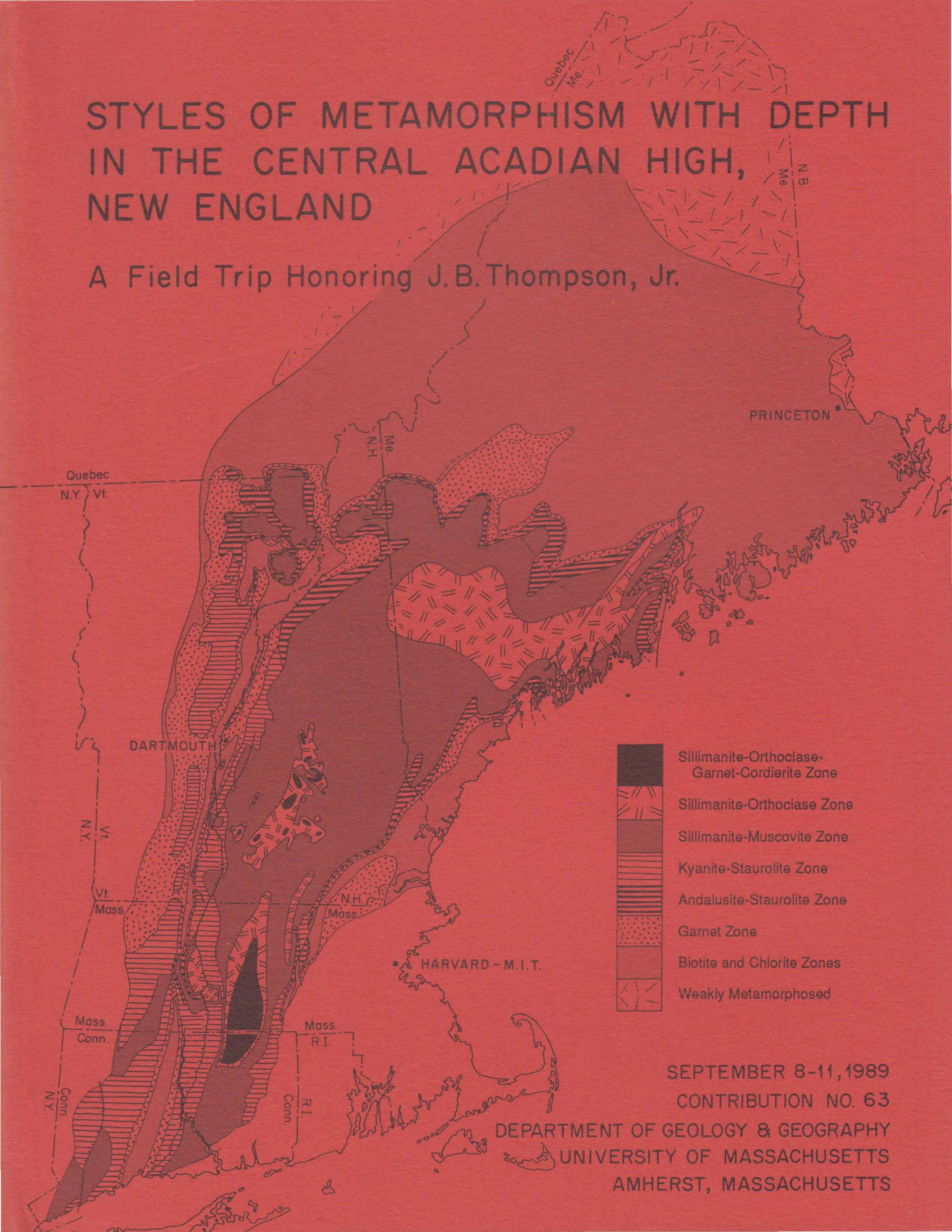
STYLES OF METAMORPHISM WITH DEPTH IN THE CENTRAL ACADIAN HIGH, NEW ENGLAND

A Field Trip Honoring J.B. Thompson, Jr.



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UNIVERSITY OF MASSACHUSETTS
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STYLES OF ACADIAN METAMORPHISM WITH DEPTH IN THE CENTRAL ACADIAN HIGH, NEW ENGLAND

A Field Trip Honoring J. B. Thompson, Jr.

by

C. Page Chamberlain[1], Henry N. Berry IV[2], John T. Cheney[3], M. P. Dickenson[4], John M. Ferry[5], Charles V. Guidotti[6], Kurt Hollocher[7], Philip H. Osberg[6], Peter Robinson[2], Douglas Rumble, III[8], Jennifer A. Thomson[2] and Robert J. Tracy[9]

C. Page Chamberlain[1] and Peter Robinson[2]
Editors

1 Department of Earth Sciences, Dartmouth College, Hanover, NH 03755
2 Department of Geology and Geography, University of Massachusetts, Amherst, MA 01003
3 Department of Geology, Amherst College, Amherst, MA 01002
4 EC Jordan, 261 Commercial St., P. O. Box 7050, Portland, ME 04112
5 Department of Earth and Planetary Sciences, The Johns Hopkins University, Baltimore MD 21218
6 Department of Geological Sciences, University of Maine, Orono, ME 04469
7 Department of Geology, Union College, Schenectady, NY 12308
8 Geophysical Laboratory, Carnegie Institution, 2801 Upton St., NW, Washington, D. C. 20008
9 Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA 24061

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Contribution No. 63, Department of Geology and Geography
University of Massachusetts, Amherst, Massachusetts

INTRODUCTION

For over three decades, Jim Thompson has led students on field trips across the northern Appalachians of New England. Typically, this trip was run across the tectonic grain, extending from the Grenville basement of the Adirondacks to the high-grade Devonian gneisses of New Hampshire. Because JBT's field trips were across strike traverses the emphasis was both on the metamorphism and tectonics of the Appalachians. During the organization of the field trip that we are now on we discussed whether or not to run the same field trip in honor of JBT. But, as rebellious students of, or students of students of, JBT's we decided to break the tradition and run the field trip along tectonic grain within the same metamorphic terrane. For what purpose ?

The purpose of the trip is to examine the metamorphic processes that operate at different levels in the crust within a single orogen. Kearsarge - Central Maine synclinorium or Merrimack synclinorium was selected because the rocks range from those metamorphosed at depths of ~13 km exposed in Maine to those metamorphosed at depths of ~25 km exposed in central Massachusetts. Because of the present interest and enthusiasm of the geologic community concerning the subject of the role of volatiles in the crust, special emphasis was given to this subject. Some of the trips, however, do not deal directly with metamorphic fluids but are necessary to point out the different styles of metamorphism that occur in an along-strike traverse.

It is our hope that by assembling a large number of experts in metamorphic petrology we will generate discussion and foster new ideas about metamorphic processes in the crust. If nothing else it will be a good time to see old friends and make new ones.

C. Page Chamberlain
and
Peter Robinson

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REGIONAL METAMORPHISM AND FLUID-ROCK INTERACTION IN A BUCHAN TERRANE, WATERVILLE-AUGUSTA AREA, MAINE

John M. Ferry

and

Philip H. Osberg

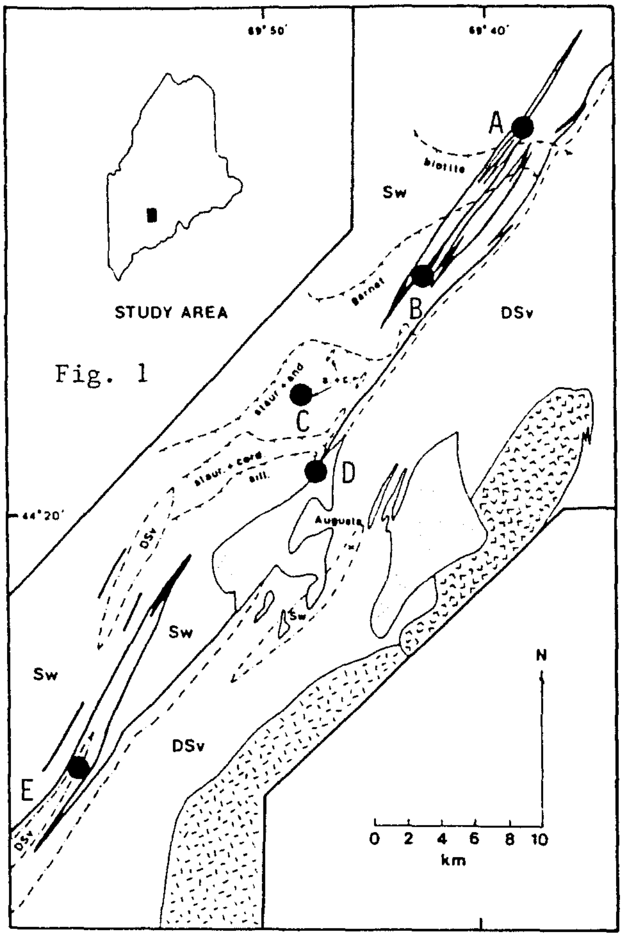
INTRODUCTION

This trip will enable participants to examine progressive metamorphism and the effects of synmetamorphic fluid-rock interaction in a Buchan-type terrane in the vicinity of Waterville and Augusta. All of the field stops are within the Waterville and the Sangerville Formations (Osberg, 1968; 1988). The parts of the Sangerville Formation we will visit have formerly been named the Vassalboro Formation. The Waterville Formation consists of pelite in beds 1-4 cm thick intercalated with quartzite and locally slightly calcareous quartzite in beds a few millimeters to 2 cm thick. A prominent limestone member forms a distinctive mapping unit within the Waterville Formation. The Sangerville Formation is represented by slightly calcareous arenite in beds 15-40 cm thick and generally dark pelite in beds 2-10 cm thick. Locally the arenite and pelite form graded beds.

The metamorphism in the Waterville-Augusta area is the product of at least two and possibly three distinct local thermal events, m1, m2, and m3(?) (Novak and Holdaway, 1981; Holdaway et al., 1982; Osberg, 1988). M1 is overprinted by m2 everywhere in the Waterville-Augusta area. Even so, m1 is recognized throughout the area by the alignment of muscovite, and to a lesser extent chlorite and/or biotite, in a well-developed schistosity that is approximately parallel to the axial surfaces of large upright, tight to isoclinal folds in the stratified rocks. Whether or not m1 displayed progressive changes in mineralogy in response to changes in intensive parameters is unknown due to the imposition of m2. The schistosity that is defined by muscovite, chlorite, and biotite of m1 origin is cut by up to meter-thick dikes of metamorphosed andesite and dacite and by the plutons of the region.

M1 predates the emplacement of the Hallowell and Togus plutons, and their ages provide a minimum age for m1. Dallmeyer and Van Breeman (1981) dated the emplacement of these plutons by Rb-Sr whole rock isochrons, giving the following results: Togus, 394$\pm$8 Ma and Hallowell, 387$\pm$11 Ma.

M2 is responsible for the progressive changes in mineralogy throughout the Waterville-Augusta area. Low-grade metamorphic rocks are present north of Waterville, but to the south isograds have been mapped on the first appearance of biotite, garnet, staurolite + andalusite, staurolite + cordierite, and sillimanite in pelitic schists of the Waterville Formation (Osberg, 1968, 1988; Figure 1), and on the first appearance of biotite, Ca-amphibole, clinozoisite, and diopside in the carbonate-bearing rocks of the



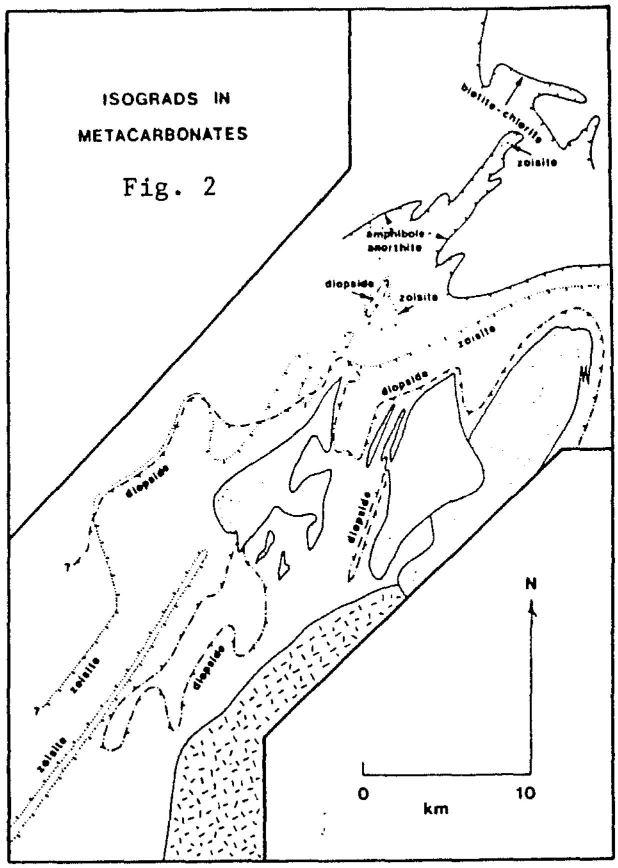


Figure 1. Geologic sketch map of the Waterville-Augusta area, Maine. Sw = Waterville Formation; DSv = Vassalboro Formation (now considered equivalent to the Sangerville Formation; black = limestone member of Waterville Formation. Stippled pattern: granite stocks; random check pattern: granodiorite stock; random dash pattern: mostly Hutchins Corner Formation. Dash-dot lines: stratigraphic boundary between Waterville and Vassalboro (Sangerville) Formations; dashed lines: isograds mapped from assemblages in pelitic schists (staur. + and. = staurolite + andalusite; staur. + cord., s. + c. = staurolite + cordierite; Sill. = sillimanite). From Ferry (1980a) and Osberg (1988). Field trip will visit locations A-E marked by large circles.

Figure 2. Isograd map for metamorphosed carbonate rocks (from Ferry, 1980a). Other features as in Fig. 1.

Sangerville Formation and the limestone member of the Waterville Formation (Ferry, 1976a, 1980a; Fig. 2). Prophyroblasts produced in m2 cut across the schistosite containing the m1 cuscovite in a generally static fashion so that elements of the early schistosite are commonly preserved within the porphyroblasts. Moreover, a second population of muscovite plates, smaller than those produced in m1, define a second cleavage in the rock that makes an angle to the dominant schistosity. Elements of the second cleavage are never found in m2 porphyroblasts. Compositions of muscovite, chlorite, and biotite are always uniform throughout a rock sample regardless of whether they are situated in the dominant schistosity or in the second cleavage, indicating that the minerals of m1 origin have been reconstituted in m2.

The andesitic and dacitic dikes are metamorphosed to mineral assemblages appropriate to m2. Furthermore, the dikes are cut by the second cleavage, and m2 minerals are aligned in this cleavage. Both the stratified rocks and these dikes are deformed by small folds that have axial surfaces essentially parallel to the second cleavage.

M2 isograds, although they have an approximate relationship to plutons, locally cut across them. Osberg (1968) mapped the sillimanite isograd as cutting through the north end of the Hallowell pluton, and Ferry (1976a) mapped a scapolite isograd through the Togus pluton on the basis of mineral assemblages found in xenoliths of country rock. Thin sections of both the Hallowell and Togus plutons show that the original fabric has been altered by granulation and partial recrystallization producing a mortar board texture. Alteration of biotite to chlorite and plagioclase to secondary muscovite + calcite and/or epidote are common. Compositions of microcline and coexisting plagioclase typically record temperatures of 400°-450°C, and the oxygen isotopes have been altered from likely primary igneous values to as high as $\delta^{18}O = 13.8$ ‰ (Ferry, 1978; Rumble et al., 1986). Finally, thin dikes of Hallowell granite cut across upright folds and dominant schistosity, but are themselves folded and cut by the second cleavage (Osberg, 1988). Thus, the mineralogical, textural, isotopic, and structural data all are consistent with m2 recrystallization of the Hallowell and the Togus plutons and significant granite-fluid interaction at subsolidus temperatures.

The age of m2 is best documented by West et al. (1988), who have dated hornblendes from the Casco Bay Group by the $^{40}Ar/^{39}Ar$ method. The Casco Bay Group is thought to be Early Ordovician or older (Hussey, 1988), but in its northern outcrop to have been recrystallized by m2. Hornblendes collected from Palermo, northeast of Augusta, record plateau ages of 372±4 and 368±4 Ma. M2 hornblende from the Sangerville Formation just north of Augusta gives a pleateau age of 374±4 Ma (West, 1989, written communication).

Pressure at the peak of m2 in the staurolite + andalusite and sillimanite zones was near 3500 bars. Temperature ranged from near 400°C in the chlorite zone in the northern part of the area to 550°-575°C in the southern part of the area.

M3(?) has been postulated by Novak and Holdaway (1981) and Holdaway et al. (1982) to affect rocks to the south and west of Augusta. Most of these rocks contain sillimanite or sillimanite + K-feldspar. Some are retrograded to chlorite-bearing assemblages. The main arguments for m3(?) west and south of Augusta are (1) too many phases for the number of components that are assumed to control the system and (2) the presence of pseudomorphed andalusite and staurolite. More work needs to be directed at determining the relationships between m3(?) and previous metamorphic assemblages.

Numerous petrologic studies of the metamorphic and plutonic rocks in the area have been made by Barker (1964), Osberg (1968, 1971, 1974a,b, 1988), Ferry (1976a,b, 1978, 1979a,b, 1980a,b, 1981, 1982, 1983a,b,c,d,e, 1984, 1986, 1987, 1988), Novak and Holdaway (1981), Holdaway et al. (1982), Rumble et al. (1986), and Cashman and Ferry (1988).

DESCRIPTIONS OF LOCALITIES

The excursion consists of five localities which are shown in Figure 1. Three of the stops, A, B, and D, are intended to show progressive metamorphic changes in the limestone member of the Waterville Formation. Stop C is in the pelitic part of the Waterville Formation and stop E is in the Sangerville Formation. A road log accompanies the description of the localities.

0.0 miles. Stop A (0.2 km south of Waterville-Winslow bridge on east shore of Kennebec River; locality 7 in Ferry's publications).

The limestone member and adjacent pelites of the Waterville Formation are exposed near the hinge of a downward-facing, northward-plunging, upright antiform. Rocks at this location constitute the "baseline" against which all effects of progressive metamorphism at higher grades are measured. The outcrop is located in the chlorite zone with respect to isograds in pelitic schists and at the biotite isograd with respect to impure carbonate rocks. Most samples of the limestone member contain ankerite + calcite + muscovite + albite + quartz + rutile + graphitic material. Approximately a third of the carbonates contain traces of biotite in addition which developed by the reaction:

(W1) muscovite + ankerite + albite + HCl = biotite + anorthite (in plagioclase) + quartz + calcite + CO_2 + H_2O + NaCl.

Two felsic dikes cut across upright folds and dominant schistosity in the limestone member, but are in turn cut by a second cleavage along which micaceous minerals are aligned. The dikes have been metamorphosed to albite + quartz + muscovite + chlorite + carbonate, and were, therefore, intruded prior to the peak of low-grade metamorphism at this locality.

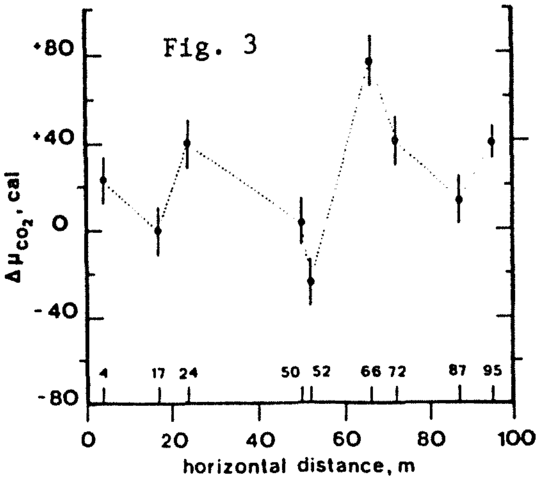
Pelitic schists of the Waterville Formation adjacent to the limestone member contain muscovite + chlorite + albite + quartz + ankerite + siderite + rutile + graphitic material. Although the pelites immediately adjacent to the limestone member are slightly atypical of the larger part of the Waterville Formation, carbonates are ubiquitous in all lithologies in the lowest-grade portion of the formation. Pelites more representative of the Waterville Formation are thinner bedded and contain less abundant carbonates and graphitic meterial than do those exposed at this locality.

Return to vehicles. Proceed north on Route 201.

0.2 mile. Stop light. Turn left across Waterville-Winslow bridge.

0.1 mile. Stop light. Turn left onto Water Street.

0.9 mile. Turn right onto Grove Street.



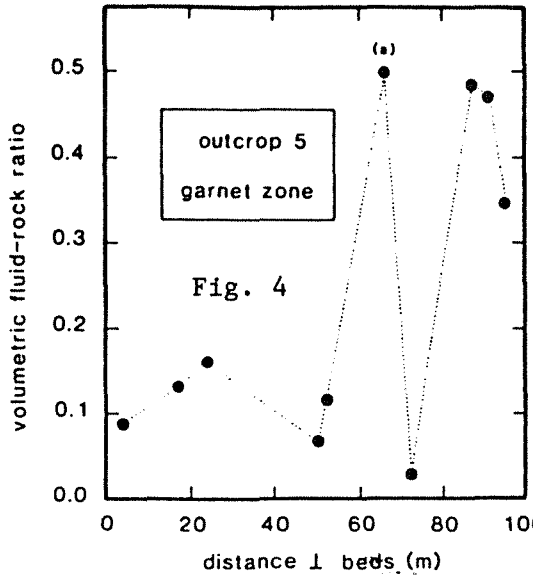
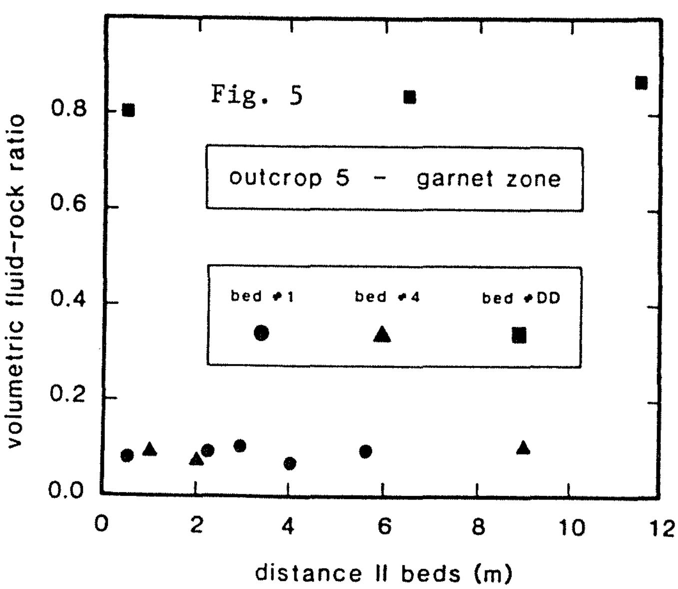


Figure 3. Differences in the chemical potential in CO2 between samples collected along a traverse across layering at outcrop 5 (stop B). All values arbitrarily taken with reference to chemical potentials recorded by sample 5-3. Each symbol corresponds to a single sample from a different lithologic layer. Vertical bars represent the estimated error in each measurement. From Ferry (1979b).

Figure 4. Calculated fluid-rock ratios for samples collected along a traverse across bedding at outcrop 5 (stop B). Same traverse as shown in Fig. 3. From Ferry (1987).



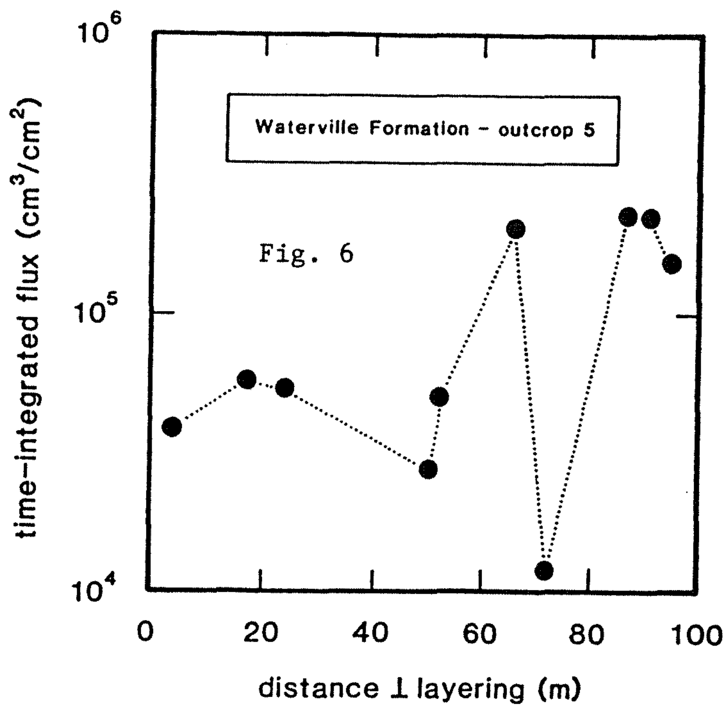


Figure 5. Calculated fluid-rock ratios for samples collected along three traverses parallel to bedding at outcrop 5 (stop B). Each data point refers to a single sample. From Ferry (1987).

Figure 6. Calculated time-integrated fluxes for samples collected along a traverse across bedding at outcrop 5 (stop B). Same traverse as in Figs. 3 and 4. From Baumgartner and Ferry (1989).

0.5 mile. Stop sign. Turn left on Silver Street. At bridge Silver Street becomes Kennedy Drive. Proceed west on Kennedy Drive.

1.6 miles. Route 95. Proceed south on Route 95 toward Augusta.

6.9 miles. Exit 32. Turn off of Route 95 and continue across Lyons Road into Blue Rock quarry property.

0.1 mile. Stop B. Blue Rock quarry (location 5 in Ferry's publications).

The quarry exposes the limestone member of the Waterville Formation near the hinge of a large upward-facing, south-plunging, upright antiform. Several nearly isoclinal, flattened, parasitic folds are exposed in the quarry. The locality lies within the garnet zone with respect to isograds in pelitic schists and in the amphibole zone with respect to impure carbonate-bearing rocks.

Almost all lithologic layers contain the assemblage ankerite + calcite + calcic plagioclase + quartz + muscovite + chlorite + biotite + rutile + graphitic material. Rarely the biotite-rich layers lack muscovite. A few carbonate layers contain calcic amphibole + ankerite + calcite + calcic plagioclase + quartz + chlorite + biotite + rutile + graphitic material, and they can be identified by their nubbly weathered surfaces. Amphibole developed by the following reaction:

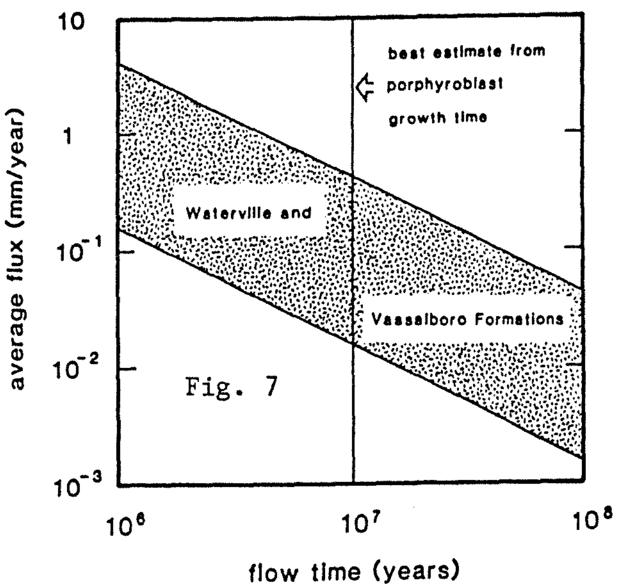
(W2) biotite + ankerite + quartz + plagioclase + H_2O = amphibole + calcite + CO_2.

Minor pelitic layers are composed of muscovite + biotite + chlorite + garnet + plagioclase + quartz + ilmenite.

The Fe/(Fe+Mg) have characteristic values for biotite, chlorite, muscovite, calcite, and ankerite that are constant within each layer, but they differ from layer to layer. These differences in composition have been quantitatively interpreted to record differences in the chemical potentials of CO_2, H_2O, and all other C-O-H fluid species at the peak of metamorphism (Fig. 3). These relationships require the buffering of metamorphic fluid composition by mineral-fluid equilibria on a layer-to-layer scale.

The amount of biotite also varies dramatically from layer to layer. These differences in biotite content have been interpreted to be due to differences in extent of reaction (W1), which must be controlled by differences in fluid-rock ratio (Fig. 4); fluid-rock ratios vary in the range 0.03-1.5 over the outcrop. Biotite contents and calculated fluid-rock ratios, however, are uniform within a given lithologic layer (Fig. 5). The pattern suggests that metamorphic fluid flow was channelized into more permeable layers that acted as metamorphic aquifers (the biotite-rich layers in the outcrop). The intervening, less permeable layers acted as metamorphic aquitards (biotite-poor layers).

Using a new model for fluid-rock interaction, biotite contents of metamorphosed carbonate rocks can be interpreted as measurements of time-integrated fluxes. Time-integrated fluxes at this locality vary between 10^4 cm^3/cm^2 in the biotite-poor aquitards to $4x10^5$ cm^3/cm^2 in the biotite-rich aquifers (Fig. 6). Average fluxes during metamorphism of 0.01-1 mm/yr and average permeabilities of 0.01-1 microdarcies (Figs. 7 and 8) can be estimated using a reasonable duration of regional metamorphism and acceptable values for



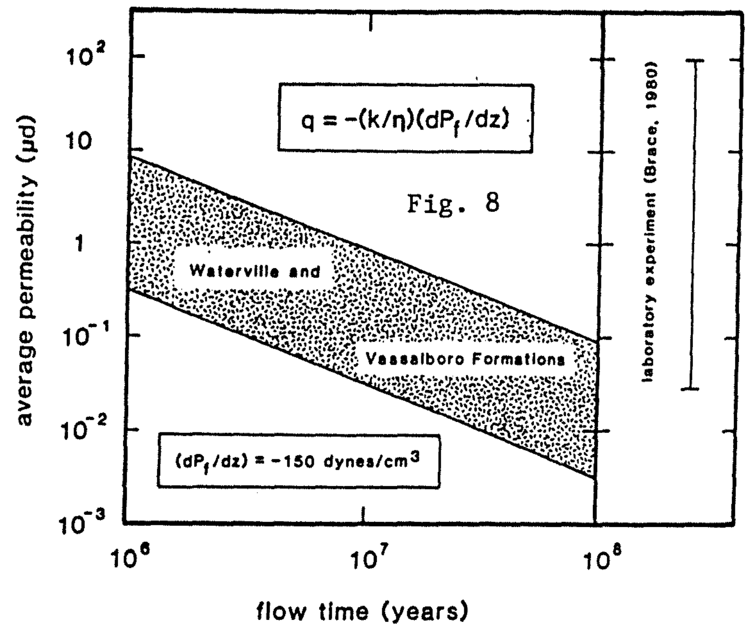


Figure 7. Calculated average flux through the Waterville and Vassalboro (now Sangerville) Formations as a function of total flow time during metamorphism. Patterned area encompasses results for all samples studied from outcrop 5 (stop B) and two other outcrops of the Vassalboro (Sangerville) Formation nearby. Time of porphyroblast growth taken from Christensen et al. (1989). From Baumgartner and Ferry (1989).

Figure 8. Range in average permeability of the Waterville and Vassalboro (Sangerville) Formations as a function of total flow time calculated from Darcy's Law and the average fluxes in Fig. 7. Fluid viscosity from Walther and Orville (1982) and fluid pressure gradient from Norton and Taylor (1979). Patterned area encompasses results from all samples studied from outcrop 5 and two outcrops of the Vassalboro (Sangerville) Formation nearby. From Baumgartner and Ferry (1989).

Fig. 9

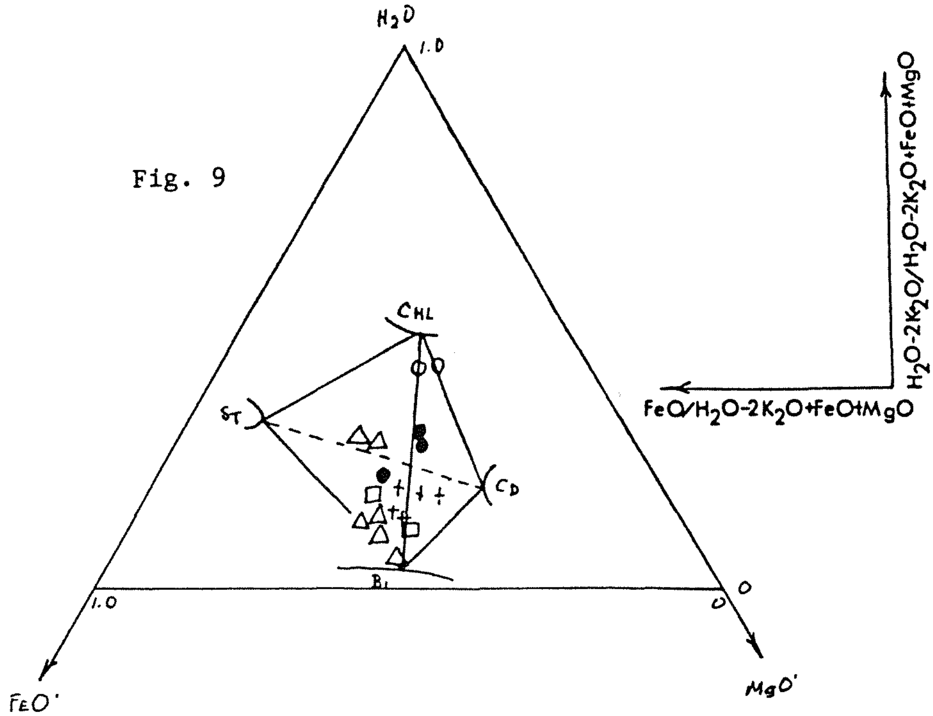


Figure 9. Al2O3-projection showing bulk chemical analyses of Waterville pelites. Open circles--chlorite zone; closed circles--garnet zone; triangles--staurolite-bearing rocks in staurolite-cordierite zone; squares--cordierite-bearing rocks in staurolite-cordierite zone; and crosses--sillimanite zone. chemographic relations between chlorite, staurolite, cordierite, and biotite are also shown for locality D.

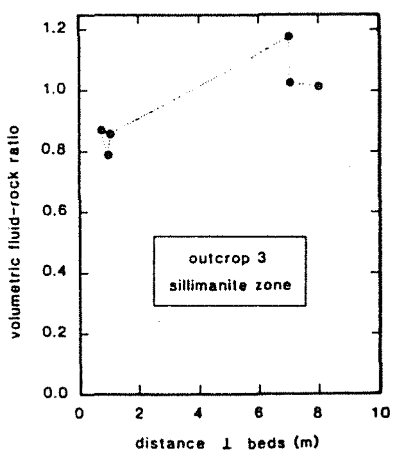


Fig. 10. Calculated fluid-rock ratios for samples collected along a traverse across bedding in outcrop 3 (stop D). Each data point refers to a single sample. From Ferry (1987).

synmetamorphic fluid pressure gradients. These are the first estimates for fluxes and permeabilities during regional metamorphism made from petrologic data.

Turn right onto Lyons Road and proceed west.

0.9 mile. Stop sign. Turn left onto Middle Road, travelling south.

2.1 miles. Center of Old Sidney village. Turn right onto Shepard Road.

1.0 mile. Turn left at "T"-intersection onto Quaker Hill Road.

2.3 miles. Stop sign. Turn left onto Route 23.

0.6 mile. Stop sign. Bear left onto Routes 8 & 11.

0.1 mile. Stop C. West Sidney. Outcrop is on left hand side of road behind telephone pole.

The outcrop is Waterville Formation. Alternating thin beds of quartzite and pelite, so typical of the formation, are clearly displayed. Less evident are packets of beds, 10-30 cm thick, that contain slightly different assemblages of minerals due to differing bulk compositions. A nearly isoclinal fold located directly behind the telephone pole produces a duplication of the beds. Two calc-silicate-bearing beds in the south end of the outcrop were originally calcareous quartzites.

Beds in the north part of the outcrop and again just south of the telephone pole contain quartz + plagioclase + muscovite + biotite + chlorite + garnet + staurolite + cordierite + andalusite + ilmenite + sulfide + minor magnetite. Garnet is small and euhedral. Staurolite and andalusite stand out in positive relief, whereas cordierite weathers out in negative relief. Chlorite occurs as well developed porphyroblasts scattered through the rock and locally concentrated at the perimeter of cordierite crystals. The MgO/(FeO+MnO+MgO) are: cordierite = 0.639, chlorite = 0.521, biotite = 0.487, staurolite = 0.151, and garnet rim = 0.090. Beds in the southern part of the outcrop contain fewer minerals, and these minerals have slightly different Mg/(Fe+Mn+Mg) compositions when compared to the beds containing 12 phases.

Osberg (1971, unpublished paper) suggested that the 12-phase beds contain a (c_i+1)-phase assemblage in the system SiO_2-TiO_2-Al_2O_3-FeO-MnO-MgO-CaO-Na_2O-K_2O-H_2O-SO_2. All of these components are thought to be internally buffered. Although progressive metamorphism generally involves dehydration and decarbonation, this is not necessarily so in middle grade reactions in terranes that have previously been dehydrated by a prior episode of metamorphism. Figure 9 shows bulk compositions of Waterville pelites from various grades of metamorphism plotted on a Rumble projection. Rocks of the greenschist facies are more hydrated than the middle grade rocks, but the latter show no pronounced trend for dehydratio regardless of the reactions that have taken place. It is not until the middle of the sillimanite zone that a dehydration trend (not plotted on the diagram) becomes apparent again. Therefore, in these circumstances, some reactions may conserve H_2O. The development of porphyroblastic chlorite in these rocks is interpreted to be a result of such a reaction. The chemographic relations are shown in Figure 9. Dehydration of the rock does not occur unless the buffering capacity of the rock is exceeded.

An alternate explanation for the mineral assemblages in these rocks is suggested by the work of Novak and Holdaway (1981). They have interpreted somewhat similar rocks to be polymetamorphic and to contain relict minerals. Chlorite, in this case, would be a retrograde mineral produced by an m3(?) event. Although chlorite locally has a reaction relationship to cordierite, its MgO/(FeO+MnO+MgO) is not equal to that of cordierite and so other minerals are involved in the reaction. Moreover, chlorite is in apparently stable association with all other minerals including garnet, staurolite, biotite, and andalusite. I interpret the demise of cordierite to be due to the switch in tie-lines as indicated in Figure 9.

Continue south on Routes 8 & 11.

2.8 miles. Turn right onto Leighton Road.

0.9 mile. Turn left onto Townsend Road.

0.2 mile. Park at end of road and walk into locality for Stop D (location 3 in Ferry's publications).

The outcrop is in the limestone member of the Waterville Formation approximately within 50 m of its contact with the Hallowell granite. The locality is in the sillimanite zone with respect to isograds in pelitic schists and in the diopside zone with respect to impure carbonates. With very few exceptions, carbonate rocks have recrystallized to marbles that contain diopside + calcic amphibole + calcite + calcic plagioclase + quartz + clinozoisite + K-feldspar + biotite + sphene + graphite. Diopside, clinozoisite, K-feldspar, and sphene developed by the reaction:

(W3) biotite + calcite + quartz + albite (in plagioclase) = diopside + K-feldspar + clinozoisite + sphene + amphibole + anorthite (in plagioclase) + H_2O + CO_2.

The amounts of decarbonation in all samples collected from this location are very similar. The similarity has been quantitatively interpreted as a relative uniformity in fluid-rock ratio throughout the outcrop (Fig. 10). Evidently the obvious distinction between metamorphic aquifers and metamorphic aquitards seen at stop B has disappeared. With increasing grade of metamorphism in the stratigraphic unit, metamorphic fluid flow appears to have become less channelized and more pervasive and uniform.

Retrace steps to transportation and proceed west on Townsend Road.

0.2 mile. Turn left on Leighton Road.

1.9 miles. Yield sign. Bear right on Manchester Road.

0.2 mile. Stop sign. Turn right onto Route 11 toward Manchester.

2.1 miles. Stop light. Turn left onto Pond Road.

8.2 miles. Stop sign. Bear right to Routes 9 & 126.

0.1 mile. Litchfield Corners. Bear left to Routes 9 & 126 towards Tacoma Lakes.

1.9 miles. Stop sign in Batchelders Corner. Turn right onto Routes 9 & 126.

1.9 miles. Stop E, just west of Tacoma Lakes (location 987 in Ferry's publications).

The outcrop contains Sangerville Formation, which here is caught in the core of a large, tight, upright syncline. Waterville pelite is exposed both east and west of this outcrop, and a thin unit of carbonaceous sulfidic pelite, stratigraphically at the base of the Sangerville, lies at the west end of the outcrop. A large pegmatite intrudes the central part of the exposure.

The calc-silicate lenses do not define original bedding heterogeneities, but must be produced by metamorphic processes.

The Sangerville Formation in this outcrop is in the sillimanite zone with respect to isograds in pelitic schists and in the diopside zone with respect to impure carbonates. Almost all impure carbonate rocks contain diopside + calcic amphibole + calcic plagioclase + calcite + quartz + clinozoisite + sphene. Rarely they contain Ca-Fe-Mn garnet or lack diopside. Interlayered psammitic rocks contain biotite + plagioclase + quartz +/- muscovite. With progressive metamorphism carbonate rocks in the Sangerville Formation developed biotite by the reaction:

(S1) muscovite + chlorite + ankerite + ilmenite + albite + HCl = biotite + anorthite (in plagioclase) + quartz + calcite + CO_2 + H_2O + NaCl.

Calcic amphibole is produced by the reaction:

(S2) muscovite + biotite + calcite + plagioclase + HCl = amphibole + quartz + ilmenite + CO_2 + H_2O + NaCl + KCl,

and diopside by the reaction:

(S3) biotite + calcite + plagioclase + quartz + ilmenite + HCl = diopside + clinozoisite + sphene + amphibole + CO_2 + H_2O + NaCl + KCl.

Measured progress of reactions (S1) - (S3) in samples from this locality have been quantitatively interpreted in terms of volumetric fluid-rock ratios (Fig. 11). Calculated values fall in the narrow range 1.7 - 2.0, and fluid flow during metamorphism, therefore, was pervasive and uniform. Study of other exposures of the Sangerville Formation from low to high grade show similar results; taken together they indicate that metamorphic fluid flow was uniform through the formation. The aquifers observed in the limestone member of the Waterville Formation evidently did not develop in the Sangerville Formation. The difference in behavior is likely related to the relative lithologic homogeneity of the Sangerville and the relative lithologic heterogeneity of the Waterville Formation.

There is a dramatic decrease in the alkali content of the carbonate rocks from the Sangerville Formation with increasing grade of metamorphism (Fig. 12). The mechanisms of alkali loss are the prograde reactions (S1) - (S3), which involve hydrolysis as well as decarbonation and dehydration. Mineralogically the K-loss is expressed by the occurrence of muscovite in all low-grade samples and the absence of any K-rich mineral (e.g., biotite or K-feldspar) in carbonate rocks from this location. Mineralogically the Na-loss is expressed by the occurrence of albite in low grade samples and the occurrence of bytownite or anorthite in samples from this locality (the amount of plagioclase does not significantly change with increasing metamorphic grade). The prograde depletion of the Sangerville Formation in K and Na is independent geochemical evidence for infiltration of the formation by substantial quantities of reactive fluid during regional metamorphism. In contrast to the chemical behavior of the Sangerville Formation, the limestone member of the Waterville Formation shows no evidence for less of K during metamorphism and only equivocal evidence for loss of Na (Fig. 12). The difference in behavior may be explained by (a) greater Cl-content of metamorphic fluids in the Sangerville Formation and/or (b) different temperature dependence of fluid composition in the two formations.

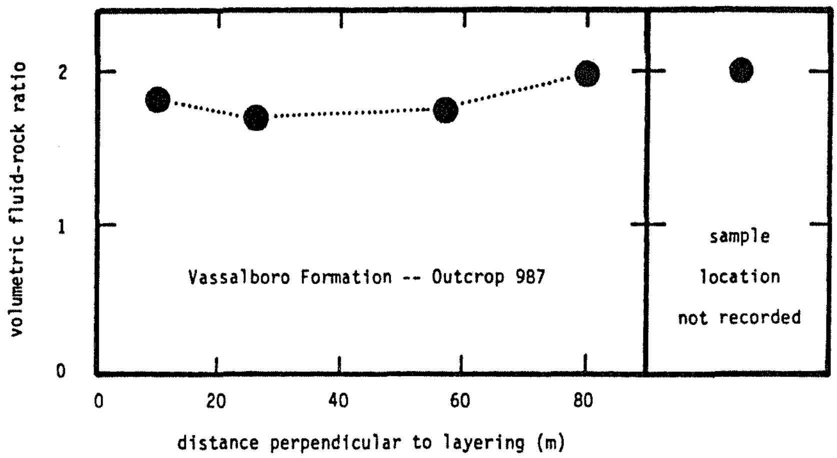


Figure 11. Calculated fluid-rock ratios for samples collected across lithologic layering in outcrop 987 (stop E). Each data point refers to a single sample. From Ferry (1983c).

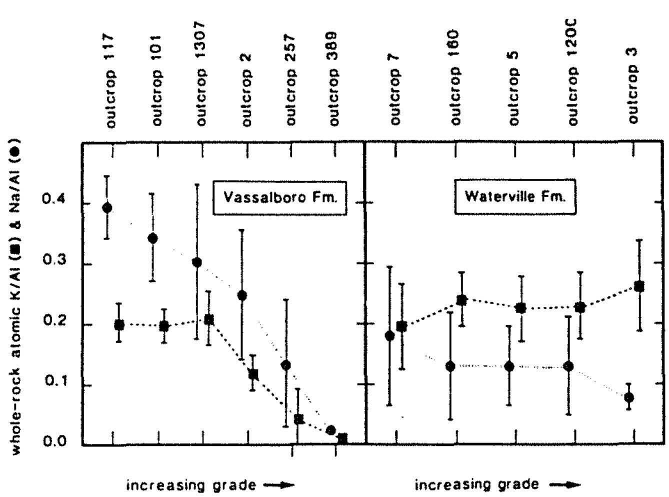


Figure 12. Average whole-rock atomic K/Al and Na/Al for metamorphosed carbonate rocks from 11 outcrops in the Vassalboro (Sangerville) Formation and the limestone member of the Waterville Formation. Symbol is mean value for all samples from each outcrop; length of bar represents 2 standard deviations. From Ferry (1988).

REFERENCES

Barker, D.S. (1964) The Hallowell granite, south-central Maine. Am. J. Sci., 262, 592-613.

Baumgartner, L.P. and Ferry, J.M. (1989) A physical model for infiltration-driven mixed volatile reactions and its application to regional metamorphism. Contrib. Mineral. Petrol. (in review).

Cashman, K.V. and Ferry, J.M. (1988) Crystal size distribution (CSD) in rocks and the kinetics and dynamics of crystallization. III. Metamorphic crystallization. Contrib. Mineral. Petrol., 99, 401-415.

Christensen, J.N., Rosenfeld, J.L., and DePaolo, D.J. (1989) Rates of tectonometamorphic processes from rubidium and strontium isotopes in garnet. Science, 244, 1465-1469.

Dallmeyer, R.D. and Van Breeman, O. (1981) Rb-Sr whole-rock and $^{40}K/^{39}Ar$ mineral ages of the Togus and Hallowell quartz monzonite and Three Mile Pond granodiorite plutons, south-central Maine: Their bearing on post-Acadian cooling history. Contrib. Mineral. Petrol., 78, 61-73.

Ferry, J.M. (1976a) Metamorphism of calcareous sediments in the Waterville-Vassalboro area, south-central Maine: Mineral reactions and graphical analysis. Am. J. Sci., 276, 841-882.

Ferry, J.M. (1976b) P, T, f_{CO_2}, and f_{H_2O} during regional metamorphism of calcareous sediments from the Waterville-Vassalboro area, south-central Maine. Contrib. Mineral. Petrol., 57, 119-143.

Ferry, J.M. (1978) Fluid interaction between granite and sediment during metamorphism, south-central Maine. Am. J. Sci., 278, 1025-1056.

Ferry, J.M. (1979a) Reaction mechanisms, physical conditions, and mass transfer during hydrothermal alteration of mica and feldspar in granitic rocks from south-central Maine, USA. Contrib. Mineral. Petrol., 68, 125-139.

Ferry, J.M. (1979b) A map of chemical potential differences within an outcrop. Am. Mineral., 64, 966-985.

Ferry, J.M. (1980a) A case study of the amount and distribution of heat and fluid during metamorphism. Contrib. Mineral. Petrol., 71, 373-385.

Ferry, J.M. (1980b) A comparative study of geothermometers and geobarometers in pelitic schists from south-central Maine. Am. Mineral., 65, 720-732.

Ferry, J.M. (1981) Petrology of graphitic sulfide-rich schists from south-central Maine: An example of desulfidation during prograde regional metamorphism. Am. Mineral., 66, 908-930.

Ferry, J.M. (1983a) Mineral reactions and element migration during metamorphism of calcareous sediments from the Vassalboro Formation, south-central Maine. Am. Mineral., 68, 334-354.

Ferry, J.M. (1983b) A comparative geochemical study of pelitic schists and metamorphosed carbonate rocks from south-central Maine, USA. Contrib. Mineral. Petrol., 80, 59-72.

Ferry, J.M. (1983c) Regional metamorphism of the Vassalboro Formation, south-central Maine, USA: A case study of the role of fluid in metamorphic petrogenesis. J. Geol. Soc. Lond., 140, 551-576.

Ferry, J.M. (1983d) On the control of temperature, fluid composition, and reaction progress during metamorphism. Am. J. Sci., 283-A, 201-232.

Ferry, J.M. (1983e) Applications of the reaction progress variable in metamorphic petrology. J. Petrol., 24, 343-376.

Ferry, J.M. (1984) A biotite isograd in south-central Maine, U.S.A.: Mineral reactions, fluid transfer, and heat transfer. J. Petrol., 25, 871-893.

Ferry, J.M. (1986) Infiltration of aqueous fluid and high fluid:rock ratios during greenschist facies metamorphism: A reply. J. Petrol., 27, 695-714.

Ferry, J.M. (1987) Metamorphic hydrology at 13-km depth and 400-550°C. Am. Mineral., 72, 39-58.

Ferry, J.M. (1988) Contrasting mechanisms of fluid flow through adjacent stratigraphic units during regional metamorphism, south-central Maine, USA. Contrib. Mineral. Petrol., 98, 1-12.

Holdaway, M.J., Guidotti, C.V., Novak, J.M., and Henry, W.E. (1982) Polymetamorphism in medium- to high-grade pelitic metamorphic rocks, west-central Maine. Bull. Geol. Soc. Am., 93, 572-584.

Hussey, Am. II (1988) Lithotectonic stratigraphy, deformation, plutonism, and metamorphism, greater Casco Bay area, southwestern Maine. In: Tucker, R.P. and Marvinney, R.G. (eds.) Studies in Maine Geology, Vol. 1, Me. Geol. Surv., p. 17-34.

Norton, D. and Taylor, H.P., Jr. (1979) Quantitative simulation of the hydrothermal systems of crystallizing magmas on the basis of transport theory and oxygen isotope data: An analysis of the Skaergaard intrusion. J. Petrol., 20, 421-486.

Novak, J.M. and Holdaway, M.J. (1981) Metamorphic petrology, mineral equilibria, and polymetamorphism in the Augusta Quadrangle, south-central Maine. Am. Mineral., 66, 51-69.

Osberg, P.H. (1968) Stratigraphy, structural geology, and metamorphism of the Waterville-Vassalboro area, Maine. Maine Geol. Surv. Bull. 20.

Osberg, P.H. (1971) An equilibrium model for Buchan-type metamorphic rocks, south-central Maine. Am. Mineral., 56, 570-586.

Osberg, P.H. (1974a) Isochemical metamorphism, south-central Maine. Trans. Am. Geophys. U., 55, 450.

Osberg, P.H. (1974b) Buchan-type metamorphism of the Waterville pelite, south-central Maine. In: Osberg, P.H. (ed.) Geology of East-Central and North-Central Maine. Univ. Me., Orono, p. 212-222.

Osberg, P.H. (1979) Geologic relationships in south-central Maine. In: Osberg, P.H. and Skehan, J.W. (eds.) The Caledonides in the U.S.A., Weston Observ. Boston Coll., p. 37-62.

Osberg, P.H. (1988) Geologic relations within the shale-wacke sequence in south-central Maine. In: Tucker, R.P. and Marvinney, R.G. (eds.) Studies in Maine Geology, Vol. 1, Me. Geol. Surv., p. 51-74.

Rumble, D., Ferry, J.M., and Hoering, T.C. (1986) Oxygen isotope geochemistry of hydrothermally altered granitic rocks from south-central Maine, USA. Contrib. Mineral. Petrol., 93, 420-428.

West, D.P., Jr., Lux, D.R., and Hussey, A.M., Jr. (1988) $^{40}Ar/^{39}Ar$ hornblende ages from southwestern Maine: Evidence for late Paleozoic metamorphism. Maritime Seds. Atlantic Geol., 24, 225-239.

Walther, J.V. and Orville, P.M. (1982) Volatile production and transportation in regional metamorphism. Contrib. Mineral. Petrol., 79, 252-257.

METAMORPHISM IN WESTERN MAINE: AN OVERVIEW

C.V. Guidotti

and

J.T. Cheney

Introduction

The goal of this write up is to provide an overview of the metamorphism in Western Maine (Fig. 1) that coincides with the context of this field trip guide. Specifically, we shall first provide a brief history of metamorphic studies in the area (supplemented by references). Then we will describe the styles of the three main metamorphic events that have occurred in Western Maine so that they can be compared more closely with the styles that have occurred directly along strike to the SW in the orogen. Finally, we will provide a brief description of the metamorphism in the East Andover 7 1/2' quadrangle plus a road log for several outcrops there which show some aspects of the metamorphic style in Western Maine.

History of Metamorphic Studies

The first modern metamorphic study published on part of this area was by Green (1963). In addition, several Harvard Ph.D. studies were carried out within this area, Green (1960), Milton (1961), Guidotti (1963), Pankiwskyj (1964), Harwood (1966), Warner (1967), and Drake (1968). Professor Thompson was directly involved in advising all of these thesis studies.

During the decade of the 60's a number of metamorphic mineralogy-petrology studies were published on various portions of Western Maine (see supplemental reference lists, SRL). The studies of Guidotti and Evans (1964) and Evans and Guidotti (1966) are notable because they were among the first studies using electron microprobe data in a broad scale petrologic study. Also notable was the paper of Thompson and Norton (1968) which showed that the metamorphism in Western Maine forms the northern end of a high-grade terrane that extends southward to Long Island Sound. Moreover, it specifically called attention to the spatial relationships between the granitic plutons and high-grade metamorphism in Western Maine.

In the 1970's a number of detailed metamorphic mineralogy-petrology studies were carried out by Guidotti and students from the University of Wisconsin (see SRL). Notable results from these studies were the recognition of polymetamorphism, (though first suggested by Pankiwskyj, 1964) and description of very systematic compositional variation of solid solution phases in response to metamorphic grade changes. In addition, the report by Guidotti (1970A) represented an early attempt to use the procedures developed by Eugster and Skippen (1967) for calculating the composition of metamorphic fluids.

Other notable aspects of studies in the 1970's include Foster's (1975) and (1977) use of irreversible thermodynamics to ascertain the detailed ionic transfer reactions that produced the textures in sillimanite grade rocks of the Rangeley area. In addition, the work of Burruss (1977) and Hollister et al.(1979) represented an attempt to interpret the fluid inclusions present in high-grade metapelites.

In the 1980's the efforts aimed at understanding the metamorphic mineralogy and petrology increased markedly with the development of a major program by M.J. Holdaway and students from S.M.U. Their efforts were focussed east of the area shown in Fig. 1. Of some note is that this work served as the foundation for much of the recent mineralogical work on staurolite by Holdaway, Dutrow, and colleagues (see SRL).

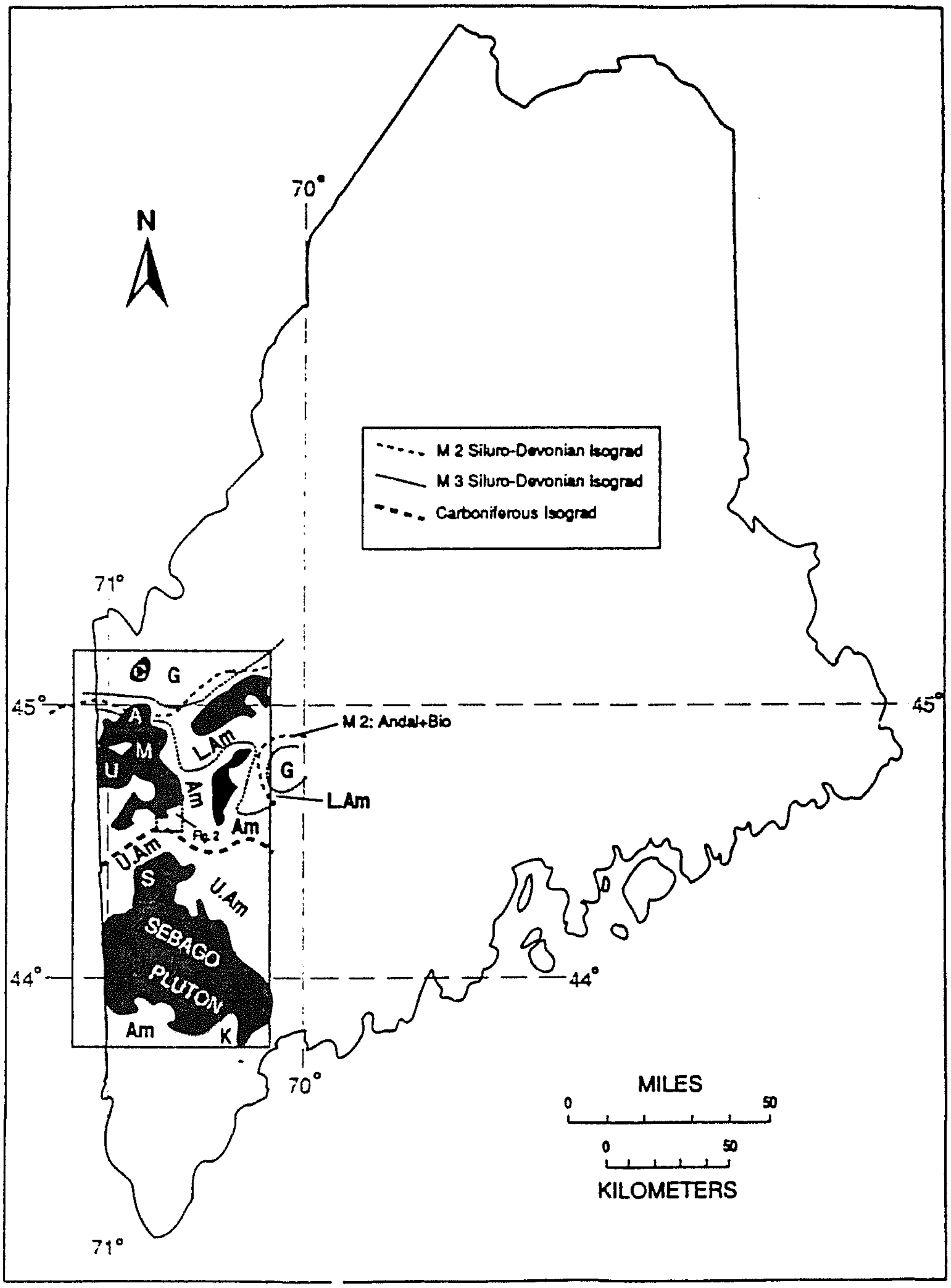


Fig. 1. Location of Area designated as Western Maine and the East Andover 7 1/2' Quadrangle. Isograds modified from Guidotti (1985). Metamorphic facies designations: G - greenschist, L.Am - Lower Amphibolite, Am - Amphibolite, U.Am - Upper Amphibolite. Other abbreviations: C - Cupsuptic pluton, A - Adamstown granite, M - Mooselookmeguntic pluton, S - Songo pluton, U - Umbagog pluton, K - Kyanite-bearing rocks. For all isograds the high-grade portion is on the west or south side as the case may be.

An especially important aspect of the work in the 1980's is that it started to focus on synthesis and integration of metamorphic information with larger geologic questions. Early reports include Holdaway et al.(1982), Guidotti et al. (1983), and Guidotti (1985). These reports clearly recognized the largely post-tectonic, multiple nature of the metamorphism that affected Western Maine and the fairly close spatial relationships between high metamorphic grade and granitic plutons, but cause and effect aspects were left as somewhat ambiguous. Basically, the heat source for the metamorphism was discussed (assigned) at least in part as due to conductive heat transfer from depth. For example, Guidotti et al.(1983) called attention to the enormous areas of migmatitic, sillimanite + K-feldspar zone rocks. Noting that none of these rocks attain granulite facies, they talked in terms of thermal buffers -- possibly due to partial melting.

During the early to mid 1980's there had also occurred an extensive expansion in our general understanding of the theoretical aspects for generation of heat during orogenesis, e.g. England and Thompson (1984,1986). There was a rapid development of sophisticated views on how heat for metamorphism can be generated and its relationship with megatectonic activity. Some of the theoretical work was even couched in the context of areas in the northern Appalachians e.g. Chamberlain and England (1985).

Coincidentally, in the early 1980's, the University of Maine expanded its bedrock geology group (via NSF funding through the EPSCOR program). A goal of this NSF funded expansion was to develop time-depth cross sections in northern New England with special emphasis on the thermal aspects of the terrain. Drawing upon the expanded theoretical understanding noted above, the U. Maine group has presented a number of reports which integrate the tectonic, metamorphic, and igneous aspects of the geology in Western Maine into a coherent model, e.g. Lux et al. (1986), DeYoreo et al. (1989A), DeYoreo et al. (1989B).

As discussed in more detail below, their model involves heat pulses due to emplacement of sheet-like plutons at intermediate depths. The resulting metamorphism is thus a product of geologically instantaneous, isobaric, thermal "spikes" with cooling to ambient temperatures between the "spikes".

Finally, in the context of this field trip it is germane to note that a very large fraction of the mineralogic and petrologic work in Western Maine has been carried out within the theoretical framework established by Professor Thompson for the AKNa and AFM systems Thompson (1957, 1961), Thompson and Thompson (1976). Such work has provided the foundation for the recent broad-scale syntheses.

Styles of Metamorphism in Western Maine

Recently, Guidotti (1989) has reviewed in some detail the metamorphic history of Western Maine. Here, the discussion will focus only on those major points relevant to the goal of this field trip, the change in style of metamorphism along strike of the orogen. Although some of the area designated herein as Western Maine may have been affected by pre-Silurian metamorphism, we will consider only Siluro-Devonian to Carboniferous, "Acadian", metamorphic events.

With the goal stated above in mind, one can note that to the northeast of the area designated as Western Maine (Fig. 1) one encounters mainly Siluro-Devonian greenschist-grade metamorphism. Higher grade rocks occur only in narrow contact zones developed around bulbous or cylindrical shaped granitic plutons. The contact rocks commonly contain cordierite + andalusite and only rarely staurolite or garnet. Hence, they probably reflect high-level intrusions that served as very localized heat sources.

Coming southwestward into Western Maine one encounters lobes of regionally developed Siluro-Devonian metamorphism that attained sufficiently high grades that the AFM, biotite + Al-silicate join, was established. Drawing from earlier reviews of Holdaway et al. (1982) and Guidotti et al. (1983), Guidotti (1989) noted the following generalizations having relevance (and here paraphrased) for this report:

(a) The Siluro-Devonian regional metamorphism involves several, largely post- tectonic recrystallization events that overlap in space and time. Due to these overlaps one commonly finds areas in which an earlier event is overprinted by a later event. This results in prograded rocks in some areas and retrograded rocks in others. A textural result of such overlaps is the development of numerous types of pseudomorphs (prograde and retrograde; see Guidotti, 1970B).

(b) In a very general way it appears that two dominant Siluro-Devonian events (M_2 and M_3) can be defined (See Fig. 1). Both involve grades high enough to establish the AFM join connecting Al-silicate (andalusite or sillimanite) and biotite (i.e., amphibolite facies) (In the north-central part of the Rumford quadrangle, M_3 may have reached upper amphibolite facies, i.e. sillimanite + K-feldspar zone). Hence, they are relatively low P, andalusite/sillimanite types of metamorphism. They were preceded by an earlier event (M_1) which attained only greenschist facies and appears to have been syntectonic.

(c) M_3 probably occurred at slightly higher P than M_2 (0.5 to 1 Kb increase). Holdaway et al. (1982) and Guidotti et al. (1983) suggested that the non-tectonic increase in P was possibly due to extrusion of a few Km of volcanic rocks or emplacement of plutons above the level of the present erosion surface.

(d) Relationships between prograde and retrograde minerals suggest that the rocks at a given locality cooled to ambient T between M_2 and M_3.

(e) The higher metamorphic grades of the metamorphic lobes are clearly related spatially to the distribution of Devonian plutons (see Figure 1). In most places the isogradic surfaces appear to have gentle dips, thereby forming the broad, NE-plunging lobes.

Drawing upon the more recent work, Guidotti (1989) noted the following additional generalizations (here paraphrased):

(a) Isotopic studies by Hayward and Gaudette (1984), Aleinikoff (1984) and Aleinikoff et al. (1985), and isotopic/ petrologic work by Lux and Guidotti (1985) strongly suggest that the highest grade rocks in the Western Maine metamorphic lobes (upper amphibolite facies, see Figure 1) are related to the thermal effects of the Carboniferous, Sebago pluton (see below).

(b) Lux et al. (1986) and DeYoreo et al. (1989) have developed quantitative models showing that the regional Siluro-Devonian M_3 recrystallization was actually a "deep", (12-16 km), contact metamorphism. Its large areal extent results from the gently dipping, tabular nature of the granitic plutons (the heat sources). Thus, the high-grade metamorphism was due directly to convective heat transfer. In the absence of the plutons the metamorphic grade would not have exceeded greenschist facies. In this sense, the geologic significance of this metamorphism shrinks because the important question becomes the source of heat to produce the magmas at greater depths.

(c) In the context of plutons serving as heat sources for the Siluro-Devonian regional metamorphism, Lux and DeYoreo (1987) have emphasized the alternative means for the above-mentioned, non-tectonic P increase between M_2 and M_3. Specifically, emplacement of sheet-like plutons above the now exposed rocks would cause an increase in P just as readily as would extrusion of a blanket of volcanic rocks. Of course, both may have occurred. Indeed, some such pluton may have served as the heat source that produced the M_2 metamorphism. This is an "attractive" idea because no other heat source has been found for that event -- at least not in the area covered in Fig. 1.

(d) Ar40/Ar39 work on the northern parts of the Mooselookmeguntic pluton (Lux and Guidotti, 1985; Lux et al., 1986) strongly supports (verifies ?) the above mentioned cooling to ambient T between M_2 and M_3. Hornblende cooling ages there are essentially identical with the crystallization age of the pluton reported by Moench and Zartman (1976) (corrected for new decay constants). Hence, the Mooselookmeguntic pluton (i.e. the heat source for M_3) must have intruded cold M_2 metamorphic rocks, and then, along with the M_3 metamorphic rocks, cooled below the 500 °C closure T of hornblende in a matter of a very few million years.

(e) As shown by Lux et al. (1986) and DeYoreo et al. (1989A) the metamorphism associated with the sheet-like Siluro-Devonian plutons was nearly isobaric. In terms of diagrammatic PT space, the metamorphism occurred essentially as a geologically instantaneous thermal "spike" resulting in prograde reactions. DeYoreo et al., (1989B) present a similar model for the Carboniferous metamorphism associated with the Sebago batholith.

Given such a thermal history, the recrystallization history is relatively simple and closely approaches equilibrium. In this respect, such low P, high T metamorphism contrasts strongly with recrystallization along the P-T-time (P-T-t) loops commonly associated with deeper level Barrovian metamorphism in which convective heat transfer is presumably negligible (e.g. see England and Thompson, 1984).

(f) The extensive area of low grade rocks to the northeast of the high grade lobes may be a manifestation of the same early heating event (M_1) that produced only greenschist facies rocks in the area now affected by the higher grades of M_2 and M_3. This remains an open question. Nonetheless, one can speculate that such a heating event was associated with a tectonic thickening which set the stage for later production of magmas at depth and which, upon movement to higher levels, caused M_2, M_3 and eventually the Carboniferous metamorphism.

Regardless of such speculation, it now appears that the transition from the high grade, regional Siluro-Devonian metamorphism northeastward, to the low grade areas containing plutons surrounded by narrow contact aureoles involves a change in pluton geometry. The plutons associated with the high grade areas occur as abundant, gently dipping sheets. In contrast, the plutons intruding the low grade rocks to the northeast seem to be more nearly equant in shape (e.g. Moore, 1960; Tewhey, 1975; Hodge et al., 1982; and Carnese, 1983). Lux and DeYoreo (1987) and DeYoreo et al. (1989A) argue that from southwest to northeast, the rocks now exposed at the earth's surface are from successively higher structural levels. Presumably, in conjunction with the resultant, systematically varying ambient T and P this would influence the shapes attained by the intruding plutons, and hence, the nature of the contact metamorphic effects.

Continuing to paraphrase from Guidotti (1989): The first strong evidence for a post-Devonian regional metamorphism in Western Maine came from the isotopic age dating of the Sebago batholith at 325 Ma. (Hayward and Gaudette, 1984; Aleinikoff, 1984). Noting that the highest grade metamorphism in western Maine is spatially related to the Sebago batholith, Aleinikoff et al. (1985) briefly speculated on the possibility that this metamorphism was the same age as the pluton.

Simultaneously Lux and Guidotti (1985) presented a much more complete picture combining geologic, petrologic and 40Ar/39Ar hornblende cooling ages for various rock types north of the Sebago batholith. Their results make a compelling argument that the extensive area of migmatitic, upper amphibolite facies rocks directly contiguous with the batholith on its northern perimeter is due to heat imparted to the country rocks by the intruding Sebago batholith. More recently, Guidotti et al. (1986) have presented further details of the effects of this Carboniferous metamorphism on both the metasedimentary strata and Devonian plutons that occur to the north of the Sebago.

Briefly, the Carboniferous metamorphic event was regionally extensive as shown merely by the distribution of upper amphibolite facies, migmatitic rocks (See Fig 1). These sillimanite + K-feldspar zone rocks were previously thought to be due to Devonian metamorphism (Guidotti, 1985) but would now be re-assigned to a younger age. Thus, the highest grade shown on Figure 1 is Carboniferous in age rather than Devonian. This event involved a post-tectonic heating that was superimposed on rocks metamorphosed earlier during the Devonian. To the north this recrystallization resulted in re-adjustment of the grades attained during the M_3 event, and in the north-central part of the Rumford 15' quadrangle produced a retrogression of the grades previously attained.

J.J. DeYoreo (in Guidotti et al. (1986) and DeYoreo et al. (1989B)) has modelled the thermal effects of the Sebago pluton intruded as a relatively thin sheet (Hodge et al. (1982) at about 15 Km and with a gentle dip. DeYoreo's results show that the Sebago batholith could have provided adequate heat to produce the observed metamorphism. Clearly the nature of this Carboniferous metamorphism is quite similar to the Devonian metamorphism in that it involves post-tectonic recrystallization related to heat imparted by a gently-dipping, sheet-like granitic pluton. As with the Devonian plutons, the intrusion occurred in rocks which had essentially cooled to ambient temperature.

Guidotti et al. (1986) considered the heating associated with the Sebago intrusion to be a late effect of the Acadian Orogeny. DeYoreo's thermal modelling shows that the Sebago intrusive event was likely a delayed response to deep-level melting that also produced the Devonian plutons subsequent to Acadian crustal thickening. In this sense the metamorphic zonation shown by Guidotti (1985) for Western Maine is all Acadian metamorphism but it includes recrystallization events separated by as much as 55 million years. Moreover, the "late stage Acadian" Sebago batholith has intruded earlier Acadian plutons such as the Songo granodiorite and thus had a reheating effect on them. The possible consequences of such reheating were considered by Lux and Guidotti (1985) and Guidotti et al. (1986).

A further interesting aspect of the Carboniferous metamorphism in western Maine is that kyanite-bearing assemblages were formed on the south side of the Sebago batholith, presumably its under side. As reported by Thomson (1986) and Thomson and Guidotti (1986,1989), true Barrovian pressures appear to have prevailed there during the recrystallization. The metamorphic grades, ranging into amphibolite facies, show a clear spatial relationship with the contact of the Sebago batholith. Moreover, the metamorphism involved mainly post-tectonic recrystallization superimposed on earlier Devonian metamorphism.

In contrast with the metamorphism on the north (upper) side of the pluton (highly migmatitic, upper amphibolite facies), the grade on the south (lower) side only reached amphibolite facies and the rocks are not migmatitic. Presumably this difference reflects greater heat transfer upward due to upward movement of escaping hot fluids, as the Sebago cooled. Although some pegmatites occur close to the Sebago pluton along its southern margin, it is on the northern side that there is a spectacular development of pegmatites. Variable sized bodies of pegmatite occur in a broad zone extending out into the country rocks for about 30 Km. This zone includes many famous pegmatite quarries that have yielded vast quantities of tourmaline, beryl, lepidolite, and other minerals. From the spatial and petrologic considerations discussed above, it is likely that the Sebago pluton might be the ultimate source for these pegmatites.

The areal extent of the Carboniferous recrystallization has not yet been determined except for the areas cited above. Presumably the distribution of upper amphibolite facies grade rocks shown on Figure 1 in Western Maine, provides a rough indication of its potential areal extent north of the Sebago pluton.

Due to time constraints, we will be able to see outcrops only of the Carboniferous metamorphism at grades ranging from lower sillimanite zone to upper sillimanite zone on this field trip. In the context of textures seen at these outcrops it will be argued that this Carboniferous metamorphism has overprinted rocks probably affected previously by both the M_2 and M_3 Siluro Devonian metamorphisms described above. All of the outcrops to be seen are in the East Andover 7-1/2 topographic sheet. Most of the comments made below are drawn from the descriptions presented in Guidotti et al. (1986) which were based on work by Cheney (1975), Cheney and Guidotti (1979) and Guidotti and Cheney (in prep).

Metamorphism in the East Andover 7-1/2' Quadrangle

Metamorphism in this area ranges from lower sillimanite zone to sillimanite + K-feldspar zone (see Figure 2). Based upon the work of Lux and Guidotti (1985) the isograds are interpreted as due to a Carboniferous event superimposed upon the earlier Devonian M_2 and M_3 events.

The transition from the upper sillimanite zone (USZ) to sillimanite + K-feldspar (SKZ) was studied and described in detail by Evans and Guidotti (1966). The upper stability limit of muscovite was never exceeded and so assuming that P-total was about 3.5 Kb (see below) and that PH20 was high, they estimated T-max to be in the range somewhat below 650 °C. More recently, Cheney and Guidotti (1979), using the exchange geothermometer based upon the reaction:

$$(NaAl_3Si_30_{10}(0H)_2)^{Musc} + (Si0_2)^{Qtz} = (NaAlSi_30_8)^{Plag} + (Al_2Si0_5)^{Sill} + H_20$$

obtained more quantitative estimates (623 - 643 °C for the SKZ). Based on fluid inclusion studies on these rocks, Burruss (1977) obtained very similar estimates. In the USZ and SKZ the rocks are so recrystallized that there is little evidence of earlier Devonian textural features. However, in portions of the East Andover quadrangle and S.E. Old Speck Mountain quadrangle (on Puzzle Mountain) the rocks are lower sillimanite zone (LSZ) (Cheney, 1975) and it is believed that some

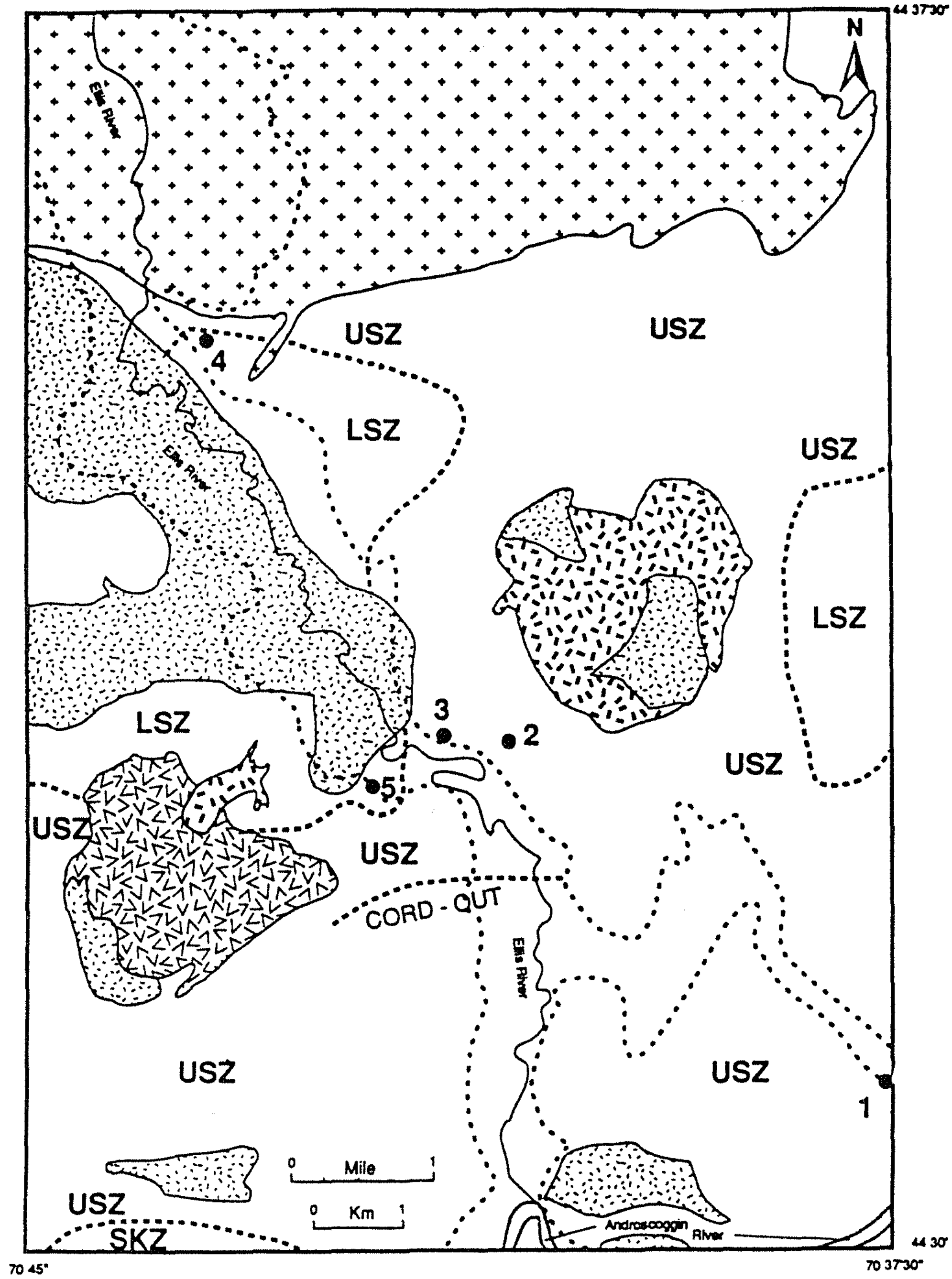


Fig. 2. Simplified Geologic Map of the East Andover 7 1/2' quadrangle. Geology based on Moench and Hildreth (1974). Metamorphic grades: LSZ = Lower sillimanite zone, USZ = Upper sillimanite zone, Crd-out = breakdown of cordierite in the presence of muscovite, SKZ = sillimanite + K-feldspar zone; Isograds from (work in progress), Cheney (1975) and Evans and Guidotti (1966). Plutons: Crosses = granodiorite, tonalite, quartz diorite, Double hachures = pegmatitic granite; Stippled = two mica granite; Random V's = gabbro, diorite, and some ultramafic. Uncolored rocks = Siluro - Devonian metapelites etc. Dotted lines = boundary of Qal deposits. Numbers 1 - 5 = Field trip stops.

of the textural features (annealed slip cleavages, partial pseudomorphs of staurolite etc) could be remnants from earlier Devonian metamorphism(s). For example, the LSZ rocks are textually extremely similar to those seen in the Rangeley area (Guidotti 1970B, 1974) where only Devonian metamorphic events have occurred. It should also be noted that these textures are essentially identical with those seen on the eastern slopes of the White Mountains in New Hampshire, (Wall and Guidotti, 1986; and Hatch and Wall, 1986). However, because the Devonian and Carboniferous metamorphism are identical in style, they should produce similar textures. Hence, textural aspects can not be related rigorously to the age of the metamorphism.

Possibly the most interesting aspect of the Carboniferous metamorphism in the East Andover quadrangle is that it affects the very sulfide-rich Smalls Falls formation as well as the Perry Mountain, Rangeley etc. formations. Hence, it involves cordierite-bearing parageneses in the highly sulfitic Smalls Falls formation and staurolite-bearing parageneses in the less or non-sulfitic units. Guidotti and Cheney (manuscript in preparation) have shown that staurolite breaks down before cordierite by means of the reactions:

(1) St + Musc + Qtz = Sill + Bio + Garn + H_2O

Subsequently, end member Mg-Cordierite breaks down (in a more Mg-rich portion of composition space) by the reaction:

(2) Musc + Mg-Cord = Sill + Phlog + Qtz + H_2O

At still higher T's the SKZ is attained via the AKNaCa reaction:

(3) Qtz + Musc + Sodic-Plag = Sill + Sodic-Ksp + Ca-richer Plag + H_2O

As developed in Guidotti and Cheney (Ms), at higher P, the order of reaction (1) and (2) would probably be reversed. It should be noticed from the order of reaction (2) and (3) (See Figure 2 also), that in the muscovitic rocks of this area, cordierite breaks down before the SKZ is attained. At grades exceeding the stability of muscovite, as in Massachusetts, end member Mg-cordierite becomes stable again in K-feldspar-bearing rocks (Robinson, pers. comm.) and via the subsequent continuous reaction it becomes Fe-richer. Thus, in muscovite-bearing rocks, cordierite becomes Mg-richer until its demise via reaction (2). In contrast, in higher grade K-feldspar-bearing rocks it comes in again as end member Mg-Cordierite which then becomes Fe-richer as T increases further.

The mineral facies diagrams determined for the Carboniferous metamorphism in the East Andover quadrangle and SW portion Old Speck Mountain quadrangles respectively, and southward through the Bryant Pond quadrangle are shown in Figure 3. Figure 4 shows the relevant PT curves and suggested PT path (i.e. metamorphic field gradient) for the Carboniferous metamorphism. It should be noted that in the context of metamorphism due to the thermal effects of the Sebago Batholith that the PT path is nearly isobaric and because the rocks underwent nearly static recrystallization, the P-T-t paths of individual horizons were also nearly isobaric.

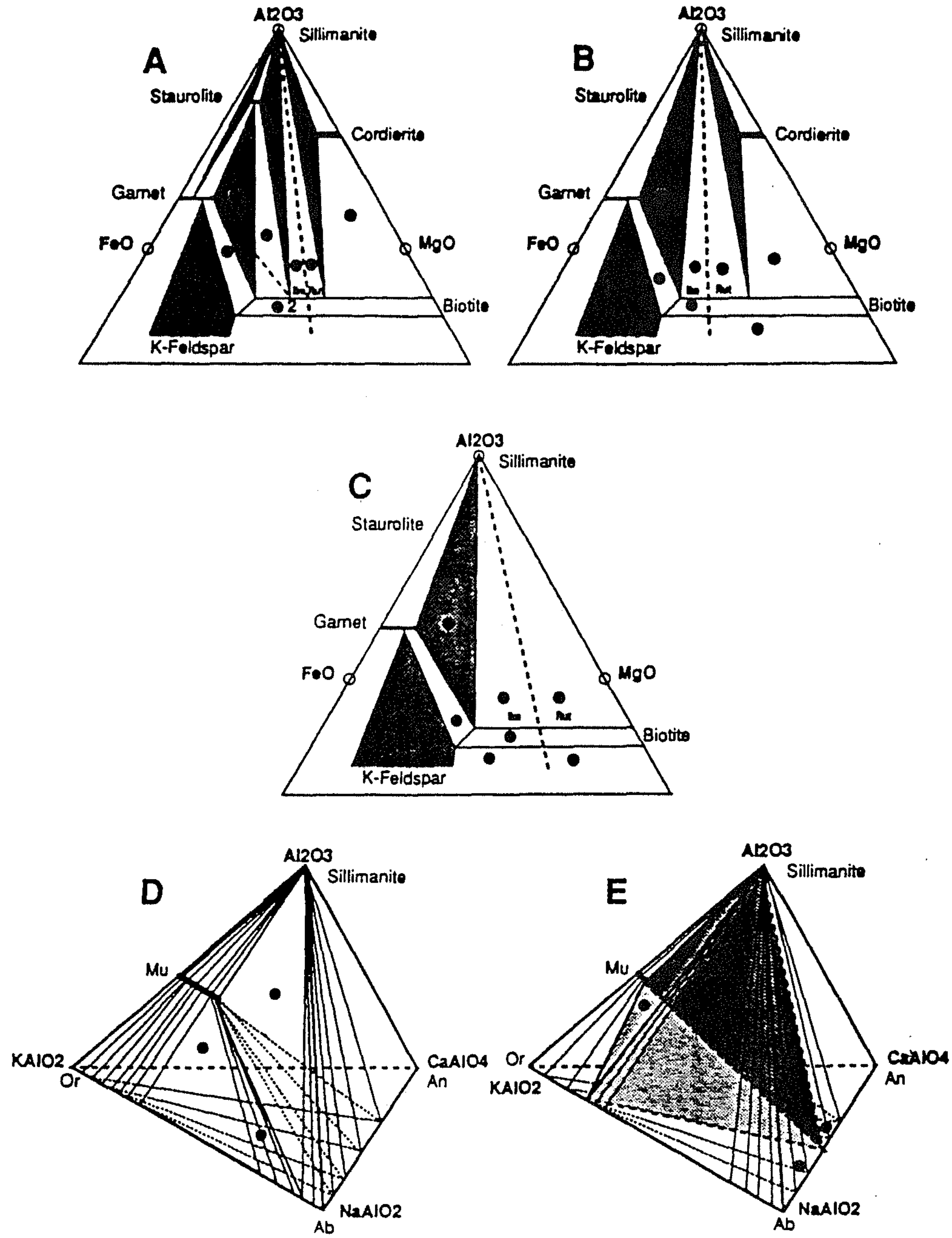


Fig. 3. Mineral facies diagrams for the East Andover quadrangle (from Guidotti et al., 1986). (A)-(C) - AFM projections; quartz, muscovite, and uniform a(H2O) in all assemblages; Ilm: Rut indicates boundary between assemblages with ilmenite vs rutile as the Ti-saturating phase. (A) Lower Sillimanite zone; Line 1-2 to indicate garnet in assemblages not shown by AFM projection. (B) Upper sillimanite zone, below the breakdown of cordierite. (C) Upper sillimanite zone and Musc-bearing sillimanite + K-feldspar zone, above cordierite break-down. (D) AKNaCa mineral facies diagram for lower and upper sillimanite zones. (E) AKNaCa mineral facies diagram for the K-feldspar + sillimanite zone. Dots on (A)-(E) = observed assemblages.

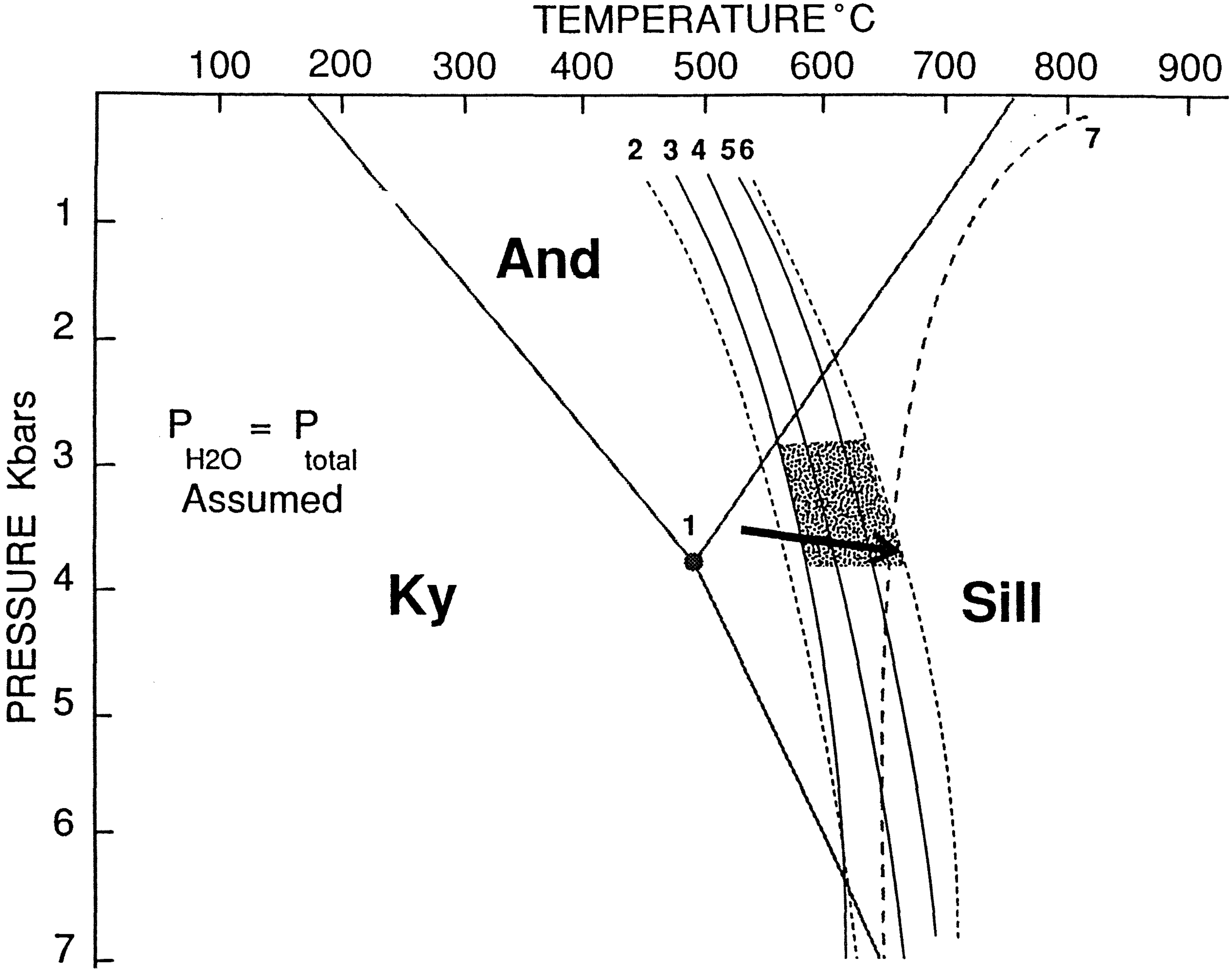


Fig. 4. PT grid of equilibria relevant to this report. (1) Al-Silicate curves, Holdaway (1971). (2) Paragonite breakdown, Chatterjee (1972). (3) St + Chl = Al-Sil + Bio, Guidotti (1974). (4) St = Al-Sil + Bio, Hoschek (1969). (5) Ab + Mu = Al-Sil + Ksp, Chatterjee and Froese (1975). (6) Muscovite breakdown, Chatterjee and Johannes (1974). (7) Granite minimum, Tuttle and Bowen (1958). Stippled area - suggested range of PT conditions for metamorphism considered herein. Heavy arrow - suggested PT path.

Itinerary for Field Stops

The assembly point for this field trip is at the Madison Motel on Rte 2, several miles west of Rumford.

Mileage

0.0 Head west (left) on Rte 2.

0.4 Small outcrops of pegmatite and adamellite

1.2 Turn right (north) on paved road

1.55 **STOP 1:** Roadside outcrops of Perry Mountain fm consisting of crinkled, laminated metapelites interbedded with biotite granulites. The metamorphic grade is upper sillimanite zone and the metapelite assemblage is Sill + Gn + Bio + Musc + Plag + Po + Gr + Ilm. Commonly sillimanite amounts to 10-15 modal %. This outcrop was included in the report of Evans and Guidotti (1966) as sample No 13. Probe data were given for the Musc, Plag, and Bio.

Some textures to note include: Small, irregular aggregates of coarse muscovite are probably pseudomorphs after staurolite. The loose blocks at the north end of the crop provide the best display of the muscovite aggregates. At higher grades these tend to recrystallize into single large spangles. Sillimanite tends to grow in the old tectonic fabric and at the north end of the crop there are some spectacular sillimanite knots. Two foliations are present, one being subparallel to bedding and the other at about 30 degrees to the bedding. This outcrop is not migmatitic but some small pegmatite veins are present. The sillimanite + K-feldspar isograd passes about five miles to the south of this outcrop. On the hill to the SW a hornblende has been studied which gives a 324 Ma. cooling age -- thereby supporting the proposed Carboniferous age for the last high grade metamorphism.

2.9 View to north of Whitecap Mtn. It consists mainly of pegmatite plus some aplite. As discussed in Guidotti et al. (1986) arguments can be made for the rocks there having a Carboniferous age.

3.15 Stay left at fork.

3.3 Continue straight ahead.

4.3 Continue straight to north.

4.6 Cliffs to the west on Mt. Dimmock consist of Madrid formation (calc-silicate and biotite granulite).

5.4 Roadside outcrops of Small Falls fm. and pegmatite.

5.7 Bear right on road to East Andover

5.9 **STOP 2:** Turn right on woods road - park in woods at 6.0 miles. TAKE CARE NOT TO GET STUCK IN THE SAND. Walk to the SE on the clear, brush-free road. Shortly we will encounter large road bed crops of Smalls Falls fm. It consists of interbeds of metapelite and quartzite. The brown rust clots on the clean surfaces give some idea of the modal abundance of pyrrhotite in this unit.

Note some of the loose blocks of other units which are staurolite bearing. Inspect the details of the staurolite textures as they involve aggregates of euhedral crystals. These textures are fairly typical of places where early staurolite grade rocks are affected by later events also at staurolite grade.

These outcrops of Smalls Falls are unusual in that they are fresh. Some of the loose blocks at the top of the road crop may have cordierite (now pinite).

At the top of the crop, take the obscure path to the left into the woods. It leads in about 150' to the old Nickel Mine. STAY OUT OF THE CAVERNS. Head to the south wall of the pit.

The big slabs at the base of the south wall contain altered cordierite and micaceous (muscovite and phlogopite) pseudomorphs after cordierite. As you work your way up the south wall and your eyes get keyed in, you'll find that cordierite is (or was) very abundant in these outcrops. This reflects the presence of the abundant pyrrhotite serving to tie up much of the Fe in the rock so that the silicate assemblage is one in a very Mg-rich portion of composition space.

Unfortunately such sulfitic rocks weather very badly and so it is very difficult to get fresh cordierite. Some has been found, the freshest of which comes from some deep drill cores recovered from an inclusion in the pegmatites of Whitecap Mtn. The assemblage there is Sill + Crd + Phlog + Musc + Plag + Gr + Po + Rut. As seen on Fig. 3, rutile instead of ilmenite is the Ti-saturating phase in these highly sulfitic rocks. This change is readily understood in terms of the graphics developed by Thompson (1972).

Return to cars and head back to tar road.

6.1 Tar road - turn right (north). The mine dumps seen on the ridge to the west are from the famous Newry pegmatite quarries.

6.5 **STOP 3:** Outcrop of Perry Mountain fm. in woods, 50' to the east on Howe Hill. Highly folded with an excellent axial plane cleavage. Also, excellent graded beds showing the coarser portion in the "upper" part of the bed.

This outcrop is remarkable because it is at upper sillimanite zone but still retains many delicate early sedimentary and fabric features. Note how some of the thin quartzite bedlets have been dragged along the axial plane cleavage into the more pelitic portions of graded beds.

Sillimanite knots are abundant and some bedlets have extreme concentrations of sillimanite. Most of the sillimanite is in the groundmass and the thin bedlets but some concentrations occur along thin ptygmatic quartz veins.

The assemblage is Sill + Gn + Bio + Musc + Plag + Ilm + Gr. Some weak development of muscovite spangles occurs.

This outcrop is a few hundred feet south of locality No. 9 in the study of Evans and Guidotti (1966). They presented microprobe data for the Musc, Plag, and Bio at that locality. Return to cars.

6.6 More outcrops of folded Perry Mountain formation.

6.8 Small, roadside outcrops of Smalls Falls fm.

9.7 Bear to the right at the fork and continue north.

9.95 Use wide entrance to woods road to make a loop and reverse direction so you are heading south.

10.2 **STOP 4:** Carefully pull off to the left and park. Large outcrop of well bedded Perry Mountain fm to the west in the woods. Note the lineation crinkles.

This exposure is in the lower sillimanite zone and contains Sill + St + Gn + Bio + Andal. The andalusite was produced during the Devonian M_2 or M_3 metamorphism. By analogy with the observations made in the Rangeley area (Guidotti 1970B, 1974) it is considered to be a metastable relic.

The best view of the mineralogy and texture present can be had from the top of the crop.

One can find staurolite and andalusite partially pseudomorphed by coarse muscovite aggregates. Note the abundant sillimanite knots. In some cases one can find the andalusite going directly to sillimanite. This is fairly unusual in western Maine - except in cases involving narrow, higher level contact hornfels (e.g. around the Reddington Pluton) where the heating was very rapid. In this outcrop the andalusite going directly to sillimanite may reflect rapid heating during the superposition of M_3 on M_2. This outcrop is very close to the contact with the Mooselookmeguntic pluton.

Some chlorite after biotite is present in this outcrop indicating some late retrogression, possibly associated with faulting. Return to cars and drive south.

10.4 Turn to the right (west).

10.9 Covered bridge - ONE CAR AT A TIME on this ONE WAY BRIDGE (Note weight limits).

11.1 Stop sign for Route 5. Turn left (south) on Route 5.

13.1 Large outcrops of the Mooselookmeguntic pluton. This outcrop was written up as a stop for the 1986 NEIGC (Guidotti et al. 1986). Continue going south.

15.1 **STOP 5:** Large outcrop of Perry Mountain fm in the lower sillimanite zone. The assemblage is Sill + Gn + Bio + Staur. The staurolite can be seen only in thin section and occurs mainly within the well developed muscovite pseudomorphs after staurolite that occur in this outcrop. Sillimanitie is also present within these pseudomorphs.

Sillimanite is modally abundant in coarse knots. Note that the muscovite in the pseudomorphs is coarse grained and unoriented. A slip cleavage is present in this outcrop but has no effect on the pseudomorphs or the original staurolite.

The best views of the pseudomorphs can be seen near the middle of the rock face.

Good graded bedding is present in this outcrop. Some of the few sandy and calcareous beds have been boudinaged. Return to cars and head south.

15.35 The road on the right leads to the Newry pegmatite mines. For a fee one is allowed to do collecting there. These mines are famous for tourmaline. In the mid 70's a pocket was opened which was "reported" to have produced four million dollars worth of tourmaline. Continue driving to the south.

15.80 Outcrops of Smalls Falls fm - continue on to the south.

18.65 Stop sign for intersection with Route 2. Turn right (west) to head to New Hampshire.

References

Aleinikoff, J.N. (1984) Carboniferous uranium-lead age of the Sebago batholith, southwestern Maine, Geol. Soc. Amer. Abstracts with Programs, 16, p.1.
Aleinikoff, J.N., Moench, R.H., and Lyons, J.B. (1985) Carboniferous U-Pb age of the Sebago batholith, southwestern Maine. Geol. Soc. Amer. Bull. 69, p.990-996.
Boone, G.M. (1973) Metamorphic stratigraphy, petrology, and structural geology of the Little Bigelow Mountain map area, western Maine. Maine Geol.Surv. Bull. 24, Dept. of Conservation, Augusta, Maine. 136 p.
Burruss, R.C. (1977) Analysis of fluid inclusions in graphitic metamorphic rocks from Bryant Pond, Maine and Khtada Lake, British Columbia: Thermodynamic basis and geologic interpretation of observed fluid compositions and molar volumes. Ph.D. thesis, Princeton Univ., Princeton NJ. 156 p.
Carnese, M.J. (1983) Gravity studies of intrusive rocks in west-central Maine. M.S. Thesis, Univ. of New Hampshire, Durham, NH. 97 p.
Chamberlain, C.P. and England P.C. (1985) The Acadian thermal history of the Merrimack synclinorium in New Hampshire. Jour. Geol. 93, p.593-602.
Chatterjee, N.D. (1972) The upper stability limit of the assemblage paragonite and quartz and its natural occurrence. Contr. Mineral. Petrol. 34, p.288-303.
Chatterjee, N.D. and Johannes W. (1974) Thermal stability and standard thermodynamic properties of synthetic $2M_1$ muscovite K $Al_2(AlSi_3O_{10})(OH)_2$. Cont. Miner. Petrol., 48, p.89-114.
Chatterjee, N.D. and E. Froese (1975) A thermodynamic study of the pseudobinary join muscovite-paragonite in the system $KAlSi_3O_8$ - $NaAlSi_3O_8$ - Al_2O_3 - SiO_2 - H_2O. Amer. Mineral. 60, p.985-993.
Cheney, J.T. (1975) Mineralogy and petrology of lower sillimanite through sillimanite + K-feldspar zone, pelitic schists, Puzzle Mountain area, Northwest Maine. Ph.D. thesis, Univ. of Wisconsin, Madison, Wisconsin, 291 p.
Cheney, J.T. and Guidotti, C.V. (1979) Muscovite-plagioclase equilibria in sillimanite + quartz-bearing metapelites, Puzzle Mountain area, northwestern Maine. Amer. Jour. Sci. 279, p.411-434.
DeYoreo, J.J., Lux, D.R. and Guidotti, C.V. (1989) A thermal model for Carboniferous metamorphism near the Sebago batholith in western Maine. Maine Geol. Survey, C.T. Jackson Commemorative Volume.
DeYoreo, J.J., Lux, D.R., Guidotti, C.V., Decker, E.R. and Osberg, P.H. (1989) The Acadian thermal history of western Maine. J. Metamorphic Geol., 7 p. 169-190.
Drake, J.C. (1968) The composition of almandine and coexisting minerals in the pelitic schists of the Errol quadrangle, New Hampshire-Maine. Ph.D. thesis, Harvard University, Cambridge, Mass.
England, P.C. and Thompson, A.B. (1984) Pressure-temperature-time paths of regional metamorphism I. Heat transfer during the evolution of the thickened continental crust. Jour. Petrol., 25, p.894-928.
England, P.C. and Thompson, A.B. (1986) Some thermal and tectonic models for crustal melting in continental collision zones. In: [(eds. Coward, M.P., & Ries A.C.)] Collision Tectonics. Geol. Soc. Spec. Pub., 19, p.83-94.
Eugster, H.P. and Skippen, G.B. (1967) Igneous and metamorphic reactions involving gas equilibria. in Abelson, P.H. ed., Researches in Geochemistry, 2, p.492-520. New York: John Wiley and Sons.
Evans, B.W. and Guidotti, C.V. (1966) The sillimanite-potash feldspar isograd in western Maine, USA. Contr. Mineral. Petrol., 12, p.25-62.
Foster, C.T. Jr. (1975) Diffusion controlled growth of metamorphic segregation in sillimanite grade pelitic rocks near Rangeley, Maine, USA. Ph.D. thesis, The Johns Hopkins University, Baltimore, MD.

Foster, C.T. (1977) Mass transfer in sillimanite-bearing pelitic schists near Rangeley, Maine. Amer. Mineral., 62, p.727-746.
Green, J.C. (1960) Geology of the Errol quadrangle, New Hampshire-Maine. Ph.D. thesis, Harvard Univ.
Green, J.C. (1963) High-level metamorphism of pelitic rocks in northern New Hampshire. Amer. Mineral., 48, p.991-1023.
Guidotti, C.V. (1963) Geology of the Bryant Pond quadrangle, Maine. Ph.D. thesis, Harvard University.
Guidotti, C.V. (1970A) The mineralogy and petrology of the transition from lower to upper sillimanite zone in the Oquassoc area, Maine. Jour. Petrol., 11, p.277-336.
Guidotti, C.V. (1970B) Metamorphic petrology, mineralogy, and polymetamorphism in a portion of N.W. Maine. In: Boone, G.M. ed. 1970 New England Inter. Coll. Geol. Conf. 62nd Ann. Mtg. Field Trip B-1, p.1-29.
Guidotti, C.V. (1974) Transition from staurolite to sillimanite zone, Rangeley quadrangle, Maine. Geol. Soc. Amer. Bull., 85, p.475-490.
Guidotti, C.V. (1985) Metamorphic Map of Maine in Osberg P.H. et al. eds., Bedrock Geologic Map of Maine. Maine Geol. Survey, Augusta, Maine.
Guidotti, C.V. (1989) Metamorphism in Maine: An Overview. Maine Geol. Surv. Studies in Maine Geology: 3, p.1-17.
Guidotti, C.V. and Evans, B.W. (1964) Petrochemical study of pelitic schists across the sillimanite + K-feldspar isograd in western Maine. Abst. Geol. Soc. Amer. Prog. Ann. Meeting, Miami, p.79.
Guidotti, C.V., Trzcienski, W.E. and Holdaway M.J. (1983) A northern Appalachian metamorphic transect - Eastern Townships, Quebec to the central Maine coast. in Schenk, P.E. ed., Regional trends in the geology of the Appalachian - Caledonide - Hercynian - Mauritanide Orogen. D. Reidel Pub. Co., Dordrecht, Holland, p.235-247.
Guidotti, C.V., Gibson, D., Lux, D.R. and DeYoreo, J.J. and Cheney, J.T. (1986) Carboniferous metamorphism on the north (upper) side of the Sebago Batholith. in Newberg, D.W. ed., New England Intercoll. Geol. Conf. Guidebook for field trips in southwestern Maine. p.306-341.
Harwood, D.S. (1966) Geology of the Cupsuptic quadrangle, Maine. Ph.D. thesis, Harvard University, Cambridge, Mass.
Hatch, N.L. and Wall, E.R. (1986) Stratigraphy and metamorphism of the Silurian and Lower Devonian rocks of the western part of the Merrimack Synclinorium, Pinkham Notch area, east-central New Hampshire. in Newberg, D.W. ed.:NEIGC Guidebook for field trips in southwestern Maine, p.138-163.
Hayward, J.A. and Gaudette, H.E. (1984) Carboniferous age of the Sebago and Effingham plutons, Maine and New Hampshire. Geol. Soc. Amer. Abstracts with Programs. 16, p.22.
Hodge, D.S., Abbey, D.A., Harkin, M.A., Patterson, J.A., Ring, M.J. and Sweeney, J.F. (1982) Gravity studies of subsurface mass distributions of granite rocks in Maine and New Hampshire. Amer. J. Sci., 282, p.1289-1324.
Holdaway, M.J. (1971) Stability of andalusite and sillimanite and the aluminum silicate phase diagram. Amer. Jour. Sci., 271, p.97-131.
Holdaway, M.J., Guidotti, C.V., Novak, J.M. and Henry, W.E. (1982) Polymetamorphism in medium-to high-grade pelitic metamorphic rocks, west-central Maine. Geol. Soc. Amer. Bull., 93, p.572-584.
Hollister, L.S., Burruss, R.C., Henry, D.L., and Hendel, E.M. (1979) Physical conditions during uplift of metamorphic terranes as recorded by fluid inclusions. Bull. Mineral., 102, p.555-561.
Hoschek, G. (1969) The stability of staurolite and chloritoid and their significance in metamorphism of pelitic schists. Contr. Mineral. Petrol., 22, p.208-232.
Lux, D.R. and Guidotti C.V. (1985) Evidence for extensive Hercynian metamorphism is western Maine. Geology, 31, p.696-700.

Lux, D.R., DeYoreo, J.J., Guidotti, C.V., Decker, E.R. (1986) The role of plutonism in low pressure/high temperature metamorphic belt formation. Nature, 323, p.794-797.

Lux, D.R. and DeYoreo, J.J. (1987) Magmatic erosion of the Acadian orogenic root. Abst. with Programs Ann. Mtg. N.E. Section Geol. Soc. Amer., 19, p.27.

Milton, D.J. (1961) Geology of the Old Speck Mountain Quadrangle, Maine. Ph.D. thesis, Harvard Univ., Cambridge, MA, 190 p.

Moench, R.H. and Hildreth, C.T. (1974) Geologic map of the Rumford Quadrangle, Oxford and Franklin Counties, Maine. U.S. Geol. Survey, Map GQ-1272.

Moench, R.H. and Zartman, R.E. (1976) Chronology and styles of multiple deformation, plutonism, and polymetamorphism in the Merrimack synclinorium of western Maine. in: Lyons, P.C. and Brownlow, A.H. eds., Studies of New England Geology: Geol. Soc. Amer. Mem., 146, p.203-238.

Moore, John M., Jr. (1960) Phase relations in the contact aureole of the Onawa pluton, Maine. Massachusetts Inst. Technology, Ph.D. Thesis 187p.

Pankiwskyj, K.A., (1964) The geology of the Dixfield quadrangle, Maine. Ph.D. Thesis, Harvard Univ. Cambridge, MA.

Tewhey, J. (1975) The controls of chlorite-garnet equilibria in the contact aureole of the Cupsuptic Pluton, west central Maine; and the two-phase region in the CaO-SiO2 system: experimental data and thermodynamic analysis. Ph.D. thesis, Brown University, Providence, RI 147 p.

Thompson, J.B., Jr. (1957) The graphical analysis of mineral assemblages in pelitic schists. Amer. Mineral., 42, p.842-858.

Thompson, J.B., Jr. (1961) Mineral facies in pelitic schists. in: Physico-chemical problems in the formation of rocks and mineral deposits (Sokolov, G.A. ed.) Korzhinskii, D.S. commemorative vol., p.313-325. Moscow: Akad. Nauk. SSSR (in Russian with English Summary)

Thompson, J.B., Jr. (1972) Oxides and sulfides in regional metamorphism of pelitic schists. 24th Int. Geol. Cong. (Montreal) Section 10, p.27-35.

Thompson, J.B. and Norton, S.A. (1968) Paleozoic regional metamorphism in New England and adjacent areas. in: Zen, et al. eds. Studies of Appalachian Geology, Northern and Maritime, John Wiley and Sons, New York, p.319-327.

Thompson, J.B. and Thompson, A.B. (1976) A model system for mineral facies in pelitic schists. Contr. Mineral. Petrol, 58, p.243-277.

Thomson, J.A. (1986) The occurrence of kyanite in southern Maine and its metamorphic implications, M.S. thesis, Univ. of Maine, Orono, ME, 129 p.

Thomson, J.A. and Guidotti, C.V. (1986) The occurrence of kyanite in southern Maine and its metamorphic implications. NE Sectional Meeting Geol. Soc. Amer. 18, 1, p.72.

Thomson, J.A. and Guidotti, C.V. (1989) Carboniferous Barrovian metamorphism in southern Maine. Maine Geol. Survey, C.T. Jackson Commemorative volume, 28 p. MS.Tuttle, O.F. and Bowen, N.L. (1958) Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$ - $KAlSi_3O_8$ - SiO_2 - H_2O, Geol. Soc. Amer. Mem 71, 153 p.

Wall, E.R. and Guidotti, C.V. (1986) Occurrence of staurolite and its implications for polymetamorphism in the Mt. Washington area, NH. Abstract - Northeast section, Geol. Soc. Amer. Ann. Meeting, 18, 1, p.74.

Warner, J.L. (1967) Geology of the Buckfield quadrangle, Maine. Ph.D. thesis, Harvard University, 232 p.

Supplemental List of References

References to papers dealing specifically with aspects of the metamorphic petrology and mineralogy of Western Maine

Papers in the 1950's and 1960's

Boone, G.M. (1955) Petrology of the Farmington, Maine area in: Trefethen, J.M.; Report of the State Geologist, 1953-54. Maine Development Commission, Augusta, Maine, p. 46-62.

Drake, J.C. (1968) The composition of almandine and coexisting minerals in the pelitic schists of the Errol quadrangle, New Hampshire-Maine. Ph.D. thesis, Harvard University, Cambridge, MA.

Evans, B.W. (1969) Chlorine and fluorine in micas in pelitic schists from the sillimanite-orthoclase isograd, Maine. Amer. Mineral., 54, p. 1209-1211.

Evans, B.W. and Guidotti, C.V. (1966) The sillimanite-potash feldspar isograd in western maine, U.S.A. Contr. Mineral. Petrol., 12, p. 25-62.

Faul, H., Stern, T.W., Thomas, H.H., and Elmore, P.L.D. (1963) Ages of intrusion and metamorphism in the Northern Appalachians. Amer. Jour. Sci., 261, p. 1-19.

Fisher, I.S. (1962) Petrology and structure of the Bethel area, Maine. Bull. Geol. Soc. Amer., 71, p. 1395-1420.

Green, J.C. (1960) Geology of the Errol quadrangle, New Hampshire-Maine. Ph.D. thesis, Harvard Univ.

Green, J.C. (1963) High-level metamorphism of pelitic rocks in northern New Hampshire. Amer. Mineral, 48, p. 991-1023.

Green, J.C. (1963) Alkali metasomatism in a thermal gradient, two possible examples. J. Geol., 71, p. 654-657.

Green, J.C. and Guidotti, C.V. (1968) The geology of the Boundary Mountains Anticlinorium in northern New Hampshire and northwestern Maine: in Zen, et al., eds., Studies of Appalachian Geology, Northern and Maritime. John Wiley & Sons, p. 255-266.

Guidotti, C.V. (1963) Geology of the Bryant Pond quadrangle, Maine. Ph.D. thesis, Harvard University.

Guidotti, C.V. (1963) Metamorphism of the Pelitic Schists in the Bryant Pond Quadrangle, Maine: Amer. Mineral., 48, p. 772-791.

Guidotti, C.V. (1965) Geology of the Bryant Pond Quadrangle, Maine. Maine Geol. Surv., Quadrangle Mapping Series No. 3, 116 p.

Guidotti, C.V. (1966) Variation of the basal spacings of muscovite in sillimanite-bearing pelitic schists of northwestern Maine. Amer. Mineral., 51, p. 1778-1786.

Guidotti, C.V. (1968) Prograde muscovite pseudomorphs after staurolite in the Rangeley0Oquossoc areas, Maine. Amer. Mineral, 53, p. 1368-1376.

Guidotti, C.V. and Evans, B.W. (1964) Petrochemical study of pelitic schists across the sillimanite + K-feldspar isograd in western Maine. Abst. Geol. Soc. Amer. Prog. Ann. Meeting., Miami, p. 79

Harwood, D.S. (1966) Geology of the Cupsuptic quadrangle, Maine. Ph.D. thesis, Harvard University, Cambridge, MA.

Harwood, D.S. and Larson, R.R. (1969) Variations in the delta-index of cordierite around the Cupsuptic pluton, west-central Maine. Amer. Mineral, 54, p. 896-908.

Milton, D.J. (1961) Geology of the Old Speck Mountain Quadrangle, Maine. Ph.D. thesis, Harvard University, Cambridge, MA, 190 p.

MIlton, D.J. and Ito, J. (1961) Gedrite from Oxford County, Maine. Amer. Mineral., 46, p. 734-740.

Pankiwskyj, K.A. (1964) The geology of the Dixfield quadrangle, Maine. Ph.D. thesis, Harvard University, Cambridge, MA.

Thompson, J.B. and Norton, S.A. (1968) Paleozoic regional metamorphism in New England and adjacent areas: in Zen, et al., eds. Studies of Appalachian Geology, Northern and Maritime: p. 319-327.

Warner, J.L. (1967) Geology of the Buckfield quadrangle, Maine. Ph.D. thesis, Harvard University, Cambridge, MA. 232 p.

Warner, J.L., Doyle, R.G. and Hussey, A.M. II (1967) Generalized map of regional metamorphic zones: in Doyle, R.G. Preliminary geologic map of Maine. Maine Geological Survey, Augusta, Maine.

Papers in the 1970's

Al-Mishwt, Ali (1972) Contact metamorphism and polymetamorphism in northwestern Oquossoc quadrangle, Maine. M.S. thesis, University of Wisconsin, Madison, Wisconsin.

Boone, G.M. (1970) Tectonic and metamorphic history of Lower Devonian rocks in the Carrabassett valley north of Kingfield, Maine: in Boone, G.M. ed., NEIGC Guidebook for field trips in the Rangeley Lakes-Dead River Basin region, western Maine. Kennebec Journal Press, Augusta, ME., A-2, p. 1-18.

Boone, G.M. (1973) Metamorphic stratigraphy, petrology, and structural geology of the Little Bigelow Mountain map area, western Maine. Maine Geol. Surv. Bull. 24, Dept. of Conservation, Augusta, ME. 136 p.

Burruss, R.C. (1977) Analysis of fluid inclusions in graphitic metamorphic rocks from Bryant Pond, Maine and Khtada Lake, British Columbia: Thermodynamic basis and geologic interpretation of observed fluid compositions and molar volumes. Ph.D. thesis, Princeton University, Princeton, NJ. 156 p.

Bussa, K.L. (1973) Sulfide-silicate equilibria in rocks of staurolite grade in N.W. Maine. M.S. thesis, Univ. of Wisconsin, Madison, Wisconsin. 79 p.

Carmichael, D.M. (1978) Metamorphic bathozones and bathograds: a measure of the depth of post-metamorphic uplift and erosion on the regional scale. Amer. Jour. Sci., 278, p. 769-797.

Cheney, J.T. (1975) Mineralogy and petrology of lower sillimanite through sillimanite + K-feldspar zone, pelitic schists, Puzzle Mountain area, Northwest Maine. Ph.D. thesis, University of Wisconsin, Madison, Wisconsin. 291 p.

Cheney, J.T. and Guidotti, C.V. (1973) Paragonite contents of coexisting, but texturally different muscovites from pelitic schists of the Puzzle Mountain area, Maine. Amer. Mineral, 58, p. 1076-1079.

Cheney, J.T. and Guidotti, C.V. (1979) Muscovite-plagioclase equilibria in sillimanite + quartz-bearing metapelites, Puzzle Mountain area, northwestern Maine. Amer. Jour. Sci., 279, p. 411-434.

Conatore, P.D. (1974) Mineralogy and petrology of the transition from the garnet zone to the lower sillimanite zone, north-central Rumford area, Maine. M.S. thesis, University of Wisconsin, Madison, Wisconsin. 139 p.

Cullers, R.L., Yeh, L., Chaudhuri, S., and Guidotti, C.V. (1974) Distribution of the rare earths in a series of pelitic schists from N.W. Maine: A sedimentary problem. Geochim. Cosmochim. Acta., 38, p. 389-400.

Foster, C.T., Jr. (1975) Diffusion controlled growth of metamorphic segregations in sillimanite grade pelitic rocks near Rangeley, Maine, U.S.A. Ph.D. thesis, The Johns Hopkins University, Baltimore, MD.

Foster, C.T. (1977) Mass transfer in sillimanite-bearing pelitic schists near Rangeley, Maine. Amer. Mineral, 62, p. 727-746.

Guidotti, C.V. (1970A) The mineralogy and petrology of the transition from lower to upper sillimanite zone in the Oquossoc area, Maine. Jour. Petrol., 11, p. 277-336.

Guidotti, C.V. (1970b) Metamorphic petrology, mineralogy, and polymetamorphism in a portion of N.W. Maine: In Boone, G.M. ed., 1970, New England Inter. Coll. Geol. Conf. 62'nd Ann. Mtg. Field Trip B-1, p. 1-29.

Guidotti, C.V. (1973) Compositional variation of muscovite as a function of metamorphic grade and assemblage in metapelites from N.W. Maine. Contr. Mineral. Petrol., 42, p. 33-42.
Guidotti, C.V. (1974) Transition from staurolite to sillimanite zone, Rangeley quadrangle, Maine. Geol. Soc. Amer. Bull., 85, p. 475-490.
Guidotti, C.V. (1978) Compositional variation of muscovite in medium- to high-grade metapelites of northwestern Maine. Amer. Mineral, 63, p. 878-884.
Guidotti, C.V. (1978a) Muscovite and K-feldspar from two-mica adamellites in northwestern Maine: Composition and petrogenetic implications. Amer. Mineral., 63. p. 750-753.
Guidotti, C.V., Herd, H.H. and Tuttle, C.L. (1973) Composition and structural state of K-feldspar from K-feldspar + sillimanite grade rocks in northwestern Maine. Amer. Mineral., 58, p. 705-716.
Guidotti, C.V., Cheney, J.T. and Conatore, P.D. (1975) Interrelationship between Mg/Fe ratio and octahedral Al content in biotite. Amer. Mineral., 60, p. 849-853.
Guidotti, C.V., Cheney, J.T. and Conatore, P.D. (1975) Coexisting cordierite + biotite + chlorite from the Rumford quadrangle, Maine. Geology, 3, no. 3, p. 147-148.
Guidotti, C.V. and Cheney, J.T. (1976) Margarite pseudomorphs after chiastolite in the Rangeley area, Maine. Amer. Mineral, 61, p. 431-434.
Guidotti, C.V., Cheney, J.T., and Guggenheim, S. (1977) Distribution of titanium between coexisting muscovite and biotite in pelitic schists from northwestern Maine. Amer. Mineral., 62, p. 438-448.
Harwood, D.S. (1973) Bedrock Geology of the Cupsuptic and Arnold Pond quadrangles, west-central Maine. U.S. Geol. Surv. Bull. 1346, 90 p.
Henry, W.E. (1974) Metamorphism of the pelitic schists in the Dixfield Quadrangle, northwestern Maine. Ph.D. thesis, University of Wisconsin, Madison, Wisconsin. 127 p.
Hollister, L.S., Burruss, R.C., Henry, D.L., and Hendel, E-M. (1979) Physical conditions during uplift of metamorphic terranes as recorded by fluid inclusions. Bull. Mineral., 102, p. 555-561.
Lonker, S.W. (1975) Geology of the Mt. Zircon Quadrangle, northwestern Maine. M.S. thesis, University of Wisconsin, Madison, Wisconsin. 178 p.
Mallio, W.J. and Gheith, M.A. (1972) Textural and chemical evidence bearing on sulfide-silicate reactions in metasediments. Mineral Deposita, 7, p. 13-17.
Miller, D.W. (1979) Petrology and origin of migmatites in the Bryant Pond area, northwestern Maine. M.S. thesis, University of Wisconsin, Madison, Wisconsin. 157 p.
Moench, R.H. and Zartman, R.E. (1976) Chronology and styles of multiple deformation, plutonism, and polymetamorphism in the Merrimack synclinorium of western Maine: in Lyons, P.C. and Brownlow, A.H., eds., Studies of New England Geology: Geol. Soc. Amer. Mem. 146, p. 203-238.
Moody, W.C., Jr. (1974) Origin of the Plumbago Mountain Mafic-Ultramafic pluton in the Rumford Quadrangle, Maine. M.S. thesis, University of Wisconsin, Madison, Wisconsin. 90 p.
Tewhey, J. (1975) The controls of chlorite-garnet equilibria in the contact aureole of the Cupsuptic Pluton, west central Maine; and the two-phase regions in the CaO-SiO2 system: experimental data and thermodynamic analysis. Ph.D. thesis, Brown University, Providence, Rhode Island. 147 p.

Papers in the 1980's

Aleinikoff, J.N. Moench, R.H. and Lyons, J.B. (1985) Carboniferous U-Pb age of the Sebago batholith, southwestern Maine. Geol. Soc. Amer. Bull., 69, p. 990-996.
DeYoreo, J.J., Lux, D.R., Guidotti, C.V., Decker, E.R., and Osberg, P.H. (1988) The Acadian thermal history of western Maine. J. Metamorphic Geol., 7, p. 169-190.
DeYoreo, J.J., Lux, D.R. and Guidotti, C.V. (1989) A thermal model for Carboniferous metamorphism near the Sebago batholith in western Maine. Maine Geol. Survey, C.T. Jackson Commemorative Volume, 30 p.

Dickerson, R.P. (1984) A study of polymetamorphism and mineral chemistry of the Little Bigelow Mountain and Bingham quadrangles, Maine. M.S. thesis, Southern Methodist University, Dallas, Texas.

Dutrow, B.L. (1985) A staurolite trilogy: III. Evidence for multiple metamorphic episodes in the Farmington quadrangle, Maine. Ph.D.thesis, Southern Methodist University,Dallas,Texas.

Eusden, J. Dykstra and Barreiro, B. (1989) The timing of peak high-grade metamorphism in central-eastern New England. Mar. Sed. and Atlant. Geol., 24, p. 241-255.

Guidotti, C.V. (1985) Metamorphic Map of Maine in Osberg, P.H. et al. eds., Bedrock Geologic Map of Maine. Maine Geol. Survey, Augusta, Maine.

Guidotti, C.V. (1989) Metamorphism in Maine: An Overview. Maine Geol. Surv. Studies in Maine Geology: 3, 17 p. Ms.

Guidotti, C.V., Trzcienski, W.E. and Holdaway, M.J. (1983) A northern Appalachian metamorphic transect - Eastern Townships, Quebec to the central Maine coast: in Schenk, P.E. ed., Regional trends in the geology of the Appalachian- Caledonide- Hercynian- Mauritanide Orogen. D. Reidel Pub. Co., Dordrecht, Holland, p. 235-247.

Guidotti, C.V., Gibson, D., Lux, D.R. and DeYoreo, J.J. and Cheney, J.T. (1986) Carboniferous metamorphism on the north (upper) side of the Sebago Batholith: in Newberg, D.W. ed., New England Intercoll. Geol. Conf. Guidebook for field trips in southwestern Maine, p. 306-341.

Guidotti, C.V., Cheney, J.T. and Henry, D.J. (1988) Compositional variation of biotite as a function of metamorphic reactions and mineral assemblage in the pelitic schists of western Maine. Amer. Jour. Sci., 288A, p. 270-292.

Henry, D.J. (1981) Sulfide-silicate relations of the staurolite grade pelitic schists, Rangeley Quadrangle, Maine. Ph.D. thesis, University of Wisconsin, Madison, Wisconsin. 437 p.

Henry, D.J. and Guidotti, C.V. (1985) Tourmaline as a petrogenetic indicator in staurolite-grade metapelitic rocks. Amer. Mineral., 70, p. 1-15.

Holdaway, M.J., Guidotti, C.V., Novak, J.M. and Henry, W.E. (1982) Polymetamorphism in medium- to high-grade pelitic metamorphic rocks, west-central Maine. Geol. Soc. Amer. Bull., 93, p. 572-584.

Holdaway, M.J. Dickerson, R.P., and Dutrow, B.L. (1986) Petrology and field relations of M_2 metamorphism in west-central Maine: in Newberg, D.W. ed., New England Intercoll. Geol. Conf. Guidebook for field trips in southwestern Maine, p. 240-253.

Holdaway, M.J., Dutrow, B.L. and Hinton, R.W. (1988) Mineral chemistry and metamorphic conditions of M_3 and M_5, west-central Maine: the muscovite-almandine-geobarometer; the staurolite problem revisited. Amer. Mineral, 74, p. 20-47.

Hussey, A.M. II, Bothner, W.A. and Thomson, J.A. (1986) Geologic comparisons across the Norumbega Fault zone, southwestern Maine: in Newberg, D.W., ed. NEIGC guidebook for field trips in southwestern Maine, p. 53-79.

Kimball, K.L. (1981) Phase relations in coexisting Fe-Mg and Ca Amphiboles. Ph.D. thesis, University of Wisconsin, Madison, Wisconsin. 227 p.

Lux, D.R. and Guidotti, C.V. (1985) Evidence for extensive Hercynian metamorphism in western Maine. Geology, 31, p. 696-700.

Lux, D.R., DeYoreo, J.J., Guidotti, C.V., Decker, E.R. (1986) The role of plutonism in low pressure/high temperature metamorphic belt formation. Nature, 323, p. 794-797.

Lux, D.R., and DeYoreo, J.J. (1987) Magmatic erosion of the Acadian orogenic root. Abst. with Programs Ann. Mtg. N.E. Section Geol. Soc. Amer., 19, p. 27.

Teichmann, F. (1988) The geochemistry of chlorites as a function of metamorphic grade, mineral assemblage and bulk composition in the Rangeley and Rumford areas, Maine. M.S. thesis, University of Maine, Orono, Maine. 175 p, plus appendices.

Thomson, J.A. (1986) The occurrence of kyanite in southern Maine and its metamorphic implications. M.S. thesis, University of Maine, Orono, Maine. 129 p.

Thomson, J.A. and Guidotti, C.V. (1986) The occurrence of kyanite in southern Maine and its metamorphic implications. NE Sectional Meeting, Geol. Soc. Amer., 18, no. 1, p. 72.

Thomson, J.A. and Guidotti, C.V. (1989) Carboniferous Barrovian metamorphism in southern Maine. Maine Geol. Survey, C.T. Jackson Commemorative Volume. 28 p. Ms.
Wood, C. (1981) Chemical and Textural Zoning in Metamorphic Garnets, Rangeley area, Maine. Ph.D. Thesis, University of Wisconsin, Madison, Wisconsin. 586 p.

A FIELD TRIP GUIDE TO BLACK MOUNTAIN, WILDWOOD ROADCUT AND BEAVER BROOK, MT. MOOSILAUKE AREA, NEW HAMPSHIRE

Douglas Rumble, III

and

M.P. Dickenson

INTRODUCTION

The purpose of the field trip to Black Mountain, Wildwood Roadcut and Beaver Brook is to study three localities that show well exposed evidence of different types of fluid-rock interactions. The rocks of Black Mountain and Wildwood roadcut show examples of the bed-by-bed control of fluid compositions by mineral assemblage buffering. Infiltration, in contrast, is the prevalent mode of fluid-rock behavior at Beaver Brook. An important question to keep in mind throughout the field trip is "Why did Beaver Brook experience infiltration during metamorphism while Black Mountain and Wildwood did not?"

MAPS AND LITERATURE

The area of the field trip is covered by two 7-1/2', 1/24,000, topographic maps: East Haverhill, N.H. and Mt. Moosilauke, N.H. (Fig. C-1). A key reference on the geology and petrology of the area is the seminal work of Marland P. Billings "Regional metamorphism of the Littleton-Moosilauke area, N.H." (1937). The distribution of Al_2SiO_5 minerals in the region is described by Rumble (1973). Pressure-temperature (P-T) estimates are given by K.V. Hodges and F.S. Spear (1982). The Black Mtn. and Beaver Brook localities are discussed, respectively, by Rumble (1978) and by Rumble, Ferry, Hoering, and Boucot (1982). The Wildwood roadcut is presented in this guidebook by M.P. Dickenson.

STRATIGRAPHY AND GEOLOGIC HISTORY

The oldest stratified rocks in the Mt. Moosilauke region are the Ammonoosuc Volcanics of Ordovician age. There are at least several kilometers of felsic and mafic volcanics in the Ammonoosuc. A more accurate estimate of thickness cannot be made because the bottom of the unit is cut out by magnetite-bearing granitoids of the Ordovician Oliverian magma series. Intrusive rocks of the Oliverian series occupy the cores of domal uplifts and are mantled by Ammonoosuc volcanics and younger rocks. An Ordovician magmatic arc composed of Ammonoosuc Volcanics and Oliverian plutons is known to 38extend parallel to regional strike trends from Connecticut, across Massachusetts, along the 38western edge of New Hampshire, and finally curving eastward into Maine. The arc is now exposed within a Devonian structure, the Bronson Hill anticlinorium.

Rocks of the Ordovician magmatic arc are overlain unconformably by (1) Silurian conglomerates and quartzites of the Clough Quartzite; (2) marbles and calc-silicates of the Silurian Fitch Formation; and (3) pelitic mica schists and micaceous quartzites of the Lower

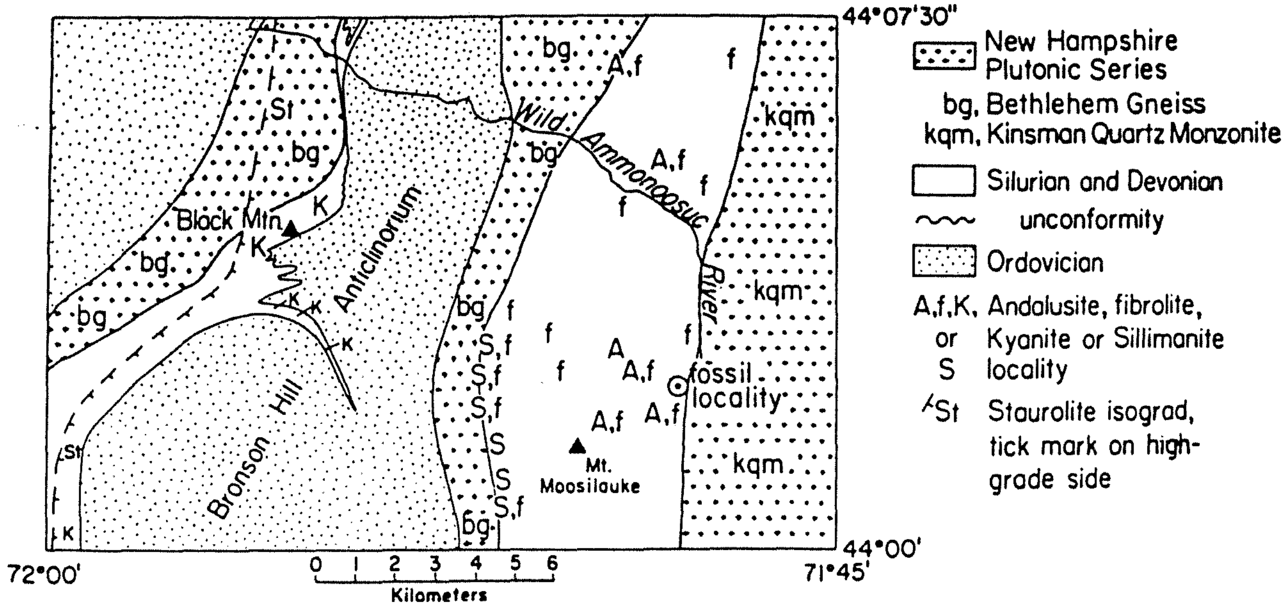


Figure 1. Location of Black Mountain, Beaver Brook fossil locality, and distribution of Al2SiO5 polymorphs in southern half of Mt. Moosilauke 15' quadrangle. Wildwood roadcut is near the first "o" in the label "Wild Ammonoosuc River".

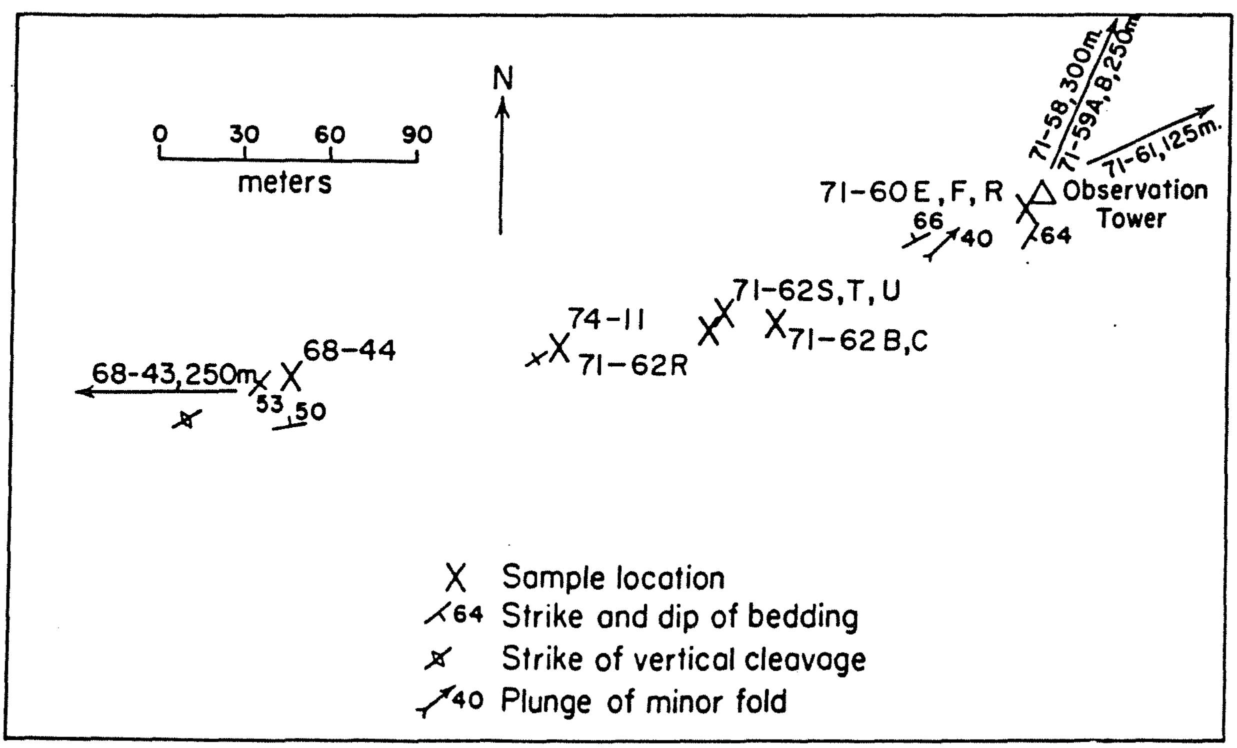


Figure 2. Location of samples collected along summit ridge of Black Mtn. Field trip route begins at site of observation tower and proceeds SW, towards 68-44.

Devonian Littleton Formation. The thicknesses of the Clough and Fitch are highly variable: in some areas the units are absent because of lack of deposition or boudinage; in other areas, as at Black Mountain, the exposed outcrop breadth of the Clough is hundreds of meters, however, some of this apparent thickness may be due to repetition by isoclinal folding. The Littleton Formation may be several kilometers thick but its top is not exposed owing to uplift and erosion. Sections of the Littleton show well developed graded beds, 10-20 cm thick; in other outcrops only massive mica schists are seen. The Littleton is intruded by foliated sill-like plutons of ilmenite-bearing granitoids, the Bethlehem Gneiss and Kinsman Quartz Monzonite of the New Hampshire series. The plutons show neither a contact metamorphic aureole nor chill zones. These features, combined with the presence of primary igneous muscovite, suggest emplacement during active regional metamorphism at depths of 10 km, or more.

The age of regional metamorphism and deformation is constrained by the presence of unaffected Mesozoic intrusives and volcanics of the White Mountain magma series in nearby Franconia Notch. Metamorphism and deformation were traditionally assigned to the Devonian, an age confirmed by the Rb-Sr dating of the syn-metamorphic Bethlehem and Kinsman at 405-415 m.y. Deformation and metamorphism were synchronous during early stages of tectonism: porphyroblasts grew and the preferred orientation of schistosity was established during emplacement of large nappes. Later, metamorphic intensity waned but folding continued as the Bronson Hill anticlinorium and its component domes were built. The sense of transport during the nappe episode was from E to W; in consequence, hotter rocks were stacked on top of cooler rocks. The domes of the Bronson Hill not only refold the nappes but also deform isograds.

BLACK MOUNTAIN

Access

Black Mtn. may be reached by following New Hampshire route 25 to the village of East Haverhill (about 5 miles E. of the intersection of route 10 and route 25). A short detour, a few hundred meters E. of the village on rt. 25, reveals an impressive view of Black Mtn., Sugarloaf Mtn., and the Hogsback, all composed of white Clough quartzite. the arcuate quartzite cliffs of the Hogsback delineate the northerly plunge of the Owls Head dome. The thick accumulations of quartzite at the Sugarloaf and Black Mtn. illustrate the effects of isoclinal folding in the nappes. Returning to East Haverhill village, turn N. and follow Lime Kiln Road to Lime Kiln Camps. The IMA trip will follow Chippewa Trail to the top of Black Mtn. Those visiting at another time can ask for directions in the camp office.

Hiking Route

The group will proceed directly to the summit of Black Mtn. We will begin detailed study of outcrops at the summit and then retrace our steps slowly downward so that we can stay together for good discussions. The rock types that will be seen are conglomerates, quartzites, and quartz-mica schists of the Clough Quartzite.

The summit was formerly marked by an observation tower (Fig. C-2). One can still find the location of its concrete foundation. Facing S. from the site of the tower, the back sides of Hogsback and Sugarloaf are visible. The northerly plunge of the Owls Head Dome beneath Black Mtn. can be easily visualized. The structure to the N. is not well displayed but mapping shows a southerly plunging dome. Thus, Black Mtn. lies in a structural saddle formed by a plunge depression between two domes. Other summits underlain by Clough Quartzite stand to the S. including Piermont Mtn., Mt. Cube, and Croydon Peak. They trace the continuation of the Bronson Hill anticlinorium.

Oxygen isotope geothermometry of Black Mtn. rocks (quartz-magnetite) gives a temperature estimate of 495°C (±10°C). Pressure is estimated to lie within 0.5 kbars of the Al_2SiO_5 triple point, i.e. 3.8 kbars according to Holdaway.

In ledges a few meters S. of the tower, there is abundant kyanite occurring as felted layers defining schistosity. Note how kyanite has been folded. the minor folds that fold kyanite are parasitic to the domes.

Turn back N. and look down at the tower foundations. Here you will find chloritoid-staurolite-chlorite assemblages of 71-60F, 71-60F). The Fe-rich, garnet-bearing assemblages have reduced oxides with $FeTiO_3$ in ilmenite of 95%, or more. (Data on mineral assemblages summarized in figures C-3 through C-10.)

Walk just to the E., beyond the tower site and you will find tourmaline-filled joints. Careful searching may find beds of staurolite-chloritoid-garnet assemblages that have been almost completely replaced by tourmaline, within a few cm of the joints.

Before beginning the descent, look SE. Mt. Moosilauke dominates the skyline. Beaver Brook, the third stop of today's trip, lies on the E. slopes of Mt. Moosilauke, and Wildwood roadcut at the N. end of the mountain. Mica schists on the slopes of the mountain contain abundant andalusite and sillimanite. The site of the Al_2SiO_5 triple point is between where we are standing on Black Mtn. and Mt. Moosilauke.

Turn towards the SW and begin to descend towards the cars. Remember to keep along the rocky spine of the ridge in order to see the best outcrops. The first steep scramble downward, just a few 10's of meters from the tower site, leads past abundant dome-stage minor folds and kyanite schists. the kyanite schists typically have the assemblage quartz-muscovite-kyanite-ilmenohematite-magnetite: they are highly oxidized.

Beyond the steep scramble one reaches a knob on the ridge. Just S. of the knob is a gently-dipping, tournaline-coated joint. The knob is the site of a chlorite schist with staurolite-chloritoid-chlorite-magnetite-ilmenite ($FeTiO_3$ = 98%). Adjacent to the chlorite schist are outcrops of kyanite-staurolite-chloritoid-chlorite-magnetite-ilmenohematitem ($FeTiO_3$ = 28%) schist (71-62B, 71-62S, 71-62T, 71-62U). Here, separated by a meter or so, are rocks whose fluids differed by 20-30% in X_{H2O} and by the same amount (antipathetically) in X_{H2}. Oxygen isotope analyses of co-existing quartz and magnetite show the minerals to have the same partitioning of $^{18}O/^{16}O$ but the quartz and magnetite of 71-62U and 71-62B are 1°/oo enriched in $\delta\ ^{18}O$ relative to 71-62R. thus, inferences of fluid inhomogeneity based on mineral assemblages are supported by oxygen isotope analyses.

Turn SW from the knob and continue to descend, following blazed trail through the scrub to avoid a steep place on the ridge. Shortly, the trail emerges on broad, open, bare-rock ledges. The kyanite-chloritoid-staurolite-paragonite quartzite of 68-44 is located in low ledges N. of the trail. In thin sections it may be seen that kyanite chloritoid, and staurolite are intergrown epitaxially, presumably along the chains of aluminum octahedra they all share.

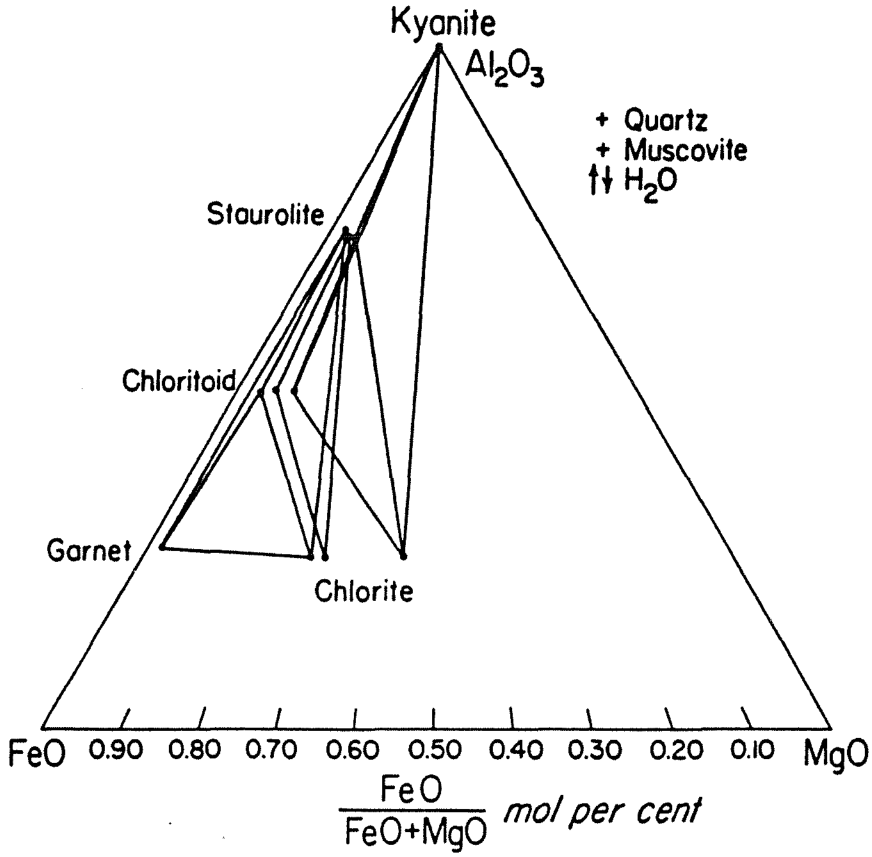


Fig. 3. Thompson projection of kyanite, staurolite, chloritoid, garnet, and chlorite compositions (ubiquitous minerals quartz and muscovite; also, projection through H_2O).

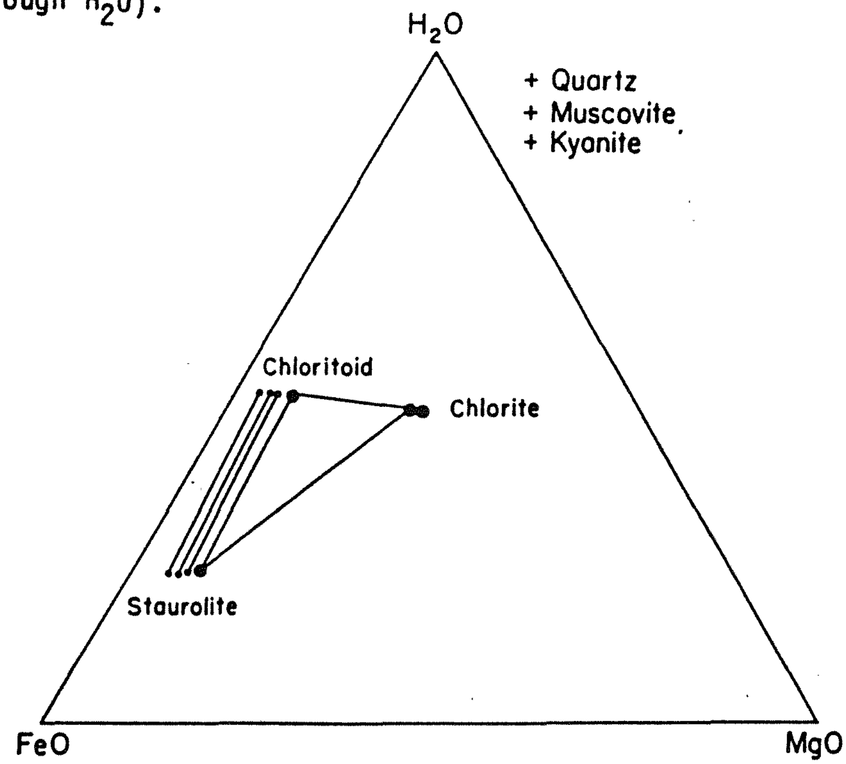


Fig. 4. Projection of chloritoid-chlorite-staurolite compositions from the ubiquitous minerals quartz-muscovite-kyanite.

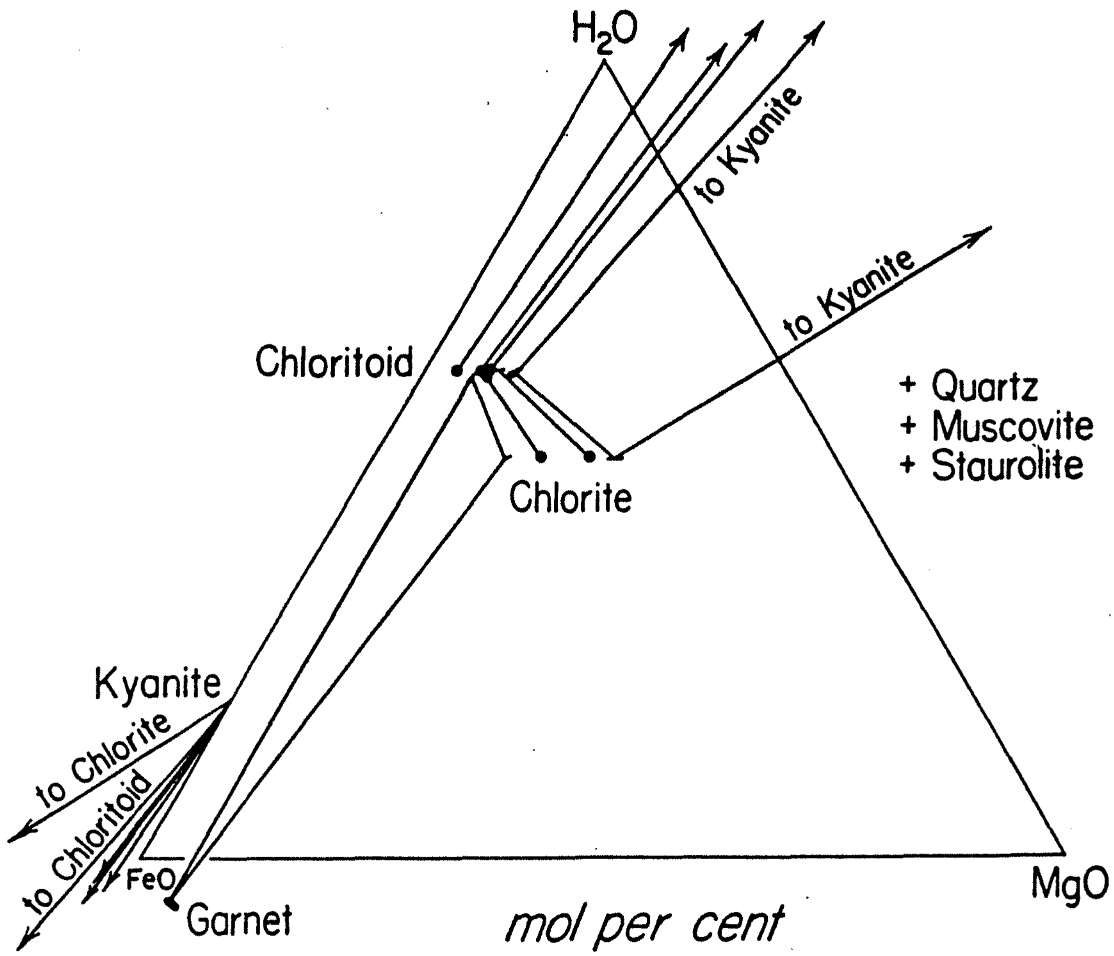


Fig. 5. Projection of chloritoid-chlorite-garnet-kyanite from the ubiquitous minerals quartz-muscovite-staurolite.

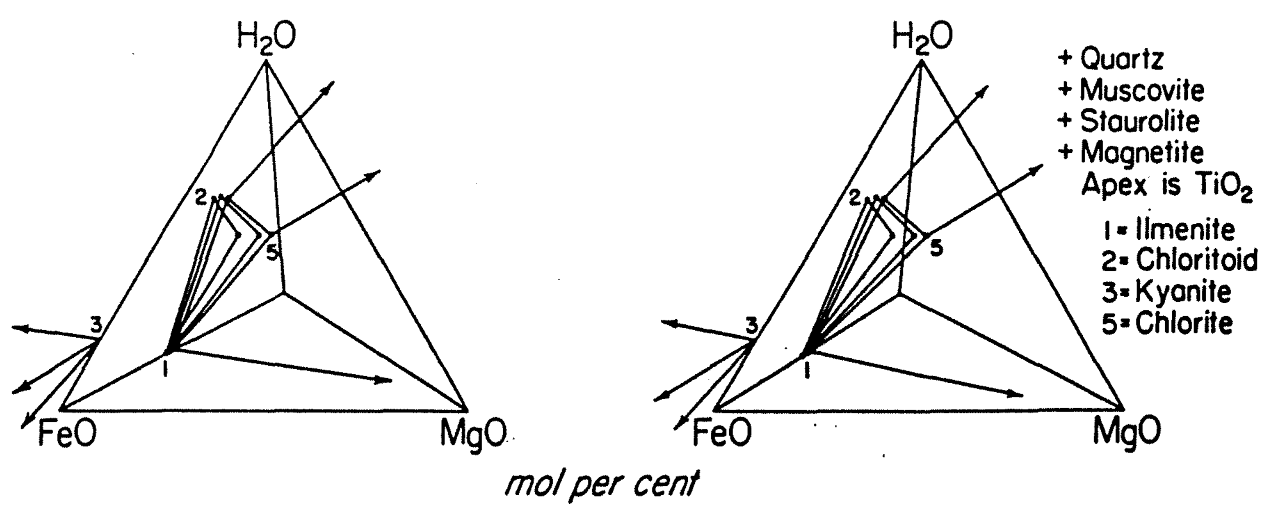


Fig. 6. Stereographic pair of tetrahedra showing projection of ilmenite-chloritoid-chlorite-kyanite from ubiquitous minerals quartz-muscovite-staurolite-magnetite. Apex is TiO_2.

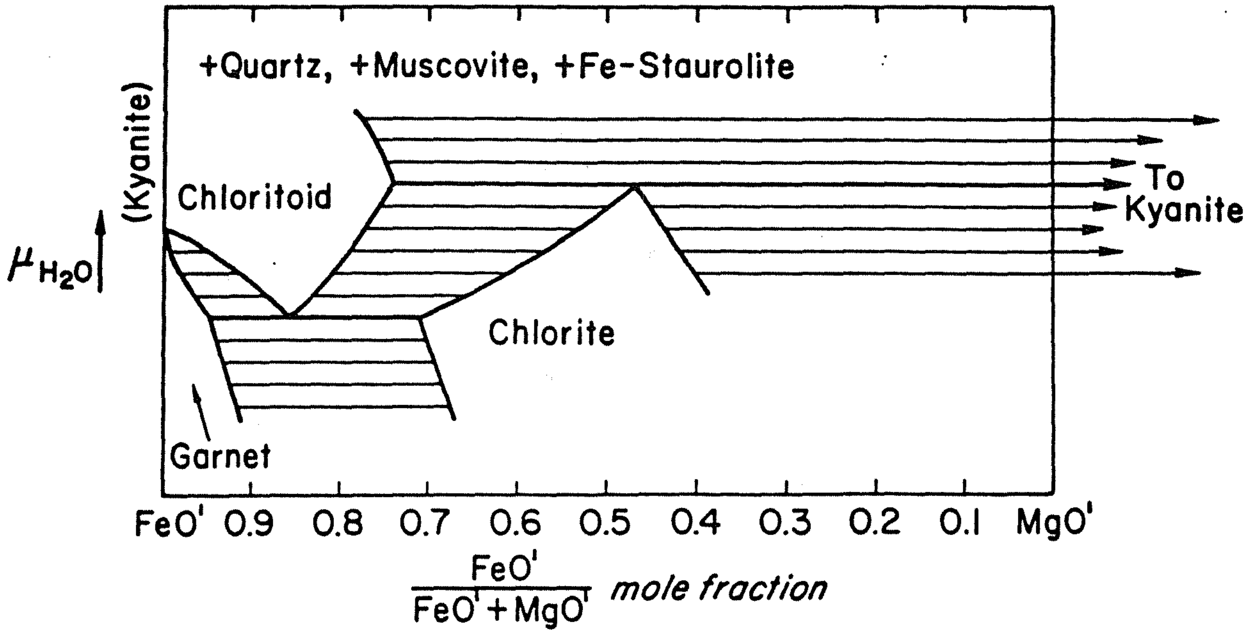


Fig. 7. Isothermal, isobaric variation of μH_2O with mineral assemblage and mineral composition (cf. Fig. C-5).

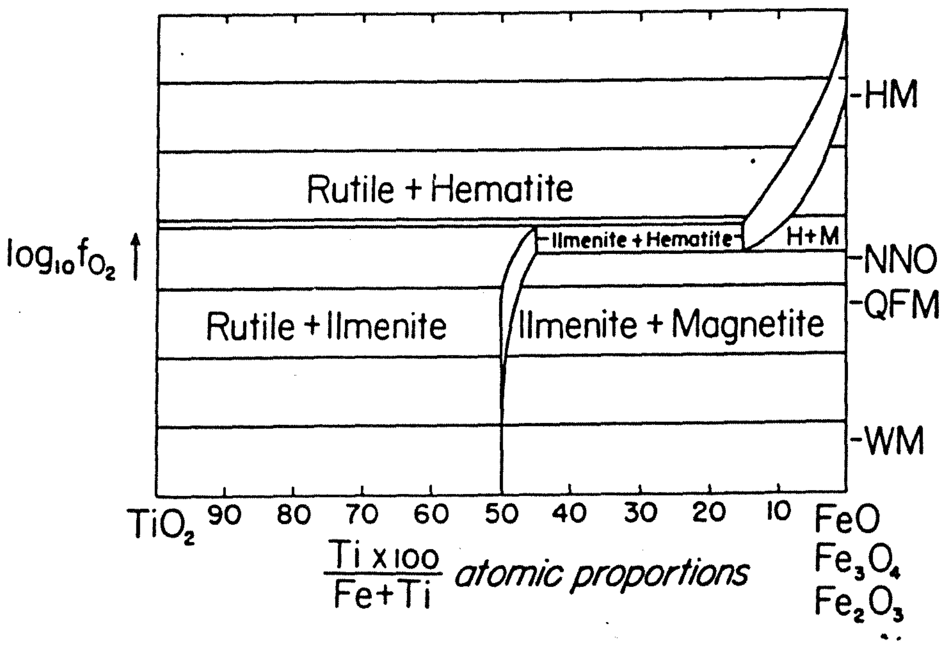


Fig. 8. Isothermal, isobaric variation of fO_2 with mineral assemblage and mineral composition.

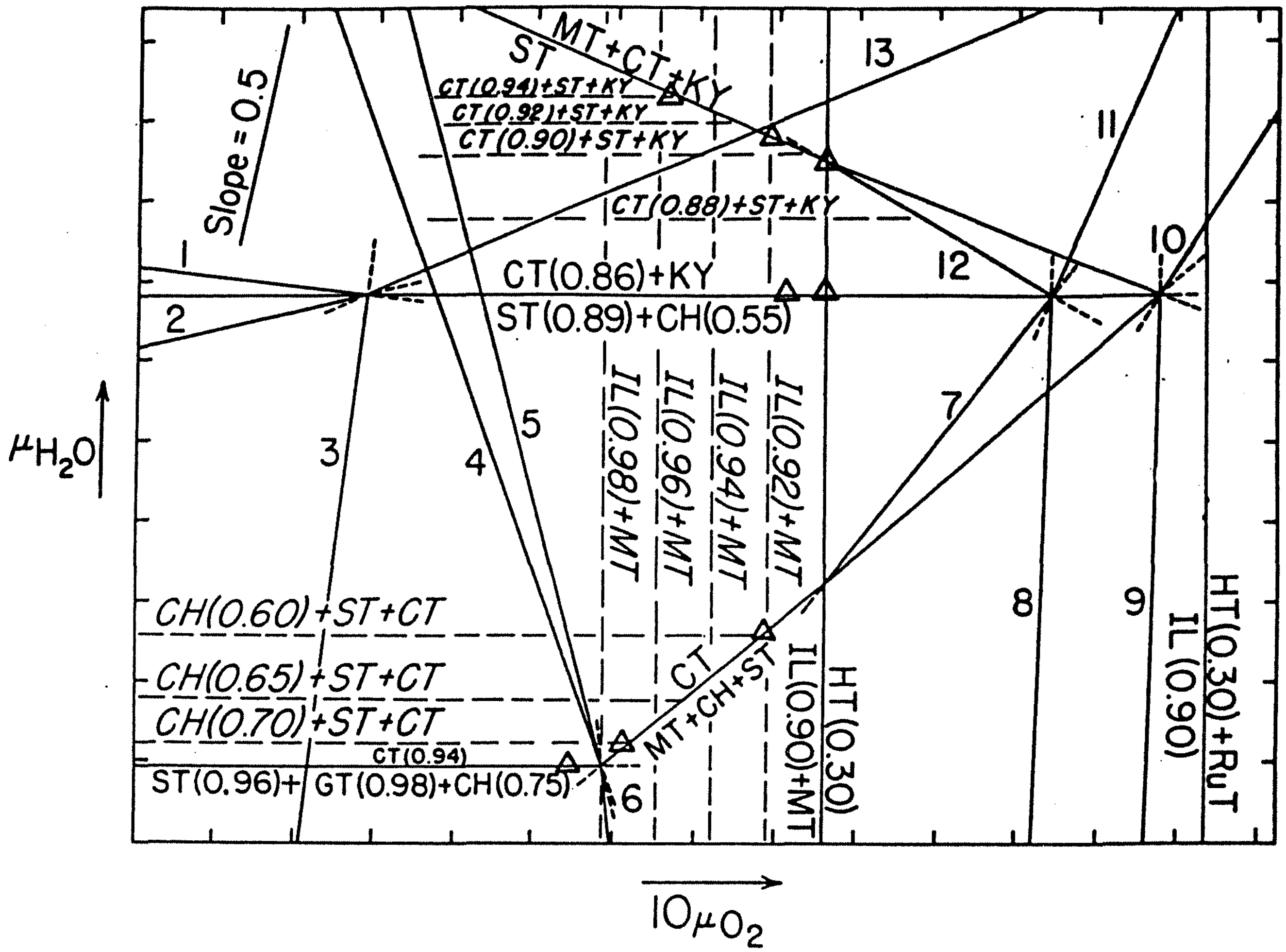


Fig. 9. Variation of μH_2O and μO_2 with mineral assemblage and mineral composition. Triangles show relative positions of observed assemblages.

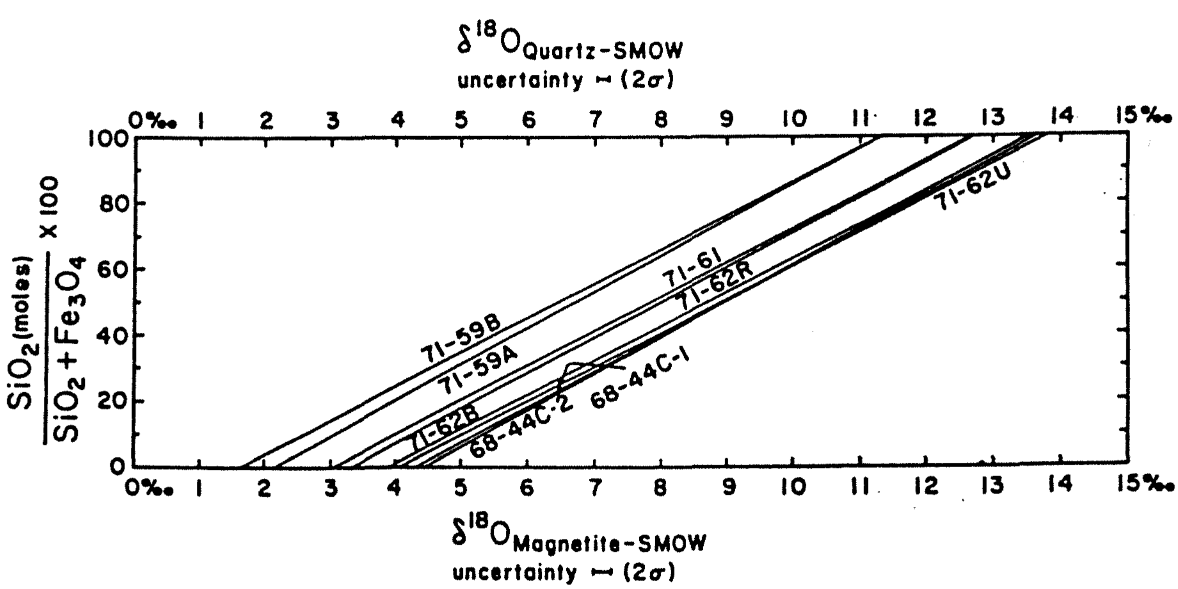


Fig. 10. Oxygen isotope exchange equilibrium diagram for coexisting quartz and magnetite.

Turn SW and continue to follow the blazed trail through the ledges of quartzite. Be on the lookout for current bedding, tourmaline-covered joints, and quartz-mica schist with very Fe^{2+}-rich garnets. The garnets may be readily recognized because of their tendency towards rusty weathering.

Return to cars for the drive to Wildwood roadcut.

WILDWOOD ROADCUT

Access

Wildwood roadcut is located on New Hampshire route 112, 1.2 km E. of its intersection with route 116. The roadcut is near the first "o" in the label "Wild Ammonoosuc River" on Fig. C-1.

Location of Rock Types

The phase petrology of rocks in the roadcut is discussed by M.P. Dickenson in the following sections. The andalusitechlorite-biotite-staurolite assemblages are located at the E. end of the roadcut on both sides of the road. Andalusite occurs as dark, non-descript porphyroblasts. An excellent example of reaction textures may be reached by proceeding to the W. end of the roadcut and climbing up the N. wall of the roadcut. Here, on westward dipping schistosity surfaces, may be seen pseudomorphs after cigar-size, cigar-shape aluminosilicate porphyroblasts (prismatic sillimanite? staurolite?). The pseudomorphs now consist of quartz, a mesh of muscovite, small garnets, and clusters of mm-sized euhedra of staurolite. M.P. Billings first described these textures in 1937. The pseudomorphs are found scattered on the W. slopes of Mt. Moosilauke. At Hurricane Mtn. (about 4 km S. of Mt. Moosilauke) the pseudomorphs contain andalusite euhedra rather than staurolite.

Elsewhere in the roadcut "coticules" (i.e. spessartine garnet rocks) may be found.

Introduction

The classical model of metamorphism assumes that mineral assemblages on an outcrop are in equilibrium with a reservoir of external fluid of fixed composition. Thus all mineral assemblages are assumed to have equilibrated with a common volatile or fluid phase. Recent studies, however, have shown that along a single outcrop the composition of the volatile phase, as deduced from petrologic/mineralogic data, varies with changes in the mineral assemblage (Ferry, 1979; Rumble, 1974, 1978; others). Furthermore, at fixed temperature and pressure within a given mineral assemblage, the change in mineral composition for various solid solution minerals is also related to changes in the composition of the volatile phase (Rumble, 1976a, 1978; Dickenson, 1984). The studies clearly demonstrate that the classical model for metamorphism is not always valid.

This study presents mineralogical and petrological data from a single outcrop of amphibolite grade metamorphosed pelite schist. The results indicate several important concepts related to pelite metamorphism: (1) Differences in the chemical potential of H_2O are recorded in bed-to-bed relationships; (2) these differences are probably related to initial compositional parameters and the buffer capacity of the mineral assemblage; and (3) the magnitude of chemical potential differences is small, suggesting little or no infiltration of externally derived fluid.

Geologic Setting

Samples were collected along a single outcrop of amphibolite grade pelitic schist on Route 112 in the Moosilauke Quadrangle, N.H. The rock unit was originally mapped as Littleton Formation, an early Devonian stratigraphic unit (Billings, 1937).

The outcrop lies within a N-S band of pelitic rocks called the Moosilauke septum. this band is typically of medium amphibolite grade, commonly including staurolite-andalusite and/or sillimanite pelitic assemblages. Mineral textures reveal a complex reaction history that is most likely a function of a complicated history of changes in intensive parameters (pressure, temperature, and fluid composition) in response to tectonic evolution. This study examines only the latest texturally equilibrated mineral assemblages and does not address the complicated reaction history.

Analytical Methods

Mineral analyses were carried out on a CAMECA automated electron microprobe located at Harvard University. All analyses were collected using 15 kV accelerating potential with a beam current of approximately 15 nA, measured by a retractable Faraday Cup. X-ray intensities were converted to oxide weight percents using the methods of Albee and Ray (1970), and simple silicates and oxides were used as standards. Reported mineral compositions represent the average of several separate grains with two to four analyses collected on each grain.

Petrography and Mineral Chemistry

The mineral assemblages described in this study occur in amphibolite grade pelitic schists. the critical assemblages addressed include three varieties (all with quartz, muscovite, garnet, ilmenite and graphite):

(1) chlorite + biotite + staurolite
(2) chlorite + biotite + staurolite + andalusite
(3) chlorite + biotite + andalusite.

Although the assemblages record a history of crystallization, mineral textures indicate that the above assemblages are the last texturally equilibrated assemblages. For example, andalusite in assemblage (3) contains inclusions of staurolite but no staurolite in the matrix. Likewise, chlorite in all assemblages appears as late porphyroblasts, crosscutting the latest foliation.

Mineralogy

Chlorites analyzed in this study are solid solutions of $F_{4.5}Al_3Si_{2.5}O_{10}(OH)_8$-$Mg_{4.5}Al_3Si_{2.5}O_{10}(OH)_8$. Biotites are solid solutions of the components $KFe_{2.5}Al_2Si_{2.5}o_{10}(OH)_2$-$KMg_{2.5}Al_2Si_{2.5}O_{10}(OH)_2$ and contain up to 1.75 wt % TiO_2. Muscovites are solid solutions of $KAl_3Si_3O_{10}(OH)_2$ - $NaAl_3Si_3O_{10}(OH)_2$ -

$KFeAlSi_4O_{10}(OH)_2$ - $KMgAlSi_4O_{10}(OH)_2$ with relatively constant Na/K, constant (Fe + Mg)/Si and containing up to 1.1 wt % FeO + MgO and 0.40 wt % TiO_2. Both biotite and muscovite contain deficiencies of alkalies (Guidotti, 1974; Rumble, 1978).

Staurolites display a limited amount of solid solution of the components $Fe_2Al_9Si_4O_{24}H_2$-$Mg_2Al_9Si_4O_{24}H_2$-$Fe_{1.625}Al_{9.25}Si_4O_{24}H_2$, with the Al_2O_3 content decreasing with increasing MgO/FeO in assemblage (1) (Fig. C-11). The mineral contains up to 0.15 wt % MnO with little ZnO (0.25 wt %) and TiO_2 (0.50 wt %).

Andalusites contain approximately 0.25 wt % Fe_2O_3 (measured as FeO) with minor TiO_2. Ilmenites are nearly pure $FeTiO_3$ with minor amounts of SiO_2, Al_2O_3, MgO and MnO. No measurable amounts of hematite components are present. The negligible amounts of observed Fe_2O_3 component are consistent with the ubiquitous presence of graphite in all samples. the ilmenite-graphite oxide assemblage is indicative of reducing conditions and suggests that all Fe-Mg silicates have low to negligible amounts of Fe_2O_3 (Rumble, 1973; 1976b).

Garnets are the only mineral with appreciable amounts of CaO (1.2 wt %) and MnO (2 wt %) but are composed predominantly of almandine component (~83-85 mol %).

Geothermometry and Geobarometry

Temperatures calculated for the sample area are calibrated using the Ferry and Spear (1978) Gt-Bt geothermometer. Pressures are estimated using the Ghent (1976) Plag-Als-Gt-Qtz geobarometer, as several assemblages do contain plagioclase (these are not considered for the analysis of phase equilibria). Results indicate that a temperature of ~250°C and pressures of 3.5-4 kbar are associated with the last texturally equilibrated assemblages. These results are in good agreement with data from Hodges and Spear (1982) compiled from pelitic rocks in the general area of the Moosilauke septum.

Phase Equilibria

Geometric relationships. As discussed above, the classical model of metamorphism assumes that all mineral assemblages in a single outcrop attained equilibrium with a single externally derived fluid of fixed composition (i.e., the chemical potential of water, μH_2O, is constant). This assumption is also implicit in the Thompson AFM projection (Fig. C-11) displaying the various mineral assemblages. Tie-lines for the three mineral assemblages are displayed in Figure C-11 which is a projection from the minerals quartz, muscovite and ilmenite, and assumes a constant value of μH_2O for all assemblages. Garnet is not displayed in the phase diagram in Figure C-11 and will not be considered further due to the solubility of significant amounts of CaO and MnO, which effectively remove Gt off the AFM plane, relegating it as an extra phase in the model KFMASH system. In the model KFMASH system assemblages (1) and (3) are trivariant, while assemblage (2) is divariant. Thus, at constant values of pressure and temperature, assemblage (2) is invariant to changes in intensive variables, while assemblages (1) and (3) are univariant.

As viewed in the AFM projection, various examples of crossing tie-lines are observed. The presence of crossing tie-lines can be caused by several relationships: (1) the mineral assemblages crystallized at different pressure and/or temperature; (2) chemical components other than those considered occur in one or more minerals; (3) the chemical

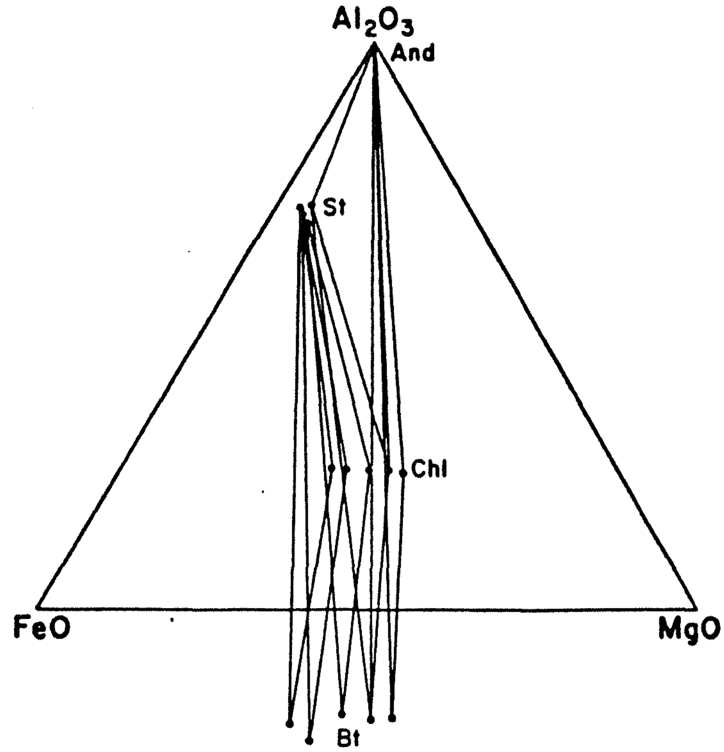


Fig. 11. AFM projection of Wildwood mineral assemblages (And = andalusite, St = staurolite, Chl = chlorite, Bt = biotite, plus quartz and muscovite).

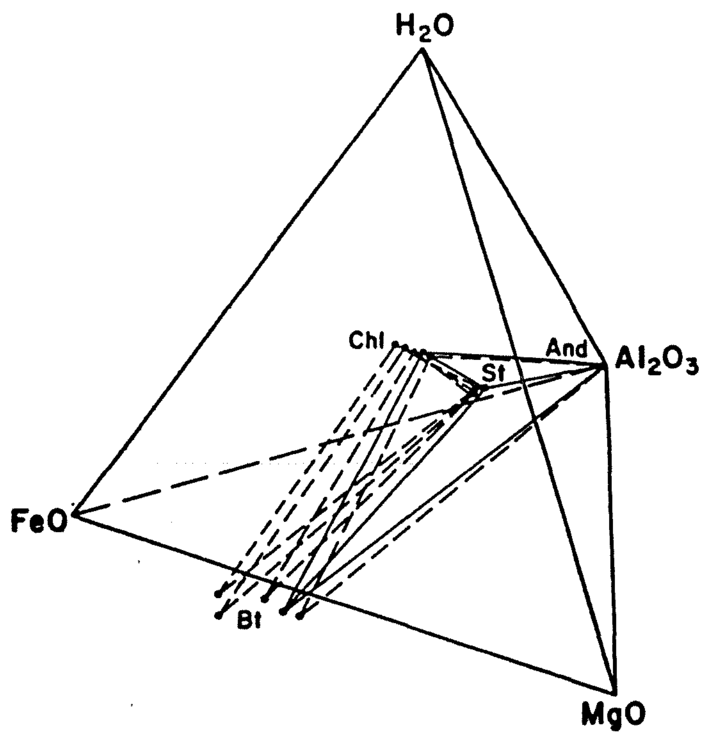


Fig. 12. AFM - H_2O phase diagram of Wildwood assemblages. Three-phase assemblages define planes (Chl-Bt-St and Chl-And-Bt); while the four-phase assemblage (And-St-Chl-Bt) forms a sub-volume within the AFM-H_2O tetrahedron.

potential of water (μH_2O) is not constant for all mineral assemblages. Since all assemblages have crystallized in one outcrop (suggesting constant P and T) and reside mainly in the system KFMASH (minimizing the effects of extra chemical components), only a variable μH_2O appears likely as the cause for crossing tie-lines. The effects of variable μH_2O can be rigorously evaluated by allowing H_2O to be a component of the phase diagram (Fig. C-12). This four-component phase diagram (AFM-H_2O) effectively allows mineral assemblages to be associated with variable values of μH_2O. Inspection of Figure C-12 demonstrates that the problem of crossing tie-lines is resolved by considering H_2O as a component of the phase diagram. Planes of the St-Bt-Chl assemblages lie to the Fe-side of the four-phase volume, St-Bt-Chl-And, while planes of And-Chl-Bt lie to the Mg-side of the four-phase volume (Fig. C-12). This arrangement of three-phase planes and four-phase volume demonstrates that the trivariant, three-phase assemblages [(1) and (3)] may equilibrate with variable values of μH_2O (each separate three-phase plane is associataed with a unique value for μH_2O), while the divariant, four-phase assemblages (2) occurs at a unique and fixed value of μH_2O.

The compositional relationships between the three mineral assemblages in Figure C-12 demonstrate that μH_2O varies from assemblage to assemblage. The diagram can also provide information constraining the relative change in μH_2O for these three assemblages. In a phase diagram like Figure C-12 any ray drawn from one of the component apices describes a decrease in that component as one proceeds along the arbitrary ray away from that component (Korzhinskii, 1959). For example, a ray drawn from the H_2O apex describes decreasing μH_2O along the ray. Construction of such a ray illustrates that the ray would first intersect the most Fe-rich St-Chl-Bt plane and would successively pierce more Mg-rich St-Chl-Bt planes. This geometric relationship is due to the fact that Chl is always the most magnesian phase and acts to "lean" the planes over towards the Mg-side. Thus, in the St-Ch-Bt assemblage decreasing μH_2O is associated with increasing X_{Mg} in the three phases. Similar arguments can be made for assemblages (2) and (3) such that the four-phase volume (assemblage 2) is associated with a value for μH_2O intermediate to that for assemblages (1) and (3), and that in assemblage (3) decreasing μH_2O is also associated with increasing X_{Mg}.

Analytical representation. The change in both mineral composition and chemical potential of water illustrated in the phase diagram in Figure C-12 can be viewed in a more quantitative fashion. The phase equilibria can be analytically formulated using the Gibbs Technique (Spear et al., 1982). This approach quantitatively relates changes in mineral composition to changes in various intensive parameters. In Figure C-13 the change in mineral composition in assemblages (1) and (3) is plotted versus $\Delta\mu H_2O$. Differences in μH_2O are calculated relative to assemblage (2) that has arbitrarily been fixed to a value of 0. The calculations are performed for mineral compositions in the model KFMASH systems. Components such as Na_2O and TiO_2 have been omitted due to the fact that the Na/K ratios in both biotite and muscovite remain constant, and TiO_2 contents are both small and constant in all mineral phases. Ferry (1979) has also shown that small amounts of Na_2O and TiO_2 in biotite and muscovite do not significantly effect calculations of this type.

The diagram in Figure C-13 was calculated by solving for the variable $(\partial\mu H_2O/\partial X^{Bt}_{Fe})$ at constant P and T. This variable was then integrated over values of ΔX_{Fe} of 0.01 and resolved iteratively over the given range of X_{Fe}. Initial mineral compositions for assemblages (2) are the starting points for the calculations in both assemblages (1) and (3). Mineral K_D's are assumed to be constant and equal to those measured in assemblage (2). The data collected from various examples of assemblages (1) and (3) are also plotted in Figure C-13. These data points were calculated by solving for the variable $(\partial\mu H_2O/\partial X^{Chl}_{Fe})_{P,T}$ and multiplying by the observed ΔX^{Chl}_{Fe} relative to that in assemblage (2). Chlorite was used to independently evaluate the accuracy of the phase diagram which was calculated using $(\partial\mu H_2O/\partial X^{Bt}_{Fe})_{P,T}$. The agreement between calculated and measured values is very good and confirms the conclusions concerning mineral composition and chemical potential of water.

Discussion and conclusions.

The data illustrated in Figures C-12 and C-13 clearly demonstrate that interbedded amphibolite grade pelitic metasediments preserve differences in the chemical potential of water. Furthermore, the algebraic analysis presented in Figure C-13 suggests that the magnitude of the differences is on the order of 100 cal. These observations have implications concerning both the behavior of a volatile phase during the amphibolite grade metamorphism and the ultimate processes that led to the formation of these differences.

The first and most obvious conclusion is that these interbedded pelites have not equilibrated with an external reservoir of fluid of fixed composition. In contrast, the data suggest that various beds have equilibrated with fluids of variable composition. This is inconsistent with massive infiltration of externally derived fluid. Such an infiltration process would most certainly eliminate such small preserved differences. This is due to the fact that the buffer capacity of trivariant assemblages is relatively small. These types of assemblages <u>respond</u> to changes in intensive variables; they do not effectively <u>resist</u> changes as do divariant assemblages.

Several other processes, however, could be responsible for the observed differences in μH_2O. These include (1) initial compositional variations, (2) small amounts of infiltration varying from bed to bed, and (3) small amounts of infiltration of differing fluid composition. The third possibility can probably be ruled out because it is geologically unrealistic to have several different fluid compositions selectively moving through various beds. The criteria for differentiating between the first two possibilities are not straightforward, as the data are consistent with both processes. Thus, either small amounts of differing infiltration of externally derived fluids or initial differences in bulk composition are most likely responsible for the observed differences in the chemical potential of water on this single outcrop.

BEAVER BROOK FOSSIL LOCALITY

Access

Drive along New Hampshire rt. 112 to a point approximately 0.75 km W. of Lost River. Here there are signs marking the Appalachian Trail. Park cars under the trees, on the S. side of the road, near Appalachian Trail signs. The Beaver Brook fossil locality may

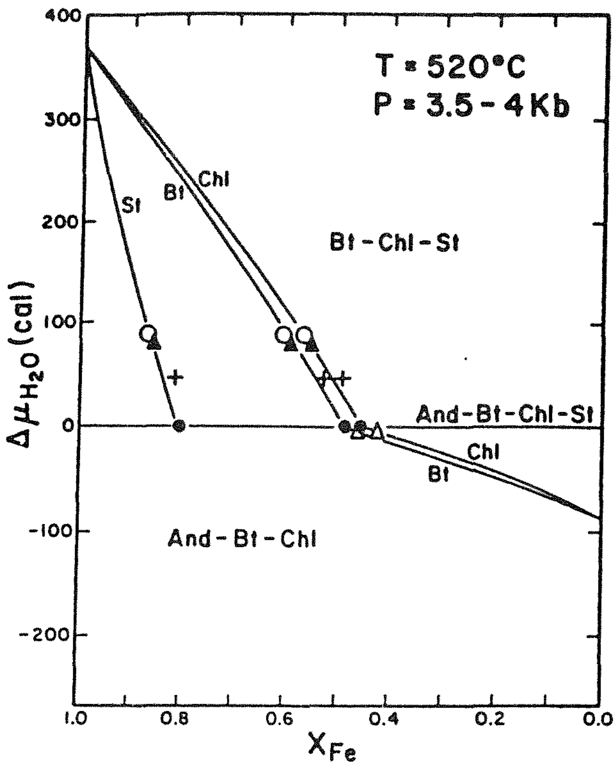


Fig. 13. Calculated μ H_2O vs. X_{Fe} diagram for Wildwood assemblages. Pairs and triples of identical symbols depict co-existing minerals.

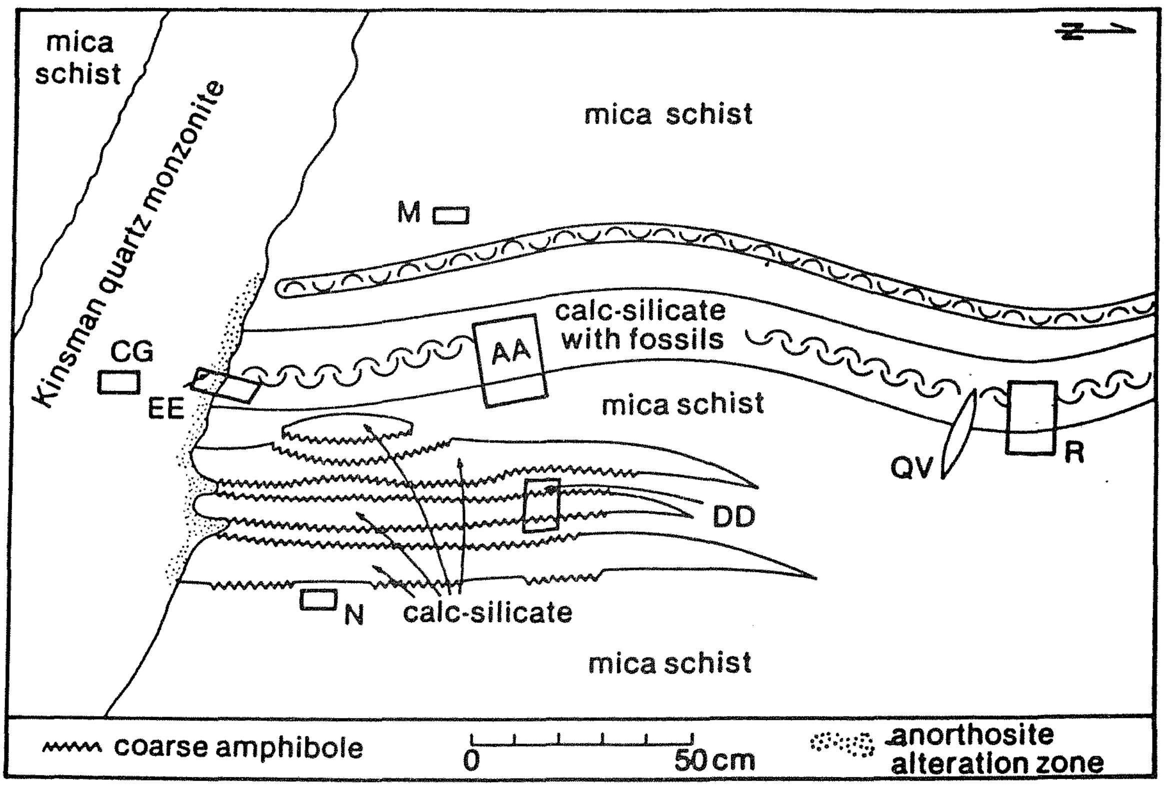


Fig. 14. Outcrop map of Beaver Bk. fossil locality. Sample locations given by letter symbols.

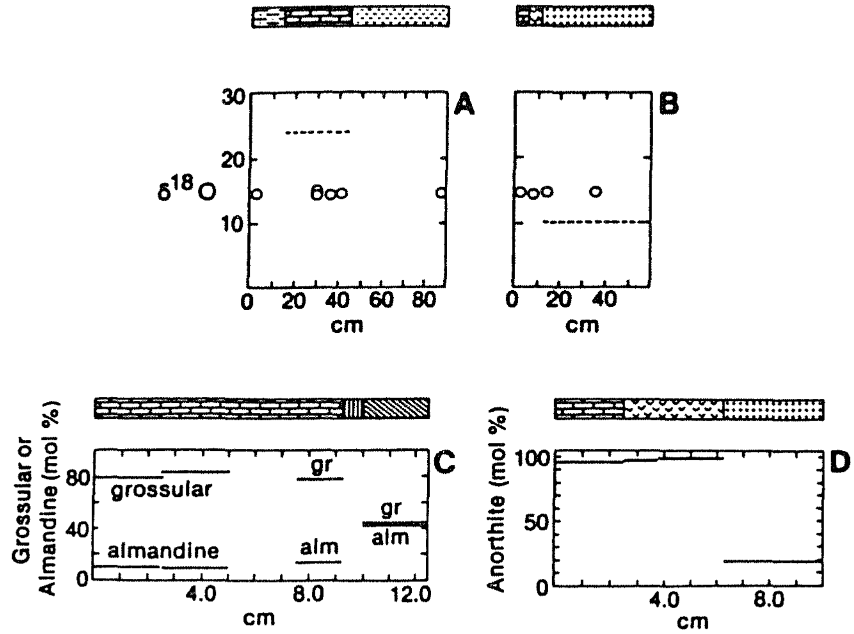


Fig. 15. (A) shows $\delta^{18}O$ values vs. distance perpendicular to bedding for traverses AA and R, combined, and samples M and N. Circles show measured data. Dashed line gives inferred pre-metamorphic isotopic composition. Dashed pattern denotes mica schist, brick work is calc-silicate. (B) is a plot of ^{18}O vs. distance perpendicular to dike/country rock contact for traverse EE and CG. The "V" pattern is anorthite replacement zone, the "+" pattern is Kinsman Quartz Monzonite. (C) shows garnet composition vs. distance perpendicular to bedding for traverse AA. Vertical ruled pattern is non-garnetiferous calc-silicate, diagonal ruled pattern is amphibolite. (D) is a plot of plagioclase composition vs. distance perpendicular to the dike/country rock contact for traverse EE and CG.

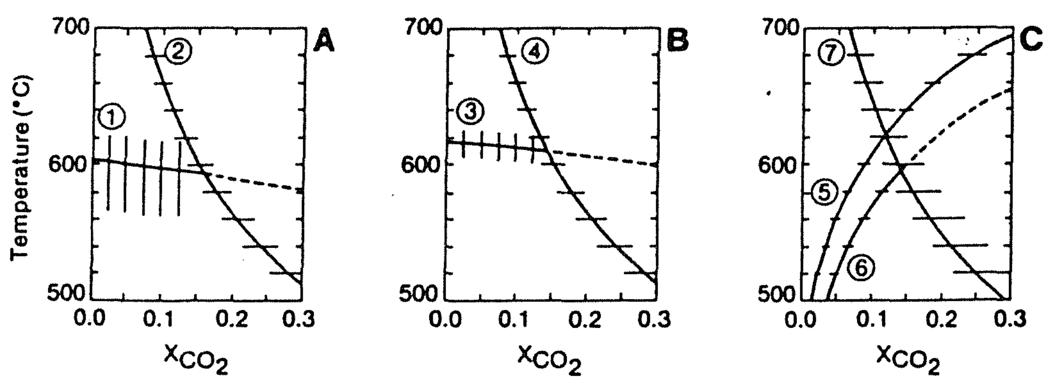


Fig. 16. Thermobarometry on T- XCO_2 diagrams at 3.5 kbars. Short horizontal and vertical bars give uncertainties in location of equilibrium curves for a range of 2 about average values of equilibrium constants. In (A) equilibrium 1 is Amp + Czo + 2Q = 5 Cpx + 9 Pl + $4H_2O$, (samples AA, R, DD, and EE) and 2 is Czo + CO_2 = 3 Pl + Cc + H_2O. (B) shows equilibrium 3, 4 Czo + Q = 5 Pl + Gt + 2 H_2O (sample AA) and equilibrium 4, Czo + CO_2 = 3 Pl + Cc + H_2O. In (C) equilibrium 5 is Cc + Q = Wo + CO_2 (sample R) equilibrium 6 is 2 Czo + 5 Cc + 3Q = 3Gt + 5 CO_2 + H_2O. Dashed lines are metastable extensions. (Amp = amphibole, Czo = clinozoisite, Q = quartz, Cpx = clinopyroxene, Pl = plagioclase, Cc = calcite, Gt = garnet, Wo = wollastonite).

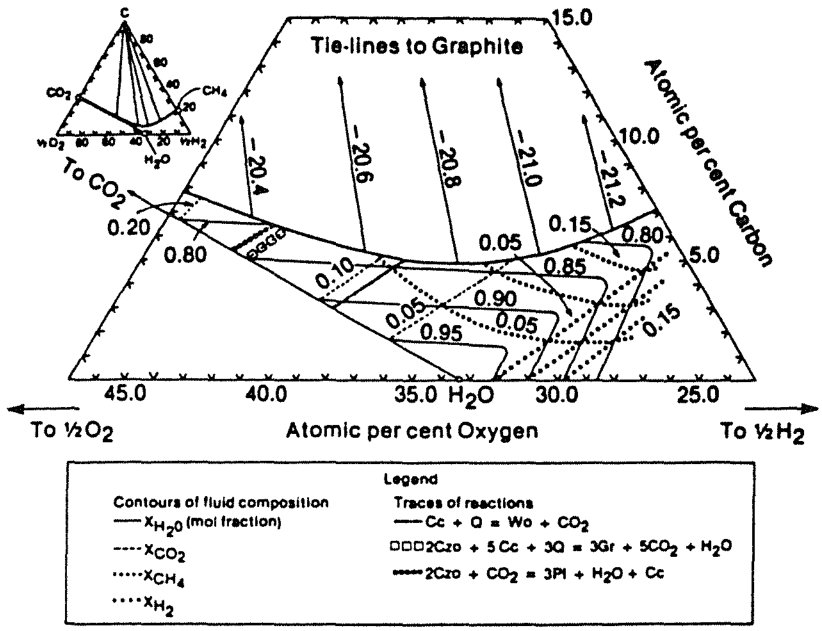


Fig. 17. Phase diagram of system C-O-H at 600°C, 3.5 kbars. Small triangle shows entire C-O-H diagram. Heavy line is graphite saturation curve and straight lines from it to graphite corner are two-phase lines. Large trapezoidal diagram is enlargement of H_2O-rich region. Contours of fluid composition drawn in one-phase fluid field. Tie-lines labeled with log fo_2 shown in two-phase, graphite-fluid region. Traces of three decarbonation of combined decarbonation-dehydration reactions are given. Dehydration reactions lie parallel to contours of XH_2O.

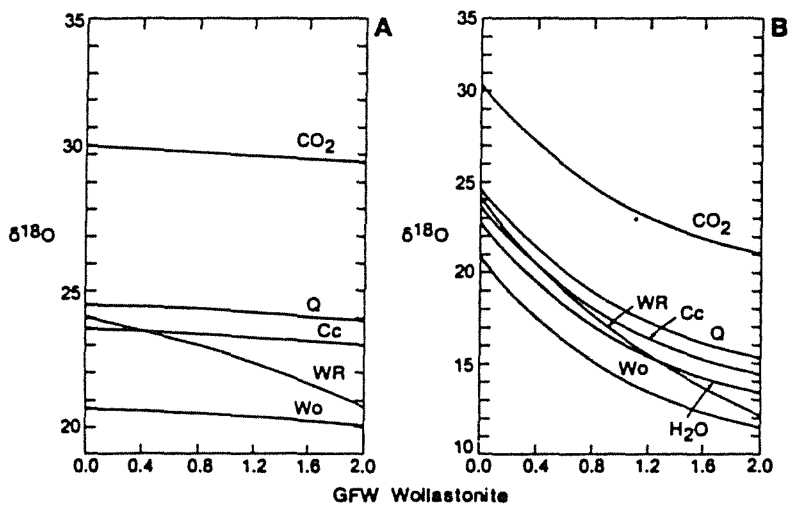


Fig. 18. Plots of $\delta^{18}O$ vs. number of gram formula weight (gfw) units of wollastonite formed in the reaction Cc + Q = Wo + CO_2. (A) shows expected changes in $\delta^{18}O$ as function of reaction progress when Rayleigh distillation is operative. Fluid evolved is pure CO_2. (B) gives expected changes in $\delta^{18}O$ in a rock infiltrated by H_2O with a $\delta^{18}O$ value of 12.5°/oo. Initial whole rock $\delta^{18}O$ = 24°/oo.

be reached by following the white-blazed, Appalachian trail S., across the floor of Kinsman Notch, and climbing about 0.5 km up the lower slopes of Mt. Moosilauke.

Hiking Route

WARNING: CASCADES OF BEAVER BROOK ARE VERY SLIPPERY. EXERCISE EXTREME CAUTION.

Hike S. on the trail (white blazes) for about 0.5 km to the first cascades of Beaver Brook. At your feet are large feldspar phenocrysts (up to 10 cm) in the Kinsman Quartz Monzonite. Visible in the cascades are mica schists of the Littleton Formation cross-cut by aplite dikes. A knife-sharp contact between Kinsman and Littleton may be found at the base of the ledge, down in the Brook **(CAUTION: SLIPPERY WHEN WET!).**

Continue following the white-blazed trail climbing alongside the cascades. Do not attempt to climb straight up the cascades because the rocks are very slippery. Over the course of the next 0.5 km you will notice that the Littleton mica schists are not nearly so aluminous as those at Wildwood roadcut. There are some muscovite-garnet schists but most rocks lack muscovite, have biotite with intermediate Fe/(Fe + Mg), and contain calcic-plagioclase. In at least one case (sample 70-184-N) there are two coexisting plagioclases: An 83 and An 67. The fossiliferous calc-silicate is usually interbedded with calcic mica schists rather than with aluminous pelites such as those at Wildwood roadcut.

There are no obvious landmarks to help locate the fossil locality. It will be marked for the IMA trip so you can't miss it. We will have to cross the Brook, then climb a gently sloping ledge to the base of a cascade. **USE EXTREME CAUTION: ROCKS ARE VERY SLIPPERY.**

Fluid-rock behavior at Beaver Brook was dramatically different from that at Black Mountain or Wildwood roadcut. You will be able to see two different types of evidence for fluid infiltration with your own naked eyes. Still other evidence for infiltration is given in the accompanying figures and will be discussed on the outcrop. Take a moment for orientation and note the color contrasts between pink-orange and green banded calc-silicates, dark gray mica schists, and pale gray-dirty white dikes of aplite and quartz monzonite. Thermobarometry for these rocks gives metamorphic conditions of 3.5 kbars and 600°C (Fig. C-14).

There are two pieces of evidence for infiltration that may be seen in the ledge before you. These are (1) a sharp, infiltration metasomatic front of anorthite in the margin of the quartz monzonite dike, on the left side fo the ledge; and (2) fossil brachiopods replaced by wollastonite, on the right side.

A close inspection of the quartz monzonite dike shows a creamy, translucent zone (2-4 cm thick) in the dike margin where it truncates calc-silicate beds. The zone is absent where the dike crosscuts mica schist. Microscopy shows that the marginal zone consists of myrmekitic anorthite (An 95-98) - quartz pseudomorphs after the quartz monzonite's primary oligoclase and microline. A microprobe traverse across the replacement front reveals a change in plagioclase composition from An 95 to An 19 over a distance of 1-2 mm. The anorthite zone has features diagnostic of infiltration metasomatism (as discussed by Korzhinskii and Hofmann) including (1) one-way mass transfer (i.e. Ca from calc-silicate to dike rock) and (2) a sharp replacement from separating mineral assemblages that are not in chemical equilibrium. The anorthite zone documents a component of the direction of fluid flow. Calcic aqueous solutions flowed from calc-silicate wall rocks into the quartz monzonite dike.

Fossil brachiopods replaced by wollastonite are located on the right side of the ledge. The original discovery of fossils (by C.V. Guidotti) was made on an exposed portion of the ledge where weathering had etched out their distinctive forms. Weathered material has since been quarried away for examination by A.J. Boucot who identified four genera of Early Devonian brachiopods including Acrospirifer, Leptocoelia, Atrypa, and either Leptostrophia or Protoleptostrophia. The present outcrop surface has not as yet weathered very deeply. You should be able to see, however, the characteristic cross-section of brachiopod shells, e.g. a shape like a flattened parenthesis.

The presence of the assemblage wollastonite-quartz-calcite in the fossil shells is evidence of fluid infiltration during metamorphism. Fluid/rock ratios may be calculated with the model of J.M. Ferry. Consider the decarbonation reaction Quartz + Calcite = Wollastonite + CO_2. At the conditios of metamorphism at Beaver Brook, 3.5 kbars and 600°C, the equilibrium XCO_2 for this reaction is 0.09. But, the reaction produces pure CO_2. The decarconation reaction would saturate a rock of normal porosity with CO_2 after only a miniscule amount of reaction. Thus, reaction would cease after formation of an undetectable amount of wollastonite. We are faced with a problem: How does one make wollastonite in detectable amounts at temperatures well below its stability limit in pure CO_2? The answer is to infiltrate the rock with a fluid whose XCO_2 is less than the equilibrium value. The infiltrating fluid sweeps away product CO_2 and makes it possible for decarbonation reactions to operate until reactants are exhausted. Calculated fluid/rock ratios at Beaver Brook range from 1.5 to 5.0, with a value of 4.0 (by volume) in the wollastonite bed.

Additional evidence of infiltration include possible hydrothermal graphite in the quartz monzonite dike and changes in $\delta^{18}O$ inferred to have occurred during metamorphism. These features cannot be seen in outcrop but will be discussed with the aid of the accompanying figures (Figs. C-15 through C-18).

Why <u>did</u> Beaver Brook experience infiltration during metamorphism while Black Mountain and Wildwood did not?

References

Albee, A.L., and Ray, L., 1970, Correction factors for electron probe microanalysis of silicates, oxides, carbonates, phosphates, and sulphates. Anal. Chem., 42, 1408-1414.

Billings, M.P., 1937, Regional metamorphism of the Littleton-Moosilauke Area, New Hampshire. Bull. Geol. Soc. Amer., 48, 463-566.

Dickenson, M.P., 1984, Local and regional differences in the chemical potential of water in amphibolite grade pelitic rocks, GSA abstr. with prog., 16, 488.

Ferry, J.M., 1979, A map of chemical potential differences within an outcrop, Am. Mineral., 64, 966-985.

Hodges, K.V., and Spear, F.S., 1982, Geothermometry, geobarometry and the Al2SiO5 triple point at Mt. Moosilauke, NH., Am. Mineral., 67, 1118-1134.

Rumble, D.III, 1973, Fe-Ti oxide minerals from regionally metamorphosed quartzites of western New Hampshire. Contr. Mineral. Petrol., 42, 181-195.

Rumble, Douglas III, 1974, Gibbs phase rule and its application in geochemistry, Jour. Washington Acad. Sci., 64, 199-208.

Rumble, Douglas III, 1976a, The use of mineral solid solutions to measure chemical potential gradients in rocks, Am. Mineral., 61, 1167-1174.

Rumble, Douglas III, 1976b, Oxide minerals in metamorphic rocks, Reviews in Mineralogy, v. 2, MSA, Washington D.C.

Rumble, D. III, 1978, Mineralogy, petrology, and oxygen isotopic geochemistry of the Clough Formation, Black Mountain, western New Hampshire, USA, Jour. Petrol., 19, 317-340.

Rumble, D. III, Ferry, J.M., Hoering, T.C., and Boucott, A.J., 1982, Fluid flow during regional metamorphism of the Beaver Brook fossil locality, New Hampshire. Am. Jour. Sci., 282, 886-919.

FLUID-ROCK INTERACTION AT MID-CRUSTAL DEPTHS, CENTRAL NEW HAMPSHIRE

C. Page Chamberlain

and

Douglas Rumble III

Introduction

The purpose of the field trip to central New Hampshire is to examine the role that fluids have played during Acadian regional metamorphism. We will visit two localities. The first area, located at Bristol New Hampshire, was infiltrated by hot metamorphic fluids during the later stages of the Acadian orogeny. The second area, located in Nelson New Hampshire, was infiltrated by cooler fluids during the earliest stage of the Acadian orogeny. Both localities that we will visit are sites where graphite was mined in the mid 1800's.

The field trip stops are both located in the high-grade terrane of the Kearsarge Central Maine Synclinorium (hereafter abbreviated KCMS). Metasediments in this terrane belong to a thick sequence of Silurian and Devonian pelites, quartzites, and calc-silicates (Hatch et al., 1983). These metasediments have been folded three times. The first stage of foldings produced west vergent nappe structures west of the axis of the KCMS and east vergent nappes east of the axis of the KCMS (Eusden, 1988). Later deformation produced open folds with northeast trending axes and isoclinal folds with northwest trending axes (Lyons et al., 1982).

Three types of Acadian plutonic rocks are found within the synclinorium: the Kinsman Quartz Monzonite, the Spaulding Series, and the Concord Granite (Lyons et al., 1982). The Kinsman Series are early Acadian, syntectonic, sheet-like plutons. The largest of these is the Cardigan pluton that runs throughout the central portion of Figure 1. The Spaulding Series are late Acadian, syntectonic plutons. The Concord plutons are two-mica granites that were intruded after deformation.

The metasediments and the Kinsman plutons have been metamorphosed to sillimanite-muscovite up to sillimanite-alkali feldspar-cordierite zone. Isograd and isotherm patterns in this terrane show an unusual metamorphic pattern consisting of a broad metamorphic high of sillimanite-muscovite zone rocks (T=550-600 C) that is puncuated by a series of metamorphic hot spots. In these hot spots pelitic rocks were metamorphosed to sillimanite-alkali feldspar-cordierite zone (T=650-700 C) (Chamberlain and Lyons, 1983). Measured pressures both across the broad metamorphic high and within the hot spots are 4 kilobars (Day and Chamberlain, 1988). All of this metamorphism is Acadian (400-380 myr) in age (Eusden and Barreiro, 1988).

The cause of the long wavelength thermal anomaly in the synclinorium is a matter of considerable debate. However, recent calculations show that the thermal and baric pattern might be due to tectonic thickening of a basin that contained sediments with an unusually high amount of heat producing elements (Chamberlain and Sonder, in prep.). Measurements of Th, U, and K show that the Rangeley and Littleton schists within the synclinorium are remarkably enriched in these elements relative to metasediments in adjacent terranes (Chamberlain and Sonder, in prep.).

The 'hot spots' are not caused by simple tectonic thickening, however. Chamberlain and Rumble (1988 and 1989) have suggested that the hot spots formed by the focusing of hot metamorphic fluids through fracture systems within quartzite layers. Evidence for infiltration are: 1) a network of quartz-graphite veins in the hot spots; and 2) an oxygen isotope alteration halo that is coincident with the isograd-isotherm high.

There is also evidence for infiltration of cool fluids early in the metamorphic history of the KCMS. For example, in Nelson New Hampshire infiltration of fluids soon after nappe emplacement caused the nappe sheets to undergo retrograde metamorphism. During retrograde metamorphism graphite was deposited within the fluid conduits (Chamberlain and Rumble, 1986).

Our observation that the KCMS was first infiltrated by fluids that were cooler than the nappes and were later infiltrated by fluids that were hotter than the nappes is consistent with results from thermal models of the New England Appalachians (Chamberlain and England, 1985). According to our calculations, the earliest fluids were derived from dewatering of structruallly lower and cooler nappe sheets. These fluids rose upward through the stack of nappe sheets and caused local retrogression. This scenario assumes that the nappes caused a large-scale thermal inversion during deformation. Later during the Acadian orogeny, the structurally lower nappes evolved fluids that were hotter than the overlying nappe sheets. Focusing of these hot fluids could have caused the local thermal perturbations that formed the hot spots (Brady, 1988).

On this trip we will visit two separate localities where we believe that early cool fluids infiltrated the nappes (Nelson, NH) and where later hot fluids infiltrated the nappes (Bristol, NH) (Figure 1).

Description of Localities

Drive from Black Mountain to Bristol, New Hampshire. This is done by returning to East Haverhill and taking Route 25 south to west Plymoth (~25 miles). Here the road bifurcates. Follow the right fork, which is 3A south. Route 3A will bring you along the east edge of Newfound Lake. Follow Rt. 3A south to Bristol (~13 miles). At Bristol turn east on Route 104. Take 104 ~1.5 miles east to large outcrops on the south side of the highway. Park on the shoulder. Drives might want to turn the vehicles around after letting off passengers. Approximate driving time from Black Mountain to Bristol Stop A is 50 minutes.

Bristol

Stop A:

The highly graphitic, Rangeley schists are exposed in two road cuts. Rocks at this location are on the east edge of the Bristol hot spot (Figure 2). They contain the assemblage: sillimanite-cordierite-alkali feldspar-plagioclase-quartz-graphite. The δ13C of graphite in these rocks is ~24 per mil (Figure 3). The carbon isotope values of graphite from the adjacent quartzites of the Perry Mountain suggest that the carbon in the lower part

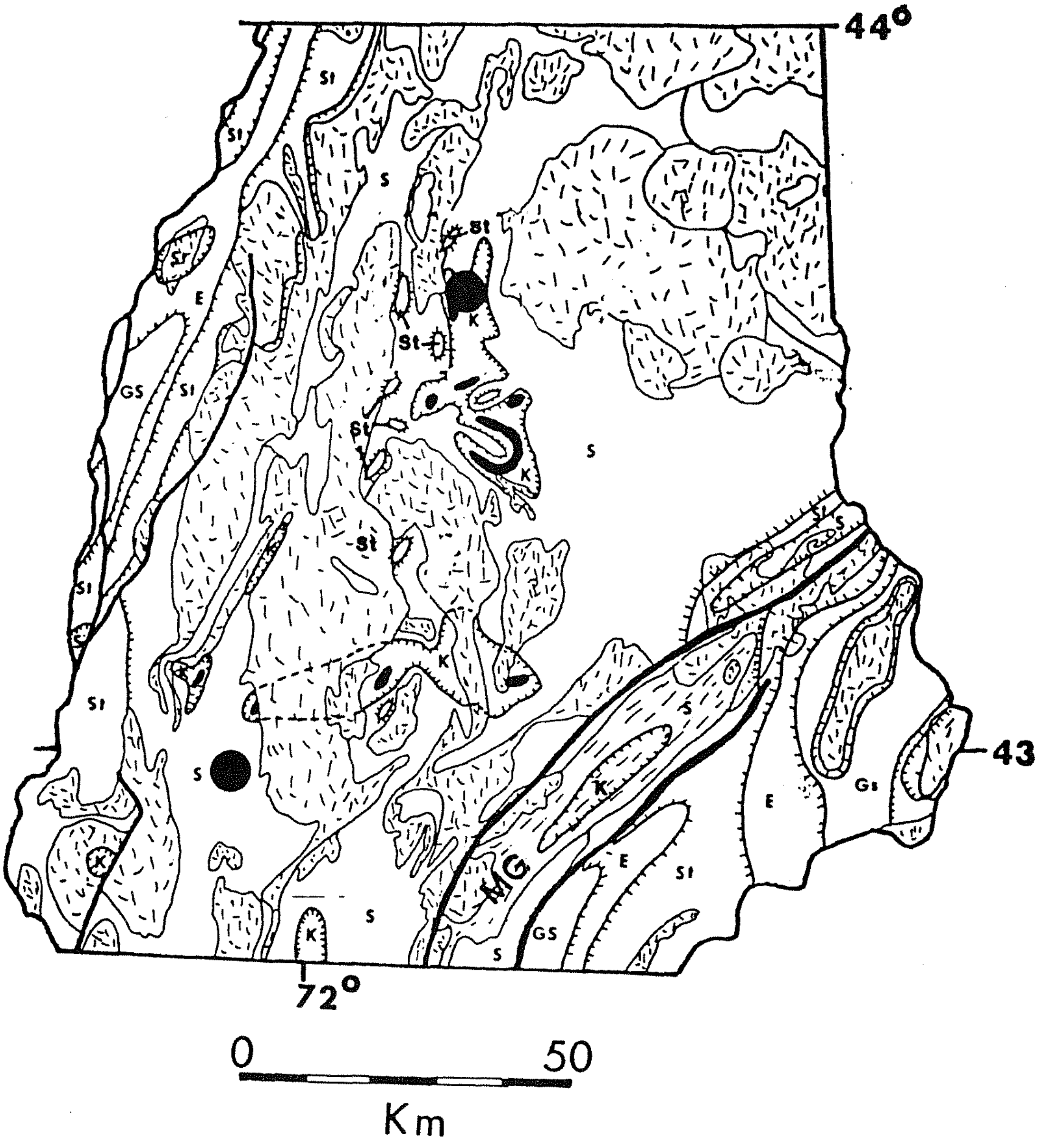


Figure 1: Metamorphic map of the Kearsarge Central Maine Synclinorium. The two black dots show the two localities we will visit, The northern dot is the Bristol Hot Spot. The southern dot is the Nelson field trip stop. Isograds are shown as the hatchured lines. GS: garnet grade; St: staurolite grade; S: sillimanite-muscovite grade; K: alkali feldspar-sillimanite zone; black areas: cordierite-alkali feldspar zone. Dashed areas are the plutonic rocks.

of this unit was derived from carbon that was mobilized during devolatilization of the Rangeley schists and infiltated the quartzite.

The oxygen istope values of quartz in these rocks show that they are depleted in δ18O by about 1 per mil relative to adjacent rocks outside the hot spot (Figure 4).

Stop B: To drive to Stop B head west on Route 104 about 0.45 miles. Take the first right and then immediately take the dirt road that forks off to the left. This road will lead you up through a new development. Drive about 0.4 miles and pull over when the road turns 90 degrees to the right. You should be parked at the base of a powerline. Park the cars and proceed up the powerline. Approxiamte driving time 10 minutes.

The powerline cut exposes a section from the base of the Perry Mountain Formation where it is in contact with the Rangeley schists to the top of the Perry Mountain where it is in contact with the Madrid calc-silicates. The Perry Mountain is an interbedded quartzite-schist unit. It lies in the heart of the hot spot and contains the graphite veins that were mined in the mid 1800's.

Examination of outcrops along the powerline shows that there are at least two generation of veins: an early set of veins that are concordant to bedding and foliation and a later set of veins that are discordant to bedding. The graphite is generally concentrated in the early veins. The vein density decreases toward the contact with the overlying Madrid Formation.

We suggest that graphite was deposited when fluids derived from the underlying pelites ($CH_4>CO_2$) mixed with fluids derived from the overlying calc-silicates ($CO_2>CH_4$) within the fracture system in the Perry Mountain. Fluid mixing is evidenced by gradients in the chemical potential of water and the δ13C graphite across the powerline outcrops. The chemical potential of water decreases and the δ13C graphite increases from the contact of the Perry Mountain and Rangley to the contact of the Perry Mountain with the Madrid calc-silcates (Chamberlain and Karabinos, 1989).

Stop C: Head back to Route 104 and turn west on 104. Continue through Bristol on 104 about 1.5 miles to the first outcrops on the north side of the road. The drivers may want to turn the cars around while the passengers look at the rocks. Approximate driving time 20 minutes.

These road cuts are made up of the graphitic, calc-silicates belonging to the Madrid Formation. The assemblage in these rocks consist of diopside-actinolite±calcite±zoisite-plagioclase-quartz-graphite. The δ13C graphite in these rocks ranges from -22 to -14 per mil. It is our opinion that the carbon in the upper part of the Perry Mountain in the hot spot was derived from carbon mobilized from these calc-silicates during prograde metamorphism.

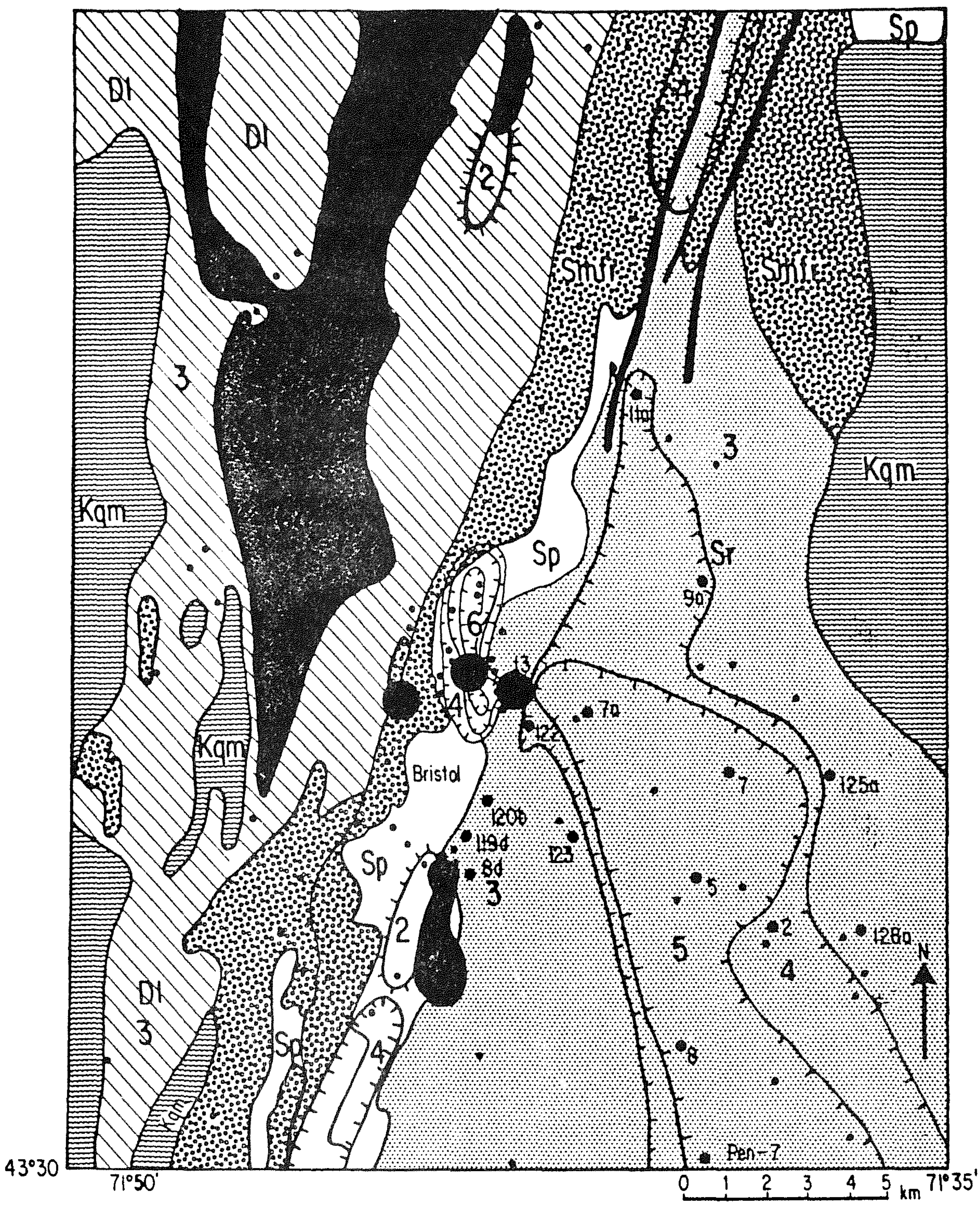


Figure 2: Geologic map of the Bristol area. Isograds are shown as hatchured lines. Zone 6: is cordierite-alkali feldspar; 5 is alkali feldspar-sillimanite; 4 is alkali feldspar-muscovite-sillimanite; 3 is sillimanite-muscovite; and 2 is sillimanite-staurolite. Co is the Concord granite; Kqm is the Kinsman; Dl is the Littleton Fm.; Smfr is the Madrid-Francestown Fm., Sp is the Perry Mountain Fm.; Sr is the Rangeley Fm. Heavy lines are faults. Heavy dots show location of the field trip stops.

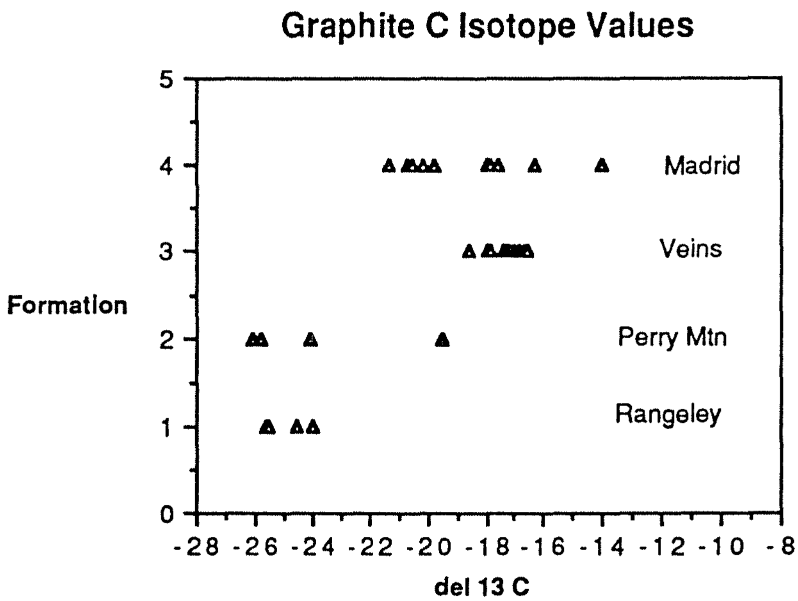


Figure 3: Plot of carbon isotope values for graphite in the Rangeley (Sr), Madrid (Smfr) and Perry Mountain (Sp). Values in per mil relative to PDB.

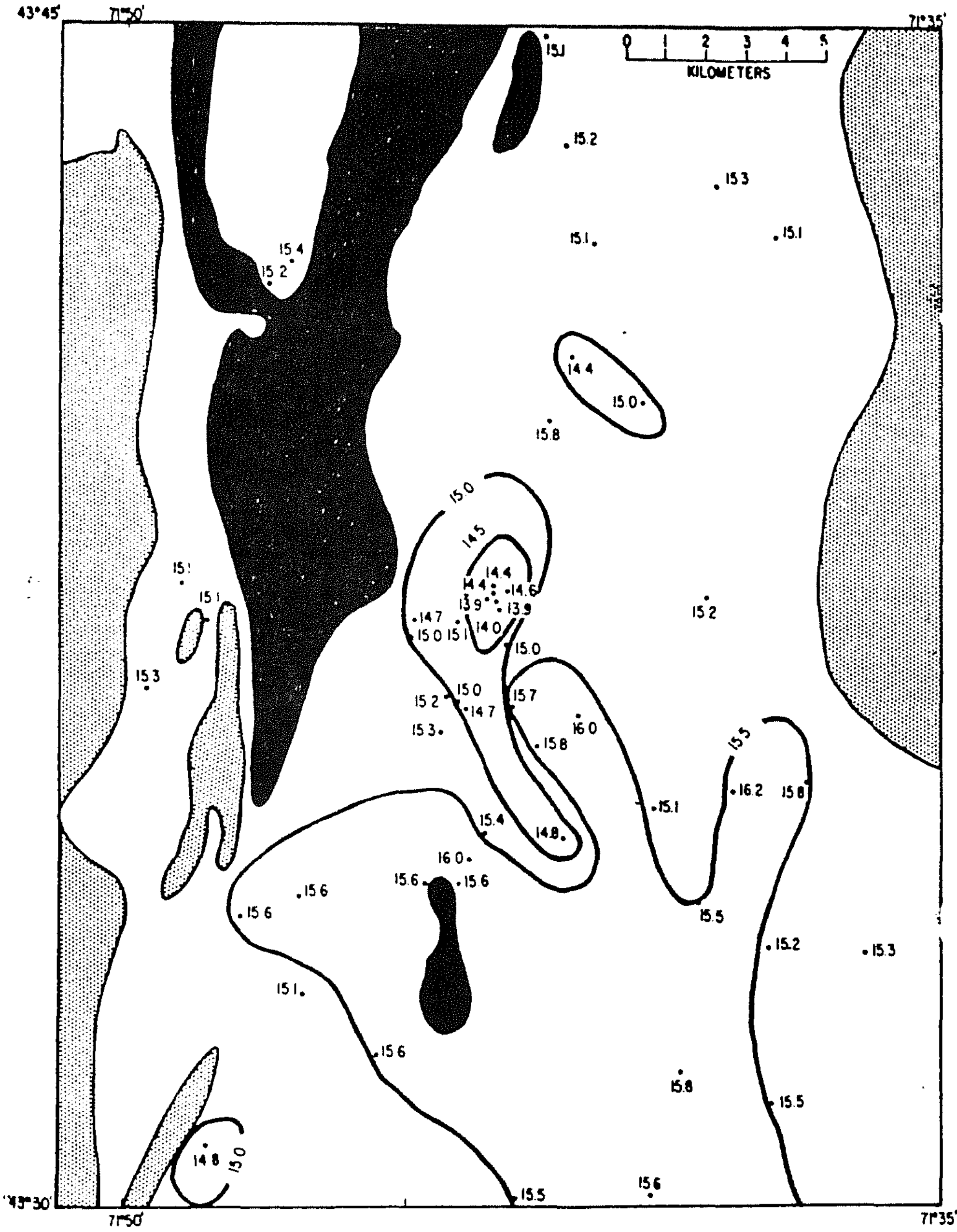


Figure 4: Map contoured for equal values of δ18O quartz (from Chamberlain and Rumble, 1988)

Nelson

To get to Nelson drive east on Rt. 104 through Bristol to Rt. 93. Take 93 south to Concord. At Concord take 89 north to Rt. 9 south. Take 9 south to Granite Lake. Immediately past Granite Lake across the village store turn south. This is the road to Nelson. Continue about two miles to the village green of Nelson. Do not continue to the right, downhill on the paved road. Proceed east, passing the twn barn, then curve to the northeast on the gravel road. Park cars along the road 1 1/2 miles from the Green. The Osgood Mine is a few hundred yards to the SE, and the end of a well overgrown haulage road. Wait for the trip leaders to guide you in to the mine. Approximate driving time is 1 1/2 hours.

The Osgood Mine in Nelson, NH is said to have furnished as much as 30 tons of plmbago in three months of mining. The presently exposed workings are the largest known in New Hampshire. Graphite ore occurs in a calc-silicate bearing, pelitic unit of the Rangeley Formation. The host rock is sulfidic and poorly bedded, consequently it is difficult to observe structures in these outcrops.

The top of the graphite ore body shown in Figure 5 is an undulating surface that strikes N-S and dips 40-45 degrees westward. The ore is truncated along its northern boundary by a fault that strikes n63 to 74 E and dips 45 to 49NW. The ore body is vriable in thickness, but averages 10 cm. The ore consists of graphite and ubiquitous hercynite, with traces of ilmenite and pyrrhotite-chalcopyrite intergrowths.

Host rocks are rusty weathered mica schists with calc-silicate nodules. The primary assemblage of pelites is cordierite-andalusite-biotite-garnet. Porhyroblasts of cordierite are replaced by chlorite and radiating sprays of muscovite. Euhedral prisms of staurolite and needles of sillimanite occur in with chlorite. There are also rare needles of kyanite.

Calc-silicates occur as nodules enclosed in the mica schist. Primary assemblage of calc-silicates include quartz-plagioclase-diopside-calcite-actinolite. These peak metamorphic assemblages have been altered such that they contain zoisite and no calcite in the rocks adjacent the vein.

Graphite ore is always found to replace aluminous pelite. It has never been observed replacing calc-silicates. The most massive ore consists of a intergrowth of schistose graphite embedded with hercynite. Hercynite is an important participant in the replacement reaction, for it is not found outside the ore. Retrogradation of relict porphyroblasts is always present in the ore. The breakdown of cordierite, andalusite, and biotite to more hydrous phases occurs outside the ore as well, however.

It is not known why graphite mineralization was restricted to the pelite. Perhaps there is a contrast in mechanical properties between the schist and calc-silicate that lead to fracturing and infiltration of ore fluids.

Carbon isotope values of graphite ore range from -24.6 to 22.9 per mil (PDB) suggesting an affinity with sedimentary reduced organic matter.

The textures and mineral reactions seen in these rocks suggest that the graphite was deposited when cooler fluids infiltrated the rocks during metamorphism. The host rocks belong to the upper plate of nappe sheet that was thrust over cooler rocks of the

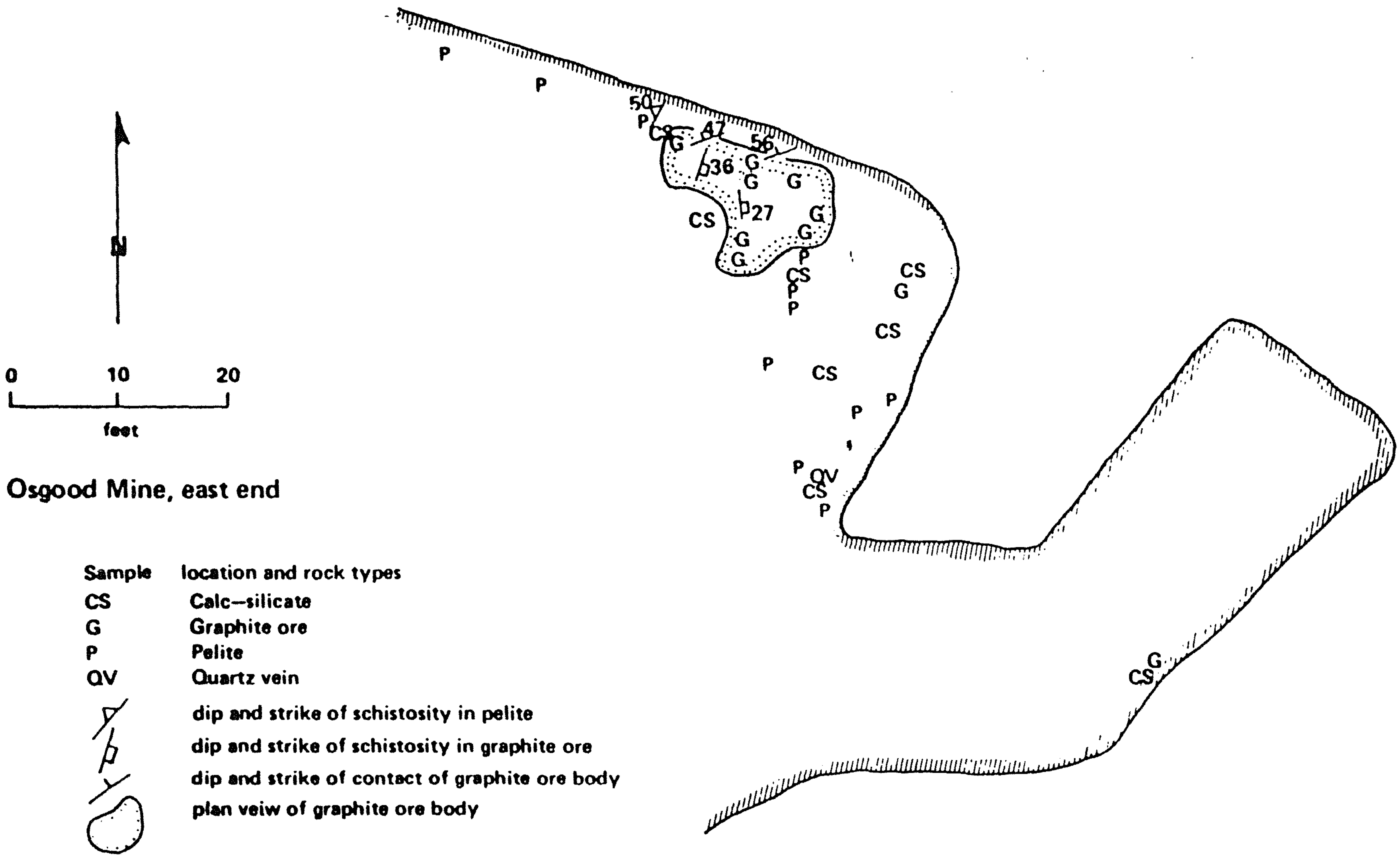


Figure 5: Geologic map of the Osgood Mine (from Rumble and Chamberlain, 1988)

Monadnock sequence (Chamberlain, 1986). Devolatization reactions in the lower plate are believed to have provided the fluids that deposited the graphite (Chamberlain and Rumble, 1986).

Keene

To get to Keene proceed back to Route 9 the way you came in to the Osgood Mine. Head south on Route 9 all the way to Keene. The Days Inn is located south of Keene in the shopping center. Approximate driving time is 30 minutes.

References

Brady, J. B., 1988, The role of volatiles in the thermal history of metamorphic teranes, J. Petrol., v. 29, 1201-1214.

Chamberlain, C.P., 1986, Evidence for the repeated folding of isotherms during regional metamorphism. J. Petrol., v. 27, 63-89.

Chamberlain, C.P., and England, P.C., 1985, The Acadian thermal history of the Merrimack synclinorium in New Hampshire. J. Geol., 93, 593-602.

Chamberlain, C.P., and Karabinos, P., 1989, Kilometer scale mixing of fluids at mid-crustal depths. Geol Soc. Amer. Abstr. with Prog., in press.

Chamberlain, C.P., and Lyons, J.B., 1983, Pressure, temperature, and m,etamorphic zonation studies of pelitic schists in the Merrimack synclinorium, south-central New Hampshire. Am. Mineral., v. 68, 530-540.

Chamberlain, C.P., and Rumble, D. III, 1988, Theraml anomalies in a regional metamorphic terrane: An isotopic study of the role of fluids. J. Petrol., v. 29, 1215-1232.

Chamberlain, C.P., amd Rumble, D. III, 1989, The influence of fluids on the thermal history of a metamorphic terrane: New Hampshire. J. Geol. Soc. London (Special Issue Dublin Mtg. on Metamorphism), in press.

Chamberlain, C.P., and Rumble, D. III, 1986, Fluid flow and graphite deposition during synmetamorphic thrusting. Amer. Geophys. Union Spring Mtg.

Day, H.W., and Chamberlain, C.P., 1989, Implications of the thermal and baric structure for controls on metamorphism northern New England, USA. J. Geol. Soc. London (Special Issue Dublin Mtg. on Metamorphism), in press.

Hatch, N.L., Moench, R.H., and Lyons, J.B., 1983, Silurian-Lower Devonian stratigraphy of eastern and south-central New Hampshire: Extensions from western Maine. Am. J. Sci., v. 283, 739-761.

Eusden, J.D., 1988, Stratigraphy, structure, and metamorphism across the "dorsal zone" central New Hampshire. NEIGC Field Trips to New England, 40-59.

Eusden, J.D., and Barreiro, B., 1988, The timing of peak high-grade metamorphism in central-eastern New England. Maritime Sediments and Atlantic Geology, v. 24, 241-255.

Lyons, J.B., Boudette, E.L., and Aleinikoff, J.N., 1982, The Avalonian and Gander Zones in central New England. In: St-Julien, P. and Beland, J. (eds.) Major Structural Zones and Faults of the Northern Appalachians. Geological Association of Canada Special Paper, 24, 44-66.

Rumble, D. III, and Chamberlain, C.P., 1988, Graphite veins of New Hampshire. NEIGC Field Trips to New England, 241-255.

BASEMENT AND COVER IN THE ACADIAN METAMORPHIC HIGH OF CENTRAL MASSACHUSETTS

by
Peter Robinson, R.J. Tracy, Kurt Hollocher, H.N. Berry, IV, and J.A. Thomson

INTRODUCTION

Acadian regional metamorphism in New England reaches its culmination in the granulite facies in central Massachusetts. Not only is the intensity of metamorphism higher than observed to the north in New Hampshire and Maine, but the relations of metamorphism to deformation seem to be different. Peak-metamorphic mineral assemblages and features due to partial melting are here overprinted by moderate to strong deformational fabrics not seen farther north. Again as in central New Hampshire, the relations between metamorphic zones and igneous intrusions are less clear than they are in northwestern Maine, although one or two examples of metamorphosed contact aureoles are known. However, the disposition of isograds seems to be much more regular than in central New Hampshire, hinting at a much more even distribution of thermal energy, with, however, the same type of puzzling problems as to the energy sources.

The purpose of this segment of the trip is to acquaint participants in one day with some of the most salient features of the central Massachusetts high as well as some recent research results in the region. Contrasts in the style of metamorphism and relations to deformation will be emphasized, as compared to metamorphic rocks seen on the first three days. The stops are designed for visits by a large group and should facilitate informative disscussion. The highlights and objectives of each stop are as follows:

Stop 1: Extensive exposures of Monson Gneiss, the typical "basement" of several gneiss domes in southern New England; evidence for the plutonic roots of a Late Ordovician magmatic arc with Acadian intrusions and partial melting.

Stop 2: Sulfidic schists and subordinate volcanics of the Late Ordovician Partridge Formation; Sillimanite-Orthoclase-Muscovite Zone (IV) in the Greenwich stratigraphic syncline studied in detail by Tracy (1975, 1978); examples of stromatic migmatites, better developed on more inaccessible islands of Quabbin Reservoir. In this region garnets show continuous Mg-depleted, Mn-enriched rims suggesting a continuous retrograde net-transfer reaction through the medium of an intergranular water-bearing fluid.

Stop 3: Typical metamorphosed tholeiitic volcanics the Partridge Formation, Greenwich syncline (Zone IV) with assemblages containing hornblende and gedrite or cummingtonite. Hornblende-cummingtonite amphibolite contains migmatitic patches with coarse cummingtonite.

Stop 4: "Big garnet syncline"; Sillimanite-Orthoclase-Garnet-Cordierite Zone (VI). Pavement exposure shows a network of feldspathic veins with large garnets believed to have formed by fluid-absent melting.

Stop 5: The extremely sulfidic schist of the Middle to Late Silurian Smalls Falls Formation. Mg end-member cordierite and biotite, H_2S-bearing cordierite. Opportunity to observe and discuss effects of granulite-facies metamorphism on shales deposited in an environment with extreme activity of S-reducing bacteria.

Stop 6: Normal pyrrhotite-garnet-cordierite schist of the granulite facies lacking obvious migmatitic features, but containing typical retrograde garnet rims produced by local ion exchange with cordierite and biotite, and lacking evidence of a pervasive metamorphic fluid.

Stop 7: Migmatitic garnet-cordierite gneisses and schists with a thin layer of mylonite containing new microporphyroblasts grown under a different facies at lower T and higher P(?).

Stop 8: Debris and outcrop along the pipeline trench. Observe and collect supremely fresh, coarse, migmatitic, highly deformed, garnet-sillimanite-cordierite gneiss typical of the granulite facies in southern Massachusetts. Evidence of a P-T path from metamorphosed cordierite pegmatites and andalusite pseudomorphs.

Stop 9: Granulite-facies metamorphism of intermediate to mafic igneous rocks. Augite-orthopyroxene-plagioclase assemblages produced by total breakdown of hornblende; orthopyroxene-orthoclase assemblages from biotite breakdown; orthopyroxene-bearing migmatitic veins.

ACKNOWLEDGEMENTS

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REGIONAL SETTING

The bedrock of the area consists of stratified and intrusive rocks of Late Precambrian, Ordovician, Silurian, and Devonian age, as well as Triassic-Jurassic strata in adjacent rift basins, and Jurassic and Cretaceous diabase dikes and sills. The main structural development of the region took place in the Ordovician Taconian Orogeny and the Devonian Acadian Orogeny. In terms of plate tectonics as presently understood, the area lay east (present geography) of the early Paleozoic ocean that closed during the Taconian Orogeny and east of the pertinent subduction zone. It also lay on the west side of a major Silurian-Lower Devonian marine sedimentary trough that closed at the beginning of the collisional tectonics of the Acadian Orogeny. Recently Osberg et al. (in press) have argued that this trough may have been located within a tectonically thinned portion of the Avalon continental plate that had been juxtaposed against North America during the Taconian. If this is true, then the collision of the Acadian only involved the closing of this trough and perhaps not the closing of an ocean. In conflict with this is recent paleomagnetic data from Newfoundland (Johnson et al., 1989) indicating that although the volcanic arc rocks were closely juxtaposed to a low latitude North America in the Ordovician, rocks of Avalon were still at high southern latitudes, leaving a major ocean yet to be closed in the Acadian.

Acadian tectonic development involved early west-directed fold nappes of regional extent that have recently been shown to be truncated by a later set of major west-directed thrust nappes (P.J. Thompson, 1985; Elbert, 1986; Robinson et al., 1986a; Robinson, 1987; Thompson, Elbert, and Robinson, 1987; Berry, 1987a, 1987b). The middle portion of Acadian tectonic development, loosely classed under backfolding, involved complex longitudinal transport of some gneiss bodies, eastward overturning of axial surfaces of fold nappes and of west-directed thrust nappes, and the formation of extensive mylonites with west-over-east movement sense superimposed on peak metamorphic fabrics (Robinson et al., 1977, 1982; Berry, 1987a, 1987b; Finklestein, 1987). The culminating deformation, with tight folding and overprinting of a north-south trending linear fabric on previous features, is temporally associated with the gravitational rise of gneiss domes in the Bronson Hill anticlinorium. The pattern of metamorphic isograds, as well as the reconstructed P-T trajectories of individual samples or groups of samples, bears a complex relationship to these deformations that needs to be better understood to develop a clearer and chronologically more precise tectonic and thermal history of the belt (Berry, 1987a; Elbert, 1987; Schumacher et al., 1987, 1988). In such work, intrusive rocks have played an important role because of their effects

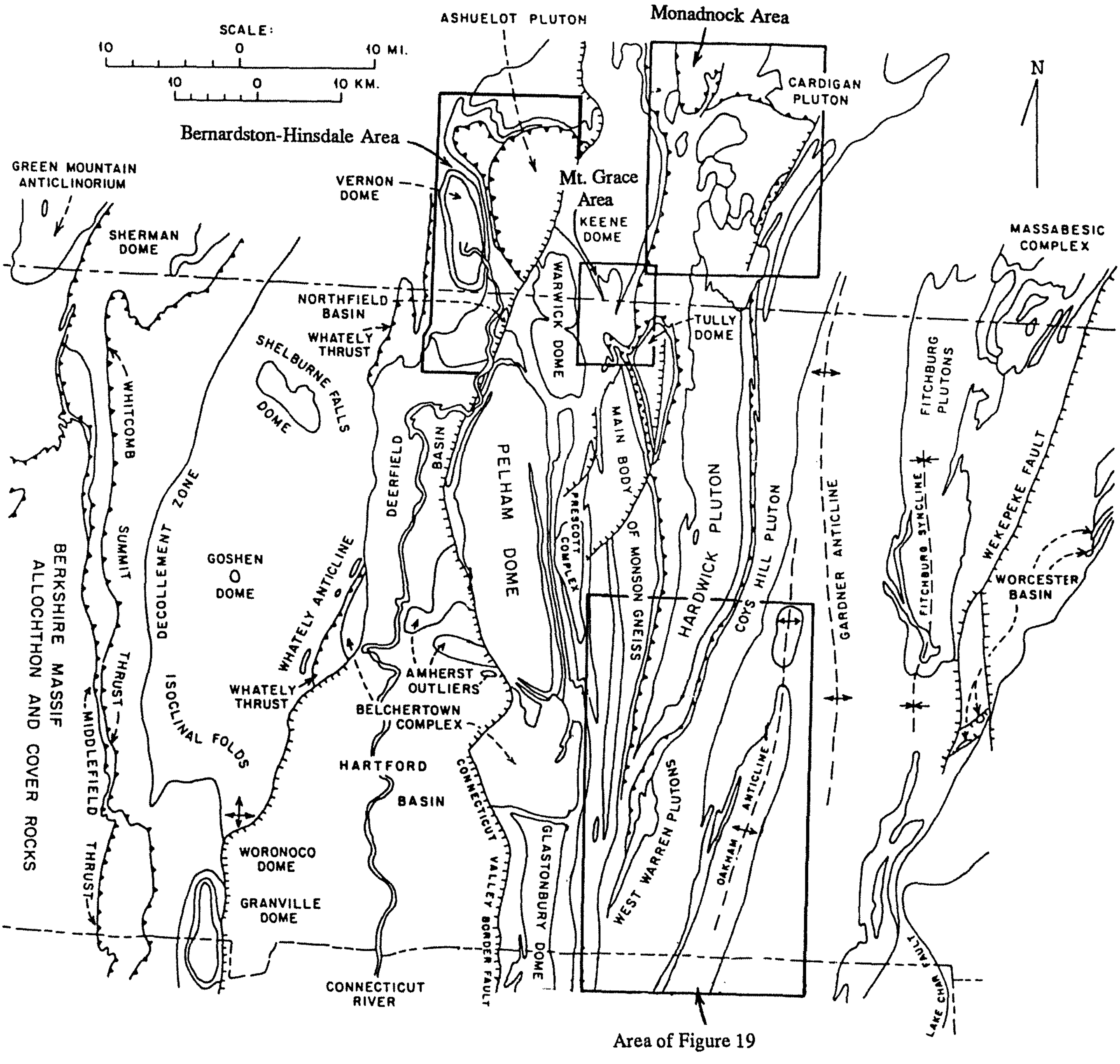

Figure 1. Geologic index map of central Massachusetts and adjacent states showing outlines of smaller areas discussed in text.

on patterns of metamorphism, because they have given crucial information about times of deformation, and because their chemistry can give some clues as to the global tectonic setting in which they were emplaced.

The reader seeking further background on the regional picture should examine the Bedrock Geologic Map of Massachusetts (Zen et al., 1983). Summary papers by Robinson, Thompson, and Rosenfeld (1979), Robinson and Hall (1980), Hall and Robinson (1982), as well as more detailed discussions on regional structural interpretation by Robinson (1979), Robinson, Field, and Tucker (1982), Robinson et al.(1988) are also available. Recent interpretations of the patterns of metamorphism and phase relations include Robinson, Tracy, Hollocher, and Dietsch (1982), Robinson (1983), and Robinson et al. (1986b). Stratigraphic and structural interpretations are changing rapidly. Rather large areas previously assigned to the Ordovician Partridge Formation are now reassigned to the Lower Silurian Rangeley Formation, and major westward thrusting has proved to be a much larger component of the early Acadian nappe phase than previously recognized.

STRATIGRAPHY

Introduction: The strata of the region may be divided into five major groups as follows: 1) Late Precambrian microcline gneisses, quartzites, schists and related rocks in the core of the Pelham gneiss dome that appear to have Avalon affinities; 2) Late Ordovician plagioclase gneisses and amphibolites structurally overlying the Late Precambrian gneisses in the Pelham dome and forming the cores of the other domes; 3) Late Ordovician Ammonoosuc Volcanics and Partridge Formation; 4) Silurian-Lower Devonian stratified rocks of the Connecticut Valley and Merrimack belts; and 5) Triassic-Jurassic sedimentary rocks and basalts of the Connecticut Valley Mesozoic basins. These are cut by a variety of pre-tectonic, syn-tectonic, and post-tectonic Silurian-Devonian intrusions ranging from gabbro to granite, and by Jurassic and Cretaceous diabase dikes. General stratigraphic relations and stratigraphic problems have been covered extensively elsewhere and only features related to the present field trip are summarized here. The origin of the Late Precambrian and Late Ordovician basement rocks in the gneiss domes, and their relations to the Late Ordovician cover rocks continues to be a problem (Robinson, 1981; Robinson et al., 1989).

Late Precambrian Rocks: The stratified gneisses in the core of the Pelham dome (Figure 1) are considered Late Precambrian on the basis of zircon ages (Naylor et al., 1973, Zartman and Naylor, 1984; Tucker et al., 1988) and lithic and geochemical similarities to rocks in southeastern Connecticut. Hodgkins (1983, 1985) has shown that the dominant felsic gneisses of this sequence,with low normative anorthite and extremely low MgO/(MgO+FeO) ratios, have a major and trace element chemistry consistent with interpretation as a sequence of chemically evolved alkali rhyolites that might have erupted in a rifting environment. The relics of an apparently pre-Devonian granulite-facies metamorphism that survived Acadian kyanite-muscovite overprinting (Robinson, Tracy and Ashwal, 1975; Roll, 1986, 1987) are still of unknown age, though tentatively assigned to the Late Precambrian (Robinson, 1983).

Monson and Related Gneisses: The most abundantly exposed rock unit in the Bronson Hill anticlinorium of central Massachusetts is the Monson Gneiss, most extensively exposed in the main body of Monson Gneiss (Emerson, 1917; Robinson, 1979; Zen et al., 1983). The Monson Gneiss forms the local "basement" for the overlying Ordovician- Devonian stratified sequence that includes the Ammonoosuc Volcanics at the base, the Partridge Formation, and younger units. Generally similar gneisses occur in the Fourmile Gneiss of the Pelham dome (Ashenden, 1973), between the Late Precambrian rocks in the core of the dome and the overlying stratified sequence, and in other domes in the Bronson Hill anticlinorium.

With the exception of very rare diopside calc-silicate and quartzite layers, the Monson Gneiss consists of rocks of broadly igneous composition (Robinson, 1963, 1967c; Hollocher, 1987,

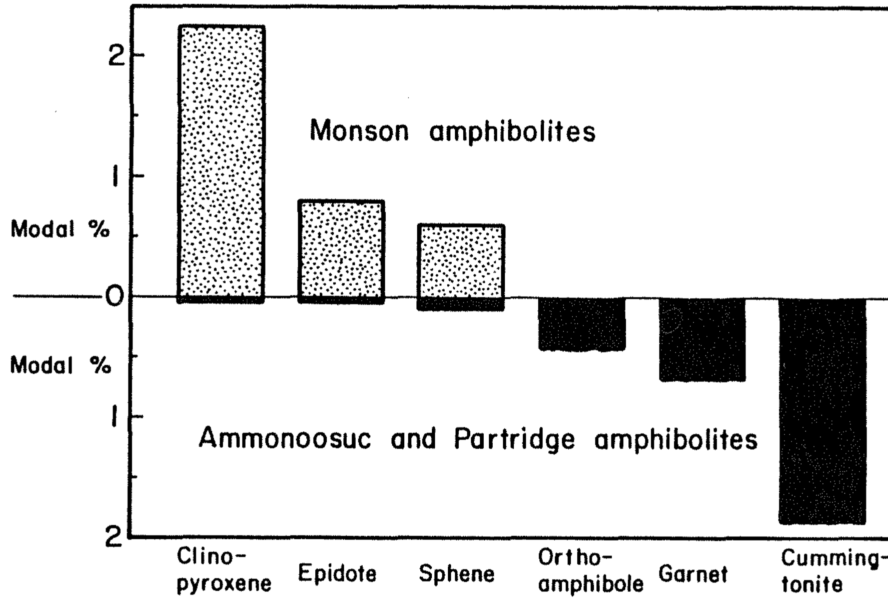

Figure 2. Histogram showing the average modal abundance of each of three Ca- rich minerals and three Fe-Mg minerals in 26 amphibolites from the Monson Gneiss and 26 amphibolites from the Ammonoosuc Volcanics and PartridgeFormation that have relatively unaltered chemical compositions. In these relatively unaltered rocks, the Ca-rich minerals are much more abundant in theMonson amphibolites, and the Fe-Mg minerals are much more abundant in the overlying volcanics. From Hollocher and Lent (1987).

1988). The rock types range from alaskitic to ultramafic, representing a nearly complete range of compositions from granite to peridotite. These are described in more detail under Stop 1. The mineralogy of the vast majority of these rocks is relatively dull. The common tonalitic gneisses rocks have the assemblage quartz-plagioclase-biotite-zircon-allanite-ilmenite-magnetite-epidote±hornblende±garnet±microcline±sulfides. The amphibolites have the assemblage hornblende-plagioclase-epidote±biotite±quartz±sphene±augite±ilmenite±magnetite±allanite± zircon+-sulfides. Retrograde epidote is ubiquitous in felsic and mafic rocks alike, and small quantities of retrograde chlorite are also common. These monotonous assemblages are in contrast to the interesting and varied assemblages in volcanics of the overlying Ammonoosuc Volcanics and Partridge Formation that contain gedrite, anthophyllite, sillimanite, staurolite, and garnet, which are notably scarce or absent in the Monson. To illustrate this mineralogical distinction, Figure 2 shows a histogram with the average modal abundance of six minerals in amphibolites from the Monson Gneiss and from the overlying volcanics: the Ca-rich minerals augite, epidote, and sphene, and the Fe-Mg minerals orthoamphibole, garnet, and cummingtonite. All of the samples have been chemically analyzed and have been chosen to represent relatively unaltered igneous compositions. The 26 amphibolites from the Monson Gneiss have only the calcic minerals and no Fe-Mg minerals. In contrast, amphibolites from the overlying volcanics contain abundant Fe-Mg minerals, whereas the calcic minerals are rare in rocks with unaltered compositions.

Based on a brief examination, Fred Barker of the U.S.G.S. (pers. comm. 1985) suggested that the Monson might be comparable to the calc-alkaline Central Gneiss Complex of the Coast Range batholith, that is thought by him (Barker and Arth, 1984) to be the deep-seated and metamorphosed plutonic root zone of a major volcanic arc. If true, such a scenario would support the concept of the Bronson Hill zone as a volcanic arc in Late Ordovician time.

Ordovician Cover Units: The Ammonoosuc Volcanics of presumed Late Ordovician age, is the basal unit of the cover sequence and its detailed stratigraphy is crucial to understanding the basement-cover relationship. In the early 1970's a basal quartzite and conglomerate lens was found where the Ammonoosuc overlies the Monson Gneiss in the Orange quadrangle, and in 1983 thin lenses of quartzite were found precisely on the same contact in two localities in the Quabbin Reservoir area. Despite numerous suggestions that the massive batholithic-looking gneisses are intrusive into the Ammonoosuc (Leo, 1985) and the radiometric dating that appears to make this permissible (Zartman and Leo, 1984), there remains no good documentation of intrusive gneisses substantially truncating a well defined Ammonoosuc stratigraphic sequence. The Ammonoosuc is overlain by the Partridge Formation, dominated by metmorphosed black shale and graywacke, but with significant intercalated volcanics.

The research of Schumacher (1983, 1987,Schumacher and Robinson, 1986,) and Hollocher (1983, 1985) shows that the volcanics of the Ammonoosuc and the overlying Late Ordovician Partridge Formation are quite similar in their major and trace elements to the low-K tholeiites, andesites, and dacites, as well as K-bearing rhyolites of modern island arcs such as Tonga and New Britain. An important aspect of their conclusions is that the K-bearing peraluminous rhyolites that characterize the upper part of the Ammonoosuc and continue into the Partridge, are not the product of melting of subducted North American continental crust as suggested by Robinson and Hall (1980), nor melting of subducted pelitic sediments, but were produced by melting of amphibolite or granulite of tholeiitic basalt composition within the magmatic arc complex.

Relations between Plagioclase Gneisses and Ordovician Cover: Historically there have been two ideas for the relationships between the Ammonoosuc Volcanics and underlying felsic gneisses. The first holds that the Monson Gneiss lies unconformably beneath the Ammonoosuc Volcanics (e.g., Robinson, 1979, 1981). The stratigraphic continuity of the well defined internal stratigraphy of the Ammonoosuc Volcanics and the presence of quartzite and quartz-pebble conglomerate at a few locations at the base of the Ammonoosuc support this. The much-debated alternative is that the Ammonoosuc Volcanics represent the volcanic pile overlying subvolcanic plutons, now the dome gneisses, and that these plutons are partly intrusive into the Ammonoosuc (e.g., Leo 1985; Zartman and Leo, 1984). This hypothesis is difficult to reconcile with the field relations in Massachusetts.

Recent high-precision lead isotope work done on zircons from the Monson Gneiss, the Fourmile Gneiss in the Pelham dome, the Pauchaug Gneiss in the Warwick dome, and the Swanzey Gneiss in the Keene dome have yielded ages from about 454 +4-2 to about 443 million years (Tucker et al., 1989, Robinson et al., 1989). Hard won zircons from the upper felsic member of the Ammonoosuc Volcanics on the west side of the Connecticut Valley border fault have yielded an age of 453 ±2 MY, and zircons from a felsic tuff horizon in the Partridge horizon yields an age of 449 +3/-2 MY. The ages of the Ammonoosuc Volcanics and volcanics in the Partridge Formation therefore overlap ages of the underlying Monson Gneiss and related dome gneisses.

The overlap in age as indicated by the zircon data, and the geochemical dissimilarities between the Ammonoosuc and Partridge volcanics above and the dome gneisses below suggest that both earlier ideas are incorrect. An easy resolution is not in sight, but elsewhere we have proposed that the contact could be a detachment fault that developed late in the history of the volcanic arc, bringing volcanic cover in contact with intrusive basement . A second alternative, suggested by Kohn and Spear (1989; personal communication from Kohn, 1989) as part of a completely different study of metamorphism in western New England, is that the Ammonoosuc and underlying dome gneisses may have been juxtaposed during Acadian or post-Acadian west-directed thrusting. The origin, original tectonic setting, and the paleogeography of the Monson is a problem that must be cleared up before the complete Ordovician and older history of the region can be properly written.

Stratified Silurian-Devonian Rocks: Stratified Silurian-Devonian rocks occur in two main belts, the Connecticut Valley belt to the west and the Merrimack belt to the east (Zen et al., 1983). Along the eastern edge of the Connecticut Valley belt, there has been a significant advance in the paleontological control of the thin strata characteristic of the Bronson Hill anticlinorium. Elbert, et al. (1988), have identified a rich conodont fauna in garnet-grade marble of the Fitch Formation in the inverted limb of the Bernardston nappe at Bernardston. This marble lens lies stratigraphically above, though structurally below, fossiliferous Clough Quartzite containing a poor fauna suggesting Lower Silurian age (Boucot et al., 1958). The marble lies structurally above, though stratigraphically below, gray garnet phyllites of the Devonian Littleton Formation, and appears to be truncated laterally by an unconformity at the base of the Littleton which is more generally in direct contact with the Clough. The conodont fauna gives a strong indication that the Fitch marbles in this location are earliest Devonian and not middle or upper Silurian as has been found elsewhere.

A key stratigraphic feature for the present studies is the eastward thickening of the Silurian sequence from the thin section of Clough Quartzite and local Fitch Formation in the Bronson Hill anticlinorium, at the east edge of the Connecticut Valley belt, into the thick sequence of the Merrimack belt. This consists of Rangeley Formation (lower Silurian), Perry Mountain Formation (middle Silurian), Francestown Formation (middle Silurian), and Warner Formation (upper Silurian) as identified by Peter Thompson (1985) in the Monadnock area (Figure 1), and correlated in detail by him with Silurian sequences in central New Hampshire and northwestern Maine (Hatch, Moench, and Lyons, 1983). It is the distinctive character of the well defined Monadnock sequence and its differences with the thin Bronson Hill sequence that makes possible the mapping of the early west-directed thrust nappes that have become an important part of recent tectonic reconstructions. One of these thrust-nappes carried rocks of the Monadnock sequence westward into the Hinsdale, N.H. area, where Elbert has discovered a distinctive horizon of garnet quartzite, and magnetite-cummingtonite iron formation at the top of the middle Silurian Perry Mountain Formation. This has made possible a new interpretation of the Mt. Grace area, Massachusetts (Figure 1) where identical rocks had previously been assigned to the Lower Devonian Littleton Formation (Huntington, 1975).

The Monadnock sequence appears to extend from southwestern New Hampshire across Massachusetts in a very tight zone east of the main body of Monson Gneiss and west of the Coys Hill pluton (Figure 1). In Massachusetts it has the distinctive graphite-pyrrhotite calc-silicate rocks of the middle Silurian Francestown Formation, but generally lacks the upper Silurian Warner Formation. Eastward from the Coys Hill pluton and into the Sturbridge area, studied by Berry, there are more subtle changes in stratigraphic facies. Strata tentatively correlated with the Lower Silurian Rangeley Formation are generally dominated by finer-grained pelites and lack the conglomerates observed farther west in lower Silurian strata. In addition, there are continuous beds, rather than merely boudins, of calc-silicate granulite, as well as layers of metamorphosed ribbon limestone, that suggest these rocks may more closely approximate the Lower Silurian Sangerville Formation in central Maine. Perry Mountain correlatives seem to be scarce or absent. The position of the Francestown Formation is taken by the distinctive graphite-pyrrhotite-cordierite schists and gneisses of the Smalls Falls Formation that is a key marker unit in this region, but several lithically identical layers are believed to be members of the Rangeley Formation (Berry, 1989b) Calc-silicate granulites stratigraphically above the Smalls Falls, in some belts previously assigned to the Paxton Formation of Massachusetts (Zen et al., 1983), appear to correlate with the Warner Formation in New Hampshire and the Madrid Formation in Maine. It should be emphasized that these stratigraphic distinctons are recognized in rocks metamorphosed in the granulite facies as compared to the greenschist facies to lower amphibolite facies in central and western Maine, but some of the lithic distinctions are so marked that they can be recognized through even this intense metamorphism. Stratigraphy still provides the essential control for local and regional structural and tectonic studies.

STRUCTURAL GEOLOGY

Sequence of Acadian Deformations: Acadian deformations in central Massachusetts and adjacent New Hampshire have been summarized elsewhere (Robinson, 1979; Robinson and Hall, 1980; Hall and Robinson, 1982). These are broadly divided into an early nappe stage, an intermediate backfold stage, and a late gneiss dome stage, each with many complications.

Recent research in the Monadnock area, New Hampshire (P.J. Thompson, 1985), the Hinsdale area, New Hampshire (Elbert, 1986), the Mt. Grace area, Massachusetts (Robinson, 1987), and the Sturbridge area, Massachusetts and Connecticut (Berry, 1987a, 1987b) has shown that the early west-directed fold nappes are truncated by a slightly later set of west-directed thrust nappes. In southwestern New Hampshire and adjacent Massachusetts the two major nappe-stage thrust faults are: the Brennan Hill thrust carrying previously folded Monadnock sequence strata across Bronson Hill sequence strata containing the Bernardston fold nappe, and the Chesham Pond thrust carrying Rangeley Formation and Kinsman Granite over previously folded rocks of the Monadnock sequence. The Brennan Hill thrust appears to trace southward from the Monadnock area through the Mt. Grace area and along the east margin of the main body of Monson Gneiss to Connecticut. The Chesham Pond thrust appears to trace southward from the Monadnock area along the west margin of the Coys Hill pluton and also may extend into Connecticut. In the Sturbridge area in southern Massachusetts and adjacent Connecticut the thrust nappes form a series of west-directed imbrications involving pre-Silurian basement rocks and their Silurian-Devonian cover, already isoclinally folded, in a major thrust complex.

The structural features tentatively assigned to the backfold stage include the following, not necessarily in chronological order:

1) Longitudinal flowage of the main and Tully bodies of Monson Gneiss, from a position south of Quabbin Reservoir, 50 km northward to a position overlying strata in the Mt. Grace area, causing recumbent folding that involuted the axial surfaces of early fold nappes as well as the trace of the Brennan Hill thrust.

2) East-southeast directed recumbent folding of the basement-cover contact in the Pelham gneiss dome (Ashenden, 1973) and the Keene gneiss dome (Robinson, 1963, 1967b; Schumacher and Robinson, 1986). Amphibole lineation associated with this phase has been identified locally in the Keene gneiss dome and in gedrite gneisses overgrown by cordierite believed to have been produced by tectonic unloading during the gneiss-dome stage. At the south end of the Pelham gneiss dome these structural features are truncated by the Belchertown Quartz Monzodiorite intrusion that has a zircon age of 380 million years (Ashwal et al., 1979). The rocks of both the dome and the intrusion are structurally overprinted by the dome stage of deformation. Unfortunately no field relations have yet been found that show the relative age of these east-directed recumbent folds to the west-directed fold or thrust nappes, or to other features tentatively included in the phase of backfolding.

3) Eastward overturning of axial surfaces of west-directed fold nappes and west-directed thrust surfaces. P.J. Thompson (1985) and H.N. Berry (1987b) have independently suggested the possibility of grand overturning of the entire eastern part of the orogen from the Monson Gneiss across the entire Merrimack belt and possibly beyond. Tentatively associated with this eastward overturning, but not conclusively linked with it, is an E-W trending linear fabric and E-W trending minor folds in metamorphosed sedimentary rocks and a wide variety of tonalitic through granitic intrusions (Robinson, 1979, Peterson, 1984). These are progressively overprinted westward by north- or northeast-trending folds and fabrics definitely related to the dome stage.

4) Development of a series of mylonites in metamorphosed sedimentary and intrusive rocks containing an E-W trending lineation believed to be related to the shear direction parallel to the E-W

lineation described under 3) above. The mylonites cut across coarse-grained migmatitic schists and gneisses formed during peak granulite-facies metamorphism, and are deformed by northeast-trending minor folds related to the dome stage. In an early structural correlation (Robinson, 1979, Robinson, Field, and Tucker, 1982) it was proposed that the mylonites formed during late stages of backfolding following peak metamorphism. With discovery of early west-directed thrust nappes by Berry (Robinson et al., 1986b) it was alternatively suggested that the mylonites are related to the thrust nappes which were then tentatively considered to post-date the peak granulite facies metamorphism. This alternative has now been definitively disproved for three reasons: 1) Several thrust surfaces related to thrust nappes are truncated by tonalite intrusions (Berry, 1987a); 2) Many tonalite intrusions are involved in the mylonites, some of them extensively so (Finkelstein, 1987; Berry, 1987b); and 3) The shear sense of the mylonites studied so far, including mylonitized tonalites, indicate a consistent west-over-east shear sense inconsistent with the west-directed thrusting. It is thus now concluded that mylonite formation was part of the phase of backfolding as originally suggested, and unrelated to thrust nappes.

There is a swirling pattern of lineations related to the gneiss dome stage of deformation, superimposed on all previous features. In the Pelham dome there is conclusive evidence that this lineation, trending N-S parallel to the dome axis, is parallel to the transport direction of a series of sheath folds (Ashenden, 1973, Onasch, 1973, Robinson, 1979, Reed and Williams, 1989). There is a suspicion that the pattern of lineations associated with the dome stage, with its evidence of the gravitational rise of low density buoyant basement strata, may also reflect a deep-seated ductile regime of longitudinal shear related to the terminal phase of Acadian collisional tectonics (cf. Ellis and Watkinson, 1987).

Post-Acadian Deformation: Subsequent to the dome stage of Acadian structural development, large areas of central Massachusetts seem to have remained relatively quiescent, but there is evidence for local late Paleozoic igneous activity, metamorphic recrystallization, and ductile deformation (Robinson, 1963, 1967b; Halpin, 1965; Hollocher, 1981; P.J. Thompson,1985) On Mt. Monadnock, P.J. Thompson (1985) mapped two microdiorite dikes that appear to truncate all Acadian fold structures but are themselves slightly metamorphosed and also intimately involved with intrusions of Fitzwilliam Granite.

The region was subjected to large-scale extensional faulting during Triassic-Jurassic time. Most important was the Connecticut Valley border fault that is thought to be listric in character. At the New Hampshire-Vermont line, reconstructed cross sections of pre-Mesozoic rocks suggest the west side was downthrown about 5 km and indications are that displacement on the same fault at the Connecticut line may have been as much as 8 km, allowing observation of the metamorphic rocks at widely different tectonic levels. A second indication of Mesozoic tectonic activity is a group of diabase dikes and sills, until recently all presumed to be Jurassic. A recent study combining paleomagnetism, geochemistry, and K-Ar dating has shown that they range in age from Early Jurassic through Early Cretaceous (McEnroe and Brown, 1987; McEnroe et al., 1987; McEnroe, 1988, 1989).

METAMORPHISM

Metamorphic Zones: The pattern of regional metamorphic zones in central Massachusetts based on the petrology of pelitic schists is shown in Figure 3. Details of the metamorphic zones are given by Tracy et al., (1976), Tracy (1978), Robinson, Tracy, Hollocher and Dietsch (1982), and Robinson et al., (1986b), although much detailed analytical data has yet to be published. Hollocher (1985) and Renate Schumacher (1986) have given details of progressive reactions in amphibolites in the same region, and Schumacher and Robinson (1986, 1987) have given details concerning the formation of cordierite in gedrite gneisses during progressive unloading in the Keene gneiss dome. Recently Berry (1989b) has documented the occurrence of wollastonite as a product of regional metamorphism in the granulite facies of the Brimfield-Sturbridge area. The

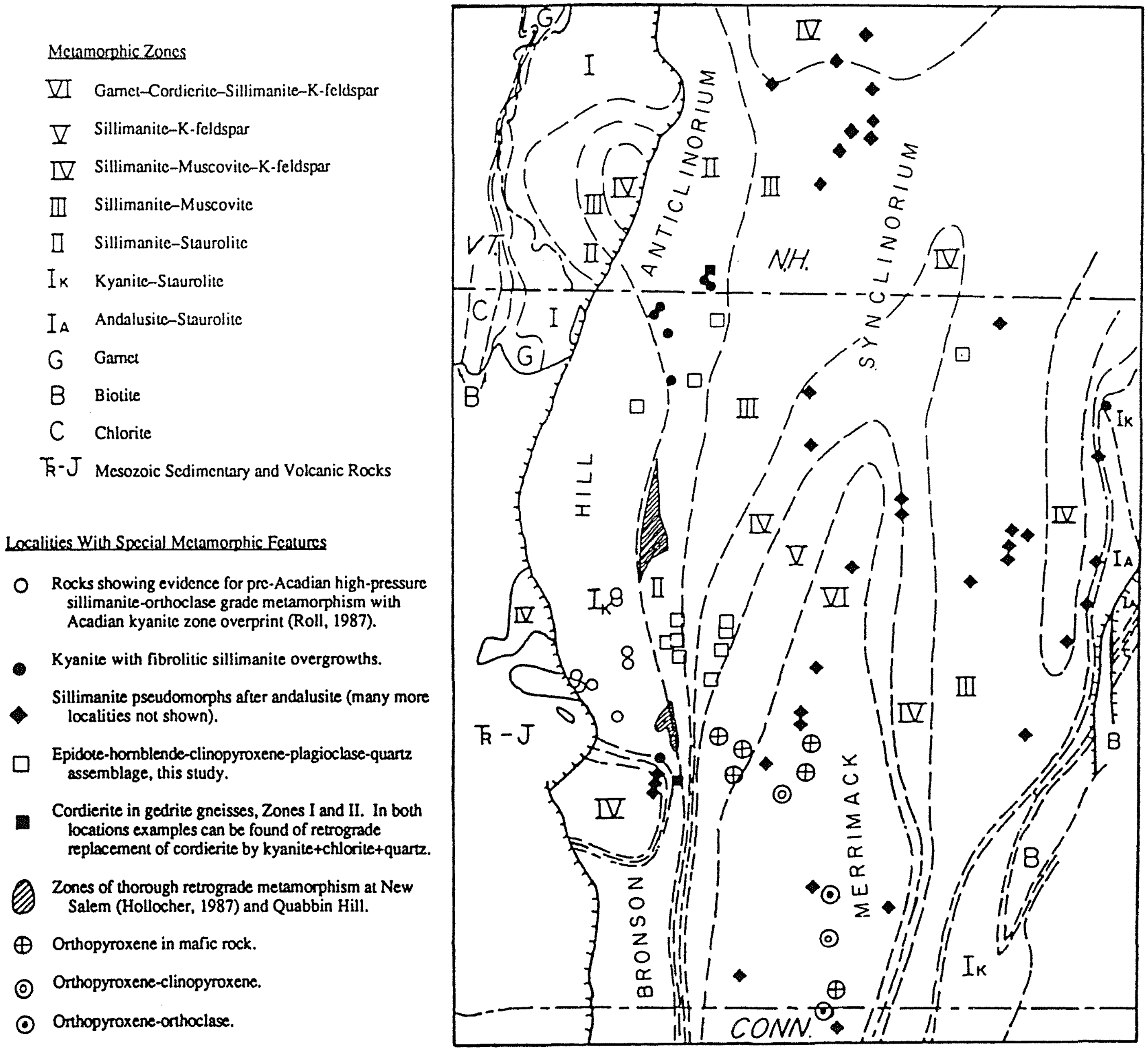

Figure 3. Generalized metamorphic map of west-central Massachusetts and adjacent states showing distribution of special metamorphic features (from Schumacher, Schumacher and Robinson, 1988).

relations of metamorphic zones and metamorphic reactions to the structural evolution of the region is still poorly understood despite an impressive amount of detail in specific locations.

The six highest grade zones and their typifying assemblages in pelitic schists are as follows:

I: Kyanite-Staurolite. Quartz-muscovite-biotite-garnet-staurolite±kyanite-ilmenite-graphite.

II: Sillimanite-Staurolite. Quartz-muscovite-biotite-garnet-staurolite-sillimanite-ilmenite-graphite.

III:Sillimanite-Muscovite. Quartz-muscovite-biotite-garnet -sillimanite-graphite-ilmenite±pyrrhotite+plagioclase, similar to staurolite- free rocks rocks in the upper part of Zone II.

IV:Sillimanite - Muscovite - K feldspar. Quartz-muscovite-orthoclase-plagioclase(An_{20-35})-biotite-garnet-sillimanite-graphite-ilmenite±pyrrhotite. At any one locality rocks with more sodic plagioclase than in the typical assemblage have sillimanite - orthoclase without muscovite, whereas rocks with more calcic plagioclase have sillimanite-muscovite without orthoclase (Tracy 1975, 1978). This is the lowest grade zone to be visited on this field trip.

V: Sillimanite - K feldspar. Quartz-orthoclase-plagioclase-biotite-garnet-sillimanite-graphite-ilmenite±pyrrhotite and no muscovite is found in prograde rocks of any available composition.

VI:Sillimanite-Orthoclase-Garnet-Cordierite Zone. Quartz-orthoclase-plagioclase-biotite-garnet-cordierite-sillimanite-graphite-ilmenite±pyrrhotite.
At certain straatigraphic levels in Zone VI as well as locally in Zones IV and V there are extremely sulfide-rich rocks lacking garnet, commonly with rutile in place of ilmenite, and in rare instances pyrite together with pyrrhotite. The genesis of these rocks is discussed in detail below.

Sequence of Reactions in Pelitic Schists: The change from sillimanite-muscovite assemblages in Zone III through sillimanite-muscovite-orthoclase assemblages in Zone IV to sillimanite-orthoclase assemblages in Zone V has been studied extensively by Tracy (1975, 1978). The controlling reaction is a continuous one essentially involving continuous increase in muscovite and K-feldspar K/Na ratios and plagioclase Ca/Na ratios:

1) Na-richer Muscovite + Na-richer Plagioclase + Quartz =
K-richer Muscovite + K-feldspar + Sillimanite + Ca-richer Plagioclase + H_2O

The Na lost from muscovite and from plagioclase in this reaction goes to make a larger amount of K-feldspar, which is more sodic than muscovite. The Zone III-Zone IV boundary is defined by the first appearance of sillimanite-K feldspar in schists with the most sodic plagioclase composition, in practice about An_{20}. The Zone IV-Zone V boundary is defined by the last appearance of muscovite in schists with the most calcic plagioclase, in practice, about An_{33}. Within Zone IV muscovite compositions range from 7 down to about 2% paragonite component in the highest grade rocks. The highest grade muscovites contain 10-20% of a celadonite component and .04 to .08 Ti ions per 11 oxygens as compared to .01 to .05 in lower grade muscovites.

So long as muscovite is stable in Zones III and IV there is little prograde change in the Fe/(Fe+Mg) ratios of garnets and biotites in sillimanite-garnet-biotite assemblages. A fairly obvious reason is to be found in the controlling reaction:

2) Garnet + Muscovite + Mg-richer Biotite = Sillimanite + Fe-richer Biotite + Quartz

First inspection suggests that since the K + Na of muscovite must be exactly balanced by the K + Na of biotite, the H_2O must balance and the reaction is fluid conservative and hence is likely to have a very small ΔS. However, C.V. Guidotti (personal communication, 1982) has called attention to the well known fact that muscovites have higher Na than coexisting biotites and hence that albite component of plagioclase should be added to the right-hand side of the equation. This would increase the ΔS of the reaction and would slightly favor the sillimanite + biotite side of the reaction with increasing grade.

Once muscovite has been entirely replaced by K-feldspar in Zone V the situation is entirely reversed. K-feldspar takes the place of muscovite in reaction 2, the amount of H_2O produced on the left hand side is large and so is the ΔS. This is rewritten as

3) Sillimanite + Fe-richer Biotite + Quartz =Garnet + K-feldspar + Mg-richer Biotite + H_2O.

This powerful dehydration reaction is responsible for major changes in mineral compositions. The Mg contents of garnets and biotites in the sillimanite-biotite-garnet-orthoclase assemblage increase progressively from values typical of Zone V (Figure4) to the higher values characteristic of Zone VI (Figure 5). In Zone VI this garnet reaction "collides" with a similar powerful dehydration reaction involving cordierite and moving in the reverse direction:

4) Sillimanite + Mg-richer Biotite + Quartz = Cordierite + K-feldspar + Fe-richer Biotite + H_2O

This "collision" produces the sillimanite - orthoclase - garnet - cordierite - biotite assemblages typical of Zone VI.

Geothermometry and Geobarometry: To date most geothermometry in the region has been based on Fe/Mg partioning between the pairs garnet-biotite and garnet-cordierite, using the empirical calibrations of A.B. Thompson (1976b) and the experimental calibration of Ferry and Spear (1978). Before calibrations can be realistically applied, the patterns of internal zoning in garnets must be understood, as discussed by Tracy et al. (1976) and in more detail by Robinson, Tracy, Hollocher and Dietsch(1982c), Tracy (1982), Spear and Selverstone (1983), and Chamberlain (1985). Garnets in pelitic schists in central Massachusetts have been divided into three types, A, B, and C (Robinson, Tracy, Hollocher and Dietsch, 1982c). Type A garnets, characteristic of zones I and II show a compositional record of prograde growth by continuous and discontinuous reactions with matrix minerals. Temperatures were less than 650 °C so that solid diffusion within garnet was too slow to homogenize the recorded compositions. Type B garnets, characteristic of zone IV and most of zone V, have homogeneous interiors (Figure 6), which either grew that way originally or were homogenized by diffusion at temperatures of 650 to 675 °C. In addition they have continuous rims believed to have formed by retrograde continuous hydration reactions with matrix minerals in the presence of a fluid phase. Type C garnets, characteristic of part of zone V and all of zone VI (Figure 7), also have homogeneous interiors. However, rim compositions differ only where they are in direct contact with another ferromagnesian mineral such as biotite or cordierite. Chemical gradients within the biotite and cordierite show that the garnet rims were produced by localized ion exchange during cooling. The lack of rims where Type C garnets don't touch ferromagnesian minerals suggests that retrograding did not involve a pervasive metamorphic fluid as it did in type B garnets. The implications of these observations are that prograde metamorphism at 675 to 730 °C. eventually drove away most metamorphic fluids, so that they were not generally available for retrograde reactions. This would be consistent with ideas that many of the migmatite features in zone VI are products of fluid-absent melting (see Stop 4 of this field trip). We believe such implications are related to the fact that the mylonites in zone VI (Stops 7 and 9, this field trip) have the finest grain size and smallest degree of metamorphic recrystallization of any in the region.

To date geobarometry has been based primarily on the Fe-Mg continuous reaction cordierite = quartz + sillimanite + garnet, despite the extended debate concerning the sensitivity of the calibration of this equilibrium to the water content of cordierite (Hensen and Green, 1973;

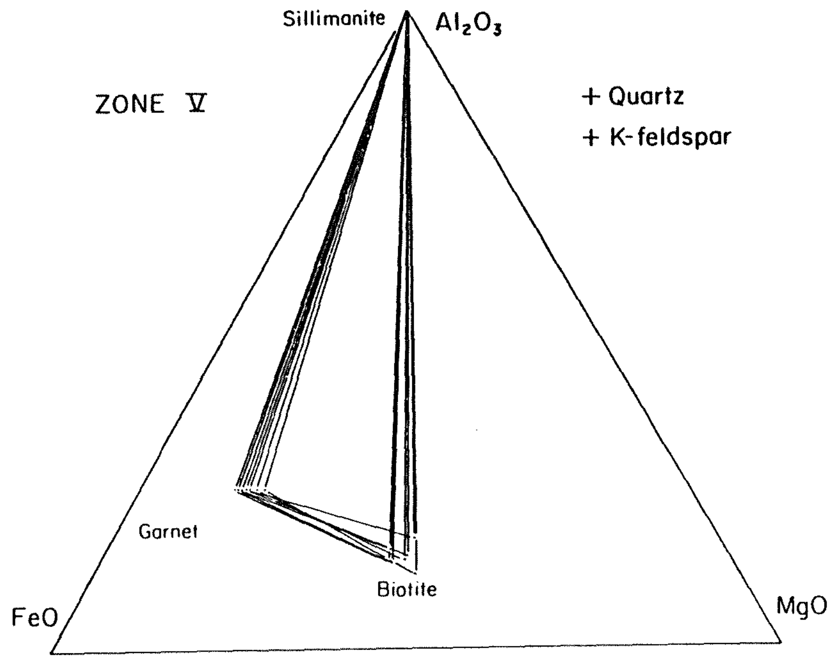

Figure 4. Quartz + K-feldspar projection of pelitic schist assemblages in Zone V. Only garnet core compositions are shown.

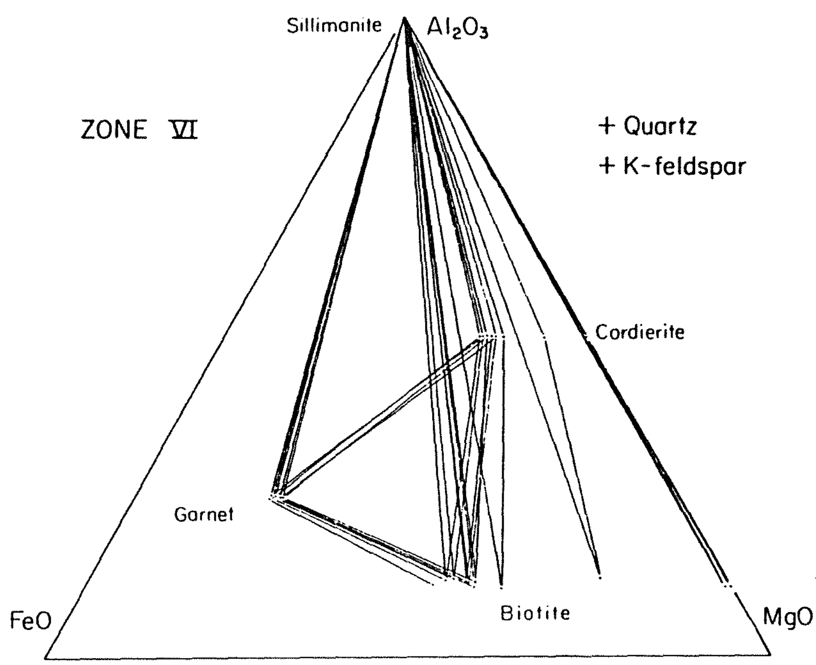

Figure 5. Quartz + K-feldspar projection of some pelitic schist assemblages in Zone VI. Only garnet core compositions are shown and some more recent garnet assemblage data is not included. The extremely magnesian assemblages are all indeed three-phase assemblages!

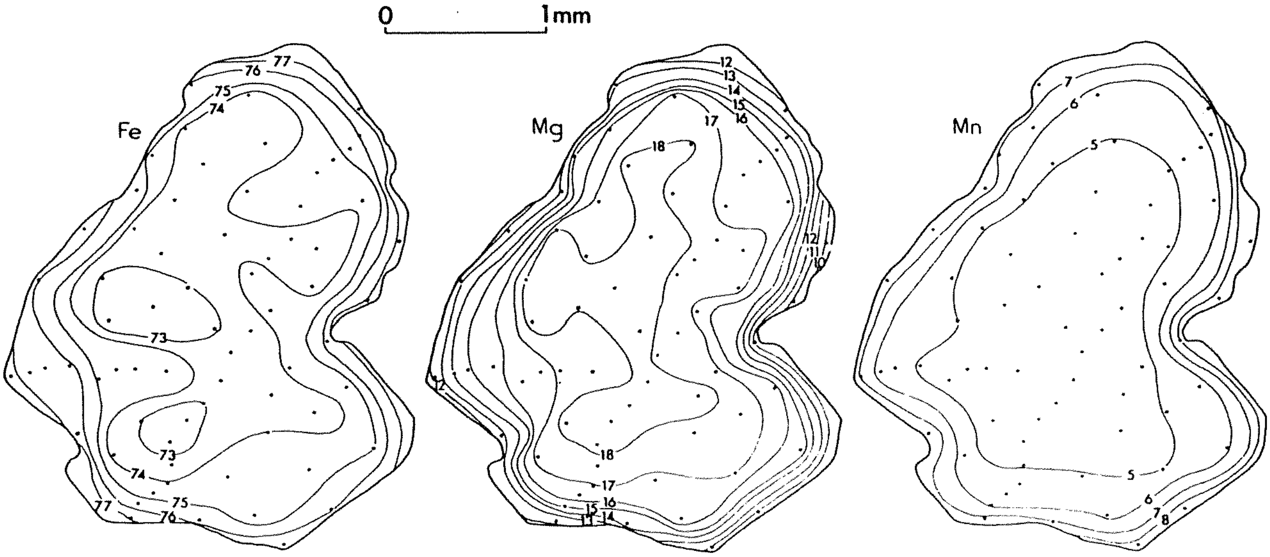

Figure 6. Contour maps of mole % Fe, Mg, and Mn in garnet 933B, Zone IV, a typical type B zoned garnet. Ca content varies little. The assemblage is quartz-orthoclase-biotite-garnet-sillimanite. Dots are analysis points.

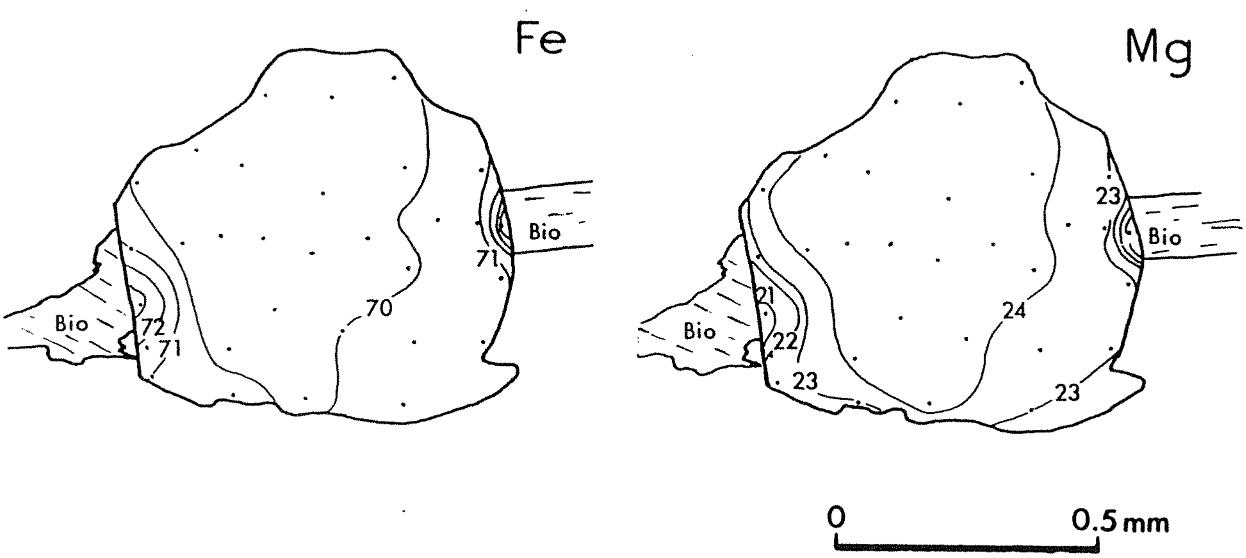

Figure 7. Contour maps of mole % Fe and Mg in garnet FW 407, Zone VI, a typical type C zoned garnet. Ca and Mn content varies little. The assemblage is quartz-orthoclase-biotite-garnet-cordierite-sillimanite. Dots are analysis points.

Weisbrod, 1973; Thompson, 1976b; Hensen, 1977; Holdaway and Lee, 1977; Newton and Wood, 1979; Martignole and Sisi, 1981; Lonker, 1981). The use of this reaction may be justified, however, because the real chemistry of the rocks is very close to that of the calibration system. The actual cordierite assemblages are found in pelitic schists only in Zone VI, also in a few gedrite gneisses in lower grade zones (Schumacher and Robinson, 1986, part G, Table G-8), and in one well documented K-poor pelite in the Monadnock area (Thompson, 1985). They commonly yield estimated pressures of about 6 kbar. Elsewhere the garnet compositions in sillimanite (or kyanite) - garnet - quartz assemblages are used to estimate minimum pressures which are 6 kbar or less. In this guide, we also report the results of the application of the widely used "GAsP" or "QGAP" barometer, based on the reacton anorthite=Al_2SiO_5 + quartz + grossular (Ghent, 1976; Newton and Haselton, 1981; Koziol and Newton, 1988). We have found that the results of the GASSP calculations aare quite consistent with pressures estimated from garnet-cordierite equilibria. The GASP barometer has been dramatically improved in precision in recent years, largely because of the re-calibration of Koziol and Newton (1988) and refinement in garnet activity models (e.g. Ganguly and Saxena, 1984; Hodges and Crowley, 1985). The models for grossular activity in metamorphic garnets and anorthite activity in plagioclase are critical because the naatural materials are quite far in composition from the calibration system. This dependence on activity models also affects the widely applicable equilibrium anorthite + annite = grossular + almandine + muscovite (Ghent and Stout, 1981; Hodges and Royden, 1984). An additional widely used barometer, GRAIL, based on the equilibrium garnet + rutile = Al_2O_3 + ilmenite (Bohlen, Wall and Boettcher, 1983) is not strictly applicable to the central Massachusetts rocks because the absense of garnet-rutile assemblages. However, the common co-existence of garnet + sillimanite + ilmenite allows calculation of maximum pressure. Typical results of this calculation for samples discussed in this guide yield upper pressure limits of about 7 kbar.

Silicate-Sulfide-Oxide-Fluid Reactions in High-Grade Pelitic Schists: Sulfidic-graphitic schists are abundant in central Massachusetts. These include the well known pyrrhotite schists characteristic of the Ordovician Partridge Formation, parts of the Lower Silurian Rangeley Formation and also the still more sulfidic Middle to Late Silurian Smalls Falls Formation. The latter was first defined as the White Schist Member of the Paxton Formation by Field (1975) and is now known to extend along strike more than half way across the state. Regional studies now suggest it is certainly correlative with the Smalls Falls of northwestern Maine (Berry, 1985) where Guidotti et al. (1975, 1977, personal communication, 1982) have done detailed petrologic studies. The first quantitative reports of these rocks in Massachusetts were given by Tracy et al., 1976; Robinson and Tracy, 1977; and Tracy and Rye, 1981, and the following discussion is based mainly on those reports plus a recent comprehensive report (Tracy and Robinson, 1988) and data and calculations in preparation for publication by Tracy.

The silicate assemblages in some sulfide- and graphite-bearing pelitic schists from Zones IV and V (dashed tie lines) and Zone VI (solid tie lines) are shown in quartz + K-feldspar projection in Figure 8. Zone V assemblages include sillimanite - garnet - biotite - orthoclase in the Partridge and Rangeley Formations and one sillimanite - cordierite - biotite - orthoclase assemblage from the Smalls Falls. The figure shows only one Zone IV assemblage of sillimanite - cordierite - biotite - orthoclase - muscovite from the Smalls Falls. Zone VI assemblages include sillimanite - garnet - cordierite - biotite - orthoclase assemblages from the Partridge and Rangeley Formations, and a variety of sillimanite - cordierite - biotite - orthoclase assemblages with different Fe/Mg ratios from the Smalls Falls. It is important to note that in this projection, assuming no variation in non-AFM components of represented phases, under one condition of pressure, temperature, composition of projective phases, and activity of H_2O or fluid composition, there should be only one equilibrium triangle for the sillimanite - cordierite - biotite assemblage, not the whole array shown. The variation shown, hence, must be due either to different P-T conditions or different activities of H_2O at different outcrops, or to non-AFM variations in solid-phase compositions. In the lack of any evidence of other variations, a model based on variable aH_2O is considered below.

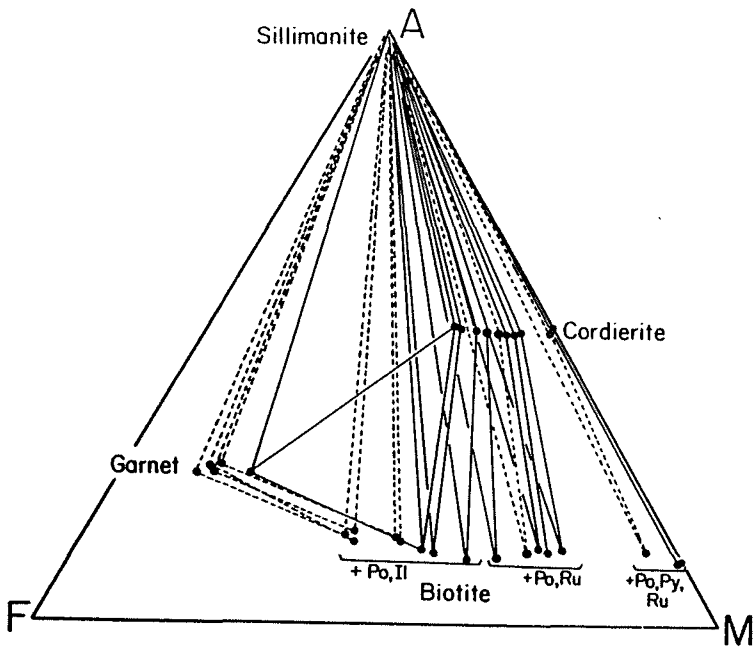

Figure 8. AFM (Quartz + K-feldspar) projection showing the data from Tables 2 and 3 of Tracy and Robinson (1988), along with garnet and biotite data presented elsewhere (Tracy, Robinson, and Thompson, 1976). All assemblages indicated with solid lines are from metamorphic zone VI and formed under very similar conditions of T and P. Assemblages shown using dashed lines are from zones IV and V and formed at temperatures up to 50° C lower than the other samples. Note that accessory minerals occurring in each assemblage type are indicated and correspond with phase relations shown in Figure 5 of Tracy and Robinson (1988).

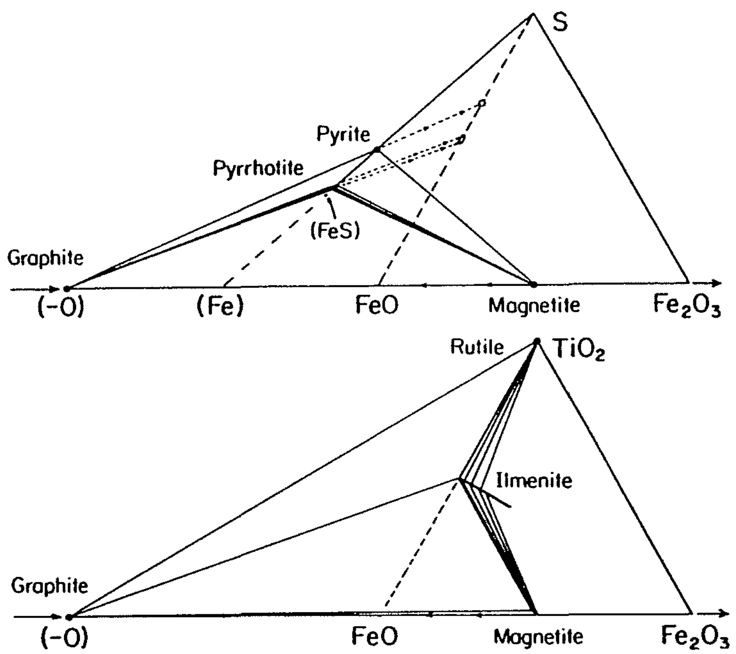

Figure 9. Projections of mineral compositions from C-O-H fluid in the system $FeO-Fe_2O_3-S$ and $FeO-Fe_2O_3-TiO_2$ using the method of Thompson (1972).

Letter symbols in Figure 8 indicate the characteristic sulfide and Ti-oxide assemblage to be found in each group of silicate assemblages. The more Fe-rich silicate assemblages contain pyrrhotite and ilmenite. Modestly Mg-rich assemblages contain pyrrohtite and rutile, and the most magnesian assemblages contain pyrite, pyrrhotite, and rutile. This systematic relationship is explored below.

Thompson (1972) has shown how assemblages in the system Fe-O-C-H can be conveniently projected from H_2O and CO_2 onto the line FeO-Fe_2O_3. In this scheme of things, represented by the bases of both triangles in Figure 9, graphite projects at -1 Fe_2O_3 and native Fe at -1/2 Fe_2O_3. To this base has been added S so that we can see the relations between graphite, magnetite, pyrrhotite and pyrite, and TiO_2 so that we can see the relations between graphite, magnetite, ilmenite, and rutile in equilibrium with C-O-H fluid. In the upper part of Figure 10 the two parts of Figure 9 are combined in a tetrahedron showing all significant graphite assemblages. These significant graphite assemblages may also be shown more easily using a graphite projection onto the triangular plane TiO_2-S-FeO as in the lower part of Figure 10. To this convenient graphite projection MgO may now be added to form the tetrahedron TiO_2-S-FeO-MgO (Figure 11) in which ferromagnesian minerals may be treated. The universal ferromagnesian silicate in these assemblages is biotite which would appear on or close to the FeO-MgO line by projection from quartz and K-feldspar. Aluminum saturation of biotite in such assemblages is provided by garnet, sillimanite, cordierite or any combination of the three. The volume of Figure 11 is filled by four-phase assemblages and three-phase tie planes as follows, beginning in the Fe-rich corner. The most FeO-rich rocks would have (graphite) - magnetite - pyrrhotite - ilmenite - biotite. We have not definitely observed this assemblage in Zone VI in central Massachusetts, but a comparable assemblage (graphite) - magnetite - pyrrhotite - ilmenite - grunerite - olivine (no quartz) has been described in Zone II by Huntington (1975). Next comes a large array of three-phase tie planes (graphite) - pyrrhotite - ilmenite - biotite corresponding to the middle region of Figure 8. This is followed by the four-phase volume (graphite) - pyrrhotite - ilmenite - rutile - biotite and then the array of tie planes (graphite) - pyrrhotite - rutile - biotite. For still more magnesian compositions there is the four-phase volume (graphite) - pyrrhotite - pyrite - rutile - biotite and finally the array of tie planes (graphite) - pyrite - rutile - biotite. As will be discussed below, the composition of biotite in equilibrium with ilmenite and rutile (front face of tetrahedron) is a function of a dehydration reaction in which the biotite becomes more Mg-rich with increasing grade. Similarly the composition of biotite in equilibrium with pyrrhotite and pyrite (base of tetrahedron) is also a function of a prograde reaction, in this case a mixed volatile reaction.

In order to understand more about the origin of these sulfide-rich rocks we will now take a brief excursion into sedimentary geochemistry using the bottom of Figure 11 as a chemographic work space. Consider an environment slightly below the seawater/sediment interface in which organic matter, sulfur-reducing bacteria, and detrital ferromagnesian minerals are interacting. For simplicity's sake, detrital grains can be considered as two extreme kinds, fine-grained highly reactive grains and coarser detrital grains that resist equilibration with interstitial fluid until low-grade metamorphic conditions are reached. To simplify further, assume that all detrital grains have the same Mg/(Mg + Fe) ratio of 0.50 as illustrated in Figure 12-A. In the early diagenetic stages reactive detrital grains are attacked by fluids or sulfur-reducing bacteria to produce an assemblage of pyrite plus iron-depleted Mg-rich silicate grains having bulk composition on the line between the detrital composition and S. Between diagenesis and low grade metamorphic re-equilibration the pelite would contain three significant components: pyrite, reacted Mg-rich oxide and silicate, and non-reacted Fe-Mg oxide and silicate. During low grade metamorphism the reacted and nonreacted silicate and oxide components would equilibrate to produce a mean ferromagnesian silicate and oxide composition coexisting with pyrite. Sulfur isotope data (Tracy and Rye, 1981; Tracy and Robinson, 1988) is consistent with a model in which diagenetic reaction occurs between sedimentary sulfur, reactive detrital ferromagnesian grains and organic carbon. All of the analysed sulfides from the Smalls Falls samples have d34S ranging from -25 per mil to -29 per mil. These very light sulfur values could only have originated through bacterial reduction of porewater sulfate

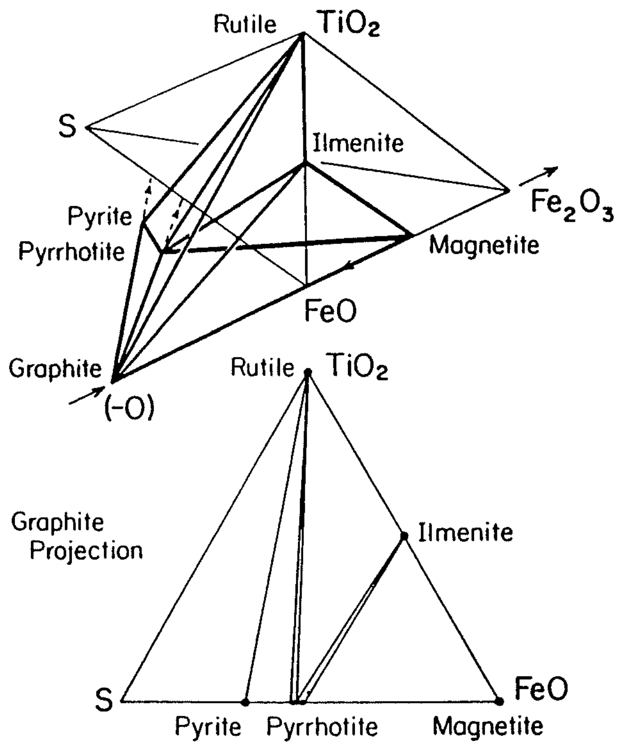

Figure 10. Top: Combination of triangles of Figure 9 into a tetrahedron FeO-Fe_2O_3-S-TiO_2. Assemblages with graphite shown with heavy lines. Bottom: Projection of graphite-bearing assemblages from graphite onto the plane FeO-S-TiO_2.

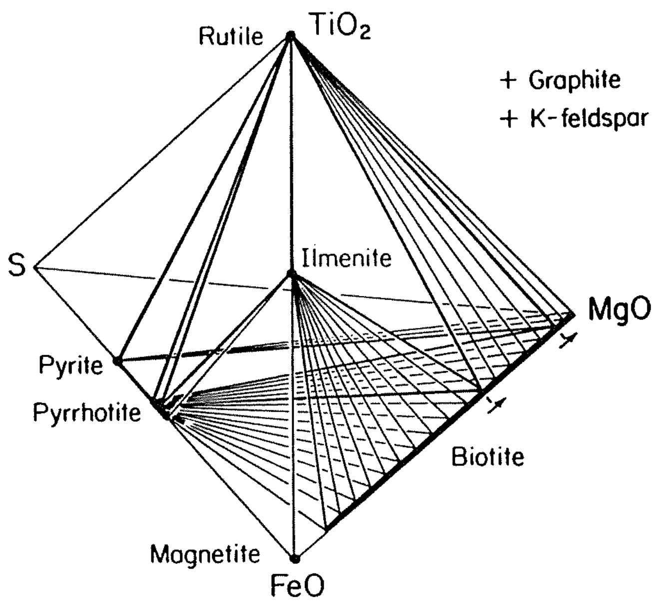

Figure 11. Graphite-projected plane FeO-S-TiO_2 with MgO added to show ferromagnesian minerals. Further projection from quartz and K-feldspar permits portrayal of biotite.

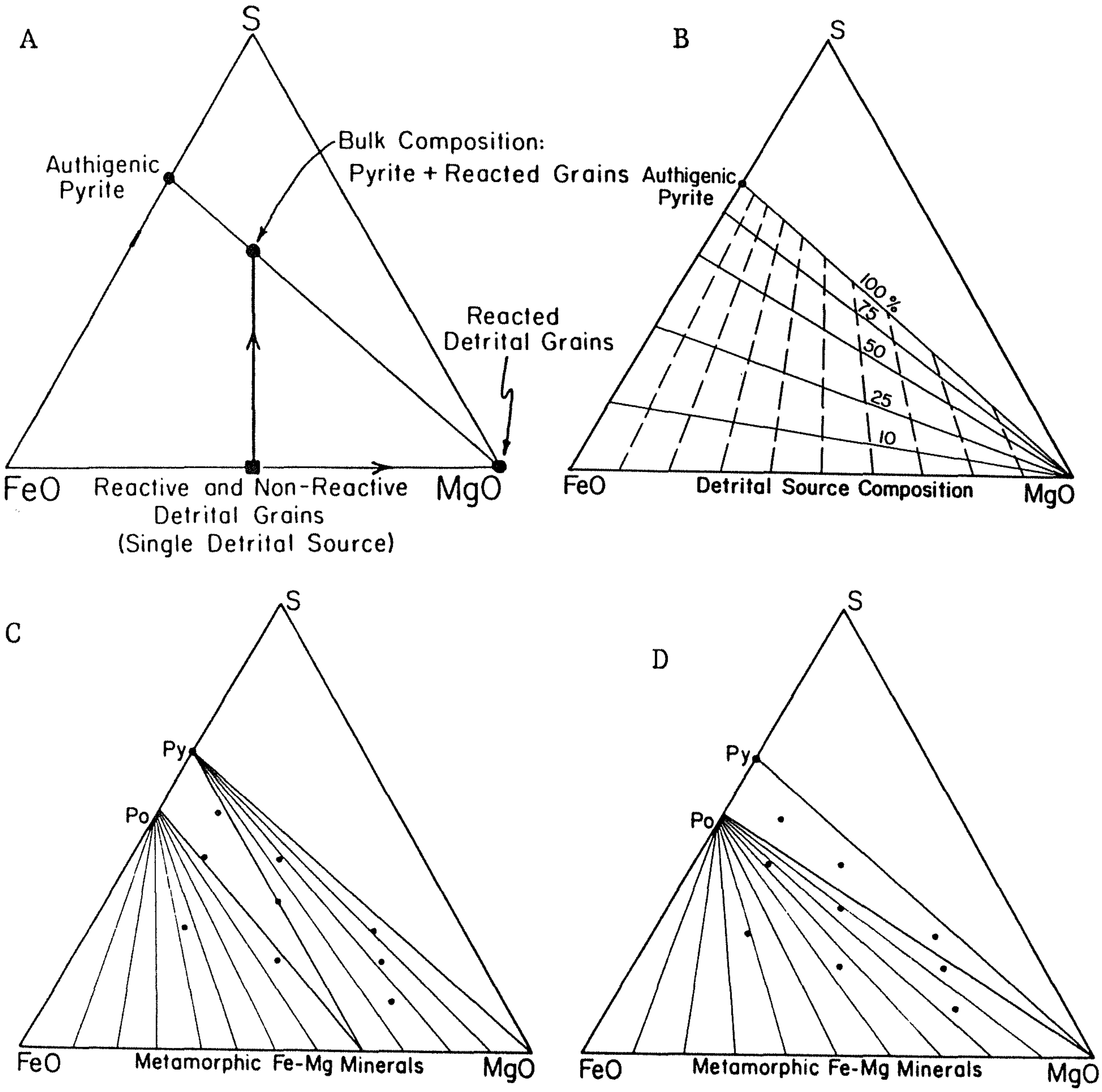

Figure 12. Diagenetic and metamorphic evolution of sulfidic shale portrayed in graphite projection on the plane FeO-S-MgO.

A) Reaction of detrital grains with sulfur (H_2S, etc.) to produce sediment with three components; pyrite, reacted detrital grains, and non-reacted detrital grains.

B) Range of low grade schist bulk compositions dependent on detrital source composition and percent of reactive grains.

C) Progress of continuous devolitilization reaction Fe-richer biotite + pyrite + graphite = pyrrhotite + K-feldspar + Mg-richer biotite + $2H_2O$ + $3CO_2$ across various bulk compositions.

D) Termination of the continuous reaction when all Fe is removed from biotite. Note that pyrite is still retained in some bulk compositions and is even more abundant than pyrrhotite in some.

in an open-system sedimentary environment. Sedimentary sulfides form from reaction of bacterially-produced H_2S and reactive detrital iron-bearing minerals (Berner, 1971). A modern analogue of this process can be found in the highly reducing deep zones of the Black Sea, where sedimentary sulfides have virtually identical d34S values to those in the Smalls Falls Formation.

Figure 12-B illustrates the possible variability of low grade pyrite plus ferromagnesian silicate rocks produced by the processes described above. Dashed lines illustrate the variable bulk compositions attainable from fixed detrital compositions of Mg0 to Mg100 in increments of Mg10, and proportions of reactive grains ranging from 0 to 100%. The solid lines illustrate variable bulk compositions attainable with constant proportions of reactive grains as labeled, but with variable detrital compositions. The combination of variable detrital compositions and variable proportions of reactive grains can produce a very wide range of low-grade bulk compositions in the FeO-MgO-S triangle available for higher-grade reactions.

The progress of the biotite-pyrite-pyrrhotite equilibrium through a range of bulk compositions is illustrated in Figure 12-C and -D. This permits one to observe directly the effect of bulk composition, and particularly proportion of primary reactive grains, on the sulfide-silicate assemblages produced in later metamorphism. In Figure (12-C the pyrite-pyrrhotite equilibrium has progressed part way across the diagram, and there are three assemblages: pyrrhotite - biotite, pyrrhotite - pyrite - biotite, and pyrite - biotite. Note in particular that in the pyrrhotite - biotite assemblage in rocks of the same bulk Fe/Mg ratio, those with a large amount of pyrrhotite will have more magnesian silicates. This effect has been well noted in modes by Henry and Guidotti (1981), and Mohr and Newton (1981). In Figure 12-D the pyrite-pyrrhotite reaction has gone virtually to completion so that pyrrhotite and pyrite coexist with essentially end-member silicates. It will be noted that for some bulk compositions originally very rich in reactive silicate grains there is not enough Fe in the bulk silicate composition to make more than a token amount of pyrrhotite even at very high grade. This appears to have been the case at the outcrop to be visited at Stop 5, where we originally reported that there was no pyrrhotite at all (Tracy et al., 1976).

We now think that the variable compositions in sillimanite - cordierite - biotite - orthoclase assemblages may be a direct consequence of their variable sulfide content and variable progress of a mixed volatile pyrrhotite-producing reaction which dilutes H_2O with other species in the fluid phase and thereby reduces μH_2O The controlling silicate reaction (see reaction 4 above) is shown schematically in terms of XMg (biotite), temperature and fluid composition in Figure 13-A, temperature and XMg of reacting phases in Figure 13-B, and quantitatively in terms of μH_2O as a function of XMg of reacting phases in the isothermal, isobaric diagram of Figure 13-C (see Tracy and Robinson, 1988). This is a continuous dehydration reaction and hence attains a maximum temperature in the presence of a pure H_2O fluid. Because of the Fe/Mg distribution between biotite and cordierite this reaction begins at lowest temperature in pure Mg compositions and proceeds thence to more Fe-rich compositions. This is reflected in biotite composition isopleths on the fluid composition diagram. The differences in μH_2O of various cordierite assemblages may also be viewed in composition projection in Figure 14.

Although we have previously described the metamorphic fluid as a C-O-H fluid, it actually contains hydrogen sulfide. Fluid composition in FW-882E, an assemblage containing pyrite + pyrrhotite + graphite, was calculated by Tracy and Robinson (1988); at 700 °C and 6 kilobars, the log fO2 was -18.1 and the mole fractions of species in the fluid were: H_2O 0.35, CO_2 0.13, H_2S 0.51, CH_4 0.01, H_2 <0.01. The calculation of fluid composition for the assemblage pyrrhotite + graphite under the same conditions (except for slightly lower fO_2 and fS_2) shows that the ratio of H_2O and CO_2 increases dramatically, but XH_2S drops to about 0.04. The calculations indicate the following mole fractions: H_2O 0.79, CO_2 0.05, H_2S 0.04, CH_4 0.11, H_2 <0.01. High H_2S in the first fluid is a direct result of the high fS_2 which is buffered by pyrite + pyrrhotite. Ongoing fluid composition calculations for high-temperature sulfidic rocks indicate that the rare presence of pyrite in such rocks, typically the result of anomalously high S/Fe ratios in protoliths, causes

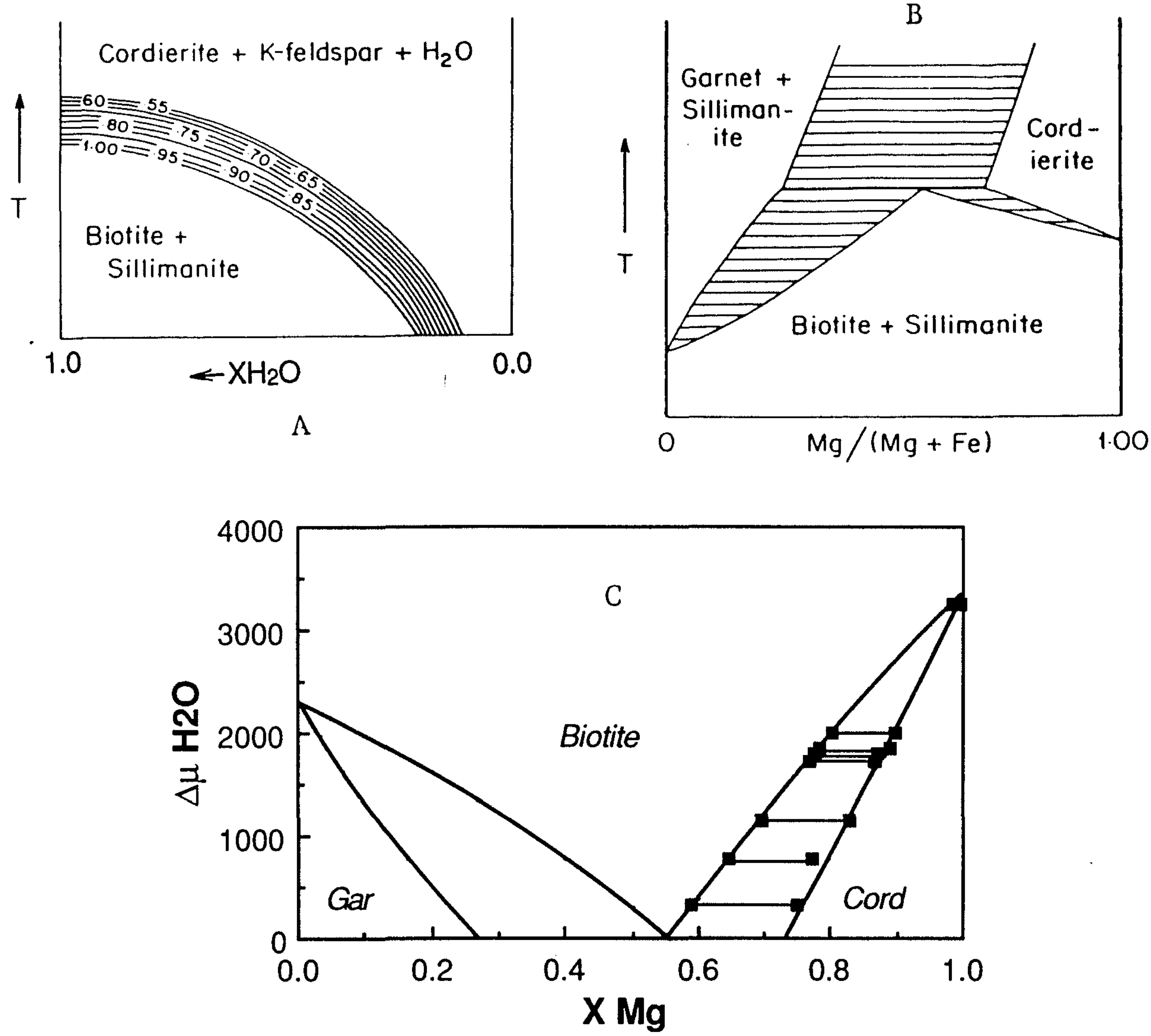

Figure 13. Diagrams illustrating relationships of composition and intensive variables for the controlling silicate reaction (4) in sulfidic schists.

A) Schematic temperature-fluid composition diagram for reaction (4) with isopleths for X_{Mg} of biotite.

B) Temperature-composition diagram for the cordierite join in quartz-, K-feldspar and water-projected system KFMASH, showing three T=X loops. On left, biotite + sillimanite= garnet + Kfeldspar + H_2O (reaction 3). On right: biotite + sillimanite = cordierite + Kfeldspar + H_2O (reaction 4). In middle: Biotite + sillimanite = garnet + cordierite + H_2O.

C) $\Delta\mu$H2O - XMg diagram for reactions (3) and (4) calculated at 700°C and 6 kbar, showing the biotite-garnet-sillimanite and biotite-cordierite-sillimanite loops representing equilibria (3) and (4), respectively. The biotite-cordierite pairs shown on the latter loop are analyzed compositions from the central Massachusetts sulfidic schists (Tracy and Robinson, 1988).

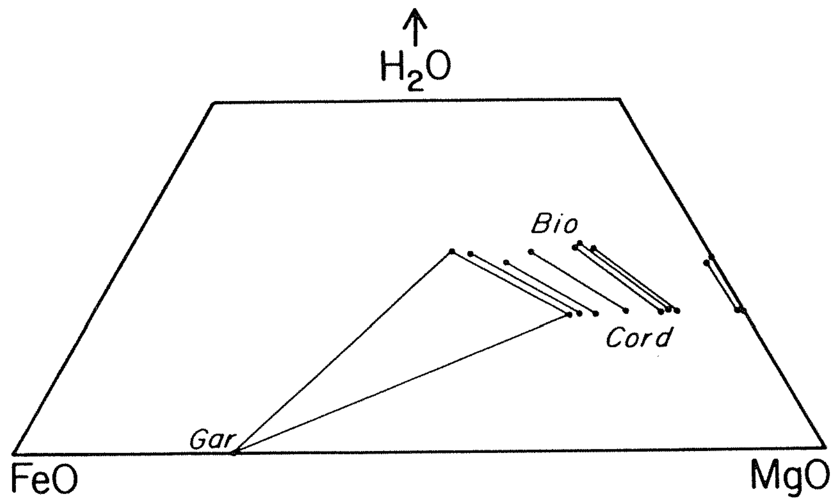

Figure 14. Projection onto the FMH plane from sillimanite, Kfeldspar and quartz in the system KFMASH (after Rumble, 1974). All sillimanite-bearing assemblages shown in solid lines in Figure 8 are shown here projected to illustrate that the crossing tielines in the three-phase and four-phase assemblages from zone VI can be explained by local variations in μH_2O, with lowest μH_2O in the more Fe-rich assemblages and highest in the Mg-rich, pyrite-pyrrhotite assemblage.

unexpected fluid behavior. At 700°C and 6 kbar, the presence of pyrite with pyrrhotite buffers a log fS_2 of about 0. At log fO_2 below that of the H_2O maximum (Ohmoto and Kerrick, 1977) fluids in equilibrium with pyrite + pyrrhotite + graphite can contain up to 90 or more mole percent H_2S. These fluids are particularly odd in that, although reduced and saturated with graphite, they contain very little of C-bearing species: CO_2 is inhibited by low fO_2, and CH_4 is replaced as the reduced species in the fluid by H_2S at such high fS_2.

An indication of the much higher XH_2S in rocks containing pyrrhotite + pyrite occurs in the form of sulfur-bearing cordierite. Probe analyses of cordierites in all the pyrite + pyrrhotite assemblages yield H_2S of about 1 to 1.5 weight %. On the other hand, sulfur has not been found in cordierite from any rock in which only pyrrhotite occurs, no matter how much modal pyrrhotite there is. Apparently the cordierite is able to accomodate H_2S in its structural channels, but only when the XH_2S of the fluid is exceptionally high, as in pyrite + pyrrhotite-bearing rocks. Recent research at V.P.I. and S.U. has shown that fluid inclusions are quite abundant in many of the sulfidic schist samples (Tracy and Sheets, 1988). Fluid inclusions in quartz in the vicinity of pyrite + pyrrhotite show heating/freezing and homogenization behavior that indicates the presence of significant H_2S (estimated at 40-60 mol. percent) and suggests that these may well be preserved primary inclusions given their agreement with the fluid composition calculations. Laser Raman spectroscopy of these inclusions at V.P.I. has confirmed the presence of major proportions of H_2S in them. Interestingly, the large K-feldspar porphyroblasts in at least one of the sulfidic schist samples contain apparent primary, very water-rich fluid inclusions which constrain a rather low-pressure P-T isochore. This isochore intersects the muscovite dehydration equilibrium in the andalusite stability field, raising the fascinating possibility that the apparent andalusite porphyroblasts pseudomorphed by later sillimanite may have formed along with the K-feldspar in an early muscovite dehydration event at rather low temperature and pressure. If true, this early Buchan-type metamorphism is in accord with the anti-clockwise P-T path suggested for this terrane by Tracy and Robinson (1980).

Metamorphic Reactions in Mafic Rocks: Amphibole dehydration reactions in central Massachusetts and southwestern New Hampshire from kyanite grade (Zone I) to the granulite facies (Zone VI) have been studied in some detail (e.g., Robinson and Jaffe, 1969a, b; Huntington, 1975;Wolff, 1978; Schumacher, 1983; Hollocher, 1985). The phase relations involving subaluminous rocks with compositions broadly similar to basalts are shown in Figure 15. These phase relations are partly based on partially constrained positions of some three-phase fields, and probable locations of other three-phase fields based on theory.

Four reactions are particularly important in converting medium-grade amphibolites into pyroxene granulites. In rocks with tholeiitic bulk compositions, these four reactions form a mineral assemblage facies series with increasing grade. Tholeiitic compositions, projected from plagioclase onto the AFM plane (Figure 15; also see Figure 33), generally lie within the hornblende field or between the hornblende field and the Fe-Mg join. In order of increasing grade, the most important facies assemblages and bounding reactions are:

Hornblende-Orthoamphibole (Mg-rich rocks)

Orthoamphibole + hornblende + quartz = cummingtonite + plagioclase + H_2O

Hornblende-Cummingtonite

Cummingtonite + plagioclase = hornblende + orthopyroxene + quartz + H_2O

Hornblende-Orthopyroxene

Hornblende + quartz = augite + orthopyroxene + plagioclase + H_2O

Orthopyroxene-clinopyroxene

Augite + garnet + quartz = orthopyroxene + plagioclase

Clinopyroxene-Garnet (Fe-rich rocks)

Mg-rich refers to bulk compositions with Xmg of about 0.55 and higher, and Fe-rich refers to rocks with Xmg of about 0.4 and lower, at the approximately 6 kbar pressure estimated for peak Acadian metamorphism in Central Massachusetts. This facies series occurs in central Massachusetts over a temperature range of about 600 to 715^{0}C, estimated using the Thompson (1976b) garnet-biotite geothermometer (Tracy et al., 1976).

Partial Melting in Mafic Rocks: Quartz-bearing mafic rocks (and other rocks) in metamorphic Zones IV, V, and VI commonly contain coarse-grained leucocratic layers, veins, dikes, pods, and larger bodies of generally tonalitic composition. These leucocratic segregations have long been interpreted as locally derived, crystallized partial melts. These crystallized melts can be divided into two general types based on grain size: coarse-grained varieties with grains typically 10-20 mm, and fine-grained varieties with grains typically 1-2 mm (beware, since most of both varieties have undergone some post-crystallization deformation and grain size reduction). The coarse-grained crystallized melts may have resulted from fluid-present (vapor-saturated) melting, or deposition from H_2O-rich solutions. The fine-grained types may have resulted from fluid-absent (vapor-undersaturated) melting. A Penrose Conference field trip(Geological Society of America Penrose Conference, Migmatites and Crustal Melting, June 1986) visited some of these outcrops, and the nature of the leucocratic segregations was strongly argued on many fronts. No consensus was reached.

In metamorphic Zones IV and V the melt segregations rarely make up more than a few percent of any outcrop, but can occupy over 15% of outcrops in Zone VI. The melt zones in mafic rocks typically consist of quartz, gray plagioclase, and several percent mafic minerals that can include cummingtonite, hornblende, garnet, biotite, orthopyroxene, augite, and oxides. K-feldspar is generally rare or absent, indicating that these crystallized melts were not derived from local schists or biotite breakdown, which result in K-feldspar-rich melts. The tonalitic melts must have formed

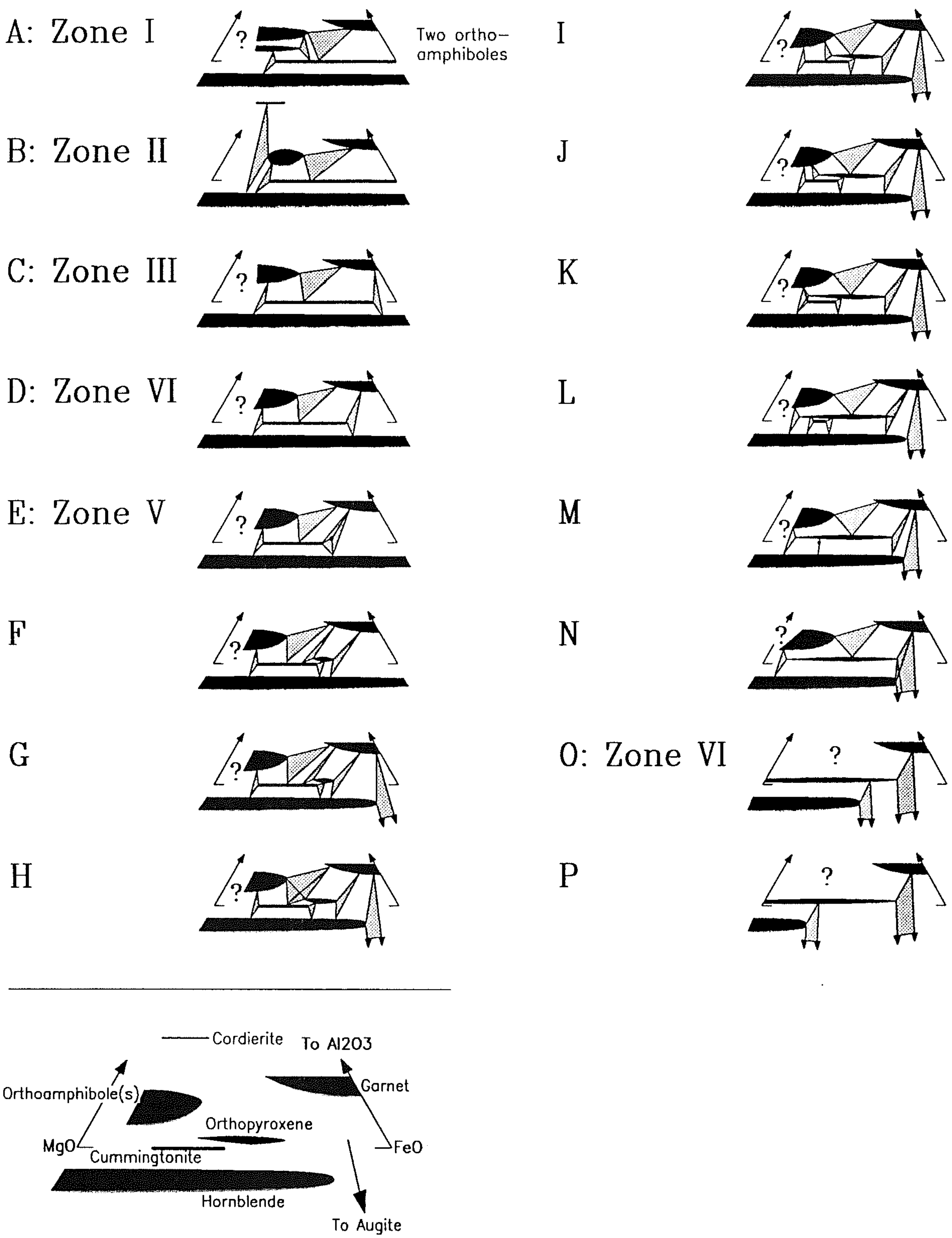

Figure 15. Plagioclase projections onto the AFM plane schematically showing the phase relations of subaluminous amphibolites and pyroxene granulites in central Massachusetts and southwestern New Hampshire. Adapted from Hollocher (1985).

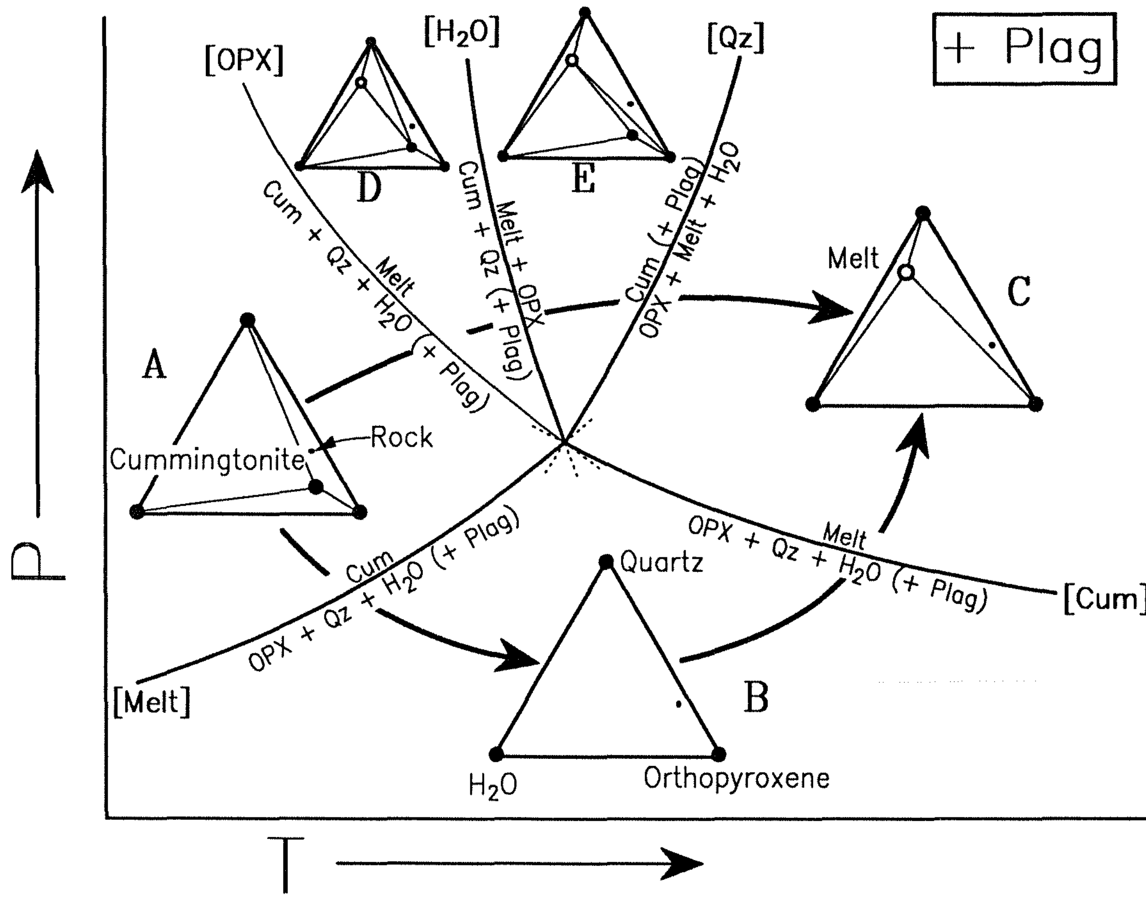

Figure 16. P-T diagram showing melting and other reactions for a simplified mafic rock having the initial assemblage cummingtonite-quartz-orthopyroxene-plagioclase, plus other minerals that do not participate in the reactions. The four reactions that change the phase relations in the example rock, labeled, are shown in heavy lines. The remaining reaction that is part of the Schreinemacher net but has no effect on the rock assemblage is shown with a light line. Abbreviations: Cum = cummingtonite, OPX = orthopyroxene, Qz = quartz, Plag = plagioclase, H_2O = H_2O-rich vapor phase, Liquid = hydrous tonalitic liquid.

as a result of dehydration reactions involving the amphiboles, principally hornblende and cummingtonite. The reactions would be similar to ordinary metamorphic dehydration reactions, except that tonalitic liquid is one of the reaction products instead of H_2O.

To illustrate partial melting, consider the simplified case of a mafic rock composed of plagioclase, cummingtonite, quartz, and other minerals such as hornblende that do not participate in the reaction examples. These minerals can be shown in the ternary plagioclase-present chemical system quartz-H_2O-orthopyroxene. Cummingtonite occurs within the triangle, as does hydrous tonalitic melt. Figure 16 is a Schreinemacher net of the phase relations in this simplified system. Note the position of the hypothetical mafic rock near the quartz-orthopyroxene side of the triangle. Triangle A (Figure 16) represents upper amphibolite grade conditions (Zone V) in which the cummingtonite-orthopyroxene-quartz-plagioclase (+hornblende) assemblage has actually been found.

There are three cummingtonite-bearing assemblages in triangle A, but common rocks are H_2O-poor and have only the cummingtonite-quartz-orthopyroxene assemblage. At low pressures (lower arrow in Figure 16), increasing temperature would move the rock across the [Melt] line, which is a vapor-producing dehydration reaction that is terminal for cummingtonite and results in the phase relations in triangle B. If the vapor (metamorphic fluid) escapes from the rock, the rock composition would move over to the anhydrous quartz-orthopyroxene sideline. With increasing temperature or pressure, if the vapor phase remains or if H_2O-rich vapor infiltrates from elsewhere, the rock would cross the [Cum] reaction line and undergo fluid-present melting. The phase relations in triangle C show that under these conditions the rock contains the assemblage quartz-orthopyroxene-melt. This melt could crystallize as leucocratic segregations.

At higher pressures (upper arrow in Figure 16), the rock traverses a different set of reactions. Starting from triangle A, the melting reaction [OPX] is crossed first and produces the phase relations in triangle D. However, this reaction has no effect on the rock because the rock is shielded from the reaction assemblages by a quartz-cummingtonite tie line. When the rock crosses reaction [H_2O] the rock undergoes fluid-absent melting to produce a quartz-orthopyroxene-melt assemblage shown in triangle E. Further increases in temperature will eventually result in a fluid-present melting reaction along the line [Qz], which is also terminal for cummingtonite like reaction [melt]. The resulting phase relations are those in triangle C, with the rock having the assemblage quartz-orthopyroxene-melt.

Remember that in real rocks dehydration and melting reactions are more complicated because of differing bulk compositions, mineral assemblages, and mineral compositions, changes in mineral composition during reaction, and variations in metamorphic fluid composition from place to place and with reaction progress. Despite these caveats, keep the differences between fluid- present melting, fluid-absent melting, and non-melting dehydration reactions in the following outcrop descriptions and in the field.

P-T Evolution in Gneiss Domes: In the Ammonoosuc Volcanics at the south end of the Keene gneiss dome amphibolites and gedrite gneisses contain an amphibole lineation parallel to the axes of local backfold-stage southeast-directed recumbent folds. The gedrite lineation is overgrown by cordierite that forms complex reaction rims around sillimanite and other aluminous minerals as a result of unloading believed to have taken place during the rise of the gneiss dome. Recently Seifert and Schumacher (1986) have developed a pressure calibration based on the equilibrium between cordierite, zincian spinel, and quartz, that has been applied to the assemblages in these aluminous enclaves, and indicates pressures of 6.3 to 3.7 kbar for the time of enclave formation. Gedrite-garnet gneisses commonly contain both kyanite followed by sillimanite, and also conclusive evidence for kyanite forming as a retrograde product with chlorite and quartz after cordierite (Robinson, 1963; J.C. Schumacher, pers. comm. 1987). Garnet with up to 32% pyrope content in some of these rocks may be a relict of an earlier, higher pressure history before the cordierite-forming reactions (Schumacher and Robinson, 1986, Table G-8). Collecting in summer 1987 has led to discovery of previously unrecognized assemblages of coexisting Mn-rich orthopyroxene and augite in the middle Garnet-Amphibole Quartzite Member of the Ammonoosuc (Schumacher, 1988) where more normal Fe-Mg rich assemblages contain only amphiboles. This appears to be an example where abundant MnO stabilizes an assemblage normally considered characteristic of the granulite facies into P-T conditions of the amphibolite facies for more normal rock compositions.

P-T Path in the Granulite Facies Region: The granulite facies area near Sturbridge, Massachusetts appears to be the area of most intense Acadian metamorphism in North America and its complex tectonic, mineralogic, and thermal evolution requires much more intense study. The widespread occurrence of sillimanite pseudomorphs after andalusite indicates an early low-pressure metamorphic history. That this low-pressure history involved high temperatures and partial melting is suggested by the rare occurrence of andalusite pseudomorphs in pegmatites and the

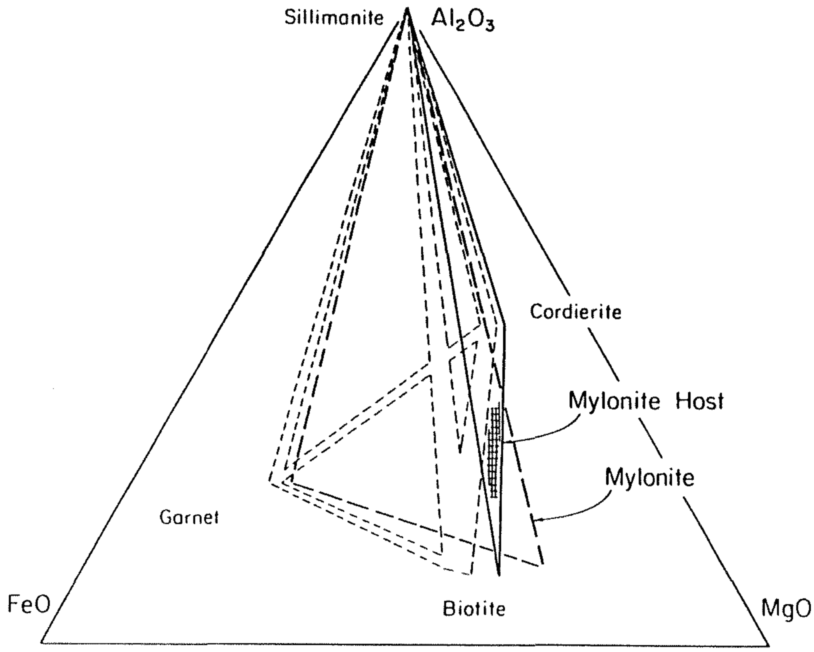

Figure 17. Quartz + K-feldspar projection of mineral compositions at Stop 7, including garnet-cordierite gneisses (light dashed lines), the mylonite host, and the mylonite.

widespread occurrence of cordierite-bearing pegmatites. In the one sample of cordierite pegmatite that has been studied in detail (Tracy and Dietsch, 1982) the cordierite has a 10% higher Fe/(Fe+Mg) ratio than typical cordierite in sillimanite-garnet-cordierite-quartz assemblages in the adjacent granulite facies pelites. This is taken to indicate that the partial melting that produced the pegmatite took place at considerably lower pressures than those indicated by peak granulite facies assemblages. Further, this cordierite is shown to have been in the process of breaking down on its own composition to an intergrowth of sillimanite, garnet, and quartz plus more Mg-rich cordierite. Similar evidence is being found by J. A. Thomson in a pristine suite of 48 samples of such cordierite pegmatites collected from the pipeline trench.

Peak-metamorphic sillimanite-garnet-cordierite-quartz-biotite assemblages in pelites in the region suggest temperatures of 685-780 °C and pressures of 5.6 - 6.3 kbar. Hollocher (1985; Robinson et al., 1986b) reported several occurrences of the classic granulite facies assemblage of orthopyroxene-orthoclase-quartz (Figure 3) and several other occurrences where orthoclase adjacent to orthopyroxene has been replaced by a high temperature retrograde symplectite of Ti-rich biotite and quartz. Several occurrences of the assemblage orthopyroxene-garnet-plagioclase-quartz (Hollocher, 1985; Robinson et al., 1986b) can be applied to the pressure calibration of Perkins and Chipera (1985) and yield pressure estimates of 5.2 to 7.0 kbar at 700 °C.

The late metamorphic history of this region is also intriguing. Keys to this late history are found in assemblages developed in mylonites that cross cut the peak metamorphic fabrics. Many of the mylonitic rocks are not distinct petrologically from sheared versions of the peak metamorphic assemblages. However, a few have undergone such severe grain size reduction that they have been able to undergo complete recrystallization to new fine-grained assemblages representative of a different metamorphic facies (Robinson et al, 1977; Robinson, Tracy, Hollocher and Dietsch, 1982; Robinson et al.,1986b). A mylonite in one host rock consisting of coarse quartz, K-feldspar, sillimanite, cordierite and biotite has recrystallized to a new assemblage of quartz, K-feldspar, sillimanite(?), garnet, and Mg-richer biotite (Figure 17). Garnet-biotite Fe-Mg exchange thermometry suggests the mylonite recrystallization took place at around 550 °C and the Mg-rich garnet composition (approx. 30% pyrope) suggests crystallization at a minimum pressure of 7-8 kbar. If these indications are correct, then it would appear that the rocks of the central Merrimack belt may have continued to be compressed as they cooled beyond the peak of metamorphism, a situation that would appear to require special tectonic conditions. This proposed "counterclockwise" P-T path for the Merrimack belt is in sharp contrast to the more traditional "clockwise" path (England and Thompson, 1984; Thompson and England, 1984) for the adjacent Bronson Hill anticlinorium (Figure 18). Tectonic correlations between the two regions would suggest that subsequent to tectonic loading in the nappe stage, rocks of the Bronson Hill anticlinorium were progressively unloaded, while those of the Merrimack belt were being progressively loaded, even in late stages where they were already undergoing cooling.

TECTONIC-METAMORPHIC MODEL FOR BRIMFIELD-STURBRIDGE AREA

The Brimfield-Sturbridge area is in the heart of the Acadian granulite-facies region of south-central Massachusetts and northern Connecticut (Figures 1, 3, and 19). Following suggestions to the northwest by Field (1975), to the north by Tucker (1977), and to the northeast by Robinson (1981b), Berry suggested that units correlative with the Silurian-Devonian Rangeley sequence are present in this area (Berry, 1985). Detailed mapping has demonstrated the distributions and contact relationships of these units (Figure 20), leading to two major results. One is the identification of a sequence of pre-Silurian"basement" rocks, perhaps as old as Precambrian. The other is a new structural and tectonic model for the Acadian evolution of this part of the granulite facies terrane.

The pre-Silurian"basement" rocks are dominated by light gray, quartz-plagioclase-biotite±magnetite gneisses, commonly with boudins or thin layers of hornlende amphibolite. In some places, thin, mappable horizons of metamorphosed sedimentary rocks are bounded on both sides by the plagioclase gneisses (Figure 20). Earlier, Berry supposed that these horizons might be remnants of Silurian-Devonian strata preserved in isoclinal synclines or thrust slices within the"basement" (see Fig. H-7 of Robinson et al., 1986b). These rocks are now thought to belong to the pre-Silurian package. The lithic assemblage and sequence of rock types suggest these pre-Silurian rocks may correlate with Ordovician(?) to Precambrian rocks of southeastern Connecticut, east-central Massachusetts, or south-central Maine (Berry, 1989b). If correct, this would be one of the few areas in which Silurian-Devonian strata of the Merrimack synclinorium are exposed together with their pre-Silurian continental basement.

A structural and tectonic model for the Acadian evolution of south-central Massachusetts has followed directly from application of these stratigraphic interpretations to the map (Figures 21 and 22; Berry, 1987a, 1987b). An early phase of low-P, high-T metamorphism produced andalusite, preserved in pseudomorphs, over a broad area of the Merrimack synclinorium including central Massachusetts and northern Connecticut (Robinson, Tracy, Hollocher, and Dietsch, 1982; Figure 3). This metamorphism also produced pre-peak-metamorphic cordierite-bearing pegmatites and dehydration in south-central Massachusetts. The earliest structural features are west-directed fold-nappes and thrust-nappes of the Acadian nappe stage of deformation (Figure 22). The thrust-nappes comprise the imbricate stack of thin slices containing basement and cover rocks which dominates the map pattern (Figure 20). The several slices contain similar rocks, so they are presumed to have come from a common source area, as symbolized by the thrust-duplex structure in Figure 22.

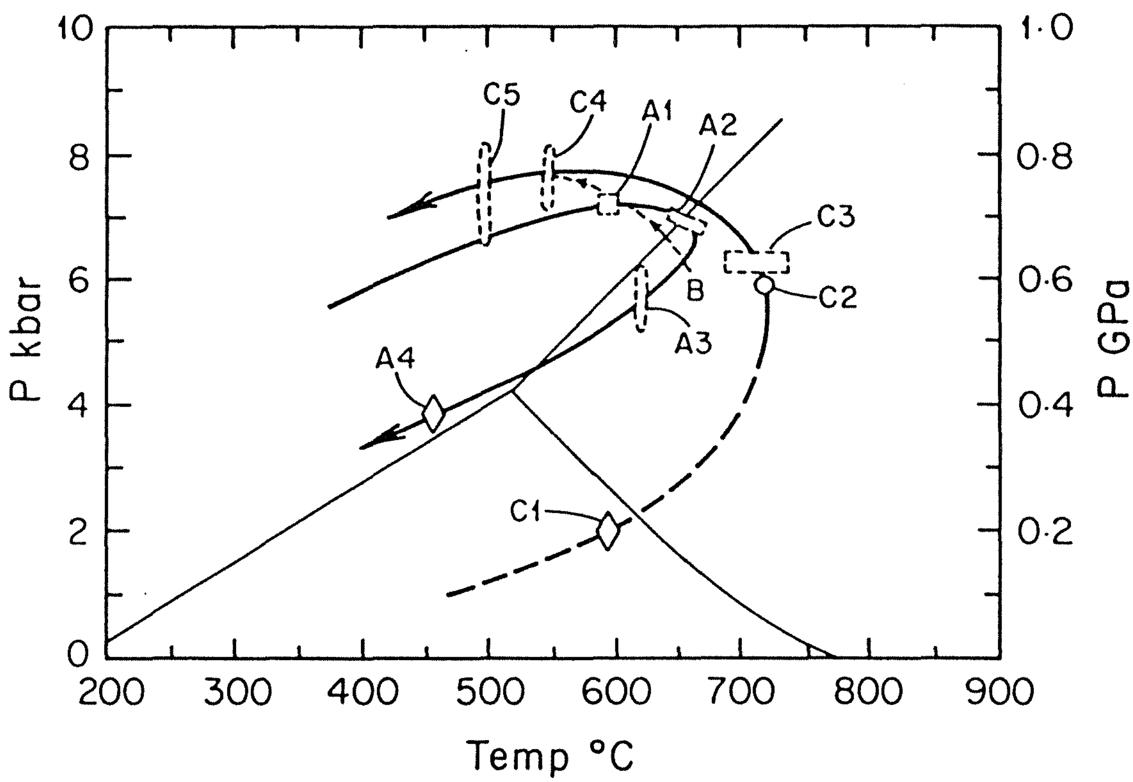

Figure 18. P-T trajectories from central Massachusetts and southwestern New Hampshire. A) central Bronson Hill anticlinorium near Keene dome based on the following: A1 - specimen I34I, garnet-biotite and garnet-cordierite thermometry, sillimanite-kyanite-quartz-garnet-cordierite barometry (Schumacher and Robinson, 1986, Table G-8); A2 - specimen 7AOBX, garnet-biotite and garnet-cordierite thermometry, kyanite-sillimanite-quartz-garnet-cordierite barometry (Schumacher and Robinson, 1986, Table G-8); A3 - cordierite enclaves, cordierite-spinel-quartz barometry with estimated temperature of 625°C (Schumacher and Seifert, 1986); A4 - kyanite-chlorite-quartz intergrowths in coarse cordierite (Robinson and Jaffe, 1969). B) eastern edge of Bronson Hill anticlinorium, Quabbin Reservoir area. Path of epidote formation, (Schumacher et al., 1988). C) Merrimack synclinorium, south-central Massachusetts based on the following: C1 - sillimanite pseudomorphs after andalusite, some of these occur in pegmatite indicating high T formation; C2 - beginning of breakdown of Fe-rich cordierite to quartz-sillimanite-garnet aggregate (data of Tracy and Dietsch, 1982); C3 - range of peak metamorphic T and P estimates from Zone VI (data of Tracy et al., 1976 and Robinson et al., 1982a); C4 - estimated P and T of mylonite recrystallization (Robinson et al., 1982a, 1986); C5 - estimated P and T of garnet rim re-equilibration in quartz-sillimanite-garnet aggregate inside zoned cordierite (data of Tracy and Dietsch, 1982).

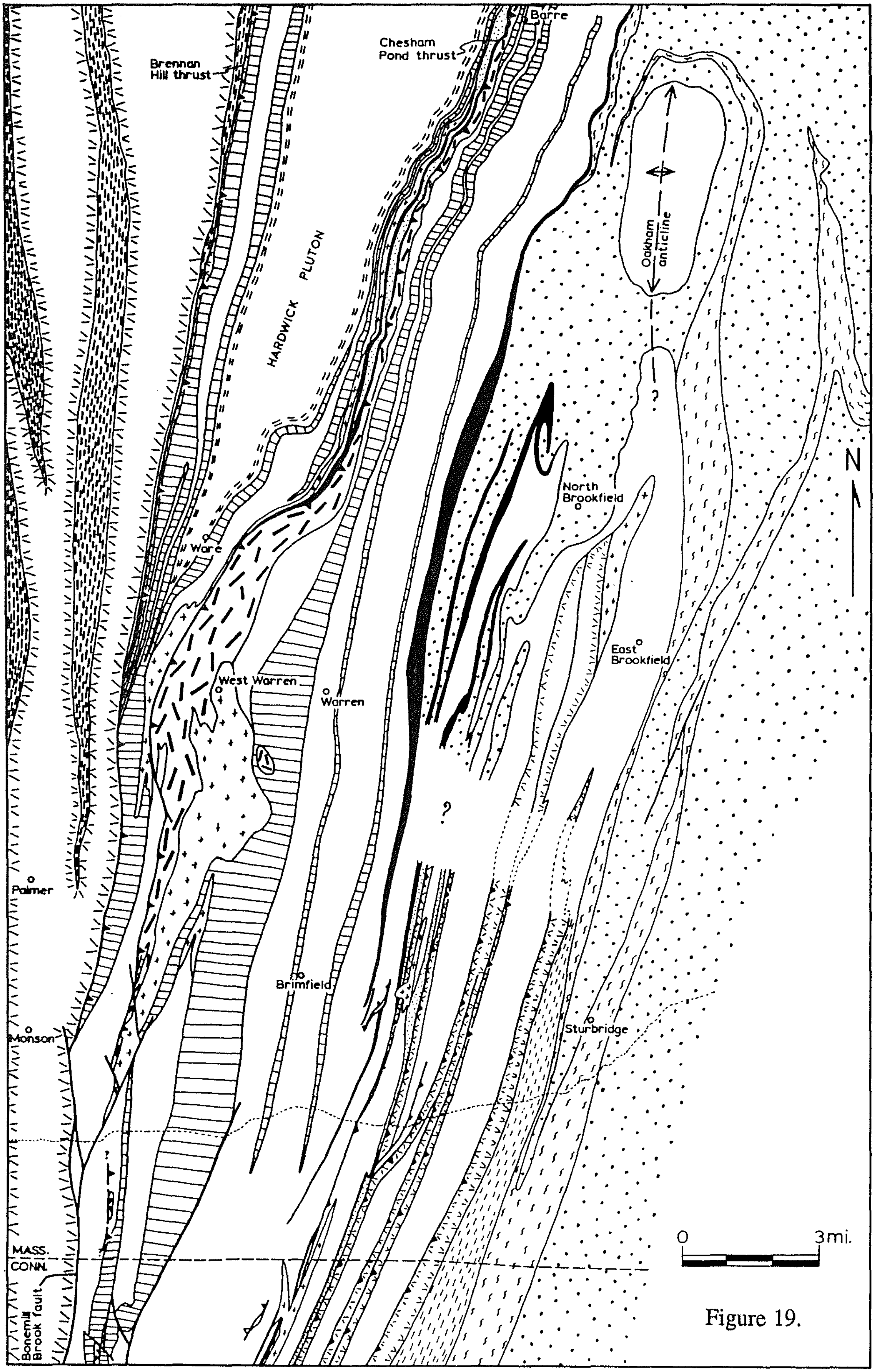
Barre
Chesham
Pond thrust
Brennan
Hill thrust
HARDWICK PLUTON
Oakham
anticline
?
N
North
Brookfield
Ware
West Warren
Warren
East
Brookfield
?
Palmer
Brimfield
Sturbridge
Monson
0
3mi.
MASS.
CONN.
Bonemill
Brook fault

Figure 19.

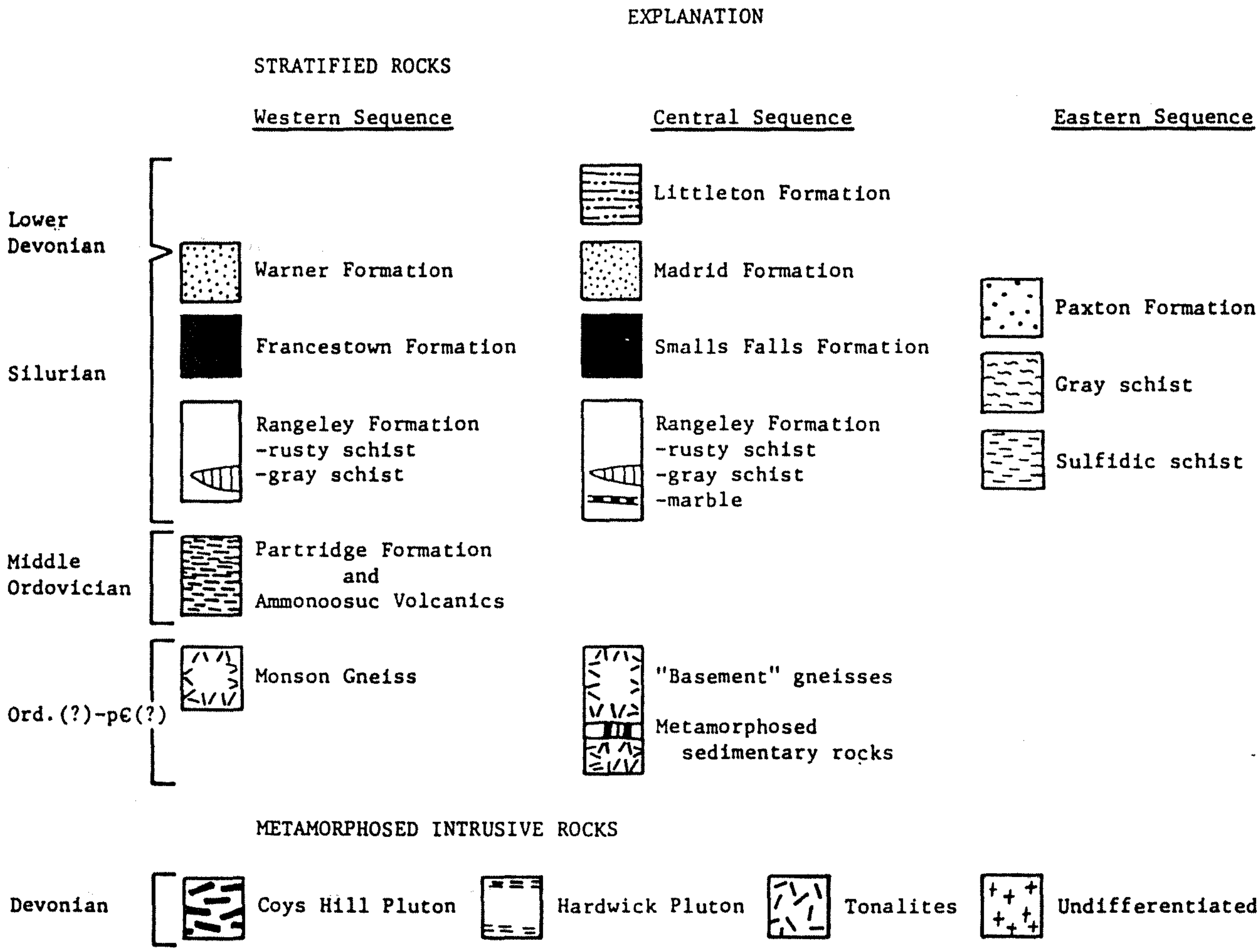

Figure 19. Geologic map of central and south-central Massachusetts modified from Zen et al. (1983). This interpretation shows the proposed southward extensions of the Brennan Hill and Chesham Pond thrusts, and reinterpretations of the Brimfield-Sturbridge area (see also Figure 20). Dashed line south of Monson and Sturbridge shows location of the pipeline trench which was extensively sampled in 1985 and 1987. Rocks west of the Coys Hill plution are referred to as the western sequence in the explanation. Silurian rocks of the western and central sequences are closely related. Units of the eastern sequence are thought to be Silurian, but have not been satisfactorily correlated with units to the west. A "?" indicates where new interpretations in the Brimfield-Sturbridge area have not been reconciled with the geology in the North Brookfield area as shown by Zen et al. (1983). The same patterns are used on Figures 19, 20 and 21.

Figure 20. Geologic map of the Brimfield-Sturbridge area. Symbols are: m=mylonitic tonalite; fm=mylonite zone studied by Finklestein (1987); a=amphibolite; SDu=undifferentiated Silurian and Devonian rocks. These rocks are shown separately in the detailed map of Figure 21. See Figure 19 for location. Patterns same as Figure 19.
Abbreviations for 7-1/2 minute quadrangle names in Massachusetts are: PA=Palmer; AN=Warren; EB=East Brookfield; MO=Monson; WL=Wales; and SO=Southbridge. In Connecticut: SS=Stafford Springs; WF=Westford; EF=Eastford. Faults with black teeth are interpeted as west-directed nappe-stage thrusts overturned to the east (see Figure 21). Open teeth indicate east-directed mylonite zones.

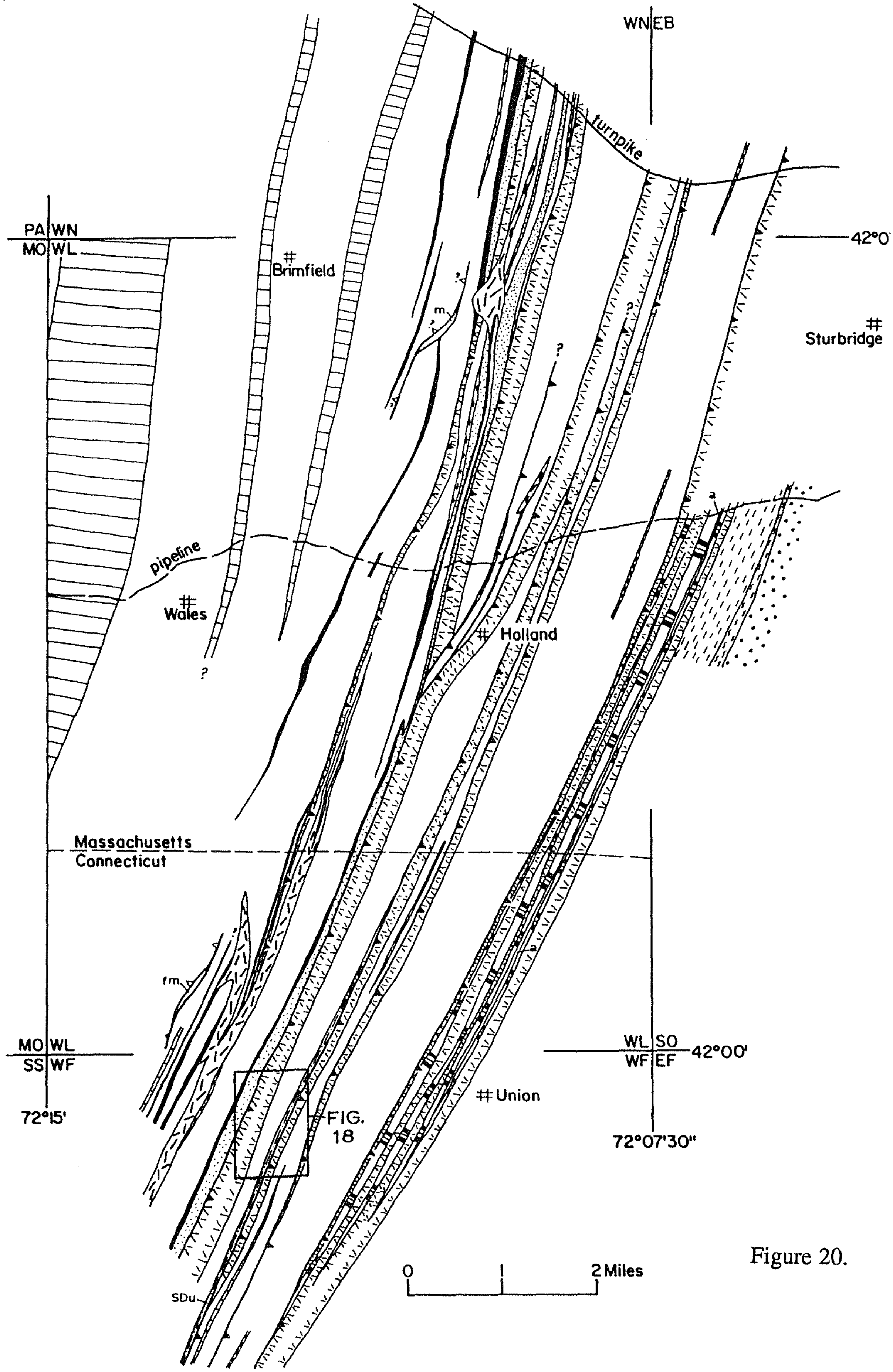
WN EB
turnpike
PA WN
MO WL
42°0'
Brimfield
Sturbridge
m
pipeline
Wales
Holland
a
Massachusetts
Connecticut
fm
MO WL
SS WF
WL SO
WF EF
42°00'
72°15'
Union
FIG. 18
72°07'30"
0
1
2 Miles
SDu

Figure 20.

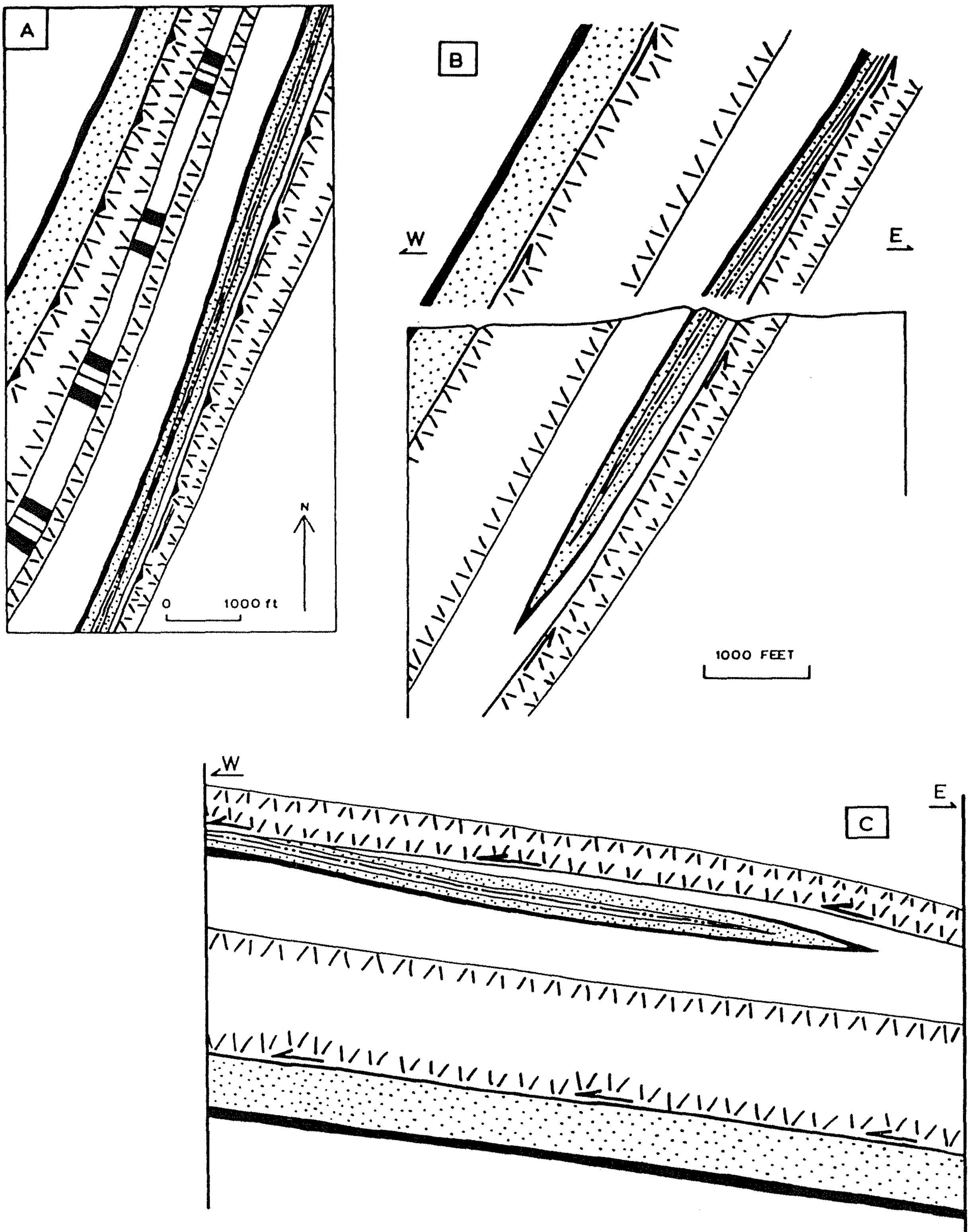

Figure 21. A) Detailed map from the southern part of Figure 20, showing repetition of Silurian units in an isoclinal syncline cut on the east by a fault. B) East-west cross section through the map in A, showing the west-dipping fold axial surface, faults, and downward-facing strata. C) Schematic diagram illustration the interpretation of originally west-directed nappe-stage folds and thrusts, repeating dominantly upright strata. These were subsequently backfolded to the east into their present orientation, shown in B.

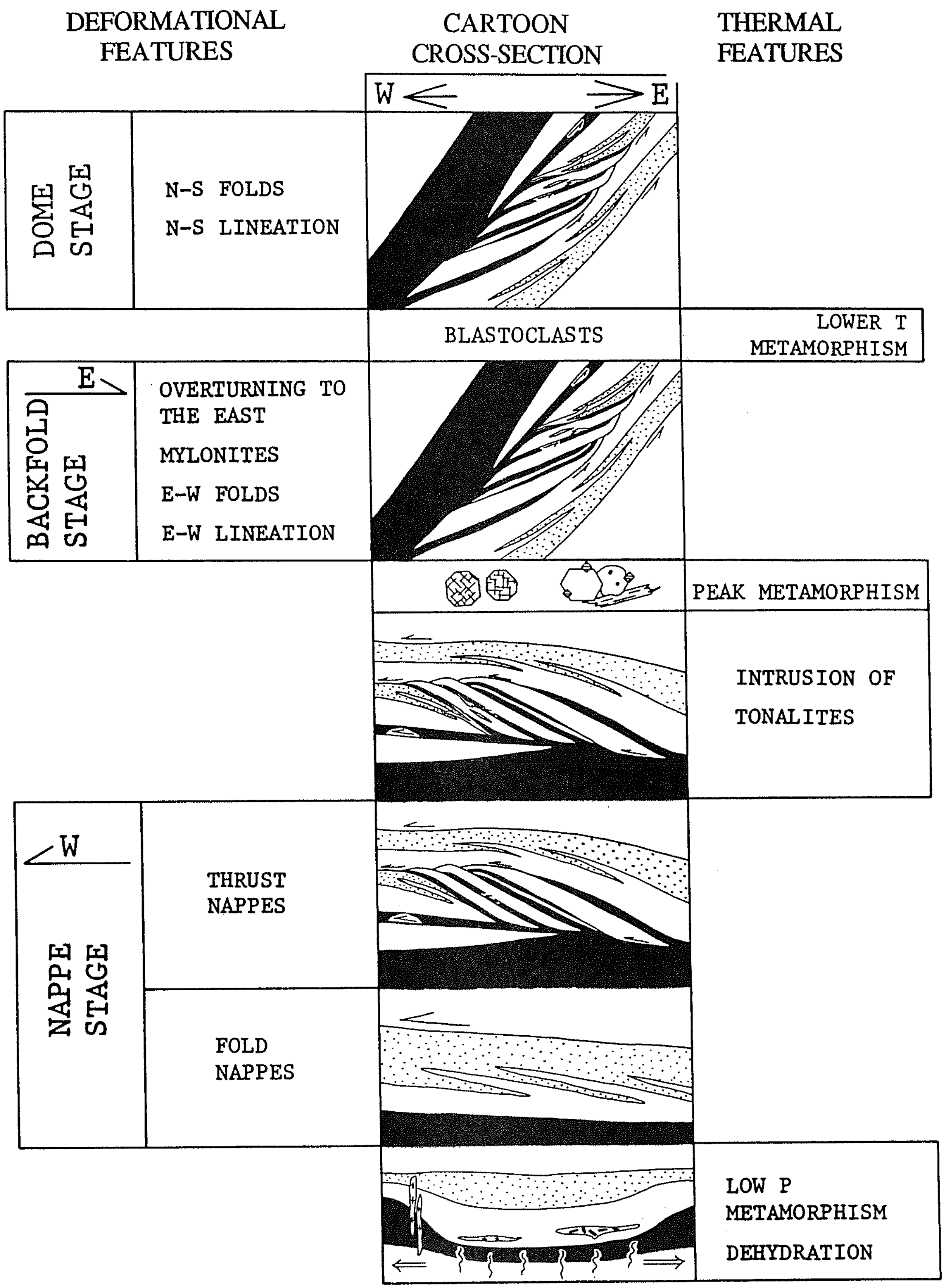

Figure 22. Acadian tectonic model for the Brimfield-Sturbridge area, showing relative ages of deformational and thermal features, with the oldest at the bottom.

Intrusive tonalite bodies are interpreted to crosscut the thrust-nappes at map-scale (Figure 20), demonstrating that they are younger. Because the tonalites were affected by the peak granulite-facies metamorphism, peak metamorphism must have been also younger than the nappe stage features. Structural features which followed peak metamorphism include an easterly trending mineral lineation, isoclinal folds in foliation, and fine-grained mylonites containing porphyroclasts of peak-metamorphic minerals. These mylonites and related structural elements are assigned to the stage of regional backfolding, during which originally west-directed nappe stage features were overturned to the east on a regional scale (Figure 21 and 22). An unusually thick zone of mylonitic tonalite and pelitic schist in the southwest part of the Wales quadrangle (Figure 20) was sampled and studied as a senior undergraduate project by David Finkelstein. He concluded from asymmetric foliation fabrics and deformed mineral grains that motion across the west-dipping mylonite was west-side-up (Finkelstein, 1987), supporting its assignment to the stage of east-directed backfolding. Asymmetric, upright folds with shallow plunges are common in the region. They are considered to be coeval with the final emplacement of the gneiss domes along the Bronson Hill anticlinorium. No significant map-scale features of this age have been identified in south-central Massachusetts.

In contrast to aspects of previous models, including that of Robinson et al. (1986b), the present work suggests that 1) the Silurian rocks of the Merrimack synclinorium rest on continental rather than oceanic crust, 2) the early transport was west-directed rather than east-directed, 3) the metamorphic peak occurred after rather than before the west-directed thrust-nappes, 4) the outcrop-scale mylonites are later than rather than synchronous with the map-scale thrust-nappes, and 5) regional backfolding may have overturned the majority of east-central Massachusetts.

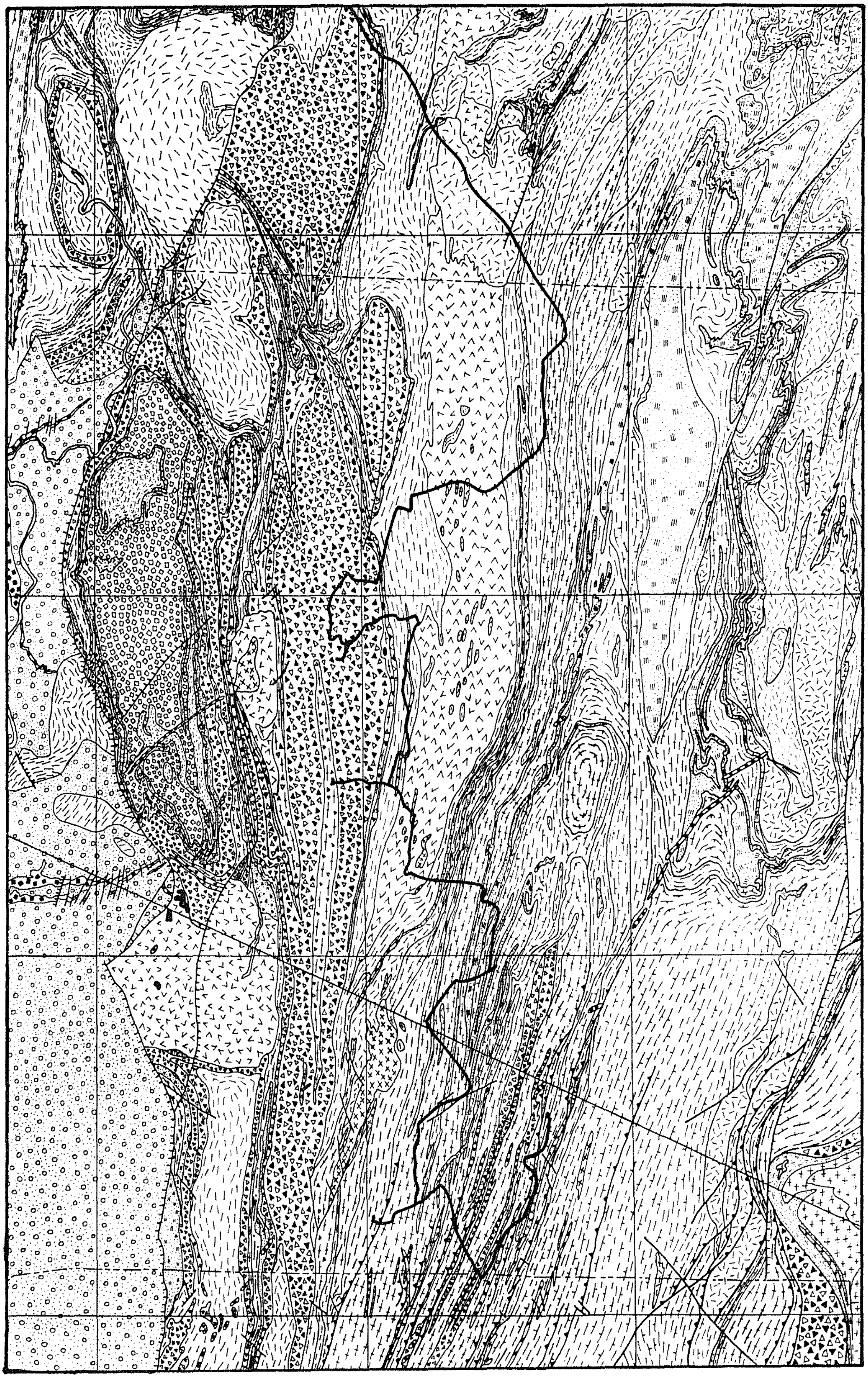

Figure 23. Generalized geologic map showing overall route of field trip on day 4.

ROAD LOG

The entire route of the field trip is shown on Figure 23 on the lithically patterned base used for Continent-Ocean Transect E-1 (J. B. Thompson et al., in review). All of the stop locations are shown with metamorphic isograds on Figure 24, and the detailed route with stops is shown on the geology in Figures 26 and 34.

0.0 Road log begins at junction of Rtes. 9, 10, and 12 in southwestern part of Keene. Proceed east on Rtes. 9 and 12.

0.8 Stop light. Turn right on Rte. 12. Proceed southeast through basin eroded in gneisses in core of Keene dome.

1.6 Rte. 32 turns right. Stay on 12.

5.2 Long road cut on left at East Swanzey. This exposes the full sequence of units of the Keene dome including Swanzey Gneiss, Ammonoosuc Volcanics, Partridge Formation, Clough Quartzite, and Littleton Formation (see detailed description by Robinson et al., 1979). Just east of this exposure we cross the early Acadian Brennan Hill thrust and enter an extensive area of Silurian strata of the Monadnock sequence dominated by the Lower Silurian Rangeley Formation (P.J. Thompson, 1985, 1988).

6.8 Road cuts both sides. Rangeley Formation augen schist. This is Stop 1 of P.J. Thompson (1988).

9.7 Center of Troy, N.H. Views of Mt. Monadnock to left. Stay on Rte. 12.

13.6 Junction of Rtes. 12 and 119, Fitzwilliam, N.H. Stop on Rte. 12. The Fitzwilliam Granite is a late, little deformed muscovite-biotite granite (Shearer, 1983) of probable late Devonian to Carboniferous age (P.J. Thompson, 1985).

19.3 Massachusetts State Line. Enter Winchendon.

22.4 Junction Rtes. 12 and U.S. Rte. 202 in Winchendon. Bear sharp right (SW) on Rte. 202.

23.5 Village of Waterville, Massachusetts!

28.6 Village of Baldwinville. Stay on Rte. 202.

31.9 Junction Rte. 202 and Mass. Rte. 2A. Stay on Rte. 202.

32.4 Junction Rte. 202 and Mass. Rte. 2. Turn left onto entrance ramp for Rte. 2 West.

34.1 Exposures, both sides of tonalites, granite porphyties, and schist inclusions of the Hardwick pluton. First outcrop on right beyond underpass has conspicuous sillimanite pseudomorphs after andalusite.

36.8 Exit onto Rte. 32 South.

37.9 Right off Rte. 32 onto Woodlawn Rd. (Lyon Hill Rd. and Doe Valley Rd. on topographic sheet)

38.9 Sharp left (south) on New Sherborn Rd.

43.0 Stop sign. Junction with Rte. 122. Turn right (west).

44.3 New Salem Town Line.

45.1 Turn left (south) on South Athol Rd., Dana Rd.

46.2 Gate 35, entrance to Quabbin Reservation. Proceed through locked gate. Stops 1, 2, and 3 are being visited by special arrangement under a research permit from Massachusetts Water Resources Commission, Quabbin Division, James Holeva, Acting Superintendant.

I
II
III
IV
V
VI
Pelham Dome
Prescott Complex
Main Body of Monson Gneiss
Hardwick Pluton
Oakham Anticline
Gardner Anticline
Coys Hill Pluton
Warren Plutons
Connecticut Valley Border Fault
Glastonbury Dome
Dbp
Dbd
1
2
3
4
5
6
7
8
9
10
11
0 5 MILES
0 5 KM
MASS.
CONN.

Figure 24. Generalized bedrock geologic map of south-central Massachusetts showing metamorphic zones, location of cross section, Figure 25, and location of field trip stops. No attempt is made to pattern all very small areas. Metamorphic zones are described in the text.

EXPLANATION

Use also for Figures 25, 26, and 34.

Jurassic	Diabase dikes.
Jurassic-Triassic	Conglomerate, sandstone, shale, and basalt.
	Foliated muscovite granite gneiss, biotite gneiss.
	Coys Hill Porphyritic Granite.
	Granodiorite, tonalite.
	Gabbro, diorite.
Lower Devonian	Erving Formation.
	Littleton Formation. Volcanics where separately mapped.
Silurian	Fitch Formation.*
	Clough Quartzite.
	Paxton Formation: Spss-Sulfidic Schist Member
	Sp-Granulite Member
	Spsq-White Schist Member, Spqr-Quartzite-Rusty Schist Member
Middle Ordovician	Partridge Formation. Separately mapped felsic volcanics in different pattern.
	Ammonoosuc Volcanics in Subzone A. Also includes separately mapped mafic volcanics in Partridge Formation in Subzones B and C.
Ordovician	Massive gneiss of plutonic derivation in dome core.
Ordovician? or older.	Layered gneiss, Monson Gneiss, Fourmile Gneiss.
Late Precambrian	Poplar Mountain Gneiss, Mt. Mineral Formation.
	Poplar Mountain Quartzite, Pelham Quartzite.
	Dry Hill Gneiss.
	Mesozoic normal fault, hachures on downthrown side.

* Fitch Formation shown solid black W. of Coys Hill Granite.

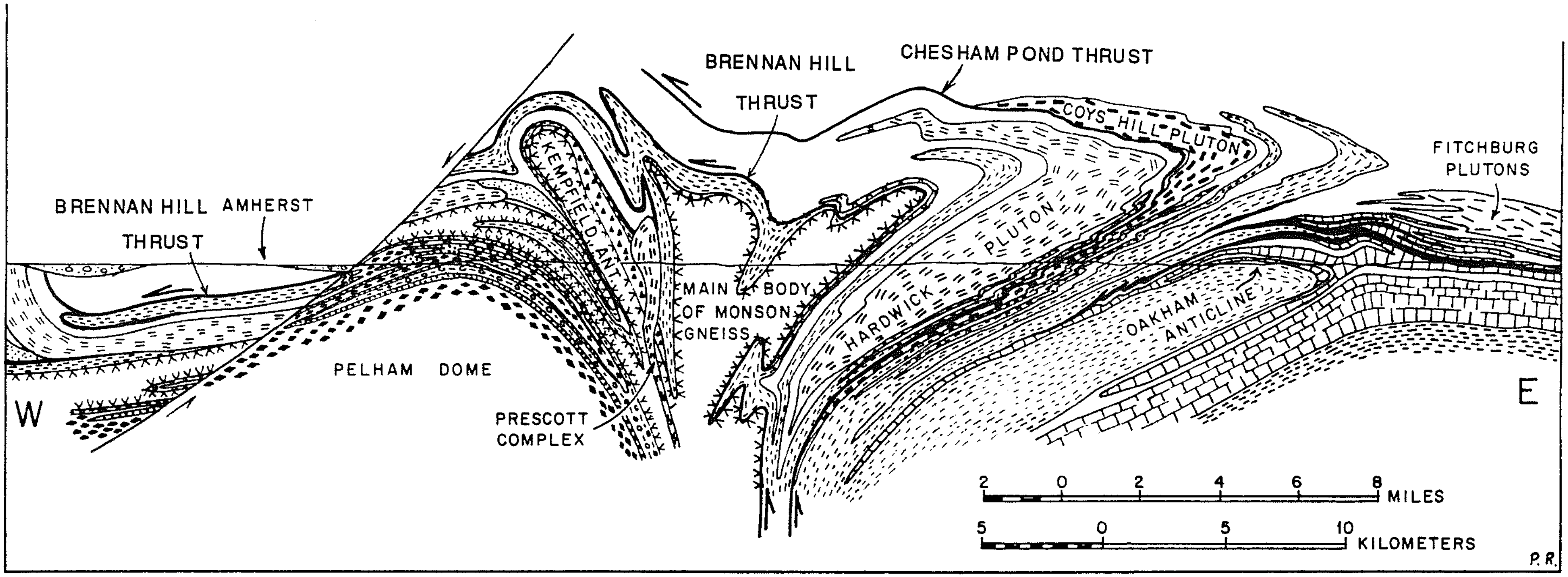

Figure 25. Bedrock geologic cross-section across central Massachusetts. Line of section (although not full length of it) is shown in Figure 24. In the two eastern straight segments the section follows the line of Quabbin Aqueduct Tunnel described in detail and sampled by Fahlquist (1935). Detailed sections on parts of this line are given by Field (1975) and Tucker (1977).

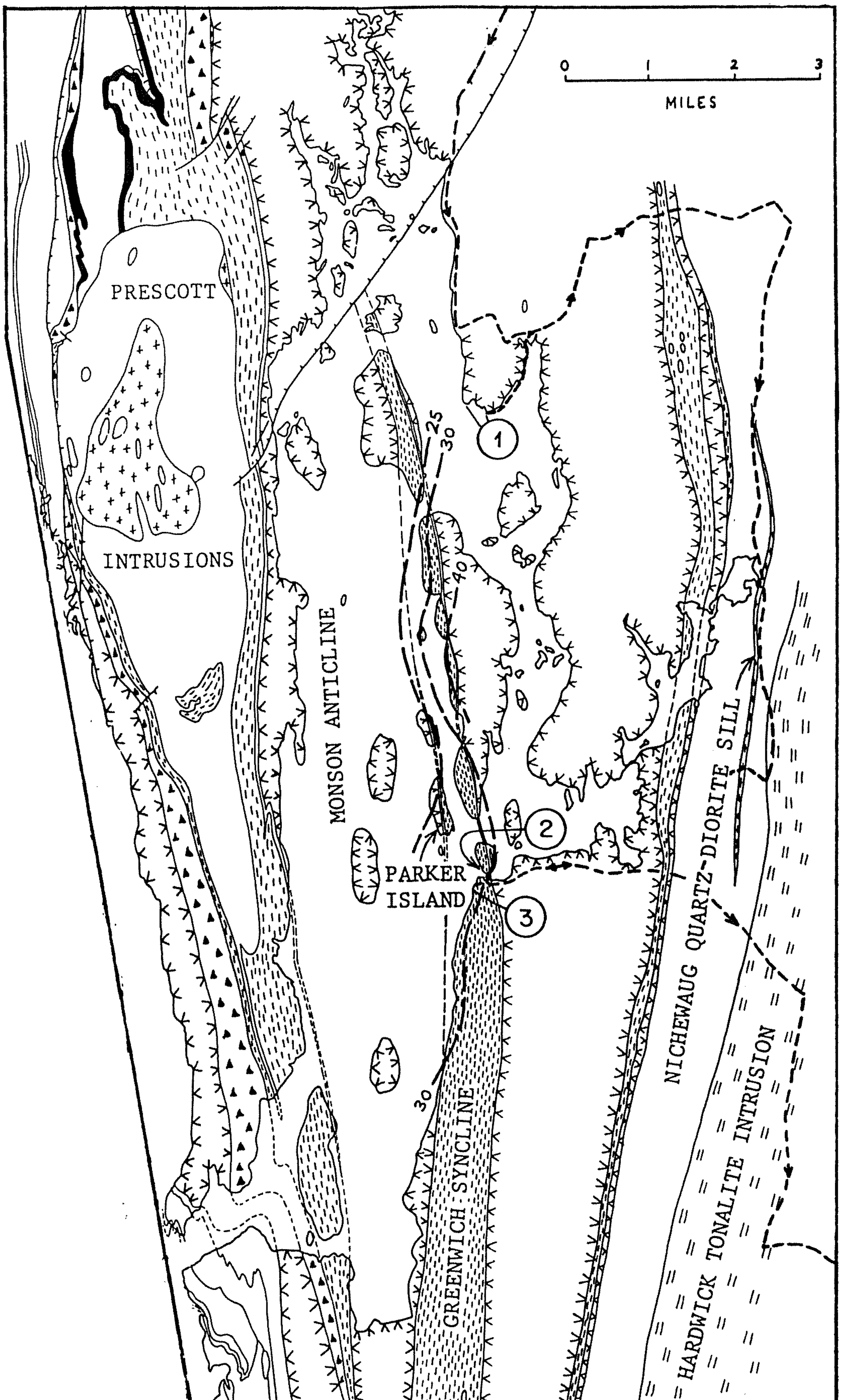

Figure 26. Generalized geologic map of the eastern part of the Quabbin Reservoir area showing location of stops 1, 2, and 3. Dashed lines numbered 25, 30 and 40 are plagioclase composition isopleths for the assemblage quartz-muscovite-sillimanite-orthoclase-plagioclase after Tracy (1978).

Just beyond Gate 35 the road turns onto the abandoned bed of the Boston and Albany, Athol Branch.

46.8 Shore of Quabbin Reservoir, northeastern arm. High hills to right front are Mt. L (to north) and Mt. Zion (to south), held up by Partridge Formation pelitic schists and amphibolites of the Greenwich Syncline studied by Tracy. The boundary between his Zones III and IV passes between these mountains as did the Athol railroad. Low areas and goes on old paved highway.

49.3 Sharp left turn (east) of pavement onto gravel road that leads over south ridge of Soapstone Hill.

50.7 Sharp right turn (southwest) onto old paved road that runs parallel to shore, in part below high water mark.

51.7 Pavement goes into water. Outcrops of Stop 1 lie on shore 0.2-0.5 miles west from this point.

STOP 1: MONSON GNEISS, SOAPSTONE HILL AREA
(60 MINUTES) (Quabbin Reservoir quadrangle)

The outcrops in this area are from metamorphic Zone IV, deep within the main body of Monson Gneiss and far from contacts with the cover rocks. The outcrops are principally composed of gray tonalitic gneisses, amphibolite, hornblendite, and gabbroic anorthosite. First impressions at these outcrops can lead observers to wildly divergent opinions about the origin of the rocks. At one end of the spectrum the well layered character of some outcrops (Figure 27) suggests that the Monson might be predominantly a layered volcanic sequence, possibly even conformable with the overlying Ammonoosuc Volcanics (Naylor, "stratified core gneiss", 1969). At the other end of the spectrum, many of the rocks are coarse-grained, massive, and contain abundant and varied inclusions suggesting an intrusive igneous origin. Marland Billings (pers. comm. 1962) pronounced one outcrop as "typical Oliverian" granite.

With the exception of very rare diopside calc-silicate and quartzite layers, the Monson Gneiss consists of rocks of broadly igneous composition (Robinson, 1963, 1967c; Hollocher, 1987, 1988). The following rock types have been identified:

Coarse- to fine-grained felsic gneisses, generally massive and intrusive-looking, commonly containing angular xenoliths of other rock types:

- Dark-gray biotite-hornblende-andesine tonalite.
- Medium-gray biotite-oligoclase tonalite.
- Light-gray biotite-garnet-oligoclase tonalite.
- Light-gray to pink microcline-hornblende augen gneiss.
- Rare white microcline-albite-quartz-biotite-garnet alaskite gneiss.

Mafic and ultramafic rocks occurring as large and small blocks, commonly included in or interlayered with the more felsic rock types:

- Ultramafic rocks including serpentine and talc-bearing lenses, olivine hornblendite, augite hornblendite, and hornblendite (Tracy et al.,1984). A few of these, particularly small patches of hornblendite, may be residues of partial melting.
- Coarse-grained amphibolites that look like massive gabbros. Some have relict ophitic texture.
- Fine-grained amphibolites commonly containing monomineralic plagioclase or hornblende patches that are probably recrystallized plagioclase or mafic mineral phenocrysts. Some of these rocks cut metamorphic foliation and layering and are clearly dikes.
- Anorthosite and gabbroic anorthosite gneiss, the best examples of which are found as glacial boulders with attached normal Monson Gneiss at Richards Ledges on the

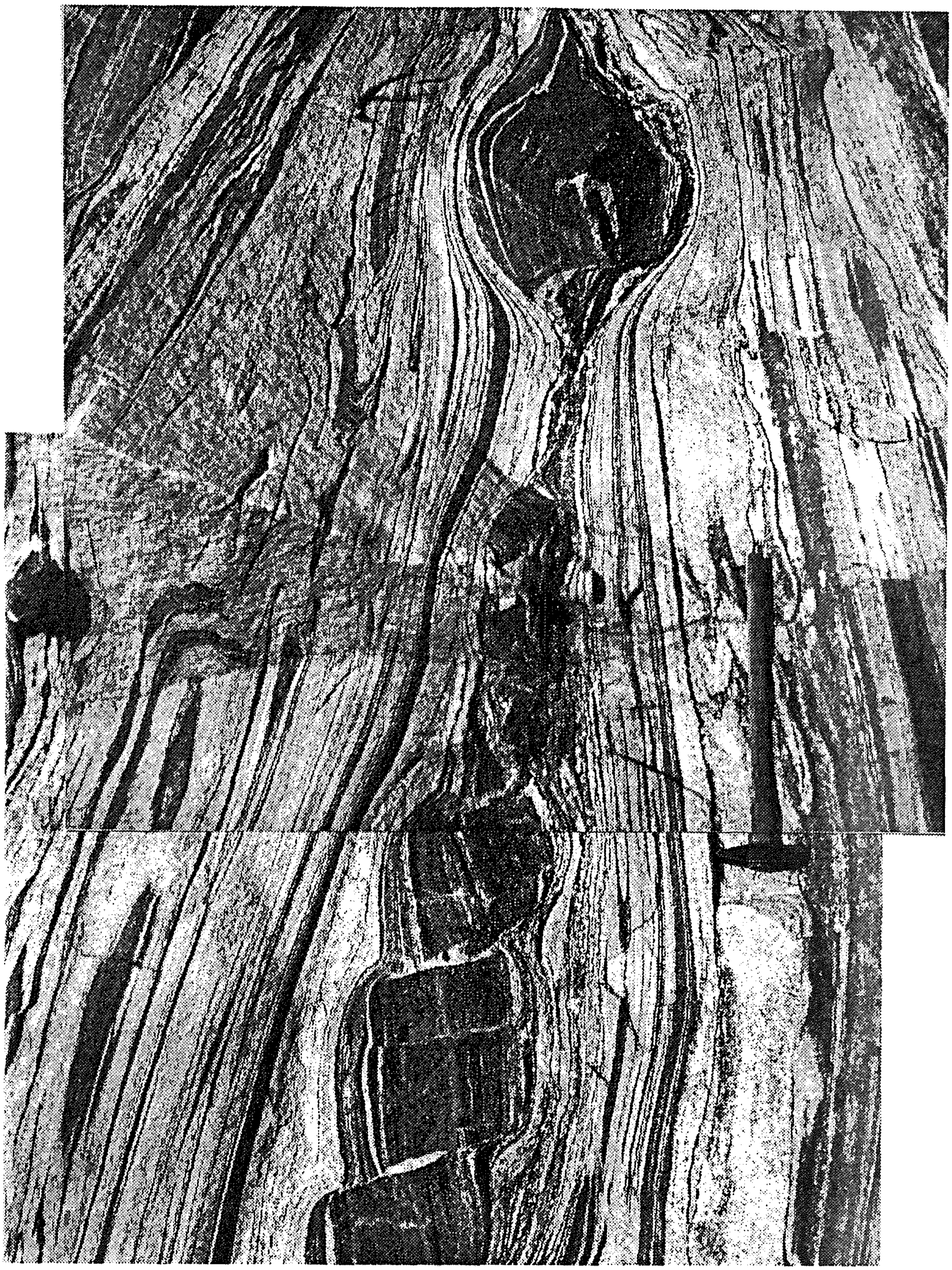

Figure 27. Composite photograph, shore outcrop of Monson Gneiss at Mt. Pomeroy, Quabbin Reservoir, looking down to north. Boudinage of layered amphibolite is a dome-stage feature, as is south-plunging fold to left, where hinge in amphibolite intersects outcrop surface four times. Left of upper boudin, earlier isoclinal fold is contained within slab truncated by shear zone. In upper left, layering in amphibolite truncated by shear zone. Outcrop is better layered and less chaotic than typical exposures at Mt. Pomeroy. Photo 1965 by the late John Haller.

Quabbin shore. Gabbroic anorthosite gneiss also occurs as layers in the Soapstone Hill area, and is identified by its subtle bluish-gray color, best seen on cloudy days.
Composite mafic and intermediate blocks with gabbro, various amphibolites,hornblendites, and biotite gneiss all enclosed in coarse-grained felsic rock.
Rare hornblende-gedrite-garnet-plagioclase-biotite rock.
Well stratified amphibolite that looks most like metamorphosed layered mafic tuffs.

Felsic rocks that occur as small intrusions that are probably or possibly synchronous with Acadian metamorphism:
Very coarse-grained, white biotite-tonalite pegmatite.
Medium-grained, white to pink biotite-garnet tonalite gneiss ("spaghetti rock") similar in many ways to migmatitic layers in adjacent cover rocks.
Medium-grained, white to pink biotite-hornblende tonalite gneiss ("hornblende spaghetti rock").
Fine-grained, medium-gray biotite-epidote tonalite.

The Monson Gneiss outcrop on Parker Island (Figures 28, 29, 30) is a structural "rosetta stone" for the Quabbin region (Robinson 1967b). It lies only 150 feet below the base of the cover sequence which is now known to consist of 150 feet of Ammonoosuc Volcanics (Schumacher, 1983) overlain by Partridge Formation. The outcrop contains about 70% of the known exposures of diopside calc-silicate rock and quartzite in the Monson, which provide useful markers for local structural study. The earliest structural "event" definitely recorded in the outcrop is the intrusion of an amphibolite dike "X", now highly folded, that cuts across a diopside calc-silicate layer in the southern part of the map area. All other obvious structural features can reasonably be ascribed to the Acadian orogeny. The earliest of these is a dominant foliation that appears to have formed parallel to axial planes of early isolinal folds. These folds cannot be seen in Figure 28, but have been mapped at a more detailed scale of 1' = 1" and the hinge of one in quartzite is just visible 1" from the top and 2" from the left margin of Figure 29. This foliation was then folded by vertically plunging, asymmetric folds with right-hand shear sense that dominate the outcrop (Figure29, left; and Figure 30, lower left). In nearby outcrops of the Partridge Formation, migmatitic melt segregations in schist that formed near the peak of Acadian metamophrism had already solidified and behaved as relatively competent layers during this same second-stage folding. Shortly after, coarse biotite pegmatite layers were intruded along the axial planes of the second-phase folds as shown by points marked "Y" in Figure 28. Barbara Barriero of Dartmouth College (personal communication, 1987) has obtained a preliminary Pb/Pb age on zircon from this pegmatite of 370 m.y. The sample shows evidence of some lead loss, but no inheritance, and an extrapolated intercept on concordia suggests a primary crystallization age close to 400 m.y. These pegmatites and all previous structural features were then cut by a series of ductile shear zones, now zones of well recrystallized gneiss, with apparent left-lateral shear sense, best illustrated by mapped offsets up to 30 feet in the western part of Figure 28 and shown close up in Figure 30. In the last deformation all previous structural features, including shear zones, were folded about sub-horizontal dome-stage axes and had the prominent dome-stage mineral lineation superimposed on them. This phase is best illustrated by the doubly plunging "bathtub" fold in the upper left of Figure 28 and is the most prominent structural phase in most outcrops of Monson Gneiss throughout the region. Although correlation of this local structural sequence to the Acadian stages of regional deformation is not easy, tentative correlations suggest that the earliest folds and foliation belong to the nappe stage, the second folds and the ductile shear zones to the backfold stage when the main and Tully bodies of the Monson Gneiss moved into their present strange positions, and the third folds to the dome stage.

Recent field, petrologic, and geochemical studies by Hollocher have shown this to be a complex of metamorphosed intrusive rock types including tonalite, diorite, gabbro, gabbroic anorthosite, and diabase porphyry with only a miniscule supracrustal component. If the Monson Gneiss is part of a plutonic complex, as implied here, then the well layered character of many

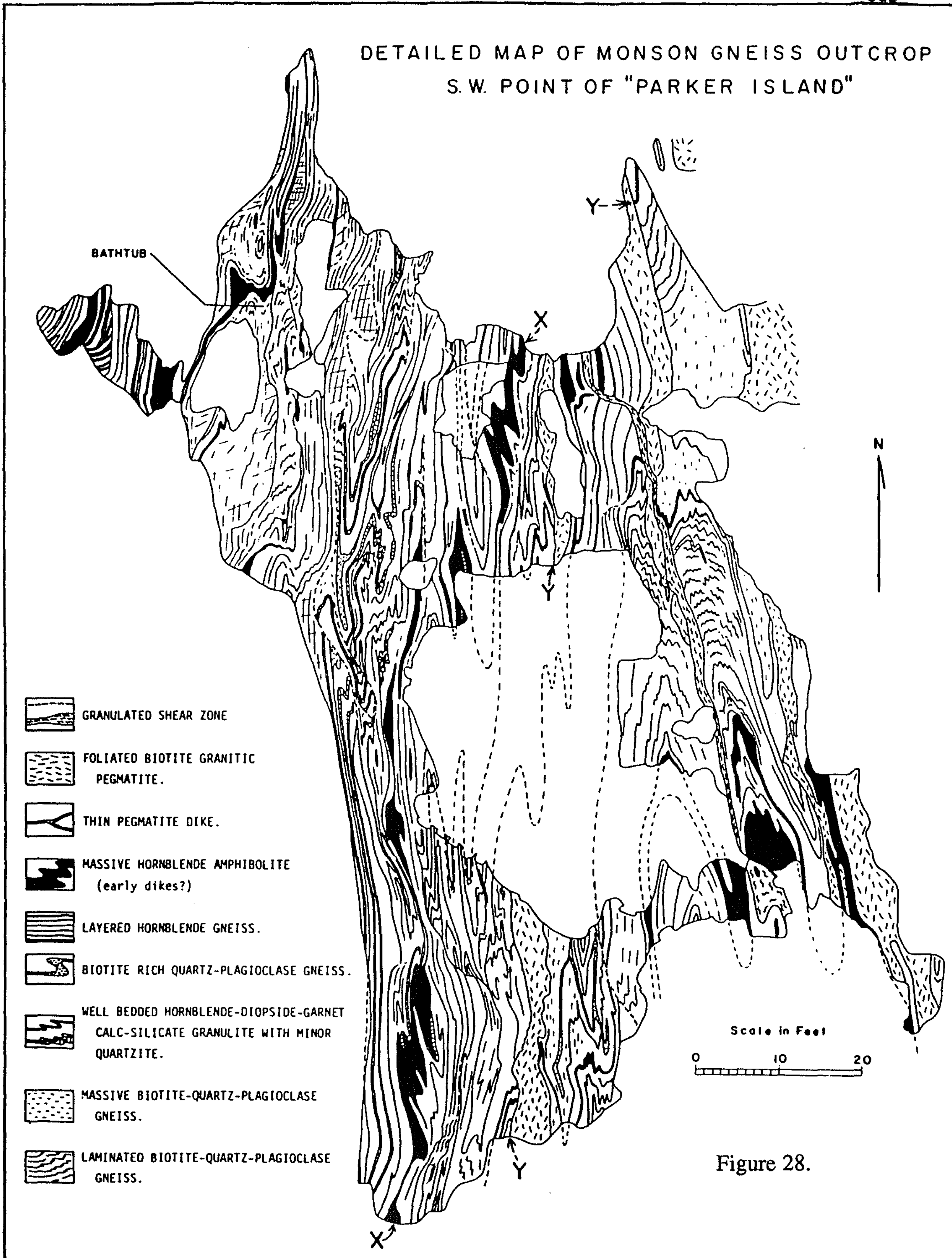
DETAILED MAP OF MONSON GNEISS OUTCROP
S.W. POINT OF "PARKER ISLAND"
BATHTUB
X
Y
N
GRANULATED SHEAR ZONE
FOLIATED BIOTITE GRANITIC PEGMATITE.
THIN PEGMATITE DIKE.
MASSIVE HORNBLENDE AMPHIBOLITE (early dikes?)
LAYERED HORNBLENDE GNEISS.
BIOTITE RICH QUARTZ-PLAGIOCLASE GNEISS.
WELL BEDDED HORNBLENDE-DIOPSIDE-GARNET CALC-SILICATE GRANULITE WITH MINOR QUARTZITE.
MASSIVE BIOTITE-QUARTZ-PLAGIOCLASE GNEISS.
LAMINATED BIOTITE-QUARTZ-PLAGIOCLASE GNEISS.
Scale in Feet
0
10
20

Figure 28.

Figure 29. Close up view of Figure 28. Looking northwest. Foreground, boudinaged diopside calc-silicate bed. To left is hinge region of second phase fold within the same sequence of calc-silicate and quartzite. Tiny first phase fold hinge is just visible in quartzite about 1.3" from the top and 2.2" from the left margin.

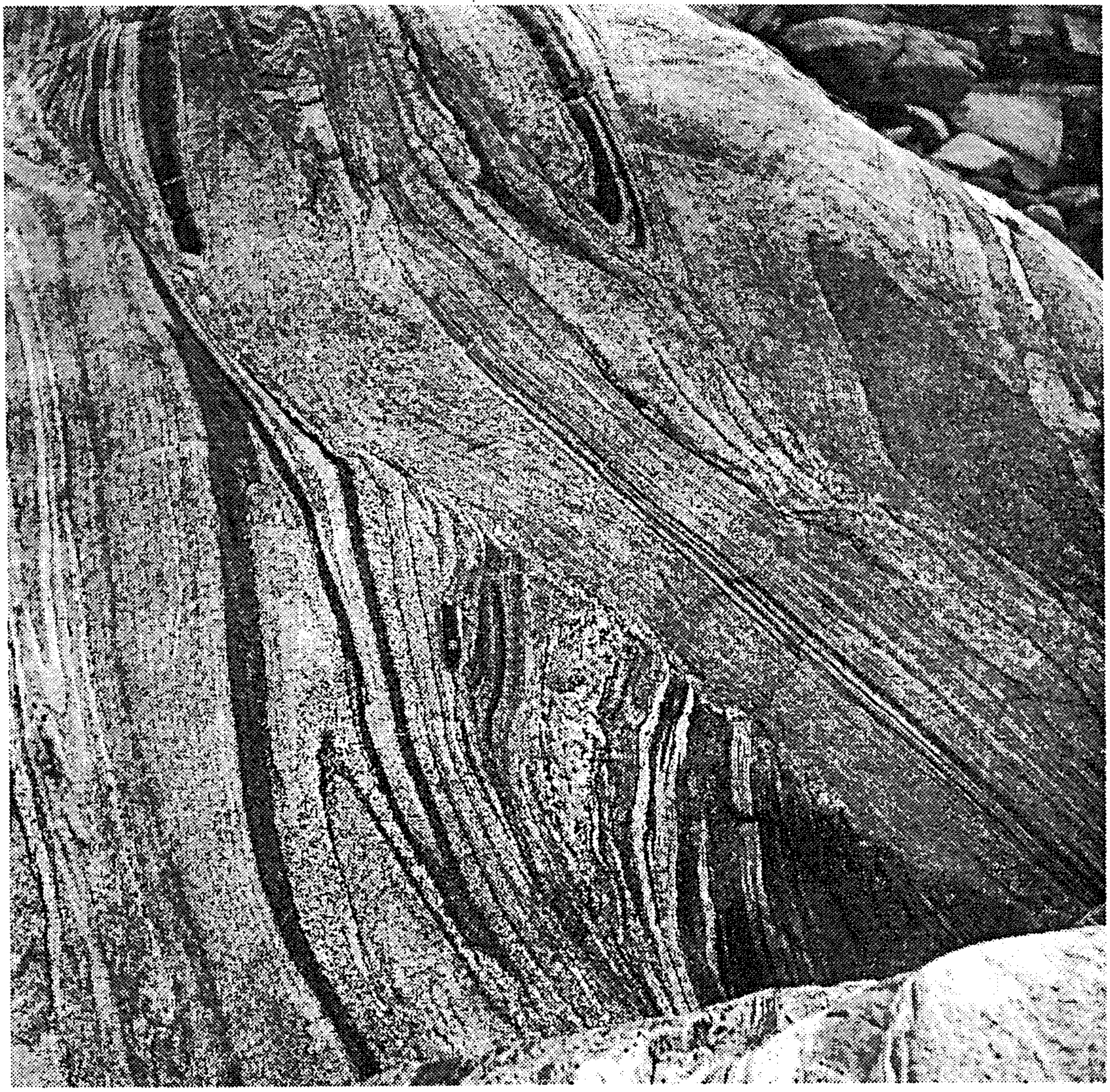

Figure 30. Close up view of Figure 28. Looking southwest across two mapped ductile shear zones. Prominent second-phase folds in the amphibolite are truncated by shear zone in foreground. Massive dark gray biotite-quartz-plagioclase gneiss to right, beyond second shear zone is key marker unit in the western part of Figure 28, and has been successfully sampled. Photographs, 1965 by the late John Haller.

outcrops would be attributed to severe deformation, either during the Acadian or earlier. It is possible that some of the layered gneisses really are volcanics, whereas the majority appear to be highly strained plutonic rocks. The rocks are petrologically and geochemically distinct from the metamorphosed volcanics in the physically overlying Ammonoosuc Volcanics and Partridge Formation, but they closely resemble rocks in other orogens described as in the roots of a calc-alkaline island arc (Barker and Arth, 1984; Hamilton 1988).

Recent precise radiometric dating of a variety of these rocks by R.D. Tucker at the Royal Ontario Museum has shown an age range from 454 to 443 m.y. (+3/-2 on six best samples, +9/-3 and +11/-8 on two others). His dates of 453 +/-2 on a peraluminous rhyolite from the Ammonoosuc and 449+3/-2 on a peraluminous rhyolite from the Partridge indicate a temporal overlap between the basement and its immediately overlying cover. These results conflict with the unconformity hypothesis long proposed by Robinson on the basis with fieldrelations and a basal conglomerate that is in turn in conflict with the intrusive hypothesis supported by Zartman and Leo (1984). An easy resolution is not in sight, but elsewhere we have proposed that the contact could be a detachment fault that developed late in the history of the volcanic arc, bringing volcanic cover in contact with intrusive basement.

The Monson Gneiss outcrops contain evidence of Devonian igneous activity in the form of gray biotite tonalite and amphibolite dikes. These arebelieved to be middle Acadian because they truncate early Acadian folds and metamorphic foliation, but are themselves deformed by late Acadian structural features (Figure 31). Such a dike in the Keene gneiss dome has been dated at 381+3/-2, but a structurally similar dike exposed at Stop 1 (Figure 32) has not yet yielded an acceptible date due to substantial zircon inheritance.

The Monson Gneiss has been intruded by later leucocratic melts that are thought to have been Acadian. The ultimate source of these melts, based on geochemical modeling, proximity, and mineralogy, is the Monson Gneiss itself (Hollocher, 1988). Some hornblendites are closely associated with some of the melts and it is possible that the hornblendite may be restite material left after melting of amphibolite. It is more likely, however, that the melts came from elsewhere and filled fractures in the brittle hornblendite during metamorphism. It is notable that all of these Acadian partial melts are white or pink, whereas dikes at other localities that are more clearly derived from amphibolite are dark-gray.

Return northeast along old paved road.

52.7 Junction where we came from left earlier. Proceed straight and cross culvert over Fever Brook.

52.8 Junction bear left.

53.3 Gate 37. Exit from Quabbin Reservation.

54.1 Four corners - turn right (east).

56.7 Junction Rte. 122. Turn right (southeast).

57.1 Junction Rte. 122 and Rte. 32A. Turn right (south) on Rte. 32A.

59.4 Village of Nichewaug. A nappe has not yet been named for this but the name has been applied to an augite diorite sill (Shearer, 1983) on which a metamorphosed contact aureole has been identified (Shearer and Robinson, 1980).

63.7 Sharp right (northwest) turn on Breen Road which loops west then southwest becoming Mellon Road.

65.8 T junction with Greenwich Road. Turn right (west).

66.3 Gate 43, Quabbin Reservation. Proceed straight ahead through locked wooden bar, and

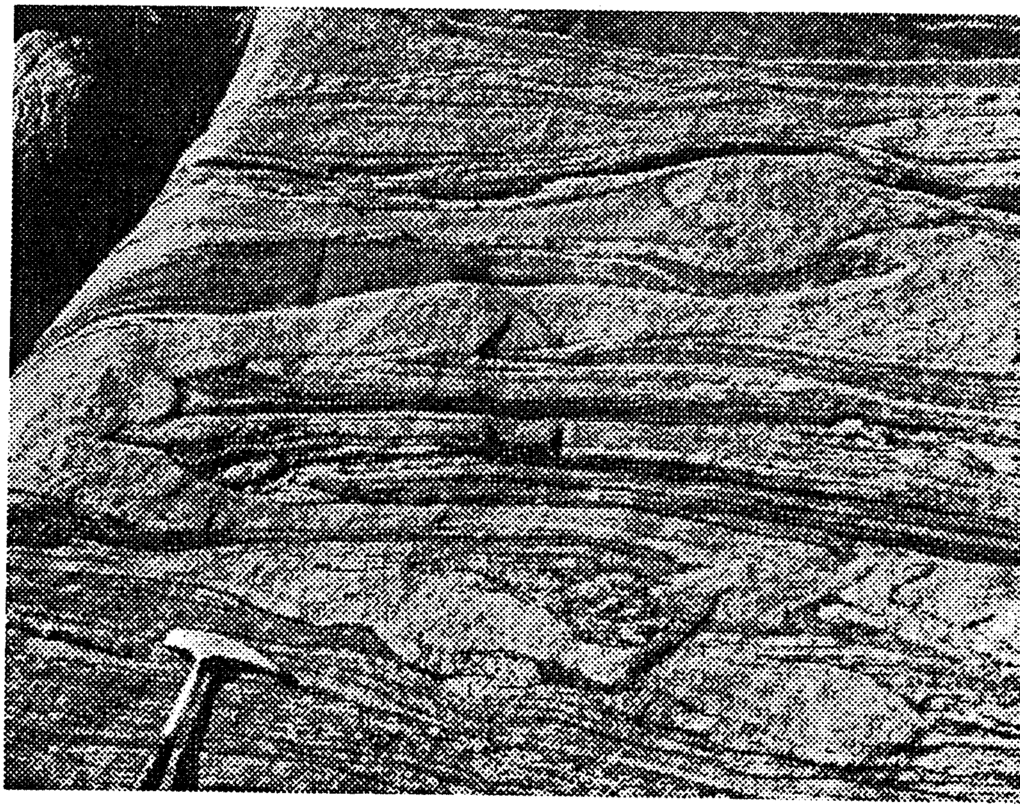

Figure 31. Close-up of exposure of Monson Gneiss at stop 1 showing folded layer of leucogneiss that truncates older foliation. The leucogneiss is believed to have been an Acadian partial melt.

Figure 32. Features of Monson Gneiss on point southwest of Soapstone Hill, Quabbin Reservoir quadrangle. Gray tonalite dike with marginal pegmatite cuts earlier gneissic layering and foliation, as well as the axial surface of early folds in foliation that probably correlate with the second phase of folding at Parker Island. The tonalite dike is itself folded about south (left) plunging folds correlated to the dome stage and has an axial plane foliation both in the matrix biotite and in the orientation of small leucogneiss xenoliths. If these features are properly interpreted, it is difficult to escape the conclusion that the tonalite dike was intruded during Acadian deformation.

west on road to Intake Works.

68.3 Turn right (north) onto Baffle Dam Road and drive across south Baffle Dam.

68.5 Park on left and walk northwest along southwest shore of Baffle Dam Island.

STOP 2: PELITIC MIGMATITES, BAFFLE DAM ISLAND, QUABBIN RESERVOIR (40 MINUTES) (Winsor Dam quadrangle)

This stop, (Figure 26) provides the most easily accessible example of the development of stromatic migmatites in pelitic schists of the Ordovician Partridge Formation in the sillimanite-muscovite-orthoclase zone (IV) of Acadian metamorphism. More spectacular and better studied exposures occur on the south end of Parker Island (Tracy, 1978) but are exposed only at times of low water, and are then only accessible by boat. As seen on the island, the stromatic migmatites must have formed early in the sequence of deformation and metamorphism because they have been isoclinally folded. The outcrops at Stop 2 are within a few hundred feet of the Monson Gneiss along the east margin of the Greenwich syncline and no more than a few feet, if any, of the Ammonoosuc Volcanics occur at the contact. Subordinate rock types include massive and layered amphibolites, and well bedded garnet-rich quartzose granulite, in this case probably a calcareous variety of the rock known locally as coticule.

Point-counted modes for pelitic schists in this vicinity reported in Tracy (1978) indicate that most samples are dominated by quartz, biotite and sodic plagioclase (An 20-32) with lesser amounts of muscovite, sillimanite, garnet and K-feldspar in addition to traces of graphite, pyrrhotite and Fe-Ti oxides. Most of the schist samples contain the quasi-univariant mineral assemblage with both muscovite and K-feldspar, as was also noted in northwestern Maine by Evans and Guidotti (1966). In contrast to Maine, however, Tracy (1978) was able to demonstrate that An content of plagioclase controlled the distribution of the muscovite + K-feldspar assemblages within the narrow (several km) zone separating sill + mus assemblages from sill + Ksp assemblages (Figure 26).

Garnets in the Baffle Dam vicinity are notably more magnesian than lower grade garnets (pyrope contents up to 15-20 mol%), indicating that the presence of K-feldspar in these schists marks the onset of the biotite dehydration reaction to form garnet and K-feldspar. Water liberated by this reaction, as well as that produced through muscovite breakdown, probably allowed melting and was dissolved in resultant melts. The relatively limited volume of melting produced in the pelitic rocks, along with the lack of exhaustion of critical melting phases in restitic zones adjacent to migmatites, indicate that melting was interrupted in progress here. We believe the critical factor in limiting the melting was a limitation on the amount of aqueous fluid locally present. At the temperatures documented for these outcrops (650-675°C at about 6-7 kbar), only water-saturated, or nearly water-saturated melting could have occurred. The extent of the melting therefore probably depended principally upon the availability of water from the break-down of hydrates locally. Migmatite formation at this locality can thus be taken as evidence that aqueous fluids were not abundant during high grade metamorphism here, and in fact were probably quite limited.

The structural history of the stromatic migmatites is best shown at Parker Island though many essential features are visible at this stop. On the island, the migmatite leucosomes appear to have behaved as competent layers relative to the enclosing schist during formation of the steeply plunging second-generation isoclinal folds. The lack of evidence of crushing or granulation in the coarse leucosomes suggests folding of molten or only partly solidified material, barring extensive post-folding annealing. In the Parker Island outcrops which contain the folded stromatic migmatite, the subhorizontal dome-stage sillimanite lineation is virtually at right angles to the axes of vertical second-generation folds. Another common feature of these outcrops is boudinage with necklines roughly normal to, and clearly deforming, dome-stage mineral lineations. It is common in the vicinity of boudin necklines to find localized pockets of various melt products as well as massive hydrothermal quartz. Thus all evidence points to sillimanite-orthoclase metamorphism and migmatite formation before the second folding, but also to continuation of high-temperature conditions appropriate to sillimanite recrystallization and continued partial melting during the dome-stage third folding.

Return across south Baffle Dam to road to Intake Works.

68.7 Turn right (west) toward Intake Works.

70.0 Intake Works. Follow trail about 200 feet south to shore exposures.

STOP 3: AMPHIBOLITES AT QUABBIN INTAKE WORKS

The rocks exposed at this stop lie in the Ordovician Partridge Formation in the Greenwhich syncline, in the upper part of the sillimanite-muscovite- orthoclase zone (Zone IV). The low outcrops just south of the intake works have three rock types of interest: hornblende-cummingtonite-plagioclase-quartz amphibolite, hornblende-gedrite-plagioclase amphibolite, and hornblende-gedrite ultramafic rock (scattered boulders only).

The northwesternmost outcrop is a very mafic hornblende-gedrite amphibolite that is the most mafic rock from the Partridge volcanics analyzed by Hollocher (1985). This rock is Si-poor, and rich in MgO, Ni, and Cr, and has very calcic plagioclase (An85.4). The hornblende-gedrite ultramafic boulders that are scattered about have recently been analyzed, and are chemically similar to the amphibolite. The mineralogical, textural, and geochemical similarities between the two gedrite-bearing rocks indicate that they may be parts of the same differentiated picritic lava flow. The mineral compositions for the hornblende-gedrite northwestern outcrop and the hornblende-cummingtonite southeastern outcrop are as follows:

	Mg/(Mg+Fe): Hornblende	Cummingtonite	Gedrite	Biotite	Anorthite: Plagioclase
NW outcrop	0.659		0.645	0.726	85.4
SE outcrop	0.616	0.599		0.689	40.7

The southeastern outcrop is largely composed of cummingtonite and quartz-bearing hornblende amphibolite. This rock probably represents a more silica- and iron-rich tholeiitic flow that may have undergone some subaerial weathering, as suggested by the abundant magnetite that is rare in the Partridge. This outcrop also contains coarse-grained tonalitic partial melt zones that have cummingtonite crystals up to 4 cm long. The abundance of cummingtonite in the melt segregations suggests that it was produced during the melting reaction. If comments above about migmatite grain size being related to fluid-absent vs. fluid-present melting, then the coarse grains in these leucocratic segregations suggest that the melting reaction in this rock was a fluid-present reaction. A possible melting reaction to produce these cummingtonite migmatite zones from the enclosing hornblende amphibolite is:

Hornblende + plagioclase + quartz + H_2O = cummingtonite + magnetite + hydrous tonalitic melt.

Figure 33 shows the approximate location of the rocks at this outcrop in T-XMg space with respect to reaction loops that define the mineral assemblages in rocks with compositions similar to tholeiitic basalt.

Leave Intake Works. Return east to Gate 43.

72.3 Gate 43 - Pass through locked barway. Turn left (north) on Hell Huddle Rd. to picnic area for lunch (40 minutes) and public toilet at fishing area about 1.5 miles to the north. Resume road log at Gate 43 after lunch.

72.3 Head east on Greenwich Road.

75.0 Center of Hardwick. Turn right on Rte. 32A.

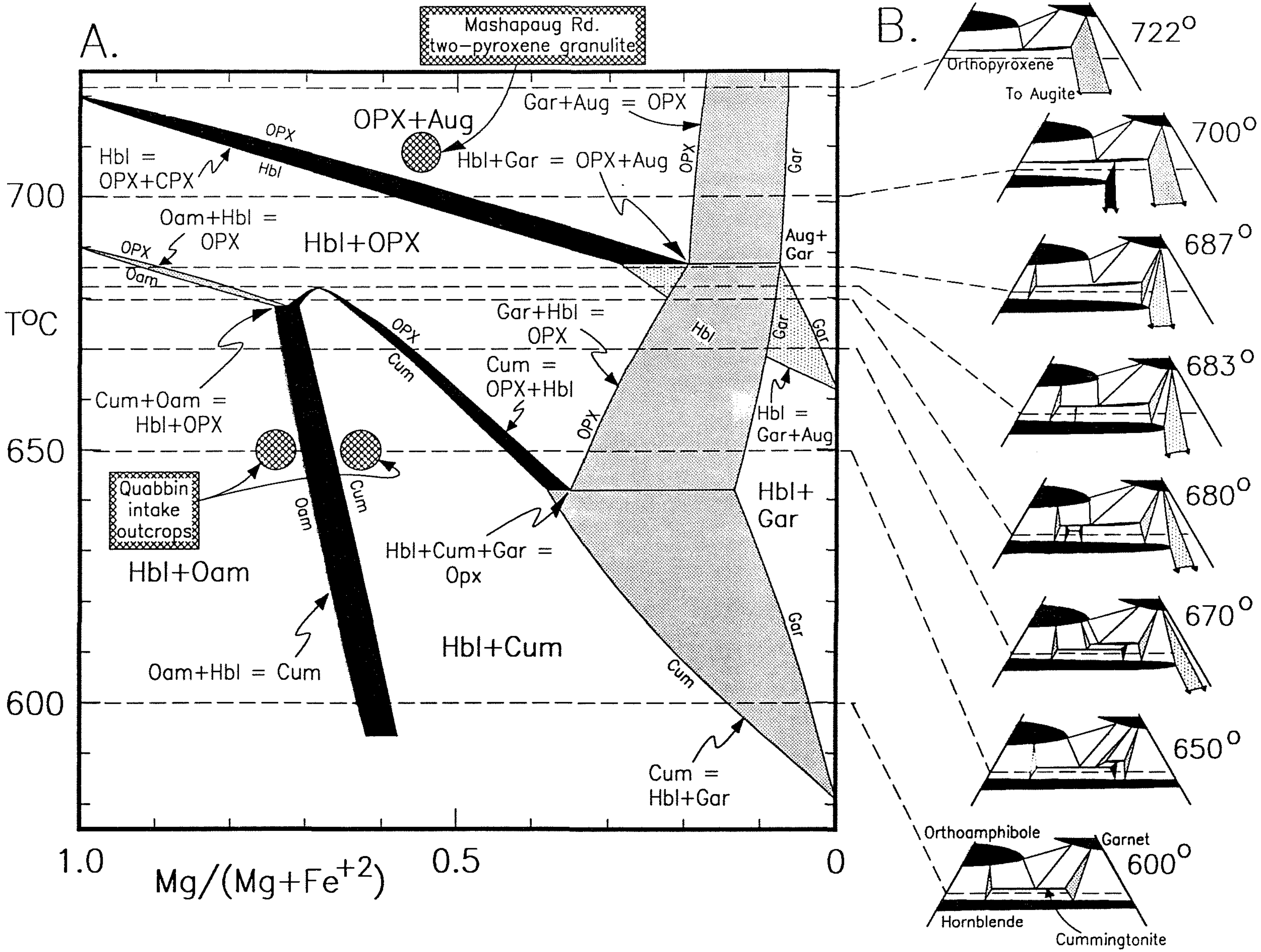

Figure 33. T-XMg diagram showing the phase relations in subaluminous mafic rocks. The black three-phase reaction loops are relatively well constrained by mineral compositions and estimated temperatures, and the gray loops are in part conjectural, and are idealized in the Fe-rich part of the diagram to ignore the stabilizing effects of Ca and Mn on garnet. Two-phase fields and three- and four-phase reactions are labeled with their appropriate assemblages. The sides of the loops are labeled with the minerals whose compositions actually constrain the loops. The approximate locations of rock bulk compositions for the Quabbin Intake and Mashapaug Road outcrops are shown. Adapted from Hollocher (1985).

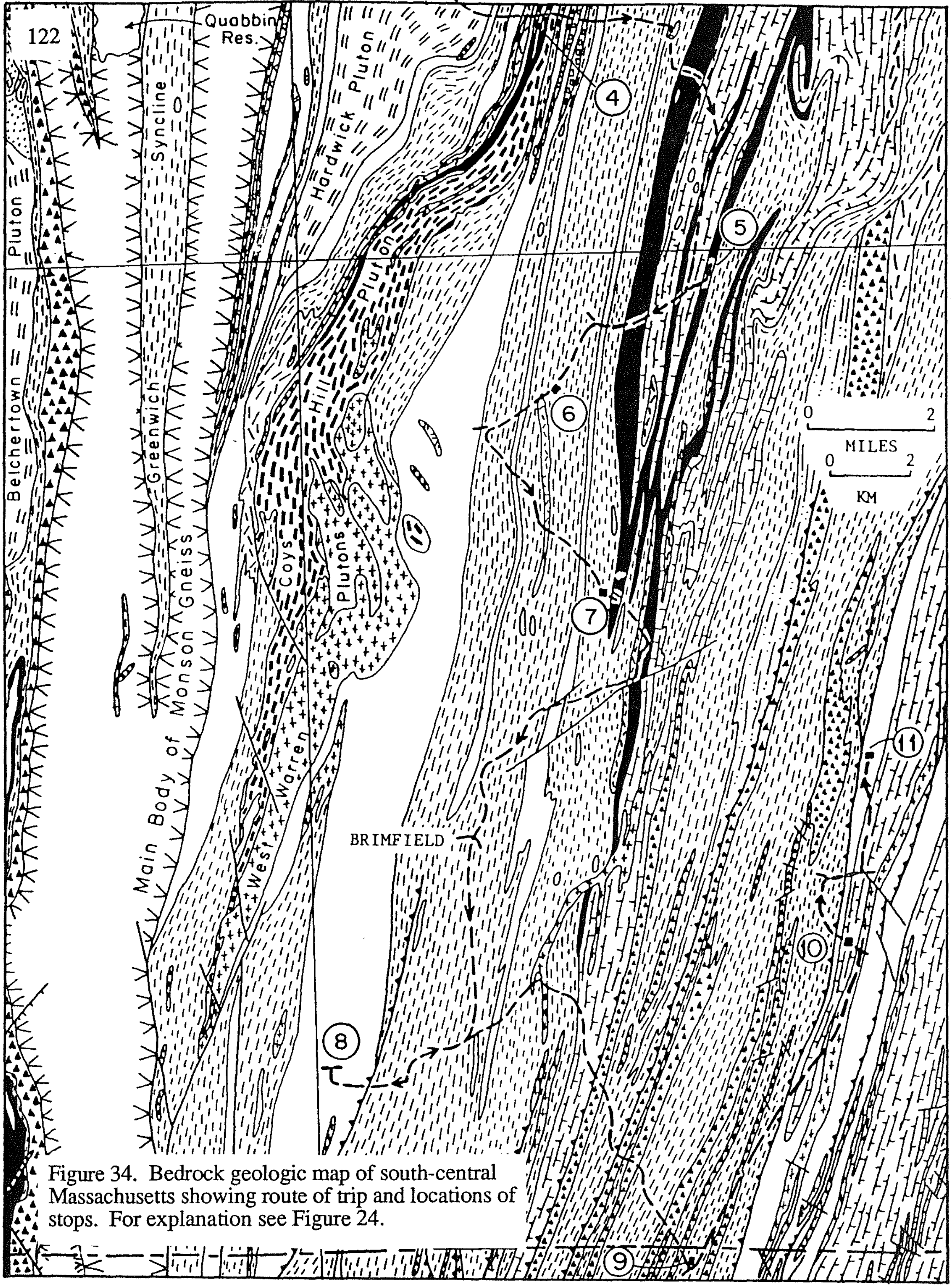

Figure 34. Bedrock geologic map of south-central Massachusetts showing route of trip and locations of stops. For explanation see Figure 24.

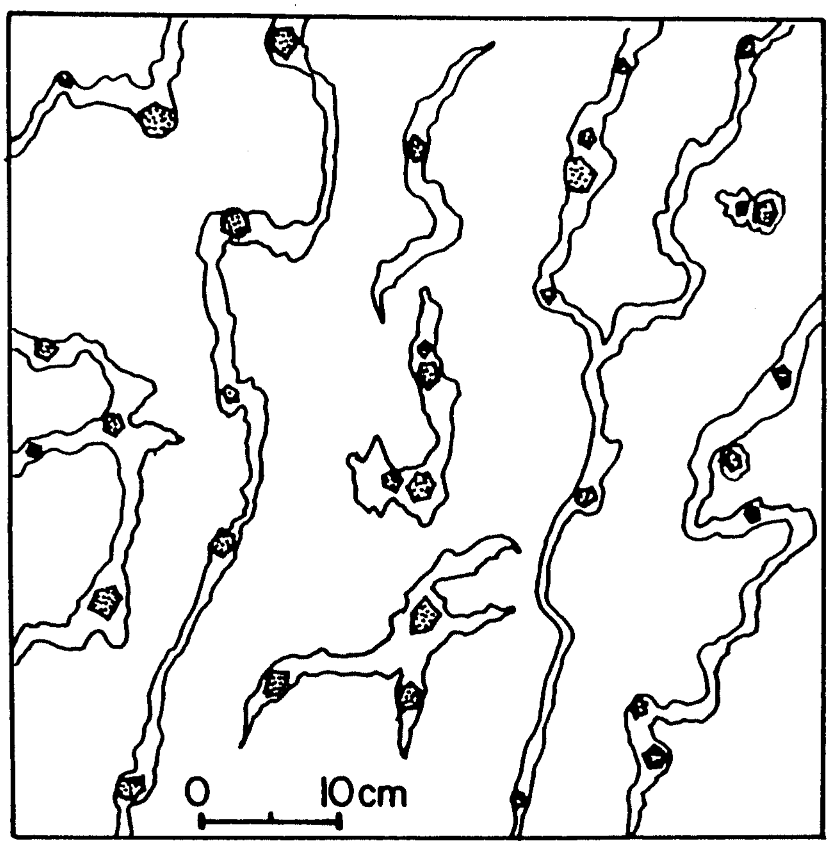

Figure 35. Sketch of part of the pasture outcrop at Stop 4 containing vein-type migmatite. The veins are medium to coarse feldspathic material containing fine-grained cordierite and sillimanite and very large garnets, shown by the stippled polyhedrons. The veins clearly crosscut prominent bedding in the host rock.

77.5 Gilbertville. Junction Rtes. 32A and 32. Turn right on 32.

77.7 Cross bridge over Ware River. Take sharp left turn (east) onto New Braintree Rd.

79.6 Park in barnyard on right and walk across hay field on right (south) to prominent natural outcrop.

STOP 4: LITTLETON FORMATION IN THE BIG GARNET SYNCLINE
(25 MINUTES) (Ware quadrangle)

This rock unit has been traced more or less continuously from the Ware Quadrangle at least as far north as Route 2. Its structural setting in an early isoclinal syncline between layers of Francestown Formation (Fitch Formation) can be seen on the geologic map (Figure 34). This rock type has been dubbed "pastureite" by Alan Thompson in honor of this beautiful glacially-smoothed outcrop. No Hammers Please!

Aside from a few layers of calc-silicate granulite, the rock in this outcrop can be described in two parts: a medium-grained schist or gneiss consisting of quartz, orthoclase, plagioclase, garnet,

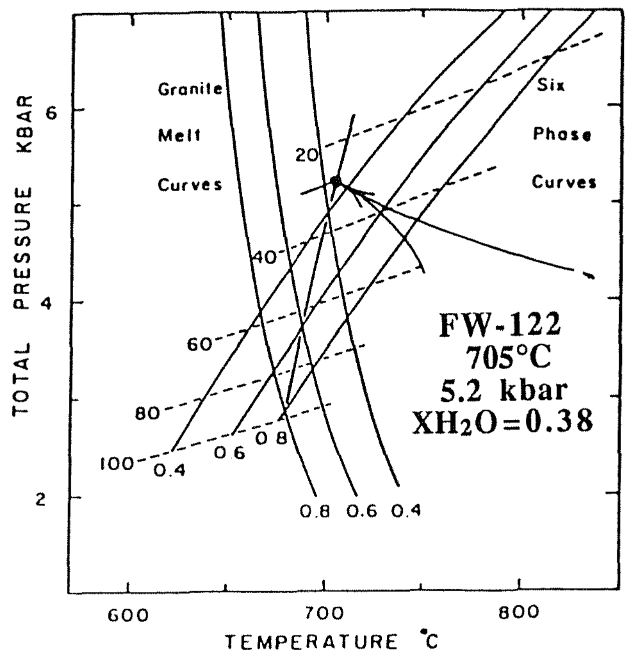

Figure 36. P-T diagram from Holdaway and Lee (1977) showing intersections of six-phase curves (quartz-orthoclase-sillimanite-garnet-biotite-cordierite) and granite melt curves to produce a fluid-absent melt curve. Dashed lines are X_{Fe} of cordierite. Suggested conditions of fluid-absent melting at Stop 4 are indicated. XH_2O in the fluid phase would be 0.38 only in the special case at the intersection of the three sets of curves.

cordierite, sillimanite, biotite, ilmenite and graphite, and a network of slightly deformed cross-cutting felsic veins consisting of quartz, orthoclase, plagioclase and cordierite with striking 2 to 4 cm euhedral garnets (Figure 35). All the garnet is remarkably uniform in composition except where in direct contact with biotite or cordierite, and has the following compositions for core and (rim): Almandine 69.6 (70.9), Pyrope 24.2 (22.8), Spessartine 2.8 (2.9) and Grossular 3.4 (3.4). The biotite has XMg of 0.57 and Ti/11 oxygens of 0.236. The cordierite has XMg of 0.71, and the plagioclase has XAn of 0.27.

Using the unmodified temperature calibrations of Thompson (1976b) and Ferry and Spear (1978), the garnet-biotite pair for this outcrop yields an estimated temperature of 726°C (T) and 746°C (F&S). Redoing the calculations using the almandine activity model of Hodges and Crowley yields 738°C (T) and 764°C (F&S). Pressure estimation using the GAsP barometer of Newton and Hasleton (1983) with the most recent Po and activity modifications of Koziol and Newton yields 5.8 kbar at 730°C. Interestingly, a sample from this outcrop contains coarse sillimanite that appears to be pseudomorphous after andalusite, as do numerous other samples from Mt. Monadnock to this area and on into northern Connecticut. If correctly interpreted, this texture implies that the belt underwent early low-P, high-T Buchan-type conditions followed by late compression with pressures several kbar higher.

An obvious explanation suggested by several visitors for the cross-cutting feldspathic veins as well as for the two sizes of garnets, and discussed by Tracy and Robinson (1983), is that the veins were formed by segregation of melt and that the large euhedral garnets grew in contact with melt. M.J. Holdaway has suggested that any ambient aqueous fluid may have been removed in an earlier stage of melting and melt extraction, and that the present texture is due to fluid-absent melting accompanying break-down of biotite in the six-phase assemblage quartz - sillimanite - K-feldspar - garnet - cordierite - biotite. Figure 36 shows Holdaway's calculated model where "six-phase curves" for different fictive XH_2O values in fluid intersect granitic melt curves for different XH_2O in melt to produce a fluid-absent melting curve. Dashed lines are isopleths of XFe in cordierite in the six-phase assemblage. Taking the cordierite with XFe of 0.29 from this outcrop and assuming the outcrop lay on the fluid-absent melt curve, the rock yields an estimated temperature of 705°C and a pressure of 5.2 kbar. The diagram also shows that if the rock had been exactly at an intersection of all three curves so that fluid could be present, it could have had an XH_2O no higher than 0.38.

Return to vans and proceed east on Gilbertville Road toward New Braintree.

81.7 Stop sign at T junction. Turn right (south) on West Brookfield Road. Directly east of junction are outcrops of the New Braintree ultramafic body (see Robinson, Tracy, Hollocher and Dietsch, 1982, p. P3-P43).

82.3 Bear left at Y junction.

82.6 Town line. Cross from West Brookfield into North Brookfield.

82.9 Junction. Bear right toward North Brookfield.

83.1 Junction. Bear right (south) on Barrett Road which becomes Wigwam Road upon crossing town line from North Brookfield back into West Brookfield.

86.3 Stop sign. Junction of Wigwam Road with Route 67. Sharp left turn (northeast) on Route 67.

86.7 Turn left (north) into barnyard and park vans. Walk north on farm road along southeast side of long outcrop.

Stop 5: SMALLS FALLS FORMATION, SULFIDE-RICH SCHIST (45 MINUTES)
(Warren quadrangle)

The outcrop has an irregular smoothed surface covered by a thick crust of iron oxides and sulfates. The outcrop surface contains many 3-5 cm pits inside which fresh-looking pyrite is typically visible. In spite of extensive searching, we have found only traces of megascopic pyrrhotite in the outcrop but are still inclined to believe that much of the outcrop's character is due to the weathering of this mineral. Partly weathered rock just beneath the crust looks white because of the abundance of colorless silicates. Much fresher rock has a bluish look. The field name "White Schist" came partly from the appearance of the broken outcrop and partly by analogy with the Mg-rich kyanite-talc rocks studied by Schreyer (1974) in Tanzania and Afghanistan.

The outcrop consists of two main rock types, sillimanite and biotite-bearing quartzites, and aluminous schists with variable proportions of sillimanite, biotite and cordierite. The biotites vary from from very pale reddish-brown Fe-bearing ones to colorless Mg end-members. Our first impression was that these were muscovites in a very retrograded fault zone but their nearly uniaxial interference figures, as well as subsequent microprobe analyses, showed they are biotites. The assemblage in two analysed samples from this outcrop (FW-882E and FW-882Y of Tracy and Robinson, 1988) is quartz-orthoclase-plagioclase-biotite-cordierite-sillimanite-rutile-pyrite-pyrrhotite-graphite. The XMg of biotite in these samples is 0.995 and 0.999 (less than 0.05 weight % FeO) and they contain 0.065 and 0.074 Ti per 11 oxygens. The K-feldspar has a composition Or 91.5, Ab 8.4, An 0.1 and plagioclase is An 32.7, Ab 57.6, Or 0.8.

The cordierites, which are full of quartz, sulfide and graphite inclusions, appear as black to bluish lumps. They are essentially pure Mg end-members (very low to undetectable FeO and MnO) and lack pleochroic haloes around monazite inclusions, presumably due to lack of iron to be oxidized by alpha bombardment. These two cordierites contain between 1.0 and 1.5 weight percent H_2S (sulfur analyzed by probe and reported as H_2S) as also does one other cordierite from another locality reported by Tracy and Robinson (1988). These sulfur-bearing cordierites only occur with pyrrhotite + pyrite assemblages which buffer very high sulfur activities that in turn produce significant molar proportions of H_2S in ambient high-temperature fluids (up to 50 mol. percent or higher). Because of the extreme narrowness (to put it mildly) of the sillimanite-biotite-cordierite three-phase "field" on the AFM (Ksp) diagram, this rock essentially lies on the univariant reaction Mg-biotite + sillimanite + quartz = Mg-cordierite + K-feldspar + V and can be considered divariant only because of the Na content of the K-feldspar. Preservation of the evidence of highly unusual localized fluid compositions in unusual rock types, for example the H_2S-rich fluids characteristic of parts of this outcrop, can be taken as a persuasive argument that there was a very limited volume of metamorphic fluid at peak conditions and that its chemistry was locally rather than externally controlled.

Return to vans. From barnyard turn right (west) on Route 67.

87.1 Drive past end of Wigwam Road.

87.6 Junction of Routes 9 and 67 in West Brookfield. Continue west on Rtes 9 and 67.

88.6 Turn left (southwest) on Route 67.

89.9 Rusty roadcut on left. Park on right and cross highway with care. During the Stop stay off the pavement except while crossing.

STOP 6: PYRRHOTITE-GARNET-BEARING SCHIST IN THE PLEASANT BROOK ANTICLINE (10 MINUTES APPROXIMATELY) (Warren quadrangle)

This is typical fine- to medium-grained Zone VI schist with the assemblage quartz - orthoclase - plagioclase - biotite - garnet - cordierite - sillimanite - graphite - ilmenite - pyrrhotite (sample FW-407). Although coated with a thin oxide crust, excellent fresh material can be collected here. The zoned Type C garnet illustrated by Tracy et al. (1976) was collected at this locality (Figure 7). The most magnesian garnet cores and most retrograded rims adjacent to biotite ( ) have the following compositions: Almandine 69.7 (77.7), Pyrope 24.7 (16.0), Spessartine 2.2 (3.0), and Grossular 3.4 (3.3). Note the delicate purplish tint of these pyrope-rich garnets. Matrix biotite has X_{Mg} of .572 and Ti/11 oxygens of .245 and biotite at garnet contacts goes to X_{Mg} .600. The cordierite has X_{Mg} of .714. Compositions of garnet core and matrix biotite yield temperature estimates of 724°C using the unmodified Thompson (1976b) calibration; garnet core and cordierite yield 680°C without pressure correction. The assemblage quartz-sillimanite-garnet-cordierite gives an estimated pressure of 6.1 kbar Retrograde garnet and biotite rims indicate a retrograde temperature of 530°C.

Continue southwest on Route 67.

90.2 Turn left off Route 67 at General Henry Knox monument. In the winter of 1775-1776 a group from the Continental Army passed this way dragging cannon captured by Ethan Allen and his "Green Mountain Boys" from the British at Fort Ticonderoga on Lake Champlain. The cannon were used by General George Washington during the siege of the British in Boston until they evacuated the city on March 17, 1776. The cannon were then hauled back to Fort Ticonderoga in time to be recaptured by the British in spring 1777, during the early stages of "Bœuf" Burgoyne's ultimately fatal southward drive fom Canada, ending at the Battle of Saratoga. General Knox later became the first U.S. Secretary of War under President Washington.

90.8 Stop sign. Revolutionary tree in middle of road. Go straight.

91.1 Dangerous intersection. Bear left toward Breezelands.

93.8 Park on right on steepest downslope where there are low road cuts on both sides.

STOP 7: SULFIDIC SCHIST AND MYLONITE IN THE WICKABOAG POND ANTICLINE (15 MINUTES) (Warren quadrangle)

This is locality WN-1 where J.S. Pomeroy discovered the mylonite that we have studied in detail and is discussed in the text. This is also the northernmost garnet-cordierite locality studied by Paul Hess (1971). The mylonite is less than 1 cm thick and is exposed on a very rotten part of the outcrop that could be damaged by hammering. In that vicinity it is possible to see the internal E-W lineation in the mylonite caused by elongate quartz rods and to see a parallel sillimanite lineation in the country rocks. The later NE-trending sillimanite lineation may also be seen. The folding of the mylonite foliation cannot be seen in the outcrop, but appears in thin section as a west-over-east overfold. A clear west-over-east shear sense is also easily determined from a variety of tiny boat shaped asymmetric porphyroclasts including monazite that gives pleochroic haloes in the enclosing biotite.

Several samples of coarse-grained gneiss from this outcrop with garnet-biotite, sillimanite-garnet-biotite-cordierite and sillimanite-biotite-cordierite assemblages have been partially analyzed (Figure 17). Garnets range up to 25.7% pyrope content. Biotite coexisting with sillimanite, garnet and cordierite has X_{Mg} of .604 and Ti/11 oxygens of .290. A nearby sample of the same assemblage without garnet contains biotite with XMg of .591 and Ti/11 oxygens of .230. This evidence of slightly crossing tie lines suggests some slight retrograde re-equilibration.

It is the dream of all petrologists to obtain numbers from outcrops. This outcrop yielded an instantaneous and direct determination on a field trip in spring, 1981, thanks to the sharp eyes of a graduate student. The number is 8 and is formed by two coalescing hollow globular clusters of graphite plates inside an orthoclase crystal in a cordierite pegmatite. This sample will be shown and advice and assistance sought.

94.9 Junction. Bear right, then bear right at second junction.

95.3 Warren/Brimfield Town Line.

96.0 Underpass beneath Massachusetts Turnpike.

99.5 Stop light in center of Brimfield. Cross Route 20 and proceed south on Route 19 toward Wales. The name Brimfield refers to "brimstone," an archaic name for sulphur, and relates to the abundant sulfates produced by the weathering of pyrrhotite schists in this region.

103.0 Intersection of Route 19 and Monson Road at Wales Post Office. Turn sharp right on Monson Road.

103.7 Monson Road bears left. Stay straight on McBride Road.

104.0 Three-way junction. Turn right (north) on Mt. Hitchcock Road.

104.2 Tennesee Gas Pipeline. Site of trench in fall 1985.

STOP 8: WELL BEDDED COARSE GRAY SCHIST OF THE RANGELEY (?) FORMATION NEAR MT. PISGAH(40 minutes) (Wales quadrangle)

This is an opportunity to collect very fresh material fairly similar to the rock in the pasture at Stop 4 though probably richer in sillimanite and cordierite and much more strongly deformed. Two samples of normal garnet-cordierite-sillimanite gneiss from this vicinity have been studied by microprobe. One sample from the small road cut on McBride Road yielded the following data: A typical garnet composition is Almandine 67.2, Pyrope 29.0, Spessartine 1.1, and Grossular 2.7. The biotite has X_{Mg} of .629 and Ti/11 oxygens of .252. The cordierite is not yet analyzed . The garnet and biotite yield an estimated temperature of 710°C and the garnet composition in the quartz - sillimanite - garnet - cordierite assemblage gives an estimated pressure of 6.4 kbar. A second sample from the pipeline trench has garnet composition Almandine 66.1, Pyrope 28.9, Spessartine 1.8, Grossular 3.1. The biotite has X_{Mg} of .60, the cordierite X_{Mg} of .24, and the plagioclase is An_{34}. This yields an estimated temperature of 758°C (Thompson, 1976b), a pressure of 6.3 kbar based on garnet-cordierite compositions, and a pressure of 5.6 kbar based on the GAsP barometer. A third sample lacking cordierite has garnet Almandine 68, Pyrope 23, Spessartine 5, Grossular 4, biotite with X_{Mg} of .52, and plagioclase $An_{35.5}$. This yields a temperature estimate of 784°C, and a pressure estimate of 6.9 kbar based on the GAsP barometer.

Samples collected from a tiny road-bed outcrop since obliterated near the intersection of Mt. Hitchcock Road and McBride Road are gneisses consisting of biotite, sillimanite, garnet, K-feldspar, plagioclase, and quartz. No cordierite was observed although it is present in an adjacent road cut 100 feet to the east. Within these gneisses there are felsic veins consisting primarily of K-feldspar and quartz with euhedral garnets up to 2 cm in diameter. In addition, there are abundant sillimanite pseudomorphs after andalusite up to 3 cm in diameter and 9 cm long, which are characterized by a zig-zag (010) cleavage in end-section under the petrographic microscope. The garnets within the felsic veins are homogeneous with typical compositions: Almandine 71, Pyrope 25, Spessartine 2, and Grossular 3. The coexisting biotite has X_{Mg} of 0.558 and Ti/11 oxygens of 0.300. The peak granulite facies temperature has been estimated at 739°C using Thompson's (1976b) calibration using garnet core - matrix biotite pairs.

Pegmatites consisting of K-feldspar, quartz, plagioclase, cordierite, sillimanite, garnet, and biotite are intercalated with the gneisses along the pipeline. Cordierite in the pegmatites (0.5 mm-10 cm) shows several textural and chemical features: 1) Most grains are partially rimmed by symplectic intergrowths of pale-green low-Ti biotite (Ti/11oxygens < 0.02)+sillimanite+quartz near large K-feldspar grains. 2) Locally, within interiors of large grains, there are symplectic intergrowths of garnet (0.25-3 mm)+sillimanite+quartz. 3) Some samples show symplectic intergrowths of skeletal sillimanite+quartz between adjacent cordierite and garnet. In these locations the garnet commonly shows euhedral overgrowths. 4) The pegmatite cordierite typically has Fe/(Fe+Mg) = 0.30-0.35, except directly adjacent to garnet where it is 0.16-0.20. This is in contrast to cordierite within adjacent gneisses where Fe/(Fe+Mg) = 0.22-0.24, except down to 0.18 where adjacent to garnet.

The cordierite pegmatites appear to be the product of in situ partial melting. The unusually Fe-rich cordierite compositions preserved suggest that this took place at lower pressure than the peak granulite facies conditions recorded in the surrounding gneisses. Analyses from a sample of pegmatite containing garnet, cordierite, biotite, sillimanite, quartz and plagioclase gave the following data: garnet - Almandine 70.5, Pyrope 23.6, Spessartine 2.8, Grossular 3.0; biotite X_{Mg} .575; cordierite X_{Mg} .68; plagioclase $An_{28.9}$. Conditions of pegmatite genesis have been estimated as follows: T based on Thompson (1976b) garnet-cordierite = 714°C; T based on P-corrected Thompson (1976b) garnet-biotite = 700°C; P based on Tracy et al., 1976 sillimanite-garnet-cordierite = 5.5-6.1 kbar; P based on Bhattacharya (1986) garnet-cordierite (assuming anhydrous cordierite) = 4.8-5.2 kbar; and P based on GAsP = 5.3 kbar. In the symplectites garnets in contact with cordierite with X_{Mg} .79-.84 have Almandine 62.6-68.2, Pyrope 26.7-33.2, Spessartine 1.8-2.2, and Grossular 2.4-3.1. Using the Thompson (1976b) garnet-cordierite thermometer without pressure correction these yield temperatures of 562-601°C and the sillimanite-garnet-cordierite barometer of Tracy et al., 1976 suggests pressures of 7-7.5 kbar. It appears that the large Fe-rich pegmatite cordierites did not re-equilibrate during peak granulite facies conditions recorded in the gneisses, but did respond by symplectite formation during cooling at still higher pressures. Together these rocks record part of a P-T path in which compression with heating appears to have been followed by further compression with cooling.

Turn around and return south on Mt. Hitchcock Road.

104.4 Turn left on McBride Road.

104.8 Bear left (east) on Monson Road.

105.5 Wales Post Office. Turm left (north) on Route 19.

106.7 Bear right (east) on Holland Road at intersection that is difficult to see.

107.4 Wales/Holland Town Line.

108.3 T junction. Turn right on Brimfield Road in Holland.

108.7 Road cut on right contains features of regional interest (see Robinson, Tracy, Hollocher, and Dietsch, 1982, p. P3-47). It is predominantly coarse-grained gneiss with the assemblage quartz - orthoclase - plagioclase - biotite - garnet - cordierite - sillimanite - graphite - ilmenite - pyrrhotite with variable amounts of sillimanite. There are several mylonites but none so fine-grained as the one at Stop 7. They contain porphyroclasts rather than porphyroblasts of new minerals.

Probe analyses of garnet yielded compositions of core and rim adjacent to biotite ( ) as follows: Almandine 65.2 (73.7), Pyrope 29.5 (21.2), Spessartine 1.4 (1.6), and Grossular 3.9 (3.4). Matrix biotite has X_{Mg} of .584 and Ti/11 oxygens of .275, whereas biotite against garnet has XMg as high as .604. Garnet core and matrix biotite yield a temperature estimate of 790°C, the highest reliable estimate we have obtained in the region, whereas garnet rim and biotite rim yield 620°C. Composition of core garnet in the quartz-sillimanite-garnet-cordierite assemblage suggests a pressure of 6.3 kbar.

110.4 Crossroads at Holland with windmill ! Go straight.

111.1 Causeway across Hamilton Reservoir. A sample of cordierite pegmatite for paper by Tracy and Dietsch (1982) was collected from rubble by Robinson and Klepacki in 1979.

112.5 New road cut on left at Connecticut State Line Monument. Well layered hornblende-clinopyroxrene-plagioclase-scapolite gneiss with secondary epidote and with cross cutting quartz-plagioclase-clinopyroxene pegmatite dikes.

112.7 Beginning of interchange at Mashapaug Road beyond truck depot. Park on grass strip to right and cross north to south end of large outcrop.

STOP 9: MASHAPAUG ROAD (20 MINUTES) (Wales quadrangle))

This outcrop (Figure 37) is located in Zone VI in the lower granulite facies. The dominant rock is a dark-gray two-pyroxene granulite that occupies the central part of the outcrop. On the northwestern end of the outcrop are exposed a felsic garnet-biotite tonalitic gneiss and an intermediate garnet- orthopyroxene-biotite gneiss. On the southeast end of the outcrop are a variety of calc-silicate granulites and calcite-diopside-quartz-sphene-scapolite-apatite-biotite pegmatite (NO HAMMERS ON THESE PEGMATITES, PLEASE).

The two-pyroxene granulite in the central part of the outcrop has the proper bulk composition to have been a quartz-bearing amphibolite at lower metamorphic grade, but all amphibole has broken down leaving an orthopyroxene- clinopyroxene-plagioclase-quartz-biotite assemblage. The feldspars in this rock have the compositions:

	An	Ab	Or	Cn
Plagioclase	43.7	55.5	0.9	0
Coarse exsolution lamellae of orthoclase	0.4	10.7	85.0	3.9

The celsian component in the exsolution lamellae is unusually high and may have promoted exsolution.

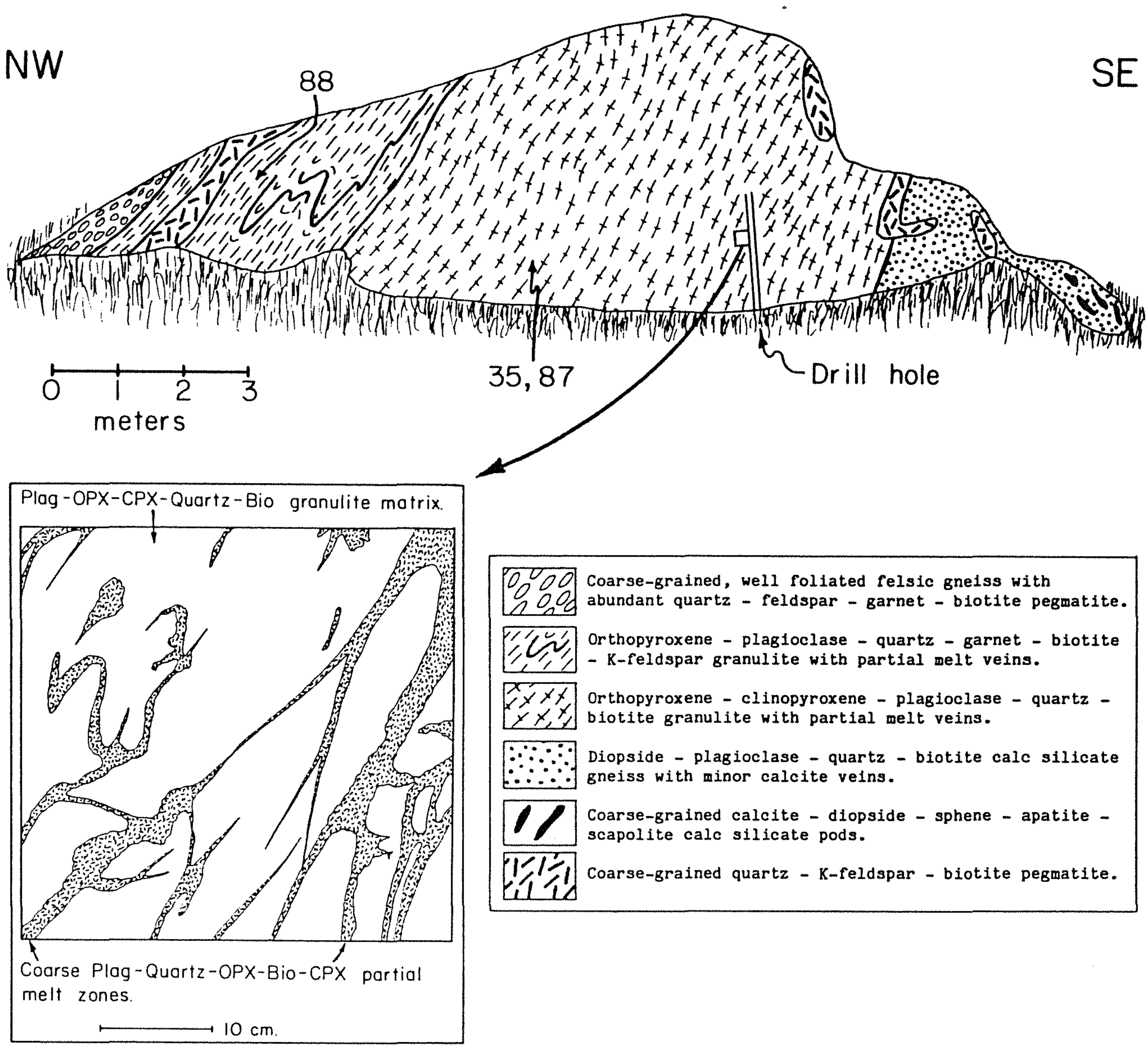

Figure 37. Schematic sketch of the southwest-facing outcrop at Stop 9, Mashapaug Rd., showing main lithologic units and sample locations. The inset is enlarged schematic sketch of the anastomosing partial melt vein network in the two-pyroxene granulite unit. Modes of samples 35, 87, and 88, and a bulk chemical analysis of sample 35 are given in Robinson et al. (1986). Electron microprobe analyses of minerals in samples 87 and 88 are also given in Robinson et al. (1986).

The mafic minerals have the following compositions:

	$Mg/(Mg+Fe^{2+})$	$Ca/(Ca+Fe^{2+}+Mg)$	
Augite	0.666	0.440	
Orthopyroxene	0.522	0.014	
Biotite	0.536		0.273 Ti/11 oxygens

The biotite is red-brown, typical for high-Ti biotite at this metamorphic grade. The original hornblende in this rock may have broken down by the continuous Fe-Mg a Ca-Na reaction:

Hornblende + quartz = orthopyroxene + augite + plagioclase + H_2O

as shown in Figure 33. Similar quartz-bearing but more magnesian assemblages to the north still contain hornblende together with its breakdown products, and quartz-free assemblages commonly contain hornblende even at this grade. The two-pyroxene granulite has apparently undergone partial melting, with crystallized melts forming a 3-dimensional network of anastomosing coarse-grained tonalitic veins containing pyroxenes, quartz, and plagioclase (Figure 37). The veins are visible on the fresh face of the roadcut and on the weathered surface on top of the outcrop. Melting may have taken place by the fluid-absent melting reaction:

Hornblende + quartz = orthopyroxene + augite + plagioclase + hydrous melt.

However, the pyroxenes are nearly as abundant in the coarse segregations as they are in the fine-grained matrix, and orthopyroxene is more abundant. It is possible that orthopyroxene was more soluble in the tonalitic liquid than was augite, but this seems unlikely. Alternatively, the segregations may have formed earlier, perhaps during a cummingtonite breakdown reaction that produced orthopyroxene + melt. Hornblende may have broken down later with or without additional melting. The two-pyroxene granulite does not contain K-feldspar, but the garnet-bearing intermediate gneiss just to the northwest does contain the granulite facies orthopyroxene-biotite-K-feldspar-quartz assemblage that formed by the partial completion of the reaction:

Biotite + plagioclase + quartz = K-feldspar + orthopyroxene + garnet + H_2O.

End of Formal Field Trip: Those wishing to remain will be conducted to Stops 10 and 11.

Bear right (south) from parking place.

112.8 Stop sign. Turn left (east) on bridge across Interstate 84.

113.0 Turn left (north) on entrance ramp to Interstate 84 North.

118.3 Bear right on Exit 3 for Route 131.

118.5 Stop sign at end of ramp. Turn left (west) on Old Sturbridge Village Road, crossing over Interstate 84.

118.6 Junction. Turn right on Old Sturbridge Village Road.

118.8 Wide place in road with outcrop to right at back entrance to Old Sturbridge Village.

STOP 10 (10 MINUTES) (Southbridge quadrangle)

Gray biotite-garnet granulites, and calc-silicate granulites of the Paxton Formation with abundant pegmatite and tight isoclinal folds.

Proceed through gate and continue on Old Sturbridge Village Road.

119.9 Junction with Route 20 at front entrance of Old Sturbridge Village. Turn right (east) on Rte 20.

120.2 Stop lights. Stay straight on Rte 20.

120.4 Turn sharp left (north) on New Boston Road (sign for "Hairport").

122.0 Bridge across Massachusetts Turnpike. Park on right beyond bridge. Step over fence on right and climb down under bridge on north side of Turnpike to large road cut behind protective railings. Stay in ditch and away from railings at all times.

STOP 11: COARSE CORDIERITE-GARNET-SILLIMANITE GNEISS AND SILLIMANITE PEGMATITE (40 MINUTES) (East Brookfield quadrangle)

This rock lies in a very narrow belt of gray-weathering aluminous schist with subordinate calc-silicate that has been assigned to the Rangeley(?) Formation. After descending to road level, walk quickly west to far end of outcrop and then work your way slowly back to bridge. The dominant rock type here is beautifully coarse quartz - orthoclase - plagioclase - biotite - garnet - cordierite - sillimanite - graphite gneiss in which cordierite, garnet, and sillimanite can be exceedingly coarse.

Within the gneiss are two sills of garnet and garnet-sillimanite pegmatite usually showing very strong deformational fabric. This portion of the pegmatite apparently contains no sillimanite and no primary biotite, but does contain retrograde low Ti biotite at contacts between garnet and K-feldspar. Next to the retrograde green biotite, the garnet shows extremely abrupt zoning. Compositions of garnet core and of garnet rim next to green biotite ( ) are as follows: Almandine 70.7 (82.5), Pyrope 26.0 (11.8), Spessartine 2.4 (3.9), and Grossular 0.9 (1.8). The retrograde green biotite has X_{Mg} of .464 and .013 Ti/11 oxygens. When paired with the garnet rim this green biotite gives a temperature estimate of 545 °C. In the pegmatite at the far end there are local patches of graphite and of fairly coarse white mica that appears to be muscovite (!?). Elsewhere the sillimanite-rich portions of the pegmatite may be metamorphosed equivalents of muscovite pegmatite.

Along the contacts of the pegmatites are dark layers up to 0.3 meters thick consisting almost exclusively of biotite, garnet, sillimanite, and cordierite. These appear to be either restite layers left behind during melting of the pegmatite or metamorphosed reaction rims between pegmatite and country rock (Note high specific gravity of specimens). In a few places there are layers of pure cordierite up to 4 cm thick at the contact between these layers and the pegmatite. The garnet is fairly homogeneous except there in contact with surrounding biotite ( ) yielding the following compositions: Almandine 72.0 (82.1), Pyrope 24.9 (14.6), Spessartine 1.2 (2.0), and Grossular 1.9 (1.4). Typical biotite has X_{Mg} of .597 and .230 Ti/11 oxygens. Garnet core and matrix biotite yield an estimated temperature of 682 °C. In the absence of quartz the sillimanite-garnet-cordierite assemblage can be used to estimate a maximum pressure of 6.3 kbar.

Vans turn around and proceed south on New Boston Road.

123.7 Junction with Route 20. Turn left (east) on Route 20.

124.0 Exit left (north) for Massachusetts Turnpike.

124.9 Massachusetts Turnpike toll booth.

125.0 Fork. Bear left for Springfield, Bradley Field, Albany, and NORTH AMERICA. Bear right for Boston, AVALON and beyond. AU REVOIR and BON VOYAGE !

REFERENCES CITED

Ashenden, David D. (1973) Stratigraphy and structure, northern portion of the Pelham Dome, North-Central Massachusetts. Dept. of Geology, Contribution No. 16 (M.S. Thesis) University of Massachusetts, Amherst, 132 p., 3 plates

Ashwal, L.D., G.W. Leo, Peter Robinson, R.E. Zartman, and D.J. Hall (1979) The Belchertown quartz monzodiorite pluton, west-central Massachusetts, a syntectonic Acadian intrusion. American Journal of Science, 279, p. 936-969.

Barker, Fred and Arth, J.G. (1984) Preliminary results, Central Gneiss Complex of the Coast Range batholith, southeastern Alaska: the roots of a high-K, calc-alkaline arc? Physics of the Earth and Planetary Interiors, v. 35, p. 191-198.

Berner, R.A. (1970) Sedimentary pyrite formation. American Journal of Science, v. 268, p. 1-23.

Berry, H.N., IV (1985) The Silurian Smalls Falls Formation in south-central Massachusetts and adjacent Connecticut. Geological Society of America, Abstracts with Programs, v. 17, p. 4.

Berry, H.N., IV (1987) Imbricate thrust slices of pre-Silurian basement and Silurian-Devonian cover in the Merrimack belt, southern Massachusetts and northern Connecticut. Geological Society of America Abstracts with Programs, v. 19, p. 5.

Berry, H.N., IV (1987) Structure and tectonics of the Acadian granulite-facies region, Merrimack belt, south-central New England. Geological Society of America, Abstracts with Programs, v. 19, p. 587.

Berry, H.N., IV (1989b) A new stratigraphic and structural interpretation of granulite-facies metamorphic rocks, Brimfield-Sturbridge area, Massachusetts and Connecticut. Unpublished Ph.D. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.

Berry, H.N., IV, and Robinson, Peter (1989) The granulite-facies region of the Merrimack synclinorium Massachussetts, USA: an open and shut case. TERRA Abstracts, v. 1, p. 359.

Bhattacharya, A. (1986) Some geobarometers involving cordierite in the FeO-Al_2O_3-SiO_2 ($\pm H_2O$) system: refinements, thermodynamic calibration, and applicability to granulite facies rocks. Contributions to Mineralogy and Petrology, v. 94, p. 387-394.

Bohlen, S.R., Wall, V.J., and Boettcher, A.L. (1983) Experimental investigation and application of model garnet granulite equilibria. Contributions to Mineralogy and Petrology, v. 83, p. 52-61.

Boucot, A.J., MacDonald, G.J.F., Milton, C., and Thompson, J.B., Jr. (1958) Metamorphosed middle Paleozoic fossils from central Massachusetts, eastern Vermont, and western New Hampshire: Geological Society of America Bulletin, v. 69, p. 855-870.

Chamberlain, C.P. (1985) Tectonic and metamorphic history of a high-grade terrane, southwestern New Hampshire. Ph.D. thesis, Harvard University, Cambridge, Massachusetts, 207 p.

Elbert, D. C. (1986) Recognition and implications of a structurally inverted Monadnock-western Maine stratigraphy directly above the core of the Bernardston nappe, Hinsdale, New Hampshire. Geological Society of America, Abstracts with Programs, v.18, p. 15.

Elbert, D.C. (1987) P-T path convergence based on garnet Fe-enrichment growth during prograde staurolite breakdown in tectonically juxtaposed pelites from southwestern N.H. Geological Society of America, Abstracts with Programs, v. 19, p. 653-654.

Elbert, D.C., A. G. Harris and K. E. Denkler (1988) Earliest Devonian Conodonts from marbles of the Fitch Formation, Bernardston nappe, north-central Massachusetts. American Journal of Science, v. 288, p. 684-700.

Ellis, Michael and Watkinson, A.J. (1987) Orogen-parallel extension and oblique tectonics: The relation between stretching lineations and relative plate motions. Geology, v. 15, p. 1022-1026.

Emerson, B.K. (1917) Geology of Massachusetts and Rhode Island. U.S. Geological Survey Bulletin 597, 289 p.

England, P.C. and Thompson, A.B. (1984) Pressure-temperature-time paths of regional metamorphism, Part I. Heat transfer during the evolution of regions of thickened continental crust. Journal of Petrology, v. 25, p. 894-928.

Evans, B.W. and Guidotti, C.V. (1966) The sillimanite-potash feldspar isograd in western Maine, U.S.A. Contributions to Mineralogy and Petrology, v. 12, p. 25-62.
Fahlquist, F.E. (1935) Geology of region in which Quabbin Aqueduct and Quabbin Reservior are located; Appendix to Report of Chief Engineer, Annual Report of the Metropolitan District Water Supply Commission, Commonwealth of Massachusetts, Public Document No. 147.
Ferry, J.M., and Spear, F.S. (1978) Experimental calibration of the partitioning of Fe and Mg between biotite and garnet. Contributions to Mineralogy and Petrology, v. 66, p. 113-117.
Field, M.T. (1975) Bedrock geology of the Ware area, central Massachusetts. Contribution No. 22 (Ph.D. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 186 p.
Finkelstein, David B. (1987) Sense of shear in a major mid-Acadian mylonite zone in granulite-facies metamorphosed tonalite and Rangeley Formation of the Merrimack belt, northeastern Connecticut. Geological Society of America, Abstracts with Programs, v. 19, p. 11.
Ganguly, J. and Saxena, S.K. (1984) Mixing properites of alluminosilicate garnets: constraints from natural and experimental data, and applications to geothermobarometry. American Mineralogist, v. 69, p. 88-97.
Ghent, E.D. (1976) Plagioclase-garnet-Al_2SiO_5-quartz: a potential geobarometer-geothermometer. American Mineralogist, v. 61, p. 710-714.
Ghent, E.D. and Stout, M.Z. (1981) Geobarometry and geothermometry of plagioclase-biotite-garnet-muscovite assemblages. Contributions to Mineralogy and Petrology, v. 76, p. 92-97.
Guidotti, C.V., Cheney, J.T., and Guggenheim, Stephen (1977) Distribution of titanium between coexisting muscovite and biotite in pelitic schists from northwestern Maine. American Mineralogist, v. 62, p. 438-448.
Hall, Leo M. and Peter Robinson (1982) Stratigraphic-tectonic subdivisions of southern New England: p. 15-41, In St.-Julien, P. and J. Beland, (ed.) Major Structural Zones and Faults of the Northern Appalachians, Geological Association of Canada Special Paper 24.
Halpin, David L. (1965) The geometry of folds and style of deformation in the Quabbin Hill area, Massachusetts, M.S. Thesis, 93 p.
Hamilton, W.B. (1988) Plate tectonics and island arcs. Geological Society of America Bulletin, v. 100, p. 1594-1615.
Hatch, N.L., Jr., Moench, R.H., and Lyons, J.B. (1983)Silurian-Lower Devonian stratigraphy of eastern and south-central New Hampshire: extensions from western Maine. American Journal of Science, v. 283, p. 739-761.
Henry, D.J., and Guidotti, C.V. (1981) Implications of the sulfide-silicate interactions in staurolite grade metapelites in NW Maine. (abstract) EOS, v. 62, p. 435-436.
Hensen, B.J. (1977) Cordierite-garnet bearing assemblages as geothermometers and barometers in granulite facies terranes. Tectonophysics, v. 43, p. 73-88.
Hensen, B.J., and Green, D.H. (1973) Experimental study of the stability of cordierite and garnet in pelitic compositions at high pressures and temperatures. Contributions to Mineralogy and Petrology, v.38, p. 151-166.
Hess, P.C. (1971) Prograde and retrograde equilibria in garnet-cordierite gneisses in south-central Massachusetts. Contributions to Mineralogy and Petrology, v. 30, p. 177-195.
Hodges, K.V. and Crowley, P.D. (1985) Error estimation and empirical geothermobarometry for pelitic systems. American Mineralogist, v. 70, p. 702-709.
Hodges, K.V., and Royden L. (1984) Geothermometry and geobarometry of retrograded metamorphic rocks: an indication of the uplift trajectory of a portion of the northern Scandinavian Caledonides. Journal of Geophysical Research, v.89, p.7077-7090.
Hodgkins, C.E. (1983) Major element geochemistry and petrology of the Dry Hill Gneiss, Pelham dome, central Massachusetts, Geological Society of America Abstracts With Programs. v. 15, p. 141.
Hodgkins, C.E. (1985) Major and trace element geochemistry and petrology of the late Precambrian Dry Hill Gneiss, Pelham Dome, central Massachusetts. Contribution No. 48 (M.S. thesis), Dept. of Geology and Geography, University of Massachusetts, Amherst, 135 p.

Holdaway, M.J., and Lee, S.M. (1977) Fe-Mg cordierite stability in high grade pelitic rocks based on experimental, theoretical, and natural observations. Contributions to Mineralogy and Petrology, v. 63, p. 175-198.
Hollocher, K.T. (1983) Major element geochemistry and possible origin of volcanic rocks in the Middle Ordovician Partridge Formation, central Massachusetts. Geological Society of America Abstracts with Programs, v. 15, p. 172.
Hollocher, K.T. (1985) Geochemistry of metamorphosed volcanic rocks in the Middle Ordovician Partridge Formation, and amphibole dehydration reactions in the high-grade metamorphic zones of central Massachusetts. Contribution No. 56 (Ph.D. thesis), Deptartment of Geology and Geography, University of Massachusetts, Amherst, 275p.
Hollocher, K.T. (1987) Petrology and geochemistry of the Monson Gneiss and related rocks, Bronson Hill anticlinorium, Massachusetts. Geological Society of America, Abstracts with Programs, v. 19, p. 19.
Hollocher, Kurt (1988) Geochemical comparisons of the Monson Gneiss and associated tonalitic Acadian plutons, and overlying Ordovician volcanics, Massachusetts: Geological Society of America, Abstracts with Programs, v. 20.
Hollocher, Kurt and Lent, Andrew D. (1987) Comparative petrology of amphibolites in the Monson Gneiss and the Ammonoosuc and Partridge Volcanics, Massachusetts. Northeastern Geoloby, v. 9, p. 145-152.
Hollocher, Kurt T. (1981) Retrograde metamorphism of the Lower Devonian Littleton Formation in the New Salem area, west-central Massachusetts. Contribution No. 37 (M.S. Thesis) Department of Geology and Geography, Univerisity of Massachusetts, Amherst, 265 p.
Huntington, J.C. (1975) Mineralogy and petrology of metamorphosed iron-rich beds in the Lower Devonian Littleton Formation, Orange area, Massachusetts. Contribution No. 19,)M.S. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 106 p.
Johnson, Rex, Van der Voo, Rob, and Van der Pluijm, Ben (1989) Paleomagnetism of the Morton's Harbour Group: constraint on the early Ordovician paleolatitude of an island arc terrane in the Newfoundland Appalachians. EOS Transactions, American Geophysical Union, v. 70, No. 15, p. 312.
Kohn, M.J. and Spear, F.S. (1989) Empirical calibration of geobarometers for the assemblage garnet + hornblende + plagioclase + quartz. American Mineralogist, v. 74, p. 77-84.
Koziol, A.M. and Newton, R.C. (1988) Redetermination of the anorthite breakdown reaction and improvement of the plagioclase-garnet-Al_2SiO_5-quartz geobarometer. American Mineralogist, v. 73, p. 216-223.
Leo, G.W. (1985) Trondhjemite and metamorphosed quartz keratophyre tuff of the Ammonoosuc Volcanics, western New Hampshire and adjacent Vermont and Massachusetts. Geological Society of America Bulletin, v.96, p.1493-1507.
Lonker, S.W. (1981) The P-T-X relations of the cordierite-garnet-sillimanite-quartz equilibrium. American Journal of Science, v. 281, p. 1056-1090.
Martignole, Jacques, and Sisi, J-C. (1981) Cordierite-garnet-H20 equilibrium: A geological thermometer, barometer, and water-fugacity indicator. Contributions to Mineralogy and Petrology, v. 77, p. 38-46.
McEnroe, S.A. (1989) Paleomagnetism and geochemistry of Mesozoic diabase dikes in wesst-central Massachusetts. Contribution No. 64 (M.S. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts, 169 p.
McEnroe, S.A. andd Brown, L.L. (1987) Paleomagnetism of diabase dikes and sills intruding pre-Mesozoic basement rocks near and east of the Connecticut Valley Mesozoic basins, west-central Massachusetts. EOS, Transactions, American Geophysical Union, v. 68, p. 295.
McEnroe, Suzanne A. (1988) Diversity in paleomagnetism, geochemistry and age in Mesozoic diabase intrusions, central Massachusetts. Geological Society of America, Abstracts with Programs, v. 20.
McEnroe, Suzanne A., Robinson, Peter, Brown, Laurie L., and Foland, Kenneth A. (1987) Paleomagnetic discovery of Cretaceous diabase sills in central Massachusetts. Geological Society of America, Abstracts with Programs, v. 19, p. 763.

Mohr, D.W., and Newton, R.C. (1981) Role of sulfur fugacity in metamorphism of mica schists. (abstract) EOS, v. 62, p. 436.

Naylor, R.S., Boone, G.M., Boudette, E.L., Ashenden, D.D., and Robinson, Peter (1973) Pre-Ordovician rocks in the Bronson Hill and Boundary Mountains anticlinoria, New England, U.S.A. (abstract) Transactions of the American Geophysical Union, v. 54, p. 495.

Newton, R.C., and Haselton, H.T. (1981) Thermodynamics of the garnet-plagioclase-Al2SiO5-quartz geobarometer. In Newton, R.C., et al.,(eds.), Thermodynamics of minerals and melts, p. 131-147. Springer-Verlag, New York.

Newton, R.C., and Wood, B.J. (1979) Thermodynamics of water in cordierite and some petrologic consequences of cordierite as a hydrous phase. Contributions to Mineralogy and Petrology, v. 68, p. 391-405.

Ohmoto, H. and Kerrick, D.M. (1977) Devolatilization equilibria in graphitic systems. American Journal of Science, v. 277, p. 1013-1044.

Onasch, Charles M. (1973) Analysis of the minor structural features in the north-central portion of the Pelham Dome: Dept. of Geology, Univ. of Mass., Amherst, Contribution No. 12 (M.S. Thesis), 87 p., 6 plates.

Osberg, P.H., Robinson, Peter, Tull, J.F., Hon, Rudolph and Butler, J.R. (in press) The Acadian orogen: in Hatcher, R.D., Editor, Geology of the U.S. Appalachians. Geological Society of America, Decade of North American Geology.

Perkins, Dexter, III, and Chipera, S.J. (1985) Garnet-orthopyroxene-plagioclase-quartz barometry: refinement and application to the English River subprovince and the Minnesota River valley. Contributions to Mineralogy and Petrology, v. 89, p. 69-80.

Peterson, Virginia L. (1984) Structure and stratigraphy of the bedrock geology in the Ashburnham-Ashby area, north-central Massachusetts: Contrib. No. 47 (M.S. Thesis), Dept. of Geology and Geography, Univeristy of Massachusetts, Amherst, 182 p., 6 plates.

Reed, Robert M. and Williams, Michael L. (1989) Petrofabric kinematic indicators in the northern portion of the Pelham Dome, western Bronson Hill anticlinorium, central Massachusetts. Geological Society of America, Abstracts with Programs, v. 21, p. 60.

Robinson, Peter (1963) Gneiss domes of the Orange area, west-central Massachusetts and New Hampshire. Ph.D. dissertation, Harvard University, Cambridge, Massachusetts, 253 p.

Robinson, Peter (1967a) Guidebook for Field Trips in the Connecticut Valley of Massachusetts: 59th Annual Meeting of the New England Intercollegiate Geological Conference, Amherst, Massachusetts, 194 p.

Robinson, Peter (1967b) Gneiss domes and recumbent folds of the Orange area, west-central Massachusetts. In Robinson, Peter, (ed.), Guidebook for field trips in the Connecticut Valley of Massachusetts. NEIGC 59th Annual Meeting, p. 17-47. Amherst, Massachusetts.

Robinson, Peter (1967c) Geology of the Quabbin Reservoir area, central Massachusetts: In Robinson, Peter, (ed.), Guidebook for field trips in the Connecticut Valley of Massachusetts. NEIGC 59th Annual Meeting, p. 114-127, 6 figures.

Robinson, Peter (1979) Bronson Hill anticlinorium and Merrimack synclinorium in central Massachusetts. In Skehan, J.W., S.J., and Osberg, P.H., (eds.), The Caledonides in the U.S.A.: geological excursions in the northeast Appalachians. IGCP Project 27, Caledonide Orogen, p. 126-150. Weston Observatory, Weston, Massachusetts.

Robinson, Peter (1981) The basement-cover enigma in the gneiss domes of central New England, U.S.A. (abstract) Uppsala Caledonide Symposium. Terra Cognita, v. 1, p. 70.

Robinson, Peter (1981b) Siluro-Devonian stratigraphy of the Merrimack synclinorium, central Massachusetts--Review based on correlations in Maine. Geological Society of America, Abstracts with Programs, v. 13, p. 172.

Robinson, Peter (1983) Realms of regional metamorphism in southern New England, with emphasis on the eastern Acadian metamorphic high. In Schenk, P.E., (ed.), Regional trends in the geology of the Appalachian-Caledonian-Hercynian-Mauritanide orogen, p. 249-258. Reidel press, Holland.

Robinson, Peter (1987) Root zones of the Acadian Bernardston nappe and Brennan Hill thrust, Mt. Grace area, north-central Massachusetts. Geological Society of America, Abstracts with Programs, v. 19, p. 54.

Robinson, Peter, and Hall, L.M. (1980) Tectonic synthesis of southern New England. In Wones, D.R., ed., The Caledonides in the U.S.A. IGCP Project 27, Caledonide orogen, p. 73-82. Blacksburg, Virginia.

Robinson, Peter, and Jaffe, H.W. (1969a) Aluminous enclaves in gedrite-cordierite gneiss from southwestern New Hampshire. American Journal of Science, v. 267, p. 389-421.

Robinson, Peter, and Jaffe, H.W. (1969b) Chemographic exploration of amphibole assemblages from central Massachusetts and southwestern New Hampshire. In Papike, J.J., ed., Pyroxenes and amphiboles: crystal chemistry and phase petrology. Mineralogical Society of America Special Paper 2, p. 251-274.

Robinson, Peter, Elbert, D.C., Tracy, R.J., Thompson, P.J., and Hollocher, K.T. (1986a) P-T trajectories of Acadian metamorphism in central Massachusetts and southwestern New Hampshire. Geological Society of America Abstracts with Programs, v.18, p. 63.

Robinson, Peter, Field, M.T. and Tucker, R.D.(1982) Stratigraphy and structure of the Ware-Barre area, central Massachusetts: Guidebook, New England Intercollegiate Geological Conference, 74th Ann. Meeting, Storrs, Connecticut, p. 341-373.

Robinson, Peter, Huntington, J.C., McEnroe, S.A., and Springston, G.C. (1988) Root zone of the Bernardston nappe and the Brennan Hill thrust involuted by backfolds and gneiss domes in the Mount Grace area, north-central Massachusetts. In Bothner, W.A., ed. Guidebook for Field Trips in Southwestern New Hampshire, Southeastern Vermont, and North-central Massachussetts, p. 293-334, New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, New Hampshire.

Robinson, Peter, Robert J. Tracy, and J.S. Pomeroy (1977) High pressure sillimanite-garnet-biotite assemblage formed by recrystallization of mylonite in sillimanite-cordierite-biotite-orthoclase schist, central Massachusetts. Geological Society of America, Abstracts with Programs, v. 9, p. 1144-1145.

Robinson, Peter, Robert J. Tracy, and Lewis D. Ashwal (1975) Relict sillimanite orthoclase assemblage in kyanite-muscovite schist, Pelham Dome, west-central Massachusetts. Transactions American Geophysical Union, v. 56, p. 466.

Robinson, Peter, Thompson, J.B., Jr., and Rosenfeld, J.L. (1979) Nappes, gneiss domes, and regional metamorphism in western New Hampshire and central Massachusetts. In Skehan, J.W., S.J., and Osberg, P.H., (eds.), The Caledonides in the U.S.A.: geological excursions in the northeast Appalachians. IGCP Project 27, Caledonide Orogen, p. 126-150. Weston Observatory, Weston, Massachusetts.

Robinson, Peter, Tracy, R.J., Hollocher, K.T., and Dietsch, C.W. (1982) High grade Acadian metamorphism in south-central Massachusetts. In Joesten, Raymond, and Quarrier, S.S., (eds.), Guidebook for fieldtrips in Connecticut and south central Massachusetts. NEIGC 74th Annual Meeting, p. 289-340. The University of Connecticut, Storrs.

Robinson, Peter, Tracy, Robert J., Hollocher, Kurt T., Schumacher, John C., and Berry, Hernry N., IV (1986b) The central Massachusetts metamorphic high,guidebook: Regional metamorphism and metamorphic phase relations in northwestern and central New England: Field Trip B-5, International Mineralogical Association, 14th General Meeting at Stanford University, Contribution No. 59, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts, p. 195-266.

Robinson, Peter, Tucker, R.D., and Hollocher, K.T. (1989) The Bronson Hill volcanic arc and its collision with North America: A radical new interpretation. Geological Society of America, Abstracts with Programs, v. 21, p.62.

Roll, M.A. (1986) Effects of Acadian prograde kyanite-zone metamorphism on relict garnet from pre-Acadian granulite facies metamorphism, Mt. Mineral Fm., Pelham Dome, Massachusetts. Geological Society of America, Abstracts with Programs, v.18, p. 63.

Roll, M.A. (1987) Effects of Acadian kyanite-zone metamorphism on relict granulite-facies assemblages, Mount Mineral Formation, Pelham dome, Massachusetts: Contribution No. 60, (M.S. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 202 p.

Rumble, D., III (1974) Gradients in the chemical potentials of volatile components between sedimentary beds of the Clough Formation, Black Mountain, New Hampshire. Carnegie Institute of Washington Year Book 73, p. 371-380.

Schreyer, Werner (1974) Whiteschist, a new type of metamorphic rock formed at high pressures. Geologische Rundschau, v. 63, p. 597-609.

Schumacher, J.C. (1983) Stratigraphy, geochemical, and petrologic studies of the Ammonoosuc Volcanics, north-central Massachusetts and southwestern New Hampshire. Ph.D. dissertation, University of Massachusetts, Amherst, Massachusetts, 237 p.

Schumacher, J.C. (1987) Stratigraphy and geochemistry of the Ammonoosuc Volcanics, central Massachusetts and southwestern New Hampshire: American Journal of Science, v. 288, p. 619-663.

Schumacher, J.C. (1988) Phase relations of multiple, Mn-bearing amphibolite and pyroxene assemblages: previewing the granulite facies in quartzites at "Amphibole Hill" southwestern New Hampshire. Geological Society of America, Abstracts with Programs, v. 20, p. A160.

Schumacher, J.C. (1988) Stratigraphy and geochemistry of the Ammonoosuc Volcanics, central Massachusetts and southwestern New Hampshire. American Journal of Science, v. 288, p. 619-663.

Schumacher, J.C., and Robinson, Peter (1986) Gedrite-cordierite gneisses and related metamorphosed volcanic rocks, guidebook: Regional metamorphism and metamorphic phase relations in northwestern and central New England. Field Trip B-5, International Mineralogical Association, 14th General Meeting at Stanford University, Contribution No. 59, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts, p. 144-149.

Schumacher, J.C., and Robinson, Peter (1987) Mineral chemistry and metasomatic growth of aluminous enclaves in gedrite-cordierite gneiss, southwestern New Hampshire. Journal of Petrology, v. 28, p. 1033-1073.

Schumacher, J.C., Schumacher, Renate, and Robinson, Peter (1987) Acadian metamorphism in central Massachusetts and southwestern New Hampshire: Evidence for contrasting P-T trajectories. Abstract for Conference Program "Evolution of Metamorphic Belts" at Dublin University, September 1987, sponsored by Metamorphic Studies Group of the Geological Society and IGCP Project 235, Metamorphism and Geodynamics.

Schumacher, J.C., Schumacher, Renate, and Robinson, Peter (in review) "Evolution of Metamorphic Belts", Geological Society London.

Schumacher, Renate (1986) Prograde metamorphism of hornblende- and epidote-bearing rocks, lower to upper amphibolite facies, central Massachusetts, U.S.A. Absttracts, 14th General Meeting of the International Mineralogical Association at Stanford University.

Seifert, Friedrich, and Schumacher, J.C. (1986) Cordierite-Spinel-Quartz assemblages: a potential geobarometer. Bull. Geol. Soc. Finland, v. 58, part 1, pp. 95-108.

Shearer, C.K., and Robinson, Peter (1980) Constant temperature metamorphism of an H_2O activity gradient in a pre-existing contact aureole, Petersham, Massachusetts. (abstract) EOS, v. 61, p. 389.

Shearer, Charles K. (1983) Petrography, mineral chemistry, and geochemistry of the Hardwick Tonalite and associated igneous rocks, central Massachusetts: Contrib. No. 51, (Ph.D. Thesis), Dept. of Geology and Geography, University of Massachusetts, Amherst, 210 p.

Spear, F.S. and J. Selverstone (1983) Quantitative P-T paths from zoned minerals: Theory and tectonic applications, Contributions to Mineralogy and Petrology, v.83, p 348-357.

Thompson, A.B. (1976b) Mineral reactions in pelitic rocks: II. Calculation of some P-T-X(Fe-Mg) phase relations. American Journal of Science, v. 276, p. 425-454.

Thompson, A.B. and England, P.C. (1984) Pressure-temperature-time paths of regional metamorphism, Part II: their inference and interpretation using mineral assemblages in metamorphic rocks. Journal of Petrology, v. 25, p. 929-95.

Thompson, J.B., Jr. (1972) Oxides and sulfides in regional metamorphism of pelitic schists. Proceedings of the 24th International Geologic Congress, Section 10, p.27-35. Montreal.

Thompson, P.J. (1985) Stratigraphy, structure, and metamorphism in the Monadnock Quadrangle, New Hampshire. Contribution No. 58 (Ph.D. thesis), Deptartment of Geology and Geography, University of Massachusetts, Amherst, 191p., 8 plates; 2 in color.

Thompson, P.J. (1988) Geology of Mount Monadnock. In Bothner, W.A., ed. Guidebook for Field Trips in Southwestern New Hampshire, Southeastern Vermont, and North-central Massachussetts, p. 268-273. New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, New Hampshire.

Thompson, P.J., Elbert, D. C., and Robinson, Peter (1987) Thrust nappes superimposed on fold nappes: A major component of early Acadian tectonics in the central Connecticut Valley region, New England: Geological Society of America, Abstracts with Programs, v. 19, p. 868.

Tracy, R. J., Robinson, Peter, and Wolff, Robert A. (1984) Metamorphosed ultramafic rocks in the Bronson Hill anticlinorium, central Massachusetts. American Journal of Science, v. 284, pp. 530-558.

Tracy, R.J. (1975) High grade metamorphic reactions and partial melting in pelitic schist, Quabbin Reservoir area, Massachusetts. Contribution No. 20 (Ph.D. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 127 p.

Tracy, R.J. (1978) High grade metamorphic reactions and partial melting in pelitic schist, west-central Massachusetts. American Journal of Science, v. 278, p.150-178.

Tracy, R.J. (1982) Compositional zoning and inclusions in metamorphic minerals. In Ferry, J.M., (ed.), Characterization of Metamorphism through Mineral Equilibria. Reviews in Mineralogy, v. 10, p. 105-152.

Tracy, R.J., and Dietsch, C.W. (1982) High temperature retrograde reactions in pelitic gneiss, central Massachusetts. Canadian Mineralogist, v. 20 p. 425-439.

Tracy, R.J., and Robinson, Peter (1980) Evolution of metamorphic belts: information from detailed petrologic studies. In Wones, D.J., (ed.), The Caledonides in the U.S.A., IGCP Project 27, Caledonide orogen, p.189-195. Virginia Polytechnic Institute and State University, Blacksburg, Virginia.

Tracy, R.J., and Robinson, Peter (1983) Acadian migmatite types in pelitic rocks in central Massachusetts: p. 163-173 In Atherton, M.P. and Gribble, C.D., (eds.), Migmatites, Melting and Metamorphism, Proceedings of the Geochemical Group of The Mineralogical Society, Shiva Publishing Ltd., Nantwich, England, 326 p.

Tracy, R.J., and Robinson, Peter (1988) Silicate-sulfide-oxide-fluid reactions in granulite-grade pelitic rocks, central Massachusetts. American Journal of Science, v. 288-A, p. 45-74.

Tracy, R.J., and Rye, D.M. (1981) Origin and mobility of sulfur in graphitic schists, central New England. Geological Society of America Abstracts with Programs, v. 13, p. 569.

Tracy, R.J., and Sheets, R.W. (1988) H2S-rich fluids in high-grade sulfidic schists: evidence from fluid inclusions and from fluid calculations based on metamorphic equilibria. Abstract Volume, First V.M. Goldschmidt Conference, Baltimore, p. 79.

Tracy, R.J., Robinson, Peter, and Thompson, A.B. (1976) Garnet composition and zoning in the determination of temperature and pressure of metamorphism, central Massachusetts. American Mineralogist, v. 61, p.762-775.

Tucker, R.D. (1977) Bedrock geology of the Barre area, central Massachusetts. Contribution No. 30, Department of Geology and Geography, Univerisity of Massachusetts, Amherst, Massachusetts, 132 p.

Tucker, R.D., Robinson, Peter, and Hollocher, K. (1988) U-Pb zircon, titanite and monazite dating in "basement" rocks of the Bronson Hill anticlinorium, central Massachusetts. Geological Society of America, Abstracts with Programs, v. 20, p. A216.

Tucker, R.D., Robinson, Peter, and Hollocher, K.T. (1989) The Bronson Hill volcanic arc: Old interpretations and new radiometric ages. Geological Society of America, Abstracts with Programs, v. 21, p. 71.

Weisbrod, Alan (1973) Refinements of the equilibrium conditions of the reaction Fe-cordierite = almandine + quartz + sillimanite + H_2O. Carnegie Institution of Washington Year Book, v. 72, p. 515-522.
Wolff, R.A. (1978) Ultramafic lenses in the Middle Ordovician Partridge Formation, Bronson Hill anticlinorium, central Massachusetts. Contribution No. 34 (M.S. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 162 p.
Zartman, R.E. and Leo, G.W. (1984) New radiometric ages on Oliverian core gneisses, New Hampshire and Massachusetts. American Journal of Science, v. 285, p. 267-280.
Zartman, R.E. and Naylor, R.S. (1984) Structural implications of some radiometric ages of igneous rocks in southeastern New England. Geological Society of America Bulletin, v. 95, p. 522-539.
Zen, E-an, Goldsmith, Richard, Ratcliffe, N.L., Robinson, Peter, and Stanley, R.S. (1983) Bedrock geologic map of Massachusetts. U.S. Geological Survey, Washington, D.C., scale 1:250,000.