Intrusive History and Petrogenesis of the Ash Mountain Complex, Sierra Nevada Batholith, California (USA)

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ABSTRACT

The Ash Mountain Complex (AMC) in the western Sierra Nevada batholith (SNB; California, USA) is an exposure of six compositionally diverse intrusive lithologies with clear crosscutting relationships that permit a focused investigation of magma source characteristics and the relative roles of petrogenetic processes on the evolution of the system. We use new field observations, zircon U-Pb dates, major and trace element data, and Sr-Nd-Pb isotopic data to develop a model that can be applied to similar SNB intrusive suites. Stage 1 units, emplaced ca. 105 Ma, consist of two gabbros, a gabbrodiorite, and a granite. Stage 2 and stage 3 units were emplaced ca. 104 Ma and ca. 103 Ma, respectively, and are granites. We suggest that stage 1 gabbroids were derived by partial melting of lithospheric mantle, whereas coeval felsic magmas were derived by partial melting of a mafic, juvenile crustal source. Stage 2 and stage 3 granitoids were derived from similar sources that generated stage 1 granitoids, but there was greater input from evolved crust. Fractionation and/or assimilation played only a minor role in system evolution. Past studies of SNB magmas have come to conflicting conclusions about the petrogenesis of intermediate magmas that dominate the batholith; we hypothesize that mafic and felsic end members of the AMC could represent end members in mixing processes that generate these magmas. The timing of emplacement of the AMC coincides with a transition of mafic intrusive style in the SNB, from smaller volume mafic magmatic suites with mixed mantle and crustal sources to larger volume magmatic suites derived from greater proportions of crust.

INTRODUCTION

The interplay of source melting dynamics, fractional crystallization, and magma hybridization by mixing and assimilation generates a wide variety of granitic magmas. Granitoid rocks of calc-alkaline batholiths, forming at the fringes of continents, are particularly intriguing as they can chemically record the variable reworking of preexisting crust, accreted island arc terranes, mafic underplate, and/or lithospheric mantle. Although the net contribution to crustal growth by continental arc magmatism remains uncertain, there is growing evidence that continental arc processes act to transform accreted terranes into bona fide continental crust (e.g., Lee et al., 2007; Lackey et al., 2012). Discoveries that texturally homogeneous granodiorite plutons may be incrementally constructed over time scales of several million years (e.g., Coleman et al., 2004; Davis et al., 2012) highlight the need for better understanding of magma sources and mixing processes in the upper mantle and lower crust of magmatic arcs. With that goal in mind, we targeted the intrusive Ash Mountain Complex (AMC) in the foothills of Sequoia National Park (California, USA). Previously documented Sr isotopic data for samples from areas near the AMC (Chen and Tilton, 1991) show that the initial $^{87}$Sr/$^{86}$Sr (Sr) = 0.706 line, which roughly denotes the boundary between accreted terranes underlying the western Sierra Nevada batholith (SNB) and continental crust underlying the eastern SNB, is immediately west of the AMC (Fig. 1). Therefore, a range of source and contamination scenarios is possible for the petrogenesis of magmas emplaced in the region.

While workers suggest that fractional crystallization could account for the large compositional diversity exhibited within the SNB (e.g., Bowen, 1928; Bateman and Chappell, 1979), others (e.g., Reid et al., 1993; Sisson et al., 1996; Ratajeski et al., 1990; Sisson et al., 1996; Ratajeski et al., 2001; Wenner and Coleman, 2004; Pignotta et al., 2010) proposed that mixing and mingling of mafic and silicic end-member magmas were important in generating this diversity (e.g., Reid et al., 1983; Frost and Mahood, 1987; Dorais et al., 1990; Sisson et al., 1996; Ratajeski et al., 2001; Wenner and Coleman, 2004). For example, partial melting of intermediate country rock may generate high-SiO$_2$ melts (e.g., Skjerlie and Johnston, 1993; Patiño-Douce, 1997), but mafic underplate and hydrous cumulates may also partially melt to generate high-SiO$_2$ magmas (e.g., Rushmer, 1991; Ratajeski et al., 2005; Sisson et al., 2005). Growing evidence suggests that heat flux provided by magmatic recharge (e.g., Bergantz, 1989; Dufek and Bergantz, 2005; Annen et al., 2006) is sufficient to drive voluminous melting of fertile crustal sources. Such crustal rocks may be drawn down by return flow along the margins of growing plutons, and potentially be inserted into magma source regions (Saleeby et al., 2003) or underthrust into these source regions by forearc or retroarc compressional deformation (Duca, 2001; Duca and Barton, 2007; DeCelles et al., 2009).

In this study we investigate the ca. 105 Ma (Holland et al., 2010) comagmatic gabbro to granite AMC in the Kaweah River valley to evaluate felsic magma source characteristics as well as processes of magma differentiation. We use major and trace element and isotopic geochemistry of these rocks to evaluate the relative roles of partial melting, mixing, mingling, and assimilation–fractional crystallization (AFC) in the origin of AMC magmas. In addition, we interpret single zircon isotopic and trace element data to determine absolute ages and to evaluate petrogenetic processes. Memeti et al. (2010a) and
Figure 1. Geologic map of the south-central Sierra Nevada batholith (SNB; after Moore, 2001). Lower right inset shows map location in California. SIS denotes the location of the Sequoia Intrusive Suite. The black dashed line represents the regional 0.706 ($^{87}\text{Sr}/^{86}\text{Sr})_i$ isopleth (Kistler and Peterman, 1973, 1978; DePaolo, 1980, 1981; Kistler et al., 1986; Chen and Tilton, 1991). The red dashed line separates the western and eastern SNB based on Sr and O isotopes as well as relationships of rock types in roof pendants (Kistler, 1990). The blue line is the boundary of Sequoia National Park. The black box indicates the map area shown in Figure 2.

Davis et al. (2012) used detailed analyses of single zircon crystals to evaluate the emplacement history of large-volume, intermediate intrusive suites of the eastern Sierra Nevada, but a similar method has not been applied to compositionally diverse, mingled or mixed magmas in the SNB. In cases where this type of analysis has been applied to mixed magmatic systems like the zoned Spirit Mountain batholith in Nevada (Claiborne et al., 2010), it has proven especially powerful for evaluating the relative importance of magma mixing and crystallization processes.

Emplacement timing of the AMC is also significant within the context of SNB Cretaceous magmatism. The ca. 108–105 Ma period has been recognized (Lackey et al., 2012) as an important period of transition of magmatic style in the Sierra Nevada, from smaller volume magmatic suites dominated by mixed mantle and crustal sources and recycling of accreted terranes to larger volume magmatic suites derived from greater proportions of remelted, preexisting arc crust including local crust or underthrust rocks. Therefore, the AMC may provide critical insights into the nature of coeval and proximal mafic and felsic magmas during this period.

**GEOLOGIC SETTING**

The SNB formed during the prolonged eastward subduction of the Farallon plate below the North American continent (e.g., Dickinson, 1981). The system was active for more than 140 m.y., from the Triassic to the Late Cretaceous (e.g., Stern et al., 1981; Chen and Moore, 1982), but most magmatic activity occurred in two brief episodes. The first pulse of volumetrically significant magmatism lasted from ca. 180 to 165 Ma in the Jurassic, with a significantly more voluminous phase between 102 and 85 Ma (e.g., Coleman and Glazner, 1997; Duea, 2001; Wenner and Coleman, 2004; Coleman et al., 2004; Memeti et al., 2010a; Davis et al., 2012).

The majority of exposed rocks in the SNB are intermediate (granodiorite) in composition (e.g., Bateman, 1992; Fig. 1), but the common occurrence of these rocks with coeval low- and high-silica rocks (e.g., Coleman et al., 1995; Ratjeski et al., 2001; Wenner and Coleman, 2004) suggests a possible common genetic origin (Frost and Mahood, 1987). Accumulation of a thick lithospheric root beneath the Sierran arc is deduced to have occurred in at least the Late Cretaceous, as recorded by pressure-temperature-time conditions and isotopic ratios of garnet pyroxenite and garnet-hornblende granulite xenoliths in the central SNB (Dodge et al., 1986, 1988; Mukhopadhyay and Manton, 1994; Lee et al., 2000, 2006; Duea, 2001, 2002). Duea (2002) used elemental compositions and isotopic ratios for these xenoliths to argue that SNB granitoids and the xenoliths are the melts and residues, respectively, resulting from partial melting at depths exceeding ~35–40 km.

In the region surrounding the AMC, Cretaceous intrusive rocks, dominantly quartz diorite to granodiorite, with minor gabbro and diorite, range in age from 133 to 97 Ma (Sisson and Moore, 1994; Fig. 2). The Frys Point granite and the Milk Ranch Peak granodiorite are the largest local plutons, while quartz diorite, diorite, and gabbro are most commonly found as isolated bodies within the Frys Point granite (Fig. 2). Metamorphic pendants are found near several sample localities (Fig. 2).

Prior studies of these adjacent intrusive rocks were largely restricted to regional mineralogical, trace element (Ague and Brimhall, 1988), and Sr, Pb, Nd, and O isotopic studies (DePaolo, 1981; Kistler, 1990; Chen and Tilton, 1991; Wenner and Coleman, 2004; Lackey et al., 2006, 2008). These investigations included rocks proximal to (and possibly part of) the AMC, including the Frys Point granite and some gabbro bodies. Magmas in the region interacted with abundant Triassic to Cretaceous (Chen and Moore, 1982; Sisson and Moore, 1994) pendants and septa of carbonateous metamorphic wall-rocks (Figs. 1 and 2), based on mineralogical and mineral chemistry redox indicators (Ague and Brimhall, 1988; Lackey et al., 2006). Isotopic studies provide further evidence for localized contamination (Chen and Tilton, 1991; Lackey et al., 2006, 2008) and suggest that both mafic and felsic rocks were derived from sources with crustal affinities (e.g., high δ18O, high Sr, high Pb; Wenner and Coleman, 2004; Lackey et al., 2006). For example, δ18O (zircon) values for the low-SiO2 (~47 wt%) Elk Creek Gabbro body in the AMC (unit KJgb, Fig. 2) are ~8.3‰ (Lackey et al.,
2006), higher than many granites in the SNB (Lackey et al., 2008) and suggestive of melting of crustal sources.

Ash Mountain Complex

Ross (1958, p. 6) initially described the AMC in southwestern Sequoia National Park (Figs. 1 and 2) with field observations limited to an area of ~2.6 km² “in the vicinity of the Ash Mountain Park headquarters”; he identified four intrusive lithologies, including a quartz diorite, a granite, a gray, fine-grained intrusive (unclassified), and a dark fine-grained intrusive (unclassified). Ross (1958) was not able to determine a clear chronology of intrusion, in large part due to mutually crosscutting relations, with observations from different localities suggesting conflicting chronologies of intrusion. We expand the area of the AMC to include exposures along the Kaweah River, based on the similar diversity in intrusive compositions and the relative proximity of the Kaweah River exposure to the AMC as defined by Ross (1958).

METHODS

Sample Collection and Preparation

We sampled each of the six AMC units identified in fresh outcrop exposures along the Kaweah River (location 4 in Fig. 2) for petrographic and geochemical analyses. We also collected samples representing the compositional range of AMC intrusive lithologies from other localities (locations 1, 2, 3, 5, and 6 in Fig. 2) in order to provide a broader regional context for the AMC sampling.
We prepared whole-rock powders for geochemical analysis by standard chipping and pulverizing (2 min) in a Rocklabs tungsten carbide head and swing mill, which was cleaned with quartz sand between samples. We used standard hydraulic, gravimetric, and magnetic separation techniques to concentrate zircon grains, employing a Rogers concentrating table, methylene iodide (density 3.32 g/cm³), and a Frantz magnetic separator. We performed final zircon purification by hand-picking under a petrographic microscope.

**X-Ray Fluorescence Major and Trace Element Geochemistry**

We measured major and selected trace elements for 19 samples by X-ray fluorescence (XRF) (Table 1) at Pomona College. We vortex blended rock powder and flux in a 1:2 ratio (3.5 g powder to 7.0 g dilithium tetraborate), and fused these mixtures to glass beads in graphite crucibles heated at 1000 °C for 10 min. We regrind the beads and fused them a second time, polished the beads on diamond laps, and analyzed the polished beads. We determined concentrations according to calibrations based on 55 certified reference materials that span a large range of compositions (see Lackey et al., 2012, for further details regarding XRF analysis at Pomona College). Typical analytical 2 standard deviation uncertainties for analyses (in wt%) of major elements, determined conservatively as reproducibility of multiple beads produced from a sample or standard powder, are ±0.15% (SiO₂); 0.06%–0.07% (Fe₂O₃, Al₂O₃); 0.02%–0.05% (Na₂O, K₂O, MgO, CaO); <0.004% (TiO₂, MnO, P₂O₅). Trace element uncertainties (±2 standard deviations) are 1 ppm (La, Pr, Hf), 2 ppm (Nd, Cu, Ga, Nb, Pb, Rh, Th, U, Y, Zr) to 3 ppm (Cr, Sc, Sr, Y) for rare earth elements (REEs), Y and Hf (1 s/mass) (M). The mean ratios for standards analyzed over 6 days are 87Sr/86Sr = 0.710271 ± 9 (2 standard deviations, n = 39) for NBS 987, 143Nd/144Nd = 0.512069 ± 8 (2 standard deviations, n = 8) for Ames Nd, and 208Pb/204Pb = 16.934 ± 6. We corrected 87Sr for 87Rb by simultaneous measurement of 86Sr, determined 87Rb/86Rb = 0.38600 and corrected 144Nd for 144Sm by simultaneous measurement of 147Sm/144Sm = 0.206700. The mean ratios for standards analyzed over 6 mo during the time of analyses are 87Sr/86Sr = 0.710271 ± 9 (2 standard deviations, n = 39) for NBS 987, 143Nd/144Nd = 0.512069 ± 8 (2 standard deviations, n = 8) for Ames Nd, and 208Pb/204Pb = 16.934 ± 6. We corrected 87Sr for 87Rb by simultaneous measurement of 86Sr, determined 87Rb/86Rb = 0.38600 and corrected 144Nd for 144Sm by simultaneous measurement of 147Sm/144Sm = 0.206700. The mean ratios for standards analyzed over 6 mo during the time of analyses are 87Sr/86Sr = 0.710271 ± 9 (2 standard deviations, n = 39) for NBS 987, 143Nd/144Nd = 0.512069 ± 8 (2 standard deviations, n = 8) for Ames Nd, and 208Pb/204Pb = 16.934 ± 6. We corrected 87Sr for 87Rb by simultaneous measurement of 86Sr, determined 87Rb/86Rb = 0.38600 and corrected 144Nd for 144Sm by simultaneous measurement of 147Sm/144Sm = 0.206700.

**RESULTS**

**Field Relationships**

Excellent continuous and uncovered exposure of the AMC extends for more than 180 m along the north side of the Kaweah River (sample location 4, Fig. 3), with more than 2000 m² of fresh surface for detailed observation. This spectacular exposure reveals the occurrence of granite and gabbro. Six units were identified at location 4: (1) gabbro porphyry (Kgp); (2) equigranular gabbro (Kg); (3) gabbro (Kgd); (4) white granite (Kgr); (5) medium gray granite porphyry (Kp); and (6) pegmatitic granite (Kpg). Crosscutting relationships and the nature of contacts among the first five units indicate at least two stages of intrusion (Fig. 3), and subsequent U-Pb zircon age dates (see following discussion of Zircon U-Pb Age Dating) reveal a three-stage intrusion history.
| Sample | 10JH14 | 10JH15 | 10SQ21 | 10JH08 | 10JH01 | 10JH05 | 10JH05-2* | 10JH09 | 10JH20 | 10JH21 |
|--------|--------|--------|--------|--------|--------|--------|-----------|--------|--------|--------|
| Rock type | Diorite | Granite | Granite | Gabbro | Gabbro | Granite | Granite | Pegmatite | Gabbro | Gabbro |
| Location description | Sycamore Drive | Sycamore Drive | Frys Point pluton interior, Kfp | General's Highway | General's Highway | General's Highway | General's Highway | General's Highway | Kaweah River, Unit Kgp | Kaweah River, Unit Kg |
| Mineralogy† | Pl, Bt, Hbl, Qz | Kfs, Pl, Qz, Bt | Pl, Hbl, Bt, Qz | Pl, Hbl, Bt, Qz | Pl, Hbl, Bt, Qz, Kfs | Pl, Hbl, Bt, Qz | Pl, Hbl, Bt, Qz | Pl, Hbl, Bt, Qz | Pl, Opx, Cpx, Hbl, Bt | Pl, Hbl, Bt, Cpx, Qz |
| Major elements (wt% oxide) | | | | | | | | | | |
| SiO₂ | 58.67 | 75.03 | 72.61 | 47.76 | 47.27 | 69.70 | 69.66 | 76.91 | 47.76 | 48.69 |
| TiO₂ | 0.62 | 0.07 | 0.26 | 1.25 | 1.07 | 0.44 | 0.43 | 0.04 | 1.67 | 1.15 |
| Al₂O₃ | 16.86 | 13.55 | 14.08 | 17.54 | 20.94 | 14.97 | 15.08 | 13.39 | 17.48 | 17.37 |
| Fe₂O₃ | 7.99 | 1.46 | 2.56 | 12.32 | 9.52 | 3.71 | 3.82 | 0.52 | 11.2 | 11.45 |
| MnO | 0.15 | 0.03 | 0.05 | 0.20 | 0.15 | 0.06 | 0.06 | 0.01 | 0.19 | 0.19 |
| MgO | 3.94 | 0.11 | 0.33 | 6.78 | 6.11 | 0.87 | 0.84 | 0.07 | 7.28 | 7.54 |
| CaO | 7.00 | 0.59 | 1.39 | 10.54 | 12.03 | 2.67 | 2.64 | 1.07 | 12.28 | 10.65 |
| Na₂O | 3.05 | 3.94 | 3.62 | 2.09 | 1.92 | 3.36 | 3.35 | 3.54 | 1.54 | 2.07 |
| K₂O | 1.44 | 4.69 | 4.84 | 0.44 | 0.44 | 3.81 | 3.76 | 4.30 | 0.38 | 0.58 |
| P₂O₅ | 0.09 | 0.05 | 0.06 | 0.17 | 0.27 | 0.11 | 0.12 | 0.03 | 0.07 | 0.09 |
| Total | 99.81 | 99.52 | 99.80 | 99.09 | 99.72 | 99.70 | 99.76 | 100.08 | 99.85 | 99.78 |
| Trace elements (ppm) | | | | | | | | | | |
| Ni | 20 | nd | 3 | 23 | 44 | nd | 3 | 3 | 27 | 41 |
| Cr | 18 | 4 | 8 | 74 | 86 | 7 | 7 | 4 | 129 | 118 |
| V | 200 | 5 | 10 | 295 | 201 | 26 | 24 | 4 | 335 | 337 |
| Sc | 27 | 5 | 6 | 45 | 37 | 10 | 11 | 3 | 44 | 40 |
| Zn | 77 | 41 | 65 | 97 | 59 | 80 | 63 | 17 | 71 | 80 |
| Ga | 15 | 16 | 20 | 18 | 17 | 18 | 16 | 15 | 17 | 17 |
| Ba | 470 | 1141 | 1132 | 193 | 1356 | 1501 | 1619 | 452 | 143 | 153 |
| Rb | 47 | 100 | 176 | 6 | 6 | 116 | 113 | 127 | 8 | 12 |
| Th | 6 | 10 | 13 | 4 | 5 | 12 | 15 | 18 | 4 | 3 |
| Sr | 308 | 50 | 134 | 454 | 469 | 270 | 264 | 94 | 401 | 388 |
| Zr | 37 | 81 | 237 | 72 | 43 | 227 | 225 | 59 | 42 | 57 |
| Nb | 9 | 6 | 11 | 9 | 7 | 15 | 13 | 10 | 10 | 8 |
| Hf | 2.3 | 4.0 | 6.8 | 3.1 | 1.3 | 6.8 | 6.5 | 4.1 | 2.1 | 2.6 |
| Ta | 16 | 21 | 31 | 9 | 5 | 47 | 47 | 15 | 7 | 17 |
| Ce | 26 | 55 | 69 | 21 | 14 | 108 | 102 | 27 | 12 | 25 |
| Pr | 3.8 | 5.1 | 7.3 | 2.8 | 2.0 | 10.2 | 10.6 | 4.0 | 2.6 | 4.4 |
| Nd | 12 | 15 | 25 | 21 | 5 | 35 | 33 | 7 | 11 | 18 |
| Y | 20 | 29 | 41 | 26 | 19 | 21 | 22 | 21 | 17 | 27 |
| U | 3 | 3 | 5 | 2 | 5 | 3 | 6 | 8 | 1 | 2 |
| Pb | 8 | 26 | 20 | 8 | 3 | 16 | 15 | 33 | 5 | 5 |
| Cu | 37 | 3 | 4 | 34 | 29 | 6 | 3 | 5 | 45 | 35 |

(continued)
TABLE 1. MAJOR AND TRACE ELEMENT COMPOSITIONS OF ASH MOUNTAIN COMPLEX SAMPLES (continued)

| Sample       | 10JH18 | 10JH22 | 10JH17 | 10JH19 | 10JH16 | 10JH11 | 10SQ17 | 10SQ15 | 10SQ16 |
|--------------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| Rock type    | Gabbro | Gabbrodiorite | Granite | Granite | Granite | Granite | Gabbro | Gabbro | Gabbrodiorite | Granite |
| Location     | Kaweah River, Unit Kgd | Kaweah River, Unit Kgd | Kaweah River, Unit Kp | Kaweah River, Unit Kp | Kaweah River, Unit Kgr | Elkh Creek | Trail to Marble Falls | Trail to Marble Falls | Trail to Marble Falls |
| Mineralogy†  | Pl, Hbl, Qtz, Kfs, Bt | Pl, Hbl, Cpx, Qtz, Kfs | Pl, Qtz, Kfs, Bt | Pl, Qtz, Kfs, Bt | Pl, Cpx, Qtz, Kfs, Bt | Pl, Hbl, Qtz, Kfs, Bt | Pl, Hbl, Qtz, Kfs, Bt | Pl, Hbl, Qtz, Kfs, Bt | Pl, Hbl, Qtz, Kfs, Bt |
| Major elements (wt% oxide) | |
| SiO₂         | 50.57  | 54.05  | 70.95  | 71.42  | 72.87  | 47.27  | 50.37  | 54.13  | 71.09  |
| TiO₂         | 1.06   | 1.42   | 0.35   | 0.34   | 0.25   | 1.52   | 1.59   | 1.22   | 0.33   |
| Al₂O₃        | 19.62  | 19.99  | 14.86  | 14.6   | 14.11  | 11.55  | 11.18  | 9.80   | 3.34   |
| Fe₂O₃        | 8.1    | 7.41   | 2.79   | 2.62   | 2.61   | 11.55  | 11.18  | 9.80   | 3.34   |
| MnO          | 0.13   | 0.12   | 0.03   | 0.04   | 0.04   | 0.22   | 0.19   | 0.16   | 0.06   |
| MgO          | 4.56   | 3.87   | 0.55   | 0.51   | 0.44   | 7.55   | 4.59   | 4.20   | 0.40   |
| CaO          | 10.08  | 8.54   | 1.96   | 1.98   | 1.98   | 12.19  | 9.24   | 7.56   | 1.95   |
| Na₂O         | 2.68   | 2.72   | 3.13   | 3.01   | 3.33   | 1.47   | 3.15   | 3.07   | 3.89   |
| K₂O          | 1.22   | 1.48   | 5.00   | 5.11   | 3.99   | 0.18   | 0.59   | 1.56   | 4.15   |
| P₂O₅         | 0.13   | 0.2    | 0.13   | 0.12   | 0.07   | 0.05   | 0.26   | 0.19   | 0.06   |
| Total        | 96.15  | 99.80  | 99.75  | 99.75  | 99.69  | 99.83  | 99.76  | 99.76  | 99.67  |
| Trace elements (ppm) | |
| Ni           | 20     | 15     | nd     | 0      | nd     | 10     | 9      | 17     | 2      |
| Cr           | 53     | 9      | 7      | 7      | 11     | 9      | 5      | nd     | 5      |
| V            | 193    | 110    | 15     | 14     | 18     | 408    | 234    | 158    | 10     |
| Sc           | 33     | 15     | 8      | 7      | 4      | 55     | 39     | 24     | 9      |
| Zn           | 71     | 65     | 59     | 57     | 53     | 75     | 92     | 92     | 69     |
| Ga           | 18     | 19     | 19     | 19     | 17     | 16     | 21     | 20     | 19     |
| Ba           | 302    | 456    | 1479   | 1306   | 1402   | 87     | 287    | 595    | 1610   |
| Sr           | 56     | 90     | 159    | 175    | 139    | nd     | 12     | 50     | 109    |
| Rb           | 3      | 6      | 18     | 20     | 9      | 3      | 4      | 6      | 10     |
| Sr           | 412    | 527    | 241    | 214    | 224    | 401    | 460    | 381    | 198    |
| Zr           | 66     | 63     | 237    | 220    | 203    | 27     | 116    | 115    | 344    |
| Nb           | 14     | 13     | 17     | 21     | 14     | 6      | 13     | 13     | 10     |
| Hf           | 3.8    | 2.8    | 7.0    | 7.4    | 6.5    | 2.5    | 3.7    | 4.0    | 7.0    |
| Ta           | 15     | 4      | 71     | 66     | 26     | nd     | 19     | 19     | 42     |
| Ce           | 34     | 22     | 149    | 144    | 58     | 4      | 31     | 42     | 95     |
| Pr           | 4.0    | 1.9    | 14.9   | 13.9   | 6.0    | 1.1    | 4.5    | 4.7    | 9.4    |
| Nd           | 20     | 8      | 46     | 46     | 14     | nd     | 31     | 24     | 27     |
| Y            | 27     | 17     | 20     | 21     | 19     | 10     | 33     | 28     | 27     |
| U            | nd     | 5      | 6      | 4      | nd     | 3      | 2      | 6      | 2      |
| Pb           | 4      | 9      | 23     | 22     | 16     | 6      | 5      | 8      | 16     |
| Cu           | 24     | 20     | 24     | 4      | 8      | 37     | 29     | 15     | 9      |

Note: nd—not determined.

*Sample 10JH05-2 is a duplicate analysis of 10JH05.

Mineral abbreviations after Whitney and Evans (2010): Pl—plagioclase; Bt—biotite; Hbl—hornblende; Qtz—quartz; Kfs—potassium feldspar; Cpx—clinopyroxene; Opx—orthopyroxene.
Stage 1 units include Kgp, Kg, and Kgd, which are characterized by gradational and/or crenulated contacts with one another (cf. Figs. 3C, 3D), suggesting that earlier units had not completely solidified prior to intrusion of the subsequent unit. The single stage 2 unit, Kgr, displays ambiguous contacts with stage 1 units (Fig. 3C), with both relatively sharp (solid lines in Fig. 3D) and gradational (dashed lines in Fig. 3D) contacts. Stage 3 units Kp and Kpg have gradational contacts with one another. Unit Kp crosscuts (Figs. 3A, 3B) and contains stage 1 and 2 units as angular xenoliths (Figs. 3E, 3F), which suggests that stage 1 and stage 2 units had solidified prior to intrusion of stage 3 units. We collected samples of Kgp, Kgd, Kgr, and Kp from location 4 for zircon U-Pb dating.

**Rock Classification and Petrography**

We provide the primary mineralogy and major and trace element analyses of both AMC and nearby samples in Table 1. Figure 4 displays AMC samples from all locations (cf. Fig. 2) based on major element analyses. With one exception, the Kaweah River samples (location 4; solid symbols in Fig. 4) are classified as either gabbro or granite. The exception is a sample from stage 1 unit Kgd (10JH22), which is classified as gabbrodiorite. Another sample from this unit (10JH18) is classified as gabbro, suggesting intraunit variability. Samples from locations 1, 3, 5, and 6 are displayed as open symbols and are compositionally similar to the Kaweah River samples, with the exception of a diorite from location 1. The half-filled red diamond represents the Frys Point intrusion, which was sampled for zircon U-Pb dating and is also a granite.

AMC and local gabbros (<52 wt% SiO₂) are dominated by plagioclase and varying amounts of clinopyroxene, orthopyroxene, and hornblende with minor biotite. AMC and local intermediate rocks (~54–59 wt% SiO₂) contain plagioclase, hornblende, biotite, and either clinopyroxene or quartz. AMC and local granites contain microcline, quartz, and plagioclase with minor biotite. Some samples exhibit secondary alteration, including local replacement of biotite by chlorite and seritization of feldspars.

**Zircon Morphology**

We separated and examined zircon grains using CL images for six AMC samples and the Frys Point granite; zircon morphologies for the four samples analyzed for geochronology are summarized in Table 2. CL images show that zircon grains in Kgp are dominated by sector zoning, whereas zircon grains from the other samples are dominated by concentric zoning. In granites, concentric zones become progressively narrower toward zircon grain margins. Zircon grains from granite samples often contain inherited cores, overgrowth rims, and melt inclusions.

**Zircon U-Pb dating**

U-Pb isotopic data for the five samples analyzed are documented in Supplemental Table 1 (see footnote 1). Gabbro samples from Kaweah River stage 1 units (10JH20, Kgp, and 10JH18,
Kgd) are dated as 105.1 ± 0.9 Ma and 105.5 ± 0.7 Ma, respectively (Fig. 5). We tentatively assign an age of 104.3 ± 0.5 Ma to stage 2 unit Kgr (10JH16) based on two concordant zircon U-Pb dates. Stage 3 unit Kp (10JH17) yields a date of 102.5 ± 0.7 Ma (Table 2; Fig. 5), suggesting a three-stage intrusion history for the AMC (Fig. 3). We also dated a granite sample from the Frys Point intrusion (10SQ21) as 105.5 Ma ± 0.7 Ma (Table 2; Fig. 5; location 2 in Fig. 2), coeval with Kaweah River stage 1 intrusions and slightly older than a published U-Pb age for the Frys Point granite (younger than 99 Ma; Chen and Moore, 1982). Dates for samples 10JH20, 10JH18, and 10SQ21 confirm that ca. 105 Ma magmatism involved coeval low- and high-silica magmas.

Zircon Trace Element Geochemistry

Trace element analyses of zircon grains collected with the SHRIMP-RG (reverse geometry) are listed in Supplemental Table 2.2 We used these analyses to constrain crystallization temperatures and cooling rates (e.g., Barth and Wooden, 2010). Figure 6 illustrates zircon crystallization temperatures calculated using the Ti-in-zircon thermometer of Ferry and Watson (2007) plotted against Hf concentrations, because temperature-Hf relationships in zircons have been used to assess the extent of fractionation and rate of cooling within discrete magma batches (Barth and Wooden, 2010). We plot zircon grain data from four AMC units at the Kaweah River locality (stage 1 units Kg and Kgd, stage 2 unit Kgr, and stage 3 unit Kp) and a granite of the stage 1 Frys Point pluton (Kfp) from location 2 (Fig. 2). We adjusted activities of oxides (SiO2, TiO2) in model temperature calculations for magma compositions, as recommended by Hayden and Watson (2007), but we did not apply the pressure corrections for Ti partitioning (e.g., Ferriss et al., 2008) because we use these uncorrected model temperatures solely to evaluate relative temperature differences among samples. In the following, and in Supplemental Table 2 (see footnote 2), all temperatures listed have not been corrected for pressure.

Calculated temperatures for zircons from Kg and Kgd (gabbros) exhibit the tightest clustering. Unit Kg records the highest temperatures (748–857 °C; Supplemental Table 2 [see footnote 2]) and exhibits a relatively narrow range of Hf concentration (8755–9868 ppm), while unit Kgd records similarly high temperatures (722–857 °C) and a narrow range of Hf concentration (9367–10349 ppm). Unit Kfp (granite) records lower temperatures (670–775 °C) and a wider range in Hf concentration (8055–13099 ppm). Zircon compositions from Kgr and Kp (granites) exhibit greater scatter in temperature, but the relatively wide ranges in Hf concentration from zircon in those units (8701–13511 ppm for Kgr and 9256–14985 ppm for Kp) are similar to the range displayed by zircon from the Frys Point pluton (Kfp).

Ti-in-zircon temperatures for igneous rocks are typically lower than expected for crystallization temperatures (Fu et al., 2008). However, the relative differences between model temperatures and variations in Hf contents among AMC samples and the Frys Point granite are notable. A small change in Hf content may imply a relatively fast cooling rate because rapid cooling does not allow significant fractionation to occur (Barth and Wooden, 2010). However, samples can display different Hf-temperature trends depending on source composition and magma composition in addition to cooling rate (Wooden et al., 2007; Barth and Wooden, 2010).

The relatively small change in Hf abundance (<1400 ppm) over their apparent crystallization temperature intervals (~110 °C) might suggest that the AMC gabbros (units Kg and Kgd) underwent little fractionation. In contrast, the large range of Hf abundances in zircons from units Kfp (~5000 ppm), Kg (~4800 ppm) suggests significant fractionation permitted by slower cooling rates. Minimum temperatures for Kg and Kgd overlap with apparent maximum temperatures for units Kfp and Kg (Fig. 6). Although other
factors can affect Hf-temperature trends, we suggest that rapid cooling (>1000 °C/m.y.; Giletti, 1986, 1991) of Kg and Kgd, inferred from their low degree of fractionation, may be explained as the result of interaction (quenching?) with coeval granitic magmas, a relationship exposed in outcrop (e.g., Fig. 3).

Whole-Rock Geochemistry

Major Elements

Major element data for the AMC and adjacent plutons are given in Table 1 and illustrated in Figure 7. For comparison, we have included ranges in elemental data from compositionally diverse (gabbro to granite) rocks across a swath of the SNB between lat 36°N and 37°N studied by Wenner and Coleman (2004) and from the El Capitan Granite (Ratajeski et al., 2005). We observe a large range of composition, from 48 to 77 wt% SiO₂, with a compositional gap between ~60 and 70 wt% SiO₂. Harker diagrams display generally linear and negative trends between SiO₂ and FeO*, MgO, Mg#, CaO, MnO, and TiO₂. Both K₂O and Na₂O increase with increasing SiO₂, with K₂O displaying a more linear trend. Al₂O₃ displays an overall negative trend with increasing SiO₂. P₂O₅ exhibits the greatest scatter with changes in SiO₂, generally increasing with increasing silica to ~55 wt% SiO₂, then decreasing with increasing silica above that value. Wenner and Coleman (2004) documented a similar range in silica content as well as trends on major element diagrams (Fig. 7), but their data display an unbroken continuum of compositions, without the compositional gap that we observe in our samples. We note that compositions of AMC granites are very similar to the El Capitan Granite (Ratajeski et al., 2005).

Trace Elements

Trace element data for AMC and local samples are given in Table 1 and illustrated in Figure 8. For comparison, we include Wenner and Coleman (2004) as the light blue shaded region on each diagram. Abundances of Ni, Cr, Sc, V, and Sr are negatively correlated with SiO₂. We observe large ranges in Ni, Cr, Sc, and V among the gabbros and diorites, whereas elemental concentrations are much more limited among the granites. In contrast, the granites exhibit wide ranges (from twofold to sixfold) in Ba, Rb, Cs, La, Ce, Th, Pb, Zr, Hf, Nb, and Y contents. Abundances of those elements generally increase with increasing silica, but some mafic and/or intermediate rocks overlap with some granites in Cs, La, Ce, Zr, Hf, Nb, and Y contents. Wenner and Coleman (2004) documented generally similar trends in trace elements (Fig. 8).
**Stage 1**  
Kgp  
Sample 10JH20  
Gabbro (48.4 wt % SiO₂)  

```
Mean age:  
105.1 +/- 0.9 Ma
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**Stage 1**  
Kgd  
Sample 10JH18  
Gabbro (52.0 wt % SiO₂)  

```
Mean age:  
105.5 +/- 0.7 Ma
```

**Stage 1**  
Frys Point Granite (Kfp)  
Sample 10SQ21  
Granite (72.9 wt % SiO₂)  

```
Mean age:  
105.5 +/- 0.9 Ma
```

**Stage 1**  
Kp  
Sample 10JH17  
Granite (71.3 wt % SiO₂)  

```
Mean age:  
102.5 +/- 0.7 Ma
```

Figure 5. Concordia plots of selected stage 1 and stage 3 units (Kgp, Kgd, and Kp) from the Kaweah River site. Ellipses represent 2σ error. Also shown is a concordia plot for sample 10SQ21, from the Frys Point intrusion, which is coeval with other stage 1 units.

Figure 6. Ti-in-zircon model temperatures versus Hf concentrations for zircon grains in selected Ash Mountain Complex samples. Temperatures were calculated with the Ti-in-zircon thermometer of Ferry and Watson (2007). These temperatures were corrected for reduced oxide activity (TiO₂ and SiO₂) appropriate for bulk composition as suggested by Hayden and Watson (2007), but were not corrected for Ti partitioning related to pressure (e.g., Ferriss et al., 2008). See text for discussion.
Sr, Nd, and Pb Isotopes

Isotopic data are given in Table 3 and illustrated in Figures 9–11. Initial isotope ratios for the seven samples analyzed (six AMC samples and one Kfp sample) range as follows: \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) = 0.7051–0.7083; \(^{143}\text{Nd}/^{144}\text{Nd}\) = 0.5123–0.5125; \(\varepsilon_{\text{Nd}}\) = –6.05 to –2.05; \(^{206}\text{Pb}/^{204}\text{Pb}\) (i) = 18.89–19.95; \(^{207}\text{Pb}/^{204}\text{Pb}\) (i) = 15.657–15.722; \(^{208}\text{Pb}/^{204}\text{Pb}\) (i) = 38.590–38.839.

These values are similar to those reported by Chen and Tilton (1991) and Wenner and Coleman (2004) for the western area of their traverses of the Sierra Nevada, with western and eastern defined as those rocks west or east of the Panthalassan–North American break (Fig. 1) of Kistler (1990). Kistler (1990) defined the boundary according to isotopic shifts in granitic rocks and lithologic changes in metamorphic wall rocks. Rocks from the nearby Sequoia Intrusive Suite (Fig. 1) are plotted separately from eastern and western data fields for reference (Figs. 9–11). The Sequoia Intrusive Suite is compositionally and temporally zoned (Chen and Moore, 1982; Moore and Sisson, 1987; Hensley et al., 2010); the marginal Giant Forest granodiorite transitions into the Big Meadows granodiorite (ca. 100 Ma) and is cored by high-silica Weaver Lake granite (ca. 97 Ma). Despite compositions that include both granite and gabbros, variations in \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) and \(\varepsilon_{\text{Nd}}\) (i) ratios among AMC stage 1 samples are limited (Table 3; Figs. 9 and 10). Granite samples from stages 2 and 3 have higher \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) ratios and lower \(\varepsilon_{\text{Nd}}\) (i), relative to stage 1 rocks. AMC samples appear to capture a transition from older rocks with \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) < 0.706 to younger rocks with \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) > 0.706 (Figs. 9 and 10). The typically spatially defined \(^{87}\text{Sr}/^{86}\text{Sr}\) (0.706) line, which is to the west of the exposed AMC (Fig. 1), is thus a temporally expressed feature in AMC rocks. For the Fine Gold Intrusive Suite of the western SNB, exposed between 36°N and 37°N, it was shown (Lackey et al., 2012) that the \(^{87}\text{Sr}/^{86}\text{Sr}\) (0.706) line is also a temporal feature (cf. Fig. 10) associated with a transition in magmatic style ca. 105 Ma.

Compared to data from other studies (Ratajeski et al., 2001, and references therein), stages 2 and 3 AMC rocks are along the Sr-Nd isotopic array typical of the SNB (Fig. 10), with stage 2 and stage 3 samples defining a trend toward higher \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) and lower \(\varepsilon_{\text{Nd}}\) (i). The stage 3 granite has considerably higher \(^{87}\text{Sr}/^{86}\text{Sr}\) (i) than is typical for SNB granitoids (Figs. 9 and 10).

With one exception (a stage 1 gabbro), Pb isotopic compositions of most AMC rocks form a fairly tight cluster within the field defined by the upper mantle of the Sierra Nevada (Fig. 11). This grouping is in contrast to other SNB suites, which largely scatter between the Sierran mantle field and the field defined by the western Great Basin basalts of presumed lithospheric mantle origin (Fig. 11). Compositions of most AMC rocks are within the area of overlap between compositions of samples from the western SNB and those of the Sequoia Intrusive Suite (Fig. 11). Viewed as a whole, there is no clear distinction in Pb isotopic compositions based on AMC rock type and the Pb data are largely ambiguous in terms of distinguishing between mantle versus crustal sources. The exception is gabbro sample 10JH18, which has a significantly higher \(^{206}\text{Pb}/^{204}\text{Pb}\) ratio and a slightly higher \(^{207}\text{Pb}/^{204}\text{Pb}\) ratio compared to other AMC rocks, although the \(^{87}\text{Sr}/^{86}\text{Sr}\) (0.7057) and \(\varepsilon_{\text{Nd}}\) (−2.3) for this gabbro are similar to other AMC gabbros (cf. Table 3).
Figure 8. Abundances (in ppm) of selected trace elements versus wt% SiO$_2$ in Ash Mountain Complex rocks. Symbols as in Figure 4. Light blue fields represent data for central Sierra Nevada batholith rocks documented in Wenner and Coleman (2004).

### TABLE 3. ISOTOPE GEOCHEMISTRY OF THE ASH MOUNTAIN COMPLEX

| Rock type | 10JH16 granite | 10JH17 granite | 10JH18 gabbro | 10JH20 gabbro | 10JH21 gabbro | 10JH22 gabrodiotite | 10S021 granite |
|-----------|----------------|----------------|--------------|--------------|--------------|---------------------|---------------|
| Unit      | Kgr            | Kp             | Kg           | Kg            | Kg           | Kg                  | Kg            |
| Age (Ma)  | 104.3*         | 102.5          | 105.5        | 105.1         | 105*         | 105*                | 105.5         |
| $^{87}$Sr/$^{86}$Sr$_{t,today}$ | 0.709017 ± 6   | 0.711070 ± 6   | 0.705919 ± 6 | 0.705997 ± 6 | 0.706361 ± 6 | 0.706361 ± 6       | 0.710831 ± 6  |
| $^{87}$Sr/$^{86}$Sr$_{t,0}$  | 0.707657       | 0.708296       | 0.705832     | 0.705867     | 0.705867     | 0.705867            | 0.705867      |
| $^{143}$Nd/$^{144}$Nd$_{0}$  | 0.512384 ± 5   | 0.512500 ± 4   | 0.512514 ± 4 | 0.512514 ± 4 | 0.512514 ± 4 | 0.512467 ± 3        | 0.512490 ± 4  |
| $^{147}$Sm/$^{144}$Nd$_{0}$  | 0.1027 ± 6     | 0.0914 ± 5     | 0.1670 ± 9   | 0.1716 ± 9   | 0.1693 ± 9   | 0.1454 ± 8          | 0.2108 ± 11   |
| $\varepsilon_{Nd(i)}$  | –3.70 ± 0.10   | –6.05 ± 0.07   | –2.30 ± 0.09 | –2.09 ± 0.08 | –2.05 ± 0.08 | –2.05 ± 0.08        | –3.08 ± 0.10  |
| $^{206}$Pb/$^{204}$Pb$_{t,today}$ | 19.237 ± 2  | 19.359 ± 2     | 19.952 ± 4   | 19.196 ± 2   | 19.206 ± 3   | 19.206 ± 3          | 19.206 ± 3    |
| $^{207}$Pb/$^{204}$Pb$_{t,today}$ | 15.686 ± 2  | 15.703 ± 2     | 15.722 ± 4   | 15.680 ± 2   | 15.665 ± 3   | 15.665 ± 3          | 15.665 ± 3    |
| $^{208}$Pb/$^{204}$Pb$_{t,today}$ | 38.982 ± 8  | 38.981 ± 7     | 38.905 ± 12  | 38.847 ± 8   | 38.797 ± 10  | 38.797 ± 10         | 38.797 ± 10   |
| $^{206}$Pb/$^{204}$Pb$_{t,i}$  | 19.237       | 18.978         | 19.952       | 19.143       | 19.087       | 19.087              | 19.087        |
| $^{207}$Pb/$^{204}$Pb$_{t,i}$  | 15.686       | 15.685         | 15.722       | 15.677       | 15.662       | 15.662              | 15.662        |
| $^{208}$Pb/$^{204}$Pb$_{t,i}$  | 38.982       | 38.602         | 38.539       | 38.767       | 38.723       | 38.723              | 38.723        |

Note: Analytical uncertainties are reported as ±2σ absolute standard error, indicated by variation in the last digit.

*A preliminary age of 104.3 ± 0.5 Ma for 10JH16 is based on two concordant zircon U-Pb dates and is consistent with crosscutting field relationships. Ages of 105 Ma for mafic samples 10JH21 and 10JH22 are inferred from the age for 10JH20, as field relationships indicate their coeval nature.
†Ratio indicates ratios corrected to the U-Pb age of the rock. Ratios for Sr and Pb are calculated using Rb, Sr, U, Th, and Pb concentrations analyzed by X-ray fluorescence (Table 2).
§Sm/Nd ratio determined by isotope dilution.
$^{143}$Nd/$^{144}$Nd$_{CHUR,today}$ = 0.512638 and $^{147}$Sm/$^{144}$Nd$_{CHUR,today}$ = 0.1967 (CHUR—chondritic uniform reservoir) and measured Sm/Nd ratios.
Figure 9. Initial Sr isotope compositions versus zircon U-Pb crystallization ages for Ash Mountain Complex (AMC) rocks. Red dashed curve includes AMC data; symbols as in Figure 4. Blue filled crosses represent compositions for eastern Sierra Nevada batholith (SNB) rocks (Chen and Tilton, 1991; Wenner and Coleman, 2004) and orange filled circles represent compositions from the western SNB (Chen and Tilton, 1991; Wenner and Coleman, 2004). Pink shaded field represents compositions for 14 samples of the Sequoia Intrusive Suite (Chen and Tilton, 1991; Wenner and Coleman, 2004).

DISCUSSION
Summary of Observations

AMC rocks exposed in the Kaweah River valley formed in three stages: stage 1 magmatism, ca. 105 Ma, produced granite and gabbro and/or gabbrodiorite plutons. Zircon trace element analyses suggest that the gabbros probably cooled relatively quickly, quenched by contact with cooler adjacent granites, which cooled more slowly. Stage 2 magmatism, ca. 104 Ma, and stage 3 magmatism, ca. 103 Ma, produced granitic plutons.

Major and trace element compositions of AMC and other local rocks yield trends similar to those of other central SNB granitic and gabbroic plutons, but there is a significant gap in silica composition (between ~60 and 70 wt% SiO2) observed in the AMC. Stage 1 rocks exhibit limited ranges in isotopic compositions (Figs. 9–11) that appear to be independent of rock type and suggest derivation from isotopically similar sources. The Sr and Nd isotope characteristics are consistent with a mantle-like, relatively juvenile magmatic source for stage 1 rocks. Stage 2 and stage 3 rocks have higher 87Sr/86Sr0 and lower εNd(t) compositions compared to those of stage 1, suggesting that they had a distinct source and/or a higher degree of contamination by older continental crust (Figs. 9 and 10).

Petrogenesis of AMC Magnas

Three major processes, crystal fractionation (with or without assimilation), magma mixing, and crustal anateksis, have been proposed to play a significant role in formation of SNB magmas (Bateman and Chappell, 1979; Ratajeski et al., 2001, 2005; Sisson et al., 1996, 2005; Reid et al., 1983; Frost and Mahood, 1987; Dorais et al., 1990; Wenner and Coleman, 2004; Zeng et al., 2005; Pignotta et al., 2010; Nelson et al., 2013). Each process would generate different geochemical and isotopic signatures in the resulting rocks. To evaluate which process played the most significant role in the formation of AMC magmas, we modeled expected geochemical behavior for the magmatic products of each process and compare these compositions with those of AMC rocks. In our investigation of these processes, we included samples from the Kaweah River exposure of the AMC that show the clearest crosscutting relations and include other samples of similar rocks surrounding the Frys Point intrusion (Fig. 2); these additional samples are from locations 1, 2, 3, 5, and 6 (Fig. 2). In the following discussion, AMC rocks include all of these samples, unless otherwise indicated.

Trace Element Constraints on Fractionation Processes

Fractionation processes involve systematic depletion or enrichment of elements as minerals form during cooling of a parental magma (e.g., Arth, 1976). The parental magma may be a closed system undergoing fractional crystallization, such that magma composition moves in a single direction as crystallization occurs. The parental magma may assimilate new material while crystallization is occurring, such that the composition of the assimilant and the relative rates of assimilation and crystallization control changes in magma composition.

We used trace element characteristics of AMC rocks to evaluate the roles of both Rayleigh fractional crystallization and assimilation–fractional crystallization (AFC) processes in generating intermediate to felsic compositions from mafic magma. We used the composition of the most primitive AMC gabbro (sample 10JH20, based on SiO2, and Mg#, Table 1; Fig. 7) as the parental liquid, and modeled processes until 90% crystallization was reached.

We calculated trends in incompatible trace element abundances and ratios for a variety of fractional crystallization and AFC scenarios, varying the minerals removed from the melt, the composition of the assimilant (lower crust and upper crust compositions from Rudnick and Fountain, 1995), and the relative rates of crystallization (Rc) versus assimilation (Ra; cf. Fig. 12). Mineral-melt partition coefficients (Kp values) used in the models are given in Table 4. Calculated fractional crystallization and AFC trends yield compositions that overlap with some trace element concentrations and ratios of AMC rocks (Figs. 12 and 13). For example, AFC models involving upper or lower crust provide reasonably good matches to some AMC gabbros and gabbrodiorites. By changing the parental magma to that of a gabbro with a low La/Rb ratio (La/Rb = 0.27; sample 10JH18), we would achieve a closer match to AMC gabbros and/or gabbrodiorites on the La/Rb versus Ba.

Figure 10. Initial Sr and Nd isotope compositions of Ash Mountain Complex rocks relative to other Sierra Nevada batholith (SNB) intrusive rocks. Symbols and fields as in Figures 4 and 9; U-Pb zircon ages are indicated. Orange fields indicate compositions for plutonic rocks from the Fine Gold Suite (Lackey et al., 2012). Pink filled circles represent compositions of the Sequoia Intrusive Suite. Also shown are compositional fields for other SNB plutonic rock suites as given by Ratajeski et al. (2001, and references therein), including the Tuolumne Intrusive Suite, the Lamarck Granodiorite, and the Yosemite Valley Intrusive Suite.
Figure 11. Initial $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb ratios versus $^{206}$Pb/$^{204}$Pb ratios for selected Ash Mountain Complex (AMC) samples and other Sierra Nevada batholith intrusive suites. Symbols as in Figures 4 and 9. The pink field and pink filled circles represent compositions of the Sequoia Intrusive Suite. Also shown are fields for western Great Basin basalts and Sierra Nevada upper mantle as given by Wenner and Coleman, 2004, and references therein). NHRL is the Northern Hemisphere reference line of Hart (1984).

Compositional gaps can be the result of crystal fractionation (e.g., Brophy, 1991; Dufek and Bachmann, 2010), but the lack of mingling textures in the AMC also does not support mixing processes. This is in contrast to other suites in the SNB that contain numerous occurrences of intermediate rocks and spectacular examples of mingling textures in mafic enclaves and swarms in granodiorites, interpreted as the result of interaction between mafic and felsic end members (Barbarin, 1990, 1991; Dorais et al., 1990; Wenner and Coleman, 2004).

Intermediate rocks studied by Wenner and Coleman (2004) exhibit significant ranges in trace elements such as La, Zr, Ba, Ce, Hf, Th, and Nb (Fig. 8) that could be the result of the involvement of multiple felsic mixing end members exhibiting a wide range in those elements. Both the granites studied by Wenner and Coleman (2004) and the AMC granites (Fig. 8) exhibit wide ranges in those trace elements. Although felsic magmas apparently did not mix with mafic end members to generate intermediate rocks in the AMC, AMC felsic rocks could represent potential end-member compositions involved in the origin of other SNB intermediate rocks.

**Isotopic Constraints on Mixing Processes**

We used isotopic modeling to evaluate the relative roles of mantle and crustal sources in generating AMC magmas. We started with simple mixing models because trace element models ruled out the importance of AFC processes in generating AMC granitic magmas. In addition, AFC models typically require unrealistic amounts of fractionation and assimilation as well as impractical mineral-melt distribution coefficients to account for isotope arrays in Sierran granitoid rocks (e.g., Kistler et al., 1986; Lackey et al., 2012). Studies of plutons and their adjacent wall-rock contacts show minimal local assimilation (Mills et al., 2009; Lackey et al., 2012), including contacts within the lower crust (Saleeby et al., 2008), diminishing the probable role of assimilation processes even further. Magma mixing was unlikely to have played a major role in the petrogenesis of the AMC because of the absence of textural and compositional intermediate hybrids, but source rocks could have mechanically mixed prior to melting. Thus, isotopic mixing models may provide constraints regarding how source rocks that produced stage 1 AMC magmas are related to sources that produced stage 2 and 3 magmas, which have higher Sr, and lower $\varepsilon_{Nd}$.

Our mixing models for Sr and Nd isotopes (Fig. 15) illustrate possible contributions of crust and mantle reservoirs to AMC magmas. Table 5 provides Sr-Nd data for each rese-
The lithospheric mantle value of Kistler et al. (1986) is adopted as the end-member reservoir because even the most mafic rocks in the western Sierra (e.g., olivine gabbros in the western foothills) give little indication of an asthenospheric, depleted-mantle source (Clemens-Knott, 1992). Instead, Sierran mantle contributions appear to be primarily subarc lithospheric mantle (Menzies et al., 1983; Coleman et al., 1992; Lee et al., 2000). Crustal reservoirs shown in Figure 15 include the Sierran crustal value of Kistler et al. (1986), and upper and lower crustal basement of the ca. 1.7 Ga Mojave province (Miller et al., 2000). Also shown are average isotopic compositions for pyroxenite xenoliths and plagioclase-bearing granulite xenoliths of the SNB (Mukhopadhyay and Manton, 1994; Ducea, 2002); Kings Sequence schist from a migmatite complex in the southern Sierra (Zeng et al., 2005); individual analyses of Mojave crustal rocks (Newberry and Hill, 1992; Martin and Walker, 1992); and individual analyses of a Precambrian schist (PC30a) and a Sierran crustal xenolith (Big Creek, B37) (Kistler and Fleck, 1994).

Figure 12. Trace element compositions of Ash Mountain Complex rocks and calculated trends for fractional crystallization (FC) and assimilation plus fractional crystallization (AFC) models. (A–C) Ba (ppm) versus Rb (ppm). (D–F) Zr (ppm) versus La (ppm). Symbols as in Figure 4. Mineral-melt distribution coefficients used in the models are given in Table 4. Arrows in A and D represent closed system Rayleigh FC processes involving olivine (olv), plagioclase (plg), clinopyroxene (cpx), or hornblende (hbd). Tick marks are shown for 60%, 75%, and 90% crystallization. In A, the hornblende fractionation trend is shown in red to distinguish it from the trend for plagioclase fractionation. Trends designated by arrows in B, C, E, and F are for AFC processes, calculated using the equation given by DePaolo (1981). Except for C, each AFC trend encompasses individual vectors for removal of olivine, clinopyroxene, plagioclase, or amphibole to 90% crystallization (closed system equivalent). In C, the purple shaded arrows include individual vectors for removal of olivine, clinopyroxene or plagioclase, and the red arrows represent removal of hornblende; all trends represent to 90% crystallization (closed system equivalent). The narrow and broad red arrows represent rate of assimilation/rate of crystallization (Ra/Rc) 0.1 and from 0.5 to 1.0, respectively. See legend for the composition of the assimilant (either lower or upper crust [LC, UC]; from Rudnick and Fountain, 1995) and assumed Ra/Rc.

Also shown is a mixing trend (green dashed curve) between the AMC gabbro with the lowest \(^{87}\text{Sr}/^{86}\text{Sr}\) and highest \(\varepsilon_{\text{Nd}}\) (sample 10JH18, cf. Table 3) and a lower crustal xenolith from Big Creek in the central SNB.

Stage 1 AMC rocks occupy positions along mixing lines that suggest sources containing ~40% or 20% of Sierran or Mojave lower crust, respectively (Fig. 15). The stage 2 granite also is along the mixing curve between lithospheric mantle and Mojave lower crust, but requires more crust in the source. If the lithospheric mantle was of a more enriched nature (e.g., Coleman et al., 1992), less crust would be required to generate both stage 1 and 2 magmas.
The stage 3 granite is along the mixing trend between the AMC gabbro and Big Creek lower crustal xenolith. This trend could represent mixing of magmas with the isotopic characteristics of these end members. However, with a silica content of ~71 wt%, it is unlikely that the stage 3 granite (10JH16, cf. Table 3) is an ~45:55 mix of felsic melt (with ~70–75 wt% SiO₂) and mafic magma (with ~50 wt% SiO₂). Instead of magma mixing, we suggest that this granite was derived by melting a source containing crust that is isotopically distinct from the source or sources of stage 1 and 2 magmas. Alternatively, the source could have been similar to average Sierran granulite xenoliths (cf. Fig. 15).

Previous oxygen isotope mass balance calculations show that SNB magmas typically contain 5%–20% crust (Lackey et al., 2005, 2008, 2012), with as much as 35% in some western SNB tonalites (Lackey et al., 2012). This upper limit of crustal input is similar to the 20%–30% proportion suggested by the Sr-Nd mixing trend using Mjove lower crust reservoir and less than the ~40% proportion of Sierran crust suggested for AMC stage 1 magmas. If a significant crustal component was involved in AMC magma genesis, δ⁸O crustal values would be relatively low. For example, Sierran pyroxenite xenoliths plot at ~50:50 mixture of Sierran mantle and crust. Upper crust δ¹⁸O values of 14.5‰ are based on measurements of SNB wall rocks (Lackey et al., 2006). A mixture of 50% Sierran mantle with a δ¹⁸O value of 6.0‰ (Kistler et al., 1986), and 50% upper crust with a δ¹⁸O of 14.5‰ produces δ¹⁸O values of 10.25‰ in the derived magma, higher than the observed 7.2‰ reported for pyroxenite xenoliths (Ducaze, 2002; Lackey et al., 2005). However, use of a lower crustal δ¹⁸O of 8.5‰ results in a perfect match. Considering oxygen isotope systematics in the SNB (Lackey et al., 2005, 2006, 2008, 2012), we thus propose that the dominant crustal input to AMC sources is mafic lower crust. For rocks with δ⁸⁷Sr/δ⁶⁸Sr values >0.706, such mafic lower crust has Proterozoic model ages of 1.7–1.8 Ga (e.g., DePaolo, 1981; Kistler et al., 1986; Chen and Tilton, 1991). For rocks with δ⁸⁷Sr/δ⁶⁸Sr <0.706, such mafic crust may also include portions of Paleozoic accreted terranes (e.g., Saleeby, 2011).

To summarize, the Nd and Sr isotopic characteristics of AMC rocks suggest modest (~20%–40%, perhaps less, depending on the degree of enrichment of lithospheric mantle) crustal input into ca. 105 Ma stage 1 AMC magma sources. There is little isotopic variation among stage 1 rocks, suggesting a common source, which we propose is mafic crust. Both stage 2 and 3 granites appear involve a greater crustal component compared to stage 1 granite. The stage 3 granite appears to be derived from a distinct crustal source, which was tapped after 104 Ma.

**Crustal Anatectic**

Partial melting of continental crust may be an important process in forming felsic SNB magmas (Hildreth and Moorbath, 1988; Beard and Glazner, 1995; Ducaze, 2001; Petford and Gallagher, 2001; Wenner and Coleman, 2004). This partial melting is presumably triggered by intrusion and stagnation of juvenile, mantle-derived mafic magmas (e.g., Hildreth, 1981; Bergantz, 1989). To investigate the potential role of partial melting in the production of AMC granites, we compare major element compositions of AMC granites to a wide variety of experimentally produced partial melt compositions derived from mafic to intermediate crustal sources (Fig. 16).

Natural mafic starting compositions with <1.0 wt% K₂O yield melts (gray shaded region, Fig. 16) with K₂O contents that are lower than those in AMC granites; this observation is consistent with other studies on SNB batholiths (Sisson et al., 2005; Ratajeski et al., 2005). Figure 16 also illustrates compositions of synthetic glasses produced in partial melting experiments that involve a variety of natural materials considered to be important in the generation of high-K, granitic magmas associated with continental arcs (Skjerlie and Johnston, 1993, 1996; Skjerlie and Patiño-Douce, 2002; Patiño-Douce, 2005; Ratajeski et al., 2005; Castro et al., 2010). Figure 16 summarizes the parameters of these experimental studies. Starting compositions include low-K mafic sources (Skjerlie and Patiño-Douce, 2002), medium- to high-K mafic sources (Ratajeski et al., 2005), intermediate (tonalitic to dioritic) sources (Skjerlie and Johnston, 1993, 1996; Patiño-Douce, 2005), and mechanical mixtures of mid-oceanic ridge basalt plus sediment (Castro et al., 2010). All of these experiments yield compositions that overlap those for the AMC. Some glasses (filled orange diamonds, Fig. 16) produced from the low K₂O sources of Skjerlie and Patiño-Douce (2002) are within the range of AMC compositions, but most (open orange diamonds) have compositions with significantly lower K₂O; therefore, we do not consider this latter group further.

The best potential match for the experimental partial melt granite would be the least evolved granites (LEGs), which have not undergone subsequent crystal fractionation. In an attempt to identify LEGs in the AMC, Ba and Sr abundances were calculated for Rayleigh fractional crystallization and compared to observed Ba and Sr concentrations in AMC granites. Ba and Sr are among the few trace elements whose concentrations are largely controlled by major phases (potassium feldspar, plagioclase, biotite, hornblende) rather than accessory phases (apatite, zircon), which have very large K₂O values for other trace elements (e.g., rare earth elements, Zr, Hf). Figure 17 illustrates fractionation trends assuming a range of Ba and Sr bulk distribution coefficients appropriate for granite melts (cf. Rollinson, 1993). Three of the AMC granites with high silica contents (73–77 wt%) are along

### TABLE 4. PARTITION COEFFICIENTS FOR TRACE ELEMENT MODELING

| Coefficients used in FC and AFC models* | plagioclase | clinopyroxene | hornblende | olivine |
|----------------------------------------|-------------|---------------|------------|--------|
| Ni                                     | 0.04        | 7.75          | 6.8        | 17.45  |
| Cr                                     | 0.05        | 34            | 17         | 0.7    |
| K                                      | 0.17        | 0.038         | 0.96       | 0.007  |
| Cs                                     | 0.025       | 0.01          | 0.01       | 0.001  |
| Ba                                     | 0.23        | 0.026         | 0.42       | 0.001  |
| La                                     | 0.17        | 0.052         | 0.2        | 0.001  |
| Nd                                     | 0.08        | 0.31          | 0.045      | 0.007  |
| Zr                                     | 0.048       | 0.1           | 0.5        | 0.012  |
| Hf                                     | 0.051       | 0.263         | 0.5        | 0.013  |
| Y                                      | 0.03        | 0.9           | 1.2        | 0.01   |
| Rb                                     | 1.8         | 0.66          | 0.46       | 0.014  |
| Nb                                     | 0.07        | 0.031         | 0.29       | 0.006  |
| Th                                     | 0.01        | 0.005         | 0.8        | 0.01   |
| Y                                      | 0.01        | 0.03          | 0.25       | 0.01   |

*Fractional crystallization (FC) and assimilation–fractional crystallization (AFC); illustrated in Figures 12 and 13. Illustrated in Figures 20 and 22.

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**Table 4. Partition Coefficients for Trace Element Modeling**

| Coefficients used in batch partial melting models | plagioclase | clinopyroxene | hornblende | garnet |
|--------------------------------------------------|-------------|---------------|------------|--------|
| Sr                                                | 1.840       | 0.300         | 0.720      | 0.003  |
| Y                                                 | 0.005       | 1.235         | 1.560      | 16.000 |

*Note: Most values are from the compilation of Rollinson (1993, and references therein). The exceptions are coefficients for Sr and Y in plagioclase, clinopyroxene, hornblende, garnet, and orthopyroxene, which are from Hofman et al. (2011, Supplementary Data 5, and references therein).*
Figure 13. Trace element compositions of Ash Mountain Complex rocks and calculated trends for fractional crystallization (FC) and assimilation plus fractional crystallization (AFC) models. (A–C) La/Rb versus Ba (ppm). (D–F) Zr/Nb versus La (ppm). Symbols as in Figure 4. Mineral-melt distribution coefficients used in the models are given in Table 4. Analytical uncertainties for each trace element or ratio are represented by error bars. Gray shaded arrows in A and D labeled FC represent closed system FC processes involving olivine, plagioclase, or clinopyroxene. The length of the arrow represents as much as 90% crystallization. The red arrows in A, C, and D represent hornblende fractionation with the lengths representing as much as 90% crystallization. Trends designated by arrows in B, C, E, and F are for AFC processes, calculated using the equation given by DePaolo (1981). Each AFC trend encompasses individual vectors for removal of olivine, clinopyroxene, plagioclase, or amphibole to 90% crystallization (closed system equivalent). See legend for the composition of the assimilant (either lower or upper crust [LC, UC]; from Rudnick and Fountain, 1995) and rate of assimilation/rate of crystallization (Ra/Rc).

Figure 14. Selected incompatible trace element ratios plotted versus wt% SiO2 for Ash Mountain Complex intrusive rocks. Analytical uncertainties are represented either by the size of the symbol or error bars.
Precambrian schist (PC30a) 164(6) 47(6) 0.7383(6) –15.6(6)

Average Sierran granulite xenolith 363(3,4) 17(3,4) 0.70827(3,4)  –6.8(3,4)

Average Sierran pyroxenite xenolith 91(3)  7(3) 0.70569(3)  –6.0(3)

Sierran lithospheric mantle 555(1) 24(1) 0.7047(1) 1.0(1)

Mukhopadhyay and Manton (1994), and Kings Sequence schist (reservoir G, Zeng et al., 2005).

Composition of Sierran pyroxenite and granulite xenoliths (reservoirs E and F, Ducea, 2002; lower crustal xenolith (reservoir H, Kistler and Fleck, 1994). Colored spheres represent average reservoirs J, Martin and Walker, 1992). Also shown in the main figure is a mixing curve (green curve denoted by dashed black rectangle in inset). The gray curve represents mixing with typical Ash Mountain Complex rocks. Mixing curves from the inset are also shown in the main figure (Fig. 16). The blue and black curves represent mixing positions of the derived melts (e.g., Martin, 1987; Drummond and Defant, 1990; Rapp et al., 1991; Johnson et al., 1997; Hoffmann et al., 2011).

For example, melts derived from garnet-bearing mafic sources have low Y but high Sr/Y values because of the high KD for Y (~16; Martin, 1987) and moderate KD for Sr (~0.01; Martin, 1987) for garnet. In contrast, melts derived from garnet-free amphibolite sources have low Sr/Y because of residual plagioclase, which has a low KD for Y (~2; Martin, 1987). In the experiments used in our evaluation of AMC LEGs (Fig. 16), those run at pressures higher than 10 kbar include garnet in the residue, whereas lower pressure experiments do not.

To evaluate the role of residual source compositions on Sr and Y abundances in AMC LEGs, these elements were modeled using the batch melting model of Shaw (1970). Sr and Y abundances in partial melts produced in five of the six experiments listed in Figure 16 were calculated using mineral modes in the residue and percentage of coexisting glass. The exception is the study of Skjerlie and Patiño Douce (2002; see following). Table 4 gives KD values of Sr and Y abundances in AMC LEGs. All data sets overlap the compositions of AMC rocks to varying degrees. Given uncertainties in synthetic glass compositions (cf. Fig. 18), these data do not compel us to rule out any particular set of experiments. However, with the exception of TiO₂, the melts generated by Ratajeski et al. (2005) appear to be the best fit.

The compositions of these glasses have the most overlap with AMC granites and exhibit similar decreases in Al₂O₃ with increasing silica, as observed for AMC granites (Fig. 19).

Although the melts generated by Ratajeski et al. (2005) display slightly higher Al₂O₃ values for given SiO₂ values, the majority of synthetic glasses from other experiments display significantly higher Al₂O₃ abundances than AMC granites.

During anatexis, the residual source mineralogy strongly influences the trace element compositions of the derived melts (e.g., Martin, 1987; Johnson and Defant, 1990; Rapp et al., 1991; Johnson et al., 1997; Hoffmann et al., 2011). For example, melts derived from garnet-bearing mafic sources have low Y but high Sr/Y values because of the high KD for Y (~16; Martin, 1987) and low KD for Sr (~0.01; Martin, 1987) for garnet. In contrast, melts derived from garnet-free amphibolite sources have low Sr/Y because of residual plagioclase, which has a low KD for Y (~2; Martin, 1987). In the experiments used in our evaluation of AMC LEGs (Fig. 16), those run at pressures higher than 10 kbar include garnet in the residue, whereas lower pressure experiments do not.

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of Sr and Y values seemingly appropriate to the starting material (cf. Table 7). We focused on experimental runs that produced partial melts that most closely resemble major element compositions of AMC granites. For example, synthetic glasses produced by Skjerlie and Johnston (1993) with silica contents >73 wt% were not included in Sr-Y models. Some experimental residual mineral modes include accessory ilmenite, titanomagnetite, apatite, spinel, corundum, rutile, kyanite, zoisite, phengite, and/or a sulphide phase. We assumed that Sr and Y are largely controlled by the major residual minerals (especially garnet, amphibole, and plagioclase) and did not include accessory phases in Sr-Y models (cf. Table 6).

Figure 20 illustrates modeled Sr and Y compositions for melts derived in experiments along with observed Sr and Y compositions for AMC LEGs. Residual mineralogies that contain garnet but no amphibole are unlikely sources for AMC LEGs (Figs. 20A, 20B). Although some melts produced in experiments conducted by Skjerlie and Patiño-Douce (2002) share broadly similar major element compositions with LEGs (Figs. 18 and 19), we did not model Sr-Y in those melts because residual mineral modes and glass percentages were not documented. Sources in the Skjerlie and Patiño-Douce (2002)
experiments contain garnet, thus we assume that the derived melts would not share the low Sr/Y values characteristic of AMC LEGs.

Melts resulting from very high degrees of melting (~50%–80%; Castro et al., 2010) have Sr-Y characteristics similar to those of AMC LEGs (Fig. 20C). However, such high degrees of melting seem unlikely as melt segregation and migration appear to occur at melt fractions of ~26–40 vol% (Sawyer, 1994, and references therein) or even lower volumes (Rushmer, 1996).

Sources that contain neither garnet nor amphibole have low Sr/Y compared to AMC LEGs, although the Y contents are much higher (Fig. 20E). In order to keep Y contents low in resulting melts, it appears that amphibole, which has a $K_d$ for Y of ~1.6 (cf. Table 4), is required in the source.

Melts produced in experiments that include amphibole but no garnet in the source (Ratajeski et al. (2005))
| Reference                      | Model name       | Glass (%) | Quartz | Grit | Plagioclase | Amphibole | Clinopyroxene | Orthopyroxene | Biotite | Kfs |
|-------------------------------|------------------|-----------|--------|------|-------------|-----------|---------------|--------------|---------|----|
| Castro et al. (2010)          | Cas, Gneiss, 1   | 40        | 54.5   | 45.5 |             |           |               |              |         |    |
|                               | Cas, Gneiss, 2   | 50        | 55.6   | 11.1 | 33.3        |           |               |              |         |    |
|                               | Cas, Gneiss, 3   | 82        | 66.7   | 33.3 |             |           |               |              |         |    |
|                               | Cas, Gneiss, 4   | 60        | 64.7   | 35.3 |             |           |               |              |         |    |
|                               | Mel 0.5 A        | 40        | 34.5   | 34.5 | 8.6         | 17.2      | 5.2           |              |         |    |
|                               | Mel 0.5 B        | 50        | 40.0   |      |             | 50.0      | 10.0          |              |         |    |
|                               | Mel 0.25         | 60        | 12.5   | 25.0 | 62.5        |           |               |              |         |    |
|                               | Mel 0.75         | 70        | 66.7   | 16.7 | 16.7        |           |               |              |         |    |
| Ratajewski et al. (2005)      | 87S35a, 1591     | 12        | 51.2   |      | 46.5        | 2.3       | V             |              |         |    |
|                               | 87S35a, 1611     | 12        | 55.3   |      | 43.5        | 1.2       | V             |              |         |    |
|                               | 87S35a, 1602     | 15        | 54.9   |      | 45.1        |           | V             |              |         |    |
|                               | 87S35a, 1694     | 12        | 58.3   |      | 41.7        |           | V             |              |         |    |
|                               | 87S35a, 1619     | 12        | 61.4   |      | 38.6        |           | V             |              |         |    |
|                               | 87S35a, 1624     | 10        | 61.9   |      | 36.9        | 1.2       | V             |              |         |    |
|                               | YOS-55a, 1635    | 22        | 51.3   |      | 40.8        | 7.9       | VI            |              |         |    |
|                               | YOS-55a, 1642    | 22        | 46.1   |      | 48.7        | 5.3       | VI            |              |         |    |
|                               | YOS-55a, 1644    | 25        | 45.9   |      | 41.9        | 5.4       | 6.8           |              |         |    |
|                               | YOS-55a, 1673    | 20        | 47.4   |      | 47.4        | 1.4       | 2.8           |              |         |    |
|                               | YOS-55a, 1640    | 20        | 0.3    |      | 53.3        | 34.2      | 6.8           | 5.5          |        |    |
| Skjerlie and Johnston (1993)  | SKJ 1            | 26        | 27.9   |      | 55.7        | 12.9      | 3.6           |              |         |    |
|                               | SKJ 2            | 29        | 28.4   |      | 57.5        | 12.7      | 1.5           |              |         |    |
|                               | SKJ 3            | 30        | 28.0   |      | 57.6        | 14.4      | V             | VI, VII, VIII, IX, X | 20–35 |
| Skjerlie and Johnston (1996)  | S&J 1            | 9         | 13.3   |      | 4.4         | 51.1      | 31.1          |              |         |    |
|                               | S&J 2            | 19        | 17.3   |      | 7.4         | 46.9      | 21.0          | 7.4          |         |    |
|                               | S&J 3            | 34        | 7.8    |      | 15.6        | 46.9      | 29.7          |              |         |    |
|                               | S&J 4            | 30        | 7.2    |      | 30.4        | 46.4      | 15.9          |              |         |    |
|                               | S&J 5            | 22        | 11.7   |      | 18.2        | 39.0      | 20.8          | 10.4         |         |    |
|                               | S&J 6            | 25        | 12.3   |      | 24.7        | 37.0      | 26.0          |              |         |    |
|                               | S&J 7            | 26        | 13.5   |      | 21.6        | 35.1      | 29.7          |              |         |    |
|                               | S&J 8            | 29        | 6.9    |      | 25.0        | 38.9      | 29.2          |              |         |    |
|                               | S&J 9            | 40        | 1.7    |      | 43.1        | 36.2      | 19.0          |              |         |    |
|                               | S&J 10           | 12        | 18.4   |      | 26.4        | 29.9      | 25.3          |              |         |    |
|                               | S&J 11           | 18        | 16.0   |      | 33.3        | 24.7      | 25.9          |              |         |    |
|                               | S&J 12           | 23        | 14.3   |      | 33.8        | 28.6      | 23.4          |              |         |    |
|                               | S&J 13           | 34        | 7.5    |      | 28.4        | 44.8      | 19.4          |              |         |    |

(continued)
et al., 2005; cf. Fig. 20D) provide the best fit for AMC LEGs. Ratajeski et al. (2005) and Sisson et al. (2005) conducted experiments on biotite-hornblende gabbros from the central SNB, which they considered to be reasonable analogs of Sierran lower crust. Ratajeski et al. (2005) found that melts from these sources have major element compositions similar to those of the El Capitan Granite in the central part of the SNB. Ratajeski et al. (2005) calculated trace element abundances (REEs, Rb, Ba, Sr, Th, U) in partial melts produced in their experiments using trace element concentrations in the starting compositions, run-product modal data, and published partition coefficients (or other predictive expressions); they found good matches between their calculated trace element abundances and the observed abundances in the El Capitan Granite (Fig. 21). Both major and trace element compositions of AMC LEGs are remarkably similar to those of the El Capitan Granite (Figs. 7 and 21). Using parameters given in Table 6 for experimental melts of Ratajeski et al. (2005) and Sr and Y abundances for one of their starting materials (Table 7), modeled Sr and Y abundances are close fits to AMC LEGs (Fig. 19D). Excellent matches are achieved (Fig. 22) when somewhat lower Sr (250–300 ppm) contents are used in the models. These Sr abundances are similar to estimates for lower to middle continental crust (see the EarthRef.org Geochemical Earth Reference Model database: http://earthref.org/GERM).

The majority of apparent Ti-in-zircon temperatures calculated for AMC granites range from ~625 to ~775 °C (cf. Fig. 6). Among the partial melt experiments examined here (Fig. 16), those of Ratajeski et al. (2005) involve the lowest range in temperatures (825–850 °C), which is ~50–225 °C higher than calculated zircon temperatures. Other experiments examined here (Fig. 16) involve higher temperatures, ranging from 875 to 1150 °C. Although calculated zircon temperatures may be lower than actual crystallization (and magmatic) temperatures (Fu et al., 2008), we believe that the Ratajeski et al. (2005) experiments are the best analogues, assuming that crustal anatexis was the dominant process responsible in generating AMC LEGs.

In summary, if AMC felsic magmas are the result of crustal anatexis, the most plausible sources are amphibole-bearing assemblages that underwent moderate (10%–30%) degrees of melting at temperatures <850 °C. Sr-Y models indicate that garnet was not a residual phase in the source, thus restricting melting to depths <~30 km (cf. Fig. 2 in Saleeby et al., 2003, and references therein). Although AMC sources might have had lower Sr and Y contents than SNB gabbros considered to be analogs of...
Sierran lower crust (Ratajeski et al., 2005; Sisson et al., 2005), they otherwise share similar major element and other trace element characteristics.

**Transition ca. 105 Ma**

The increasing crustal input to AMC magmas after 105 Ma appears to coincide with a time when the Sierran arc was undergoing a significant increase in magmatic activity, as recorded by distributions of U-Pb zircon age and pluton area (Lackey et al., 2008). This period also corresponds to an apparent increase in the volume or preservation of 105–102 Ma rhyolites and dacites in pendants and septa at Ash Peaks, Redwood Mountain, and Boyden Cave in the Sequoia–Kings Canyon region (Moore and Sisson, 1987; Saleeby et al., 1990), the Erskine Canyon sequence in the southern Sierra (Busby-Spera and Saleeby, 1990; Saleeby et al., 2008). Thus, this period appears to record major redistribution of faulting in the batholith. Such a change in deformation style aligns with the proposal by Matthews et al. (2012) that a major global tectonic reorganization occurred between 105 and 100 Ma. We hypothesize that the magmatic and deformational changes during this period are linked in the SNB and that, within the AMC, the transition from stage 1 to stage 2 and stage 3 records a tectonically induced shift in magmatic source. The greater crustal participation in the production of stage 2 and 3 magmas is possibly related to increased melting of existing crust or melting of crust at higher levels in the arc. This shift may record insertion of fertile crust into source regions by underthrusting (e.g., Ducea and Barton, 2007). Whereas a 25–50-m.y.-cycle of

**TABLE 7. Sr AND Y CONTENTS USED IN CRUSTAL MELTING MODELS**

| Starting composition | Source | Sr (ppm) | Y (ppm) |
|----------------------|--------|----------|---------|
| I                    | Castro et al. (2010), Table 1 therein, Bt-gneiss | 131.8 | 33.1 |
| II                   | Castro et al. (2010), Table 1 therein, Mel 2 | 170.1 | 28.3 |
| III                  | Castro et al. (2010), Table 1 therein, Mel 1 | 189.3 | 26.0 |
| IV                   | Castro et al. (2010), Table 1 therein, Mel 3 | 151  | 30.7 |
| V                    | Sisson et al. (1996), Table 1 therein, sample 87S35a | 850 | 22.0 |
| VI                   | Sisson et al. (1996), Table 1 therein, sample 87S35a | 457 | 16.0 |
| VII                  | Hoffman et al. (2011), Supplementary Data 2; tonalite, 468001 | 227  | 9.64 |
| VIII                 | Lackey et al. (2012), Table 2 therein; representative range of Bass Lake Tonalite (Blt) compositions | 630  | 18.0 |

**Note:** The starting compositions I–X represent various crustal sources used in partial melting models involving trace elements (Sr and Y) illustrated in Figures 20 and 22.

Figure 20. Sr/Y versus Y (ppm) contents in Ash Mountain Complex least evolved granites (LEGs; red open and filled diamonds). Each panel shows calculated Sr-Y compositions for partial melts produced in experiments (cf. Fig. 16). Table 6 provides details of the parameters (residual mineral assemblages, degree of melting) assumed in the model calculations. Table 7 gives Sr and Y contents assumed for initial starting compositions. Each panel indicates the residual mineralogy. Note the change in scale for Y contents in E.
magmatic reinvigoration is proposed to explain the magmatic flare-ups in the Triassic, Jurassic, and Cretaceous (DeCelles et al., 2009), we suggest that similar recharge of magma sources by crustal additions from the forearc or retroarc may have operated on shorter timescales. We recognize that our isotopic and age-dating sampling is limited, and emphasize that these ideas should be tested further.

CONCLUSIONS

Based on field relationships and zircon U-Pb analyses of rocks within the AMC, we propose that the complex formed in three stages of intrusion dominated by mafic and felsic compositions. Stage 1 magmatism occurred ca. 105 Ma and consisted of two gabbros, a gabbro/gabbrodiorite, and a granite. Stages 2 and 3 magmatism occurred ca. 104 Ma and ca. 103 Ma, respectively, and consisted of granites. Calculated Ti-in-zircon temperatures and Hf fractionation data support a model where contact between cooler granitic magmas likely caused relatively rapid cooling of hotter mafic magmas, preventing significant fractionation of the mafic magmas, while granitic magmas cooled more slowly, permitting more fractionation.

Major and trace element data from the AMC are broadly similar to those of other studies of the SNB, suggesting that the AMC is an excellent locality to examine the petrogenesis of coeval mafic and felsic magmas. Our trace element variations among AMC gabbros and gabbrodiorite. However, fractionation models cannot account for the high concentrations or wide ranges of incompatible trace element concentrations in AMC felsic rocks (Figs. 12-14).

Crustal anatexis of an amphibole-bearing, mafic crustal source appears to best explain the observed major and trace element and isotopic compositions of AMC granitic rocks. Among the partial melting experiments considered here (Fig. 16), the experimentally synthesized glasses of Ratajeski et al. (2005), derived by melting of biotite-hornblende gabbros, best fit the measured major element compositions of AMC felsic rocks (Figs. 18 and 19). The temperatures of the Ratajeski et al. (2005) experiments (825–850 °C) are the closest fits to the majority of apparent Ti-in-zircon temperatures calculated for AMC granites (~625–775 °C). The pressure used in these experiments (7 kbar) results in an amphibolitic, garnet-free residual source mineralogy. Modeled Sr-Y variations in liquids derived from this source mineralogy are similar to those observed in AMC LEGs. We thus hypothesize that crustal anatexis took place at depths ~<30 km.

If AMC granites were derived by melting of mantle-derived mafic crust, they should exhibit mantle-like isotopic characteristics. We suggest that the low variability in isotopic compositions of stage 1 gabbros and granites [87Sr/86Sr(i) = 0.7051–0.7059; εNd(i) = –3.1 to –2.1] supports this hypothesis. The more evolved isotopic compositions of the younger stage 2 and 3 granites [87Sr/86Sr(i) = 0.7064–0.7080; εNd(i) = –6.1 to –3.7] indicate a larger input from evolved crust (Fig. 15). The AMC rocks thus appear to capture a temporal transition between the ca. 105 Ma emplacement of stage 1 rocks [87Sr/86Sr(i) < 0.7060] with more juvenile sources and the post–105 Ma emplacement of stage 2 and 3 rocks [87Sr/86Sr(i) > 0.7060] with greater involvement of older continental crust.

The AMC provides an example of coeval mafic and felsic magmatism involving different sources at different levels within the lithosphere. The brief timing and change of isotopic compositions of stage 1 relative to stage 2 and 3 units suggest that these magmas might represent a transition in sources beneath the SNB, from a more primitive mantle source to a source region involving a greater crustal component, potentially related to intra-arc compression (e.g., Ducea and Barton, 2007). We suggest that the compositional diversity of the AMC can be produced through partial melting of crustal and mantle sources, with fractionation and assimilation playing relatively minor roles. We also suggest that mafic and felsic compositions of the AMC could represent end members of mixing processes, which may have generated the intermediate magmas that dominate much of the SNB.

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