 10 m. One such interval of green schist about a meter wide between bodies of dark-gray slate is pictured in figure 7. The contacts of this body of green schist are sharp, and the rock along the contacts appears flattened and sheared. Furthermore, the body of green schist pinches out abruptly at the top of the photograph; this pinchout is believed to be tectonic. Furthermore, the tightness of the contacts (no gouge, no kink folds, no breccia, no slickensides, et cetera) supports assigning that tectonism to the synmetamorphic rather than the post-metamorphic event.

Figure 8 shows similar relations on a different scale in a small area near West Newbury (fig. 1). Figure 8A shows the geology as presented by White and Billings (1951). Figure 8B shows the same area reinterpreted here. The essential change is that I place the metavolcanic rocks of their "Orfordville" in the Ammonoosuc Volcanics of the New Hampshire sequence, I assign the slates of their "Orfordville" to the Meetinghouse and thus change their single plane Monroe fault to a more complex Monroe fault zone. Although this reinterpretation cannot be "proven," its strong points are the assignment of all the identical dark-gray slates to the same unit, the Meetinghouse, and the exclusion from the Meetinghouse of the metavolcanic rocks that are rare to absent in it elsewhere.

Discussion of ductile structures.—Of the three types of evidence presented here for ductile faulting, both the intercalation of Vermont and New Hampshire sequence rocks and the concentrations of quartz veins could be interpreted in other ways than as fault-related structures. The zones of truncations are clearly ductile faults, although the amount of total displacement cannot be determined. Taken all together, I believe they make a reasonable case for synmetamorphic faulting along the Monroe fault. The rough parallelism of the individual fault surfaces,

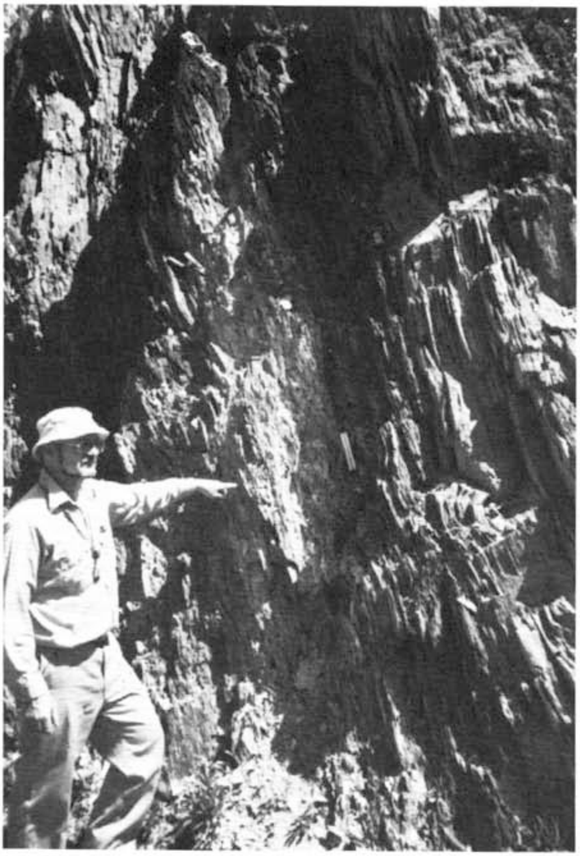

Fig. 7. Lens of green schist (where geologist is pointing) assigned to the New Hampshire sequence bounded on both sides by gray slates assigned to the Vermont sequence (Meetinghouse Slate) about 2.6 km north of Lake Morey on local paved road.

where they can be observed, with the dominant schistosity, the presence of deformed porphyroblasts in these areas, and the fact that limbs of isoclinal folds are locally sheared off by these faults suggests that the faulting, isoclinal folding, schistosity formation, and metamorphism were all closely related events. Furthermore, the gently plunging to horizontal attitude of the axes of the isoclinal folds suggests that all of these features resulted from thrust, rather than a strike-slip movement. I thus conclude that the structures described in this section resulted from thrust faulting in a compressional regime. I attribute this compression to the Acadian orogeny simply because no post-Acadian metamorphic dates have been reported from central or eastern Vermont and because the Acadian orogeny has traditionally been credited with the

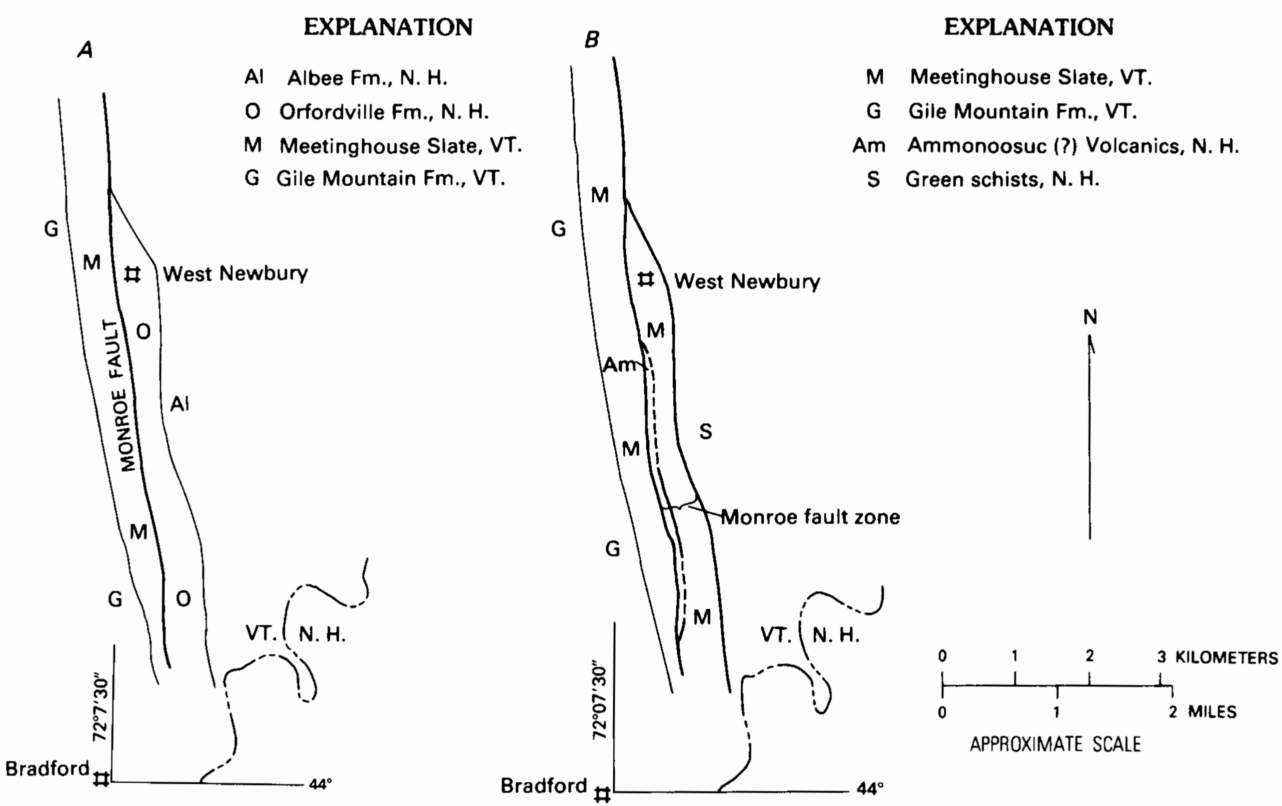

Fig. 8. Geologic map of a small area around West Newbury, Vermont. (A) as shown by White and Billings (1951, pl. 1) and (B) as mapped in the present study, showing differing interpretations of two belts of lithically similar gray slate. NH indicates unit was assigned to the New Hampshire sequence. VT indicates unit was assigned to the Vermont sequence.

deformation and metamorphism of the rocks in eastern Vermont (Thompson and Norton, 1968; Sutter, Ratcliffe, and Mukasa, 1985).

CONCLUSIONS

The preceding pages have presented evidence supportive of two periods of faulting along what has most recently been known as the Monroe line at the east margin of the Vermont sequence of metasedimentary strata in eastern Vermont. On the basis of this evidence I propose that the original designation "Monroe fault" (Eric, White, and Hadley, 1941) be reinstated. The first stage of faulting appears to be synmetamorphic. It was concentrated along a zone rarely more than a kilometer wide in which movement was distributed over many individual surfaces. Exact figures for amount of displacement cannot be determined, but the geometry of the rocks in the area of figure 8 and other comparable areas a few kilometers to the south suggests that movement on some individual surfaces must have been at least a kilometer. Net movement over a group of such surfaces could have been many kilometers. Until we have much better data on the ages of the rocks involved, more accurate figures probably will not be possible. As noted in the previous paragraph, the best available evidence suggests that the movement was in a thrust sense. The Acadian age assignment is based on the fact that fossiliferous Silurian and Lower Devonian rocks a few kilometers to the east in New Hampshire are affected by the same folding and cleavage events and on the isotopic work of Sutter, Ratcliffe and Mukasa (1985). Finally, if I am correct in interpreting the early faulting as thrusting, the present subvertical attitude of these faults must result from the later period of refolding that has been well described by White and Jahns (1950). Although White and Jahns (1950) and I believe that this later refolding was Acadian, it has not been firmly dated.

Evidence for post-metamorphic brittle faulting is even stronger. The only uncertainty is the age of the faulting. No independent evidence is available to ascertain a Mesozoic age, but the fact that all the features described are characteristic of the Ammonoosuc fault (fig. 1) and the faults bordering the Mesozoic basins in southern New England (Peter Robinson, oral commun., 1985) is at the least highly suggestive of a Mesozoic age. In fact Thompson and others (1968) extended the west-dipping east border fault of the Massachusetts and Connecticut Mesozoic basins north into Vermont to within about 55 km of Plainfield, N.H. and showed other down-to-the-west normal faults extending north to the vicinity of Plainfield. I suggest that the evidence presented here extends the Mesozoic border fault system north from the vicinity of Plainfield along the trace of the pre-existing Acadian Monroe thrust fault at least as far as Concord, Vt.—an additional 80 km. More work will be needed to ascertain the relative importance of the Monroe and the Ammonoosuc faults (fig. 1) in this northward extension of the Mesozoic rift fault system. And more work will be needed to determine exactly how far north this Mesozoic fault system extends. Finally, if the interpretations presented here are correct, they add a chapter to the model of fault reactivation in the control of Mesozoic extensional basins recently presented by Ratcliffe and others (1986). They presented evidence that Mesozoic normal faults re-used Taconian (Ordovician) thrusts in the Newark Basin of eastern Pennsylvania. This paper presents evidence in support of Mesozoic normal faults re-using Acadian (Devonian) thrusts in a similar manner.

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