

The eye of the petrographer, the mind of the petrologist

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ABSTRACT

To bring meaningful numbers out of metamorphic rocks, one must first understand what the numbers represent. An idealized objective in metamorphic studies is to demonstrate that coexisting minerals approached equilibrium. In a few rocks this can be done reasonably, for example, in coarse-grained nodules from kimberlite that were rapidly quenched from mantle conditions. More typical, and more interesting, are rocks showing disequilibrium by mineral zoning and reaction textures at various scales. These provide a view, if correctly interpreted, into a sequence of events and P - T conditions. In making quantitative estimates, a problem is to decide correctly which minerals were in equilibrium with which other minerals, and what types of reactions and processes actually took place. These decisions depend on the integration of both petrography and phase equilibrium theory. Failing this, a false interpretation of rock history can be made. Features in metamorphic rocks in central Massachusetts, including schists bearing strongly zoned garnets produced by four different types of reactions, and a newly described occurrence of wollastonite marble provide opportunities to explore these ideas.

In many examples, rock history can only be charted through study of the complete geologic setting at outcrop or larger scale. Particularly spectacular are well-preserved pyroxene granulites, gabbros, and diabases in the Scandinavian Caledonides that are locally transformed along shear zones into eclogite and amphibolite. Lacking knowledge of the surroundings, there would be no way to know that the original high T rocks survived eclogite-facies overprinting with few to no mineralogical traces. Their field setting demonstrates the importance of structural environment, kinetics, and fluids in producing different minerals from the same rock composition along the same P - T path.

INTRODUCTION

In petrologic studies, as in field geology, the observations that are possible are conditioned by the experience and training of the observer. Petrologic research is an attempt to define the sequence of events in the formation of the rocks, the conditions under which the events happened, and why they happened. Petrology can never be totally quantitative, i.e., “getting numbers out of rocks,” but the petrologist’s familiarity and deftness with measurements, calculations, and the constraints imposed by phase equilibria are essential. Also essential is observational and descriptive petrography from the atomic through the microscopic and macroscopic to the petrologic map scale, providing the framework for interpretations. Maps can only be made by integrating knowledge gained from submicroscopic and microscopic scales using the hand lens and unaided eye; thus the most truthful petrologic maps are done by those most skilled in both petrography and mapping. These ideas are illustrated by examples where specialized knowledge or background opened the eyes of the petrographer, leading to unexpected interpretations. These ideas are explored in three parts, each chosen to represent a different scale of observation and thinking but having in common the use of phase relations to try to understand the sequence of meta-

morphic events. The first part concerns the textural and phase relations between garnet grains and their immediate surroundings. The second part deals with features at the scale of a whole thin section in a wollastonite marble where variations in local bulk compositions are shown to have a dramatic effect on phase relations. The third section covers relations between rocks at the scale of a large outcrop or a mountainside, where an understanding of the total setting is vital to understanding rock history.

A theme of this paper is “getting numbers out of rocks” and concerns the need for careful and thorough petrographic observations in thin sections and in the field if one is to make meaningful maps of igneous and metamorphic rocks and if one is to combine theory and experiment to make useful petrogenetic models. In the game of getting numbers out of rocks, much of my attention has been focused on the central Massachusetts Acadian metamorphic high, shown on a metamorphic map of New England in Figure 1 and in closer detail in Figure 2, with metamorphic zones ranging from chlorite zone (C) in the extreme northwest through Zone I (kyanite-muscovite-staurolite) through Zone VI (sillimanite-orthoclase-garnet-cordierite). In this whole area I am familiar with only two “direct” numerical determinations. The first (Fig. 3) is a slightly flawed number 8 found in a granulite facies

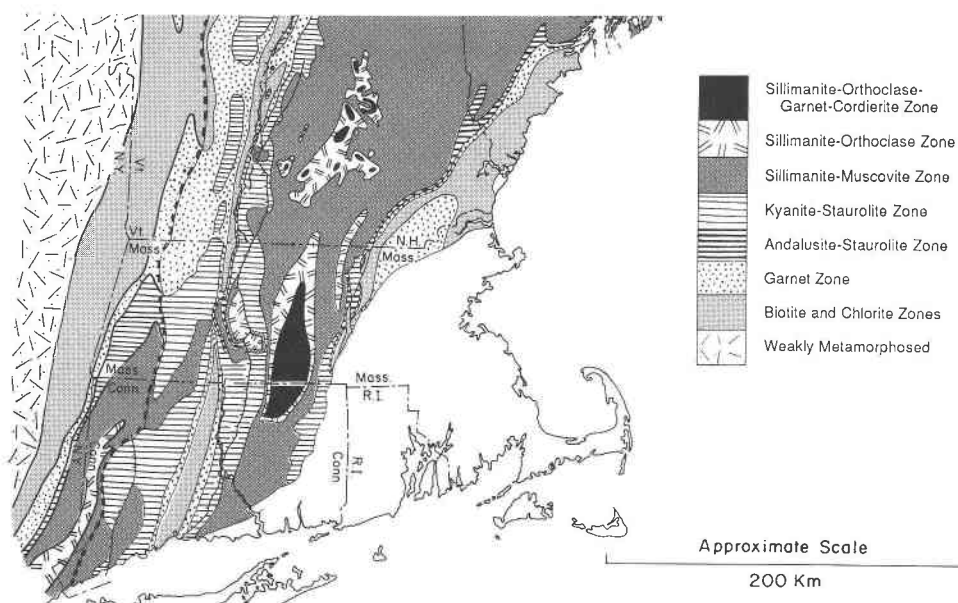


Fig. 1. Generalized map of Acadian and Taconian metamorphic zones in southwestern New England. Modified from Osberg et al. (1989) based on a compilation by Robinson (1983) and an unpublished compilation, Metamorphic Map of the U.S.-Canadian Appalachians, by G.W. Fisher and W.E. Trzcinski. Heavy dashed line represents the approximate boundary where peak metamorphism is Acadian to the east and Taconian to the west.

pegmatite formed by spherical aggregates of graphite platelets enclosed in a crystal of orthoclase. The second (Fig. 4) is a much larger number 2 discovered in a garnet-rich schist in the same area by Elaine Padovani. In considering the question of geothermometry and geobarometry in such a region, it is natural to turn to the compositions of garnets in pelites as well as to consideration of garnet stability and geothermobarometry in a newly described wollastonite-bearing marble.

INTERPRETATION OF GARNET COMPOSITIONS IN METAMORPHIC ROCKS

Aside from its aesthetic values, garnet has captured the attention of quantitative metamorphic petrologists for three reasons: (1) It holds very high concentrations of Fe and Mn relative to most other coexisting minerals, and hence distribution coefficients are sensitive to temperature. (2) It has high density compared with most coexisting minerals, and hence reactions involving it are likely to be sensitive to pressure changes. (3) Elemental diffusion is relatively slow in garnet compared to most other minerals, except at the highest temperatures, so that disequilibrium features are commonly preserved, and these provide a key, if properly interpreted, to the changing conditions of metamorphism. These disequilibrium features are the subject of this part of the paper.

The mention of metamorphic garnets in central Massachusetts opens a whole can of worms (Fig. 5). In this case the heads of the worms are the cores of the garnets,

and the tails are the rims. The roman numerals correspond to the different metamorphic zones in central Massachusetts, and the compositional trajectories are different in each of the metamorphic zones. In the subsections that follow, each of the major types of garnet zoning found in central Massachusetts is discussed, the importance of careful petrographic observations to deduce equilibria and reaction mechanisms are emphasized, and the particular pitfalls involved in making geothermometric interpretations are pointed out.

In the discussion that follows, the terms "prograde" and "retrograde" appear frequently because they are part of the everyday language of metamorphic petrologists, yet they are fraught with ambiguity in the context of complex P - T paths. A prograde path is commonly thought of as one involving a temperature increase, a dehydration, or both. Yet prograde metamorphism to a high-pressure assemblage might involve mainly a pressure increase with little or no increase of temperature. To consider prograde as the direction in which minerals grow would be entirely erroneous because it is quite possible for some minerals to grow with decreasing temperature. The ambiguity may be epitomized by considering a hypothetical equilibrium $A = B$ with a positive P - T slope. P - T trajectories, also with positive slopes, may be drawn to cross this reaction in different up-temperature directions so that A goes to B or alternatively so that B goes to A . Both senses of the reaction could independently be described as prograde. Thus, the best that one can say is that prograde metamorphism involves a movement more or less directly

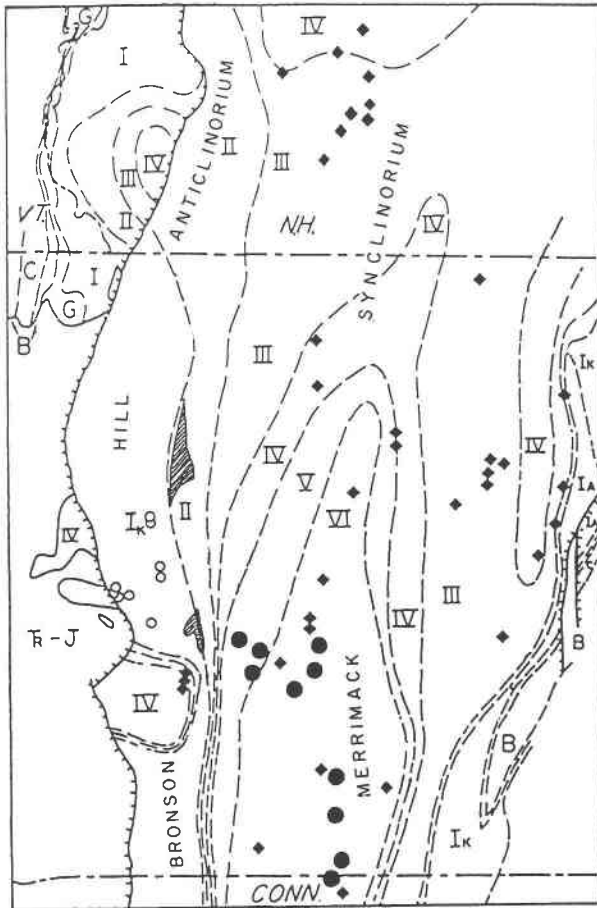


Fig. 2. Generalized map of Acadian and younger metamorphic zones in west-central Massachusetts and adjacent states. Modified from Schumacher et al. (1989). Zones I to VI after Tracy et al. (1976). Diamonds indicate locations of sillimanite pseudomorphs after andalusite; closed circles indicate orthopyroxene in metamorphosed mafic and intermediate igneous rocks; open circles indicate locations of relict granulite-facies assemblages in Pelham gneiss dome; diagonal shading indicates zones of retrograde metamorphism at New Salem and Quabbin Hill. The area of zone I between the retrograde zones and the Mesozoic Connecticut Valley fault appears to be an area of intense late Paleozoic metamorphic and tectonic overprinting.

away from the P - T origin in a way that may or may not involve significantly increasing pressure.

Zoning produced during prograde garnet growth

In the lowest grade zones in central New England, the garnets are commonly well formed, are nearly euhedral, and show a characteristic zoning pattern with Mn-rich cores zoning to less Mn-rich rims with first an increasing Mg/Fe ratio and then a decreasing ratio (Figs. 6 and 7). With some reservations, these may possibly be taken as examples of garnet growth under the simplest conditions, where diffusion in garnet is minimal to nonexistent. Only the surface of the garnet is in equilibrium with the matrix.

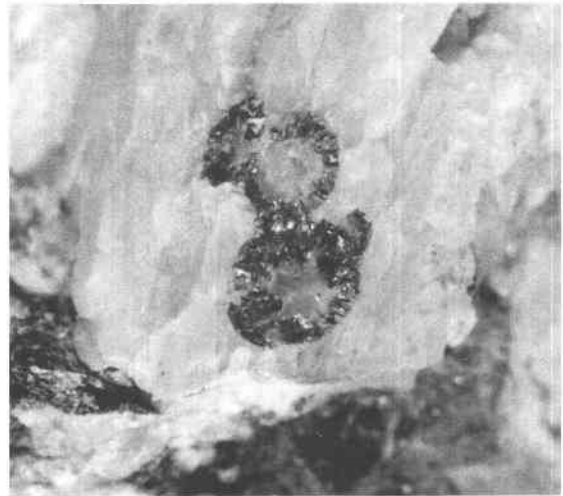


Fig. 3. The number 8 formed by spherically arranged graphitic platelets enclosed in orthoclase of a pegmatite in Zone VI, central Massachusetts. For location see Robinson et al. (1986), stop 7.

A sequence of garnet compositions can be studied that records garnet growth. Unless inclusions are preserved, only garnet rim and matrix compositions can be used for geothermometry. Inclusions may reequilibrate within garnet, but this requires diffusion. Note that garnet growth need not be prograde but may be retrograde, depending



Fig. 4. The number 2 formed by scattered garnet in sillimanite-orthoclase-garnet-cordierite-biotite schist from Zone VI, central Massachusetts. For location see Robinson et al. (1986), stop 11.

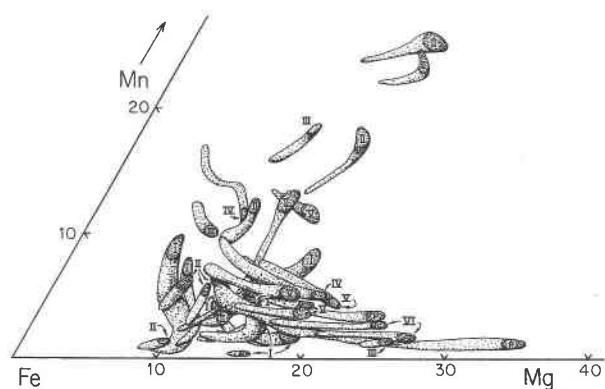


Fig. 5. Summary plot of compositional trajectories of zoned garnets from central Massachusetts in terms of Fe, Mn, and Mg. Numerals I–VI refer to metamorphic zones in central Massachusetts, and P refers to partially resorbed granulite-facies garnet from the Pelham dome. From Robinson et al. (1989).

critically on the P - T path and the reaction isopleths. This concept of growth, under either a prograde net-transfer reaction or a retrograde net-transfer reaction, is illustrated in Figure 8. In aluminous rocks, Mn is most strongly partitioned into garnet as compared with any other phase. For this reason, it is the element that appears most abundantly in early formed garnet and is also the last to be depleted from garnet in any garnet resorption reaction in

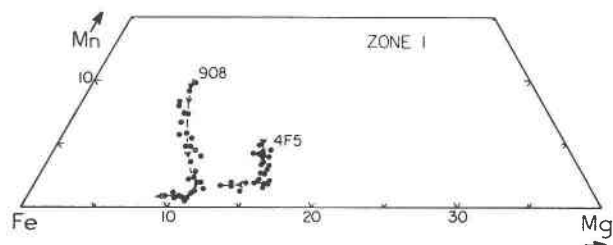


Fig. 7. Compositional trajectories of two typical zoned garnets from zone I, central Massachusetts in terms of Fe, Mn, and Mg. Core compositions are at most Mn-rich points next to specimen number. Specimen 908 is staurolite-garnet-biotite-muscovite schist; specimen 4F5 is similar but also with kyanite. From Tracy et al. (1976).

which the garnet reequilibrates. (It is interesting that in some skarns containing mostly andradite garnet, however, Mn is not partitioned into garnet but into pyroxene.)

What are the means by which examples of prograde growth zoning in garnet are destroyed or altered to obscure their story? First of all, prograde or retrograde diffusional ion exchange between garnet and its matrix (Fig. 9), here taken to be biotite, could destroy the near-rim record. A second possibility is a retrograde (or possibly prograde) diffusional net-transfer reaction (Fig. 10) by which the outside of the garnet is partially resorbed while those components retained by garnet actively diffuse to-

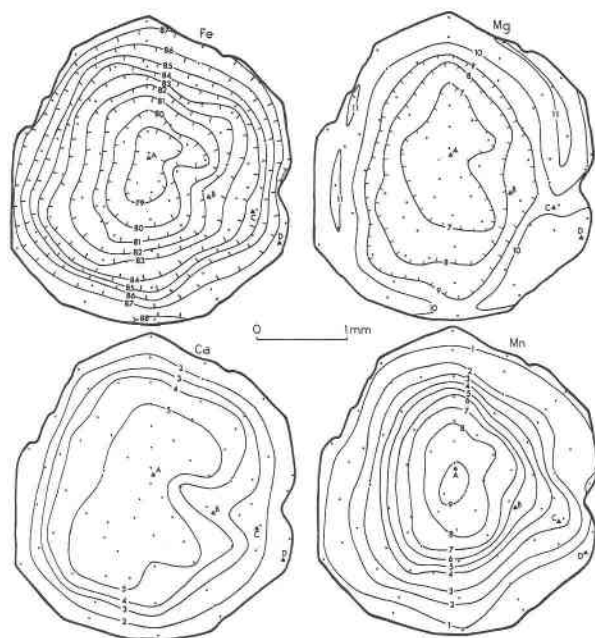


Fig. 6. Compositional maps of a zoned garnet in sample 908 from staurolite-garnet-biotite-muscovite schist of zone I, central Massachusetts. Contours based on $\text{Fe}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$, etc., are Fe 79–88%, Mg 7–11%, Ca 2–5%, and Mn 1–9%. Adapted from Tracy et al. (1976).

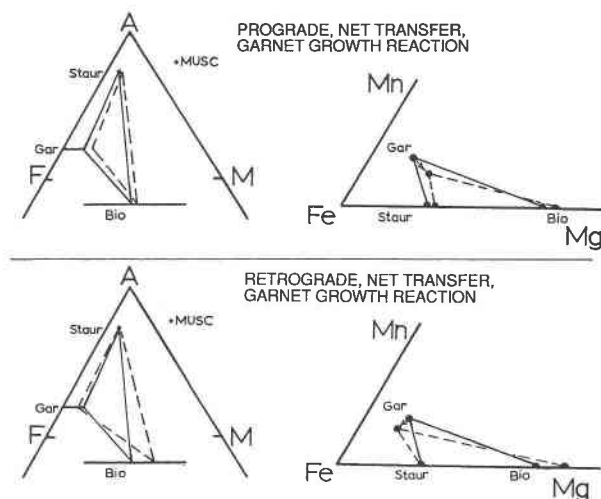


Fig. 8. AFM muscovite projections and Fe-Mn-Mg plots showing the nature of prograde and retrograde garnet-growth reactions. In each example, reaction proceeds from the solid to the dashed tie lines, and in each example there is a net increase in the modal amount of garnet. The prograde reaction involves dehydration but may not involve significant temperature increase, as shown by the nearly parallel tie lines in the AFM view. The retrograde reaction does involve a significant temperature drop, as shown by the crossing tie lines in the AFM view. Modified from Roll (1987).

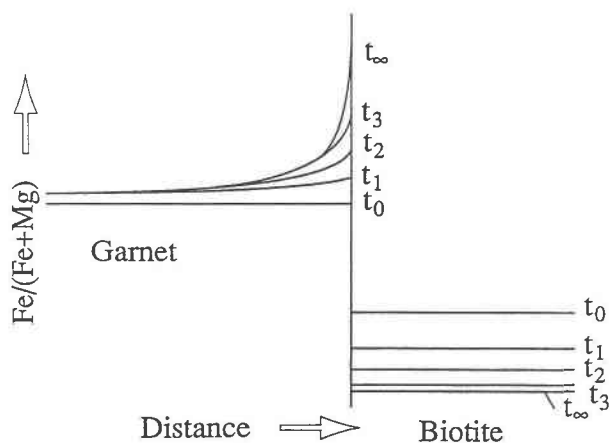


Fig. 9. Hypothetical composition profiles of garnet and biotite for retrograde ion exchange for time t_0 through t_∞ . Diffusion in garnet is a strong function of temperature, whereas biotite is assumed to homogenize throughout the period of equilibration with garnet. From Spear (1991).

ward the interior. A third possibility, recently explored by Spear and Florence (1991), is internal diffusive modification of the primary zoning profile, which would produce a new and obviously false record of original garnet growth compositions. In a thermal model (Fig. 11) of diffusional ion exchange between garnet and matrix biotite done by Spear and Florence (1991), using diffusion data from Cygan and Lasaga (1985) and assuming a quasi-infinite reservoir of biotite host, we can see the limits of diffusional reequilibration at a cooling rate of 10°C per m.y. beginning at 725°C and tailing down to a limit close to 500°C .

Karabinos (1983) has described garnet-zone pelites in

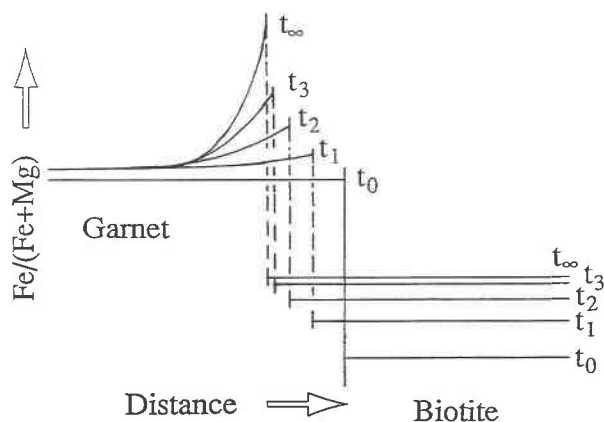


Fig. 10. Hypothetical composition profiles of garnet and biotite for a retrograde net-transfer garnet resorption reaction for time t_0 through t_∞ . Diffusion through garnet is a strong function of temperature, whereas biotite is assumed to remain homogeneous throughout the period of garnet equilibration. Because of resorption, the garnet interface retreats in favor of biotite. Modified from Spear (1991).

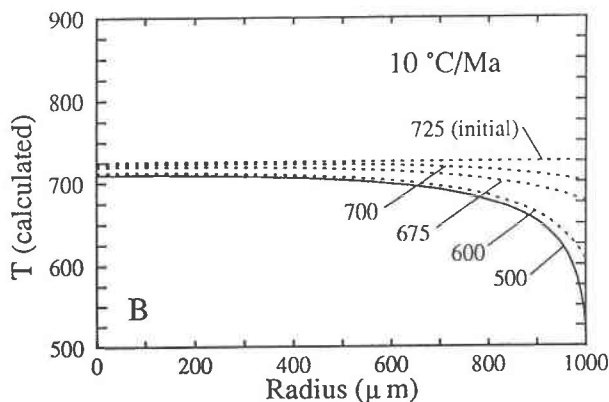
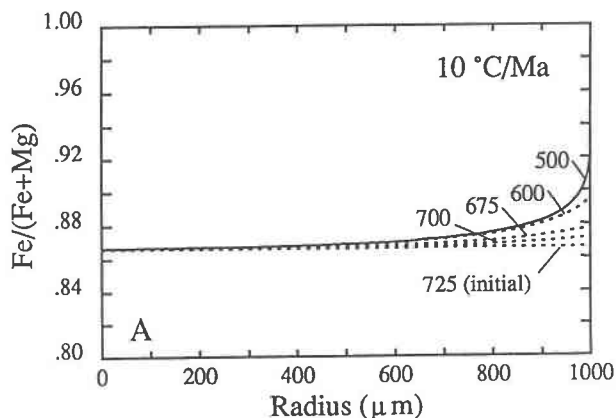


Fig. 11. Profiles of garnet composition (A) and calculated garnet-biotite temperature (B) at temperatures from 725°C to 500°C , assuming diffusional retrograde ion exchange between garnet and a major biotite reservoir. Note that although the interior of the garnet is hardly changed in composition, $T(\text{calculated})$ drops slightly as a result of a decrease in $\text{Fe}/(\text{Fe} + \text{Mg})$ of matrix biotite composition as it exchanges with the garnet rim. From Spear and Florence (1991).

Vermont that preserve two cycles of apparent prograde garnet-growth zoning (Fig. 12). These are separated by an Mn-enriched rim (Fig. 13) believed to have been produced during an intervening period of garnet resorption. Karabinos argues that the garnet itself gained an Mn-rich rim by a resorption reaction (Fig. 14) in which garnet was diffusively depleted in Mg and Fe for the production of chlorite. My own interpretation is that diffusion in garnet at the grade experienced by these rocks, which appears to have been at 500°C or less as demonstrated by apparent delicate preservation of prograde growth zoning, was too slow to permit such retrograde diffusive reequilibration. Indeed, using available diffusion coefficients, Karabinos could not account for it. I would argue that the garnet was partially resorbed by surface dissolution to produce an Mn-enriched retrograde assemblage of hydrous ferromagnesian silicates, mainly chlorite. The Mn-rich intermediate garnet rim then grew by prograde growth from the matrix at a very early stage in the second episode of garnet growth in this rock.

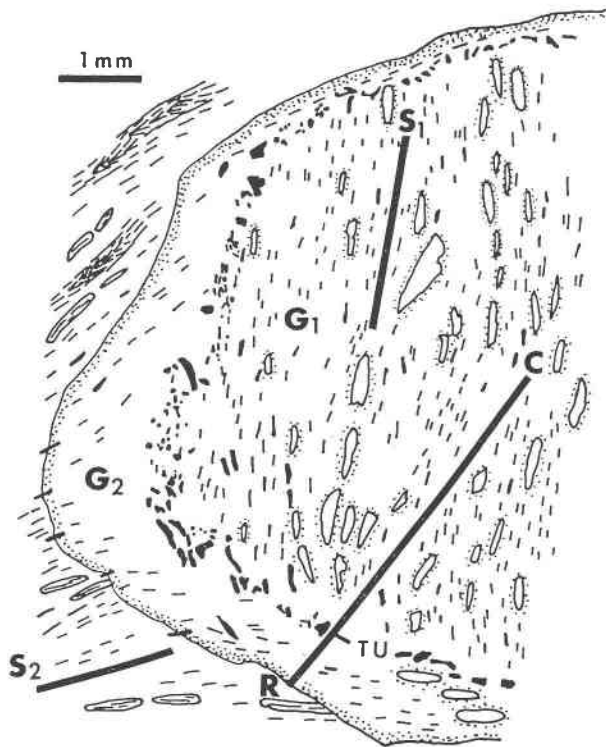


Fig. 12. Textural map of garnet from eastern Vermont, showing two cycles of prograde growth. From Karabinos (1983). Line C-R is the compositional profile shown in Figure 13, and TU is the point of textural unconformity supposed by Karabinos to be the outer limit of the original garnet at the end of the stage of retrogression. I argue that resorption proceeded slightly deeper into the crystal.

Hollocher (1981, 1987) has described just such an example of surface dissolution of garnet to produce chlorite (Figs. 15 and 16) in the New Salem retrograde zone, central Massachusetts (see Fig. 2). In this case there are detailed analyses of the full spectrum of relict primary zon-

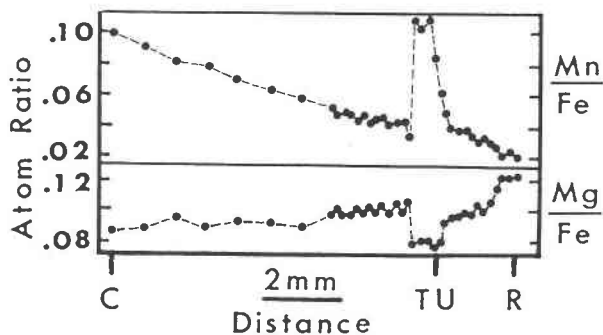


Fig. 13. Composition profile C-R in the garnet shown in Figure 12, modified from Karabinos (1983). Karabinos suggests that the zone with high Mn/Fe ratio was produced during retrograde garnet resorption; I believe it was produced early in the second stage of garnet growth.

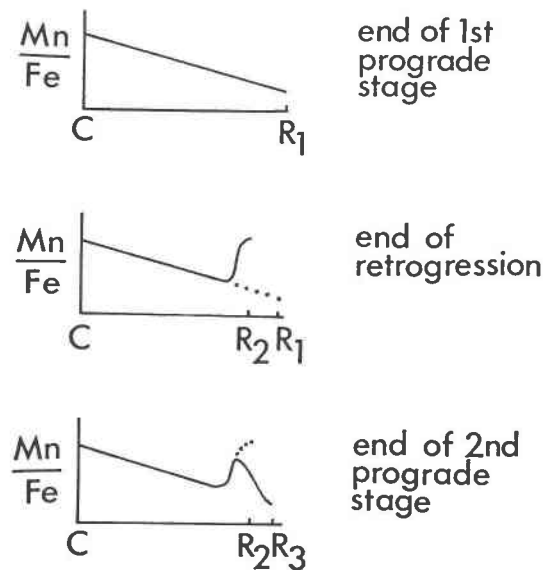


Fig. 14. History of garnet growth, resorption, and growth for the sample illustrated in Figures 12 and 13, as portrayed by Karabinos (1983).

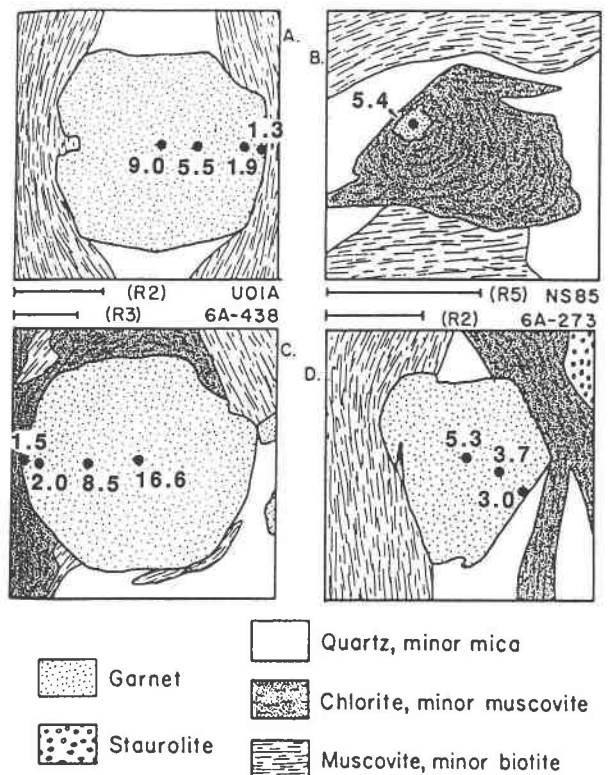


Fig. 15. Variably resorbed garnets from the New Salem retrograde zone studied by Hollocher (1987). Numbers indicate the variable spessartine content obviously left over from the period of initial prograde garnet growth.



1 mm
Sample NS 71
Retrograded Garnet

Fig. 16. Drawing of partially resorbed garnet from New Salem retrograde zone by Hollocher (1981). Much of garnet is replaced by chlorite, but where it remains it retains the spessartine contents from the much earlier stage of prograde garnet growth.

ing in garnets that are partially dissolved in a regional hydration reaction. This demonstrates that the volume of garnet that was in equilibrium with the matrix during the retrograde reaction was zero, but the surface of the garnet was being actively corroded to produce new matrix minerals.

In a more complex situation, garnet is growing, but diffusion is causing reequilibration within garnet and progressively destroying the earlier tape recorder message.

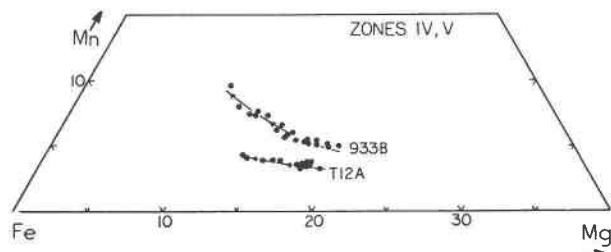


Fig. 18. Compositional trajectories of two typical zoned garnets from zones IV and V, central Massachusetts, in terms of Fe, Mn, and Mg. Core compositions are at most Mg-rich points next to specimen number. Both specimens are sillimanite-orthoclase-garnet-biotite schists in which garnet has been resorbed by the reaction garnet + orthoclase + H_2O = sillimanite + biotite. From Tracy et al. (1976).

Spear (1991) has recently explored this situation in detail and shown that the last grown part of the prograde zoned garnet, which is likely to record the highest temperature, is also the first to be destroyed by retrograde ion exchange or resorption reactions.

Zoning produced by retrograde garnet resorption

In regions of higher grade metamorphism, the record of earlier history, if there was one, was destroyed by diffusion, and the garnets are largely unzoned. What happens beyond the peak of metamorphism depends critically on environmental conditions and whether garnet resorption reactions take place. The garnet illustrated in Figure 17 is an irregularly shaped one from the sillimanite-orthoclase-muscovite zone in central Massachusetts. It appears to have a continuous exterior rim of Mn and Fe enrichment and Mg depletion wherever the garnet is touching the matrix and regardless of what phase is directly adjacent in the matrix. This is interpreted to mean that the exterior part of the garnet was equilibrated with a pervasive phase, probably an aqueous fluid, that was in

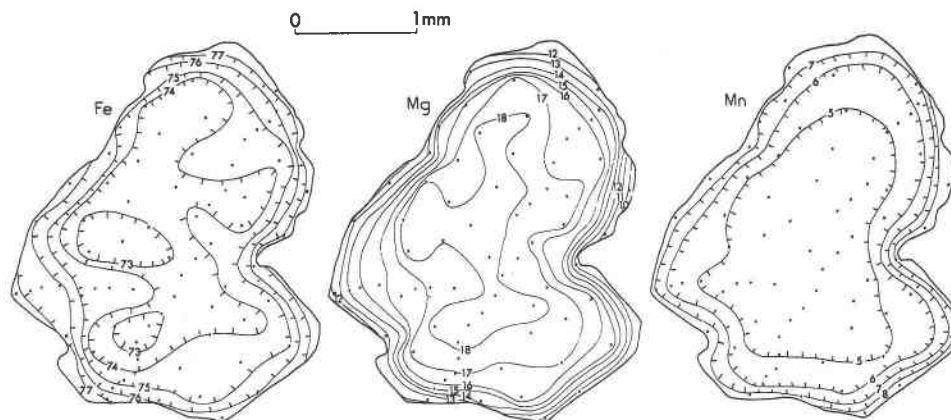


Fig. 17. Composition maps of a garnet in sample 933B from the sillimanite-orthoclase-muscovite zone (Zone IV), central Massachusetts. Contours based on $Fe/(Fe + Mg + Mn + Ca)$, etc., are Fe 73–77%; Mg 12–18%; Mn 5–8%. Continuous rims of higher Fe, lower Mg, and higher Mn around this irregular garnet are characteristic of this zone and indicate a continuous retrograde, net-transfer, hydration resorption reaction through the medium of an interstitial fluid. Modified from Tracy et al. (1976).

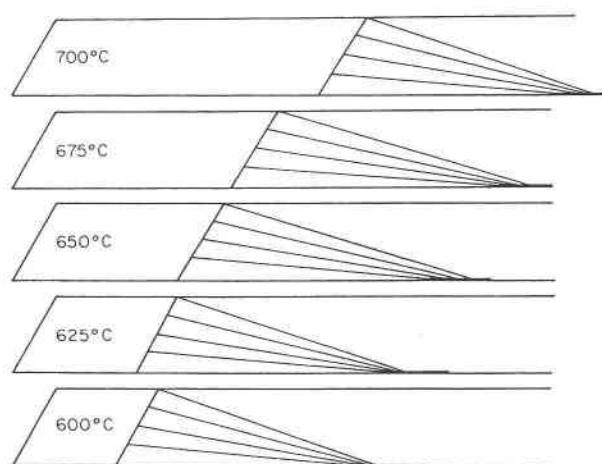


Fig. 19. A series of model sections at constant T , P , and humidity showing equilibrium compositions of coexisting garnet and biotite in an Fe-Mn-Mg ternary. From Robinson et al. (1982).

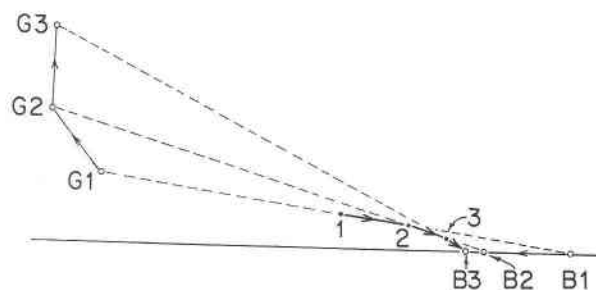


Fig. 21. Fractional garnet resorption illustrated schematically as a series of steps. At stage 1 bulk composition 1 consists of garnet G1 and biotite B1. The inner part of garnet G1 ceases to equilibrate with matrix biotite and the effective bulk composition moves to 2, where it now consists of garnet G2 and biotite B2. Still more of the garnet ceases to equilibrate with matrix, moving the bulk composition to 3, where it consists of garnet G3 and biotite B3. The process continues until the effective bulk composition contains no garnet, i.e., that diffusion in garnet is so slow that the mineral is incapable of reacting with biotite. From Robinson et al. (1982).

easy communication with other matrix ferromagnesian minerals. The fact that the Mn-content increases outward (Fig. 18) suggests the action of a net-transfer, fractional, garnet-resorption, retrograde hydration reaction, as illustrated in Figure 10, such as $\text{GAR} + \text{KSP} + \text{H}_2\text{O} = \text{SILL} + \text{BIO}$. Commonly the amount of fluid available for such a reaction would be limited and, because of limited diffusion, would involve only a limited volume of the garnet, while involving virtually all of the matrix. The reaction is fractional because the garnet interior is progressively removed from participation in the retrograde reaction, so that a decreasing volume of the exterior is subjected to more and more extreme reactions. The effective volume of reequilibration of garnet continues to fall, until it approaches zero and the garnet ceases to be

in equilibrium with the matrix minerals, although, as pointed out above, its surfaces may still be subject to surface corrosion that could contribute to the bulk chemistry of matrix assemblages.

In 1982 we modelled this process in a very simple way in an Mn-Fe-Mg ternary (Robinson et al., 1982) that is now being duplicated more elegantly in computer programs by Frank Spear and his students. The first step was to select a series of equilibrium sections (Fig. 19) corresponding to a series of changing metamorphic conditions of temperature, pressure, and humidity. The figure shows putative garnet-biotite tie lines at a series of temperatures and a series of progressively more humid conditions with falling temperature. A similar series of templates could

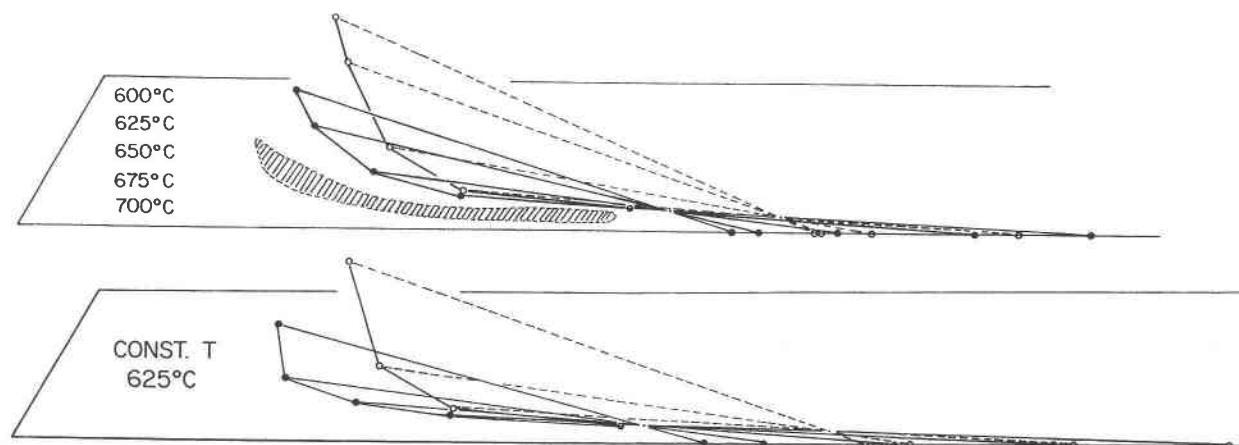


Fig. 20. Model equilibrium garnet composition Fe-Mn-Mg trajectories for two different initial garnet-biotite ratios during progress of the hydration resorption reaction $\text{garnet} + \text{orthoclase} + \text{H}_2\text{O} = \text{sillimanite} + \text{biotite}$. In the upper diagram the hydration takes place with falling temperature from 700 to 600 °C; in the lower diagram at constant temperature of 625 °C. In each case the equilibrium garnet composition trajectory is longer and more curved for the more biotite-rich bulk composition. From Robinson et al. (1982).

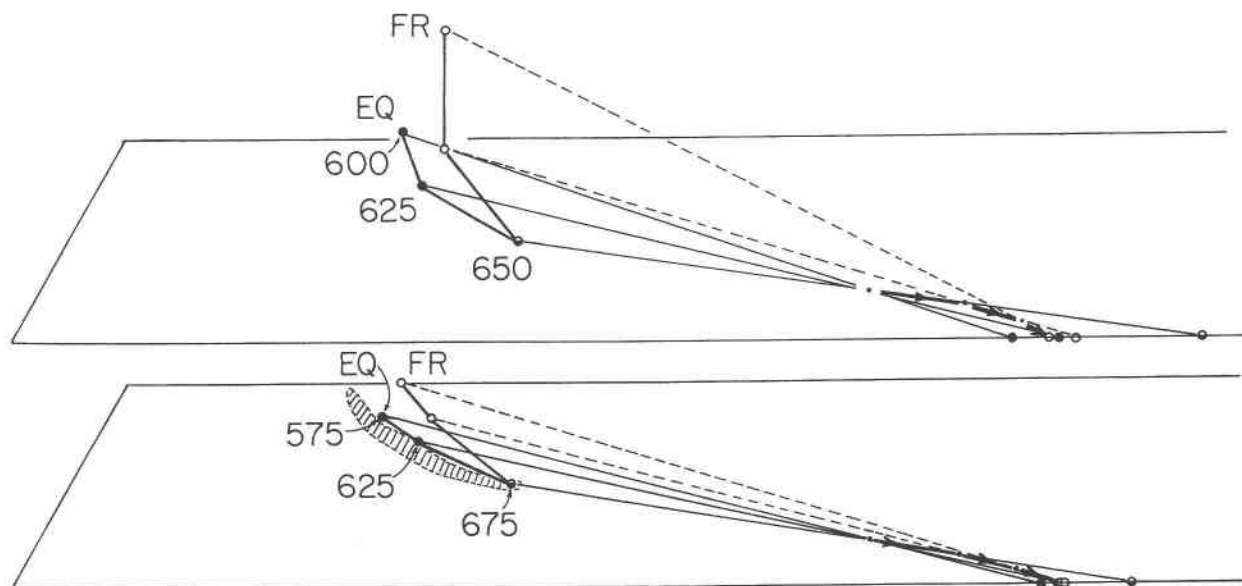


Fig. 22. Comparisons of model equilibrium composition trajectories and fractional crystallization trajectories during retrograde garnet resorption for two paths of cooling and hydration starting with the same initial bulk composition. Note that the fractional paths are more strongly curved than the equilibrium ones. The actual zoning path for the garnet in 933B (Fig. 18) is shown for comparison. From Robinson et al. (1982).

be constructed for retrograde hydration at constant temperature. Using such templates, a series of equilibrium garnet-biotite tie lines (Fig. 20) has been constructed for a series of changing conditions for two different bulk compositions, both originally on the same garnet-biotite tie line at the peak of metamorphism. One can see that the equilibrium paths of garnet for the two bulk compositions are quite different and that garnet composition changes more for bulk compositions that are poorer in garnet. This simple geometrical concept was well understood by Ferry and Spear (1978) when they designed their geothermometric experiments. In order to avoid disequilibrium problems with zoned garnet, they designed their experiments with large ratios of garnet to biotite so that there would be little tendency for composition change in garnet and they could study composition changes in biotite, which reached equilibrium easily. It should be emphasized that the equilibrium paths illustrated in this figure are only indirectly related to the zoning in natural resorbed garnet, which is produced by a fractional disequilibrium process.

The fractional aspect is illustrated in Figure 21 in a series of steps involving alternate equilibration and fractionation, in which progressively less and less garnet participates, thus decreasing the garnet to biotite ratio. As the fraction of participating garnet decreases, the change in composition of the remaining portion becomes more and more extreme, while changes in biotite composition decrease dramatically. For this reason the fractional crystallization paths (Fig. 22) are more strongly curved than the equilibrium ones, but the exact details depend critically on the extent of fractionation. Obviously, this type

of reaction ceases when the diffusion is so slow that garnet cannot participate at all.

A serious consequence of this type of reaction, in cases where garnet and biotite start out in roughly equal proportions, is that none of the original biotite composition

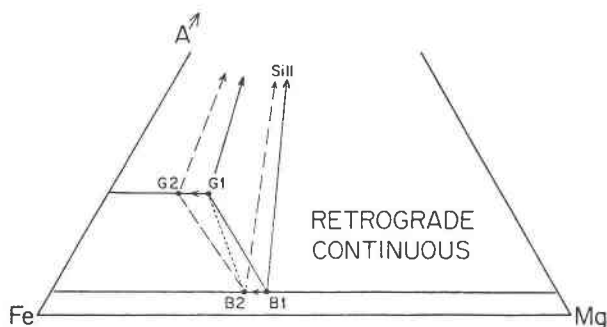


Fig. 23. Consequences for geothermometry of a diffusional retrograde continuous net-transfer garnet-resorption hydration reaction, here viewed in an orthoclase AFM projection. As the resorption reaction proceeds, equilibrium garnet and biotite compositions both move toward Fe enrichment. The original garnet composition G1 is protected from reaction in the interior, whereas all or nearly all of the original biotite B1 in the matrix is destroyed, thus removing the possibility of directly determining the original peak metamorphic temperature. Garnet G2 and matrix biotite B2 may be used to estimate the temperature under which the retrograde reaction ceased. When garnet G1 is paired with biotite B2, with which it was never in equilibrium, a completely false temperature estimate is made, which can normally be much higher, even hundreds of degrees, than the original peak temperature. From Robinson et al. (1982).

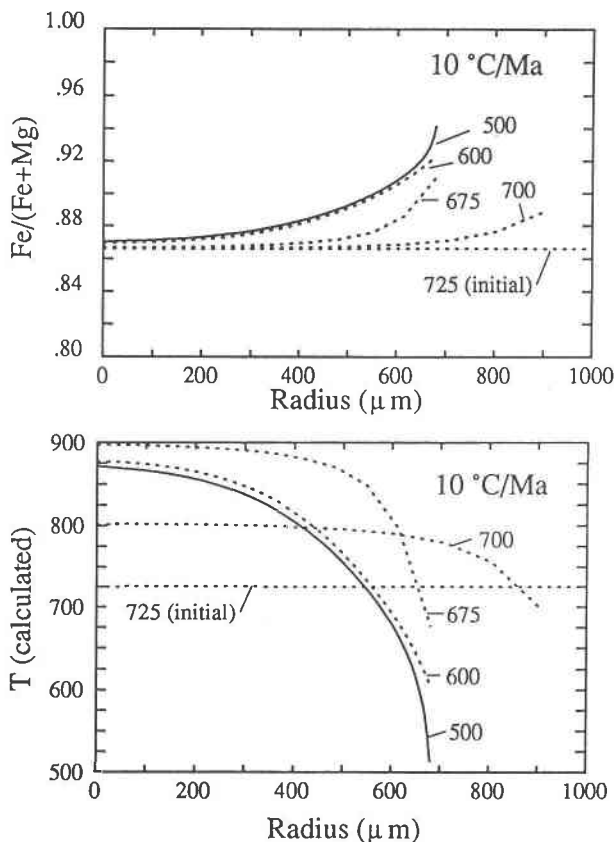


Fig. 24. Profiles of garnet composition and calculated garnet-biotite temperature at temperatures from 725 to 500 °C. In the particular model produced, the bulk garnet/biotite ratio began near 1:1, and a diffusional retrograde net-transfer garnet-resorption reaction with a major biotite reservoir was assumed to take place between 725 and 675 °C. This was followed by pure ion exchange from 675 and 500 °C. Note that although the interior of the garnet is hardly changed in composition, $T(\text{calculated})$ for the core rises dramatically from 725 to 900 °C as a result of a major increase in $\text{Fe}/(\text{Fe} + \text{Mg})$ of matrix biotite as the garnet rim is resorbed. This is followed by a decline from 900 to 875 °C as the biotite decreases in $\text{Fe}/(\text{Fe} + \text{Mg})$ during ion exchange. From Spear and Florence (1991).

remains that was originally in equilibrium with the garnet core, unless a grain remained in some unusual protected location. Thus, when an unaltered garnet core composition, such as G1 in Figure 23, is paired to make a temperature estimate with a highly altered matrix composition such as B2, the result is always higher than the original peak temperature. In some cases the result may be hundreds of degrees higher, as pointed out by Robinson et al. (1975) and illustrated in a recent calculated profile (Fig. 24) by Spear and Florence (1991).

The style of reaction just discussed is illustrated in a spectacular way by zoned garnets studied in detail in the Pelham gneiss dome, Massachusetts (Fig. 25), by Roll (1987). These rocks occur in the Mount Mineral Formation within the area of Zone I (kyanite-muscovite-

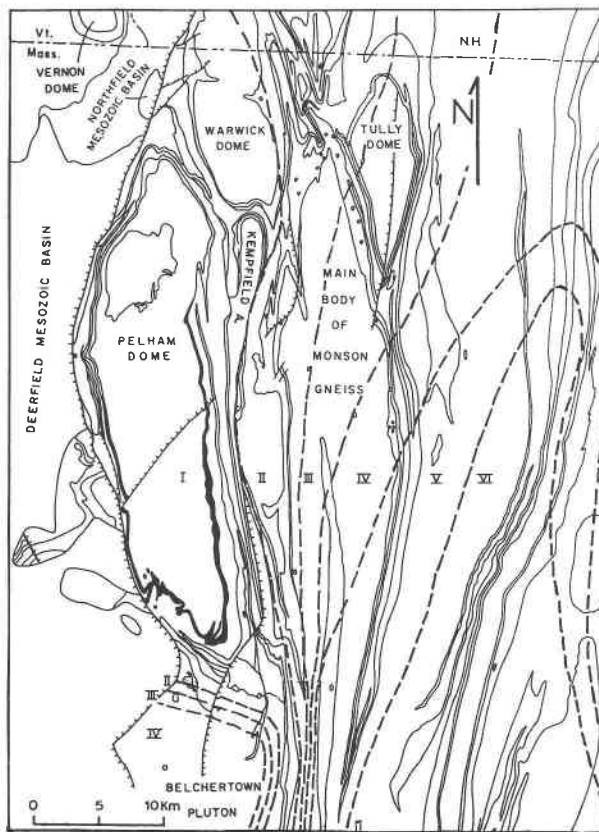


Fig. 25. Map showing the Pelham dome relative to the regional geology and metamorphic zones of west-central Massachusetts. Narrow belt shown in black in the southern part of the dome is the Mount Mineral Formation containing relict granulite-facies assemblages. Adapted from Roll (1987).

staurolite), and they retain relics of an older granulite-facies metamorphism shown by relict high orthoclase with sillimanite inclusions, garnet cores with pyrope content up to 35%, coarse sillimanite, and relict associations of

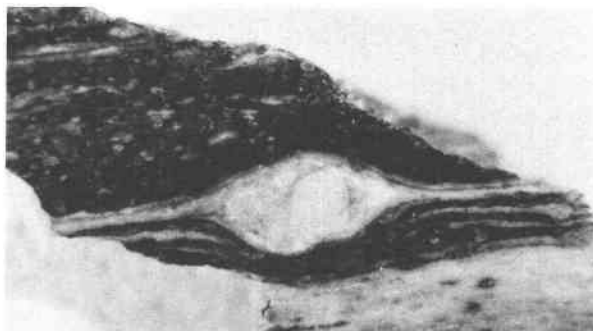


Fig. 26. Schist sample Q-P25c from the lower part of the Mount Mineral Formation, east limb of Pelham dome, central Massachusetts. Contains a sheared relic of high orthoclase about 2.5 cm long with tails of secondary plagioclase and muscovite set in a matrix of schist rich in muscovite, biotite, kyanite, and garnet.



Fig. 27. Photomicrograph of schist sample W67B from the upper part of the Mount Mineral Formation, east limb of Pelham dome. To left is the margin of a high orthoclase megacryst enclosing quartz, plagioclase, and sillimanite (stout needles far left). The immediately surrounding matrix (right) contains muscovite, biotite, and coarse kyanite (large grain close to orthoclase right center).

garnet with coarse rutile. These occur in a variety of rocks with a mylonitic character related to late shearing in the dome, and they commonly occur side by side with rocks that have been so completely reconstituted that they show few or no traces of the older metamorphism and only features of late mineral growth.

The sample in Figure 26 contains a sheared relic of high orthoclase with tails of secondary plagioclase and muscovite set in a matrix of schist rich in muscovite, biotite, kyanite, and garnet. Figure 27 shows the margin of a high orthoclase megacryst enclosing sillimanite. The matrix in the immediate area contains muscovite, biotite, and coarse kyanite. In Figure 28, large garnet megacrysts surrounded by orthoclase and sillimanite are enclosed by thick rims of biotite produced by retrograde hydration reactions. One of the samples studied in detail was a my-



Fig. 28. Schist sample 160C originally collected by L.D. Ashwal in 1972 from the middle of the Mount Mineral Formation, east limb of the Pelham dome. Garnet megacrysts up to 1.5 cm long, surrounded by orthoclase and sillimanite, are enveloped by thick rims of biotite produced by retrograde hydration reactions.

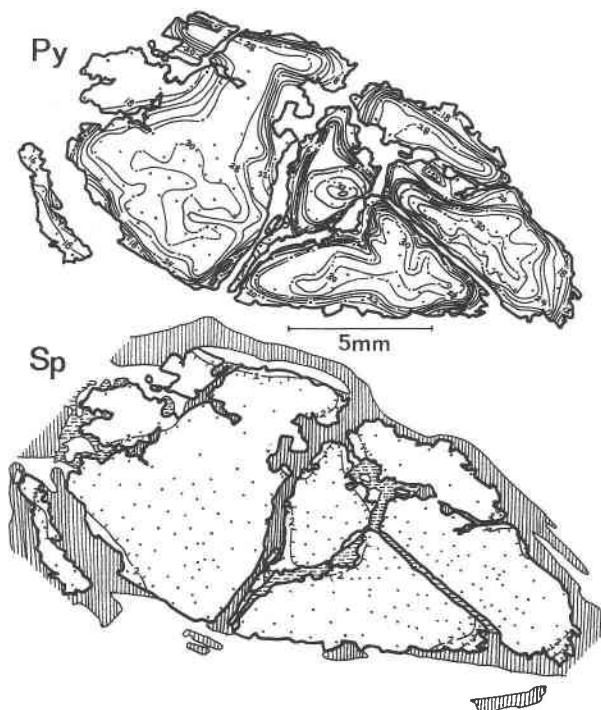


Fig. 29. Garnet composition maps from sample 160X (see previous caption) with analytical points indicated by dots. **Top** map shows contours of pyrope content ranging from 34% down to 12% at 2% intervals. **Bottom** map shows contours of spessartine content of 2% and 3%. Vertical ruling indicates areas of brown biotite; dashed horizontal ruling indicates areas of pale green biotite. Adapted from Roll (1987).

lonitic garnet gneiss, sample 160X, from which the garnet composition map in Figure 29 and the composition trajectory diagram in Figure 30 were produced. These garnets show extreme zoning in pyrope content and Fe/Mg ratio and a slight marginal increase in Mn. The pyrope composition map shows that the large garnet was being resorbed along a series of cracks that look like fjords on some island off the Norwegian Coast. The relict cores in sample 160X contain 35% pyrope and 1.1% spessartine. From the core to the rim, the pyrope content decreases continuously to 10–11% and spessartine increases smoothly to a maximum of about 3%. A second garnet from a similar rock (Fig. 31) is a more regular egg-shaped porphyroclast showing an irregular pyrope-rich interior and a thin pyrope-poor, spessartine-enriched rim and is more amenable to diffusional modelling.

Roll modeled the time required to produce such resorption rims at 580 °C, 5 kbar using three different sets of diffusion coefficients (Fig. 32). Under such conditions it would take 40 m.y. for diffusional penetration of the garnet to a depth of 0.05 cm, rather longer than appeared to be available during Acadian metamorphism. Based on these results, she suggested that the retrograde resorption zoning might have been more easily produced in two episodes, one during cooling at the end of an early, probably

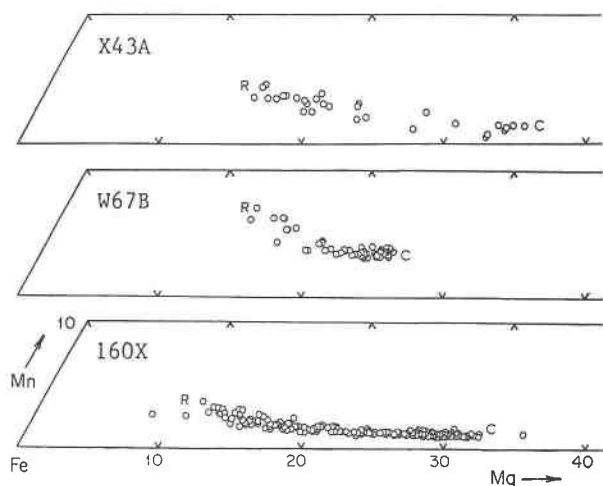


Fig. 30. Compositional trajectories of zoned garnets from the Mount Mineral Formation, Pelham dome, central Massachusetts, in terms of Fe, Mn, and Mg. C indicates core, R indicates rim. Composition maps for these garnets are given in Figure 29 (160X), Figure 31 (W67B), and Figure 33 (X43A). All come from porphyroclastic schists with other relics of original near granulite-facies conditions, in which garnet has been resorbed by the reaction garnet + orthoclase + H_2O = aluminum silicate + biotite. From Roll (1987).

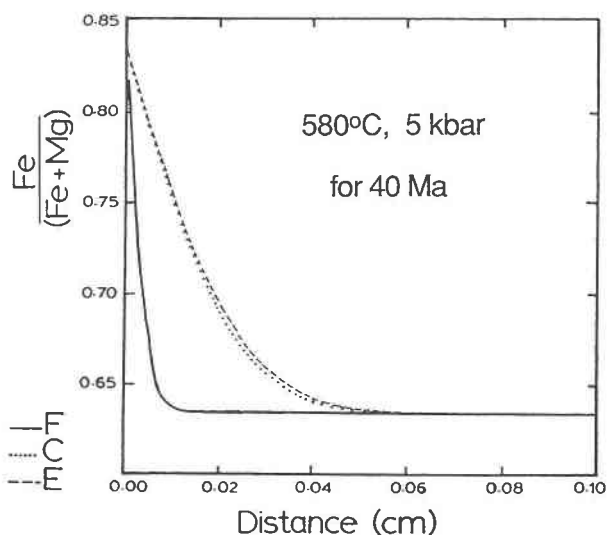


Fig. 32. Model penetration distances produced in homogeneous garnet 2 mm in diameter of 35% pyrope composition by retrograde reequilibration at 580 °C, 5 kbar for 40 m.y. Gradients are based on three sets of diffusion coefficients: F = Freer (1979); C = Cygan and Lasaga (1985); E = Elphick et al. (1985). Under such conditions it would take 40 m.y. for diffusional penetration of the garnet to a depth of 0.05 cm. The model does not take into account the movement of the garnet interface produced by resorption. From Roll (1987).

Proterozoic, high-grade metamorphism and a second during Acadian retrograde hydration. These views require reinterpretation based on new detailed isotopic studies by R. D. Tucker (Tucker et al., 1988; Tucker and Robinson, 1991; Robinson and Tucker, 1991). Zircon and monazite from a sillimanite-orthoclase pegmatite within the schist in question yielded nearly concordant U-Pb ages of 357 and 367 Ma, respectively, and monazite from the schist gave 367 Ma, precluding an early metamor-

phism much before 360 Ma. The quartzite at the top of the Mount Mineral Formation contains virtually concordant single detrital zircons of a range of ages including grains with ages of 458, 440, and 439 Ma, showing that the quartzite itself could be no older than Early Silurian. The same quartzite contains the extremely retrograde-zoned, Mn-rich garnet described below and has yielded

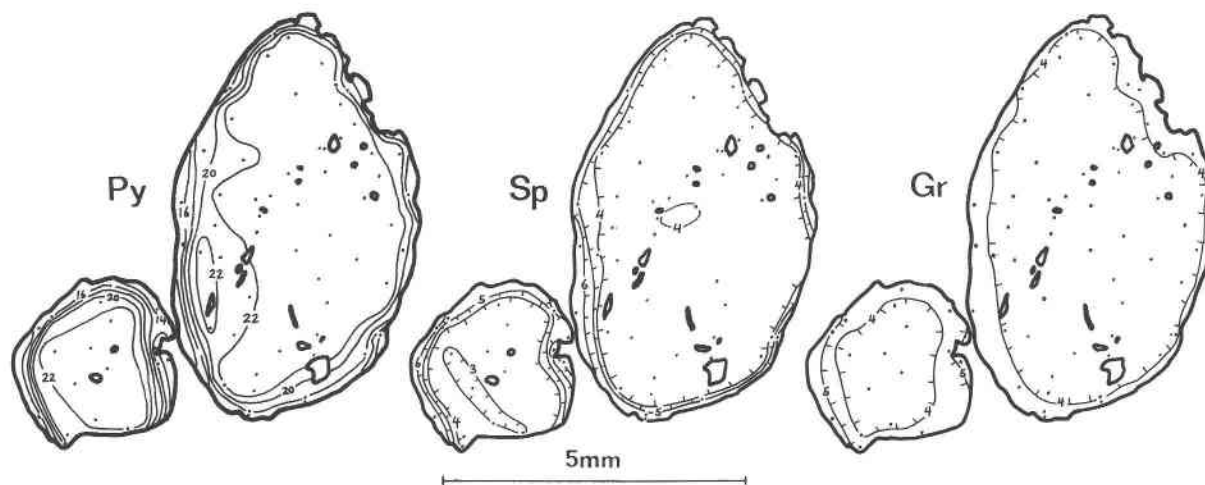


Fig. 31. Composition maps of egg-shaped porphyroclastic garnet from sample W67B of the Mount Mineral Formation (see Fig. 27). Contours based on $Mg/(Fe + Mg + Mn + Ca)$, etc., are Py 16–22% at 2% intervals, Sp 4–6% at 1% intervals, and Gr 4–5% at 1% intervals. Maps show rimward decrease in pyrope and increase in spessartine consistent with resorption of garnet by reaction with matrix through the medium of interstitial fluid. Adapted from Roll (1987).

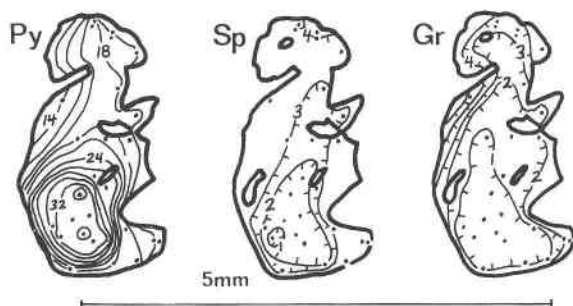


Fig. 33. Composition maps of small fragmented garnet from sample X43A from the Mount Mineral Formation on the south-west limb of the Pelham dome. Contours based on $\text{Mg}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$, etc., are Py 14–34% at 2% intervals, Sp 1–4% at 1% intervals, Gr 1–4% at 1% intervals. Although only 1×3 mm in size, it contains almost the same range of pyrope compositions as the much larger garnet in Figure 29. Probably the garnet is a sheared fragment from the interior of a once larger garnet that was only tectonically exposed to the matrix over a relatively short period near the end of the late reequilibration. Adapted from Roll (1987).

a recrystallization age of metamorphic monazite of 297 Ma, consistent with metamorphic sphene ages (Tucker and Robinson, 1990), and Rb-Sr and Nd-Sm mineral isochrons (Gromet and Robinson, 1990) nearby in the dome. Thus, it appears that the relict granulite-facies metamorphism was Acadian and the kyanite-muscovite overprint was late Paleozoic, not Acadian. The boundary between this late overprint and true Acadian isograds farther east is being actively investigated.

In sample X43 (Fig. 33) almost the same range of pyrope compositions was obtained (Fig. 30) in a grain 1×3 mm in size. In this case it seems hardly likely that such

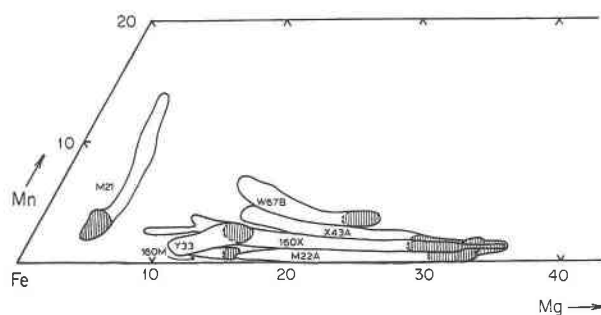


Fig. 34. Summary diagram of the zoning trajectories for garnets in seven specimens of the Mount Mineral Formation studied by Roll (1987). X43A, 160X, and W67B are from the schists described above. M22A is from a sillimanite-orthoclase pegmatite. The Fe-Mn-rich garnet is from a nearby garnet quartzite shown with contours for spessartine and grossular in Figure 35. There is little to no zoning in pyrope content, and the trajectory is consistent with a resorption reaction involving formation of Fe-rich biotite. Specimens 160M and, especially, Y33 belong to rocks that appear to have been totally reconstituted, with garnet newly grown during the late hydrous metamorphic overprint, as discussed in the text.

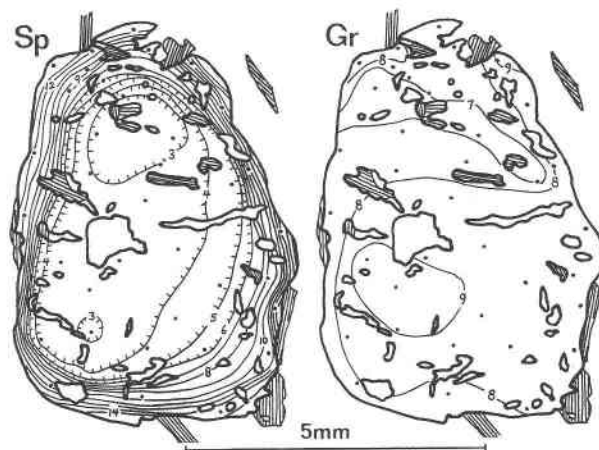


Fig. 35. Composition maps of egg-shaped porphyroclastic garnet from quartzite sample M21 near the upper contact of the Mount Mineral Formation on the east limb of the Pelham dome. Contours at intervals of 1% based on $\text{Mn}/(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})$, etc., Sp 3–14%, Gr 6–8%. Maps show rimward increase in spessartine consistent with resorption of garnet by reaction with the matrix biotite through the medium of interstitial fluid. Depression hachures are omitted from closely spaced contours near edge of garnet. Adapted from Roll (1987).

a steep gradient could have been preserved through even a single retrograde event at 580 °C, and more probably the garnet is a sheared fragment from the interior of a once larger garnet that was only tectonically exposed to the matrix over a relatively short period near the end of the late reequilibration.

Figure 34 summarizes the zoning trajectories for garnets in seven different specimens in the Mount Mineral Formation studied by Roll. Three are from the schists already described, and a fourth is similar in form but is from a sillimanite-orthoclase pegmatite. The Fe-Mn rich garnet, from the quartzite at the top of the formation, is shown in maps in Figure 35, with contours for grossular and spessartine. In this specimen there is little to no zoning in pyrope content and the trajectory is consistent with a resorption reaction involving formation of Fe-rich biotite.

In Figure 36 biotite compositions from all seven rocks are plotted in terms of Ti per 11 O atoms and the ratio $\text{Mg}/(\text{Mg} + \text{Fe})$. Ti content for a given Mg ratio has been shown to be a function of metamorphic grade in rocks saturated with ilmenite or rutile as well as aluminum silicate (Robinson et al., 1982; Schumacher et al., 1990). In these rocks variable Ti content seems to be related in part to extent of participation in retrograde resorption reactions and proximity to grains of Ti oxide minerals. The extreme example is biotite from deep within the “fjords” of the garnet 160X in Figure 29, which are colorless to pale green in thin section and contain less than 0.025 Ti per 11 O atoms. Because these fjord biotite grains were clearly produced by the reaction of a K-bearing matrix solution upon the enclosing walls of garnet, it appears

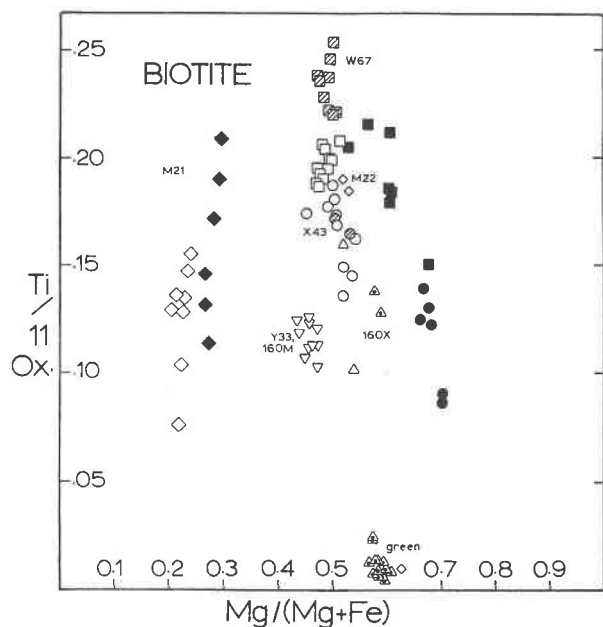


Fig. 36. Biotite compositions from all seven rocks plotted in terms of Ti per 11 O atoms and the ratio $Mg/(Mg + Fe)$. Samples shown are W67 (squares), X43 (circles), 160X (triangles), Y33 and 160M (inverted triangles), M22 (small diamonds), and M21 (large diamonds). Inclusion biotite (closed symbols) is more magnesian than corresponding rim (open symbols) or matrix (ruled symbols) biotite. Dotted triangles indicate biotite in 160X that fill cracks in garnet. From Roll (1987).

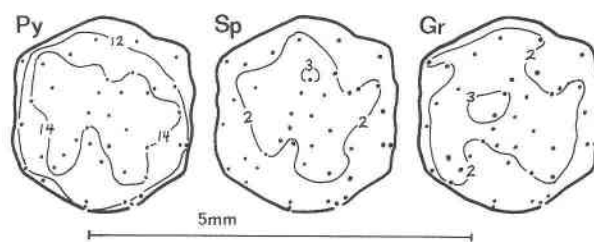


Fig. 38. Composition maps of euhedral garnet in sample Y33 from schist near the upper contact of the Mount Mineral Formation on the southwest flank of the Pelham dome. This schist is believed to have been completely reconstituted during the late metamorphic reequilibration. Maps show rimward decrease in pyrope, spessartine, and grossular content consistent with retrograde garnet growth. Adapted from Roll (1987).

that under the ambient retrograde conditions estimated at about 550 °C, the matrix solution had not dissolved or did not contain enough Ti to produce a biotite with a normal Ti content typically found in matrix locations in close physical contact with Ti-oxide grains.

In three of the samples, M21, W67, and X43, compositions were obtained both on matrix biotite and on biotite included within garnets. Inclusion biotites (Fig. 36) are consistently more magnesian than matrix biotites, in agreement with the two different processes by which

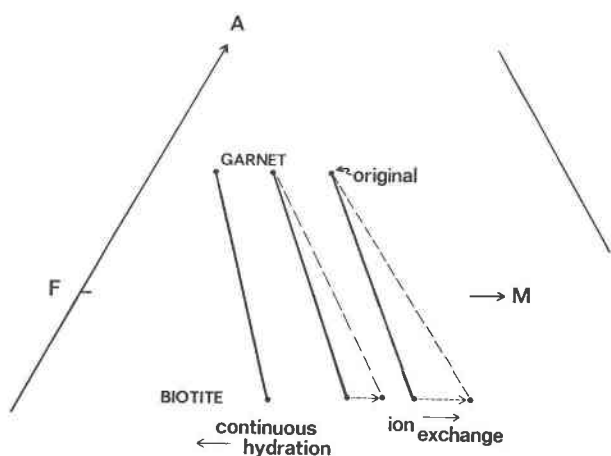


Fig. 37. Schematic AFM projection, showing the contrasting effects on biotite compositions of continuous retrograde hydration on matrix biotite and of retrograde ion exchange on inclusion biotite. As hydration proceeds, equilibrium tie lines between garnet rims and matrix biotite move to the left, whereas inclusion biotite moves to the right. The original peak biotite composition is no longer preserved but presumably had a composition intermediate between matrix biotite and biotite included in garnet cores. From a diagram prepared by M.A. Roll (personal communication, 1986).

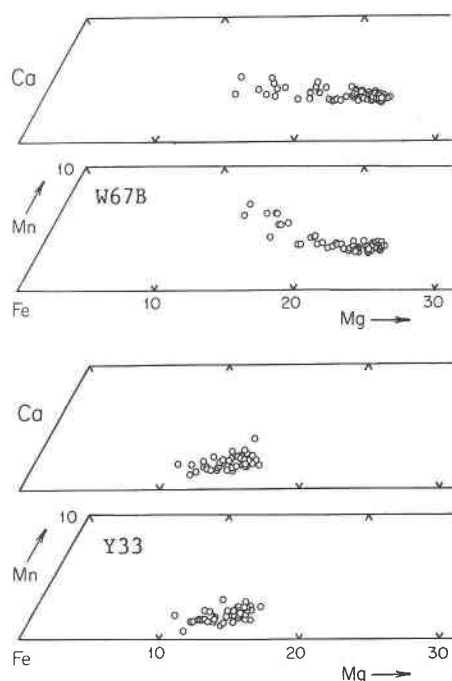


Fig. 39. Comparisons of composition trajectories of garnet in W67B and in Y33. In each example, garnet core has the most Mg-rich composition. W67B is consistent with a retrograde garnet resorption reaction, whereas Y33 is consistent with retrograde growth. Adapted from Roll (1987).

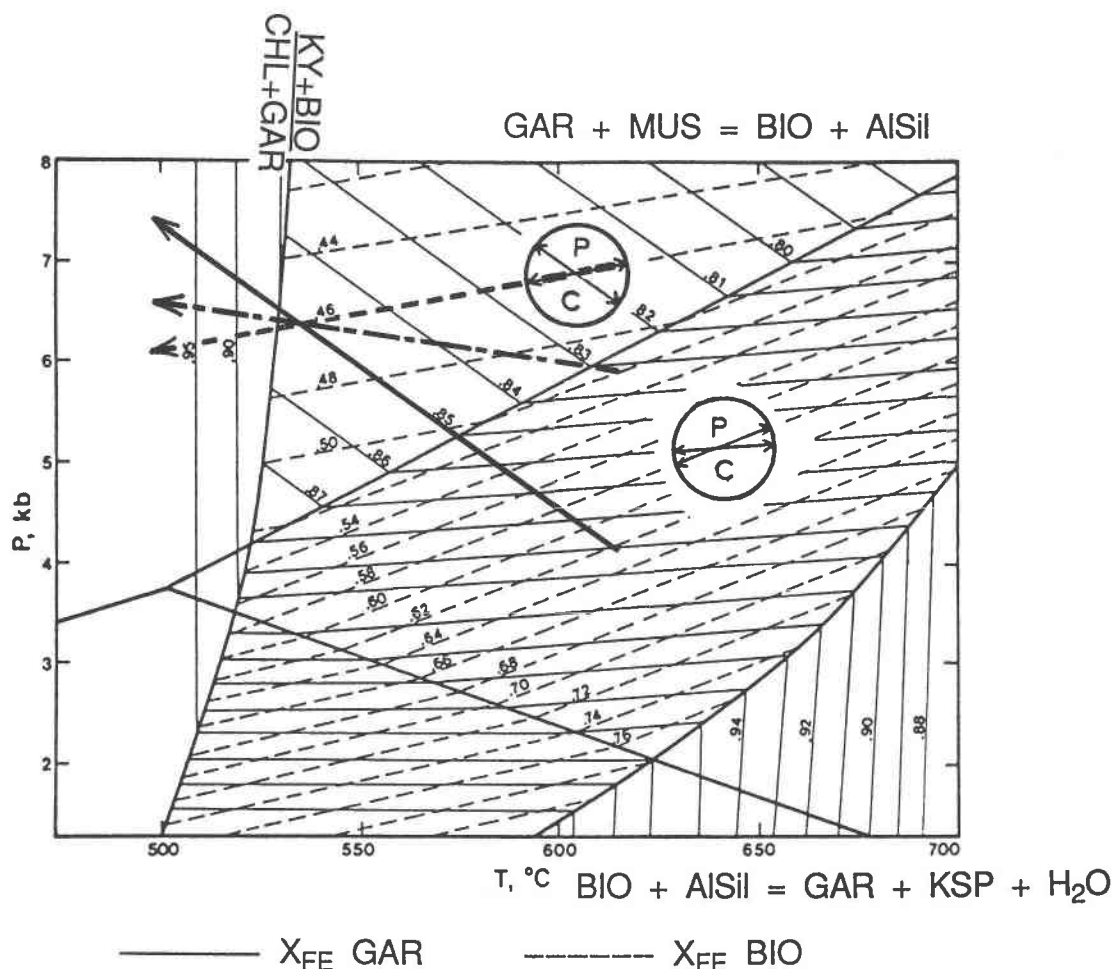


Fig. 40. P - T diagram with garnet and biotite composition isopleths for the reaction garnet + muscovite = biotite + aluminum silicate. Note that isopleths change orientation when crossing aluminum silicate field boundaries and when crossing the muscovite-out reaction in lower right. Quadrants of garnet production (P) and garnet consumption (C) are indicated for the

kyanite and sillimanite zones. Heavy lines with arrows represent three retrograde P - T trajectories illustrated in quasi-binary sections in Figure 41: one parallel to garnet isopleths (Fig. 41A), one parallel to biotite isopleths (Fig. 41C), and one intermediate (Fig. 41B). Adapted from Spear and Selverstone (1983).

their compositions were achieved. The matrix biotite formed by continuous retrograde net-transfer garnet resorption in which biotite grew in amount and became progressively more Fe enriched. The inclusion biotite did not change in amount but became Mg enriched as a result of local ion exchange with the enclosing garnet, probably producing a small Mg-impoverished compositional well in the garnet, such as that described by Spear et al. (1990).

It is apparent that if Fe-enriched matrix biotite is matched with the relict core garnet composition with which it was never in equilibrium, an artificially high-temperature estimate will be obtained (see also Fig. 24). In the case of sample 160X such a procedure gives an estimated temperature of 1000 °C. By contrast, matching of inclusion biotites with immediately surrounding garnets yields estimates of temperature of final ion exchange

equilibration of about 450 °C. A major problem is to locate and analyze, or to estimate, the composition of biotite that was originally present in the rock at the peak of granulite-facies conditions. Spear et al. (1990) managed to locate biotite inclusions within sillimanite that had apparently experienced little of the retrograde re-equilibration described here. In the Pelham dome rocks, the best estimate of original biotite composition was made by choosing one intermediate between matrix and inclusion extremes (Fig. 37) and also one that would yield an estimated temperature and pressure consistent with the presence of sillimanite in the rock. On this basis, a biotite composition chosen for sample 160X with $Mg/(Mg+Fe) = 0.66$ yields conditions of equilibration with sillimanite, garnet, and high orthoclase of 700 °C and 6.8 kbar (Roll, 1987).

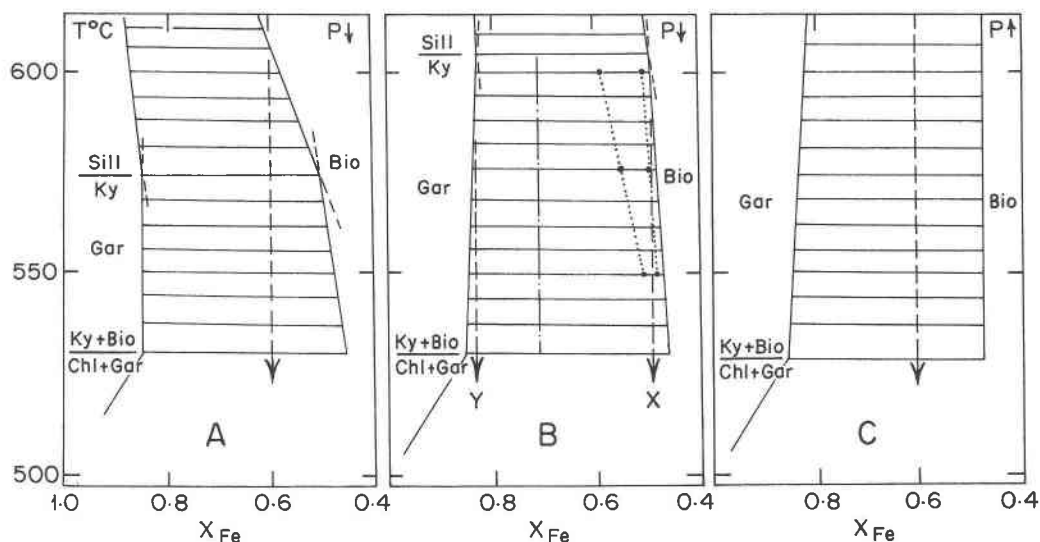


Fig. 41. Binary representations in terms of T - P and X_{Fe} of the phase relations along the three retrograde trajectories illustrated in Figure 40. For details see text. Modified and expanded from Roll (1987).

Zoning produced during retrograde garnet growth

The two garnet grains with short zoning trajectories in Figure 34 are from schists of the Mount Mineral Formation that appear to have been completely reconstituted during the second more hydrous metamorphism. These garnets are typically euhedral (Fig. 38) with Mn- and Ca-enriched cores and might be supposed to have formed by prograde growth zoning. However, unlike such garnets, they do not show a typical increase in the $Mg/(Mg + Fe)$ but rather a decrease, resembling in this way zoning profiles produced by retrograde garnet-resorption reactions. Chemical trajectories of the two types of garnet are compared in Figure 39. The garnet in sample W67B shows the typical outward increase in both Ca and Mn associated with diffusive rim resorption of the Mg component into other ferromagnesian minerals. The garnet in sample Y33, on the other hand, shows rimward decrease of Ca and Mn that can only be associated with garnet growth because there is no other mineral in the rock that could be proportionally absorbing more Mn than the garnet.

Peter J. Thompson (1985) has described interesting garnets from the Mount Monadnock area, New Hampshire, in which there is a similar strong rimward decrease in $Mg/(Mg + Fe)$ accompanied by a similar decrease in Mn but with a striking rimward increase in Ca. In this example, it appears that the pyrope content of the garnet was undergoing progressive resorption with respect to the cordierite in the rock, but at the same time the grossular content of the garnet was undergoing progressive growth with respect to the anorthite component of plagioclase that slightly overbalanced the resorption. In this example, as in all of the others discussed here, the Mn profile is the most reliable predictor of growth vs. resorption because Mn is more strongly fractionated into aluminous garnet than any other phase.

It is a common habit of metamorphic petrologists to think of garnet growth in terms of a prograde reaction and resorption as a retrograde (i.e. decreasing temperature) phenomenon. It is further habitual to visualize such reactions in terms of binary T - X loops. Study of composition isopleths for a reaction on a P - T diagram illustrates the true complexity even in an oversimplified quasi-binary situation. For this purpose, the X_{Fe} isopleths for garnet and biotite derived by Spear and Selverstone (1983) for the reaction garnet + muscovite = aluminum silicate + biotite serve as an admirable example, as shown in Figure 40. In the habitual view, one may cross the isopleths by increasing pressure and increasing temperature slightly such that garnet is produced, biotite is consumed, and progressively lower X_{Fe} isopleths for both minerals are crossed. Similarly, one may cross isopleths by decreasing pressure and temperature such that garnet is resorbed, biotite grows, and progressively higher X_{Fe} isopleths for both minerals are crossed. These two directions toward garnet production or consumption (P and C in Fig. 40) lie in only two of the possible four quadrants between the isopleths.

Relations within the remaining two quadrants are more complex. In these quadrants the directions of garnet and biotite composition change are opposite, and for each P - T trajectory, garnet production or consumption will depend critically on the garnet-biotite ratio in the rock. The relations in these quadrants are illustrated by PT - X_{Fe} sections from Figure 40 taken to follow three specific trajectories. Figure 41A shows a trajectory constrained along the Garnet $X_{Fe} = 0.85$ isopleth for the kyanite zone. This PT - X section is oriented with increasing T at the top, which is also the direction of decreasing P . Within the kyanite zone, as specified, garnet composition is constant, but along the same trajectory at lower temperature in the



Fig. 42. Pavement outcrop of typical migmatitic schist of the sillimanite-orthoclase-garnet-cordierite zone, central Massachusetts. Large garnets and leucosomes are believed to be the product of fluid-absent partial melting. See Robinson et al. (1986), stop 4.

garnet-chlorite zone and also at higher temperature in the sillimanite zone, garnet is more Fe-rich. Taking a typical bulk composition along the dashed line and following it in Figure 41A in the kyanite zone only, it can be seen that garnet is consumed with increasing temperature (i.e., the bulk composition on the garnet-biotite tie line lies progressively closer to the biotite end) and is produced as temperature decreases.

Figure 41C shows a trajectory constrained along the biotite isopleth $X_{\text{Fe}} = 0.45$ for the kyanite zone. This section is also oriented with increasing T at the top, but in this case, this is also the direction of slightly increasing P . Taking the same typical bulk composition along the

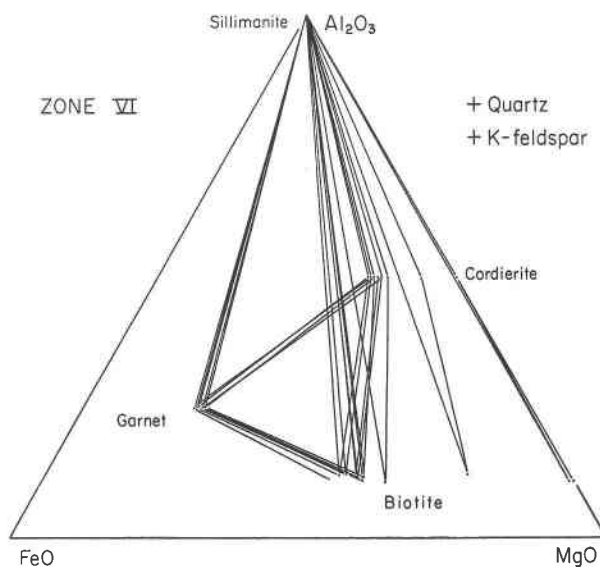


Fig. 43. AFM projection showing mineral compositions in typical assemblages of pelitic schist of Zone VI, central Massachusetts. From Robinson et al. (1982).

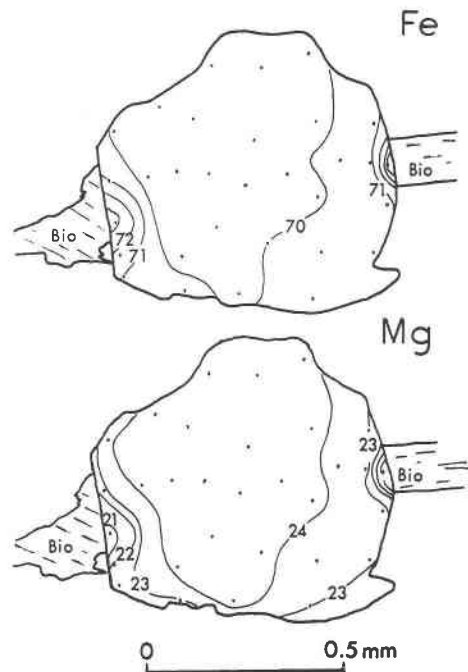


Fig. 44. Composition maps of garnet in sample FW407 of sillimanite-orthoclase-garnet-cordierite-biotite schist from Zone VI, central Massachusetts. Garnet shows significant Fe enrichment and Mg depletion only where in direct contact with adjacent biotite, permitting retrograde ion exchange. From Tracy et al. (1976). See Robinson et al. (1986), stop 6.

dashed line and following it in Figure 41C, it can be seen that garnet is produced in an up-temperature direction (i.e., the bulk composition lies progressively closer to the garnet ends of the tie lines) and is consumed in a down-temperature direction.

Figure 41B is a PT - X_{Fe} section for a trajectory between the two bounding examples in 41A and 41C. On this trajectory, with increasing temperature, garnet X_{Fe} decreases while biotite X_{Fe} increases, consistent with the decreasing distribution coefficient between these two phases. Note that for Fe-poor compositions (X), garnet is consumed with increasing temperature even though the garnet itself decreases in X_{Fe} until the bulk composition consists only of biotite. Similarly for Fe-rich compositions (Y), garnet is produced with increasing temperature. With falling temperature along the same trajectory, garnet is produced in Fe-poor compositions (X) and is consumed in Fe-rich compositions (Y).

The garnet-biotite two-phase region in a profile such as that in Figure 41B is divided into regions of production and consumption by a vertical dash-dot line, which is positioned differently for each trajectory. To the left of this boundary, as in composition Y , garnet is produced with increasing temperature and consumed with decreasing temperature. To the right of this boundary, as in composition X , garnet is consumed with increasing temperature and produced with decreasing temperature. By

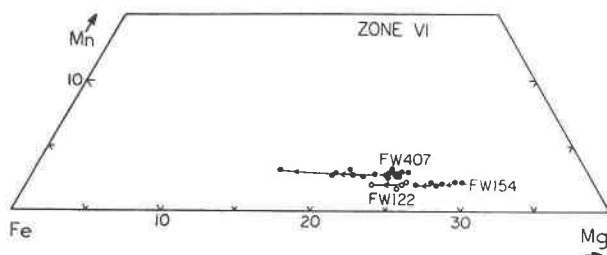


Fig. 45. Compositional trajectories of zoned garnets from Zone VI, central Massachusetts, in terms of Fe, Mn, and Mg. Typical compositions are the most Mg rich, and zoning to more Fe-rich compositions occurs only where garnet is in direct contact with biotite or cordierite. From Tracy et al. (1976).

comparing Figures 41B and 41C, it will be seen that the more nearly parallel the P - T trajectory is to the biotite isopleths, the larger the region for retrograde garnet consumption and the smaller the region for retrograde production, until, in the extreme example parallel to the biotite isopleth, there is no region for retrograde production.

In the trajectory of Figure 41B, the potential for retrograde garnet production is proportional to the distance from the production-consumption boundary. This is illustrated in two calculated examples showing the change from 600 to 550 °C, in which X_{Fe} garnet changes from 0.832 to 0.845. For a bulk composition $X_{\text{Fe}} = 0.593$, the mole percent garnet will increase from 30 to 33, whereas for a bulk composition $X_{\text{Fe}} = 0.507$, the mole percent garnet will increase from 5 to 10.5. In the case of garnet growth by fractional crystallization, the interior of the garnet will be effectively locked away from the exterior bulk assemblage so that the effective bulk composition will become progressively more biotite rich. This will drive the effective bulk composition toward lower X_{Fe} , further into the region of retrograde production of garnet with decreasing temperature. If, in the above examples of cooling from 600 to 550 °C, there is fractional crystallization during retrograde growth such that half of the garnet is taken out of reaction at 575 °C (see dotted lines in Fig. 41B), then the effective bulk composition will be even

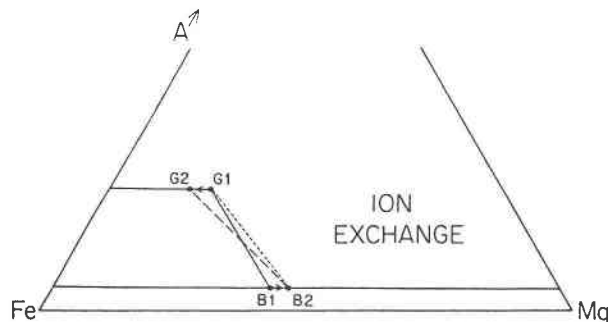


Fig. 46. AFM orthoclase projection of garnet and biotite compositions, illustrating the relatively simple interpretation required in cases of retrograde ion exchange. From Robinson et al. (1982).

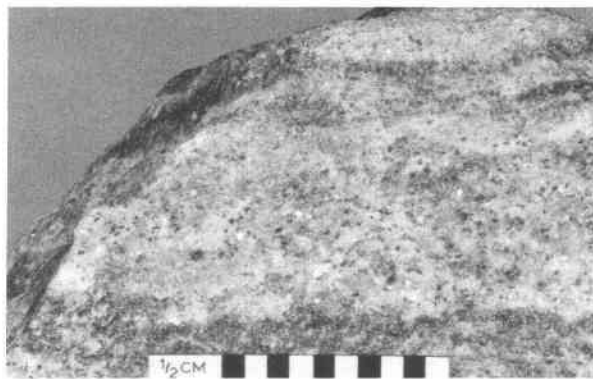


Fig. 47. Photograph of even-grained white quartzose marble with brightly reflecting elongate grains of white wollastonite. Specimen was collected from a gas pipeline trench on the east side of Rattlesnake Mountain by Berry (1989), from the heart of zone VI, central Massachusetts.

more biotite rich and still more capable of growing garnet; in the two cases, from 30 to 33.7% or from 5 to 11.7%.

In Figure 41B, it will be seen that the best opportunity for retrograde growth zoning with a substantial garnet composition change will be in a trajectory at a slight angle to biotite isopleths where a small biotite-rich composition region will be subject to garnet growth. Fractional crystallization will continue to maintain the effective bulk composition in this biotite-rich region where garnet growth can continue. Such a situation of garnet-growth zoning with falling temperature appears to be the origin of the zoned garnets in completely reconstituted schists such as sample Y33 (Fig. 39).

Zoning produced during retrograde ion exchange

The pelites of the sillimanite-orthoclase-garnet-cordierite zone (Zone VI) in central Massachusetts are typified by leucosomes of quartz, feldspar, and cordierite commonly containing garnet up to 2 to 3 cm in diameter (Fig. 42). Such textural features have been described as the product of fluid-absent melting (Tracy and Robinson, 1983). Typical assemblages and mineral compositions are shown in an orthoclase projection in Figure 43. Garnet is typically about 5–10% more magnesian than garnet in Zones IV and V, indicating considerable progress of the strong dehydration reaction biotite + sillimanite = orthoclase + garnet + H_2O . Very locally there is evidence of late retrograde hydration reactions between garnet or cordierite and orthoclase to produce sillimanite + biotite, but this is not typical. The common four-phase assemblage in projection may indicate that H_2O was behaving as an inert component in the absence of an aqueous fluid. Typically both larger and smaller garnets are homogeneous from core to rim except at points where the garnets are in direct contact with another ferromagnesian mineral, either cordierite or biotite (Fig. 44). This style of zoning was first described in this region by Hess (1969,

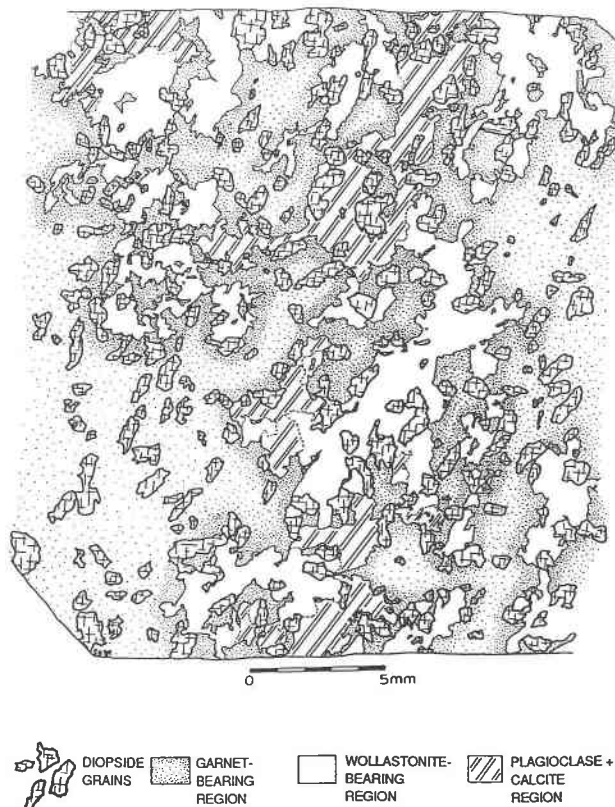


Fig. 48. Thin-section sketch of wollastonite marble, showing subdivision of the rock into regions. Wollastonite-bearing regions have a veinlike character, separated by areas with quartz-calcite. Plagioclase-calcite regions occur predominantly away from diopside, whereas close to diopside, plagioclase and calcite are separated by garnet. From Berry (1989).

1971). In these locations, the Fe-enriched garnet rim is accompanied by zoned biotite or cordierite that is enriched in Mg as compared to biotite or cordierite that is in the matrix away from the garnets. The local chemical zoning trajectories (Fig. 45) are virtually flat with respect to Mn, strongly indicating that there has been neither garnet resorption nor garnet growth. This type of zoning is most easily explained as the product of local retrograde ion exchange between garnet and surrounding ferromagnesian minerals where they are in direct contact and without participation of an interstitial fluid. It is a strong argument that at peak conditions pervasive fluids were not present in most zone VI rocks. If no aqueous fluid was present, then we must appeal to straight diffusion through the matrix or localized ion exchange at grain-to-grain contacts. The latter was first observed and studied in central Massachusetts by P. C. Hess (1969, 1971), and his observations have been repeated countless times.

The interpretation of these kinds of garnets in terms of geothermometry is apparently much simpler. The homogeneous garnet composition can be compared with the matrix biotite composition away from the garnet to estimate peak conditions (Fig. 46). The garnet rim com-

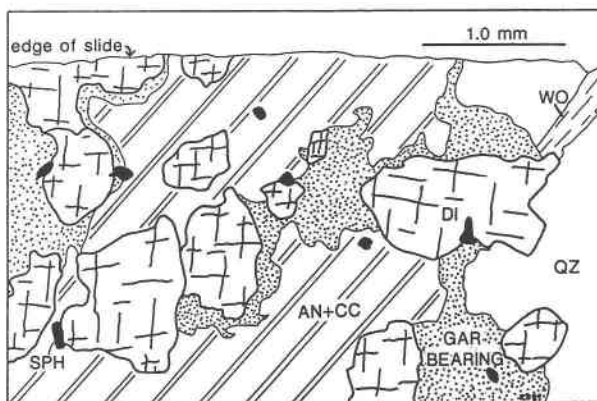


Fig. 49. Thin-section sketch showing garnet-bearing regions close to diopside and preservation of anorthite + calcite away from diopside. From Berry (1989).

position and the adjacent biotite or cordierite composition can be used to estimate the temperature of last retrograde ion exchange. The only question might be whether some other process intervened between the peak metamorphism and the retrograde ion exchange that could have effectively changed the typical matrix biotite composition without also producing concentric zoning in the garnet. In my opinion this is unlikely.

Conclusion concerning garnet zoning

It seems fairly clear that at temperatures below about 500 °C, most garnet zoning can be interpreted in terms of simple growth zoning. At higher temperatures, however, all sorts of contradictions are produced by diffusion, which commonly makes it very difficult to select the correct mineral compositions for quantitative geothermometry and geobarometry. If all of the conditions and considerations outlined here are correctly accounted for, and one happened to be confronted with appropriate rocks, then it may be possible to construct detailed *P-T* paths such as those of Spear et al. (1990).

PHASE RELATIONS IN A WOLLASTONITE-BEARING MARBLE

In the next section of this paper, I want to show how detailed petrographic observations in a single thin section led to a new conception for a petrogenetic grid. This conception comes from the Ph.D. thesis of Henry Berry (1989, 1991) and concerns one aspect of his interpretation of the first documented occurrence of regional metamorphic wollastonite from the granulite-facies in New England, near Sturbridge, in the center of Zone VI of Figure 2. This section is based on a small highlight of a more comprehensive study that Berry is preparing for separate publication.

Petrography

The rock is an even-grained, foliated, white quartzose marble with brightly reflecting elongate grains of white

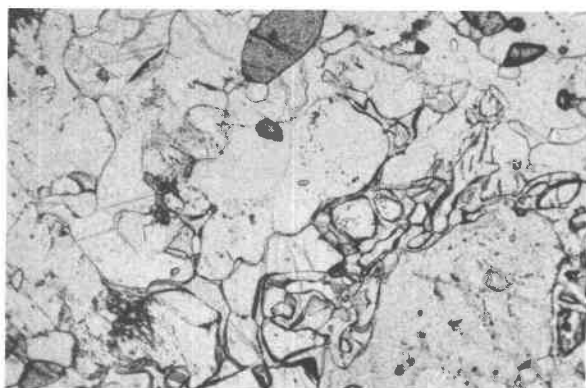


Fig. 50. Photomicrograph showing abundant garnet coronas between anorthite and calcite close to large diopside grain in lower right and absence of such coronas between anorthite and calcite in upper left. Dark grains are sphene. Width of view, 3 mm approximately. Photograph provided by H.N. Berry.

wollastonite (Fig. 47). The rock consists mainly of quartz, calcite, anorthite (An₉₈), wollastonite, calcic scapolite (~Me₈₆), ferroan diopside [$\text{Fe}/(\text{Fe} + \text{Mg}) = 0.40\text{--}0.45$], grossular-rich garnet (Gross₈₈–₉₀), and sphene. Locally wollastonite appears to be concentrated along veinlike surfaces that may have allowed concentration of early metamorphic fluids.

An unusually detailed petrographic study of a single thin section by Berry indicated different local chemical environments that require treatment in different chemical systems (Fig. 48). In the first place, the rock can be divided into regions containing quartz + calcite (mainly

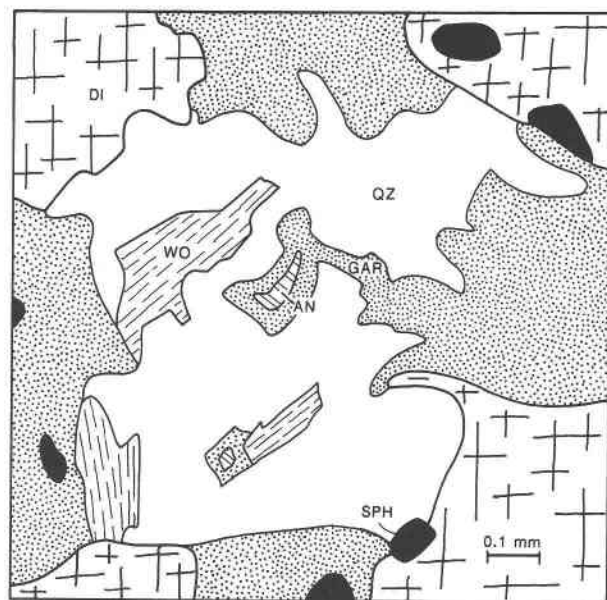


Fig. 51. Thin-section sketch in wollastonite-bearing region with abundant diopside, showing garnet rim between anorthite and wollastonite. Sketch provided by H.N. Berry.

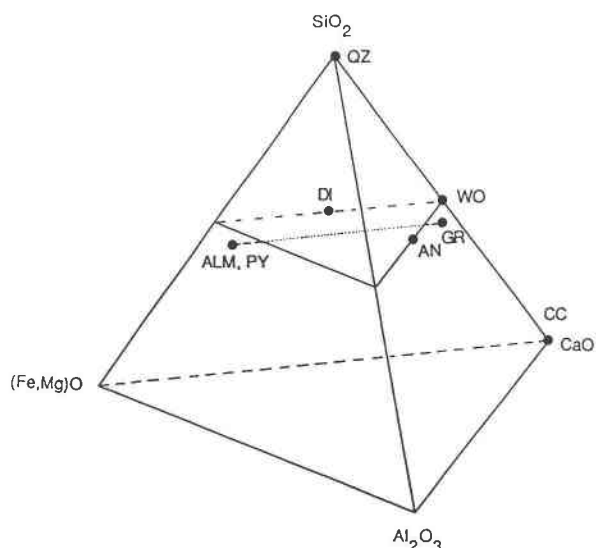


Fig. 52. The chemical tetrahedron $\text{SiO}_2\text{--CaO--Al}_2\text{O}_3\text{--(Fe, Mg)O}$, showing the compositions of quartz (QZ), anorthite (AN), wollastonite (WO), diopside (DI), calcite (CC), and the garnet solid solution grossular (GR) to almandine-pyrope (ALM, PY). From Berry (1989).

plagioclase + calcite regions of Fig. 48) and those containing quartz + wollastonite (wollastonite regions of Fig. 48). Clearly conditions of the calcite + quartz = wollastonite reaction had been reached in some parts of the rock but not in others, suggesting a very locally variable X_{CO_2} of the fluid may have obtained at the peak of metamorphism.

Secondly, the rock can be divided into regions where plagioclase and calcite are in equilibrium and there is no garnet, and other regions where plagioclase and calcite are separated by abundant garnet. The regions of abundant garnet are invariably close to the widely scattered grains of ferroan diopside. This is illustrated in Figure 49 and more dramatically in Figure 50 where garnet has grown at the expense of anorthite + calcite only in the vicinity of diopside. The necessary participation of ferroan diopside in garnet-forming reactions is shown by the analysed garnet composition: grossular 88–90, almandine 5, pyrope 0.5–0.7, spessartine 1.3–1.8, andradite 2.7–3.4. An analogous reaction has taken place in wollastonite-bearing regions near diopside, where wollastonite and anorthite are not in contact and anorthite is rimmed by garnet (Fig. 51). These critical textural observations emphasized the role of the diopside component in all the garnet-producing reactions in this rock and led to consideration of equilibria in a more complex system and the derivation of a specific new petrogenetic grid.

Chemography

The consideration of calcite-quartz-wollastonite equilibria requires $\text{SiO}_2\text{--CaO--H}_2\text{O--CO}_2$, the addition of anorthite and grossular reactions requires the addition of

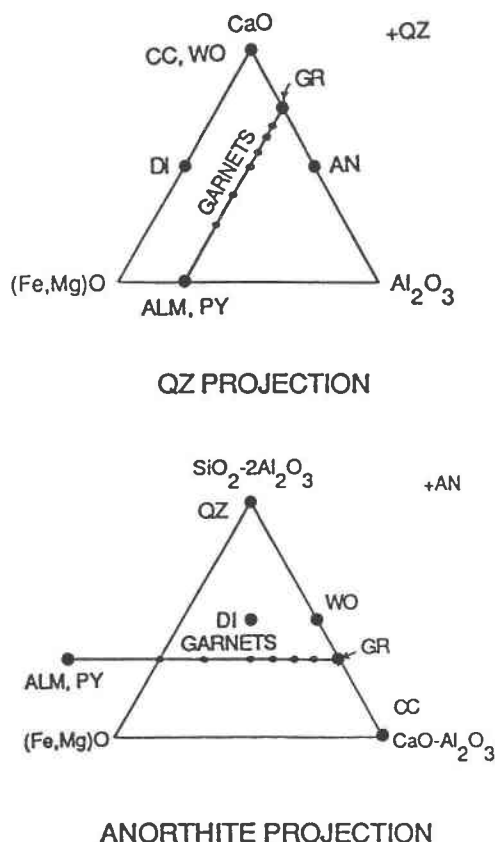


Fig. 53. Quartz projection of phase compositions onto the base of the tetrahedron of Figure 52 and anorthite projection of the same compositions onto the plane $\text{SiO}_2\text{-CaO-(Fe, Mg)O}$. In the anorthite projection ALM, PY garnets plot at negative values of CaO. From Berry (1989).

Al_2O_3 , and the addition of ferroan diopside and the real garnet compositions requires the addition of $(\text{Fe,Mg})\text{O}$, here lumped for simplicity as one component. The resulting chemical tetrahedron is shown in Figure 52. All the phases plot at point compositions except garnet, which has a theoretically complete range from pure grossular to pure almandine-pyrope. When projected from the quartz apex onto the base of the tetrahedron, the phase compositions appear as shown in Figure 53 (top). A more helpful projection is from anorthite onto the plane $\text{SiO}_2\text{-CaO-(Fe,Mg)O}$ (Fig. 53 bottom), which already contains the phases quartz, wollastonite, calcite, and diopside. Garnet projects as a line below the diopside-wollastonite join. The grossular end of the series projects positively between calcite and wollastonite, but the almandine-pyrope end projects to negative values of CaO well outside the triangle of projection. The big advantage of this projection is that it allows direct observation of the special garnet compositions that are coplanar with other phases in the tetrahedron, and it is used in all discussion that follows.

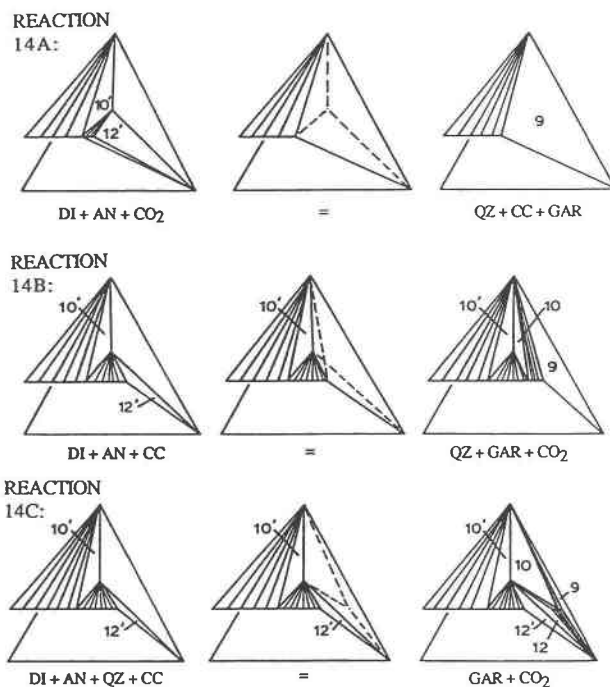


Fig. 54. Anorthite projections showing three chemographies of the equilibrium $\text{DI} + \text{AN} + \text{QZ} + \text{CC} + \text{GAR} + \text{CO}_2$ (reaction 14 of Berry, 1989) depending on the equilibrium composition of garnet. Other numbers indicate continuous reactions described in detail by H.N. Berry. From Berry (1989).

A new petrogenetic grid

In Figure 54 are illustrated the chemographic relations of the equilibrium quartz + calcite + anorthite + diopside + garnet (reaction 14), all obviously within the stability of calcite + quartz. If garnet is less grossular than the quartz-diopside-anorthite plane, as in reaction 14A, then the join diopside-anorthite gives way to the assemblage quartz + calcite + garnet, consuming CO_2 , and is thus a garnet-forming reaction that would take place with decreasing T . If the garnet composition is slightly more grossular than the quartz-diopside-anorthite plane, as in reaction 14B, then the plane diopside-anorthite-calcite gives way to quartz + garnet + minor CO_2 and is a garnet-forming reaction that would take place with increasing T . If the garnet composition is more grossular than the diopside-calcite-anorthite plane, as in reaction 14C, then all the phases quartz-calcite-diopside-anorthite give way to garnet + a large amount of CO_2 . The limiting situation would be the reaction where quartz-calcite-anorthite gives way to grossular + CO_2 without participation of diopside.

Of course all of these segments are part of the same continuous reaction, which must change its P - T slope drastically in the presence of CO_2 -bearing fluid, and each segment is separated from the next by an invariant point characterized by a particular coplanarity of phases. Between reaction 14A and reaction 14B, invariant point

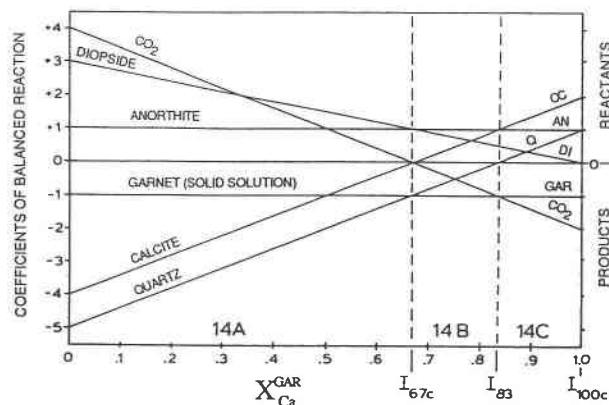


Fig. 55. Plot of reaction coefficients versus X_{Ca} of garnet for reaction 14 of H.N. Berry. Points I67c, I83, and I100c represent garnet compositions at invariant points produced by collinearities of garnet composition with other phases. From Berry (1989).

I67c represents the coplanarity of quartz-diopside-anorthite-grossular 67. Between reaction 14B and reaction 14C, invariant point I83 represents the coplanarity of diopside-calcite-anorthite. Invariant point I100 represents the limiting coplanarity of quartz-calcite-anorthite with grossular. The varied stoichiometry of reaction 14 with variable grossular content of garnet is illustrated in Figure 55, showing how CO_2 changes dramatically from a reactant

for garnet compositions below grossular 67 to a product that becomes increasingly voluminous to a maximum for pure grossular. In the vicinity of I67, the CO_2 coefficient is vanishing and there is no ΔS of devolatilization, and hence the reaction is primarily pressure dependent.

From the above considerations, Berry constructed the P - T loop shown in Figure 56 (left). At each of the three invariant points the curve of reaction 14 is tangent to a second univariant curve. At I67 this is the fluid-independent reaction grossular 67 + quartz = diopside + anorthite, which has a metastable extension to lower temperature. This reaction is critical in subsequent discussions. At I83 the second univariant curve is the decarbonation reaction diopside + calcite + anorthite = grossular 83 + quartz + CO_2 , which has a metastable extension to higher pressure. At I100 the second reaction is the decarbonation reaction calcite + anorthite + quartz = grossular 100 + CO_2 . In this case the reaction 14 curve comes in tangent to the other reaction at an end point, and there are no metastable extensions. To me, the resulting diagram far exceeds the worst nightmares encountered in J. B. Thompson's phase equilibrium course thirty years ago. My student Berry had only to experience this by proxy. Figure 56 (right) shows the same equilibria but with the addition of garnet composition isopleths for various continuous reactions within the region of stability of garnet + quartz-calcite-anorthite.

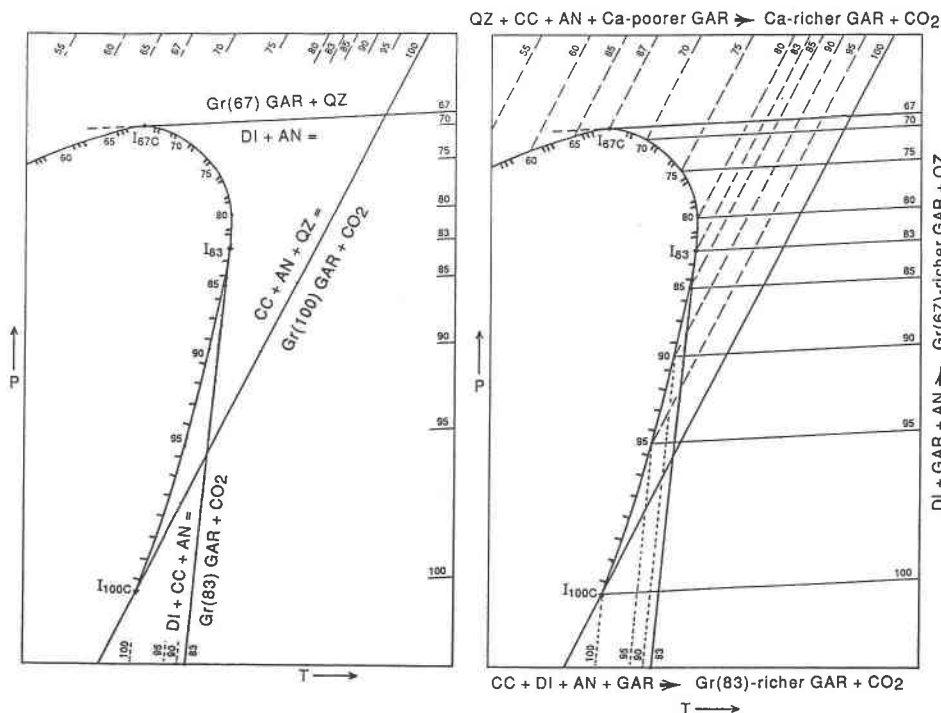
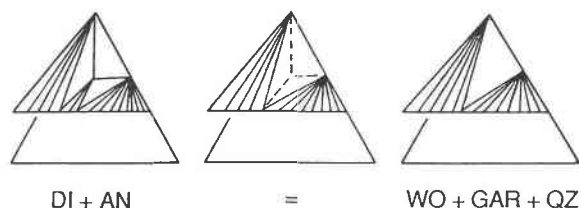


Fig. 56. P - T diagrams in presence of CO_2 -bearing fluid, showing relations of reaction 14 of H.N. Berry to other reactions involving garnet in the chemical system of Figure 54. Diagram to left shows intersections of reaction 14 with other reactions involving collinearities of garnet compositions with other phases. Hachures on reaction 14 indicate the three segments: 14A (triple hachures), 14B (double hachures), and 14C (single hachures) demarcated by invariant points I67c and I83. Diagram to right shows isopleths of garnet composition for various reactions within the stability field of garnet. Adapted from Berry (1989).

REACTION 16A:



REACTION 16B:

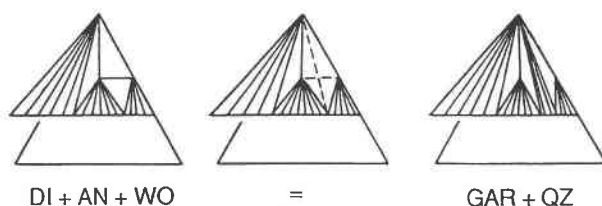


Fig. 57. Anorthite projections showing two chemographies of the equilibrium $DI + AN + QZ + WO + GAR$ (reaction 16 of Berry, 1989) depending on equilibrium composition of garnet. From Berry (1989).

An analogous but simpler reaction sequence in the field of stability of quartz-wollastonite instead of quartz-calcite is illustrated in Figure 57. If garnet has less grossular than the quartz-diopside-anorthite plane as in reaction 16A, then the join diopside-anorthite gives way to the assemblage quartz + wollastonite + garnet, a fluid-independent reaction that would take place with increasing T . If the garnet composition is more grossular than the quartz-diopside-anorthite plane, as in reaction 16B, then the plane diopside-wollastonite-anorthite gives way to quartz-garnet, a fluid-independent reaction that would take place with decreasing T . The limiting situation would be the reaction where wollastonite-anorthite gives way to grossular-quartz with decreasing T without participation of diopside.

Reactions 16A and 16B are parts of a single continuous reaction marked by two invariant points, as shown in Figure 58 (top). At each of the invariant points the curve of reaction 16 is tangent to a second univariant curve. At I67w this is the fluid-independent reaction grossular 67 + quartz = diopside + anorthite, which has a metastable extension to higher temperature. This reaction is the same one that connects at lower temperature to I67c in Figure 56. At I100 the second reaction is the fluid-independent reaction wollastonite + anorthite = grossular + quartz. In this case the reaction 16 curve comes in tangent to the other reaction at an end point, and there are no metastable extensions. Figure 58 (bottom) shows the same equilibria but with the addition of garnet composition isopleths for two continuous equilibria within the region of stability of garnet + quartz-wollastonite-anorthite. All of the equilibria in Figure 58 are fluid independent.

The equilibria in Figure 56, with quartz-calcite stable, and those in Figure 58, with wollastonite stable, are crude mirror images of each other, reflected across the fluid-

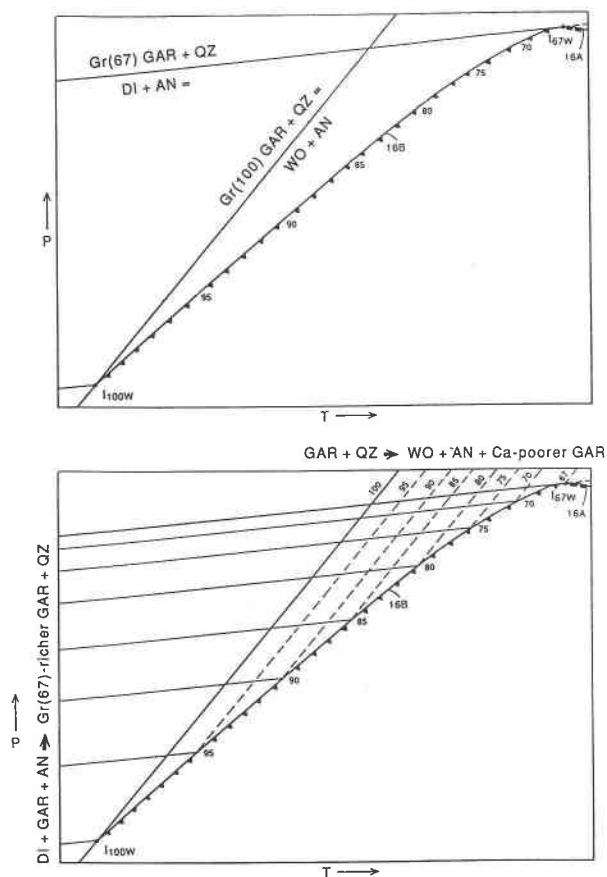


Fig. 58. P - T diagrams showing relations of reaction 16 of H.N. Berry to other reactions involving garnet in the chemical system of Figure 57. All equilibria are fluid independent. Diagram at top shows intersections of reaction 16 with other reactions involving collinearities of garnet compositions with other phases. Hachures on reaction 16 indicate the two segments: 16A (square hachures) and 16B (triangular hachures) demarcated by invariant points I67w. Diagram at bottom shows isopleths of garnet composition for reactions within the stability field of garnet. Adapted from Berry (1989).

dependent reaction quartz + calcite = wollastonite + CO_2 . These are shown in Figure 59, in which fluid-dependent and fluid-independent equilibria are keyed differently. Most of the equilibria in Figure 56 are fluid dependent; those in Figure 58 are all fluid independent, but they share the two fluid-independent equilibria connecting I67c to I67w and I100c to I100w. At low X_{CO_2} (Fig. 59A), the fluid-dependent equilibria are at low T . As X_{CO_2} increases, they and the related invariant points slide to higher temperature along the fluid-independent equilibria (Fig. 59B, C, and D), progressively shrinking the stability field of pure grossular in the center and the field of grossular-rich garnet on either side. Eventually the two invariant points I100c and I100w merge along the calcite + quartz = wollastonite + CO_2 reaction (Fig. 59C), and at higher values of X_{CO_2} (Fig. 59D), progressively less and less grossular garnet can be in equilibrium with quartz-

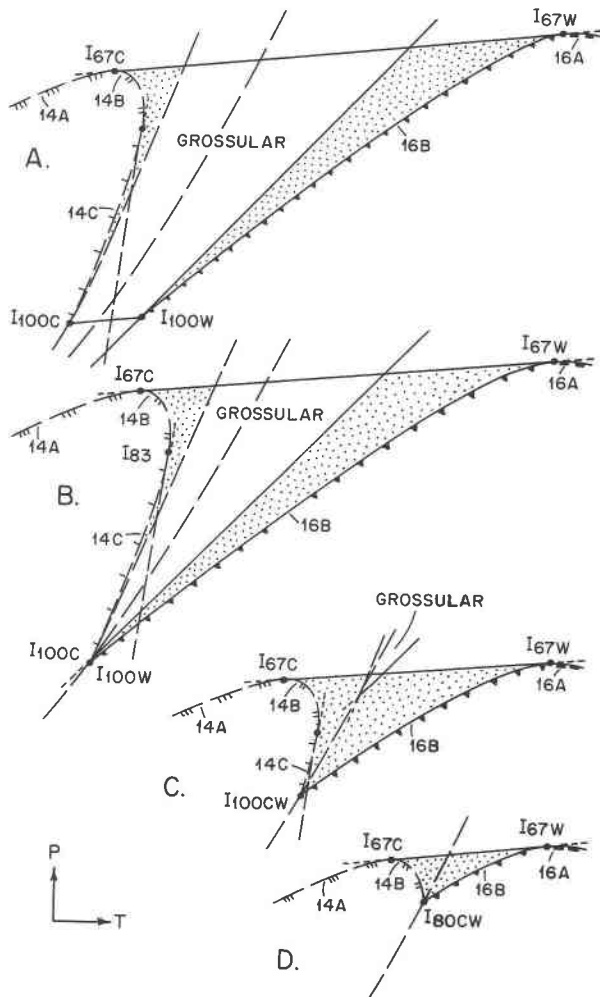


Fig. 59. Composite P - T diagrams of phase relations of DI-AN-QZ-CC-WO-GAR- CO_2 in Figures 56 and 58 separated by the equilibrium $\text{CC} + \text{QZ} = \text{WO} + \text{CO}_2$. Fluid-dependent reactions (dashed) slide toward higher temperature against fluid independent reactions (solid) as X_{CO_2} increases from A to D. Region of end-member grossular stability is marked in the center, and areas of stability of less grossular garnet are indicated by stipple. Adapted from Berry (1989).

calcite-wollastonite-anorthite-diopside. Further, as X_{CO_2} increases, the total length of the combined reactions 14 and 16 that govern the stability of garnet with anorthite-diopside and either quartz-calcite or wollastonite decreases.

Petrogenetic interpretation

Berry's petrogenetic interpretation of these rocks, based on the petrographic features and the petrogenetic grid developed above, is illustrated in Figure 60. The rocks attained the quartz + calcite = wollastonite reaction at relatively low X_{CO_2} , and once encountered, the X_{CO_2} of the fluid increased. Slightly later in this history, a fluid-independent reaction was encountered that produced

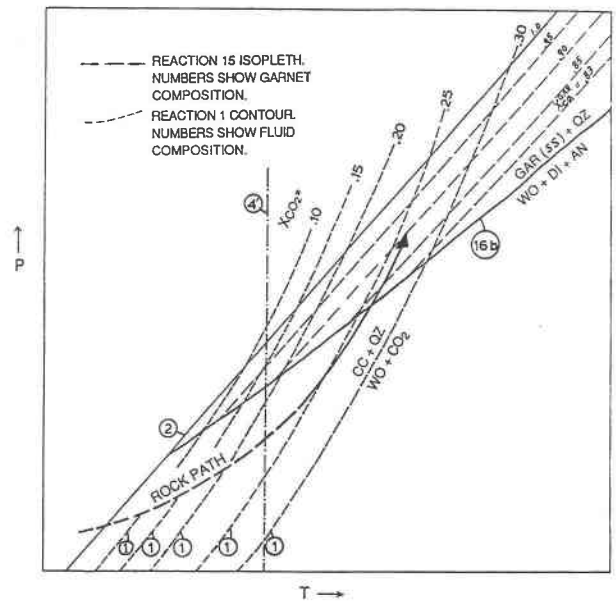


Fig. 60. P - T diagram to illustrate P - T -fluid evolution of wollastonite marble. Isopleths 1 illustrate $\text{CC} + \text{QZ} = \text{WO} + \text{CO}_2$ for various X_{CO_2} . Reaction 2 is $\text{AN} + \text{WO} = \text{GR}$. Reaction 4' is the reaction $\text{CC} + \text{PLAG} = \text{SCAP}$. Reaction 15 is the continuous fluid-independent reaction $\text{WO} + \text{AN} + \text{Ca-poorer GAR} = \text{Ca-richer GAR} + \text{QZ}$. Reaction 16b defines the limit of GAR stability with $\text{DI} + \text{AN} + \text{QZ} + \text{WO}$ as in Figures 58 and 59. From Berry (1989).

scapolite from plagioclase + calcite, which removed practically all of the albite component from plagioclase. Following a curve of increasing T and P while maintaining an increasing X_{CO_2} consistent with the quartz + calcite = wollastonite equilibrium, the rock eventually reached a pressure where calcite + wollastonite + anorthite + diopside reacted to form garnet + quartz. This would have been for a composition of grossular 83 along the fluid independent reaction 16B, which would in effect also be at invariant point I83cw in the terminology of Figures 59C and 59D. Further reaction allowed the garnet to become slightly more grossular-rich, to about grossular 90, but beyond this it appears that either compression and heating ceased or reaction rates due to fluid loss or slow diffusion did not permit garnet of higher grossular content to form. The necessary counterclockwise P - T path for this sequence of events independently supports the early part of the P - T path claimed for this region on the basis of assemblages in aluminous rocks (Schumacher et al., 1989; Robinson et al., 1989).

Summary

In the example given above, the role of the diopside component in all garnet-producing reactions is emphasized, and this leads to consideration of equilibria in a more complex system and the derivation of a specific new petrogenetic grid. In the interpretation of this rock, one can see that it is not just the phases present in the rock

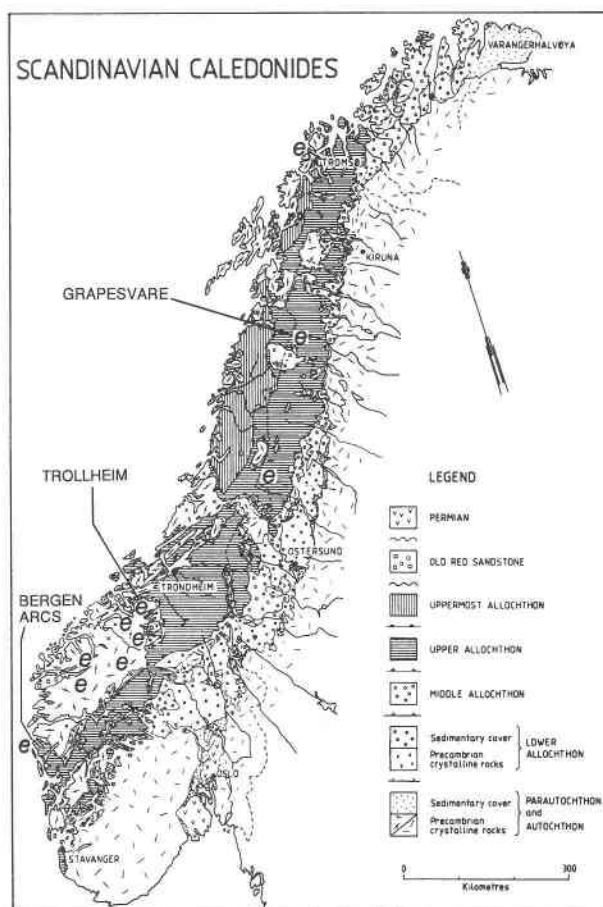


Fig. 61. Geologic index map of the Scandinavian Caledonides showing eclogite localities discussed in this paper. Adapted from a figure provided by P.-G. Andréasson.

and their compositions that are important but also the spatial and textural relations of the minerals that explain the reactions involved and their probable sequence.

METAMORPHIC HISTORY UNDERSTOOD ONLY IN TERMS OF FEATURES OF LARGER ROCK BODIES

In the last segment of this paper, I consider petrology on a larger scale. In many places, rock history can only be charted through study of the complete geologic setting of a large outcrop or even an entire mountain. Particularly spectacular are well-preserved pyroxene granulites, gabbros, and diabbases in the Scandinavian Caledonides (Fig. 61) that are locally transformed along shear zones into eclogite and amphibolite. Here a host of rock compositions all appear to have been subjected to eclogite-facies conditions, but the exact record they give of this was apparently severely dependent on very local stress and fluid availability. In some locations, the original protoliths, such as coarse-grained Proterozoic gabbros, pyroxene granulites, and diabbases, appear to have been totally protected or only slightly affected by Caledonian eclogite-facies overprinting. Elsewhere, eclogite is beau-



Fig. 62. Eclogite boudins (dark smudges) in metamorphosed sandstones of the Seve nappe, south slope of Grapesvare, Swedish Caledonides. Photograph by the author.

tifully developed with the growth of coarse metamorphic textures. Nearby, however, neither the eclogite nor its country rock may show any remaining vestiges of its eclogite-facies history, being entirely overprinted by amphibolite-facies assemblages.

Eclogitized diabase dikes in the Grapesvare area, Swedish Caledonides

My first example is from rocks of the Seve nappe in the Grapesvare area of the Swedish Caledonides (Fig. 61) described by Andréasson et al. (1985), Santallier (1988), and Mørk et al. (1988) and visited personally by me in August 1990. The rocks in question are metamorphosed Proterozoic sandstones and subordinate shales injected by diabase dikes, believed to have formed in an exten-

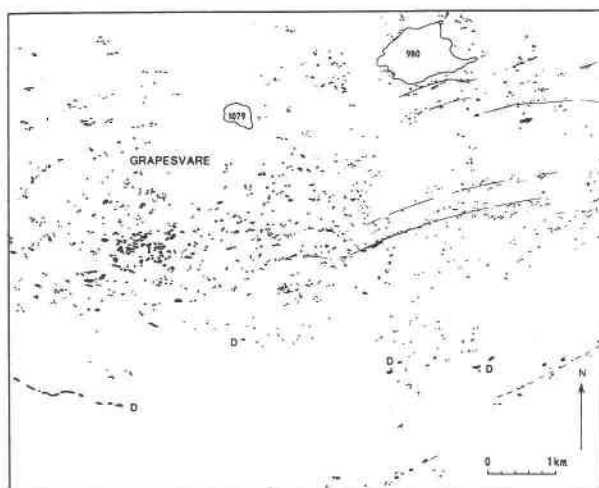


Fig. 63. Photogeologic map of distribution of metamorphosed diabase layers and boudins at Grapesvare. Boudins in northern two-thirds of map area are eclogite; those to south (indicated D) are diabase with garnet coronas and omphacitic pyroxene rims. Modified from a figure provided by P.-G. Andréasson.

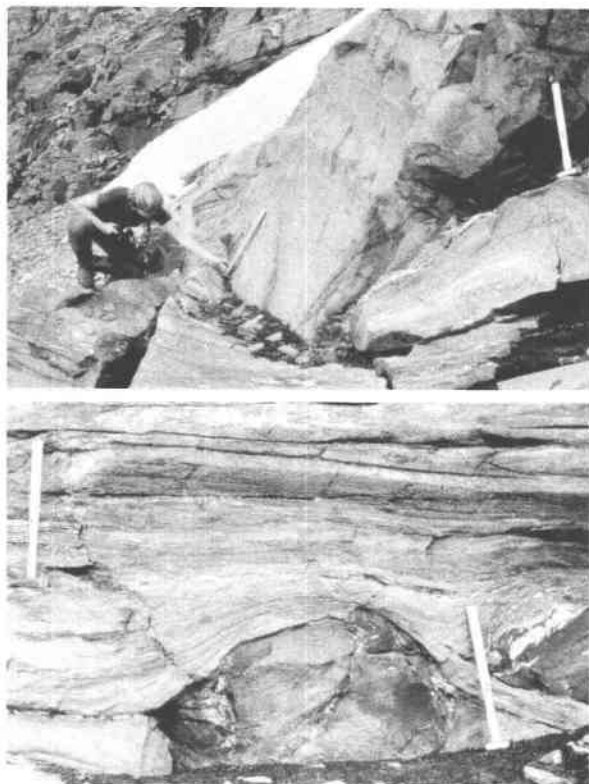


Fig. 64. Outcrop views of eclogite boudins at Grapesvare. **Top:** Lower contact of large boudin with enveloping country rock below. This boudin could not be seen on August 7, 1990, because of snow cover from the previous winter. **Bottom:** Partial views of two small boudins. Light colored area in center of boudin is garnet-rich eclogite; darker ends are amphibolitized. Photographs provided by P.-G. Andréasson.

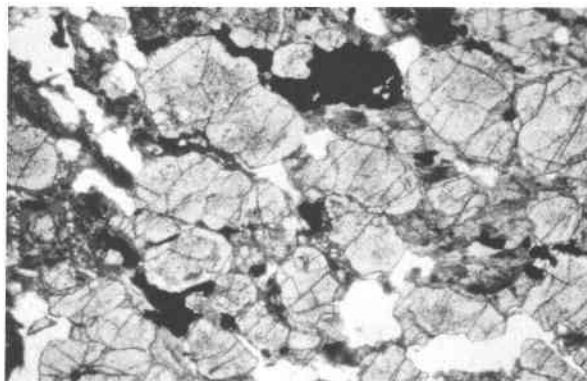


Fig. 65. Photomicrograph of eclogite from Grapesvare, showing zoned garnets about 1 mm in diameter in a matrix of quartz, omphacite, hornblende, and dark fine-grained symplectite of pyroxene and plagioclase after omphacite. Provided by P.-G. Andréasson.

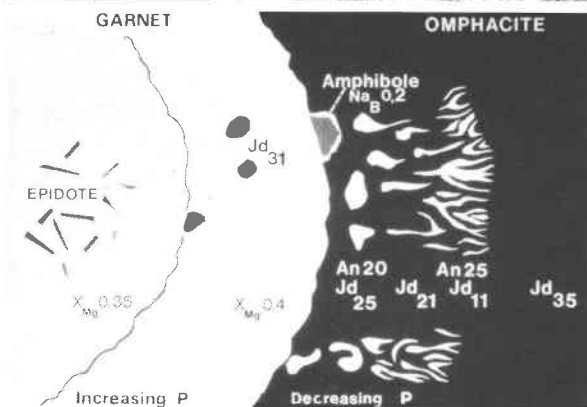
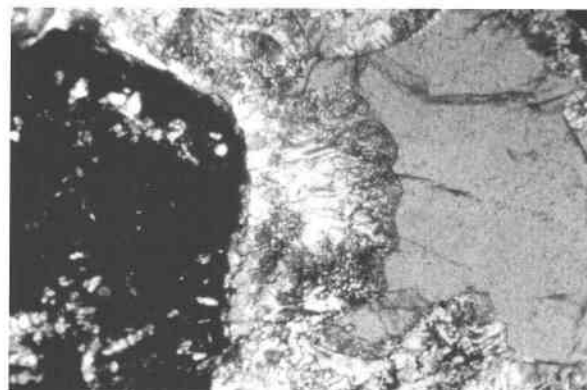


Fig. 66. Petrographic detail of Grapesvare eclogite. **Top:** Detailed photomicrograph of garnet at extinction under crossed nicols separated from fresh omphacite by a zone of plagioclase-pyroxene symplectite. **Bottom:** Sketch of the same contact region. Epidote from an original low-grade assemblage was overgrown by garnet, with progressively increasing X_{Mg} and with omphacite inclusions Jd31 that were in contact with matrix omphacite Jd35. During retrogression some omphacite broke down to pyroxene-plagioclase symplectite and was replaced locally or elsewhere completely by amphibole. Provided by P.-G. Andréasson.

sional environment along the margin of Baltica. In this they closely resemble the typical rocks of the structurally lower Särvi nappe and differ from the latter mainly in being involved in more intense metamorphism. A typical view of the field area is shown in Figure 62. The exposure is about 80%, interrupted only by a few areas of moss, small bogs, and lakes. The lighter-colored areas are highly folded and lineated metamorphosed feldspathic sandstone and minor shale with rare calc-silicate. The darker oval areas are large boudins of metamorphosed diabase up to about 15 m thick. The detailed distribution of mafic boudins over an area of about 8×10 km is shown in Figure 63. Some of the boudins occur in rows and are interpreted to be dismembered portions of individual dikes. Within the map area, there are no obvious stratigraphic or tectonic contacts, though, of course, there is a high degree of overall strain.

In the northern two-thirds of the map area, all of the mafic boudins are eclogite, usually with a sheath of retrograde amphibolite of variable thickness (Fig. 64). Eclogite is well preserved in some very small boudins (Fig. 64, bottom), whereas other large boudins are extensively



Fig. 67. Large boudin of metamorphosed diabase from the southern part of Figure 63. Note two geologists for scale to left. Photograph by the author.

amphibolitized. The typical eclogite consists of small euhedral garnets in a matrix of well-preserved omphacite, of retrograde pyroxene-plagioclase symplectite, and of secondary amphibole (Fig. 65). There is no preservation of original igneous texture except for rare light-colored patches that may be relict plagioclase phenocrysts.

A detailed study area along a garnet-omphacite contact is shown in Figure 66. The core of the garnet contains

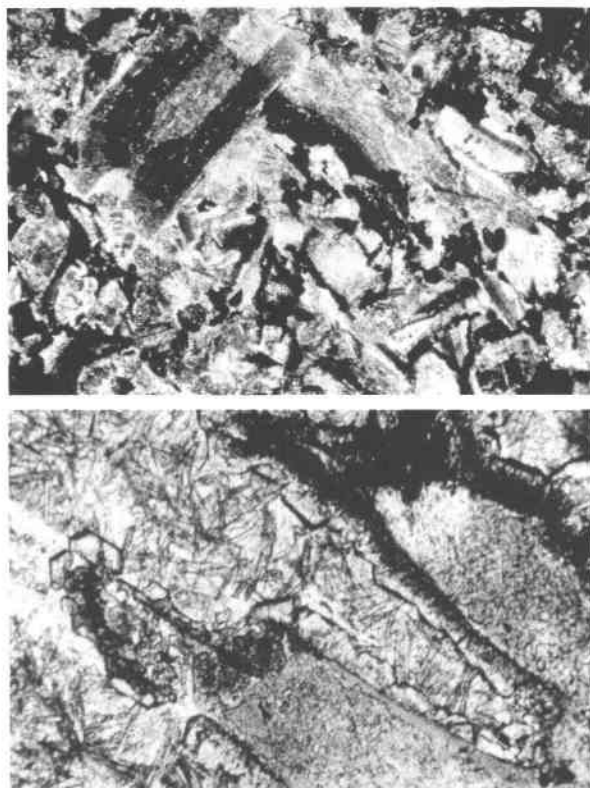


Fig. 68. Photomicrographs of metamorphosed diabase. **Top:** Ophitic texture of plagioclase laths with abundant coronas of garnet at extinction under crossed nicols. **Bottom:** Plagioclase laths with epidote needles separated from fine-grained areas of pyroxene and amphibole by coronas of high-relief garnet. Provided by P.-G. Andréasson.



Fig. 69. Photomicrograph of margin of original igneous pyroxene in metamorphosed diabase boudin. Rim area is suffused with exsolved ilmenite, and remaining pyroxene in rim area contains up to 8% jadeite content. Provided by P.-G. Andréasson.

epidote believed to have been included during prograde garnet growth. The garnet shows apparent prograde growth zoning from $X_{Mg} = 0.35$ to 0.40, and the garnet contains omphacite inclusions with composition Jd31 as compared to Jd35 in the coarse matrix omphacite. Omphacite breakdown to symplectite of less sodic pyroxene + plagioclase is localized near the garnet contact. Closest to the garnet, this consists of pyroxene Jd25 and An20; at the symplectite edge farthest from garnet, it consists of pyroxene Jd11 and An25. Very commonly the pyroxene of this reaction symplectite is replaced by hornblende. Santallier (1988) has estimated conditions of eclogite formation at about 700 °C and 19 kbar. Detailed studies of Nd-Sm mineral isochrons in the eclogites (Mørk et al., 1988) suggest that the eclogite-facies metamorphism took place at 500 Ma in the so-called Finnmarkian event, well before the Scandian final emplacement of the thrust sheets.

In approximately the southern one-third of the map area all of the mafic boudins are recognizable texturally as diabase, with a fine ophitic texture of former igneous plagioclase and, locally, plagioclase phenocrysts recognizable in hand specimen. The form of the boudins (Fig. 67) and the character of the enclosing rocks is the same. In thin sections (Fig. 68) there seems to be evidence for two metamorphic events, an early greenschist-facies metamorphism with formation of epidote in plagioclase and a later high-pressure event with the formation of delicate garnet coronas between plagioclase and mafic minerals in the matrix. Some original Ti-rich calcic igneous pyroxene is still preserved (Fig. 69) but has been attacked marginally with exsolution of Fe-Ti oxides and shows an increase in jadeite component up to 8%. It appears that these rocks were just beginning the process of eclogite formation but had gone through very much less of the process as compared with similar former dikes just a kilometer or two away to the north on the same mountain. What was it that caused these dramatic differences in reaction progress within the confines of a small area in a single thrust sheet? Andréasson and coworkers are now

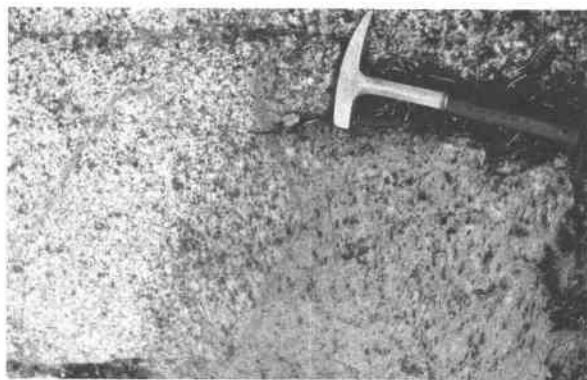


Fig. 70. Pavement outcrop at Vinddølalalen, Trollheim, Norway, showing progressive development of eclogite along a shear zone in coarse ophitic gabbro (shown to left). Gray areas are dominated by garnet, which tends to form coronas around large remaining plagioclases. Photograph by the author.



Fig. 71. Zone of Silurian eclogite (dark gray) parallel to quartz vein cutting across old foliation in Proterozoic anorthosite gneiss from the Bergen Arcs, Norway. From Boundy et al. (1992).

striving to understand the meaning of these differences by studying mafic boudins in a transitional position midway up the slope.

Eclogitized Proterozoic gabbro, Vinddølalalen, western Trollheim, Norway

My second example comes from exposures of Proterozoic basement rock about 100 m below a basement-cover contact in Vinddølalalen, western Trollheim (Fig. 61), studied by Tørudbakken (1981) and visited by me in 1981. Here, there is an extensive sequence of Caledonide thrust sheets, and both basement and tectonic cover have been very complexly refolded. Within this basement there is a boudin of coarse gabbro and metamorphosed gabbro about 180 m thick. Within the core of the boudin, the gabbro appears pristine, with large plagioclase laths commonly retaining finely exsolved opaque dust. Locally within the gabbro, there are shear zones, several of which are well exposed on polished pavements (Fig. 70). Along the margins of the shear zones the gabbro shows garnet coronas around plagioclase, and inward the rock changes to a gneissic rock with some relict plagioclase and then to an eclogite gneiss. The outer margins of the gabbro and eclogite against Proterozoic gneissic country rock are everywhere converted into garnet amphibolite or even garnet-free amphibolite, in some locations apparently produced by hydration directly from gabbro, in others from eclogite. Tørudbakken (1981) has estimated conditions of eclogite formation at 630 °C and 9–10 kbar. At Vinddølalalen it appears that the eclogite-forming reaction was catalyzed only in shear zones in the gabbro and that eclogite was preserved only where protected from later amphibolitization away from the margins of the gabbro boudin.

Eclogitized Proterozoic garnet granulite, Bergen Arcs, western Norway

My third example comes from the Bergen Arcs of western Norway. This was brought to my attention by a talk at the American Geophysical Union (Boundy et al., 1990)

and related papers (Austreim and Griffin, 1985; Austreim, 1987; Boundy et al., 1992). The rocks in question are exposed on Holsnøy, 30 km northwest of Bergen. They belong to a Caledonide thrust sheet of granulite-facies gneisses metamorphosed about 900 Ma, which have been variably subjected to eclogite-facies recrystallization during the Silurian Scandian orogeny.

Of the approximately 3 × 5 km field area mapped in detail by Boundy et al. (1990), about 30% consists of large ovoid areas of granulite-facies anorthosite gneiss with a very strong Proterozoic metamorphic foliation trending north-northeast. These areas contain numerous cracks and veins, commonly perpendicular to the old foliation, along which the anorthosite gneiss has been converted to eclogite (Fig. 71). About 15% of the area consists of well-foliated eclogite-facies rocks, with foliation trending irregularly east. Most of the remaining 55% of the area, between the areas of granulite-facies gneiss and the well-foliated eclogite-facies rocks, consists of eclogite breccia, composed of 40 to 80% foliated eclogite, enclosing blocks, some up to tens of meters in diameter, of relatively well-preserved granulite-facies anorthosite gneiss.

Conditions of the Proterozoic granulite-facies metamorphism are estimated to have been 850 °C and 8 kbar. The Silurian eclogite-facies conditions are estimated to have reached 700–750 °C and 16–21 kbar, yet some of the granulite-facies rocks show little mineralogical evidence of this later event. While it is obvious in this example that shearing has been involved in promoting the eclogite-forming reaction in the large areas of well-foliated eclogitic rocks, fluids moving along younger cracks in older foliated rocks may have been highly significant in promoting eclogite-facies reactions. The grand scale of eclogite development within a very large body of fluid-poor protolith may have minimized the effects of subsequent amphibolite-facies retrograde hydration, which is limited to a very small part of the study area.

Summary

In the last segment of this paper I have attempted to demonstrate the importance of structural environment,

kinetics, and fluids in producing different minerals from the same rock composition along the same P - T path. Lacking knowledge of the surroundings, there would have been no way to know that the original high- T rocks survived eclogite-facies overprinting with only very limited or no mineralogical traces.

CONCLUSION

What I have been preaching above might have been given from this pulpit 10–15 years ago, or at any time in the Society's history: namely, that if one is going to obtain meaningful results from analyses of minerals in metamorphic rocks, one must understand the textural relations in the rocks, the field relations of the rocks, the sequence of reactions and diffusional processes that the rocks went through, the features of the chemical space within which the rock composition lies, and the detailed way in which the analytical results relate to the metamorphic processes. These ideas were well understood by some petrologists going back nearly a century, but it seemed to me they did not touch a majority until fairly recently. It was at a meeting in Dublin, Ireland, in 1987 that I saw a homogeneous group nearly all thinking along these lines. So, I think, my message is less desperately needed but still worthwhile to impart.

In each of the examples I have given, one can see wrong avenues down which it would be easy to wander, particularly when the scope of one's thinking or information is limited. The most important point is to develop a stance or viewpoint, based on observational knowledge and experience, that will permit one to get closer and closer to the truth about what actually has happened. Although we view the rocks as our friends, they can in some places tell lies to us. To get to the truth behind these lies requires perpetual training of the eye of the petrographer and the mind of the petrologist.

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The section on the wollastonite marble is almost totally the development of former Ph.D. student H.N. Berry, my only contribution being to suggest that his weeks of graphical adventure would indeed prove very interesting. He gave permission to use a small part of his larger work as an example of the importance of petrographic observations at the scale of a whole thin section or hand specimen. The section on Scandinavian eclogites would not have been possible without the generous and timely contributions of P.-G. Andréasson of Lund University, A.G. Krill of

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REFERENCES CITED

- Andréasson, P.-G., Gee, D.G., and Sukotji, S. (1985) Seve eclogites in the Norrbotten Caledonides, Sweden. In D.G. Gee and B.A. Sturt, Eds., *The Caledonide orogen—Scandinavia and related areas*, p. 887–901. Wiley, London.
- Austrheim, Håkon (1987) Eclogitization of the lower crustal granulites by fluid migration through shear zones. *Earth and Planetary Science Letters*, 81, 221–232.
- Austrheim, Håkon, and Griffin, W.L. (1985) Shear deformation and eclogite formation within granulite facies anorthosites of the Bergen Arcs, western Norway. *Chemical Geology*, 50, 267–281.
- Berry, H.N., IV (1989) A new stratigraphic and structural interpretation of granulite-facies metamorphic rocks in the Brimfield-Sturbridge area, Massachusetts and Connecticut, 330 p. Ph.D. thesis, University of Massachusetts, Amherst, Massachusetts.
- (1991) Wollastonite-in and wollastonite-out reactions in central Massachusetts. *Geological Society of America Abstracts with Programs*, 23, 8.
- Boundy, T.M., Fountain, D.M., and Austrheim, Håkon (1990) Development of seismic anisotropy, structure, and petrofabrics of deep crustal eclogite facies shear zones in the Bergen Arcs, Norway (abs.). *Eos*, 71, 563.
- (1992) Structural development and petrofabrics of eclogite facies shear zones, Bergen Arcs, western Norway: Implications for deep crustal deformational processes. *Journal of Metamorphic Geology*, 10, 1–21.
- Cygan, R.T., and Lasaga, A.C. (1985) Self-diffusion of magnesium in garnet at 750–900 C. *American Journal of Science*, 285, 328–350.
- Elphick, S.C., Ganguly, J., and Loomis, T.P. (1985) Experimental determination of cation diffusivities in aluminosilicate garnets: I. Experimental methods and interdiffusion data. *Contributions to Mineralogy and Petrology*, 90, 36–44.
- 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*, 66, 113–117.
- Freer, R. (1979) An experimental measurement of cation diffusion in almandine garnet. *Nature*, 280, 220–222.
- Gromet, L.P., and Robinson, Peter (1990) Isotopic evidence for late Paleozoic gneissic deformation and recrystallization in the core of the Pelham dome, Massachusetts. *Geological Society of America Abstracts with Programs*, 22, A368.
- Hess, P.C. (1969) The metamorphic paragenesis of cordierite in pelitic rocks. *Contributions to Mineralogy and Petrology*, 24, 191–207.
- (1971) Prograde and retrograde equilibria in garnet-cordierite gneisses in south-central Massachusetts. *Contributions to Mineralogy and Petrology*, 30, 177–195.
- Hollocher, Kurt T. (1981) Retrograde metamorphism of the Lower Devonian Littleton Formation in the New Salem area, west-central Massachusetts, contribution no. 37, 265 p. Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.
- (1987) Systematic retrograde metamorphism of sillimanite-staurolite schists, New Salem area, Massachusetts. *Geological Society of America Bulletin*, 98, 621–634.

- Karabinos, Paul (1983) Polymetamorphic garnet zoning from southeastern Vermont. *American Journal of Science*, 284, 1008–1025.
- Mørk, M.B.E., Kullerød, Kåre, and Stabel, Arne (1988) Sm-Nd dating of Seve eclogites, Norrbotten, Sweden—Evidence for early Caledonian (505 Ma) subduction. *Contributions to Mineralogy and Petrology*, 99, 344–351.
- Osberg, P.H., Tull, J.F., Robinson, Peter, Hon, Rudolph, and Butler, J.R. (1989) The Acadian orogen. In R.D. Hatcher, Jr., W.A. Thomas, and G.W. Viele, Eds., *The Appalachian-Ouachita Orogen in the United States*. In Geological Society of America's *The geology of North America*, F-2, 179–232.
- Robinson, Peter (1983) Realms of regional metamorphism in southern New England, with emphasis on the eastern Acadian metamorphic high. In P.E. Schenk, Ed., *Regional trends in the geology of the Appalachian-Caledonian-Hercynian-Mauritanide Orogen*, p. 249–258. Reidel, The Netherlands.
- Robinson, Peter, and Tucker, R.D. (1991) Implications of new U-Pb zircon and monazite ages in the Bronson Hill anticlinorium, Massachusetts: Proterozoic source regions, a major overthrust, and Acadian versus Carboniferous metamorphism. *Geological Society of America Abstracts with Programs*, 23.
- Robinson, Peter, Tracy, R.J., and Ashwal, L.D. (1975) Relict sillimanite-orthoclase assemblage in kyanite-muscovite schist, Pelham dome, west-central Massachusetts. (abs.) *Eos*, 56, 466.
- Robinson, Peter, Tracy, R.J., Hollocher, K.T., and Dietsch, C.W. (1982) High grade Acadian regional metamorphism in south-central Massachusetts. In *New England Intercollegiate Geological Conference, 74th Annual Meeting*, Storrs, Connecticut, Guidebook, 289–339.
- Robinson, Peter, Tracy, R.J., Hollocher, K.T., Schumacher, J.C., and Berry, H.N., IV (1986) The central Massachusetts metamorphic high. In Peter Robinson and D.C. Elbert, Eds., *Field trip guidebook: Regional metamorphism and metamorphic phase relations in northwestern and central New England*, contribution no. 59, p. 195–266. Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.
- Robinson, Peter, Tracy, R.J., Hollocher, K.T., Berry, H.N., IV, and Thomson, J.A. (1989) Basement and cover in the Acadian metamorphic high of central Massachusetts. In C.P. Chamberlain and Peter Robinson, Eds., *Styles of metamorphism with depth in the central Acadian high, New England*, contribution no. 63, p. 69–140. Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.
- Roll, M.A. (1987) Effects of Acadian kyanite-zone metamorphism on relict granulite-facies assemblages, Mount Mineral Formation, Pelham dome, Massachusetts, contribution no. 60, 202 p. Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.
- Santallier, D.S. (1988) Mineralogy and crystallization of the Seve eclogites in the Vuoggatjålme area, Swedish Caledonides of Norrbotten. *Geologiska Föreningens i Stockholm Föreläsningar*, 110, 89–98.
- Schumacher, J.C., Schumacher, Renate, and Robinson, Peter (1989) Acadian metamorphism in central Massachusetts and southwestern New Hampshire: Evidence for contrasting *P-T* trajectories. In J.S. Daly, R.A. Cliff, and W.D. Yardley, Eds., *Evolution of metamorphic belts*, p. 453–460. Geological Society Special Publication No. 34, London.
- Schumacher, J.C., Hollocher, K.T., Robinson, Peter, and Tracy, R.J. (1990) Progressive reactions and melting in the Acadian metamorphic high of central Massachusetts. In J.R. Ashworth, and M. Brown, Eds., *High-temperature metamorphism and crustal anatexis*, p. 198–234. The Mineralogical Society Series, Unwin Hyman, London.
- Spear, F.S. (1991) On the interpretation of peak metamorphic temperatures in light of garnet diffusion during cooling. *Journal of Metamorphic Geology*, 9, 379–388.
- Spear, F.S., and Florence, F.T. (1991) Thermobarometry in granulites: Pitfalls and new approaches. *Journal of Precambrian Research*, in press.
- Spear, F.S., and Selverstone, Jane (1983) Quantitative *P-T* paths from zoned minerals: Theory and tectonic applications. *Contributions to Mineralogy and Petrology*, 83, 348–357.
- Spear, F.S., Selverstone, Jane, and Hickmott, D.D. (1990) Metamorphic consequences of thrust emplacement, Fall Mountain, New Hampshire. *Geological Society of America Bulletin*, 102, 1344–1360.
- Thompson, P.J. (1985) Stratigraphy, structure, and metamorphism in the Monadnock Quadrangle, New Hampshire, contribution no. 58, 191 p. Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts.
- Tørudbakken, B. (1981) Day 1. Trondheim-Surnadal. In W.L. Griffin and M.B.E. Mørk, Eds., *Eclogites and basal gneisses in western Norway*, excursion B-1, p. 13–19. Uppsala Caledonide Symposium.
- Tracy, R.J. (1982) Compositional zoning and inclusions in metamorphic minerals. In *Mineralogical Society of America Reviews in Mineralogy*, 10, 355–397.
- Tracy, R.J., and Robinson, Peter (1983) Acadian migmatite types in pelitic rocks of central Massachusetts. In M.P. Atherton, and C.D. Gribble, Eds., *Migmatites, melting and metamorphism*, p. 1163–1173. Shiva, Cheshire, U.K.
- 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*, 61, 762–775.
- Tucker, R.D., and Robinson, Peter (1990) Age and setting of the Bronson Hill magmatic arc: A re-evaluation based on U-Pb zircon ages in southern New England. *Geological Society of America Bulletin*, 102, 1404–1419.
- (1991) Age of inherited components in composite, single zircons from the Dry Hill Gneiss, central Massachusetts: Evidence concerning the “basement” of the Bronson Hill magmatic arc. *Geological Society of America Abstracts with Programs*, 23, 141.
- Tucker, R.D., Robinson, Peter, and Hollocher, K.T. (1988) U-Pb zircon, titanite, monazite dating in “basement” rocks of the Bronson Hill anticlinorium, central Massachusetts. *Geological Society of America Abstracts with Programs*, 20, A216.

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