Thermodynamic calculation of emplacement pressures for batholithic rocks, California: Implications for the aluminum-in-hornblende barometer

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ABSTRACT

The aluminum-in-hornblende (AH) barometer has been widely used to estimate the emplacement pressure (*P*) of granitic rocks. Application of the barometer has remained controversial because the controls on the aluminum content of hornblende imposed by temperature (*T*), magma composition, and fluid composition remain incompletely understood. In this paper, the *P-T* conditions of equilibration for 19 rocks from the Mesozoic batholiths of California are estimated by simultaneous evaluation of (1) the *P*-sensitive reaction tremolite + phlogopite + 2 anorthite + 2 albite = 2 pargasite + 6 quartz + K-feldspar and (2) previously published amphibole-plagioclase thermometer equilibria. *P* estimates for the batholiths range from 1.2 to 7.4 kbar and are consistent with independent geologic evidence. The highest *P* samples (5.1–7.4 kbar) are located near the San Andreas fault. Furthermore, it appears that regional *P* gradients determined by previous AH barometry studies are valid, but that the absolute values of the *P* estimates will need to be revised somewhat (±1.5 kbar). I suggest that the method of this study provides a viable, thermodynamics-based alternative to conventional AH barometry that requires no a priori assumptions about *T* or bulk composition.

INTRODUCTION

Determination of the emplacement depths of granitic magmas is critical for deciphering the baric and petrotectonic evolution of orogenic belts. Crystallization pressures are commonly estimated using thermobarometric analysis of coexisting metamorphic mineral assemblages in intruded wall rocks. A significant drawback of this approach is that large areas of many batholithic terranes are devoid of metamorphic wall rocks and roof pendants containing mineral assemblages suitable for estimation of pressure-temperature *(P-T)* conditions. Furthermore, relative movement between magma and wall rocks occurs during magma ascent and, therefore, intruded wall rocks can preserve a different baric signature than adjacent plutons (cf. Zen, 1989).

The best way to attack the problem is to estimate equilibration *P* directly from coexisting mineral assemblages in the granitic rocks. In their landmark paper, Hammarstrom and Zen (1986) made a fundamental contribution in this regard by presenting evidence that the total aluminum (Al^T) content of hornblende (Hbl) could be used a barometer for granitic rocks. Their analysis suggested that the Al^T of Hbl coexisting with quartz, plagioclase, K-feldspar, biotite, titanite, Fe-Ti oxides, supercritical fluid, and melt increases linearly with increasing *P*. The relationship between Al^T and *P* has become known as the Al-in-hornblende barometer (abbreviated here as the AH barometer). Subsequent field-based studies (e.g., Hollister et al., 1987) and experimental studies (e.g., Johnson and Rutherford, 1989; Schmidt, 1992) have demonstrated unequivocally that the Al^T of Hbl in granitic systems does increase with increasing *P*. The experimental studies suggest that the change in Al^T is governed largely by tschermak-type substitutions.

The increase in Al^T with increasing *P* has been clearly demonstrated, but application of the AH barometer in natural geologic settings has remained controversial because of several outstanding questions (e.g., Cosca et al., 1991; Anderson and Smith, 1995). (1) Must Hbl equilibrate with both supercritical fluid and melt in order for the AH barometer to work? (2) The Al^T of Hbl is also a function of *T* (cf. Hammarstrom and Zen, 1986; Schmidt, 1992; Holland and Blundy, 1994; Anderson and Smith, 1995). How do variations in *T* affect *P* estimates? (3) What impact do variations in magma bulk composition, fluid composition, mineralogy, and oxygen fugacity have on *P* estimates? (4) What are the *P*-sensitive chemical reactions that control Hbl chemistry in granitic systems?

In this paper, a first-order attempt is made to address some of these questions by examining a *P*-sensitive reaction that shows promise as an igneous barometer and that requires no a priori assumptions about fluid composition, magma composition, or *T*. Equilibration *P* is calculated using the state-of-the-art thermodynamic data and comprehensive nonideal activity model for amphibole end members of Mäder et al. (1994). Pressures are estimated for a representative suite of tonalites, granodiorites, and granites from the Mesozoic batholiths of California.

REACTION

Mäder and Berman (1992) and Mäder et al. (1994) presented thermodynamic data and a nonideal activity model for the amphibole end members tremolite ($Ca_2Mg_5Si_8O_{22}[OH]_2$), pargasite ($NaCa_2[Mg_4Al][Si_6Al_2]O_{22}[OH]_2$), tschermakite ($Ca_2 [Mg_3Al_2] [Si_6Al_2] O_{22} [OH]_2$), and the Fe analogs of these end members. Their analysis placed reasonably tight constraints on (1) the standard-state properties of tremolite and pargasite and (2) the amphibole mixing properties. The standard-state properties of tschermakite and the Fe end members were not as well constrained (Mäder and Berman, 1992), and therefore only tremolite and pargasite were considered in this study. The search for a useful barometric reaction was focused on equilibria involving tremolite, pargasite, phlogopite, quartz, and feldspars. Only solid-solid reactions were considered in order to explicitly remove the activity of water (a_{H_2O}) as a variable. The one reaction that can be written under the above constraints is

$$\text{tremolite} + \text{phlogopite} + 2\text{anorthite} + 2\text{albite} = 2\text{pargasite} + 6\text{quartz} + \text{K-feldspar}. \quad (1)$$

Mäder and Berman (1992) pointed out that this reaction (1) has a fairly small d*P*/d*T* slope (~13.4 bars/degree) and is therefore usable for barometry

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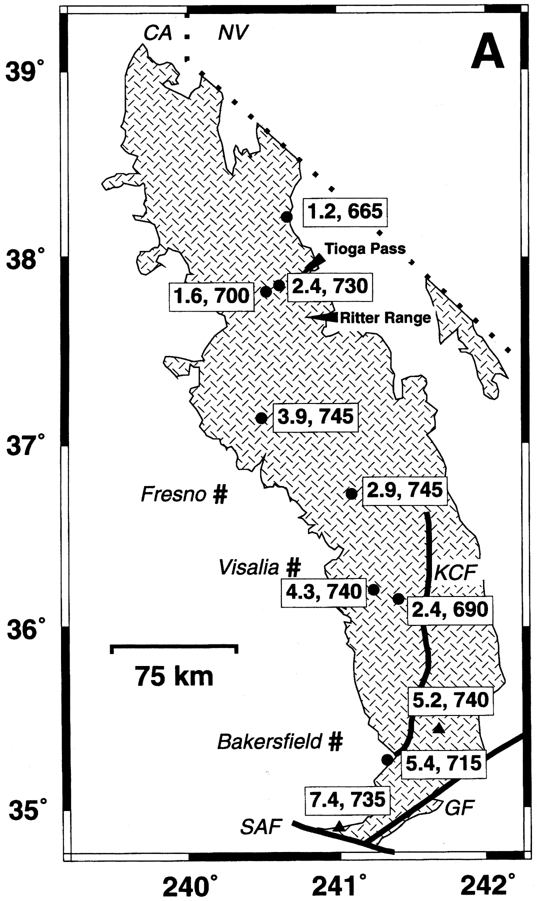

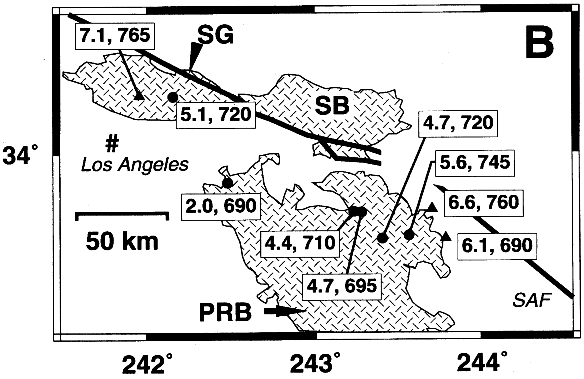

Figure 1. Sample locations and pressure (kbar), temperature (°C) estimates. Triangles show samples that contain possible magmatic epidote. A: Sierra Nevada batholith. GF = Garlock fault; KCF = Kern Canyon fault; SAF = San Andreas fault. B: Peninsular Ranges batholith (PRB), San Gabriel mountains (SG), and San Bernardino mountains (SB). SAF = San Andreas fault.

and (2) gives P estimates that are in good agreement with those based on other equilibria for felsic granulites of the Furua Complex, Tanzania. A probable link between reaction 1 and the AH barometer is that reaction 1 has the Al-free amphibole end member tremolite on the low-P (left) side of the reaction and the aluminous end member pargasite, rich in both octahedral and tetrahedral Al, on the high-P side. It is important to emphasize, however, that reaction 1 is not necessarily "the" single Al-in-hornblende barometer reaction. It is simply *one* reaction that shows promise as a barometer for rocks containing Hbl, biotite, two feldspars, and quartz—a fact pointed out by Mäder and Berman (1992).

SAMPLES AND METHODS

Reaction 1 was used to estimate emplacement P for 19 granitic rocks from the Sierra Nevada batholith, Peninsular Ranges batholith, and San Gabriel mountains of California (Fig. 1). The batholiths formed as a result of subduction beneath the western margin of North America during the Mesozoic. The specimens (1) form part of the sample suites studied by Ague and Brimhall (1988a, 1988b), Ague and Brandon (1992), and Pickett and Saleeby (1993); (2) range in composition from tonalite to granite; and (3) were selected to cover a broad range in Hbl Al^T and emplacement depth. All samples contain Hbl, biotite, plagioclase, K-feldspar quartz, magnetite and/or ilmenite, and titanite, with the exception of two samples from the Sierra Nevada batholith (1011-34, 1011-376), which lack titanite. Five samples contain epidote, the textural characteristics of which suggest a possible magmatic origin (Fig. 1; cf. Zen and Hammarstrom, 1984). Hbl with the lowest Al^T are from the northeastern Sierra Nevada batholith ($Al^T < 1.07$ atoms per 23 oxygens). Independent geologic evidence indicates that upper crustal and, in some cases, subvolcanic levels are exposed in this region (e.g., Ritter Range area; cf. Tobisch et al., 1985). Near Tioga Pass (Fig. 1A), the presence of andalusite and absence of sillimanite in roof pendants metamorphosed at ~600 °C (Kerrick, 1970) suggests $P < \sim 2.5$ kbar (Al_2SiO_5 stability fields after Berman, 1988). P was probably > ~1 kbar (the approximate minimum P required for Hbl stability in typical granitic systems). Samples from the central SNB probably equilibrated under upper-middle crustal conditions; widespread andalusite in roof pendants suggests $P <$ ~4–4.5 kbar (cf. Ross, 1985). The deepest parts of the Sierra Nevada batholith are exposed at its southern end (cf. Ross, 1985; Pickett and Saleeby, 1993). Thermobarometry results for meta-igneous and metasedimentary rocks are variable, but suggest that pressures were mostly in the range of 5 to 6 kbar east of Bakersfield (e.g., Dixon et al., 1994) and 7–8 kbar at the southernmost tip of the batholith (e.g., Pickett and Saleeby, 1993). The sample set also includes seven rocks from the Peninsular Ranges batholith (Fig. 1B). The depth of exposure generally increases eastward across the Peninsular Ranges batholith (cf. Todd et al., 1988, and references therein). Metamorphic pressures are not tightly constrained, but existing thermobarometry (done on wall rocks) suggest values of 4 to 5 kbar for the eastern half of the Peninsular Ranges batholith (cf. Grove, 1987; Todd et al., 1988; Rothstein and Manning, 1994). Two samples from the San Gabriel mountains were also studied (Fig. 1B).

Reaction 1 was evaluated with the TWEEQU software of Berman (1991), employing the thermodynamic data of Mäder et al. (1994) for tremolite and pargasite and Berman (1988) for the other components. Following Mäder et al. (1994), the calculations use the activity models of Furhman and Lindsley (1988) (plagioclase), McMullin et al. (1991) (biotite), and Mäder et al. (1994) (Hbl). The structural formulas for Hbl and biotite were computed following Mäder et al (1994) and McMullin et al. (1991), respectively. The hydroxyl mole fractions (X_{OH}) in Hbl and biotite were set to one because the thermodynamic data and activity models of Mäder et al. (1994) were derived under this assumption. K-feldspar generally loses significant albite component during cooling and, as a result, high-T compositions are rarely preserved in granitic rocks (cf. Czamanske et al., 1981). Consequently, the activity of $KAlSi_3O_8$ in K-feldspar ($a^{\text{K-feldspar}}_{\text{KAlSi}_3\text{O}_8}$) was set to a representative value of 0.80 for all calculations based on the K-feldspar compositions expected to coexist with observed plagioclase compositions at magmatic temperatures (cf. Furhman and Lindsley, 1988). The assumptions regarding $a^{\text{K-feldspar}}_{\text{KAlSi}_3\text{O}_8}$ and X_{OH} probably have relatively minor effects on P estimates (see Discussion).

The best estimate of equilibration T was taken to be the average of the results for the amphibole–plagioclase thermometers A and B of Holland and Blundy (1994). This best estimate was used in conjunction with results for reaction 1 to solve simultaneously for P and T (Fig. 2). A T error of ±50 °C translates into a P error of about ±1 kbar. Amphibole and plagioclase are central to both the barometer of this study (reaction 1) and the thermometers of Holland and Blundy. Thus, it is reasonable to assume that, for a given sample, the barometer and the thermometers equilibrated at similar times under similar P-T conditions.

Microprobe analyses of coexisting minerals (Appendixes 1–5[1]) were

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[1] GSA Data Repository item 9731, Appendixes 1–5, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301. E-mail: editing@geosociety.org.

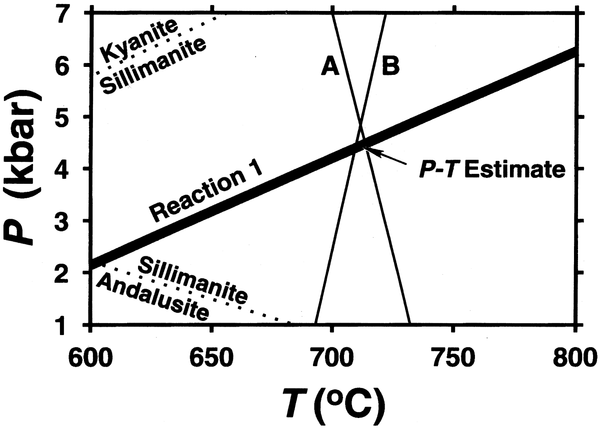

Figure 2. Pressure-temperature (*P-T*) intersection of reaction 1 with thermometers A and B of Holland and Blundy (1994) for representative sample SC-69-40.

done for this study and were also taken from Ague and Brimhall (1988a), Ague and Brandon (1992), and Pickett and Saleeby (1993). Hbl and plagioclase rim compositions were used for the calculations. For plagioclase, albite-enriched regions along grain margins and cracks that were inferred on the basis of texture to have formed during low-*T* chemical exchange with hydrothermal fluids or K-feldspar were excluded. Biotite tends to lose Ti and Ba on its rims and along bands parallel to cleavage planes during subsolidus cooling. Therefore, only those grains with the highest Ti and Ba concentrations in each rock were used for the calculations in an effort to retrieve magmatic or near-magmatic equilibration conditions.

COMPARISON OF PRESSURE ESTIMATES

The pressures calculated for the batholiths of California using reaction 1 are consistent with all the geologic evidence, including the increase in *P* southward across the Sierra Nevada batholith and the increase in *P* eastward across the Peninsular Ranges batholith (Fig. 1). The highest *P* samples (5.1–7.4 kbar) are located near the San Andreas Fault in the San Gabriel mountains, southern Sierra Nevada batholith, and eastern Peninsular Ranges batholith. Samples that contain possible magmatic epidote yield equilibration *P* of 5.2 to 7.4 kbar (Fig. 1), consistent with available field and experimental constraints that suggest that magmatic epidote only crystallizes in the middle and lower crust (cf. Zen and Hammarstrom, 1984; Schmidt, 1992). Conventional AH barometry *P* estimates computed using the experimental calibrations of the AH barometer are plotted as a function of *P* computed using the thermodynamics-based method of this study in Figure 3. It is apparent that a strong positive correlation exists between the conventional estimates and those computed using the new method. Closer inspection reveals that pressures calculated using Johnson and Rutherford's (1989) calibration of the AH barometer (abbreviated here as JR89) plot below the 1:1 line of perfect correlation (Fig. 3). The <2 kbar pressures calculated using Schmidt's (1992) AH barometer calibration (S92) also plot below the 1:1 line, but 81% of the >2 kbar *P* estimates plot above it (Fig. 3).

As emphasized by Schmidt (1992) and Anderson and Smith (1995), the Al^T of Hbl tends to increase with increasing *T* and, therefore, AH barometer *P* estimates obtained from rocks that equilibrated at temperatures that differ significantly from those of AH barometer calibration may be inaccurate. For example, if Hbl equilibrates at a temperature lower than that of the AH calibration, then *P* estimates will tend to be too low. The average *T* of equilibration for the rocks of this study is about 725 °C. The JR89 calibration was done mostly at higher temperatures of 740 to 780 °C. Thus, the systematic clustering of JR89 *P* estimates below the 1:1 line (Fig. 3) is suggested to be largely the result of *P* underestimation by the JR89 barometer. The S92 experiments followed the *P-T* path of the water-saturated tonalite solidus; *T* increased from 655 °C at 9.5 kbar to 700 °C at 2.5 kbar. Most of the *P-T* estimates for the >2 kbar samples of this study lie on the high-*T* side of the solidus. Thus, the tendency of the S92 *P* estimates to plot above the 1:1 line (Fig. 3) is attributed largely to *P* overestimation (by ~0.8 kbar, on average) by the S92 barometer. Furthermore, because the bulk of the *P-T* estimates for the >2 kbar mineral assemblages lie on the high-*T* side of the water-saturated tonalite solidus, it is likely that a_{H_2O} was less than one during the chemical equilibration of most of these assemblages. The *P-T* estimates for the three lowest *P* samples (1.2 kbar, 665 °C; 1.6 kbar, 700 °C; 2.0 kbar, 690 °C; Fig. 1A) lie on the low-*T* side of Schmidt's experimental *P-T* trend; the S92 barometer may therefore underestimate *P* for these three samples (Fig. 3).

T effects can be accounted for to some extent by using the *T*-dependent AH calibration of Anderson and Smith (1995), but errors in AH barometry can also arise from variations in fluid and rock compositions away from the compositions used for barometer calibration. It is worth noting in this regard that differences between JR89 and S92 almost certainly reflect bulk compositional factors as well as *T* effects. For example, the two studies used starting materials with significantly different $Mg/(Mg + Fe^T)$ (Fe^T = total Fe).

DISCUSSION

Use of reaction 1 to estimate equilibration *P* has six major advantages over conventional AH barometry. (1) Explicit assumptions about a_{H_2O} are eliminated. (2) Variations in the compositions of coexisting minerals are accounted for via the activity terms in the equilibrium constant for reaction 1. (3) The presence of titanite and Fe-Ti oxides is not required, thus expanding the range of rock compositions suitable for barometry. (4) Melt is not required to be part of the assemblage. Consequently, *P* estimates can be made for igneous and amphibolite and granulite facies metamorphic mineral assemblages that equilibrated below solidus temperatures. (5) The reaction can be used in low-*P* settings that are outside the *P-T* range of current experimental AH barometer calibrations. (6) *T* effects can be accounted for directly by solving for *P* and *T* simultaneously using suitable thermometers together with reaction 1.

Several other issues must also be kept in mind when using reaction 1 to estimate *P* (Hbl structural formulas and mole fractions discussed below computed following Holland and Blundy, 1994). (1) Hbl and biotite of this study have $Mg/(Mg + Fe^T)$ in the range of 0.31 to 0.68. The Ca, Al^T, and calculated $Fe^{2+}/(Fe^{2+} + Fe^{3+})$ of Hbl range from 1.81 to 1.91, 0.74 to 2.49, and 0.68 to 0.85, respectively. Whether reaction 1 yields reasonable results outside of these ranges is unknown. (2) The amphibole activity model of Mäder et al. (1994) assigns all Na to the A-site; the mole fraction of Na on M4 (X_{Na}^{M4}) is set to zero. Thus, as pointed out by Mäder et al. (1994), the model will fail for some Na-rich calcic amphiboles and all sodic-calcic amphiboles. Hbl crystals in this study have calculated X_{Na}^{M4} in the range of 0.051 to 0.075; evaluation of reaction 1 for more Na-rich Hbl remains to be done. (3) X_{OH} was set to one for Hbl and biotite in all calculations. The impact of this simplification is minimized somewhat by the fact that the X_{OH} values of coexisting biotite and Hbl are generally similar (Ague and Brimhall, 1988a) and, therefore, X_{OH} terms will tend to cancel in the equilibrium constant for reaction 1. However, nonideal interactions in very F rich and/or Cl rich hydrous silicates could potentially cause significant errors in *P* estimates. The hydrous silicates of this study have $X_{OH} > 0.89$; evaluation of reaction 1 for more halogen-rich compositions remains to be done. (4) Estimates of Fe^{2+}/Fe^{3+} based on structural formula calculation schemes are subject to significant, difficult-to-quantify uncertainties (cf. Cosca et al., 1991). The impact of these uncertainties on *P* estimates is reduced, however, by focusing on Mg end member reactions like reaction 1. (5) The value of $a_{KAlSi_3O_8}^{K\text{-feldspar}}$ was set to a representative value of 0.8. The calculation results are not highly sensitive to this simplification. For example, changing $a_{KAlSi_3O_8}^{K\text{-feldspar}}$ by +0.04 (a reasonable range for the rocks studied here) at constant *T* changes *P* estimates by only ±120 bar. (6) Reaction 1 has not been investigated experimentally. However, Mäder and Berman's (1992) thermodynamic modeling

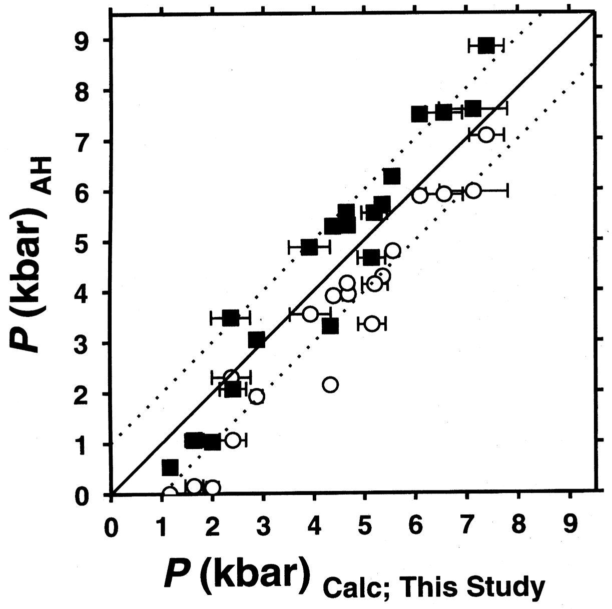

Figure 3. Conventional aluminum-in-hornblende (AH) barometer pressure (*P*) estimates computed using calibrations of Johnson and Rutherford (1989) (open circles) and Schmidt (1992) (filled squares) plotted as function of *P* estimates calculated using approach of this study. 1:1 line of perfect correlation is denoted by solid line; dashed lines 1 kbar above and below 1:1 line shown for reference. Error bars represent range in *P* corresponding to range in temperature calculated using thermometers A and B of Holland and Blundy (1994); bars are not intended to represent total uncertainty on individual *P* estimates (see text).

of amphibole end members included a broad range of data from experimental (and natural) systems. It is important that their modeling was done completely independently of all experimental and empirical AH barometer calibrations.

It is suggested that the method of this paper, which uses reaction 1, provides more consistent and reliable *P* estimates than conventional AH barometry because (1) the method is based on quantitative thermodynamics and (2) no a priori assumptions about a_{H_2O}, magma composition, or *T* are needed to use reaction 1. Of course, full confirmation of the *P-T* estimates presented herein will require additional field and experimental studies aimed ultimately at refining the thermodynamic properties and activity-composition relations for amphibole end members. The results shown in Figure 3 suggest that regional gradients in equilibration *P* determined previously for Cordilleran batholiths using conventional AH barometry (e.g., Ague and Brimhall, 1988b; Ague and Brandon, 1992, 1996) are robust, but that the absolute values of the *P* estimates need to be revised somewhat.

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SUPPLEMENTAL MATERIAL—SAMPLES AND MINERAL CHEMISTRY[1]

APPENDIX 1. SAMPLES AND PRESSURE–TEMPERATURE ESTIMATES

Sample No.	Batholith*	Longitude	Latitude	P (kbar)	T (°C)	Sample Collector
SC-69-1b	PRB	242.4783	33.8503	2.0	690	Armstrong and Suppe (1973)
SC-69-40	PRB	243.2350	33.7039	4.4	710	Armstrong and Suppe (1973)
SC-69-41	PRB	243.2814	33.7008	4.7	695	Armstrong and Suppe (1973)
SC-69-43a	PRB	243.4047	33.5672	4.7	720	Armstrong and Suppe (1973)
SC-69-46	PRB	243.5639	33.5831	5.6	745	Armstrong and Suppe (1973)
SC-69-47	PRB	243.7022	33.7161	6.6	760	Armstrong and Suppe (1973)
SC-69-153	PRB	243.7842	33.5681	6.1	690	Armstrong and Suppe (1973)
1011-34	SNB	241.4083	36.1542	2.4	690	Ague and Brimhall (1988a, 1988b)
1011-49	SNB	241.3250	35.2667	5.4	715	Ague and Brimhall (1988a, 1988b)
1011-60	SNB	240.6125	37.8500	2.4	730	Ague and Brimhall (1988a, 1988b)
1011-62	SNB	240.5250	37.8167	1.6	700	Ague and Brimhall (1988a, 1988b)
1011-175	SG	241.9542	34.2875	7.1	765	Ague and Brimhall (1988a, 1988b)
1011-188	SG	242.1542	34.2833	5.1	720	Ague and Brimhall (1988a, 1988b)
1011-263	SNB	240.6667	38.2167	1.2	665	Ague and Brimhall (1988a, 1988b)
1011-311	SNB	240.4917	37.1375	3.9	745	Ague and Brimhall (1988a, 1988b)
1011-376	SNB	241.0958	36.7250	2.9	745	Ague and Brimhall (1988a, 1988b)
1011-385	SNB	241.2417	36.2042	4.3	740	Ague and Brimhall (1988a, 1988b)
1011-415	SNB	241.6750	35.4333	5.2	740	Ague and Brimhall (1988a, 1988b)
GC-33	SNB	241.000§	34.890§	7.4	735	Pickett and Saleeby (1993)

Note: Format for longitude and latitude is degrees.decimal degrees.

*SNB = Sierra Nevada batholith; PRB = Peninsular Ranges batholith; SG = San Gabriel mountains.

§Estimated from Figure 2 in Pickett and Saleeby (1993).

APPENDIX 2. HORNBLENDE ANALYSES (wt. %)

	Ref.*	SiO_2	TiO_2	Al_2O_3	FeO	MgO	MnO	CaO	Na_2O	K_2O	F	Cl	Total
SC-69-1b	1	49.94	0.81	4.96	14.72	14.32	0.24	11.68	0.86	0.43	0.11	0.13	98.20
SC-69-40	1	43.72	1.16	9.80	20.28	8.64	0.34	11.57	1.08	1.13	0.17	0.07	97.96
SC-69-41	1	43.99	0.90	9.86	20.26	8.72	0.34	11.60	1.07	1.06	0.15	0.07	98.02
SC-69-43A	1	43.90	1.24	10.26	19.14	9.49	0.34	11.60	1.08	1.21	0.14	0.10	98.50
SC-69-46	1	42.93	0.88	11.11	19.62	9.40	0.40	11.63	1.16	1.22	0.14	0.10	98.59
SC-69-47	1	40.81	0.77	12.43	21.17	8.12	0.32	11.67	1.39	1.45	0.12	0.09	98.34
SC-69-153	1	42.14	0.44	12.52	19.56	8.36	0.37	11.35	1.43	1.31	0.14	0.06	97.68
1011-34	2	44.77	1.30	7.60	23.55	7.08	0.64	11.49	1.00	0.88	0.20	0.16	98.67
1011-49	2	45.06	1.26	10.64	16.65	10.92	0.37	11.76	1.23	0.75	0.04	0.07	98.75
1011-60†	2,4	47.54	1.05	6.21	14.58	13.44	0.88	11.75	1.28	0.63	0.34	0.05	97.75
1011-62	2	50.05	0.64	5.07	14.33	14.55	0.92	11.74	1.07	0.45	0.18	0.01	99.01
1011-175	2	39.79	0.84	12.36	22.16	6.97	1.26	11.59	1.53	1.58	0.18	0.15	98.41
1011-188	2	44.23	0.84	9.14	20.56	9.22	0.63	11.64	1.35	0.98	0.08	0.02	98.69
1011-263	2	50.74	0.58	4.38	12.58	15.28	0.57	12.29	0.75	0.41	0.29	0.03	97.90
1011-311	2	43.95	1.27	9.42	19.34	9.82	0.50	11.96	1.19	1.07	0.09	0.07	98.68
1011-376	2	45.91	1.11	7.26	19.05	10.58	0.59	11.56	1.11	0.74	0.13	0.09	98.13
1011-385	2	45.77	1.31	7.58	18.32	10.93	0.51	11.42	1.29	0.71	0.10	0.12	98.06
1011-415	2	42.23	1.78	9.97	23.41	6.53	0.50	11.41	1.35	1.39	0.22	0.11	98.90
GC-33	3	39.69	1.41	13.74	22.42	5.70	0.54	11.42	1.38	1.93	N.D.	N.D.	98.23

Notes : All Fe as FeO. N.D. = not determined. Mineral compositions measured in this study were determined using the JEOL JXA-8600 electron microprobe at Yale University, employing wavelength dispersive spectrometers, natural and synthetic standards, off–peak and fluorescence–corrected mean atomic number background corrections, and $\phi(\rho z)$ matrix corrections.

*Analysis reference: 1 = Ague and Brandon (1992); 2 = Ague and Brimhall (1988a, 1988b); 3 = Pickett and Saleeby, (1993) 4 = This study.

†Redetermined in this study to avoid chloritic areas.

APPENDIX 3. BIOTITE ANALYSES (wt. %)

	Ref.*	SiO_2	TiO_2	Al_2O_3	FeO	MgO	MnO	CaO	BaO	Na_2O	K_2O	F	Cl	Total
SC-69-1b	2	36.51	4.50	12.80	20.63	11.25	0.21	N.D.	0.03	0.10	9.16	0.22	0.32	95.73
SC-69-40	2	35.81	4.12	15.09	22.01	8.49	0.23	N.D.	0.40	0.09	8.99	0.18	0.04	95.45
SC-69-41	2	36.06	4.04	15.13	22.64	8.40	0.25	N.D.	0.50	0.11	9.05	0.22	0.07	96.47
SC-69-43A	2	36.07	4.32	14.90	21.80	9.67	0.31	N.D.	0.18	0.03	9.05	0.15	0.11	96.59
SC-69-46	2	36.56	3.28	15.23	21.09	9.85	0.33	N.D.	0.17	0.09	9.08	0.15	0.07	95.90
SC-69-47	2	35.89	2.94	15.05	21.22	10.10	0.33	N.D.	0.10	0.08	9.17	0.25	0.05	95.18
SC-69-153	2	36.79	4.23	15.93	19.60	9.36	0.41	N.D.	0.07	0.05	9.52	0.14	0.04	96.14
1011-34	1	35.36	3.40	14.55	24.11	7.78	0.35	0.01	N.D.	0.07	8.87	0.46	0.23	95.19
1011-49	1	36.81	2.74	16.67	18.14	11.54	0.14	0.01	N.D.	0.16	9.49	0.10	0.05	95.85
1011-60†	1,2	38.11	2.55	13.50	16.87	13.50	0.76	N.D.	B.D.	0.09	9.20	0.88	0.01	95.47
1011-62	1	37.12	2.94	13.51	17.61	13.30	0.57	B.D.	N.D.	0.09	9.46	0.48	0.03	95.11
1011-175†	1,2	36.61	2.39	15.28	21.50	9.16	1.00	N.D.	0.09	0.08	9.21	0.27	0.11	95.70
1011-188	1	35.80	2.81	15.22	23.09	9.87	0.42	0.03	0.31	0.09	9.27	0.18	0.02	97.11
1011-263	1	37.67	1.55	15.42	16.57	13.93	0.44	B.D.	0.22	0.06	9.63	0.75	0.03	96.27
1011-311	1	36.57	3.25	15.54	20.68	10.40	0.31	0.03	0.19	0.08	9.72	0.18	0.02	96.97
1011-376	1	36.09	4.23	14.43	21.99	9.85	0.36	0.02	0.20	0.09	9.50	0.24	0.13	97.13
1011-385	1	36.43	3.56	14.89	21.59	10.04	0.37	0.02	0.41	0.07	9.37	0.18	0.08	97.01
1011-415†	1,2	35.45	3.62	15.35	24.67	6.82	0.31	N.D.	0.19	0.08	9.27	0.24	0.05	96.05
GC-33	2	35.14	4.19	14.71	24.35	7.04	0.34	N.D.	0.38	0.07	9.10	0.08	0.22	95.62

Notes : All Fe as FeO. B.D. = below detection; N.D.=not determined. Mineral compositions measured in this study were determined using the JEOL JXA-8600 electron microprobe at Yale University, employing wavelength dispersive spectrometers, natural and synthetic standards, off–peak and fluorescence–corrected mean atomic number background corrections, and $\phi(\rho z)$ matrix corrections.

*Analysis reference: 1 = Ague and Brimhall (1988a, 1988b); 2 = This study.

†Redetermined in this study to avoid low–Ti, low–Ba areas.

APPENDIX 4. PLAGIOCLASE ANALYSES (wt. %)

	SiO_2	Al_2O_3	FeO	CaO	BaO	Na_2O	K_2O	Total	X_{Ab}	X_{An}	X_{Or}
SC-69-1b	60.06	24.71	0.17	6.50	B.D.	7.83	0.09	99.36	0.682	0.313	0.005
SC-69-40	59.04	25.35	0.07	7.25	0.05	7.26	0.19	99.20	0.638	0.352	0.011
SC-69-41	61.03	25.25	0.08	6.80	0.03	7.74	0.22	101.14	0.665	0.323	0.013
SC-69-43A	59.43	26.12	0.10	7.87	0.04	7.11	0.19	100.85	0.614	0.376	0.011
SC-69-46	58.47	25.77	0.10	7.68	0.09	7.26	0.13	99.51	0.626	0.366	0.008
SC-69-47	60.78	25.13	0.23	6.57	0.05	7.94	0.20	100.90	0.679	0.310	0.011
SC-69-153	59.95	24.72	0.08	6.44	0.04	7.80	0.14	99.17	0.681	0.311	0.008
1011-34	63.38	23.21	0.06	4.57	0.02	9.08	0.16	100.49	0.775	0.216	0.009
1011-49	59.55	26.70	0.11	8.31	0.05	7.08	0.09	101.89	0.603	0.391	0.005
1011-60	61.67	24.37	0.11	5.85	0.04	8.29	0.16	100.49	0.713	0.278	0.009
1011-62	62.25	23.62	0.14	5.55	0.04	8.07	0.18	99.86	0.717	0.273	0.011
1011-175	63.00	23.61	0.06	4.98	0.02	8.77	0.16	100.59	0.755	0.237	0.009
1011-188	62.47	23.94	0.14	5.23	0.04	8.66	0.22	100.69	0.741	0.247	0.012
1011-263	62.78	23.61	0.14	5.08	0.05	8.31	0.70	100.67	0.717	0.242	0.040
1011-311	60.29	24.92	0.10	6.64	0.04	7.82	0.18	99.99	0.674	0.316	0.010
1011-376	60.57	25.04	0.13	6.64	0.03	7.76	0.16	100.33	0.673	0.318	0.009
1011-385	61.60	24.62	0.11	6.17	0.03	8.09	0.16	100.78	0.697	0.294	0.009
1011-415	60.55	25.15	0.42	6.78	0.09	7.56	0.22	100.77	0.707	0.279	0.014
GC-33	60.06	25.00	0.05	6.60	0.13	7.44	0.75	100.02	0.642	0.315	0.043

Notes : All Fe as FeO. B.D. = below detection. All compositions determined in this study using the JEOL JXA-8600 electron microprobe at Yale University, employing wavelength dispersive spectrometers, natural and synthetic standards, off–peak and fluorescence–corrected mean atomic number background corrections, and $\phi(\rho z)$ matrix corrections.

As discussed in the text, the activity of $KAlSi_3O_8$ in K-feldspar was set to 0.8 for the calculations. Care needs to be exercised when doing calculations with the TWEEQU software of Berman (1991; version 1.02) that involve K-feldspar. Normally, the program uses the Or component of plagioclase to compute the activity of $KAlSi_3O_8$, regardless of whether or not the activity was set to the fixed value of 0.8. To get around this difficulty, one can add a "new" phase to the Berman data base that has exactly the same composition and thermodynamic properties as K-feldspar. Give this "new" phase a name other than K–feldspar, K–feldspar (high), or K–feldspar (low). This "new" phase can then be used in the calculations, and its activity can be fixed at the desired value.

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