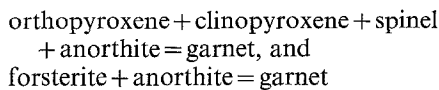


The Formation of Garnet in Olivine-Bearing Metagabbros from the Adirondacks

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Abstract. A regional study of olivine-bearing metagabbros in the Adirondacks has permitted testing of the P(pressure)–T(temperature)–X(composition) dependence of garnet-forming reactions as well as providing additional regional metamorphic pressure data. Six phases, olivine, orthopyroxene, clinopyroxene, garnet, plagioclase and spinel, which can be related by the reactions:



occur together both in coronal and in equant textures indicative of equilibrium. Compositions of the respective minerals are typically Fo_{25-72} , En_{44-75} , $\text{En}_{30-44}\text{Fs}_{9-23}\text{Wo}_{47-49}$, $\text{Pp}_{13-42}\text{Alm}_{39-63}\text{Gr}_{16-20}$, An_{29-49} and Sp_{16-58} . When they occur in the same rock, equant and coronal garnets are homogeneous and compositionally identical suggesting that chemical equilibrium may have been attained despite coronal textures. Extrapolating reactions in the simple CMAS² system to granulite temperatures and making thermodynamic corrections for solid solutions gives equilibration pressures (using the thermometry of Bohlen et al. 1980b) ranging from about 6.5 kb in the Lowlands and southern Adirondacks to 7.0–8.0 kb in the Highlands for the assemblage olivine-plagioclase-garnet. These results are consistent with inferred peak metamorphic conditions in the Adirondacks (Valley and Bohlen 1979; Bohlen and Boettcher 1981). Thus the isobaric retrograde path suggested by Whitney and McLelland (1973) and Whitney (1978) for the formation of coronal garnet in olivine metagabbros may not be required. Application of the same equilibria gives >8.7 kb for South Harris, Scotland and 0.9 kb for the Nain Complex. Disagreement of the latter value with orthopyroxene-olivine-quartz barometry (Bohlen and Boettcher 1981) suggests that the use of iron-rich rocks (olivines $\geq \text{Fa}_{50}$) results in errors in calculated pressures.

Introduction

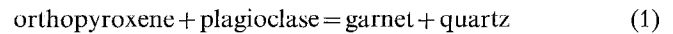
Garnet has classically been recognized as a high *P* phase in metamorphosed igneous rocks. Its formation can be described by the reaction:

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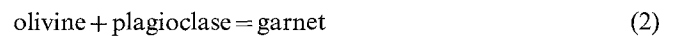
¹ Present address: Climax Molybdenum Co., P.O. Box 961, Battle Mountain, Nevada 89820, USA

² CMAS = CaO–MgO–Al₂O₃–SiO₂.

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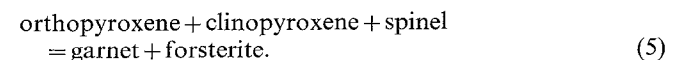
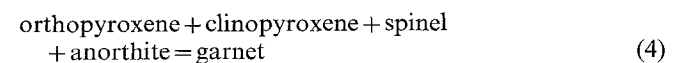
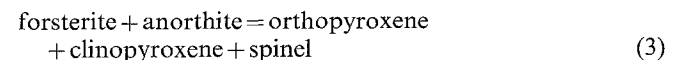
in oversaturated rocks and in undersaturated rocks as:



(Ringwood and Green 1964, 1966). In principle, assemblages such as orthopyroxene-plagioclase-garnet-quartz and olivine-plagioclase-garnet will fix *P* at a constant *T* for given mineral compositions (Wood and Banno 1973; Wood 1975; Schmid and Wood 1976). Garnet barometry is potentially widely applicable to granulitic terranes if it can be adequately calibrated. Reaction (2) is easier to adjust for variable solid solutions because olivines are well represented as binary solutions while complete description of natural orthopyroxenes requires consideration of variable MgSiO_3 , FeSiO_3 , CaSiO_3 and $\text{MgAl}_2\text{SiO}_6$ (Herzberg 1978a).

Two different kinds of experiments have been applied to garnet reactions: those on natural rock compositions and those on simple end-member reactions. Ringwood and Green (1964, 1966), Green and Ringwood (1967) and Ito and Kennedy (1971) have investigated garnet reactions in basaltic rocks but they concentrated mainly on the plagioclase-out reaction forming eclogites. Application of these experiments below 900° C is difficult in any case because of uncertainties in extrapolation: reversals are difficult to interpret in these multicomponent systems and the garnet-in boundary is unlikely to be a straight line (pace Green, Ito, Kennedy and Ringwood). Until experiments on multicomponent rock systems are completely and tightly reversed over a wide range of *T*, they will be difficult to apply to metamorphic rocks.

Garnet-forming reactions have also been investigated in the simple system CMAS. Several experimenters (Kushiro and Yoder 1966; O'Hara et al. 1971; Herzberg 1972, 1976; Herzberg and Chapman 1976; Presnall 1976; Jenkins and Newton 1979; O'Neill 1981) have investigated the reactions:



Reaction (5) has been reversed between 800 and 1,100° C (Jenkins and Newton 1979; O'Neill 1981) while reactions (3) and (4) are constrained only by half-reversals above

900° C. Obata (1976) and Herzberg (1978b) have shown that the straight line extrapolations to lower T made by McGregor (1970), Whitney and McLelland (1973) and others are inappropriate. The compositions of equilibrium CMAS pyroxenes slide with P and T due to Ca–Mg and MgSi–AlAl substitutions. Reactions (3), (4) and (5) curve markedly away from each other between 800 and 1,200° C as a result.

Whitney and McLelland (1973) and Whitney (1978) have applied phase equilibria in the CMAS system to olivine-bearing rocks in the Adirondacks. Using straight line extrapolations, they propose that reactions (3) and (4) cross at granulite T . They infer that metamorphic conditions at or near this crossing point were required to stabilize the six phase assemblage olivine-orthopyroxene-clinopyroxene-garnet-spinel-plagioclase in the Adirondacks. Sack (1980a) has calculated how several reactions, including (3) and (5), shift in P – T space with changing bulk composition. He has shown that the six phase assemblage of Whitney and McLelland (1973) is stable at $730 \pm 50^\circ$ C and 8 ± 1 kb when olivine has composition Fa_{34} and garnet has composition Gr_{18} or when olivine has composition Fa_{34} and plagioclase has composition An_{41} . Thus solid solutions can have substantial effects on the stability of phases in natural systems.

Garnet barometry based on results in the simplified CMAS system can be applied successfully to granulite terranes. This approach requires accurate phase equilibria in

the four component system at granulite T and knowledge of the effects of solid solutions to extend these results to complex natural systems.

Adirondack Olivine-Bearing Metagabbros

Quartz-undersaturated gabbros occur mainly in the eastern Adirondack Mountains as dikes, sills and small stocks from tens of meters to several kilometers in diameter (Buddington 1939). The rocks typically contain olivine, garnet, orthopyroxene, clinopyroxene and plagioclase with lesser amounts of pargasitic hornblende, biotite, green spinel and fluorapatite (Table 1). The locations of samples collected for this study are described in Appendix I. Texturally, the rocks fall into two categories. Magnesian rocks have their igneous textures preserved. Lath-like plagioclase crystals up to 5 millimeters in length are often clouded with minute inclusions of spinel (Whitney 1972; Whitney and McLelland 1973; McLelland and Whitney 1980a, b) and occasionally clinopyroxene. Coronas occur interstitial to texturally igneous pyroxenes and feldspars. Granular ortho- and clinopyroxenes rim olivine and are in turn rimmed by garnet (Fig. 1a). Coarser spinel occurs armoured by amphibole and biotite. More complete textural descriptions are in Whitney and McLelland (1973) and McLelland and Whitney (1980a, b). Mineralogically identical but more Fe-rich gabbros are completely recrystallized and have finer-

Table 1. Mineral assemblages from Adirondack metagabbros

	olivine	ortho-pyroxene	clino-pyroxene	garnet	amphibole	biotite	plagioclase	spinel	apatite	Fe–Ti oxides	sulfides
I87-1	c	i, r	i, r	c, e	×	×	i	p	×	×	×
SC-2	c	r	r	e	×	×	i, r	a, p	×	×	×
AS-1	c	i, r	i, r	c, e	×	×	i, r	a, p	×	×	×
TL-1	c	i, r	i, r	c, e	×	×	i, r	p	×	×	×
N-3	c	i, r	i, r	c, e	×	×	i	a, p			×
W-2	c	i, r	i, r	c, e	×	×	i, r	p	×		×
LP-2	c	i, r	i, r	c, e	×	×	i, r	p	×	×	×
LL-1	c	i, r	i, r	c, e	×	×	i, r	p	×	×	×
I87-4	e	r	r	e	×	×	r	p	×	×	×
233-1	c	i, r	i, r	c	×	×	i	p		×	×

i = coarse igneous appearing; r = recrystallized metamorphic appearing; c = coronal; e = equant, not associated with coronas; a = armoured by hydrous phases; p = minute inclusions in plagioclase

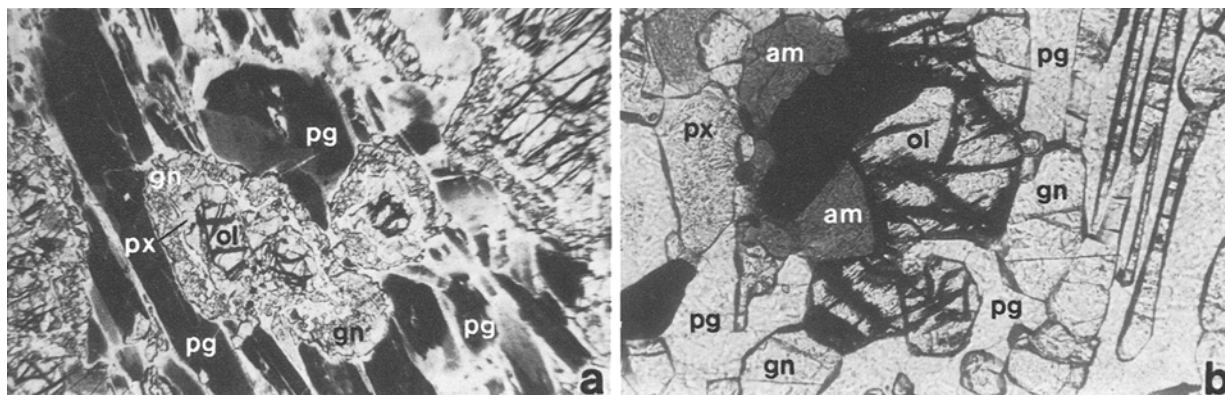


Fig. 1. a, b. (a) Well developed corona. Pyroxenes (px) surround olivine (ol) and are rimmed by garnet (gn). Euhedral plagioclase (pg) is clouded by minute spinel inclusions. Field of view is approximately 7.5 mm wide. (b) Recrystallized metamorphic fabric. Olivine, garnet, plagioclase, pyroxenes and amphibole (am) share grain boundaries. Field of view is approximately 0.8 mm wide. Both views in plane polarized light

grained, equant metamorphic textures. Garnet and olivine share grain boundaries without rimming relationships (Fig. 1b). Assumptions of equilibrium in coronal and equant garnet-olivine metagabbros will be tested below.

Analytical Procedures

Mineral analyses were obtained using the University of Michigan ARL-EMX electron microprobe analyzer equipped with LiF, PET and TAP crystal spectrometers. Standard operating conditions were 15 kV accelerating potential, 150 μ A emission current and 0.005 to 0.008 μ A sample current. Beam current was digitized. A minimum of 10,000 counts was recorded during analysis of standards which generally required counting times between 15 and 20 s. Spectrometer data for individual minerals were collected at four or more points and were averaged to obtain an analysis. Chemical heterogeneity of mineral grains, particularly garnets, was carefully monitored and was found rarely to exceed typical microprobe uncertainty, 2 or 3% of the amount of an element present. Natural, well analyzed almandine-rich garnet (C.O. Ingamells, unpubl.), Marjalahti olivine (Univ. of Michigan), Spring Mountain kaersutite (R. Binns, written comm.), aluminous clinopyroxene (A.J. Irving, written comm.), Nain labradorite (Univ. of Michigan), Gotthard adularia (Univ. of Michigan) and Tiburon albite (B.W. Evans, written comm.) were used as standards for silicates. Synthetic forsterite, corundum, rutile, ZnO, V₂O₅ and Ni₂SiO₄ in addition to naturally occurring hematite (Univ. of Michigan), chromite (M. Engi, written comm.) and almandine-rich garnet (C.O. Ingamells, unpubl.) were used as standards for spinel. All major and minor elements were identified in spinels by collecting emission spectra for 300 s using a Kevex Li-drifted silicon detector with Nuclear Data multichannel analyzer. This technique permitted detection of heavy elements ($Z \geq 11$) present in excess of approximately 0.1 wt.%. Spectrometer data

were corrected for atomic number, absorption, fluorescence, background and machine drift using the computer program EMPADR VII (Rucklidge and Gasparini 1969).

Ortho- and clinopyroxenes were normalized about 4 total cations, olivine and spinel about 3, plagioclase about 5 and garnet about 8 to obtain mineral formulae. The amount of ferric iron was inferred from stoichiometry. Garnet was assumed to have none.

Minute spinel and clinopyroxene inclusions in plagioclase, most of which were too fine to resolve using the electron microprobe, were identified and semiquantitatively analyzed using a University of Michigan JEOL JSM-U3 scanning electron microscope equipped with a Kevex Li-drifted silicon detector and multichannel analyzer. Analyses of minute spinels in plagioclase and coarser spinels enclosed in hydrous phases from the same rock are provided for samples SC-2 and W-2 (Table 2). Data from spinels in plagioclase are used in the following calculations.

Pyroxene analyses were obtained from smaller metamorphic appearing grains. The orthopyroxenes range from En₄₄Fs₅₄Wo₂ to En₇₅Fs₂₄Wo₁ with MgTs (MgAl₂SiO₆) components of 1 to 8 mol%. Clinopyroxenes vary from En₄₄Fs₉Wo₄₇ to En₃₀Fs₂₃Wo₄₇ with CaTs (CaAl₂SiO₆) contents of 4 to 12%. These pyroxenes also show Na in excess of Fe³⁺ allowing calculation of a nominal Jd (NaAlSi₂O₆) component of 0 to 5 mol% with most values at around 4%. The olivines vary from Fo₂₅ to Fo₇₂, the garnets from Pp₄₃Alm₃₉Sp₁Gr₁₇ to Pp₁₃Alm₆₃Sp₅Gr₂₀. Spinel range from Hc₄₀Sp₅₇Gh₁Mt₂ to Hc₇₆Sp₁₅Gh₃Mt₆ with very minor V, Mn, Ni, Ti and Cr in solution (Gh = ZnAl₂O₄). Plagioclases change from An₂₉Ab₆₇Or₄ to An₄₉Ab₄₉Or₁.

Phase Equilibria in the CMAS System

Herzberg and Chapman (1976) obtained a half-reversal of reaction (3) at 1,200° C and 8.9 kb. This half-reversal, the

Table 2a-f. Electron microprobe analyses of (a) olivines, (b) orthopyroxenes, (c) clinopyroxenes, (d) garnets, (e) plagioclases and (f) spinels

	I87-1	SC-2	AS-1	TL-1	N-3	W-2	LP-2	LL-1	I87-4	233-1
SiO ₂	35.42	36.69	37.27	36.76	37.09	38.20	37.66	36.56	32.81	34.80
TiO ₂	0.03	0.01	0.00	0.01	0.03	0.00	0.00	0.03	0.00	0.00
Al ₂ O ₃	0.09	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00
FeO	40.85	30.06	28.82	30.47	32.89	25.92	29.99	38.62	56.56	45.40
MnO	0.27	0.02	0.15	0.15	0.12	0.14	0.20	0.29	0.89	0.58
MgO	24.34	33.72	34.46	33.35	30.02	36.92	33.66	26.43	10.94	19.74
CaO	0.05	0.00	0.00	0.02	0.02	0.03	0.03	0.01	0.03	0.00
Na ₂ O	0.00	0.02	0.00	0.02	0.00	0.00	0.00	0.02	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	101.05	100.69	100.71	100.80	100.17	101.20	101.54	101.96	101.22	100.52
Si	1.001	0.980	0.991	0.983	1.016	0.996	0.999	1.010	1.012	1.017
Al	0.002	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000
Fe ³⁺	0.000	0.040	0.018	0.033	0.000	0.010	0.002	0.000	0.000	0.000
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Mg	1.025	1.340	1.365	1.330	1.226	1.434	1.331	1.089	0.503	0.859
Fe ²⁺	0.966	0.632	0.623	0.649	0.754	0.555	0.664	0.893	1.460	1.109
Mn	0.006	0.005	0.003	0.004	0.003	0.003	0.004	0.007	0.023	0.014
Ca	0.002	0.000	0.000	0.000	0.001	0.001	0.001	0.000	0.001	0.000
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fo	51.3	66.5	67.9	65.9	61.8	71.6	66.5	54.7	25.3	43.3

Table 2b.

	I87-1	SC-2	AS-1	TL-1	N-3	W-2	LP-2	LL-1	I87-4	233-1
SiO ₂	51.83	52.41	53.02	53.78	54.22	53.08	53.49	54.61	51.40	52.34
TiO ₂	0.10	0.09	0.04	0.06	0.05	0.02	0.07	0.06	0.06	0.07
Al ₂ O ₃	2.02	3.51	3.28	2.61	2.70	3.12	2.74	1.39	0.47	1.63
FeO	23.58	16.96	17.44	18.35	18.77	15.52	17.54	21.65	32.49	26.18
MnO	0.25	0.14	0.13	0.20	0.15	0.12	0.20	0.27	0.73	0.51
MgO	21.75	24.96	26.09	24.96	24.71	26.83	25.60	23.04	14.70	19.98
CaO	0.63	0.50	0.34	0.39	0.45	0.35	0.36	0.04	0.76	0.34
Na ₂ O	0.01	0.00	0.01	0.01	0.02	0.01	0.01	0.02	0.02	0.00
K ₂ O	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.17	98.58	100.35	100.36	101.09	99.05	100.02	101.44	100.64	101.06
Si	1.932	1.930	1.912	1.954	1.961	1.926	1.942	1.996	2.005	1.960
Al	0.068	0.070	0.088	0.046	0.039	0.074	0.058	0.004	0.000	0.040
Al	0.021	0.082	0.052	0.066	0.076	0.059	0.059	0.056	0.022	0.032
Ti	0.003	0.002	0.001	0.002	0.001	0.000	0.002	0.002	0.002	0.002
Fe ³⁺	0.042	0.000	0.034	0.000	0.000	0.015	0.000	0.000	0.000	0.003
Mg	1.208	1.370	1.403	1.352	1.332	1.451	1.385	1.255	0.855	1.115
Fe ²⁺	0.693	0.522	0.492	0.558	0.568	0.456	0.532	0.662	1.060	0.817
Mn	0.008	0.004	0.004	0.006	0.005	0.004	0.006	0.008	0.024	0.016
Ca	0.025	0.020	0.013	0.015	0.017	0.014	0.014	0.016	0.032	0.014
Na	0.001	0.000	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Wo	1.3	1.0	0.7	0.8	0.9	0.7	0.7	0.8	1.6	0.7
En	62.7	71.6	73.5	70.2	69.5	75.5	71.7	66.3	43.9	57.3
Fs	36.0	27.3	25.8	29.0	29.6	23.7	27.5	32.9	54.4	42.0

Table 2c.

	I87-1	SC-2	AS-1	TL-1	N-3	W-2	LP-2	LL-1	I87-4	233-1
SiO ₂	51.58	50.33	51.90	51.63	53.39	51.48	51.28	52.90	52.32	52.87
TiO ₂	0.37	0.72	0.35	0.26	0.11	0.26	0.66	0.25	0.17	0.16
Al ₂ O ₃	3.79	5.32	3.54	4.90	3.86	4.22	4.35	3.77	2.26	1.96
FeO	8.88	6.86	6.98	6.29	7.08	4.81	5.96	7.18	14.42	8.81
MnO	0.18	0.01	0.06	0.09	0.10	0.06	0.09	0.18	0.36	0.21
MgO	12.58	13.32	14.23	13.75	13.54	13.35	13.20	13.58	9.68	13.04
CaO	20.72	22.02	21.08	21.77	22.11	23.21	22.50	22.43	20.90	23.15
Na ₂ O	1.17	0.98	1.09	1.12	1.13	0.88	0.96	0.78	1.24	0.65
K ₂ O	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.27	99.56	99.23	99.83	101.32	98.27	98.99	101.07	101.36	100.87
Si	1.926	1.860	1.920	1.895	1.940	1.921	1.906	1.933	1.961	1.952
Al	0.074	0.140	0.080	0.105	0.060	0.079	0.094	0.067	0.039	0.048
Al	0.093	0.092	0.074	0.107	0.105	0.106	0.097	0.095	0.061	0.037
Ti	0.010	0.020	0.010	0.007	0.003	0.007	0.016	0.007	0.005	0.004
Fe ³⁺	0.047	0.078	0.064	0.068	0.031	0.023	0.030	0.015	0.058	0.049
Mg	0.700	0.734	0.785	0.752	0.733	0.743	0.731	0.739	0.541	0.718
Fe ²⁺	0.231	0.134	0.152	0.125	0.184	0.127	0.155	0.204	0.394	0.223
Mn	0.006	0.000	0.002	0.003	0.003	0.002	0.003	0.006	0.012	0.007
Ca	0.829	0.872	0.835	0.856	0.861	0.928	0.896	0.878	0.840	0.916
Na	0.085	0.070	0.078	0.080	0.078	0.064	0.069	0.055	0.090	0.047
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Wo	47.1	50.1	47.2	49.4	48.4	51.6	50.2	48.2	47.2	49.3
En	39.8	42.2	44.2	43.4	41.2	41.3	41.0	40.6	30.3	38.7
Fs	13.1	7.7	8.6	7.2	10.3	7.1	8.7	11.2	22.5	12.0

most restrictive reported in the literature, provides an upper *P* limit for the stability of forsterite + anorthite. Due to Ca – Mg and MgSi – AlAl exchange however, the run product pyroxenes were not end-member diopside or enstatite. CaTs and enstatite in the equilibrium clinopyroxene were

determined using X-ray diffraction (XRD) data (Herzberg and Chapman 1976). For the equilibrium orthopyroxene, estimation of the diopside component (Lindsley and Dixon 1976) and the assumption that the mole fraction of Al in the M1 site of the orthopyroxene ($X_{M1,opx}^{Al}$) is the same as

Table 2d.

	I87-1	SC-2	AS-1 [°]	AS-1 [°]	TL-1	N-3	W-2	LP-2	LL-1	I87-4	233-1
SiO ₂	39.36	39.97	39.68	40.20	39.31	40.52	39.90	39.91	39.99	37.36	39.51
TiO ₂	0.05	0.03	0.01	0.02	0.03	0.05	0.04	0.04	0.03	0.00	0.00
Al ₂ O ₃	21.45	21.97	22.19	21.78	20.57	22.27	21.98	21.35	21.88	21.07	21.98
FeO	24.75	21.05	20.26	20.65	21.20	21.90	18.75	20.77	24.25	27.66	25.91
MnO	0.74	0.52	0.46	0.47	0.49	0.53	0.44	0.65	0.90	2.06	1.13
MgO	7.28	10.82	10.70	10.81	10.46	10.02	11.39	9.85	7.86	3.11	5.92
CaO	6.38	6.09	6.38	6.19	6.34	6.18	6.51	6.31	6.74	6.95	7.01
Total	100.01	100.45	99.68	100.12	98.40	101.47	99.01	98.88	101.66	98.22	101.46
Si	3.037	3.000	2.995	3.026	3.019	3.029	3.014	3.055	3.024	3.014	3.028
Al	0.000	0.000	0.005	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Al	1.951	1.943	1.970	1.932	1.862	1.959	1.958	1.926	1.950	2.004	1.985
Ti	0.003	0.002	0.001	0.001	0.002	0.003	0.002	0.002	0.002	0.000	0.000
Mg	0.838	1.210	1.204	1.300	1.197	1.115	1.283	1.124	0.887	0.374	0.667
Fe	1.595	1.321	1.275	0.030	1.362	1.368	1.185	1.330	1.534	1.866	1.661
Mn	0.049	0.049	0.030	1.212	0.032	0.033	0.028	0.042	0.058	0.140	0.073
Ca	0.527	0.527	0.516	0.499	0.521	0.494	0.527	0.517	0.546	0.601	0.576
Pp	27.8	39.6	39.8	39.8	38.5	37.0	42.4	37.3	29.3	12.5	22.4
Alm	53.0	43.2	42.1	42.7	43.8	45.4	39.2	44.1	50.7	62.6	55.8
Sp	1.6	1.1	1.0	1.0	1.0	1.1	0.9	1.4	1.9	4.7	2.4
Gr	17.5	16.0	17.1	16.4	16.7	16.4	17.4	17.2	18.0	20.2	19.3

° Coronal

° Equant, not associated with coronas

Table 2e.

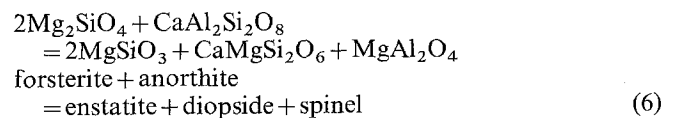
	I87-1	SC-2	AS-1	TL-1	N-3	W-2	LP-2	LL-1	I87-4	233-1
SiO ₂ *	60.99	57.25	59.64	57.50	59.21	55.40	56.97	56.26	59.31	55.50
Al ₂ O ₃ *	24.50	29.08	27.90	25.22	26.16	28.07	26.96	26.59	24.82	27.06
CaO	5.98	10.63	9.11	8.14	8.00	9.87	8.71	8.93	7.28	9.93
Na ₂ O	7.82	5.78	6.67	7.03	6.84	6.01	6.45	6.12	7.54	6.01
K ₂ O	0.72	0.21	0.35	0.48	0.33	0.21	0.48	0.26	0.35	0.16
Total	100.01	102.95	103.67	98.36	100.54	99.58	99.56	98.16	99.30	98.65
Si	2.713	2.502	2.578	2.609	2.633	2.495	2.562	2.572	2.662	2.525
Al	1.285	1.482	1.422	1.349	1.371	1.490	1.429	1.433	1.313	1.451
Ca	0.285	0.498	0.422	0.396	0.381	0.476	0.420	0.437	0.350	0.484
Na	0.674	0.499	0.559	0.619	0.590	0.525	0.562	0.543	0.656	0.530
K	0.041	0.012	0.019	0.027	0.019	0.012	0.027	0.015	0.020	0.009
O	7.996	7.980	8.000	7.960	8.008	7.970	7.982	8.010	7.982	7.980
An	28.5	49.4	42.2	38.0	38.5	47.0	41.6	43.9	34.1	47.3
Ab	67.4	49.4	55.9	59.4	59.6	51.8	55.7	54.6	63.9	51.8
Or	4.1	1.2	1.9	2.6	1.9	1.2	2.7	1.5	1.9	0.9

* Si and Al were determined by fixed stoichiometry. By this scheme approximately 85 weight percent of a plagioclase is calculated from an analysis of 15 weight percent of the total. As a result, oxide totals of 95 to 105 are roughly equivalent to totals of 99 to 101 if Al and Si had been measured

that in the clinopyroxene ($X_{M1, cpx}^{Al}$) permit estimation of the composition³.

3 This assumption, used by Herzberg and Chapman, may in fact overestimate $X_{M1, opx}^{Al}$. Data from naturally occurring pyroxene pairs in granulites and lherzolites (Wood 1975; Bohlen and Essene 1978a; Carswell et al. 1979; Brown et al. 1980; this report) seem to indicate that $X_{M1, opx}^{Al}$ is on the order of 50% larger than $X_{M1, cpx}^{Al}$. We assume, nevertheless, that $X_{M1, opx}^{Al} = X_{M1, cpx}^{Al}$ since the errors introduced in the resulting enstatite activities are small.

Knowing the reduced activities (Wood and Banno 1973) of enstatite and diopside in the experimental assemblage, one can calculate a maximum Gibbs free energy for the end-member equilibrium:



using the relation:

Table 2f.

	I87-1	SC-2 ^p	SC-2 ^a	AS-1	TL-1	N-3	W-2 ^p	W-2 ^a	LP-2	LL-1	I87-4	233-1
TiO ₂	0.02	0.00	0.03	0.02	0.02	0.00	0.02	0.02	0.03	0.00	0.00	0.00
Al ₂ O ₃	60.36	64.88	65.03	63.84	64.93	62.47	65.92	64.37	64.76	61.53	56.22	59.48
Cr ₂ O ₃	0.01	0.02	0.16	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
V ₂ O ₃	0.25	0.03	0.02	—	—	—	0.00	0.00	—	—	—	—
FeO	30.12	23.81	25.08	23.60	21.47	22.72	20.26	21.72	20.65	26.04	39.00	33.74
MnO	—*	—	—	—	—	0.00	—	—	—	0.08	0.32	0.20
MgO	9.11	13.22	12.16	14.26	14.84	12.34	14.88	14.32	15.11	10.74	3.71	6.40
ZnO	1.11	0.33	0.32	0.19	0.49	0.40	0.24	0.46	0.40	0.79	1.31	1.19
NiO	0.00	0.29	0.20	—	—	0.16	0.13	0.09	—	0.22	0.05	0.12
Total	100.98	102.58	103.00	101.91	101.75	98.09	101.45	100.98	100.95	99.40	100.61	101.31
Ti	0.000	0.000	0.001	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Al	1.923	1.967	1.977	1.949	1.962	1.982	1.990	1.965	1.967	1.958	1.881	1.929
Cr	0.000	0.000	0.004	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ³⁺	0.070	0.037	0.017	0.051	0.042	0.018	0.012	0.035	0.029	0.040	0.119	0.071
V	0.006	0.000	0.000	—	—	—	0.000	0.000	—	—	—	—
Fe ²⁺	0.611	0.475	0.524	0.521	0.418	0.493	0.422	0.436	0.416	0.548	0.807	0.706
Mn	—	—	—	—	—	0.000	—	—	—	0.002	0.008	0.005
Mg	0.367	0.507	0.467	0.470	0.567	0.495	0.568	0.553	0.580	0.432	0.157	0.262
Zn	0.022	0.006	0.006	0.007	0.009	0.008	0.004	0.009	0.008	0.016	0.027	0.024
Ni	0.000	0.006	0.004	—	—	0.004	0.003	0.002	—	0.005	0.001	0.003
Hc	57.6	46.2	51.7	49.6	39.9	48.6	41.8	41.9	40.0	53.0	75.6	67.6
Sp	36.7	51.3	46.8	47.1	57.0	49.7	57.1	55.4	57.8	43.4	15.6	26.4
Gh	2.2	0.6	0.6	0.7	0.9	0.8	0.4	0.9	0.8	1.6	2.7	2.4
Mt	3.5	1.9	0.8	2.6	2.1	0.9	0.6	1.8	1.4	2.0	6.0	3.6

* no analytical data were obtained

^a armoured by hydrous phases

^p minute inclusions in plagioclase

$$\Delta G_T^p = -RT \ln(K^0\gamma(6)/K\gamma(3))$$

where $K^0\gamma(6)$ is the equilibrium constant for end-member equilibrium (6) ($K^0\gamma(6)=1$) and $K\gamma(3)$ is the equilibrium constant for the run product assemblage. $\Delta G_{1,200^\circ\text{C}}^{8,9\text{kb}}(6)$ is at most 9.66 kJ/mol indicating that equilibrium (6) is metastable with respect to reaction (3). The location of equilibrium (6) in P - T space has been calculated by solving:

$$d(\Delta G) = \Delta V dP - \Delta S dT = 0$$

where for each phase of interest:

$$dV = -\int \beta V dP + \int \alpha V dT \quad \text{and} \quad dS = -\int \alpha dP.$$

We have done this using the computer program EQUILI (Wall and Essene unpubl.) and the results are shown in Fig. 2⁴. Note that since Fig. 2 is based on a half-reversal, it represents the high P limit for equilibrium (6).

Reaction (5) has been reversed at 900° C and 15.0 ± 0.5 kb and 1,000° C and 16.0 ± 0.5 kb (Jenkins and Newton 1979). As in the previous experiments, the run product pyroxenes undoubtedly showed some degree of mutual solubility and small amounts of Tschermak's com-

4 S_{298}^0 and S_T^0 for the phases were taken from Robie et al. (1978) with the exception of orthoenstatite, diopside and forsterite which were taken from Krupka et al. (1979), Krupka et al. (1980) and Robie (unpubl.), respectively. Values for V_{298}^0 were taken from Robie et al. (1978). Thermal expansion and compressibility data were obtained from Skinner (1966) and Birch (1966). Where data were unavailable, α and β were approximated using values for structurally similar phases. In all cases, it was assumed that $\alpha_T^p = \alpha_T^0$ and $\beta_T^p = \beta_{298}^p$.

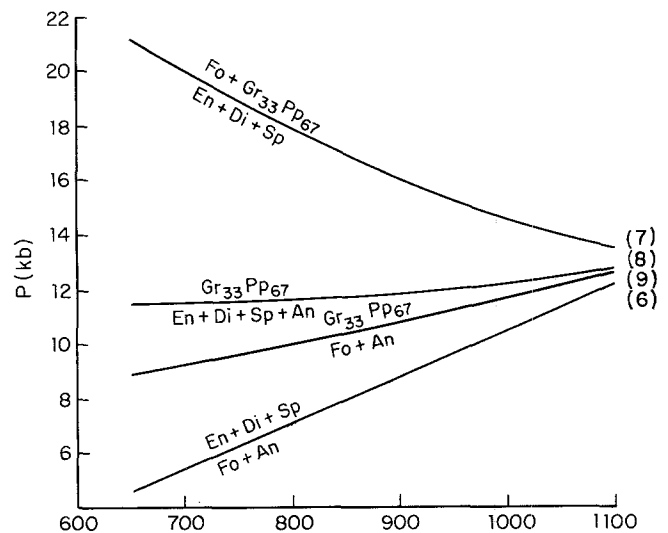
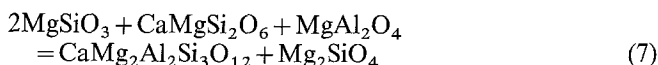


Fig. 2. Reactions (6), (7), (8) and (9) in the CMAS system. Equilibria located using experimental data and the Fortran program EQUILI (see text for explanation)

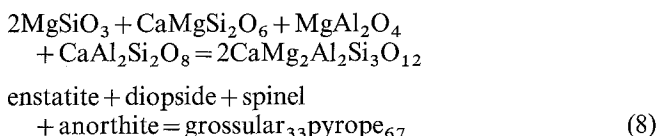
ponent. As in the above calculation, equilibrium compositions have been estimated using the diopside-enstatite solvus (Lindsley and Dixon 1976), data on the solubility of CaTs in diopside (Herzberg 1978a) and the assumption that $X_{M1, \text{cpx}}^{\text{Al}} = X_{M1, \text{opx}}^{\text{Al}}$. Garnet composition determined using XRD data was Gr₁₅Pp₈₅ (Jenkins and Newton 1979). For each reversal one can calculate a Gibbs free energy for the end-member analog of reaction (5):



enstatite + diopside + spinel
= grossular₃₃pyrope₆₇ + forsterite

$\Delta G_{900^\circ\text{C}}^{15.0\text{kb}}$ (7) and $\Delta G_{1,000^\circ\text{C}}^{16.0\text{kb}}$ (7) are 0.08 and -0.29 kJ/mol, respectively. Pyroxene activities were calculated as above; garnet activities were taken to be $(a_{\text{gr}})(a_{\text{pp}})^2$. A regular solution model ($W_{\text{gr-pp}} = 4180 - 1.2 T$ (°C)) was used to calculate grossular and pyrope activities (Perkins 1979). Equilibrium (7) has been located and extrapolated in $P-T$ space using EQUILI⁵ such that it agrees as well as possible with calculated ΔG at both 900 and 1,000° C. The slope of the equilibrium shown by EQUILI does not agree particularly well with the slope derived from experiments on reaction (5) in the 900 to 1,000° C range. The disagreement suggests an uncertainty of ± 1 kb in the location of equilibrium (7) at granulite T . It is important to note that on the univariant line the equilibrium constant for equilibrium (7), with the activity of Gr₃₃Pp₆₇ defined as above, is not unity. The equilibrium constant in fact changes along the univariant curve since grossular and pyrope activities are functions of T .

The Gibbs free energy for the reaction:



can now be calculated by summing the Gibbs free energies for equilibria (6) and (7):

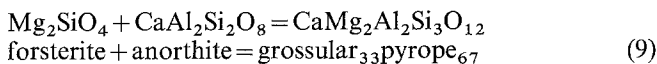
$$\Delta G_T^P(8) = \Delta G_T^P(6) + 2\Delta G_T^P(7).$$

At constant T , this reduces to:

$$\Delta G_T^P(8) = \int \Delta V(6) dP + 2 \int \Delta V(7) dP.$$

Maximum equilibrium P for reaction (8) are 11.5, 11.9 and 12.8 kb at 700, 900 and 1,100° C, respectively (Fig. 2). Again, since the garnet required to balance this reaction is not end-member in composition, the equilibrium constant (where a_{gn} is defined as $(a_{\text{gr}})(a_{\text{pp}})^2$) changes as a function of T and is nowhere equal to one.

The Gibbs free energy for the reaction:



can also be calculated by summing the free energies of equilibria (6) and (7):

$$\Delta G_T^P(9) = \Delta G_T^P(6) + \Delta G_T^P(7).$$

Solution of the equivalent equation:

$$\Delta G_T^P(9) = \int \Delta V(6) dP + \int \Delta V(7) dP$$

at 700, 900 and 1,100° C gives maximum equilibrium P of 9.2, 10.8 and 12.6 kb, respectively (Fig. 2). Reaction (9) is in fact metastable with respect to reactions (6) and (8).

Values for $\Delta G(\text{rxn})$ derived here are compared with values calculated using Robie et al. (1978) in Table 3. The

⁵ S_{298}^0 and S_T^0 for Gr₃₃Pp₆₇ were obtained by adding an ideal mixing term (15.86 J/mol) and an excess mixing term (Haselton and Westrum 1980) to a linear combination of the end-members grossular and pyrope (Robie et al. 1978). V_{298}^0 for the same phase was calculated from the partial molar volumes of Cressey et al. (1978)

Table 3. Derived and calculated (Robie et al. 1978) Gibbs free energy data

	This paper	Robie et al. (1978)*
$\Delta G_{1,200^\circ\text{C}}^{8.9\text{kb}}$ (6)	≤ 9.66 kJ/mole	438.4 kJ/mole
$\Delta G_{900^\circ\text{C}}^{15.0\text{kb}}$ (7)	0.08	-459.8
$\Delta G_{1,000^\circ\text{C}}^{16.0\text{kb}}$ (7)	-0.29	-453.5
$\Delta G_{900^\circ\text{C}}^{11.9\text{kb}}$ (8)	≤ 0.0	-474.5
$\Delta G_{900^\circ\text{C}}^{10.8\text{kb}}$ (9)	≤ 0.0	6.6

* Ideal and excess (Haselton and Westrum 1980) mixing terms were included in the free energy value for Gr₃₃Pp₆₇.

large discrepancies suggest that there are large errors in the tabulated Gibbs free energies of formation for clinoenstatite, diopside and/or spinel.

Order-Disorder in Spinel

A problem in extrapolating equilibria (6) and (7) to lower T and applying equilibria (6), (7) and (8) to rocks involves the potential for small amounts of inverse character in spinel. Synthetic spinels annealed above 800° C may show a degree of inversion of 0.10 to 0.15 while natural spinels are completely normal (Hafner and Laves 1961; Stoll et al. 1964). A discrepancy may therefore exist between the locations of equilibria (6) and (7) for synthetic phases involving a spinel with some disorder and the locations of the same reactions for more ordered natural phases.

Following the development of Navrotsky and Kleppa (1967), 10% inverse character in spinel ((Mg_{0.9}Al_{0.1})(Al_{1.9}Mg_{0.1})O₄) results in a disordering enthalpy given by:

$$\Delta H = -RT \ln \frac{x^2}{(1-x)(2-x)}$$

where x is the fraction of tetrahedral Al. At 1,200° C, this adds 6.3 kJ/mol to the spinel phase. Thus for real rocks (assuming zero disordering entropy), equilibrium (6) might lie at 17.2 kb, some 3.2 kb above the calculated location (Fig. 2) at 1,200° C. Similarly, at 900° C, equilibrium (7) might, for ordered spinel, lie at 8.2 kb, some 6.9 kb below the calculated location (Fig. 2).

Disorder would have no effect on the location of reaction (9) since the experimental errors introduced by using synthetic spinel would exactly cancel in summing $\Delta G(6)$ and $\Delta G(7)$ at the same T . The calculated location of reaction (8), since it contains spinel, may be at erroneously high P . This possible error would be on the order of 0.2 kb at granulite T and may contribute to the fact that reaction (8) gives Adirondack P which seem 1 to 2 kb high.

Phase Equilibria in Natural Systems

The equilibria in Fig. 2 have been developed for the simple CMAS system. The remaining task is to extend these results to natural systems involving many more than four components.

Each of the ten samples used in this study contains univariant assemblages corresponding to reactions (8) and (9).

Equilibration T for each assemblage was assumed to be the same as that of nearby more felsic rocks as determined by feldspar and Fe–Ti oxide thermometry (Bohlen et al. 1980b). Small corrections were made for the regional T gradient in the Adirondacks which has been documented by feldspar and Fe–Ti oxide thermometry and supported by other thermometers (Brown et al. 1978; Valley and O'Neil 1981) and T sensitive phase equilibria (Bohlen and Essene 1978b; Valley and Essene 1980a, b). At a given T , the following relation will permit calculation of equilibrium P for an assemblage of known reduced activities:

$$\int \Delta V dP = -RT \ln (K\gamma/K^0\gamma)$$

where $K\gamma$ is the equilibrium constant for the naturally occurring assemblage and $K^0\gamma$ is the equilibrium constant for an end-member reaction. This calculation is critically dependent on activity-composition relations for phases involved in the reactions.

Garnet activities were calculated using Ganguly and Kennedy's (1974) formulation for multicomponent garnets. Mixing parameters were assumed to be T dependent:

$$W_{gr-pp} \text{ (kcal)} = 4,180 - 1.2 T \text{ (}^\circ\text{C)},$$

$$W_{pp-al} = 3,480 - 1.2 T, \quad W_{al-gr} = 1,050 - 1.25 T$$

with uncertainties of approximately ± 0.5 kcal (Perkins 1979). The spessartine component in garnets from Adirondack metagabbros is typically less than 5 mol%. It was assumed to be zero in calculating $\text{Gr}_{33}\text{Pp}_{67}$ activities introducing only very small errors in the results. Forsterite activities were obtained from the data of Engi (1980). Estatite and diopside activities were calculated using interpolations of Saxena's (1973) and Froese and Gordon's (1974) activity coefficients after determining cation distributions in each case as proposed by Wood and Banno (1973). The data of Saxena (1973) differ from those of Sack (1980b) in that they predict a positive rather than a negative deviation from ideality for orthoferrosilite-enstatite solutions at granulite T . For the orthopyroxene compositions considered here, the discrepancy in calculated enstatite activities is approximately 10%. For reaction (8), P calculated using Saxena's activities agrees better with other Adirondack barometers than P calculated using Sack's activities. Anorthite activities were obtained from the data of Seil and Blencoe (1979), provided by Blencoe (written comm. 1980). These data are similar to those of Orville (1972), but provide for a small T dependence. Finally, spinel (MgAl_2O_4) activities were calculated using an ideal two-site model ($a_{sp} = (X_{Mg})(X_{Al})^2$). The similar ionic radii of Fe^{2+} and Mg^{2+} (Shannon 1976) and the fact that MgAl_2O_4 and FeAl_2O_4 show continuous solid solution suggest that hercynite-spinel solutions do not show a major deviation from ideality and an ideal ionic model may be a reasonable choice.

Table 4 shows calculated P for each of ten Adirondack rocks. The shifts in reactions (8) and (9), the lowest P garnet-forming reactions, were determined for Adirondack compositions. The regional distributions of the results are shown in Figs. 3 and 4.

Areas of uncertainty associated with these calculations are several. Jenkins and Newton (1979) constrained reaction (5) to ± 0.5 kb while Kushiro and Yoder (1966) provided only a half-reversal of reaction (3). Thus original experimental uncertainty is at least ± 0.5 kb. Errors involved in estimating run product compositions and calculating ΔG for end-member reactions are probably small. Several run

Table 4. Inferred equilibration T (Bohlen et al. 1980b), equilibrium constants and calculated P for 10 Adirondack metagabbros. $K\gamma(8)$ and $K\gamma(9)$ are equilibrium constants for the reactions: $2 \text{En} + \text{Di} + \text{Sp} + \text{An} = 2 \text{Gr}_{33}\text{Pp}_{67}$ and $\text{Fo} + \text{An} = \text{Gr}_{33}\text{Pp}_{67}$, respectively

Sample No.	T ($^\circ\text{C}$)	$K\gamma(8)$	$P(8)$	$K\gamma(9)$	$P(9)$
I87-1	725	0.105	9.9	0.273	7.8
SC-2	650	0.109	9.5	0.236	6.5
AS-1	750	0.103	9.9	0.243	7.7
TL-1	725	0.108	9.9	0.279	7.9
N-3	750	0.086	9.5	0.242	7.7
W-2	700	0.114	9.9	0.269	7.5
LP-2	750	0.069	8.9	0.215	7.3
LL-1	700	0.073	8.8	0.204	6.6
I87-4	725	0.064	8.6	0.142	5.7
233-1	700	0.056	8.1	0.148	5.6

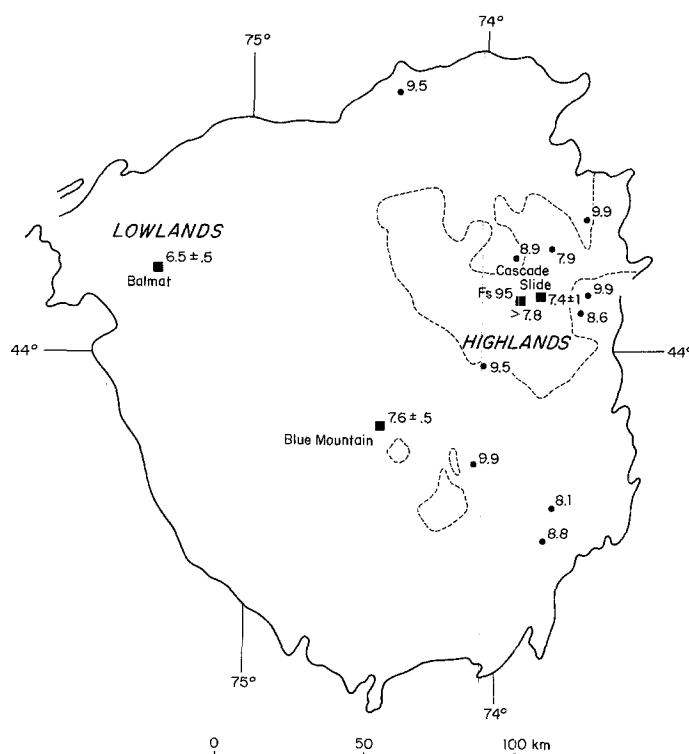


Fig. 3. Outline of Precambrian rocks in the Adirondacks. Areas enclosed by dashed lines represent anorthosite. ● indicates metamorphic P calculated using reaction (8). ■ indicates metamorphic P constraint explained in text. Fs_{95} in Mount Marcy quad, akermanite-monticellite-wollastonite at Cascade Slide, kyanite at Blue Mountain, sphalerite-pyrite-pyrrhotite at Balmat)

product components were measured by the experimenters and the run products were close enough to end-member compositions that the activity models probably give fairly accurate results. Slight disorder in experimental spinel would cause errors of several kb in the locations of equilibria (6) and (7). These errors cancel in calculating the location of reaction (9) and may in part account for the apparent errors of 1 to 2 kb in P calculated from reaction (8) as explained below. Slight disagreement of the EQUILI extrapolation of reaction (7) from $1,000^\circ\text{C}$ with ΔG for the same reaction derived from the reversal at 900°C suggests that there may be an error of perhaps ± 1 kb in the location

of reaction (7) at granulite T . A final source of uncertainty in calculated P is associated with calculating activities of components in complex natural phases. Considering these uncertainties and comparing results in the Adirondacks with independent P constraints, results from reaction (9) probably have an uncertainty of ± 1 kb; results from reaction (8) appear to be systematically in error by as much as 2 kb. These uncertainties apply to assemblages with olivine more Mg-rich than Fo_{50} .

Discussion

Metamorphic P in the Adirondacks can be constrained by several independent equilibria. Sphalerite barometry and calcite-dolomite thermometry in the Balmat-Edwards district (NW Adirondacks) indicate a metamorphic P of 6.5 ± 0.5 kb at $625 \pm 25^\circ \text{C}$ (Brown et al. 1978). The coexistence of akermanite, monticellite and wollastonite at Cascade Slide requires 7.4 ± 1.0 kb and $750 \pm 30^\circ \text{C}$ (Valley and Essene 1980a). The occurrence of ferrosilite 95 ($\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg})$) in the Mount Marcy (Jaffe et al. 1978) and Saranac Lake (Bohlen et al. 1980a) quads gives a P minimum of 7.8 kb for a T of 750°C (Bohlen et al. 1980b; Bohlen and Boettcher 1981). Further, sillimanite occurs throughout the Adirondacks and restricts P to less than 6.0 ± 0.5 kb in the Lowlands and southern Adirondacks and less than 9.0 ± 0.5 kb in the Highlands (Newton 1966; Richardson et al. 1968; Holdaway 1971). A single kyanite occurrence at Blue Mountain (Boone 1978), itself in close proximity to several sillimanite localities, requires 7.6 ± 0.5 kb for a T of $730 \pm 30^\circ \text{C}$ (Bohlen et al. 1980b). In sum, barometry in the Adirondacks seems to give results which are independent of rock type and which vary smoothly across the terrane from about 6 kb in the Lowlands and southern Adirondacks to roughly 8 kb in the Highlands. The close agreement of several different barometers suggests that peak or close to peak regional P has been retained.

P in the Highlands calculated from reaction (9) ranges from 7.3 to 7.9 kb (with one exception) and compares well with the above constraints (Fig. 4). Results in lower T areas, southeast and north of the Highlands, are 6.5 and 6.6 kb with a third assemblage giving a lower value of 5.6 kb. These values are barely consistent with the P maximum of 6.0 ± 0.5 kb inferred from ubiquitous sillimanite. Two assemblages, in samples I87-4 and 233-1, give P which seem anomalously low when compared with those from nearby assemblages. This may be due to the high Fe content of these two rocks (Table 2). One might expect activity models for solid solutions to lose accuracy as they are required to describe increasingly dilute phases.

P calculated from reaction (8) seems to be systematically high by 1 to 2 kb (Fig. 3). Values ranging from 8.9 to 9.9 kb in the Highlands disagree with an estimate of 7.4 ± 1.0 kb at Cascade Slide. Results of 9.5, 8.8 and 8.1 kb in lower T areas are inconsistent with the P maximum of 6.0 ± 0.5 kb imposed by the presence of sillimanite. As with reaction (9), Fe-rich assemblages appear to give low values. The apparent systematic error in P calculated from reaction (8) may be the result of imprecise activity models. The activities of diopside-hedenbergite and hercynite-spinel solid solutions in particular are not well constrained; the presence of both phases in reaction (8) may well contribute to inaccuracy in calculated P . Also, as mentioned above, reaction

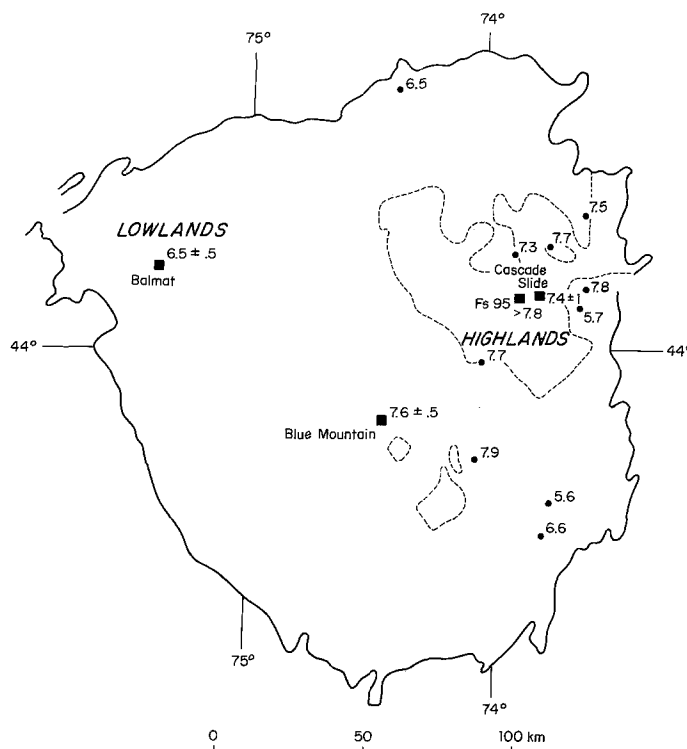


Fig. 4. Outline of Precambrian rocks in the Adirondacks. Symbols as in Fig. 3 except ● which signifies metamorphic P calculated using reaction (9)

(8) may be located here at erroneously high P as a result of disorder in experimental spinel.

Whitney and McLelland (1973) and Whitney (1978) interpret coronal garnet in some Adirondack gabbros as retrograde. They suggest that these rocks retain no peak metamorphic information, only evidence of a retrograde path. Gabbros of this textural type are common in high-grade terranes (for example, Shand 1945; Friedman 1955, 1957; Murthy 1958; Martignole and Schrijver 1971; Griffin and Heier 1973; Whitney and McLelland 1973; Grieve and Gittins 1975; van Lamoen 1979). An isobaric cooling path at moderate to high P is often invoked (for example, Martignole and Schrijver 1971; Griffin and Heier 1973; Whitney and McLelland 1973) to explain rimming relationships.

Whatever their mode of origin, the coronal minerals seem to have equilibrated chemically at or near what are inferred to be peak metamorphic conditions in the Adirondacks (Valley and Bohlen 1979). Inhomogeneity in coronal phases is very slight, on the order of routine microprobe uncertainty. For example, the maximum variation observed in garnets was 9% of the amount of an element present. There is no significant chemical zoning in any of the phases (Gasparik 1980, this report), contrary to what one might expect in rocks which texturally show partially completed reactions. Further, phases which occur in the same rock both in rimming relationships and as equant, apparent equilibrium phases, are essentially identical in composition. The two types of garnet differ by at most 2 or 3 mol% pyrope or almandine (Table 2, sample AS-1), the two types of spinel by at most 5 mol% hercynite (Table 2, samples SC-2 and W-2). McLelland and Whitney (1980a) report compositional variations in plagioclase related to spinel cloud-

ing, but the magnitude of those variations does not appear large (about 5 mol% An). Their effect on the phase equilibria is negligible. In addition, there is no K_D -type evidence for strong T gradients. The garnet-clinopyroxene K_D thermometer (Ellis and Green 1979) and the two-pyroxene thermometer (Wood and Banno 1973), both useful in granulites as semi-quantitative tools, are broadly consistent with inferred peak T (Johnson et al. 1982). P calculated (Table 4) using data from rocks showing coronal textures (AS-1, TL-1, W-2, LP-2, LL-1, 233-1) are in close agreement with those calculated using data from more equant, recrystallized rocks (I87-1, SC-2, N-3, I87-4). In addition, these P appear to be consistent with P information retained by the numerous other Adirondack assemblages discussed above. We suggest that the coronal rocks have undergone a sweeping chemical reequilibration during metamorphism while preserving a reaction texture. What is unquestionably textural disequilibrium may be preserved as a result of low water fugacities in the rocks inhibiting recrystallization and annealing.

Limited data from the literature permit estimates of metamorphic P in other high-grade terranes. Based on garnet reactions in the Fe-bearing system, Wood (1975) proposes P of 10 to 13 kb at 800 to 860° C for three metagabbros with garnet ($\text{Gr}_{20-29}\text{Alm}_{47-56}\text{Pp}_{21-32}\text{Sp}_1$) and plagioclase (An_{29-45}) from South Harris, Scotland. Reaction (9) suggests values of 7.9, 8.2 and 8.7 kb which, in the absence of olivine, represent P minima. Berg (1977; written comm.) reports coexisting garnet ($\text{Gr}_3\text{Alm}_{85}\text{Pp}_{10}\text{Sp}_2$), olivine (Fo_{14}), plagioclase (An_{40}), hypersthene and quartz in a contact aureole of the Nain Complex, Labrador. In this very Fe-rich rock, garnet has been stabilized at quite low P . Berg (written comm.) now estimates that this rock equilibrated at 825° C and 2.5 kb. Bohlen and Boettcher (1981) suggest a P of 3.2 kb based on coexisting orthopyroxene, olivine and quartz. Calculations using reaction (9) give 0.9 kb. This represents reasonably good agreement considering the extreme dilution of forsterite and $\text{Gr}_{33}\text{Pp}_{67}$ and the resulting large P shift (−9.2 kb) of the reaction.

Conclusions

A P – T – X calibration of garnet-forming reactions in silica-undersaturated systems has been developed using the best available experiments and thermodynamic data on reactant and product phases and on common solid solution series. This calibration has been field tested in the Adirondacks and shows acceptable agreement with other barometers indicating that Adirondack metamorphic conditions ranged from approximately 800° C and 7.0 to 8.0 kb in the Highlands to roughly 650° C and 6.5 kb in the margins of the terrane.

Garnet-forming reactions appear to be strongly affected by the solid solutions found in natural systems. Failure to account for these effects and uncritical application of experimental data generally lead to overestimates of P in regionally metamorphosed granulites. Garnet is stable in the Nain Complex at P as low as 2.5 (or 3.2) kb due to extreme dilution of forsterite and $\text{Gr}_{33}\text{Pp}_{67}$ with Fe. In the Adirondacks, solid solutions may have stabilized garnet in silica-undersaturated gabbros at P as low as 6.5 kb. We conclude that garnet in silica-undersaturated gabbros in the Adirondacks may be a stable phase having equilibrated at

peak metamorphic P and T . A significantly retrograde origin for this phase is not required.

Appendix I: Sample locations.

I87-1	I87, 3.1 mi S of Elizabethtown exit, Elizabethtown Quad
SC-2	Walkersville Rd, 2.6 mi S of Skerry, Santa Clara Quad
AS-1	South side of Jay Mtn, Au Sable Forks Quad
TL-1	Barton Mine pit, Gore Mtn, Thirteenth Lake Quad
N-3	West side of Pine Hill, Newcomb Quad
W-2	0.5 mi SE of Sugarloaf Mtn, Willsboro Quad
LP-2	South side of Clements Mtn, Lake Placid Quad
LL-1	Rt 418, 0.5 mi E of Thurman Sta, Lake Luzerne Quad
I87-4	I87, 8.3 mi S of Elizabethtown exit, Elizabethtown Quad
233-1	0.7 mi E of Forest Lake, North Creek Quad

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