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Geothermometry and Geobarometry Applied to Early Proterozoic "S-Type" Granitoid Plutons, Wopmay Orogen, Northwest Territories, Canada

D.R.M. Pattison,* D.M. Carmichael, and M.R. St-Onge
Department of Geological Sciences, Queen's University, Kingston, Ontario K7L 3N6, Canada

Abstract. Many of the "S-type" granitoid plutons that comprise Hepburn and Wentzel Batholiths of the Early Proterozoic Wopmay Orogen contain garnet, biotite and rarely cordierite and sillimanite. The garnet, cordierite and sillimanite are interpreted to be relict crystals brought up from the depth of origin of the magmas.

Two methods of geothermobarometry were applied to ten samples from the two batholiths: the garnet-biotite $Fe-Mg$ exchange equilibrium (Ferry and Spear 1978) and the garnet-plagioclase-sillimanite-quartz equilibrium (Ghent 1976). The intersection of the two displaced equilibrium curves on a $P-T$ plot provides an estimate of the $P-T$ conditions of equilibration of the minerals.

$P-T$ estimates in eight of ten samples ranged between 800 and 1,100 °C, and 5 and 10 kbars. These values are in distinct contrast to the typical 650 °C – 3 kb results obtained from the immediately adjacent metapelites, to which the same geothermobarometers were applied.

Numerous theoretical and practical problems are encountered both in the application of the methods and in the interpretation of the results. Two of the important theoretical problems are (1) how the anatexis of certain minerals in the equilibria influences the elemental ratios of other minerals that remain as solids, and (2) how far the elemental ratios of the minerals re-equilibrated below the maximal $P-T$ conditions. The most serious practical problem was chloritization of biotite, which generated spuriously high temperatures on the garnet-biotite geobarometer.

The significant difference in results between the batholith rocks and the metamorphic rocks indicates that the minerals did not re-equilibrate to sub-magmatic conditions. Using the $P-T$ results as a guide to the conditions of origin of the plutons, it appears that several of the plutons were generated at depths between 21 and 29 km, and rose between 8 and 18 km to the level of emplacement.

The core-to-rim zonation of the garnets produces a $P-T$ trend that is towards lower pressure, temperature or both. Such a pattern would be expected in a pluton that rose to the less extreme conditions at higher levels in the crust.

The geothermobarometers successfully distinguish between granitoid rocks that were generated at depth and those that were formed by anatexis of country rocks near the level of emplacement.

Introduction

Syn- to post-tectonic plutonism in the Early Proterozoic Wopmay Orogen has resulted in two large composite batholiths: Hepburn Batholith and Wentzel Batholith (Hoffman et al. 1980). Many of the granitoid plutons that make up the two batholiths contain the aluminum-rich minerals biotite and garnet, and local sillimanite and cordierite. The occurrence of garnet, sillimanite and cordierite and the virtual absence of hornblende are characteristic of "S-type" granites, inferred by Chappell and White (1974) to be formed by anatexis of pelitic metasediments.

Recently, numerous workers have calibrated cationic exchange equilibria and net-transfer equilibria among co-existing minerals as a function of temperature and pressure in an effort to develop practical methods of inferring the $P-T$ conditions of metamorphosed rocks (geothermometry and geobarometry). Examples of these methods are the distribution of Fe and Mg between garnet and biotite, and the net-transfer equilibrium involving grossular and anorthite in the presence of sillimanite and quartz.

Geothermobarometry has been applied most commonly to medium- and high-grade pelitic schists and gneisses, where it has proved to be a useful, if imperfect, tool for deducing $P-T$ conditions (e.g. Ghent et al. 1979; Tracy and Robinson 1976). Of particular interest to the present study are the results obtained by St-Onge (1982) on the pelites immediately adjacent to the two batholiths, which delineate a gentle pressure gradient of 1.7 kbars over 20 km.

Other granitoid batholiths, e.g. the Kosciusko Batholith in Australia (Hine et al. 1978), have been documented to contain garnet, cordierite and sillimanite, although there is no evidence in the literature of geothermobarometry having been attempted on them. Are these phases phenocrysts that crystallized from the melt as a primary phase, xenocrysts that entered the rising magma from the country rocks, or relict crystals brought up from where the magma originated? White and Chappell (1977) consider them to be refractory minerals brought up in the rising diapir as part of a hot crystal "mush". If this is true, will the minerals

* *Present address:* Grant Institute of Geology, West Mains Road, Edinburgh EH9 3JW, Scotland

Reprint requests to: D.R.M. Pattison

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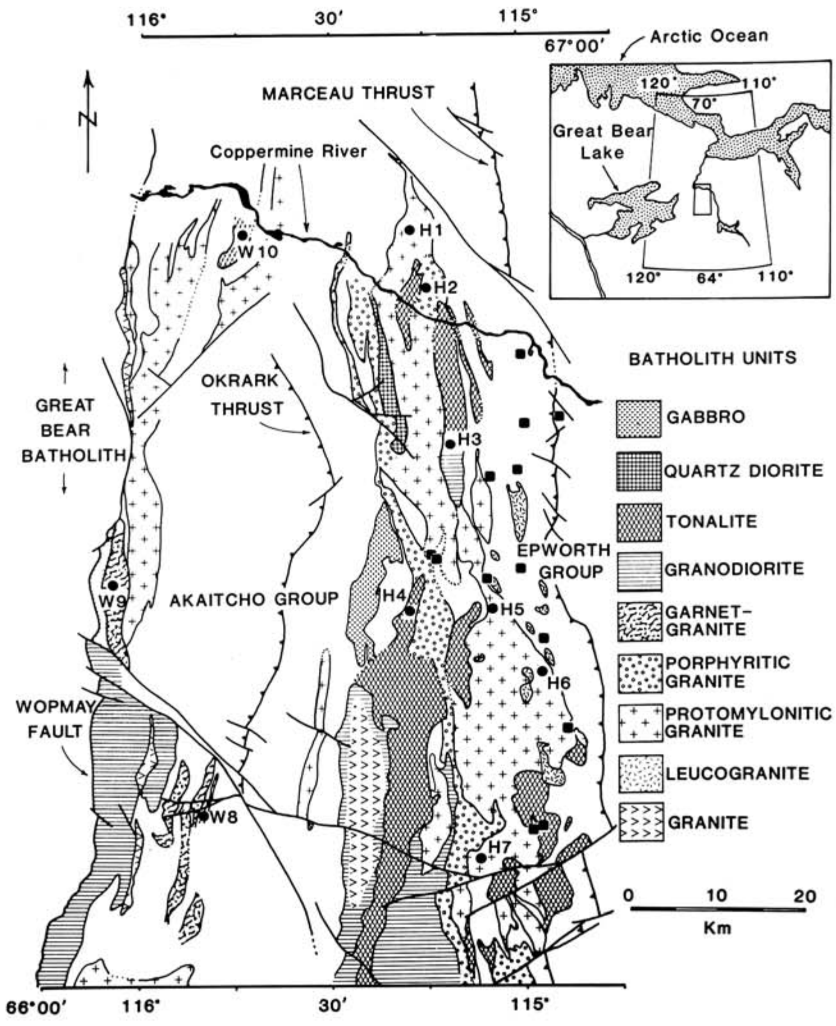

Fig. 1. Plutons and important structural features of the Hepburn Lake map-area (simplified from Fig. 1, Hoffman 1980). The eastern cluster of plutons comprise Hepburn Batholith, and the western cluster comprise Wentzel Batholith. *Solid circles* are sample localities from the batholiths. *H*: Hepburn Batholiths. *W*: Wentzel Batholith. *Solid squares* are sample localities from the metamorphic rocks of the Epworth Group (St-Onge 1982)

have retained their elemental compositions from the depth of origin, or will they have re-equilibrated as the melt rose to the lower temperatures and pressures at the level of emplacement?

White and Chappell (1977) contend that these minerals would re-equilibrate as the diapir rose, and would therefore provide no evidence of the original elemental ratios. By analysing the composition of the minerals in ten samples and applying the methods of geothermobarometry mentioned above, this study provides a critical test for the question of re-equilibration. If the temperatures and pressures are significantly higher than the temperatures and pressures of the adjacent metamorphic rocks, then re-equilibration cannot have been as complete as White and Chappel consider. Ultimately, the results should provide some insight into the origin of the plutons that comprise the two batholiths.

Two methods of geothermobarometry were applied to the Wopmay plutons: the garnet-biotite thermometer (Ferry and Spear 1978) and the grossular-sillimanite-quartz-anorthite barometer (Ghent et al. 1979). Because the garnet-biotite equilibrium has a much steeper $P-T$ slope than the grossular-sillimanite-quartz-anorthite equilibrium, the intersection of the two on a $P-T$ diagram yields an estimate of the pressure and temperature of equilibration of the minerals.

Geological Setting

The two batholiths are part of Wopmay Orogen (McGlynn 1970), a tectonic-metamorphic belt in the northwest of the Canadian Precambrian Shield (Hoffman 1980; Hoffman et al. 1978, 1980; St-Onge 1981, 1982). The syn- to post-tectonic plutons that comprise the two batholiths are situated in the central zone of the orogen. They are areally associated with early thrust faults and large, upright similar folds developed in the early rift metasediments and metavolcanics of the Akaitcho Group (Easton 1980) and the continental rise facies of the Epworth Group (Hoffman 1980). The Hepburn Batholith is situated in the centre of the zone, and the Wentzel Batholith to the west (see Fig. 1).

Over 150 individual plutons, varying greatly in size and shape, have been mapped in the field. The most abundant rock type areally is granite (Streckheisen 1967), although the compositional span ranges from granite to pyroxenite (Hoffman et al. 1980).

Cross-cutting field relationships indicate a general compositional trend whereby the plutons become more basic with time (Hoffman et al. 1980). Although in some rare instances the plutons have some "I-type" characteristics, such as the appearance of hornblende and sphene, the predominant characteristics are "S-type" (ibid.).

A study of the metamorphism of the adjacent sedimentary rocks by St-Onge (1981) has revealed distinctive tectonic differences between the two batholiths. Metamorphic grade increases towards the batholiths from quartz-plagioclase-muscovite-chlorite to quartz-plagioclase-biotite-garnet-sillimanite-cordierite-K-feldspar-granitic melt (inferred). These sequences are roughly concordant with the batholiths, which are therefore inferred to be the principal heat source for the metamorphism.

Reconstructing fault displacements, St-Onge (1981) determined that both of the sequences on either side of Hepburn Batholith dip in towards the batholith, i.e. the isograds are inverted ("hot side up"). Assuming the isograds to be a reflection of the shape of the batholith, this implies that the batholith is either a rootless synform or a broad funnel with feeder conduit at depth. The latter is favoured, as will be seen in detail later, by the presence of garnet in granitoid rocks, which implies a deep (>25 km) origin (Green 1976).

Pressure results from the intersection of the garnet-biotite geothermometer and the grossular-anorthite-quartz-sillimanite geobarometer produced a maximum pressure limit equivalent to a depth of 11–13 km. Assuming the batholith to be at the same hydrostatic pressure as the country rocks it intruded, this depth would represent the exposed floor of the funnel-shaped batholith (St-Onge 1981). Thus, a considerable volume of the body has been eroded away to reach the present exposed surface.

The prograde sequence on the western margin of Wentzel Batholith dips away from the batholith, and is therefore normal ("hot side down"). The present exposure must therefore be the roof of Wentzel Batholith (St-Onge 1981).

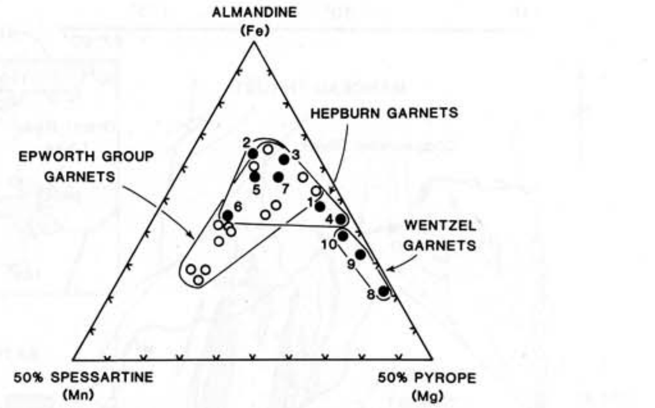

Fig. 2. Garnet compositions plotted on an Fe–Mg–Mn ternary diagram. *Solid circles*: batholith garnets; *open circles*: metamorphic garnets (St-Onge 1982)

Petrography

The texture of the plutonic rocks is hypidiomorphic-granular, medium-grained (with the exception of the fine grained leucogranites and porphyritic granites), and they possess a moderate to strong penetrative foliation defined by the orientation of tabular biotite grains.

The most common mineral assemblage is plagioclase-quartz-biotite-K-feldspar (not present in the tonalite)-garnet, with sillimanite, cordierite and muscovite as rare additions (see Table 1). Zircon is ubiquitous, and apatite is consistently present in minor amounts.

The garnet occurs as subhedral grains between 0.1 mm and 1 cm in diameter, usually between 1 and 3 mm. It is always fractured, and is spacially associated with biotite. Inclusions of quartz and biotite in the garnet are uncommon, and in many samples the garnets are inclusion free.

Origin of the Garnets

The garnets in the granitoid plutons may have one of three origins:

(a) primary crystals, crystallizing directly from the melt,

(b) xenocrysts, brought up from or near the depth of origin as relict source material, or

(c) xenocrysts that entered the melt from the country rocks.

The third possibility may be ruled out immediately. Garnets in the adjacent country rocks of the Epworth and Akaitcho Groups are relatively rare, and could not possibly account for all of the garnet found in the batholiths. The metamorphic garnets are smaller, less fractured and more euhedral than the batholith garnets. Furthermore, the majority of the metamorphic garnets have an Mn-content of 10% or higher, while all but two of the batholith garnets have an Mn-content below 10% (see Fig. 2).

The other two possibilities are more difficult to resolve. Petrographically, there is no reliable method to differentiate between the two. However, experimental data provide some important constraints that favour the xenocryst hypothesis.

Many authors contend that the solubility of Al in silicate melts in excess of the concentrations appropriate for feldspars and micas is very low (Schairer and Bowen 1955, 1956; Winkler 1979). Only in superheated melts could enough excess Al be dissolved to subsequently precipitate an Al-rich mineral such as almandine-garnet or cordierite. Given that the volume of garnet in these samples is as high as 20%, this is unlikely. Sillimanite is even less likely to be a primary magmatic mineral based on this reasoning.

Is it reasonable to assume that garnet can survive thermodynamically in a melt? Several authors (Winkler 1979; White and Chappell 1977) consider that garnet will survive in almost any crustal anatectic situation, provided that it does not disappear by reaction with accompanying phases. This contention is supported by Green's experiments on the high-temperature- and -pressure generation of garnet- and cordierite-bearing granitic liquids from a pelite-composition glass, where he found that garnet persisted to 1,060° C at H_2O and 10 kb, and 1,080° C at 5% H_2O and 10 kb (Green 1976).

Based on these arguments, it is likely that the garnet in the Wopmay plutons are xenocrysts brought up as part of a crystal "mush" from or near the depth of origin.

It is more difficult to present such a strong interpretation for the biotite and plagioclase, for they melt at lower temperatures and are common as a primary phases in plutonic rocks. The state of the biotite and plagioclase in these granitoids will be discussed in a subsequent section in the light of the results obtained from the geothermobarometry.

Sample Strategy and Microprobe Results

Ten samples were analysed on the Queen's University at Kingston ARL-AMX electron microprobe (for details of

Table 1. Modal analyses, representative garnet, biotite and plagioclase oxide analyses, and end-member compositions of the ten samples

Sample number	H1		H2		H3		H4		H5		H6		H7		W8		W9		W10	
Modal analysis	n = 225		n = 543		n = 588		n = 566		n = 758		n = 498		n = 503		n = 499		n = 508		n = 659	
Plagioclase	45%		29%		56%		60%		42%		29%		42%		17%		20%		50%	
K-feldspar	5		31		0		5		27		43		25		16		39		0	
Quartz	21		20		23		15		15		15		17		35		10		20	
Biotite	22		15		16		15		10		1		15		10		10		16	
Garnet	7		5		5		5		5		1		5		20		21		14	
Sillimanite	–		–		–		–		<1		<1		–		2		<1		–	
Rock classification	Tonalite		Porphyritic granite		Tonalite		Tonalite		Granite		Leucogranite		Granite		Granite (gneiss?)		Syenogranite		Tonalite	
Garnet composition	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim
SiO_2	38.01	37.39	36.56	36.98	37.52	37.42	38.47	38.05	37.39	37.41	36.53	35.96	37.82	36.95	39.44	38.71	39.24	38.78	38.13	37.60
TiO_2	0.06	0.00	0.11	0.00	0.10	0.00	0.00	0.34	0.11	0.07	0.00	0.07	0.13	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al_2O_3	22.20	21.86	21.34	21.09	21.00	20.89	23.02	22.18	21.22	21.32	21.35	21.08	21.74	21.06	21.92	22.40	21.51	22.20	21.93	21.60
Fe_2O_3	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
FeO	33.40	33.63	33.43	33.38	34.74	34.87	31.69	32.77	32.97	33.50	31.91	31.34	34.70	34.42	29.62	29.54	28.48	29.22	29.95	30.49
MnO	1.17	1.74	3.50	3.09	1.61	1.82	0.80	0.73	3.27	4.95	7.17	6.73	3.17	3.18	0.64	0.57	0.61	0.56	0.75	0.89
MgO	5.18	4.37	1.80	1.88	4.11	3.31	6.39	6.16	3.30	2.44	2.17	2.22	3.73	3.12	8.86	8.34	7.95	8.09	7.63	6.57
CaO	1.42	1.60	3.05	2.86	1.52	1.28	1.32	1.41	2.38	1.07	0.38	0.33	0.90	0.70	0.75	1.04	1.51	0.96	1.38	1.63
Na_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K_2O	0.04	0.00	0.00	0.00	0.01	0.05	0.02	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.12	0.00	0.00
Total	101.48	100.59	99.79	99.28	100.61	99.64	101.71	101.64	100.67	100.76	99.51	97.73	102.19	99.43	101.23	100.60	99.30	99.93	99.77	98.78
Alm	73.2	74.3	75.8	76.7	75.9	78.7	69.5	70.8	72.9	75.7	73.3	73.7	75.9	78.0	62.8	63.8	63.0	64.3	65.0	67.5
Sp	2.6	3.9	8.0	7.2	3.6	4.2	1.8	1.6	7.3	11.3	16.7	16.0	7.0	7.3	1.4	1.2	1.4	1.2	1.6	2.0
Py	20.2	17.2	7.3	7.7	16.1	13.4	25.0	23.7	13.0	9.8	8.9	9.3	14.5	12.6	33.8	32.1	31.3	31.7	29.5	25.9
Gr	4.0	4.5	8.9	8.4	3.1	3.2	3.7	3.9	6.6	3.1	1.1	1.0	2.5	2.0	1.2	2.9	4.3	2.7	3.8	4.6
And	0.0	0.0	0.0	0.0	1.2	0.5	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.9	0.0	0.0	0.0	0.0	0.0
Plagioclase and Biotite	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.	Plag.	Biot.
SiO_2	60.18	35.33	58.74	33.61	60.59	35.42	57.13	34.91	60.85	33.57	64.90	32.81	59.86	35.75	60.23	35.80	60.24	35.11	59.86	35.24
TiO_2	0.00	3.51	0.11	3.32	0.02	2.34	0.15	4.45	0.00	1.58	0.00	1.83	0.00	2.85	0.00	5.20	0.00	4.21	0.00	5.10
Al_2O_3	25.53	18.21	25.62	17.00	24.90	19.23	27.28	16.38	24.01	18.83	22.44	20.03	25.22	19.85	25.05	15.65	24.73	18.20	26.90	16.60
Fe_2O_3	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
FeO	0.11	19.43	0.13	26.43	0.18	22.56	0.18	20.82	0.00	24.79	0.00	23.45	0.00	20.56	0.00	16.63	0.00	17.70	0.00	16.88
MnO	0.08	0.00	0.00	0.38	0.00	0.21	0.06	0.28	0.05	0.53	0.00	0.36	0.24	0.21	0.00	0.10	0.00	0.00	0.00	0.21
MgO	0.50	10.11	0.00	4.24	0.00	7.01	0.44	9.14	0.00	6.10	0.31	7.62	0.24	7.46	0.00	12.08	0.00	10.59	0.31	10.95
CaO	6.52	0.00	7.98	0.00	6.50	0.00	8.35	0.15	6.05	0.00	2.49	0.11	6.82	0.11	6.79	0.00	6.08	0.00	7.61	0.25
Na_2O	7.63	0.44	6.51	0.00	7.82	0.09	6.01	0.62	7.75	0.00	9.89	0.50	7.36	0.19	7.20	0.43	7.18	0.53	6.82	0.28
K_2O	0.47	9.33	0.03	8.84	0.42	9.44	0.14	8.91	0.22	9.37	0.28	7.50	0.12	9.45	0.59	9.38	0.29	8.59	0.39	9.13
F	–	–	–	–	–	0.33	–	0.36	–	–	–	–	–	–	–	0.95	–	–	–	0.68
Anhydrous total	101.02	96.36	99.12	93.82	100.43	96.63	99.74	96.02	98.93	94.77	100.31	94.21	99.86	96.43	99.86	96.22	98.52	94.93	101.89	95.32
An	31.2		40.3		30.7		43.0		29.7		12.0		33.6		33.1		31.3		37.3	
Ab	66.1		59.5		66.9		56.1		68.9		86.3		65.6		63.5		66.9		60.4	
Or	2.7		0.2		2.4		0.9		1.3		1.6		0.7		3.4		1.8		2.3	

operating conditions, correction factors and error estimates, see St-Onge (1982)). These samples were selected to be broadly representative of the large plutons in the batholith (see Fig. 1), and also so that disequilibrium features were minimized, such as the alteration of biotite and garnet to chlorite and plagioclase to sericite. Two samples contained sillimanite and cordierite in addition to garnet, but these were rejected because the extremely chloritized condition of the biotite and cordierite would have produced meaningless analyses.

Modal analyses, representative garnet, biotite and plagioclase oxide analyses, structural formulae and end-member compositions are compiled in Table 1. Modal analyses are taken from point-counted hand specimens.

The end-member components of the batholith garnets are plotted on an Fe−Mg−Mn ternary diagram, in addition to the garnets from the adjacent Epworth Group pelites (St-Onge 1982) (see Fig. 2). There are three distinct populations, corresponding to the Epworth Group pelites and the two batholiths. The metamorphic garnets, as discussed above, generally have a higher spessartine (Mn)-content. The Wentzel garnets have 5–15% more Mg relative to Fe than the Hepburn garnets.

The garnets in the plutons are broadly homogeneous except for a consistent depletion of Mg relative to Fe near the rims (3–4%) (see Table 1). The two exceptions to this trend are sample H6, which shows a sharper decrease (7%) of Mg relative to Fe at its rim (in addition to its significantly higher Mn-content), and sample W8, which has a minor (2%) increase of Mg/Fe at its rim.

The biotite is homogeneous in the samples, and exhibits a similar bimodal variation between the two batholiths. The Wentzel biotites all have higher Mg/Fe ratios than the Hepburn biotites. Fluorine, analysed in four samples (see Table 1), accounts for between 0.33- and 0.95-weight percent.

Plagioclase ranges from An_{49} in the sample H4 tonalite, to An_{10} in the sample H6 leucogranite, generally ranging between An_{30} and An_{40} (Oligoclase). Minor normal zoning is present in most grains, never exceeding 5%, and the normal oscillatory zoning so common in many mesozonal granites is conspicuously absent.

The stoichiometry of the minerals is close to ideal, with the exception of biotite, which shows a cationic-site deficiency of up to 0.43, even with all of the Fe calculated as FeO. In some samples (e.g. H6 and W9), biotite has an abnormally large deficiency of K_2O in its structure which is accompanied by an abnormally low total oxide weight. This effect is almost certainly due to chloritization of the biotite, which lowers both the K_2O content and the total-oxide weight.

Garnet shows minor amounts (0.05 weight-percent) of Na_2O and K_2O, which are ascribed to inaccuracies in the electron microprobe analysis (St-Onge 1982).

Thermodynamics

1. Biotite-Garnet Geothermometer

This study uses the Ferry and Spear (1978) calibration of the Fe−Mg distribution between garnet and biotite.

The exchange reaction that expresses the equilibrium between the two minerals is:

$$\underset{\text{Fe-garnet}}{Fe_3Al_2Si_3O_{12}} + \underset{\text{Mg-biotite}}{KMg_3AlSi_3O_{10}(OH)_2}$$
$$= \underset{\text{Mg-garnet}}{Mg_3Al_2Si_3O_{12}} + \underset{\text{Fe-biotite}}{KFe_3AlSi_3O_{10}(OH)_2}$$

From the experimental calibration of the Fe−Mg distribution between the two minerals over a range of temperatures, Ferry and Spear's resulting polythermal, polybaric equation that expresses the Fe−Mg partitioning is:

$$12{,}454 - 4{,}662\, T(°\mathrm{K}) + 0.057\, \mathrm{P(bars)} + 3\mathrm{RT} \ln K_D = 0$$

where

$$K_D = \frac{(\mathrm{Mg/Fe})\ \text{garnet}}{(\mathrm{Mg/Fe})\ \text{biotite}}.$$

The agreement of this calibration with natural samples makes equation (1) a useful geothermometer without correction at least for $(Ca+Mn)/(Ca+Mn+Fe+Mg)<0.2$ in the garnet and $(Al^{vi}+Ti)/Al^{vi}+Ti+Fe+Mg)<0.15$ in the biotite (Ferry and Spear 1978). This study's garnets all fall within the first limit, but the biotites generally have a high Ti-content that puts the second ratio above 0.15 and as high as 0.22. No correction was made for the high Ti-content.

Ferry and Spear estimate that the geothermometer is accurate to $\pm 50°$ C (based on 1% inaccuracies in analysed mole fractions).

2. Grossular-Anorthite-Sillimanite-Quartz Geobarometer

This study uses the Ghent et al. (1979) calibration of the Ca-equilibrium between garnet and plagioclase in the presence of sillimanite and quartz.

The reaction that expresses the equilibrium is:

$$\underset{\text{grossular}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{alumino-silicate}}{2Al_2SiO_5} + \underset{\text{quartz}}{SiO_2} = \underset{\text{anorthite}}{3CaAl_2Si_2O_8}$$

Because sillimanite is the stable alumino-silicate polymorph in these samples, the resulting expression for the Ca-equilibrium between garnet and plagioclase is:

$$0 = \frac{-2{,}551.4}{T(°\mathrm{K})} + 7.1711 - \frac{0.2842(P(\mathrm{bars})-1)}{T(°\mathrm{K})} + \log K_x - 0.4, \tag{2}$$

where

$$K_x = \frac{(X^{an}_{plag})^3}{(X^{ar}_{gar})^3}, \qquad X = \text{mole fraction.}$$

The value of −0.4 is an empirical value for the log of the activity coefficients (log K) (for an explanation of the determination of this value see Ghent et al. 1979).

Ghent et al. (1979) consider that the geobarometer is accurate to ±1.6 kbar (2 standard deviations based on a pooled error estimate of temperature and mole-fraction determinations).

On a $P-T$ diagram, the two equilibria intersect at a unique point (see Fig. 3). If all the phases reached equilibrium together, then this point defines a $P-T$ estimate of the conditions of equilibration.

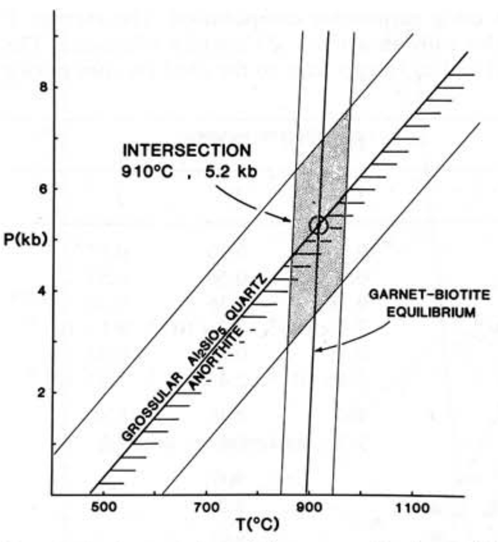

Fig. 3. The intersection of the garnet-biotite equilibrium and grossular-anorthite-sillimanite-quartz equilibrium defines a precise estimate of the $P-T$ conditions of equilibration of the minerals. The error estimate for the garnet-biotite geothermometer ($\pm 50°$ C) is taken from Ferry and Spear (1978); for the grossular-anorthite equilibrium (± 1.6 kb), from Ghent et al. (1979) (see text). If sillimanite is absent from the mineral assemblage, then the sample will plot in the half-space below the grossular-anorthite equilibrium curve, indicated by the hatch-marks. Therefore, the use of the geobarometer is sillimanite-absent assemblages generates a maximum pressure value

Use of Cores and Rims

The results of the geothermobarometry, including cationic ratios, K-values and inferred temperatures and pressures, are summarized in Table 2. The $P-T$ results are plotted in Fig. 4 along with those of the Epworth Group samples (St-Onge 1982).

With the exception of samples H6, H7, and W8, the batholith results show temperatures and pressures in the vicinity of 7–10 kb and 800–1,100° C. Clearly, these results are in distinct contrast to the typical metamorphic values of 3 kb and 650° C, which were derived using the same geothermobarometers. The significance of these inferred temperatures and pressures will be discussed in a subsequent section.

The within-sample precision of $P-T$ determinations from garnet-plagioclase-biotite triads in the same thin section is not as high as the precision obtained in the metamorphic samples, which averaged at $\pm 22°$ C and ± 0.37 kb (St-Onge 1982). The worst sample is H5 at $\pm 120°$ C and ± 1.8 kb. The average within sample precision for the batholith samples is $\pm 45°$ C and ± 0.85 kb, which is within the estimated error limits of the geothermometer/geobarometer intersection (see Fig. 3).

Two important points must be discussed before any interpretation of the results can be made. First, the variation in elemental ratios between the garnet cores and rims generate different $P-T$ values. Second, unique pressures were calculated from the garnet-plagioclase geobarometer for all ten samples, although sillimanite was not present in six of them.

The garnet rims, as mentioned above, have a lower Mg/Fe ratio than the cores, with the exception of sample W8. Because the equilibrium constant for the garnet-biotite equilibrium is equal to (Mg/Fe) garnet/(Mg/Fe) biotite, the rims generate lower apparent temperatures. Consequently, because the garnet plagioclase isopleths have a positive $P-T$ slope, lower apparent pressures are also registered.

In unaltered prograde garnets, it is commonly presumed that the garnet rim would be in equilibrium with co-existing unzoned minerals instead of the core (Ghent 1976; Tracy and Robinson 1976). In the Wopmay samples, the garnets are broadly homogeneous except for a narrow rim consistently depleted in Mg relative to Fe. It is highly unlikely that such a regular rim depletion is due to primary equilibrium crystallisation when the bulk of the garnet is homogeneous and of a different composition. More likely, it is a retrograde re-equilibration feature involving volume diffusion, a mechanism inferred from studies by Tracy and Robinson (1976) and Stephenson (1978), where garnet rims in various metamorphic facies showed consistent, Mg-depleted rims.

Therefore, although the rims are probably the only part of the garnet now in equilibrium with the biotite and plagioclase, they are not valid indicators of the maximal $P-T$ conditions experienced by the plutons. Is it reasonable to assume that the cores are?

Where the amount of garnet relative to biotite is high, then re-equilibration of any portion of the garnet would have a significant effect on the Fe–Mg ratios of the biotite. If, however, the volume of biotite far exceeds the volume of garnet, then the biotite would effectively act as an infinite reservoir, even if there is noticeable re-equilibration of the garnet rims (Tracy and Robinson 1976). The latter case is applicable to all of the Hepburn samples and to the more garnet-rich Wentzel samples if the amount of re-equilibration of the garnet is low, as it appears to be. Therefore, the use of the garnet core compositions with the homogeneous biotite should give minimal errors due to retrograde re-equilibration, and is considered to be the closest approximation to the maximal $P-T$ conditions the plutons experienced.

The average $P-T$ results of the ten samples, using both the garnet cores and rims, are plotted in Fig. 5. Although the garnet rims are not considered to be valid indicators of the maximal $P-T$ conditions of equilibration, they should provide an idea of the extent of retrograde re-equilibration below the conditions at which the garnets were homogeneous throughout.

In Fig. 5, all but sample W8 show a core-to-rim trend of decreasing temperature, pressure or both. Such $P-T$ trend would be expected in a pluton that rose from the maximal $P-T$ conditions of origin to the lower $P-T$ conditions at or near the level of emplacement.

The second important point is that the garnet-plagioclase geobarometer has been applied to all of the samples, although six do not contain sillimanite, one of the essential minerals in the equilibrium. If sillimanite is not present, then the activity of Al_2SiO_5 must be arbitrarily lower than unity. Recalling that sillimanite is on the high pressure side of the equilibrium curve (see Fig. 3), a maximum pressure will be generated, because decreasing the activity of sillimanite displaces the curve to lower pressures.

It is noteworthy that with the exception of sample H2, the apparent pressures are not significantly higher for the sillimanite-free samples than for the sillimanite-bearing ones. This suggests that the activity of Al_2SiO_5 was not

Table 2. Summary of ionic ratios, K-values and inferred temperatures and pressures using garnet-core compositions. The average T and P for each sample using the garnet rims are listed at the bottom. Those samples with an asterisk (*) contain sillimanite. The sillimanite-absent samples give maximum pressures. The activity terms for Ca (a_{Ca}(ga) and a_{Ca}(plag)) refer to the ideal on-sites mixing portion of the RTln K term (ie K_x) – see text

Sample number	H1			H2		H3		
Station	1	2	3	1	2	1	2	3
(Mg/Fe)ga	0.30	0.28	0.29	0.10	0.086	0.20	0.20	0.17
(Mg/Fe)bio	0.91	0.94	0.95	0.30	0.30	0.52	0.56	0.57
K_D(Fe–Mg)	0.33	0.30	0.30	0.34	0.28	0.38	0.36	0.30
(a_{Ca})ga	9.2×10^{-5}	6.0×10^{-5}	9.1×10^{-5}	6.8×10^{-4}	7.1×10^{-4}	7.3×10^{-5}	7.4×10^{-5}	9.1×10^{-5}
(a_{Ca})plag	0.32	0.32	0.33	0.39	0.33	0.31	0.31	0.32
K(Ca)	2.7×10^{-3}	1.8×10^{-3}	2.4×10^{-3}	1.2×10^{-2}	1.9×10^{-2}	2.4×10^{-3}	2.4×10^{-3}	2.7×10^{-3}
T (°C)	880	810	820	910	800	980	920	820
P (kb)	8.0	6.4	7.0	11.2	10.1	9.3	8.5	7.1
Avg. T (core)		840		860			900	
Avg. P (core)		7.1		10.6			8.3	
Avg. T (rim)		780		850			830	
Avg. P (rim)		6.7		10.3			6.9	

Sample number	H4		H5*			H6*		H7	
Station	1	2	1	2	3	1	2	1	2
(Mg/Fe)ga	0.34	0.36	0.17	0.16	0.15	0.14	0.12	0.18	0.18
(Mg/Fe)bio	0.78	0.81	0.44	0.44	0.51	0.56	0.56	0.67	0.65
K_D(Fe–Mg)	0.45	0.44	0.38	0.37	0.29	0.25	0.22	0.27	0.28
(a_{Ca})ga	4.1×10^{-5}	4.0×10^{-5}	2.0×10^{-4}	6.0×10^{-5}	9.3×10^{-5}	2.1×10^{-6}	1.2×10^{-6}	1.1×10^{-5}	1.3×10^{-5}
(a_{Ca})plag	0.39	0.43	0.29	0.29	0.27	0.13	0.11	0.30	0.30
K(Ca)	6.8×10^{-4}	4.8×10^{-4}	7.7×10^{-3}	2.6×10^{-3}	4.6×10^{-3}	9.5×10^{-4}	9.6×10^{-4}	3.7×10^{-4}	4.6×10^{-4}
T (°C)	1,100	1,080	990	960	800	700	650	740	760
P (kb)	8.4	7.7	11.8	9.2	7.8	3.8	3.3	4.2	3.5
Avg. T (core)	1,090			920		680		750	
Avg. P (core)	8.0			9.6		3.5		3.9	
Avg. T (rim)	1,050			870		680		720	
Avg. P (rim)	7.7			7.6		2.0		2.3	

Sample number	W8*			W9*			W10		
Station	1	2	3	1	2	3	1	2	3
(Mg/Fe)ga	0.54	0.56	0.62	0.45	0.50	0.44	0.45	0.45	0.45
(Mg/Fe)bio	1.36	1.55	1.64	1.07	1.07	1.06	1.12	1.14	1.06
K_D(Fe–Mg)	0.39	0.36	0.38	0.42	0.47	0.41	0.40	0.39	0.42
(a_{Ca})ga	8.1×10^{-6}	1.5×10^{-5}	1.6×10^{-5}	6.0×10^{-5}	5.6×10^{-5}	4.1×10^{-5}	5.1×10^{-5}	1.3×10^{-4}	9.3×10^{-5}
(a_{Ca})plag	0.34	0.33	0.32	0.32	0.31	0.30	0.38	0.39	0.39
K(Ca)	2.1×10^{-4}	4.3×10^{-4}	4.9×10^{-4}	1.9×10^{-3}	1.9×10^{-3}	1.5×10^{-3}	1.0×10^{-3}	2.2×10^{-3}	1.6×10^{-3}
T (°C)	970	910	950	1,160	1,150	1,130	1,010	1,000	1,050
P (kb)	4.6	5.3	5.9	10.0	11.2	9.1	8.0	9.4	9.6
Avg. T (core)		940			1,140			1,020	
Avg. P (core)		5.3			10.1			9.0	
Avg. T (rim)		960			1,100			910	
Avg. P (rim)		6.2			10.1			7.7	

very much less than unity in the sillimanite-free samples. It also implies that the anatectic protolith for the granitoid magmas was predominantly pelitic in composition and probably contained sillimanite, although none is found in many of the plutons.

Difficulties in Assuming Equilibrium Between Phases

The maximum temperature registered on the garnet-biotite geothermometer is 1,150° C from sample W9 which, due to its highly chloritized condition, is not considered to have

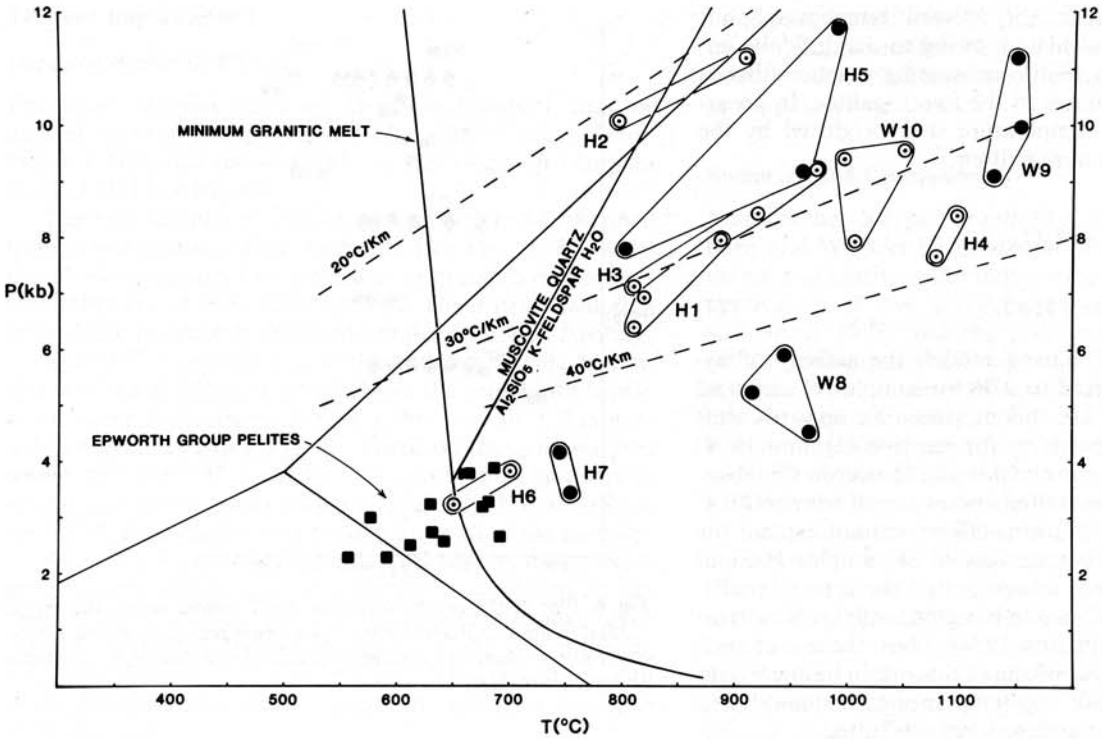

Fig. 4. Graphical representation of $P-T$ results obtained from the ten batholith samples. *Solid circles*: sillimanite present; *open circles*: sillimanite absent; *squares*: Epworth Group Samples. Each square represents the average $P-T$ result obtained from several within-sample analyses (St-Onge 1982). The minimum granitic melt curve and the muscovite-quartz-Al_2SiO_5-K-felspar-H_2O curve are from Winkler (1979), the alumino-silicate triple point from Holdoway (1971), and the 20° C/km and 30° C/km goethermal gradients from Brown and Fyfe (1970)

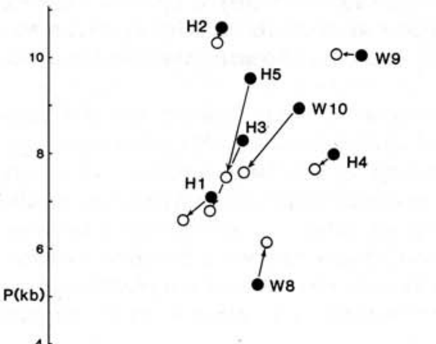

Fig. 5. Garnet core-to-rim $P-T$ trend of the batholith samples (see text for more detailed explanation). *Solid circles*: cores; *open circles*: rims

generated good analyses. The inferred temperatures of the other samples are all below 1,100° C, and are therefore compatible with the restite-hypothesis for the origin of the garnet.

Biotite constitutes a more difficult problem. Although several authors have determined approximate $P-T$ limits for the high temperature disappearance of biotite (e.g. Brown and Fyfe 1970; Robertson and Wyllie 1971), there has been little progress made in specifically characterizing these biotite-consuming reactions. A biotite-breakdown reaction for one bulk composition could be different from a comparable reaction for another bulk composition.

From experiments that have been conducted (Green 1976; Robertson and Wyllie 1971; Brown and Fyfe 1970), there is only general agreement on the upper stability of biotite. At 1.5% H_2O and 2 kb, Robertson and Wyllie placed their "biotite-out" curve from a granodiorite bulk composition at 900° C. Brown and Fyfe estimated the biotite-breakdown for a diorite bulk composition to be at 800° C and 10 kb. Green noted biotite persisting to 980° C at 2% H_2O and 10 kb from an average-pelite bulk composition. If Robertson and Wyllie's limit is extrapolated to higher pressures, the biotite-consuming curve having a positive $\partial P/\partial T$ slope, then a temperature comparable to Green's is obtained.

Paradoxically, samples H4 and W10 register inferred temperatures above 1,000° C, apparently above the upper limit of biotite stability. It is noteworthy that these two samples are both tonalites, whose temperature of formation is presumably the highest of the granitoids (Winkler 1979).

The weight percent of fluorine in these two samples is 0.36-% for sample H4 and 0.68-% for sample W10. It has been documented that fluor-biotite has a higher-temperature stability limit than pure hydroxy-biotite (Munoz and Ludington 1974). Therefore, the presence of fluorine in the hydroxyl site of the biotite will tend to lower the activity of hydroxy-biotite and, consequently, stabilize the mineral to higher temperatures.

A quantitative value for this upwards temperature shift could not be reliably calculated, owing to the difficulty encountered in writing a biotite-consuming reaction that is compatible with the phases in the two tonalites. In an attempt to estimate the temperature shift produced by the fluorine, two reactions were written:

$$\text{Muscovite} + \text{Biotite} = \text{K-feldspar} + \text{Garnet} + H_2O, \quad (3)$$

$$\text{Biotite} + \text{Quartz} = \text{K-feldspar} + \text{Enstatite} + H_2O. \quad (4)$$

Assuming an ideal mixing model, the activity of hydroxy-biotite was lowered to 0.96 for sample H4 and 0.92 for sample W10. At 8 kb, this produced an upwards shift of 5° C and 10° C respectively for reaction (3), and 18° C and 38° C respectively for reaction (4). However, the absolute temperatures of the shifted curves are all below 970° C at 8 kb, and thus this "fluorine-effect" cannot explain the extreme geothermobarometer results of samples H4 and W10. It has been shown, however, that the activity coefficient of fluorine in phlogopite is significantly greater than unity (Munoz and Ludington 1974). Thus, the use of such a non-ideal model for the mixing of fluorine in biotite would produce a higher "shift" up-temperature; a quantitative estimate using this approach was not attempted.

If the biotite were completely dissolved in the melt, then the geothermometer result is obtained from a garnet that co-existed with melt, and biotite that crystallized later at a lower temperature. Such a condition would violate the inherent assumption in the application of the geothermometer that the two minerals were in constant equilibrium as the Fe−Mg ratios varied with temperature. The geothermometer would be valid as long as any biotite remained as a solid phase. After the last grain of biotite melted, the geothermometer would read progressively less accurately with rising temperature, and would be systematically too high or too low. This argument can only be applied in the case where garnet and biotite are the only two ferromagnesian phases present.

The garnet of sample H4 shows a significantly higher Mg/Fe ratio relative to the other Hepburn garnets than does the H4-biotite relative to the other Hepburn biotites. Possibly this effect is due to the process described above.

A similar problem to the biotite occurs with plagioclase when evaluating the garnet-plagioclase geobarometer, for it enters the melt at the onset of partial melting. Because garnet and plagioclase are the only two minerals that contain significant Ca, the garnet-plagioclase equilibrium will be preserved until the last grain of plagioclase melts. If all of the plagioclase is melted, then the activity of anorthite will begin to fall below its value in the solid plagioclase as the temperature rises. If the garnet were to equilibriate with the An-component of the melt at elevated temperatures and fail to re-equilibrate to lower temperatures, then a spuriously low pressure would result, since anorthite is on the low-pressure side of the equilibrium curve (see Fig. 3).

This argument presumes that quartz remains as a solid phase throughout. If the quartz were completely dissolved in the melt, then the activity of SiO_2 would fall below unity, and similar arguments would apply. The overall effect of these conditions on the geobarometer depends upon the relative amounts of each mineral in the source rock and the amount of melting that had occurred.

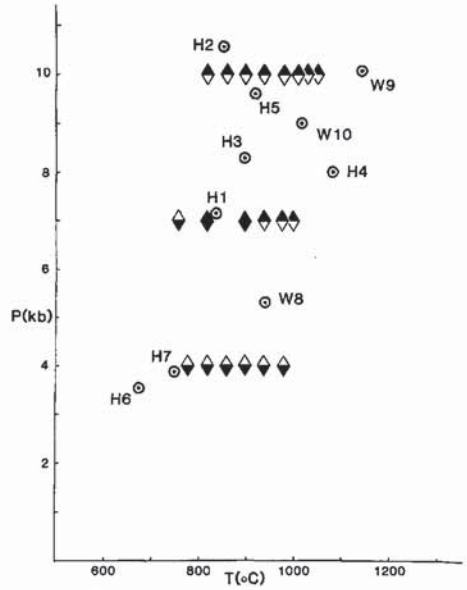

Fig. 6. Plot of the geothermometer $P-T$ results versus the experimental zones of Green (1976). *Solid upper triangles*: garnet stable. *Solid lower triangles*: cordierite stable. *Solid diamonds*: cordierite and garnet stable

Further difficulties arise in determining the extent of re-equilibration of the elemental ratios as the minerals rose to less extreme conditions near the level of emplacement. Clearly, the minerals have not re-equilibrated extensively, for in eight samples the present ratios generate temperatures and pressures well above those of the bordering metamorphic rocks. Had re-equilibration occurred up until the time the plutons were emplaced, then the geothermobarometer results would be the same as those of the metamorphic rocks.

Some re-equilibration is made evident by the consistently Mg-depleted garnet rims, but this effect is considered to be minor owing to the large volumes of biotite present in the rocks (see previous section). However, biotite is susceptible to low-grade alteration to chlorite, which has a significant effect on its Mg/Fe ratios. This effect is manifested in the extreme (1,100° C) results of sample W9, whose biotite is severely chloritized. The other samples contain fresh, unaltered biotite.

Despite the variety of bulk compositions exhibited by the batholith samples, seven of the ten plot in accordance with Green's experiments on the generation of garnet- and cordierite-bearing granitic liquids from a pelite-composition glass (see Fig. 6). He found that a 7 kb garnet and cordierite co-exist up to 1,000° C, while at 10 kb garnet is stable and at 4 kb cordierite is stable.

It is noteworthy that Green's experiments were conducted with a pelite-composition glass that had an MnO content of zero, a component that all of the Wopmay samples contain. The garnet of sample H6 has a spessartine (Mn) component of 16%, nearly twice that of the other samples, a factor that Green suggests might stabilize it to lower temperatures and pressures.

The good correlation between the experimental results and the geothermobarometry suggest that in spite of numerous theoretical and experimental problems (see also St-Onge 1982), the geothermobarometry has generated some reasonable results from the batholith plutons.

Tectonic Implications

Hepburn Batholith Plutons

The seven samples from the Hepburn batholith may be divided into two groups: the >6 kb, $>800°$ C group (samples H1–H5) and the <4.5 kb, $<775°$ C group (samples H6 and H7) (see Fig. 4).

The five samples of the former group are all obtained from major plutons of the batholith (see Fig. 1). The samples show no apparent temperature- or pressure-related pattern reflective of their different bulk chemistry, such as a progressive increase in apparent temperature with basicity.

Fyfe (1973) outlined a zone of maximum magma generation by crustal fusion situated above the muscovite breakdown curve and between the 20° C/km and 30° C/km geothermal gradients (see Fig. 4). Three of the samples plot within this zone. If, however, the geothermal gradient in tectonically active zones ranges as high as 60° C/km (Watson 1978), then such a zone encompasses all of the samples.

Assuming a geobaric gradient of 0.3 kb/km, the geothermobarometer results indicates depths of equilibration between 21 and 29 km. The bordering metamorphic rocks generate pressures corresponding to 11–13 km, implying that these plutons rose between 8 and 18 km from the depth of equilibration, possibly the depth of origin, to the depth of emplacement.

In the lower $P-T$ group, geothermobarometry from sample H6 generates at temperatures of 675° C and pressure of 3.75 kb. The pluton from which the sample was taken is one of several small discordant leucogranite stocks along the eastern margin of the Hepburn Batholith and in the bordering metamorphic pelites. The composition and euhedral form of its garnets is similar to those of many of the pelitic garnets, and its composition is close to the eutectic minimum in the $SiO_2-NaAlSi_3O_8-KAlSi_3O_8$ system (Winkler 1979). This pluton is therefore interpreted to be a late stage, high-level intrusion formed by the anatexis of the adjacent pelites at or near the level of emplacement.

Sample H7 was collected from a fine-grained zone in a large granite pluton, and registered inferred $P-T$ conditions of 4 kb and 750° C. Although its garnet has a similar composition as the other batholith samples, its fine-grained texture suggests a similar high level anatectic origin as sample H6.

Wentzel Batholith Plutons

The Wentzel Batholith, judging by the 10–15% higher Mg content of its garnet and biotite, was derived from a more Mg-rich protolith than the Hepburn Batholith.

Sample W10, a tonalite, apparently equilibrated at 1020° C and 8.8 kb, comparable to the large plutons of the Hepburn batholith.

Sample W8 generates $P-T$ conditions of 940° C and 5.2 kb. In the field this strongly foliated "granite" was intimately associated with high grade migmatites of similar mineralogy. Dip-slip motion on major faults has exposed a deeper level of the crust than in the areas bordering Hepburn Batholith, producing a broader contact zone. The core-to-rim zonation of the garnets is one of increasing temperature and pressure, possibly a faint prograde zonation.

This body is therefore interpreted to be either a very high-grade gneiss near the roof of the Wentzel Batholith, or a partial-melt pluton that rose only a short distance to the level of emplacement. The zoning pattern, although considered unreliable in the other samples, favours the high-grade gneiss hypothesis.

Summary and Conclusions

Many of the "S-type" granitoid plutons that comprise Hepburn and Wentzel Batholiths of Wopmay Orogen contain garnet and biotite, and rarely cordierite and sillimanite. The application of the garnet-biotite geothermometer (Ferry and Spear 1978) and the grossular-sillimanite-quartz-anorthite geobarometer (Ghent et al. 1979) has generated temperatures and pressures of equilibration in eight of ten samples between 800 and 1,100° C and 5 and 10 kb. These results are in distinct contrast to the average values of 650° C and 3 kb obtained from the bordering metamorphic pelites, to which the same geothermobarometers were applied.

Numerous theoretical and practical problems are encountered both in the application of the methods and in the interpretation of the results. The significant difference in the $P-T$ results between the batholith rocks and the metamorphic rocks indicates that the elemental ratios of the minerals did not completely re-equilibrate to the submagmatic conditions at which they were emplaced. Although it is difficult to determine where in the sequence of magma generation, transport, emplacement and possible re-equilibration that the present elemental ratios were finally produced, the $P-T$ results clearly indicate the hotter and deeper conditions where the plutons originated. Indeed, if the elemental ratios did re-equilibrate to lower pressures and temperatures, then these results are minimum values.

If the results can be taken as a semi-quantitative guide to the depth and temperature of origin of the plutons, then it appears that several of the plutons were generated at depths between 21 and 29 km. Given that the exposed metamorphic rocks indicate depths between 11 and 13 km (St-Onge 1982), this implies that the plutons rose between 8 and 18 km to the level of emplacement. The core-to-rim zonation of the garnets produces a $P-T$ trend of decreasing temperature, pressure or both, a pattern that would be expected from the relative upwards motion of a magma body rising to higher levels in the crust.

Moreover, the geothermobarometers have proven to be sensitive in differentiating between plutons of deep crustal origin and those formed near the level of emplacement by local anatexis of adjacent country rocks. The genetic differences, initially suggested by the geological associations and petrologic character of the bodies, are confirmed by the geothermobarometry.

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