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Magmatic garnets from the Cardigan pluton and the Acadian thermal event in southwest New Hampshire

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ABSTRACT

Samples from the Cardigan pluton preserve magmatic garnet-biotite temperatures and record an intrusive cooling history. Cardigan garnets do not seem to have formed in situ nor to have been reset by regional metamorphism. The interiors of Cardigan garnets are compositionally homogeneous, reflecting high-temperature, rapid volume diffusion. Garnet core–matrix biotite temperatures range from 750 to 900°C, with samples from the margins of the Cardigan pluton recording higher temperatures than those from the interior of the pluton. On the other hand, garnets from metapelites surrounding the Cardigan pluton have different compositions and preserve lower temperatures (500–700°C) than Cardigan garnets. The rims of Cardigan garnets are steeply zoned, reflecting incomplete re-equilibration at lower temperatures. Temperatures calculated from garnet rim and adjacent biotite compositions are approximately 400°C, reflecting the last temperature of equilibration.

These relationships provide constraints on the relative timing of emplacement of the Cardigan pluton and regional metamorphism. K-feldspar–cordierite zone metamorphism preserved in metapelites surrounding the Cardigan pluton could not have been a regional event that postdated the intrusion of the Cardigan pluton. These data contrast with previous work that suggested that high-grade Acadian isograds crosscut the Cardigan pluton.

INTRODUCTION

Southwest New Hampshire is characterized by abundant igneous and high-grade metamorphic rocks formed during the Acadian (Devonian) orogeny. It is an ideal place to study the interplay between magmatism and metamorphism and to address the long-standing problem of what provides the heat source for both processes. In highly migmatized areas, "ultrametamorphism" seems clearly responsible for the generation of granite; in other areas, contact aureoles seem to provide clear evidence that intrusion of a hot magma metamorphosed the country rocks. This study focuses on the thermal history of a large granitic pluton in southwest New Hampshire, as deduced from the compositional zoning of its garnets, and relates this thermal history to that of the surrounding metapelites.

Garnet is a useful petrogenetic indicator not only because it is stable over a wide range of temperatures and so can be employed in a number of geothermometers, but also because cations diffuse sluggishly in garnet and so a zoned garnet can record a substantial part of a rock's pressure-temperature-time path (cf. Spear et al., 1984). Recent work by Chamberlain and Lyons (1983) shows that the distribution of temperatures acquired using garnet-biotite geothermometry closely matches the distribution of metamorphic isograds in southwest New Hampshire. Chamberlain and Lyons extended K-feldspar–cordierite isograds across synorogenic plutons of the New Hampshire Plutonic Series (Fig. 1) on the basis of the facts that some of these plutons contain K-feldspar, cordierite, sillimanite, and garnet and that the few samples they had analyzed from the plutons recorded very high temperatures. The Cardigan pluton is a large body of garnet-bearing granite that lies in the middle of the proposed high-grade metamorphic ridge (Fig. 1). This study uses the garnet-biotite geothermometer and the shape of garnet zoning profiles to show that Cardigan garnets are magmatic and that high-grade Acadian isograds do not crosscut the Cardigan pluton.

Cardigan pluton

The New Hampshire Plutonic Series is a group of granitic intrusions that represents the major magmatic pulse of the Acadian orogeny in southwest New Hampshire. There are four members of the New Hampshire Plutonic Series: from oldest to youngest, the synorogenic Kinsman, Bethlehem, and Spaulding intrusive suites, and the postorogenic Concord intrusive suite (Fig. 2). The Cardigan pluton, one of the six plutons in New Hampshire composed of rocks of the Kinsman suite, is a 20 × 90 km sheet no more than 2.5 km thick (Nielson et al., 1976) that strikes parallel to the axis of the Kearsarge–Central Maine synclinorium and extends southward into Massachusetts as the Coys Hill granite. A rind of flecky gneiss

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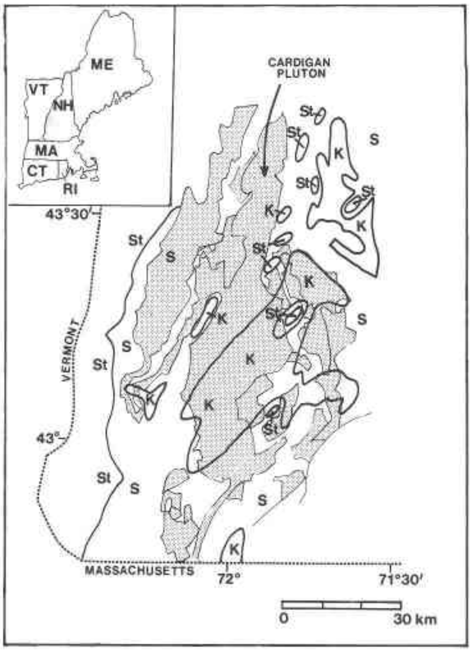

Fig. 1. Acadian metamorphic isograds in southwest New Hampshire after Chamberlain and Lyons (1983) superimposed upon New Hampshire Series plutons (stipple). St = staurolite; S = sillimanite; K = K-feldspar–cordierite zone metamorphism.

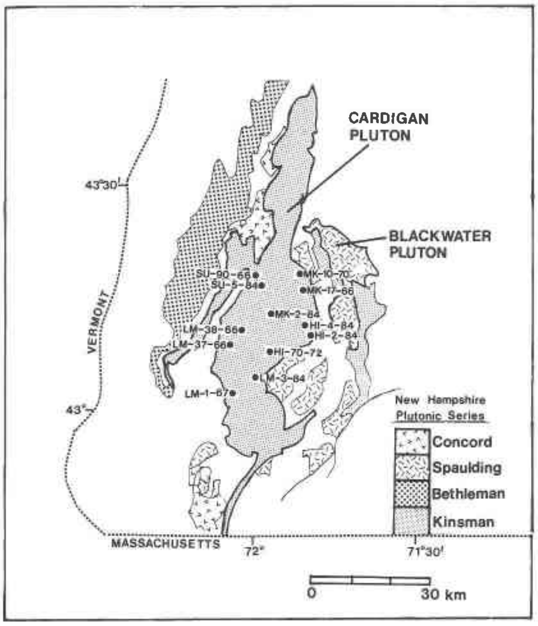

Fig. 2. New Hampshire Series plutons in southwest New Hampshire. Sample numbers refer to Cardigan pluton samples analyzed in this study.

that borders the Cardigan pluton has been interpreted as a contact aureole (Nielson, 1974).

Rocks of the Cardigan pluton range in composition from tonalite to granite (depending upon the relative abundance of K-feldspar) and are considered S-type granitoids that have been derived by partial melting of a pelitic protolith (Clark and Lyons, 1986). The Cardigan pluton is well foliated, indicating its intrusion during early stages of folding. Cardigan rocks have striking textures due to the abundance of large K-feldspar megacrysts (up to 12 cm) and garnet-biotite clots (up to 2 cm) and, in addition, contain quartz, plagioclase, muscovite, and trace sillimanite, cordierite, ilmenite, graphite, apatite, monazite, zircon, and allanite. Garnets range in size from 1–20 mm and compose typically 5 but up to 20 modal percent of the rock. Cardigan garnets are relatively inclusion free, in contrast to the seive-textured garnets in the surrounding metapelites. There are as many as five different populations of biotite within a single sample of the Cardigan pluton. Biotite is most abundant as large (up to 1 cm) red-brown grains in the matrix (typically 10–20 modal percent), but may also occur intergrown with garnet, surrounding garnet rims as biotite-sillimanite symplectites, or replacing garnet along cracks and rims as pale green secondary biotite. In the northern and southern extremities of the Cardigan pluton, biotite has completely replaced garnet in what has been interpreted as a late-stage deuteric reaction (Lyons et al., 1973). For this reason, samples analyzed in this study come from the middle of the Cardigan pluton where garnets are large, euhedral, and for the most part free of the pale parasitic biotite (see Fig. 2 for sample locations).

Of particular interest are garnet-rich pods (ranging in size from 1 m to 100 m) within the Cardigan pluton. These "garnetites" contain sillimanite, cordierite, plagioclase, and up to 65 modal percent garnet and grade into typical Cardigan rocks or pelitic xenoliths (Clark and Lyons, 1986). Clark (1972) suggested that the garnetite represents restite dragged up with the Cardigan crystal mush.

Mineral compositions

Major-element compositions of more than 150 garnets and biotites from twenty Cardigan samples were determined by electron-microprobe analysis. Most of the data were collected on a JEOL-JXA-50A microprobe at the Graduate School of Oceanography, University of Rhode Island, with an accelerating voltage of 15 kV, a beam current of 0.02 μA for biotites and 0.025 μA for garnets, and counting intervals of 20 s. Natural and synthetic silicates and oxides were used as standards (see Schilling et al., 1985, for details). Three to five biotites and at least three garnets were analyzed per sample. All rim compositions are within 10 μm of the crystal edge. Ten garnets were profiled using the microprobes at Harvard University and Lamont-Doherty Geological Observatory (under similar analytical conditions) and the microprobe at Dartmouth College (using standardless energy-dispersive spectra). Although several microprobe facilities were used, interlaboratory biases are not a concern since all quantitative analyses were gathered at Rhode Island, and the other

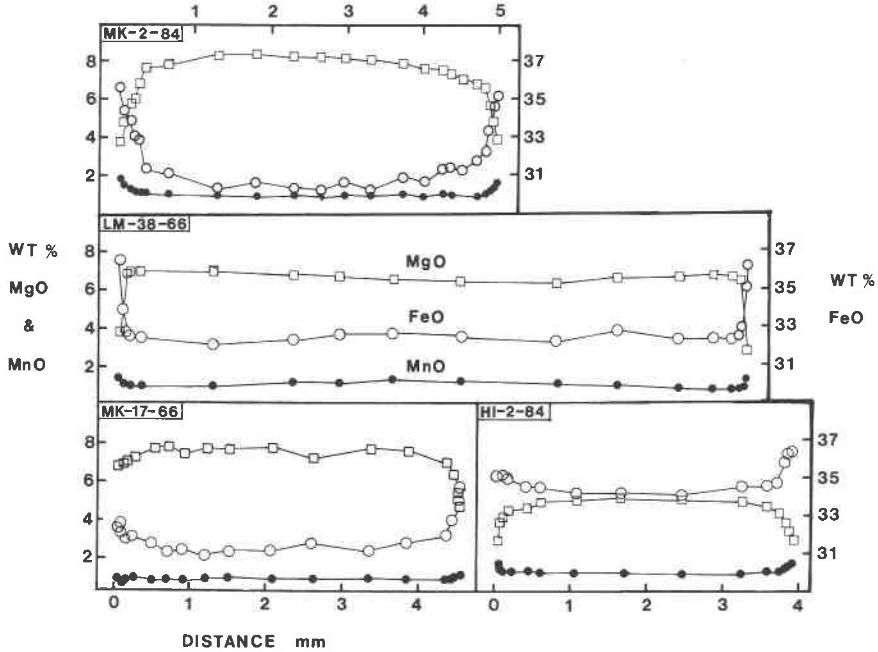

Fig. 3. Rim-to-rim profiles of four Cardigan garnets. Profiles are oriented with garnet-biotite contacts on the right and garnet-quartz, garnet-sillimanite, or garnet-plagioclase contacts on the left.

three microprobes were used only to determine the shapes of intragranular zoning profiles.

Garnet

Figure 3 shows rim-to-rim profiles of four representative garnets from the Cardigan pluton. The important feature of these profiles is the contrast between uniform interior compositions and steep compositional gradients within the garnet's outer 100 μm. In all cases, the largest garnet in a slide was profiled in an attempt to maximize the possibility of a central cut through the garnet. Even so, the effect of a noncentral cut through the garnet would be to merely broaden the compositional gradient at the rim. All ten Cardigan garnets that were profiled, including one from a garnetite pod, have flat, homogeneous cores and steeply zoned rims enriched in Fe and Mn and depleted in Mg. The compositional gradients appear steepest where the garnet is in direct contact with a biotite (e.g., the stronger Fe enrichment in the right-hand rims of HI-2-84 and MK-17-66 in Fig. 3). Ca is in low abundance (less than 1.5 wt%) and unzoned in all garnets from the Cardigan pluton.

Figure 4 compares the core compositions of garnets from the Cardigan pluton to core compositions of garnets from the surrounding metapelites. Cardigan garnets are distinctly more Mg rich and Mn poor than garnets from the metapelites. The variability in the garnets from the metapelites is due to the differences in metamorphic grade, with the high-Mn cores being characteristic of low-grade growth-zoned garnets, and the lower-Mn cores being characteristic of high grades where high-temperature diffusion homogenizes Mn throughout the garnet (cf. Tracy, 1982). It is the high-temperature garnets from the metapelites that overlap with some Cardigan garnets.

Biotite

Large red-brown biotites within the matrix, far from garnet, are richest in Fe and Ti. Biotites included within garnets are extremely magnesian and are zoned with Mg/

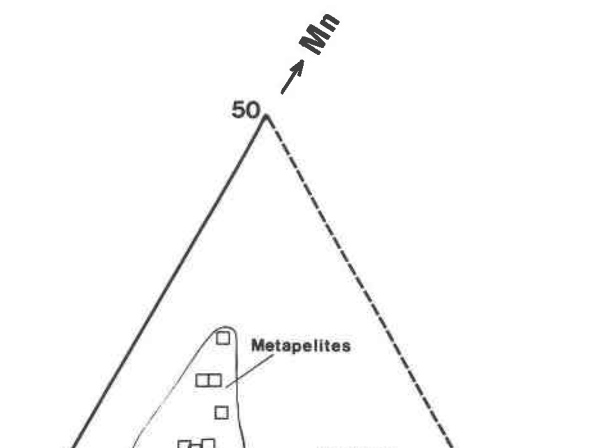

Fig. 4. Core compositions (in atom percent) of garnets from the Cardigan pluton (this study) and the surrounding metapelites (data from Chamberlain, 1981).

TABLE 1. Representative microprobe analyses of Cardigan garnets and biotites

Sample:	HI284				MK284		MK1766		LM3866				SU9066			
Domain:	GAR.C	BIO.M	GAR.R	BIO.R	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.R	BIO.R	GAR.C	BIO.M	GAR.R	BIO.R
Grain size (mm):	4 × 4	2	4 × 4	1.5	5 × 7	1	4 × 5	1	7 × 8	3	7 × 8	2	7 × 6	5	5 × 5	1
SiO_2	37.55	35.22	36.42	34.90	38.55	36.79	38.07	35.42	38.12	34.84	36.02	35.48	39.00	35.79	37.26	35.49
TiO_2	0.02	4.57		2.17		1.79	0.02	5.08		3.71		0.15	0.01	2.93		1.04
Al_2O_3	22.90	17.98	21.08	18.43	22.69	20.35	22.42	16.09	22.41	17.03	20.90	19.71	22.52	18.01	21.46	19.16
FeO	33.97	21.29	37.54	19.15	30.08	15.07	30.92	16.53	32.48	17.59	37.27	16.94	29.31	17.50	36.97	13.55
MgO	4.85	7.81	1.92	9.89	8.01	12.45	7.26	11.39	6.50	10.24	1.92	11.62	7.76	11.02	2.85	13.27
MnO	0.93		1.88	0.06	0.77		0.56		0.98	0.03	2.06	0.06	0.78		1.44	0.02
CaO	1.14		1.14		1.33		1.20		1.25		1.17	0.19	1.23		1.33	
Na_2O		0.09		0.06		0.08		0.02		0.07		0.06		0.11		0.15
K_2O		9.79		9.96		9.48		9.95		9.49		9.60		9.46		9.81
Total	101.27	96.75	99.98	94.62	101.43	96.07	100.45	94.48	101.75	93.00	99.34	93.81	100.62	94.82	101.31	92.49
	Formulae based on 22 oxygens for biotite, 12 for garnet															
Si	2.94	5.35	2.97	5.37	2.96	5.41	2.96	5.41	2.95	5.42	2.96	5.42	3.00	5.43	2.98	5.42
Ti		0.52		0.25		0.20		0.58		0.43		0.02		0.33		0.12
Al	2.11	3.22	2.03	3.34	2.05	3.53	2.06	2.90	2.05	3.12	2.02	3.55	2.04	3.22	2.02	3.45
Fe	2.22	2.70	2.56	2.47	1.93	1.85	2.01	2.11	2.10	2.29	2.56	2.16	1.89	2.22	2.47	1.73
Mg	0.57	1.77	0.23	2.27	0.92	2.73	0.84	2.59	0.75	2.37	0.24	2.65	0.89	2.49	0.34	3.02
Mn	0.06		0.13	0.01	0.05		0.04		0.06		0.14	0.01	0.05		0.10	
Ca	0.10		0.10		0.11		0.10		0.10		0.10	0.03	0.10		0.11	
Na		0.03		0.02		0.02		0.01		0.02		0.02		0.03		0.04
K		1.90		1.96		1.78		1.94		1.88		1.87		1.83		1.91

Sample:	LM3766		LM384		MK1070		HI7072		LM167		SU584		MK284G		HI484	
Domain:	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M	GAR.C	BIO.M
Grain size (mm):	5 × 6	3	4 × 3	4	3 × 3	1	1 × 2	4	6 × 5	4	5 × 4	3	3 × 4	3	12 × 7	2
SiO_2	37.85	35.02	38.07	33.77	37.60	36.10	37.52	35.14	38.46	34.86	38.68	36.03	38.29	36.41	38.23	35.96
TiO_2	0.02	3.84	0.01	2.71		3.63	0.02	3.35		3.16		2.74	0.04	2.26		4.10
Al_2O_3	22.31	17.82	21.68	17.96	21.97	16.67	21.43	17.99	22.18	18.69	21.71	17.96	21.65	19.06	21.96	18.69
FeO	31.77	18.82	34.61	21.82	30.92	18.55	35.33	23.35	31.47	19.74	29.74	19.29	33.13	18.23	32.86	19.78
MgO	5.94	9.48	4.23	7.84	6.72	10.19	3.80	7.38	5.70	8.84	7.64	10.74	6.62	11.87	6.13	9.36
MnO	0.92	0.01	1.11		0.86		1.91		1.14		0.64		1.01		0.79	0.02
CaO	1.18		1.11		1.24		0.85		1.27		1.18		1.13		1.20	0.02
Na_2O		0.13		0.10		0.04		0.05		0.07		0.12		0.25		0.05
K_2O		9.55		9.54		9.94		9.52		9.61		9.13		8.80		9.34
Total	99.98	94.67	100.82	93.74	99.31	95.13	100.85	96.78	100.22	94.94	99.59	96.01	101.87	96.88	101.17	97.32
	Formulae based on 22 oxygens for biotite, 12 for garnet															
Si	2.98	5.37	3.01	5.33	2.97	5.51	2.99	5.38	3.01	5.35	3.02	5.43	2.97	5.39	2.98	5.36
Ti		0.44		0.32		0.42		0.39		0.36		0.31		0.25		0.46
Al	2.07	3.22	2.02	3.34	2.04	3.00	2.01	3.24	2.05	3.38	2.00	3.19	1.98	3.32	2.02	3.28
Fe	2.09	2.41	2.29	2.88	2.04	2.37	2.35	2.99	2.06	2.53	1.94	2.43	2.15	2.26	2.14	2.46
Mg	0.70	2.17	0.50	1.84	0.79	2.32	0.45	1.68	0.67	2.02	0.89	2.41	0.77	2.62	0.71	2.08
Mn	0.06		0.07		0.06		0.13		0.08		0.04		0.07		0.05	
Ca	0.10		0.09		0.10		0.07		0.11		0.10		0.09		0.10	
Na		0.04		0.03		0.01		0.01		0.02		0.04		0.07		0.01
K		1.87		1.92		1.93		1.86		1.88		1.75		1.66		1.78

Note: Domains are C, core of garnet, far from inclusions; M, biotite in the matrix, far from garnet (analyses usually from middle of biotite grain); R, analyses within 10 μm of mutual contact between garnet rim and biotite. MK284G is a "garnetite" sample.

Fe increasing toward the garnet contact, indicating chemical exchange with the host garnet. Biotites along garnet rims, whether intergrown, symplectic, or parasitic, are also zoned and more magnesian than biotites wholly within the matrix.

Geothermometry

There are several calibrations for the partitioning of Fe and Mg between coexisting garnet and biotite as a function of temperature. Two commonly used calibrations are the semi-empirical calibration of Thompson (1976) and the experimental calibration of Ferry and Spear (1978). Table 1 lists representative garnet and biotite analyses for samples from the Cardigan pluton, and Table 2 lists temperatures calculated for these samples using the Thompson calibration and the Ferry and Spear calibration compensated for minor-element effects (Ti and Al in biotite; Ca and Mn in garnet) on ideal binary Fe-Mg exchange (model B of Indares and Martignole, 1985). A pressure of 3.9 kbar was chosen for use in the Ferry and Spear equation based on the proximity of the Al_2SiO_5 triple-point isobar to the Cardigan pluton as mapped by Thompson and Norton (1968); however, the geothermometer is largely insensitive to pressure. For the Car-

TABLE 2. Garnet-biotite temperatures (°C) calculated from analyses given in Table 1

Sample	K_D	T_{Th}	T_{FS}
Core			
HI284	0.39	820	760
MK284	0.32	740	745
MK1766	0.34	765	675
LM3866	0.34	770	725
SU9066	0.42	855	875
LM3766	0.37	800	760
LM384	0.34	765	750
MK1070	0.4	830	810
HI7072	0.34	765	740
LM167	0.4	830	840
SU584	0.46	900	955
MK284G	0.31	730	730
HI484	0.39	820	780
Rim			
HI284	0.099	430	365
SU9066	0.079	395	340
LM3866	0.075	385	350

Note: T_{Th} = Thompson (1976) calibration; T_{FS} = Ferry and Spear (1978) calibration with the additions of Indares and Martignole (model B; 1985); K_D as in Fig. 5.

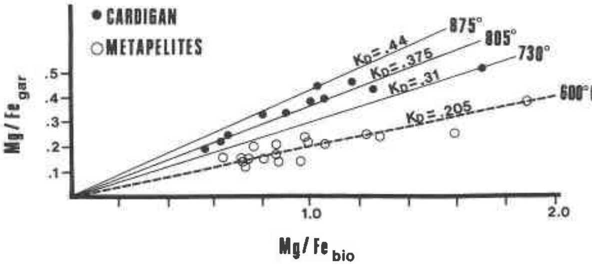

Fig. 5. Mg/Fe ratios of one garnet core–matrix biotite pair per sample. Metapelite data from Chamberlain (1981). Temperatures are calculated using the Thompson (1976) calibration. $K_D = (Mg/Fe)_{gar}/(Mg/Fe)_{bio}$.

digan samples, maximum calculated temperatures result from matching garnet core compositions with matrix biotite compositions. Values calculated with the Thompson calibration agree closely with those calculated with the compensated Ferry and Spear values, averaging 800°C for Cardigan samples. The metapelites surrounding the Cardigan pluton preserve garnet core–matrix biotite temperatures ranging from 500–700°C (Chamberlain, 1981). Within the errors of the calibrations (i.e., ±50°C for the Ferry and Spear calibration), there is little overlap between the temperatures calculated for the metapelite samples and temperatures calculated for Cardigan samples (see Fig. 5).

Compositions within 10 μm of all types of mutual garnet-biotite contacts (including garnet contacts with intergrown biotite, parasitic biotite, symplectic biotite, and included biotite) result in minimum calculated temperatures averaging 400°C for Cardigan samples.

Temperatures from the interior of the Cardigan pluton appear lower (750°C avg.) than those from the margin (850°C avg.; Fig. 6). This suggests slower cooling in the interior, allowing garnets and biotites to equilibrate at lower temperatures, whereas garnets and biotites from the margins were effectively quenched at higher temperatures.

A geothermometer is only as good as the assumptions behind choosing which mineral pairs realistically represent equilibrium. A number of observations lend support to the assumption made here that garnet cores and matrix biotites retain equilibrium compositions.

1. Cations diffuse much more sluggishly within garnet than within biotite (Lasaga, 1983). Diffusion within garnet is also much slower than intergranular diffusion within the matrix. This is why garnets are useful recorders of metamorphism; garnets store *P-T-t* information in their compositional zoning profiles whereas the matrix may undergo several discontinuous reactions. The unzoned interiors of Cardigan garnets indicate that diffusion was sufficiently rapid to homogenize Cardigan garnets and, therefore, rapid enough to keep garnets in communication with matrix biotite. In the case of Cardigan samples, diffusive exchange between garnet interiors and matrix biotite seems to have ceased below 800°C, presumably because of a rapid cooling rate. On the basis of a simple conductive cooling model for an intrusive sheet of infinite lateral extent (after Jaeger, 1968), the Cardigan pluton would cool to the ambient wallrock temperature on the order of 1 m.y., which translates to rapid cooling rates (1000–100 deg/m.y.) for a wide range of wallrock temperatures (400–700°C). The "geospeedometry" models of Lasaga (1983) predict effective blocking temperatures in the range observed (i.e., 750–850°C), for 5-mm garnets (typical for Cardigan garnets), using the above cooling rate estimates and the diffusion data of Lasaga et al. (1977).

2. Tracy et al. (1976) have pointed out that "false" high temperatures may be calculated if matrix biotites have become more Fe rich at the expense of garnet rims owing to retrograde reaction. However, because the volume of garnet available for reaction (the outer 100 μm) is small in comparison to the volume of matrix biotite in these samples, the threat of extensive re-equilibration of matrix biotite compositions due to retrograde reaction is minimal. Simple mass-balance calculations show that the compositional changes in the re-equilibrated biotite in the samples would propagate to temperature changes within the error of the geothermometry calibrations.

3. Biotites adjacent to garnets are much more magnesian than matrix biotites and are zoned approaching the garnet contact. The steep chemical gradients at the garnet rim, therefore, seem to represent re-equilibration with adjacent biotite over an ever-decreasing length scale, until diffusion finally ceased at 400°C, as indicated by the calculated rim temperatures. The Mn enrichment at garnet rims, however, cannot be explained by garnet-biotite cation exchange, because Mn does not partition into biotite. One explanation for the Mn-enriched garnet rims comes from the experimental work of Kress (1986) that

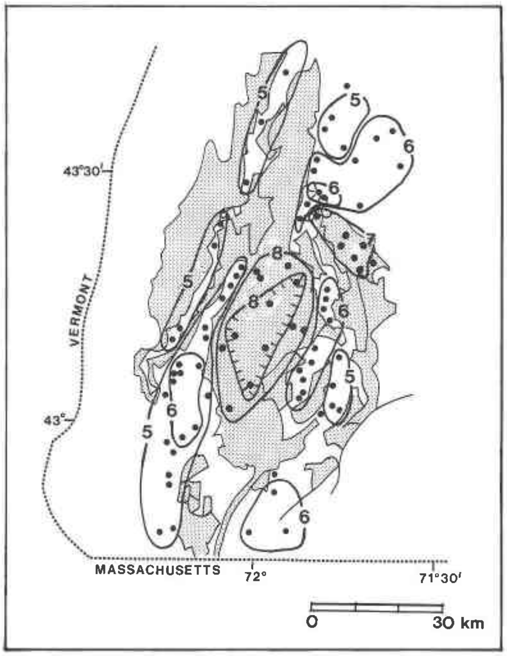

Fig. 6. Garnet core–matrix biotite temperatures (in °C/100; Thompson, 1976, calibration) of metapelites (data from Chamberlain, 1981), the Cardigan pluton (this study), and the Blackwater pluton (Chamberlain, 1981; Duke, 1984; and this study). Stipple = New Hampshire Series granites. Hachures point to lower temperature.

shows that as temperature decreases, Mn can be taken into garnet during Fe-Mn exchange with ilmenite, a phase also commonly in contact with garnet. Therefore, both the large relative volume of biotite and the seemingly local equilibration of garnet rims with neighboring phases provide support in favor for little change in matrix biotite compositions since equilibration with garnet cores.

Discussion

Origin of Cardigan garnet

Garnet-biotite temperatures recorded in Cardigan samples are higher than those of regional metamorphism, and Cardigan temperatures are zoned in a way that seems to reflect the cooling geometry of the pluton. Thus, both the magnitude and distribution of temperatures within the Cardigan pluton indicate equilibration during cooling of the magma and not during regional metamorphism. This conclusion contrasts with previous suggestions that Cardigan garnets are metamorphic (Heald, 1950; Barker, 1961; Chamberlain and Lyons, 1983). Further evidence in support of a magmatic origin for Cardigan garnets lies in their preservation of intrusive ages. A Sm-Nd garnet–whole-rock isochron yields the same age within error for the Cardigan pluton (413 ± 5 Ma; Barreiro and Aleinikoff, 1985) as both an Rb-Sr whole-rock isochron (402 ± 19 Ma; Lyons et al., 1982) and a concordant magmatic zircon (400 ± 6 Ma; M. Harrison, pers. comm.).

The problem of whether Cardigan garnets actually crystallized from a magma, or whether they represent restite, is difficult to resolve. White and Chappell (1977) concluded that restite minerals do not retain their source compositions, but are mostly in equilibrium with the granitic melt in which they are entrained. For this reason, it would be expected that both garnet phenocryst and restite preserve the same thermal history in their compositions, and so this study cannot distinguish between the two possible origins for magmatic Cardigan garnets. However, a restite origin seems most compatible with several characteristics of Cardigan garnets. Garnet phenocrysts are known to occur in peraluminous granites like the Cardigan (Miller and Stoddard, 1981; Propach and Gillessen, 1984; Clemens and Wall, 1984), but their compositions tend to follow an almandine-spessartine trend unlike that of Cardigan garnets. On the other hand, 10-kbar melting experiments of a pelitic composition (Green, 1976) predict almandine garnet (along with biotite, sillimanite, cordierite, and plagioclase) as a residual, or restite, phase in equilibrium with a granitic liquid. Garnets from the garnetite pods of the Cardigan pluton, good candidates for residues of partial melting because of their refractory mineral assemblage, are compositionally identical to typical Cardigan garnets and retain similar high temperatures and flat zoning profiles. These data are consistent with the recent work by Clark and Lyons (1986) that suggests that Cardigan garnets are restite minerals. Clark and Lyons (1986) have developed a model, based on Q-mode factor analysis of major elements and rare-earth-element modeling, that explains the compositional range of the Cardigan pluton in terms of mixtures of three components: a minimum melt granite, a garnet-rich restite, and a pelite. This model is especially appealing because all components are present in the field: the K-feldspar-rich Cardigan rocks, the garnetite pods, and the pelitic xenoliths. Thus, it is not necessary to call upon prolific crystallization of almandine garnet from a granitic melt nor metamorphic recrystallization to explain the garnet-bearing Cardigan pluton. Rather, a single partial melting–emplacement episode can explain the Cardigan pluton's unusual mineral assemblage and broad compositional range.

Thermal history of the Cardigan pluton

A great deal of information regarding a rock's cooling history can be extracted from garnet compositions by combining geothermometry with "geospeedometry" models (cf. Lasaga, 1983). Diffusion-induced zoning profiles document the extent of re-equilibration to lower temperatures, whereas a garnet core–matrix biotite temperature reflects an effective blocking temperature. Both zoning profiles and core temperatures are strongly dependent upon cooling rates. The following is a model for the thermal history of the Cardigan pluton that is consistent with both the garnet-biotite geothermometry and garnet-zonation studies.

The first stage in the thermal history of the Cardigan pluton was its intrusion and early rapid cooling. This stage was characterized by high temperatures at which diffusion was rapid enough to keep garnet interiors homogenized and in communication with matrix biotites. The temperatures recorded during this stage were primarily a function of the cooling rates within the different regions of the Cardigan pluton. Samples near the margin of the Cardigan pluton cooled fastest and preserved the highest blocking temperatures, providing the best estimate for the temperature of intrusion (850–900°C). The lowest magmatic temperatures (750–800°C) were recorded by samples from the middle of the Cardigan pluton, where cooling rates were slower than those at the pluton's margin, but still rapid compared to regional metamorphic cooling rates, which would not have permitted retention of high temperatures and homogeneous garnet cores.

The second stage in the thermal history of the Cardigan pluton was the regional metamorphic history shared with the surrounding metapelites. Even though Cardigan garnets preserve temperatures higher than those of regional metamorphism, it is unlikely that the Cardigan pluton escaped the thermal pulse felt by most of southern New Hampshire (much of which is sillimanite grade; see Fig. 1). The age of the Cardigan pluton (very early Devonian) precludes its postdating much, if any, Acadian metamorphism. The extensive zoning developed at the garnet rims is good evidence that Cardigan samples attempted to reequilibrate to the conditions of regional metamorphism. Based on the geospeedometry models of Lasaga (1983), ambient metamorphic temperatures could have not exceeded 650°C without completely resetting Cardigan garnets. Therefore, Cardigan garnet cores could have tolerated sillimanite-zone metamorphism (corresponding to temperatures of 500–600°C) without resetting, but not K-feldspar–cordierite zone metamorphism (600–700°C). This means that the K-feldspar–cordierite zone metamorphism preserved locally in the metapelites of southwest New Hampshire could not have been a regional event that postdated the intrusion of the Cardigan pluton.

There are three possibilities regarding the timing of the K-feldspar–cordierite zone metamorphism in southwest New Hampshire: either (1) it predated, (2) it was synchronous with, or (3) it was a local event that postdated the intrusion of the Cardigan pluton. Case 1 is unlikely because of the very early Devonian age of the Cardigan pluton. Evidence for Case 2 lies in the close proximity of K-feldspar–cordierite zone metapelites to New Hampshire Series plutons, suggesting that these high-grade rocks are a product of contact metamorphism. Case 3 finds support in the work of Chamberlain (1986) that shows how small-scale folding of isotherms in the root zone of the Fall Mountain nappe is responsible for local thermal highs and lows.

Because the K-feldspar–cordierite zone metamorphism evident in adjacent metapelites seems to find no expression in the Cardigan pluton, this study does not support the broad northeast-striking thermal ridge proposed by Chamberlain and Lyons (1983; see Fig. 1), but instead suggests limiting the K-feldspar–cordierite zone isograds to the metapelites surrounding the Cardigan pluton. Evidence that these highest-grade Acadian isograds do not crosscut other New Hampshire Series plutons lies in preliminary results from samples of the younger Spaulding intrusive suite from the Blackwater pluton (see Fig. 2). These latter results are based on calculated garnet core–matrix biotite temperatures for seven samples from Duke (1984), Chamberlain (1981), and this study that suggest that the Blackwater pluton also intruded at temperatures exceeding 800°C (see Fig. 6).

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