Use of trace element abundances in augite and hornblende to determine the size, connectivity, timing, and evolution of magma batches in a tilted batholith

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ABSTRACT

The tilted Wooley Creek batholith (Klamath Mountains, California, USA) consists of three main zones. Field and textural relationships in the older lower zone suggest batch-wise emplacement. However, compositions of augite from individual samples plot along individually distinct fractionation trends, confirming emplacement as magma batches that did not interact extensively. The younger upper zone is upwardly zoned from tonalite to granite. Major and trace element compositions of hornblende show similar variations from sample to sample, indicating growth from a single magma batch that was homogenized by convection and then evolved via upward percolation of interstitial melt. Highly porphyritic dacitic roof dikes, the hornblende compositions of which match those of upper zone rocks, demonstrate that the upper zone was eruptible. The central zone contains rocks of both lower and upper zone age, although in most samples hornblende compositions match those of the upper zone. The zone is rich in symplutonic dikes and mafic magmatic enclaves. These features indicate that the central zone was a broad transition zone between upper and lower parts of the batholith and preserves part of the feeder system to the upper zone. Homogenization of the upper zone was probably triggered by the arrival of mafic magma in the central zone. Continued emplacement of mafic magmas may have provided heat that permitted differentiation of the upper zone magma by upward melt percolation. This study illustrates the potential for use of trace element compositions and variation in rock-forming minerals to identify individual magma batches, assess interactions between them, and characterize magmatic processes.

INTRODUCTION

The ways in which plutons are assembled are currently the subject of vigorous debate. The occurrence of large volumes of volcanic rocks ejected during single eruptive events (e.g., Ritchey, 1980; Bacon and Druitt, 1988; Lipman et al., 1997; Bachmann et al., 2002, 2005; Christiansen, 2005; Hildreth, 2004; Christiansen and McCurry, 2008) suggests that large volumes of magma are stored in the middle to upper crust. Recent high-precision U-Pb dating of zircon indicates that it is not uncommon for intrusive systems to grow over several millions of years (e.g., Glazner et al., 2004; Matzel et al., 2006; Grunder et al., 2008; Memeti et al., 2010). Thermal modeling, however, indicates that even large batches of magma should not remain above their solidus for >1 m.y. (Glazner et al., 2004; Paterson et al., 2011). These modeling results have led to emplacement models in which plutons form in increments that do not interact extensively with one another, and have also been interpreted to indicate that plutons are not necessarily related to volcanic rocks (Coleman et al., 2004; Mills and Coleman, 2010). These emplacement models also imply that important processes in the chemical evolution of magmatic suites such as fractional crystallization, assimilation, and magma mixing (e.g., De Paolo, 1981; Bohrson and Spera, 2007; Ohba et al., 2007; Claiborne et al., 2010; McLeod et al., 2010; du Bray et al., 2011) are restricted to the lower part of the crust (Ammen, 2009; Mills and Coleman, 2010; Tappa et al., 2011). Incremental emplacement of batholith-scale intrusions has been demonstrated through high-precision geochronology (Glazner et al., 2004; Matzel et al., 2006; Walker et al., 2007; Grunder et al., 2008; Memeti et al., 2010; Miller et al., 2011) in cases where emplacement spanned hundreds of thousands to millions of years. However, some batholith-scale plutons may be emplaced in time scales of <1 m.y. (Miller et al., 2011; Coint et al., 2013). In such cases, alternative methods must be used to identify magma batches and assess their size and evolution.

High-temperature mafic minerals crystallize early in magmatic systems and are capable of incorporating trace elements in abundances large enough to be measured in situ. These minerals record the evolution of the melt composition in magmatic systems and thus have great potential to provide information that allows reconstruction of the history of intrusive and extrusive complexes. In the case of calc-alkaline magmas, augite and hornblende are of particular interest because they can crystallize at or close to the liquidus, depending on the amount of water present in the system (Piwinski, 1968; Eggler and Burnham, 1973), and because they occur throughout a wide range of compositions, from basaltic andesite to rhyolite. By studying the compositions of these minerals, we can also track the chemical evolution of the melt, which is generally not accessible when working with bulk-rock data on intrusive rocks because many plutonic rocks are partial cumulates (e.g., Deering and Bachmann, 2010).

In this study we utilize major and trace element abundances in augite and hornblende to reconstruct the assembly of a tilted calc-alkaline pluton, the Wooley Creek batholith (WCB; Barnes et al., 1986b; Barnes, 1987). Tilting and subsequent erosion have exposed ~9 km of structural relief through the intrusion, making it a good candidate for study of the organization of intrusive bodies. Furthermore, the presence of roof dikes in the structurally highest part of the system provides samples of magma that escaped the system (Barnes et al., 1986a), enabling us to address the problem of the connection between volcanic and plutonic rocks.

GEOLOGICAL SETTING

Klamath Mountains

The WCB is situated in northern California, USA, in the Klamath Mountains geologic province. The province consists of a series of north-south–oriented accreted terranes bounded by regional east-dipping thrust faults, resulting in preservation of a record of more than 400 m.y. of active subduction along the western North American margin (Smoke and Barnes, 2006).

The WCB is one of a series of plutons (Wooley Creek suite) emplaced from Middle to Late
Jurassic time (167–156 Ma) into rocks of the western Paleozoic and Triassic belt, central metamorphic belt, and eastern Klamath belt (Allen and Barnes, 2006). The Slinkard pluton, which crops out northeast of the WCb, is related to WCb magmatism (Barnes et al., 1986a). However, extensive subsolidus recrystallization prevented us from including the Slinkard pluton in this study. At the time of the plutonic event, the area was undergoing extension associated with formation of the backarc Josephine ophiolite ca. 159–164 Ma (Harper et al., 1994; Hacker et al., 1995). The Wooley Creek suite was emplaced east (modern coordinates) of the Josephine ophiolite. At the same time, subduction-related magmatism was active west of the Josephine ophiolite and is now represented by the Chetco plutonic complex and Rogue Formation. The Chetco-Rogue arc, the Josephine ophiolite, and its cover sequence, the Galice Formation, form the western Jurassic belt, which was thrust underneath the western Paleozoic and Triassic belt along the Orleans thrust (Harper et al., 1994) during the Neva- dan orogeny (153–150 Ma; Allen and Barnes, 2006). This thrusting truncated the base of at least some Wooley Creek suite plutons (Barnes, 1982, 1983; Jachens et al., 1986). Exhumation of high-pressure rocks of the Condrey Mountain Schist through a structural window tilted the WCb ~15°–30° toward the southwest (Barnes et al., 1986b), and subsequent erosion has resulted in ~9 km of structural relief through the pluton.

WCb

The WCb (Fig. 1) was emplaced between 159 and 155 Ma (Coint et al., 2013). The batholith intrudes three host terranes of the western Paleozoic and Triassic belt; from structurally higher to lower, these are (1) the eastern Hayfork terrane, a chert-argillite mélangé, (2) the western Hayfork terrane, a volcanioclastic sandstone and argillite unit, and (3) the Rattlesnake Creek terrane, a serpentinite matrix to block-on-block ophiolitic mélangé overlain by a coherent cover sequence (Wright and Wyld, 1994). The WCb can be divided into three zones, all of which display gradational contacts (Fig. 1). The lower zone is composed primarily of biotite ± hornblende two-pyroxene diorite and/or gabbro and biotite ± hornblende two-pyroxene tonalite. Two U-Pb (zircon, chemical abrasion–thermal ionization mass spectrometry, CA-TIMS) ages from this unit are identical with analytical uncertainty at 158.99 ± 0.17 Ma and 159.22 ± 0.10 Ma (Coint et al., 2013). Lower zone rocks are heterogeneous at the scale of the outcrop and locally display sheet-like organization with variable amounts of pyroxene. Numerous pyroxene-
rich blocks and dikes (pyroxenite and melagabbro), showing variably angular and gradational contacts with the host magmatic rock, as well as sparse mafic magmatic enclaves (MME), mafic dikes, and apyrite dikes and pods, are present in the lower part of the intrusion. Foliation is primarily magmatic, is oriented north-south, and dips steeply to the east or west (Coint et al., 2013) except along the northeastern contact, where tonalitic rocks display a protoclastic foliation that dips northeast, subparallel to the contact. Because deformation of these latter rocks was subsolidus, it is unlikely that the minerals preserve their igneous composition; therefore, these rocks were excluded from this study.

The central zone (Fig. 1) is mainly composed of biotite hornblende quartz diorite and tonalite; some samples contain scraps of pyroxene surrounded by hornblende. Detailed mapping in the Cuddihy Lakes basin (Fig. 1) shows that the zone consists of multiple decimeter- to meter-scale sheets of tonalite and quartz diorite (with or without MME) along with variably deformed and disrupted mafic symphytic dikes, including apyrites (Barnes et al., 1986a; Leopold and Yoshinobu, 2010; fig. 3F in Coint et al., 2013). In outcrop some sheets display characteristics of upper zone rocks and others of lower zone rocks. These distinctions are reinforced by U-Pb (zircon) ages of two central zone samples (Coint et al., 2013). The age of one sample (sample Z5; 159.01 ± 0.20 Ma) is identical within uncertainty to lower zone samples, whereas the age of the other (sample WCB-4909; 158.30 ± 0.16 Ma) is identical to samples from the upper zone (see following). The presence of samples with both upper and lower zone characteristics is reflected in the boundaries of the central zone with the two adjacent zones. Ridge traverses indicate that these boundaries are gradational and diffuse over distances of at least 500 m. Foliation in the central zone is magmatic and has variable strike and dip (Coint et al., 2013).

The upper zone ranges from structurally lower biotite hornblende quartz diorite and/or tonalite to structurally highest biotite hornblende granite (Fig. 1), and three samples have U-Pb zircon ages ranging from 158.21 ± 0.17 Ma to 158.25 ± 0.46 Ma (Coint et al., 2013). Outcrops in the upper zone are relatively homogeneous and repeated traverses along ridges and in cirque basins over a 20-yr period have failed to identify internal intrusive contacts. Swarms of MME are locally present and are most common near the contact with the central zone and along the southwestern intrusive contact. Magmatic foliation is typically weak and forms a broadly concentric pattern that is subparallel to intrusive contacts except east of Ten Bear Mountain, where the approximate east-west strike of foliation is nearly perpendicular to the intrusive contact (Coint et al., 2013). The southern part of the upper zone is intruded by a small (2 km × 0.7 km) body of hornblende biotite granite (Fig. 1) with a U-Pb (zircon) age of ca. 156 Ma (Coint et al., 2013). The majority of this unit is finer grained than the surrounding upper zone rocks, making the sharp intrusive contacts readily apparent in the field.

Selvages of two-pyroxene ± hornblende diorite and quartz diorite occur locally along the western and southern contacts of the upper zone (Fig. 1), and have U-Pb (zircon) ages of 159.28 ± 0.17 Ma and 158.32 ± 0.32 Ma, respectively (Coint et al., 2013). The contact between the southern mafic selvage and the granodiorite is lobate (fig. 7 in Coint et al., 2013). No sharp contact between the western mafic selvages and the granitic to granodioritic rocks of the upper WCB was observed.

Along the western and southwestern contact zone, a series of porphyritic dikes intrudes the host rocks and the structurally highest parts of the upper zone. Andesite dikes generally contain phenocrysts of augite, enstatite (herein orthopyroxene), and plagioclase, but some contain augite and hornblende. Dacitic dikes have phenocrysts of plagioclase and hornblende ± biotite ± quartz. Crosscutting relationships suggest that some andesite dikes were emplaced before the replacement of the upper WCB magma, because the dikes are cut by the upper granodiorite, whereas some andesitic dikes intrude the upper WCB (Barnes et al., 1986a). In contrast, dacitic dikes that cut the upper zone have not been observed. In Barnes (1987) and Barnes et al. (1990), the andesite dikes were interpreted to be derived from the lower zone, whereas the dacitic dikes were interpreted to be from the upper zone. These interpretations are tested (see following) on the basis of mineral compositions.

**Petrography**

**Lower Zone**

Other than the pyroxenites and melagabbros, rocks from the lower zone range from biotite two-pyroxene diorite and/or gabbro to biotite hornblende quartz diorite and tonalite. They display hypidiomorphic granular texture and are more or less porphyritic, with subbedal augite phenocrysts (Fig. 2A). Augite, orthopyroxene, and plagioclase are euhedral and define a weak magmatic foliation. Plagioclase compositions vary between An$_{50}$ and An$_{30}$, with normal to oscillatory normal zoning (Barnes, 1987; this paper). Biotite is subbedal and commonly shows a reaction relationship with pyroxene.

Hornblende is present in some samples as a late mafic magmatic phase, growing in continuity with pyroxenes (Fig. 2A). Quartz, where present, is interstitial. Accessory minerals are apatite, zircon, allanite, and rarely tourmaline. Actinolite, chlorite, epidote, and sericite are secondary minerals, which are variable in abundance. Where epidote displays sharp contacts with hornblende, it is interpreted as being magmatic, whereas where present as an anhedral phase associated with sericitized plagioclase and chloritized biotite, it is secondary.

Pyroxenite and melagabbro blocks and dikes-like sheets are coarse grained and have an orthocumulate texture (Fig. 2B). These rocks consist of subbedal exsolved augite and orthopyroxene, with or without olivine or olivine pseudomorphs. Pale green to brown amphibole is present as poikilitic crystals and in some samples is optically continuous with pyroxene. Hercynite is present as an accessory mineral. Talc, serpentine, and green actinolite are secondary minerals. Plagioclase (An$_{50}$ to An$_{30}$) and quartz are interstitial and occur in veins that brecciate the pyroxenite blocks.

**Central Zone**

Rocks in the central zone range from biotite hornblende quartz diorite to biotite hornblende tonalite. Although these samples contain similar mineral assemblages, they display a range of grain size and magmatic foliation intensity. Pyroxenites are rarely preserved as scarpby inclusions in hornblende (Fig. 2D) and some relic grains are evident as intergrown actinolitic hornblende and quartz blebs surrounded by magmatic hornblende. Hornblende habits vary from euhedral prismatic to poikilitic (Fig. 2C). The hornblende crystals are optically zoned, with brown cores and green rims. Plagioclase (An$_{50}$ to An$_{30}$) is subbedal to euhedral and displays oscillatory normal and oscillatory reverse zoning. Subbedal biotite, plagioclase, and amphibole define a weak magmatic foliation. Quartz and K-feldspar are interstitial. Chlorite, epidote, sericite, minor sphe set, and actinolite are secondary minerals and can locally be abundant. The association of epidote with hornblende suggests that some epidote is magmatic.

**Upper Zone**

Tonalite through granite rocks of the upper zone are hypidiomorphic granular (Fig. 2E). Euhedral to subbedal hornblende and plagioclase define a weak magmatic foliation. Plagioclase is weakly normally zoned in samples collected near the transition with the central

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Figure 2. Thin section scans and photomicrographs. Abbreviations: Cpx—clinopyroxene, Opx—orthopyroxene, Bt—biotite, Srp—serpentine, Pl—plagioclase, Hbl—hornblende, Act—actinolite, Qz—quartz, K-spar—K-feldspar, Px—pyroxene. (A) Biotite hornblende two-pyroxene quartz diorite from the lower zone. White dash lines outline pyroxene crystals. (B) Pyroxene-rich block. (C) Biotite hornblende tonalite. (D) Pyroxene surrounded by actinolite and green hornblende. (E) Biotite hornblende granodiorite from the upper zone. (F) Andesitic roof-zone dike containing phenocrysts of orthopyroxene, augite, and plagioclase. (G) Dacitic roof dike with quartz, hornblende, plagioclase, and biotite phenocrysts. (H) Contact between a mafic magmatic enclave (MME) and the host granodiorite. The groundmass in the enclave is mainly composed of poikilitic K-feldspar, optically continuous across the boundary between the granite and the enclave.
zone (\(A_n_{04}\) to \(A_n_{20}\)) and becomes strongly normally zoned (\(A_n_{04}\) to \(A_n_{43}\)) in samples collected from the structurally higher (southwestern) part of the unit (Barnes, 1987). Subhedral biotite crystals are randomly oriented. Quartz is interstitial in tonalite but has euhedral faces against K-feldspar in granodiorite and granite. Orthoclase is interstitial to poikilitic and is variably perthitic. Zircon, apatite, and allanite are accessory phases. Pyroxene is rarely preserved as remnant cores in hornblende. However, the presence of rounded quartz inclusions in hornblende suggests that pyroxene was present as a high-temperature phase. Whether these crystals grew from the upper zone magma or were inherited is not possible to determine. The upper zone is petrographically distinct from both the central and lower zones in having euhedral hornblende phenocrysts as much as 1 cm long, a seriate distribution of euhedral to subhedral hornblende, and a general lack of evidence for augite to hornblende reaction. Chlorite, epidote, actinolite, sericite, and minor sphene are secondary minerals.

Late Granite

The late granite that intrudes the southern part of the upper zone (Fig. 1) is texturally distinct, with euhedral, strongly zoned plagioclase (\(A_n_{04}\) to \(A_n_{33}\)), acicular hornblende, and acicular to prismatic biotite. Perthitic orthoclase is present as poikilitic crystals that reach 1 cm in diameter. Quartz is subhedral to interstitial and displays slight undulose extinction. Euhedral sphene is a primary mineral and is commonly associated with magnetite. The late granite is the only part of the WCB in which sphene is a primary phase. Apatite and zircon are accessory minerals.

Mafic Selvages

Two main mafic selvages rim the intrusion along the southern and western contacts. The textural complexity observed in the southern mafic selvage suggests a complex history. For example, samples MMB-236a and WCB-2408 are slightly porphyritic, with optically zoned euhedral augite microphenocrysts. Relict olivine is surrounded by orthopyroxene and secondary oxides. The groundmass consists of euhedral to subhedral plagioclase (\(A_n_{18}\) to \(A_n_{33}\)) defining a weak magmatic foliation. Apatite is present as acicular needles included in plagioclase. In contrast, MMB-257a is distinct from the other pyroxene-bearing rocks in the selvage. Pyroxenes occur as aggregates associated with a relatively large proportion of opaque minerals, and the pyroxene-oxide clusters are surrounded by a groundmass of subhedral, aligned plagioclase (\(A_n_{01}\) to \(A_n_{43}\)); the foliation defined by the latter wraps around the pyroxene glomerocrysts.

Sample MMB-594 from the western selvage (Fig. 1) is a quartz-bearing biotite two-pyroxene diorite. Plagioclase is euhedral and zoned (\(A_n_{05}\) to \(A_n_{12}\)) and augite and orthopyroxene are euhedral, whereas biotite is poikilitic to interstitial. Relict olivine has been replaced by orthopyroxene and opaque minerals. Zircon and apatite are accessory minerals. This mineral assemblage is identical to that common in the lower zone, and the bulk compositions of samples from the western selvage are similar to lower zone compositions (Coint et al., 2013).

Roof Dikes

Andesitic roof dikes are porphyritic and contain 28％-41％ phenocrysts. Optically zoned plagioclase, augite, and orthopyroxene occur as phenocrysts and glomerocrysts (Fig. 2F). Plagioclase displays oscillatory normal to oscillatory reverse zoning with compositions varying between \(A_n_{01}\) and \(A_n_{43}\) (Barnes, 1987). Pyroxene is more or less altered to actinolitic amphibole, and plagioclase is partially albitionized. Dacitic roof dikes are porphyritic (Fig. 2G), with 23％-54％ phenocrysts of euhedral hornblende, plagioclase ± biotite, and quartz. Relict pyroxenes are rare as cores in amphibole phenocrysts and as inclusions in plagioclase phenocrysts. Plagioclase is oscillatory normally zoned with compositions varying between \(A_n_{18}\) to \(A_n_{50}\) (Barnes, 1987). Glomerocrysts are composed of large zoned plagioclase and euhedral hornblende, with interstitial quartz and minor K-feldspar and biotite. Accessory minerals are zircon, apatite, and allanite. Chlorite, actinolite, epidote, and sphene are secondary minerals and partially replace hornblende and biotite.

Mafic Magmatic Enclaves and Sympol联网 Dikes

The syngenetic dikes that are characteristic of the central zone are slightly porphyritic to aphanitic. Phenocrysts and groundmass minerals are mainly hornblende and plagioclase with lesser subhedral to anhedral biotite. The MME are porphyritic, with plagioclase and hornblende phenocrysts (Barnes, 1987; Buck et al., 2010). The groundmass of the MME consists of 200-300 µm subhedral plagioclase and hornblende. In the upper zone, poikilitic quartz and/or K-feldspar are optically continuous with crystals in the host granodiorite to granite and contain numerous inclusions of hornblende, plagioclase, and biotite (Fig. 2H). Mafic magmatic enclaves from the upper zone are porphyritic, with phenocrysts of zoned plagioclase, hornblende, and subhedral biotite. Hornblende clots are common in MME; some display actinolitic cores with small opaque mineral inclusions. Zircon and apatite are accessory minerals.

ANALYTICAL METHODS

Major and Trace Element Analyses

Methods used to analyze bulk-rock major and trace element compositions were presented in Coint et al. (2013). Major element data for minerals come from a variety of sources. Some hornblende data are available in the literature (Barnes, 1982, 1983, 1987). New microprobe data collected for this project are presented in Supplemental File 1. Some of the data published here were collected on the JEOL JXA8900 electron microprobe at the University of Wyoming, with operating conditions of 15 kV accelerating voltage, 20 nA current, and a beam diameter of 1-2 µm when focused. Matrix-matched natural and synthetic standards were used for calibration. The precision for individual major component was <0.1 wt% (Li, 2008). The remaining data, mainly obtained on rocks from the central zone, were collected at the University of Oklahoma using a Cameca SX50 electron microprobe equipped with five asynchronous wavelength-dispersive spectrometers and PGT PRISM 200 energy-dispersive X-ray analyzer. Analysis was done by wavelength-dispersive spectrometry using 20 kV acceleration, 20 nA beam current (measured at the Faraday cup), and 2 µm spot size. The PAP algorithm (Pouchou and Pichoir, 1985) was used to correct from matrix effects, with oxygen content calculated by stoichiometry.

Hornblende, augite, and orthopyroxene trace element data were collected in situ, in polished sections, by laser ablation (LA) ICP-MS analysis using a NewWave 213 nm solid-state laser and Agilent 7500CS ICP-MS at Texas Tech University. Spot diameter was 40 µm with a laser pulse rate of 5 Hz and fluence of 11-12 J cm⁻². For each analysis, 25 s of background (laser off) and 60 s of signal were recorded. NIST 612 glass was used as the standard and was analyzed every 5-7 unknown mineral analyses. Precision

Supplemental File 1. New microprobe data of hornblende and pyroxene. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00931.S1 or the full-text article on www.gsapubs.org to view Supplemental File 1.
and accuracy of the LA-ICP-MS system were determined by repeated analysis of basaltic glass BHVO-2 and are given in Supplemental File 2. The contents of SiO₂ or CaO for hornblende and orthopyroxene and CaO for augite, as determined by electron microprobe, were used as internal standards to correct for variability in ablation efficiency. Reduction of the hornblende data using either CaO or SiO₂ as internal standards gave results within the uncertainty of the counting statistics. The LA-ICP-MS data set is presented in Supplemental File 3.

**Geochemistry**

A discussion of the bulk-rock major and trace element data was presented in Coint (2012) and Coint et al. (2013), in which we show that the lower zone has scattered compositions that do not define a clear-cut compositional trend. The data also show that mafic selvages along the western contact are compositionally identical to the lower zone. In contrast, rocks of the central and upper zone were interpreted to define a compositional array with inflections in the trends of a number of trace and minor elements that are indicative of fractional crystallization. The data also support a cogenetic relationship between rocks of the upper zone and dacitic roof dikes. These relationships are illustrated and expanded upon in Figures 3A and 3B, in which the following is evident. (1) The lower zone is clearly distinct from the central and upper zones in terms of Mg/(Mg + Fe) and Sr contents (Figs. 3A, 3B). (2) The compositions of lower zone tonalites do not plot on the crude trend defined by more mafic lower zone rocks (Fig. 3A); this indicates that fractional crystallization is not an adequate explanation for the compositional variation in the lower zone. (3) The central and upper zones define an array of compositions that overlaps those of dacitic roof-zone dikes. (4) Andesitic roof zone dike compositions tend to cluster in the low-SiO₂ part of the central and upper zone trends. (5) The late-stage hornblende biotite granite is compositionally distinct from the rest of the batholith.

Rare earth element (REE) patterns of most WCb samples are broadly similar across rock types. Figure 3. Bulk-rock major element diagrams. (A) Mg/(Mg + Fe) versus SiO₂ (l.z.—lower zone). (B) Sr versus MgO. (C) Rare earth element (REE) patterns normalized to chondrite (Boynton, 1984) of the central and lower zone rocks. (D) REE patterns normalized to chondrite (Boynton, 1984) of upper zone rocks. (E) REE patterns normalized to chondrite (Boynton, 1984) of the roof-zone dikes; both dacite and andesite. (F) REE patterns normalized to chondrite (Boynton, 1984) of synplutonic dikes.
Pluton assembly using trace elements in augite and hornblende

Types, with moderate negative slopes and small or no Eu anomalies (Figs. 3C–3E). Typical gabbroic through tonalitic lower zone rocks are shown as a field in Figure 3C because these samples have parallel slopes. As shown in Coint et al. (2013), the REE abundances of these rocks are positively correlated with SiO₂ contents. In contrast, pyroxenite and melagabbro blocks and intrusive sheets in the lower zone have lower REE abundances and flatter patterns, and display both positive and negative Eu anomalies (Fig. 3C). Although samples from the central and upper zones have total REE (REE tot) abundances broadly similar to lower zone samples (Figs. 3C, 3D), with increasing SiO₂, the abundances of middle and heavy REEs decrease and the REE patterns show more pronounced concave-upward shapes (Fig. 3C; Barnes, 1983).

The REE patterns of adestic and dactitic roof zone dikes are shown in Figure 3E. The adestic dike patterns are parallel to those of the least evolved (lowest SiO₂) dacies. As with upper zone samples, the REE abundances of the dactitic dikes decrease with increasing SiO₂ and the REE patterns show greater upward concavity (Fig. 3E).

Mafic syenitonic dikes, which characterize the central zone, show wide compositional scatter, but essentially overlap compositions of lower zone samples (Figs. 3A, 3B). The REE patterns of these dikes are similar to those of lower zone rocks, but show a greater range of REE abundances (Fig. 3F).

Mineral Chemistry

The major element compositions of minerals from the WCB are summarized here from previously published work (Barnes, 1987; for additional analyses, see Supplemental File 1 [see footnote 1]; for LA-ICP-MS analyses for augite, orthopyroxene, and hornblende, see Supplemental File 3 [see footnote 3]). The compositions of augite crystals in pyroxene-rich blocks have compositions slightly more magnesian (Wo 0.44–0.49, En 0.42–0.46, Fs 0.07–0.18) than the augite from the common rock types in the lower zone. Augite from the pyroxene-rich blocks for two different regions in the Zr versus Cr diagrams (Figs. 3C, 3C).

In the following section, pyroxenes are classified according to the official nomenclature (Morimoto et al., 1988). Both augite and diopside are present; however, for simplicity clinopyroxene is referred to as augite. Nonquadrilateral components vary between 2.3–11.1 mol%, with Al being the most abundant nonquadrilateral cation (Barnes, 1987). Trace element abundances are compared to Zr contents in Figure 4 because Zr is incompatible in pyroxene and behaved incompatibly in rocks of the lower zone (e.g., Barnes, 1983). Therefore, increasing Zr in a single pyroxene grain is reflective of increasing Zr in the melts from which the pyroxene grew.

Augite from the lower zone varies between wollastonite, Wo 0.42–0.47, enstatite, En 0.40–0.47, ferrosilite, Fs 0.14–0.18. In terms of trace elements, augites from the lower zone have compositions that vary widely and can be divided into two groups based on their Cr content, with one group containing >3700 ppm Cr and the other containing <2700 ppm Cr. With increasing Zr, Cr decreases exponentially from 5995 to 167 ppm (Fig. 4A). REE tot contents range from 5 to 156 ppm. Individual crystals tend to show increasing REE tot from core to rim (Fig. 4B). Strontium contents are nearly constant with increasing Zr and vary between 12.2 and 25.6 ppm (Fig. 4D). The augite crystals displaying the highest Sr content are part of the high Cr group.

Augite crystals in pyroxene-rich blocks have compositions slightly more magnesian (Wo 0.44–0.49, En 0.42–0.46, Fs 0.07–0.18) than the augite from the common rock types in the lower zone. Augite from the pyroxene-rich blocks for two different regions in the Zr versus Cr diagrams (Figs. 3C, 3C).

In the upper zone, relict augite was analyzed in only one sample, MMB-397. The few augite crystals preserved have low Cr (92–471 ppm), intermediate Zr (14.8–19.3 ppm; see Supplemental File 3 [see footnote 3]; Figs. 4E, 4F), and Sr concentrations that overlap with augite from the lower zone.

Compositions of augite in the western sylvite overlap those from the lower zone and the pyroxene-rich blocks (Figs. 4E, 4F). In contrast, augite from the southern selvage is distinct in having a broader range of Mg# (0.62–0.78 versus 0.65–0.68 from the western selvage augite; Barnes, 1987). In terms of trace element abundances, augite from the western selvage has Cr and Sr contents similar to those from the lower zone (Figs. 4E, 4F), but have higher concentrations of REEs. Augite from the southern mafic selvage has higher Sr (>22 ppm) and lower Cr compared to the augite from the western sylvite. The Zr and REE tot contents of augite from the southern mafic selvage are similar to augite from the two-pyroxene adiabatic roof dikes (Figs. 4E, 4F).

Augite from the two-pyroxene adiabatic roof dikes has higher TiO₂ and Al₂O₃ and lower CaO contents than augite from the lower part of the intrusion (Barnes, 1987). Chromium contents are relatively constant and low (<1000 ppm) compared to augite from the lower WCB (Fig. 4C).

In the following we use the amphibole nomenclature of Leake et al. (1997). We refer to brown-green amphibole as hornblende and to pale green amphibole as actinolite. Actinolitic hornblende is the transition between the two.

Amphibole compositions in the lower zone range from magnesiohornblende to actinolite (fig. 6 in Barnes, 1987). Chromium concentrations vary from 2023 to 139 ppm, with most spots below 1200 ppm (Fig. 5A); Cr is not correlated with Ti contents. Hornblende contains variable amounts of REEs (Fig. 5C); the light (L) REEs vary from 10× to 100× chondritic values, whereas heavy (H) REEs vary from 6× to 50× chondrites. Hornblende REE patterns are not necessarily parallel to each other, especially for the LREEs, where the shape of the patterns is quite variable (Fig. 5C). The size of the negative Eu anomaly increases with increasing REE abundance.

Amphiboles in the pyroxene-rich blocks are magnesiohornblende and these amphiboles have higher Mg# (0.8–0.9) than does hornblende in other lower zone samples. In contrast, Cr contents overlap with hornblende from the lower zone (Fig. 5A). Titanium and Zr are present in relatively low concentrations (<5000 ppm and <50 ppm, respectively; Fig. 5B). The REE concentrations in magnesiohornblende in the pyroxenites are lower than those from the other rocks of the lower zone, with LREE concentrations between 7× and 50× chondrites, and HREE concentrations between 2.5× and 12× chondrites. Both slightly positive and slightly negative Eu anomalies are present.
Figure 4. Diagrams displaying trace element concentrations in augite crystals. Int—intermediate parts of the augite. (A, B) Trace element composition of augite from the lower zone and from the pyroxenite-melagabbro blocks. Expected fractionation trends (F) are displayed in a schematic diagram in the lower right corners. The gray arrows in B display variable fractionation trends recorded by the augite in different samples. REE tot—total rare earth elements. (C, D) Trace element composition of augite from the roof-zone dikes. (E, F) Trace element composition of augite from the mafic selvages and upper zone.
Pluton assembly using trace elements in augite and hornblende

Figure 5. (A, B) Trace element plots of hornblende from the lower zone and pyroxenite-melagabbro blocks. The striped pattern shows the range of composition found in the upper zone hornblende. In A, the black arrows indicate analysis of hornblende close to the pyroxene core, where the reaction is potentially incomplete and Cr is inherited from the augite (see the text for more explanations). (C) Rare earth element patterns of hornblende (Hbl) from the lower zone, normalized to chondrite (Sun and McDonough, 1989).

The infl uence of orthopyroxene exsolution on augite composition is discussed in Supplemental File 4. A simple experiment in which exsolved cores and rims were analyzed demonstrates that the REE patterns of augite from which orthopyroxene has exsolved are not affected by the amount of orthopyroxene exsolution, because the proportion of the REEs that orthopyroxene can accommodate is small. Therefore, the REE contents of augite from which orthopyroxene has exsolved are similar to those of augite from which orthopyroxene has not exsolved.

Hornblende from the upper zone is magnesiohornblende to ferrohornblende in which Ti and Al decrease from core to rim as Mg# (figs. 6A and 7A in Barnes, 1987) and Sc abundances (this work) remain approximately constant. The trace elements P, Hf, Y, Sr, V, Sc, REEs all decrease toward the rims (not shown). The REE abundances are between 40× and 100× chondrites for the LREEs and 20–60× chondrites for HREEs (Figs. 8D–8F). Most of the REE patterns are subparallel to each other. Hornblende cores display medium-sized negative Eu anomalies, which decrease in the intermediate part of the crystal and increase toward rims (Fig. 8C).

Hornblende phenocrysts in the dacitic roof dikes are magnesiohornblende to actinolitic magnesiohornblende (Barnes et al., 1987). These hornblende samples are similar in major and trace element concentrations (Mg#, Ti, Cr, Zr, V, Sc, REEs) to hornblende from the upper zone (Fig. 8H) and from several samples from the central zone. This similarity includes decreasing concentrations of Zr, Ti, and Cr from crystal cores to rims.

**DISCUSSION**

The influence of orthopyroxene exsolution on augite composition is discussed in Supplemental File 4. A simple experiment in which exsolved cores and rims were analyzed demonstrates that the REE patterns of augite from which orthopyroxene has exsolved are not affected by the amount of orthopyroxene exsolution, because the proportion of the REEs that orthopyroxene can accommodate is small. Therefore, we found in the groundmass. Cr contents are anti-correlated to Zr and Ti and increase in concentration from core to rim (Fig. 7A). Small, euhedral hornblende crystals interpreted as growing late in the crystallization history of the enclaves has Cr concentrations that are highly variable, from 60 to 800 ppm (Fig. 7A). REE concentrations are correlated with Ti and Zr concentrations and decrease from core to rim. The negative Eu anomalies in the cores of hornblende phenocrysts are larger than in rims and in groundmass hornblende (Fig. 7D). Hornblende crystals in hornblende phenocrysts contain higher amounts of Cr, Zr, and Ti than are found in MME.

Hornblende from MME and synplutonic dikes is magnesiohornblende (Barnes, 1987). Phenocryst cores contain higher concentrations of Zr and Ti than in the rims or in late hornblende found in the groundmass. Cr contents are anti-correlated to Zr and Ti and increase in concentration from core to rim (Fig. 7A). Small, euhedral hornblende interpreted as growing late in the crystallization history of the enclaves has Cr concentrations that are highly variable, from 60 to 800 ppm (Fig. 7A). REE concentrations are correlated with Ti and Zr concentrations and decrease from core to rim. The negative Eu anomalies in the cores of hornblende phenocrysts are larger than in rims and in groundmass hornblende (Fig. 7D). Hornblende crystals in hornblende phenocrysts contain higher amounts of Cr, Zr, and Ti than are found in MME.

Hornblende from the upper zone is magnesiohornblende to ferrohornblende in which Ti and Al decrease from core to rim as Mg# (figs. 6A and 7A in Barnes, 1987) and Sc abundances (this work) remain approximately constant. The trace elements P, Hf, Y, Sr, V, Sc, REEs all decrease toward the rims (not shown). The REE abundances are between 40× and 100× chondrites for the LREEs and 20–60× chondrites for HREEs (Figs. 8D–8F). Most of the REE patterns are subparallel to each other. Hornblende cores display medium-sized negative Eu anomalies, which decrease in the intermediate part of the crystal and increase toward rims (Fig. 8C).

Hornblende phenocrysts in the dacitic roof dikes are magnesiohornblende to actinolitic magnesiohornblende (Barnes et al., 1987). These hornblende samples are similar in major and trace element concentrations (Mg#, Ti, Cr, Zr, V, Sc, REEs) to hornblende from the upper zone (Fig. 8H) and from several samples from the central zone. This similarity includes decreasing concentrations of Zr, Ti, and Cr from crystal cores to rims.

**DISCUSSION**

The influence of orthopyroxene exsolution on augite composition is discussed in Supplemental File 4. A simple experiment in which exsolved cores and rims were analyzed demonstrates that the REE patterns of augite from which orthopyroxene has exsolved are not affected by the amount of orthopyroxene exsolution, because the proportion of the REEs that orthopyroxene can accommodate is small. Therefore, we

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**Supplemental File 4. The role of exsolution on augite REE content.** If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00931.S4 or the full-text article on www.gsapubs.org to view Supplemental File 4.
Figure 6. Trace element plots from the hornblende of the central zone. (A) Cr versus Ti binary diagram. (B) Zr versus Ti diagram. The dashed line contours the hornblende analysis from the Granite Lakes basin. (C) Eu* versus Zr in hornblende. The back arrows points out the trend existing from core to rim. (D–H) Hornblende (Hbl) rare earth element (REE) patterns normalized to chondrite (Sun and McDonough, 1989) shown by location compared to the REE patterns of the hornblende from the upper zone (striped field) and three late hornblendes from synplutonic dikes (gray field). Sample numbers are indicated in italics.
assume that the compositions obtained by LA-ICP-MS reflect magmatic augite compositions.

Pluton Assembly Based on the Mineral Chemistry

Mineral compositions record information about the composition of the melt from which the mineral grew. The melt composition is, in turn, influenced by the minerals that precipitate from it. These minerals need not be physically fractionated (i.e., to be removed from the system) to induce compositional variations in newly formed growth zones of existing minerals. Therefore, core to rim compositional variations in minerals can result from a variety of magmatic processes such as fractional crystallization, mixing, and assimilation–fractional crystallization, assuming that intracrystal diffusion is slow enough for the zoning patterns to be preserved (see following). Bulk-rock compositions, unlike mineral ones, will also record processes such as accumulation. Therefore, both data sets need to be combined to understand in detail the effects of these different processes. Here we refer to fractional crystallization when there is chemical evidence for separation between the minerals and the residual melt, to in situ fractional crystallization when chemical variation was induced by precipitation of minerals from the melt without physical removal of the newly grown phase from the system, and to fractionation when the process responsible for the chemical variation is not identified.

The amount of a trace element incorporated into a mineral lattice is governed by the partition coefficient between the melt and the mineral, which varies with pressure, temperature, and magma composition. Inasmuch as both augite and hornblende were high-temperature phases in the various parts of the WCb, the two minerals recorded the evolution of the melt from which they crystallized, and therefore allow us to identify and map the extent of the distinct magma batches in the batholith. Distinct magma batches should produce minerals with distinct compositional and zoning patterns and magma mixing should result in discernible compositional reversals.

In this study, the core to rim variations in trace element abundances are used to understand the evolution of the melt composition time through time. Because intracrystalline diffusion of REEs in diopside is very slow under crustal conditions (Van Orman et al., 2001), the REE record preserved in both hornblende and augite is used as a proxy for the melt evolution. Elements such as Ti (Cherniak and Liang, 2012), Sr, and Cr are more sensitive to intracrystalline diffusion. However, these element data are used in this study because their intracrystalline variations are broadly correlated with the REE data, and we therefore conclude that compositional variation in the minerals is primarily a function of the magmatic record.

Magmatic Evolution of the Lower WCb Based on Augite Compositions

High bulk-rock concentrations of MgO and Sc in the lower zone indicate that some of these rocks are cumulates, and therefore that fractional crystallization affected lower zone magmas. Augite was one of the first minerals to crystallize, along with plagioclase (Fig. 2A), and may be used to track the evolution of the magma compositions. Hornblende, which is present as a rimming phase around pyroxene, grew at lower temperature conditions and is a less valuable indicator of changes in melt composition in this zone (Fig. 2A).

The abrupt core to rim decrease in Cr and the proportional increase in REETOT and associated increasing Zr concentrations in augite (Figs. 4A, 4B) are consistent with evolution of magmas in the lower zone by fractional crystallization. However, the extent of fractional crystallization is not the same in every sample. In some samples, augite displays wide compositional variations (e.g., WCB-5209), whereas in others

Figure 7. (A, B) Binary diagrams displaying chemical variations in hornblende from mafic magmatic enclaves (MME; MMB-681b and MMB-686b) and synplutonic dikes (MMB-775, MMB776, and MMB-860) compared to those found in the upper zone (striped pattern). (In the legend, “Dikes” refers to synplutonic dikes.) (C, D) Hornblende (Hbl) rare earth element patterns normalized to chondrite (Sun and McDonough, 1989) from mafic magmatic enclaves and synplutonic dikes, normalized to chondrite (Sun and McDonough, 1989).
Figure 8. Trace elements in hornblende (Hbl) from the upper zone. (A–C) Binary diagrams displaying chemical variations in hornblende from the upper zone compared to hornblende from the late granite and hornblende from the dacitic roof dikes (RZD). In C, the red arrow points out the evolution trend for hornblende from the upper zone, from core to rim, whereas the black arrow follows the trend from core to rim described by hornblende in sample MMB-397. LREE—light rare earth elements. (D–H) Hornblende REE patterns normalized to chondrite (Sun and McDonough, 1989). Black arrows show the LREE–depleted patterns (see the text for explanations) interpreted as late rims crystallizing from strongly fractionated residual melt. Qtz—quartz.
Inherited Compositions?

On the basis of textural relationships, hornblende crystals in samples from the lower zone grew late in the magmatic history (Fig. 2A), and so are more likely to have recorded processes that occurred at lower temperatures relative to those preserved in augite. The Cr contents of some of the hornblende (Fig. 5A) are anomalously high for crystals that formed from a low-temperature, evolved melt, and are much higher than the Cr concentrations found in the upper zone hornblende. These Cr concentrations partly overlap with the Cr contents of the lower zone augite crystals (cf. Figs. 4A and 5A). A few high-Cr hornblende crystals are also found in the central zone (Fig. 6A) and in one rock from the upper zone (Fig. 8B). These high Cr contents may result from incomplete reaction between melt and pyroxene to form hornblende or from direct crystallization of hornblende from a Cr-rich mafic melt. In the lower zone, the Cr-rich hornblende forms clusters or is associated with quartz beads. Both of these textural features result from pyroxene + melt reactions (Castro and Stephens, 1992; Stephens, 2001), suggesting that the high Cr content of the hornblende is inherited from the original pyroxene. The extent of this inheritance is not well constrained. Cr is present in larger concentrations in augite than in hornblende; therefore if diffusion is slow, or if the reaction takes place rapidly, complete equilibration between the newly formed hornblende and the melt might not be reached. Other cations with 3+ and 4+ valences, such as Al, Ti, and REEs, could also be inherited from preexisting augite. However, Supplemental File 1 (see footnote 1) shows that the abundances of these elements are generally higher in hornblende than in augite, which indicates that they are more readily partitioned into hornblende. Therefore, it is probable that the concentrations of these elements in the latter represent equilibrium between the growing hornblende and melt, whereas Cr concentrations reflect inheritance from precursor augite. This example shows the importance of combining textural observations with the geochemistry of mineral phases when interpreting the nature of peritectic reactions.

Pyroxene-Rich Blocks and Their Relationships with the Lower Zone

Pyroxene-rich blocks and intrusive bodies in the lower zone can be either (1) cumulates resulting from disruption of dikes, or (2) fragments of ultramafic host rocks that reacted with the host magma, resulting in the crystallization of pyroxenes. The composition of the olivine found in sample MMB-100 (forsterite, Fo71) is not appropriate for mantle rocks (as pointed out in Barnes, 1983, 1987), leaving the first hypothesis as the most likely explanation. The composition of the augite in the lower zone overlaps with augite compositions from the pyroxenite and melagabbro bodies (Figs. 4A, 4B), and this similarity indicates that augite in these bodies crystallized from magmas similar to those that formed the lower zone. The field relationships and the textural observations suggest that this overlap is mostly the result of crystal exchange between the pyroxene-rich bodies and the host magmas. For example, sample WCB-133 displays a gradational contact with a large block of pyroxene melagabbro. All analyzed augite cores and mantles in sample WCB-133 have high Cr concentrations and are interpreted as being inherited from the adjacent melagabbro, whereas rims overlap with augite compositions typical of lower zone augite (Fig. 4A). These rims are interpreted to form after augite from the pyroxene-rich block was transferred to the magma that hosted the block. The high Cr contents and high Mg# (~0.8 instead of values of 0.65–0.75 typical of most lower zone rocks; Barnes, 1987) of the augite in pyroxenite and melagabbro bodies and the high Sr contents of these augite crystals compared to the rest of the lower zone augite (Figs. 4A, 4E) suggest that parent magmas of the pyroxene-rich bodies were relatively primitive and water rich, consistent with the paucity of olivine in these rocks.

Mafic Selvage

The augite crystals found in the western selvage (e.g., sample MMB-594) display geochemical characteristics similar to those from the lower zone (Figs. 4E, 4F). This similarity and the fact that the western selvage was coeval with the lower zone are interpreted to indicate that the western selvage formed from magma batches similar to those that formed the lower zone of the pluton.

Rocks from the southern mafic selvage show much more diversity in texture, bulk-rock compositions (Fig. 3), and mineral compositions (Figs. 4E, 4F). Augite analyzed in samples MMB-257a and MMB-236a overlap with augite from the two-pyroxene andesite roof dikes in terms of Sr concentrations and display similar trends in REETOT versus Zr (Figs. 4E, 4F). Therefore, we interpret the southern mafic selvage to be related to the two-pyroxene andesitic roof dikes.

Upper Zone

Bulk-rock compositions from the upper zone of the WCB display narrow trends (Fig. 3). Therefore, fractional crystallization and associated crystal accumulation were interpreted to be the main processes responsible for bulk compositional variation (Coint et al., 2013). In the upper zone, allanite, apatite, and zircon are accessory phases. Zircon and apatite occur as inclusions in biotite, hornblende, and plagioclase, suggesting that these minerals were stable at temperatures above the solidus and were likely to have fractionated during crystallization. Fractionation of hornblende along with the accessory minerals will result in decreasing concentrations of the REEs in the residual melt, and therefore in the minerals that grow from the progressively fractionated melt. One of the consequences is that incompatible elements whose enrichment can be used to unequivocally track the evolution of the system are absent. However, the amount of Ti incorporated into hornblende depends on temperature (Otten, 1984; Ernst and Liu, 1998; Femenias et al., 2006), and therefore is used as a proxy to track the down-temperature variations of trace elements (Fig. 8).

Hornblende from the upper zone shows a limited variation in Mg# (Barnes, 1987), REE concentrations, shape of the REE patterns (Figs. 8D–8H), and Cr contents (Fig. 8B). The abundance of REE elements and the size of the Eu anomaly tend to decrease from core to mantle (Fig. 8C), but hornblende rims tend to display larger Eu anomalies than the mantle. This core to rim decrease in trace element abundances can be explained in terms of the
compatible behavior of the REEs, although the slight decrease in the Eu anomaly from core to mantle (Figs. 8D–8H) is counterintuitive. In the upper zone, fractionation of a small amount of apatite and allanite is thought to counteract the expected increase in the size of the negative Eu anomaly. We present simple models to illustrate this effect in Supplemental File 5. The presence of large (1 mm diameter) allanite crystals in the granodioritic and granitic rocks suggests that in some cases the proportion of allanite that fractionated may be greater than the value used in the model (see Supplemental File 5 [see footnote 5]). In such a case, allanite fractionation will deplete the residual melt in LREEs relative to the HREEs. This depletion effect can explain the presence of a few hornblende rims that have lower LREE concentrations than typical (Figs. 8D, 8F). The larger negative Eu anomaly associated with some of the LREE-depleted patterns, compared to the rest of the hornblende patterns from the upper zone, cannot be accounted for by allanite fractionation, but could be the result of rare instances of in situ K-feldspar fractionation (e.g., Bachmann et al., 2005). This possibility is consistent with the fact that most of the hornblende rims with relatively large Eu anomalies are from granitic or granodioritic samples.

The abundances of elements such as Zr and Ti in upper zone hornblende are well correlated and define a single trend (Fig. 8A). By comparison with the lower zone (Fig. 4B), which represents multiple magma batches with limited chemical communication, this single trend of upper zone hornblende compositions indicates that upper zone hornblende crystallized from (1) a single batch of magma or (2) multiple magma batches that were compositionally identical and had identical crystallization histories. The well-defined bulk-rock trends in Hacker diagrams (Figs. 3 and 8 in Coint et al., 2013) coupled with the decreasing An content of plagioclase (An$_{w}$ to An$_{n}$; Barnes, 1987) with increasing bulk-rock SiO$_{2}$ concentrations are consistent with either interpretation. However, the lack of intrusive contacts in the upper zone compared with the abundance of intrusive contacts in the central and lower zones, and the gradual upward change from tonalitic to granitic compositions among upper zone rocks support crystallization from a large homogeneous magma body.

Hornblende from one upper zone sample, MMB-397 (Figs. 1 and 8G), displays wider variations in REE compositions than the rest of the upper zone hornblende (cf. Figs. 8D, 8E, 8F). This rock is also distinct from other analyzed samples because of the presence of relict pyroxene as inclusions in poikilitic hornblende and plagioclase with reverse zoning (Barnes, 1987). These features suggest a distinct magmatic history compared to the rest of the upper zone. The composition of the relict pyroxene (Figs. 4E, 4F) indicates that the pyroxene grew from a magma similar to magmas that formed the lower zone and the western mafic selvage (e.g., MMB-594; Fig. 1). Hornblende cores and mantles have compositions similar to hornblende from the central and the upper zones, and are interpreted to have grown from upper zone magma. In contrast, the rims of these hornblendes display steeper REE patterns with enrichment in LREEs but without Eu anomalies. The REE patterns of hornblende rims are interpreted as resulting from crystallization from a hybrid magma that formed by mixing of upper zone magma with a crystalline lower zone-type magma batch from the adjacent mafic selvage. Most of the textural characteristics of upper zone rocks are present in this rock: medium grain size (2 mm), poikilitic to interstitial K-feldspar, large (at least 100 µm long), subequant zircons, euhedral hornblende, and glomerocrysts of zoned plagioclase. All of these features indicate that the upper zone magma accounted for much of the mass of this rock. The presence of pyroxene crystals similar in composition to ones found in the western selvage (MMB-594; Figs. 4E, 4F) as inclusions in hornblende, plus abundant hornblende grains that contain actinolite ± oxide inclusions are interpreted to be the result of reaction between magma of the upper zone with augite from the western selvage. If this is the case, it would suggest that the upper zone magma invaded mushy magma of the western selvage and incorporated some of its minerals, such as augite. This interpretation is consistent with hornblende compositional variations from core to rim. Hornblende cores grew from upper zone magma, whereas the rims crystallized from a more mafic, hybrid magma, which resulted in variations in the LREE pattern shape and smaller negative Eu anomalies than found in most upper zone hornblende.

Central Zone

The central zone of the intrusion plays a key role in understanding the link between the upper and lower zones because, as indicated above, it constitutes a transition between the two. Hornblende in the central zone displays major and trace element concentrations similar to hornblende from the upper zone, with moderate Cr concentrations and decreasing Ti and Zr from core to rim (Figs. 6A, 6B). In samples WCB-4809 and WCB-4909, hornblende REE patterns overlap perfectly with the REE patterns of hornblende from the upper zone (Fig. 6E). These data suggest that hornblende in WCB-4809 and WCB-4909 crystallized from magmas similar in composition to the upper zone magma and are consistent with the age of zircon from WCB-4909, which is identical to upper zone samples. These two samples were collected from the structurally lower part of the central zone (Fig. 1) and they are organized in a series of alternating comagmatic sheets (fig. 3F in Coint et al., 2013). They are, therefore, interpreted be part of the feeder system to the upper zone.

Five other central zone samples contain hornblende having REE patterns that show more variability. For example, hornblende cores in these five samples commonly overlap compositions of hornblende from the upper zone; however, their rim compositions have lower REE contents and display smaller Eu anomalies than in upper zone samples (Figs. 6D, 6F–6H). One sample, MMB-777b, shows a bimodal distribution of the REE patterns (Fig. 6D). The compositional overlap between hornblende cores from the central zone with hornblende from the upper zone is strongly suggestive that these hornblende cores grew from upper zone magmas. However, unlike WCB-4809 and WCB-4909 and most of the hornblende in the upper zone, the rims of central zone hornblende have lower REE contents and small negative to no Eu anomalies (Fig. 6C). These zoning features can be explained in several ways.

Fractional Crystallization

It has already been demonstrated that in the upper zone REEs were compatible and concentrations generally decreased with fractional crystallization. In the central zone, minerals such as K-feldspar and allanite, which were involved in the evolution of upper zone magma, are scant. Therefore, the variations in the rims of the hornblende in the central part cannot be associated with fractionation of such minerals. Textures of MMB-777b, WCB-6209, WCB-6309, and WCB-7109 are also distinct from the upper zone samples. In the latter samples, hornblende and biotite are poikilitic and contain numerous inclusions of plagioclase, hornblende, and apatite, and relict pyroxene cores are common. These features are rarely observed in the upper zone rocks, so it is unlikely that hornblende compositions from these samples result from simple fractional crystallization.
Mixing between MME, Synplutonic Dikes, and the Upper Zone Magma

A second possibility is that hornblende rims in samples MMB-777b, WCB-6209, WCB-6309, and WCB-7109 grew from hybrid magma. Ample evidence of mixing and mingling is present in the central zone, such as disrupted synplutonic dikes and swarms of MME (Barnes et al., 1986a). The presence of pyroxene cores surrounded by hornblende (Fig. 2D) would suggest that mixing could have occurred between lower zone and upper zone magmas.

The extent of the interactions between parent mafic magmas of MME and synplutonic dikes (mainly basaltic andesite) and central zone magmas is difficult to constrain for the following reasons. The cores of hornblende phenocrysts in the MME have compositions similar to hornblende cores from central and upper zone rocks (Fig. 7). This similarity is interpreted to indicate that hornblende phenocrysts in MME are xenocrysts from the surrounding magma that were mechanically incorporated into the MME magmas. One of the MME from the upper part of the pluton has a groundmass of centimeter-scale, poikilitic K-feldspar crystals that are optically continuous with K-feldspar crystals in the host granodiorite (Fig. 2H). This relationship suggests that many MME are complex hybrids and that mixing involved not only exchange of large crystals, but also flow of interstitial melt between enclave and host magma. For these reasons, the MME are not viewed as appropriate mixing end members.

In contrast, hornblende in the synplutonic dikes is more likely to reflect compositions of the mafic magmas because there was evidently much less interaction between the dikes and the surrounding magma, as is indicated by the sharp contacts between dikes and their host rocks (Fig. 3E in Coint et al., 2013). As is the case in the MME, the cores of hornblende phenocrysts found in synplutonic dikes have REE compositions similar to those of hornblende in the host magma (Fig. 7C), suggesting that the hornblende phenocrysts in the dikes were inherited from their hosts. However, phenocryst rims and groundmass hornblende in the synplutonic dikes have high Cr contents (Fig. 7A) yet lack evidence for reaction from pyroxene. This lack indicates that the Cr concentrations observed in the groundmass hornblende and hornblende rims are a reflection of the concentration of Cr in the dike magma. Therefore, we consider the compositions of hornblende rims and groundmass hornblende to be representative of a possible mixing end member.

If mixing between the central zone magma and the synplutonic dike magmas was important in the central zone, then the resulting hornblende compositions should be between those of the groundmass hornblende from the synplutonic dikes and the hornblende from the upper zone (Fig. 7A). However, such mixing does not explain the low Cr contents of hornblende rims, the wide range of Cr contents, and the lack of Ti increase in phenocryst rims and groundmass hornblende (Fig. 6A). These features indicate that the potential mixing end member was not primitive (in the sense of containing high Cr and Ti contents) and therefore not a synplutonic dike magma. Instead, the mixing end member was probably of intermediate composition.

Hornblende in central zone sample WCB-5709b is unlikely to be the result of mixing of upper zone magma with synplutonic dike magma because hornblende in this sample has higher LREE and lower middle REE and HREE concentrations than the hornblende from the synplutonic dikes (Fig. 6H). The low middle and heavy REE concentrations (Fig. 6H) and the small Eu anomaly suggest that the hornblende in WCB-5709b crystallized from a relative mafic magma. Relict pyroxene, now completely transformed into hornblende, are abundant in the sample, and we interpret this feature to indicate that sample WCB-5709b is a fragment of the lower zone now surrounded by central zone rocks, an interpretation consistent with the age of central zone sample Z5, which is identical to ages of lower zone samples.

Mixing between the Lower and Upper Zone Magmas

A second mixing model for the central zone involves mixing between magmas of the upper and lower zones. Texturally, the central zone rocks are possibly hybrids between the lower zone, which contains pyroxene partially reacted to actinolite hornblende or hornblende (Fig. 2D), and the upper zone, which contains coarse phenocrysts of euhedral zoned plagioclase and hornblende (Fig. 2E). The gradational contacts between the different zones of the intrusion and amphibolite and hornblende from the Ukonom Lake and Granite Lakes areas (WCB-6209, WCB-6309, WCB-7109; Figs. 8F, 8G). We therefore conclude that the rocks from the Ukonom Lake and Granite Lakes areas of the central zone formed by mixing between a pyroxene-bearing magma and upper zone magma. The nature of the pyroxene-bearing end member is difficult to constrain as no obvious mixing trends are observed in Harker diagrams (fig. 8 in Coint et al., 2013) or in Figures 6, 7, and 8. However, these samples were collected in proximity to the gradational contact with the lower zone, which is a good candidate for a mixing end member.

In summary, most of the central zone crystallized from upper zone magma, and we interpret the sheeted part of the central zone preserved in the Cuddihy basin to be part of the feeding system to the upper zone (see Coint et al., 2013). Some samples, particularly ones collected near the boundary with the lower zone (samples MMB-777b, WCB-7109, WCB-6309, and WCB-6209), are primarily the result of mixing between lower and upper zone magmas, with local complications due to further mingling and mixing with magmas parental to the MME and synplutonic mafic dikes. Among the analyzed samples, WCB-4809 and WCB-4909 are the only ones that show no evidence of input from mafic magmas.

Roof Dikes

Two-Pyroxene Andesitic Roof Dikes

Some of the two-pyroxene andesitic dikes are composite, with older external two-pyroxene andesite and inner biotite granodiorite related to the upper zone (Barnes et al., 1986a; Coint et al., 2013). Other two-pyroxene andesitic dikes cut the uppermost part of the upper zone (Coint et al., 2013). These mutual crosscutting relationships indicate that the two-pyroxene andesitic magmas were coeval with the magmas of the upper zone. The previous interpretation of the origin of the two-pyroxene andesitic dikes (Barnes et al., 1986a) was that they represented magmas that were derived from the lower zone. This interpretation was based on the broad overlap of bulk-rock compositions of the dikes with the lower zone (Barnes et al., 1986a). However, augite crystals in the andesitic roof dikes have compositions that are distinct from augite in the lower zone, with higher Sr and lower Cr contents (Figs. 4C, 4D). Augite from the andesitic dikes is also distinct in showing zoning reversals (e.g., MM-164b; Figs. 4C, 4D), whereas most lower zone augite crystals are normally zoned (Figs. 4A, 4B). The zoning reversals indicate that many of the two-pyroxene andesitic roof dike magmas underwent a mixing event, a conclusion that is consistent with the common occurrence of orthopyroxene phenocrysts rimmed by augite and zoning reversals in plagioclase phenocrysts (Barnes, 1987).
The presence of high-Al orthopyroxene in these dikes (Barnes, 1983, 1987) suggests that mixing was not a deep-crystall event and was more likely to occur in a shallower reservoir.

Although the compositions and zoning patterns in the andesitic roof dike rules out an origin from lower zone magmas, it is possible that the dike magma came from the central zone, the only part of the pluton with abundant evidence for mixing and mingling. Pyroxene in the central zone is rarely preserved, which makes comparison with pyroxene in the roof dike difficult. Zoned augite crystals in roof-zone andesite (e.g., MMB-164b) show an abrupt increase in Cr contents that suggests mixing with relatively primitive mafic magma. No evidence of such mafic input has been recorded by hornblende in the central part of the intrusion. We therefore conclude that the source of the two pyroxene andesitic magmas was a part of the batholith that is not exposed. This zone of the batholith most probably underlies the southern and southwestern portion of the upper zone.

**Dacitic and Rhyodacitic Roof Dikes**

Dacitic and rhyodacitic roof dike were interpreted as originating from the upper zone magma (Barnes et al., 1986a; Coint et al., 2013) on the basis of their bulk-rock compositions, which overlap compositions of the upper zone rocks (fig. 8 in Coint et al., 2013). In addition, hornblende phenocrysts in the dacitic and rhyodacitic roof dike rocks are compositionally identical to hornblende from the upper zone in terms of major and trace element abundances, and display the same zoning relationships (Fig. 8H). The presence of these dike structurally above the upper zone indicates that the upper zone magma was eruptible. Presumably, the compositional range of dacitic to rhyodacitic roof dike rocks reflects a temporal change in composition of differentiated magmas in the uppermost part of the upper zone.

**IMPLICATIONS FOR THE PLUTON ASSEMBLY AND COMPARISON WITH OTHER SYSTEMS**

Our results indicate that the lower zone resulted from emplacement of many individual magma batches of broadly andesitic composition. The CA-TIMS ages cited here (presented in detail in Coint et al., 2013) show that this zone was emplaced through a short period of time, from 159.22 ± 0.10 Ma to 158.99 ± 0.17 Ma. The predominantly north-south strike and steep dips of internal contacts and foliation in the lower zone are suggestive of batch-wise magma emplacement in subvertical sheets (Fig. 9A; Coint et al., 2013). The range of rock compositions and texture from sample to sample (Figs. 4A, 4B) can be explained by variation in the compositions of individual magma batches, different cooling rates of the individual magma batches, and variable segregation of interstitial melts. Local gradational contacts between textural distinctly rocks indicate that some of the magma batches were significantly above their solidus when adjacent batches were emplaced, and this relationship is exemplified by inheritance of augite from pyroxenitic blocks in adjacent quartz diorite and tonalite.

The andesitic to dacitic upper zone magmas were more buoyant than the rocks and masses in the lower zone and so were emplaced above the lower zone, into rocks of the eastern Hayfork and western Hayfork terranes (Figs. 9B, 9C). Local mixing between the lower zone and upper zone magmas occurred, such as along the western mafic selvage and in the central zone (Fig. 9C). Mixing created hybrid rocks with the compositions of which indicate an affinity with the upper zone. In the meantime, hot basaltic andesitic was emplaced into the central zone, where it locally mingled with the host magma and reheated the system (Fig. 9C). Ages obtained on the central zone are in agreement with the geochemical data presented here, in that they overlap with ages for both the upper and lower zones (Fig. 9 and Coint et al., 2013).

Several studies based on numerical modeling and natural examples have suggested that mixing due to injection of mafic magma into silicic magma is unlikely to be extensive, leaving the mafic magma as a source of heat and fluid (Robinson and Miller, 1999; Bachmann et al., 2002; Huber et al., 2009; Ruprecht et al., 2012). Bargar and Bergantz (2011) demonstrated that the heat generated by a small volume of mafic material emplaced at the base of a crystal-rich silicic magma can cause rapid remobilization of the overlying mush (unzipping), in a few days to several months after the intrusion of the mafic material. This remobilization results in homogenization of the felsic magma due to convection (Robinson and Miller, 1999). If this type of mobilization occurred in the upper zone magma, the scant evidence for resorption in plagioclase, hornblende, and quartz suggests that heating did not rejuvenate a locked mush such as in some large dacitic eruptions (Bachmann et al., 2002), but instead initiated the homogenization of magma having crystallinity low enough to permit free flow (Fig. 9D; Robinson and Miller, 1999).

Hornblende in the central zone, the inferred feeder system of the upper zone, is compositionally nearly identical to hornblende in the upper zone, and hornblende in the upper zone is essentially identical from one sample to the next over a range of rock types and structural levels. These similarities suggest that the broadly tonalitic magmas initially emplaced in the central and upper zones were very similar in bulk composition. Convective mixing inferred for the upper zone did not significantly affect the central zone, because the heterogeneous nature of the central zone (e.g., deformed synplutonic dikes, MME swarms) is preserved. Evidently, these similar tonalitic magmas in the central and upper zones gained their compositional features in structurally deeper levels of the magmatic system.

Convection in the upper zone permitted growth of hornblende and plagioclase phenocrysts from a broadly homogeneous magma body. However, once large-scale convection stopped (Fig. 9E), either due to increased crystallinity or loss of mafic heat input, upward percolation of residual melt through the mush resulted in the broad upward zoning of the upper zone, in a manner similar to that proposed for monotonous ignimbrites (Christiansen, 2005). This process resulted in the lower part of the upper zone becoming a partial cumulate, upward increase in the abundances of quartz and K-feldspar, and development of euhedral quartz. Percolation of residual melt through a mush can also explain why the upward zoning is not systematic (fig. 2 in Coint et al., 2013), because melt percolation would depend on local conditions of porosity and permeability. In contrast, upward zoning due to emplacement of successively more evolved magmas in subhorizontal zones should result in much more systematic zoning (e.g., Bartley et al., 2008). Moreover, emplacement of variably evolved magma batches would not be expected to result in crystallization of hornblende with uniform composition throughout the thickness of the upper zone.

We therefore conclude that the upper zone was once a large, convecting body of chemically and physically interconnected magma that crystallized compositionally very similar hornblende and plagioclase (Fig. 9D). Upward zoning occurred after initial crystallization of hornblende and plagioclase phenocrysts by upward percolation of residual melt (Fig. 9F).

**CONSEQUENCES FOR VOLCANIC SYSTEMS: DID WCB MAGMAS ERUPT?**

The abundance of mafic through felsic dikes in the roof zone (Figs. 2E, 2F) of the batholith leads to the conclusion that some WCB magmas escaped the system and probably erupted. The first magmas to erupt (i.e., roof dikes) are comparable to the lower part of the system, and were mainly basaltic andesitic in composition. Few clear-cut examples exist because many of the oldest mafic roof dikes were intensely altered by thermal effects of emplacement of the
Figure 9. Emplacement model for the Wooley Creek system (WCb—Wooley Creek batholith). Ages reported here are chemical abrasion–thermal ionization mass spectrometry (CA-TIMS) U-Pb zircon ages from Coint et al. (2013). They are presented as the oldest to youngest ages for each step of the pluton assembly accounting for the 2σ error. $P$—pressure. (A) Emplacement of the lower zone as individual magma batches. The black lines between the different magma batches indicate that the interaction between these particular batches with the surrounding ones were limited; however, they do not indicate the presence of a sharp contact in the field. (B) Emplacement of the central zone as individual magma batches. (C) Emplacement of the upper zone and interaction between the upper and lower zone magmas in the western mafic selvage followed by the arrival of the basaltic andesitic magma in the central zone and emplacement of the two-pyroxene andesitic batches at depth. (D) Basaltic andesite emplacement in the central zone triggers convection and homogenization in the upper zone. (E) Development of the upper zone as a large batch of magma of relatively homogeneous composition. (F) Evolution of the upper zone by fractional crystallization with development of a broad upward zoning with more felsic rocks occurring toward the top of the intrusion. Eruption of more two-pyroxene andesitic dikes and emplacement of the late and two-mica granite occurred later. The dashed line is the current level of exposure. RCt—Rattlesnake Creek terrane; eHt—eastern Hayfork terrane, wHt—western Hayfork terrane.
upper zone. However, dike samples such as fine-grained gabbro MMB-590 (Barnes et al., 1990) indicate that lower zone magmas reached high levels of the system and were capable of erupting. In typical volcanic systems, such magmas would have formed scoria cones and lava flows; however, the volume ejected would have been limited (Figs. 9A, 9B). The lateral extent of the lower zone suggests that volcanism was unlikely to be focused at one volcano, but rather formed a volcanic field with several eruptive vents, releasing magmas of variable compositions.

As the dactitic magmas that formed the upper zone arrived in the system, it is possible that these batches erupted, forming part of the roof dike system observed today. The emplacement of several batches of dactitic magma would suggest that eruptions would not necessarily be focused in a single edifice, but probably at several vents. At the same time, batches of two-pyroxene andesite were emplaced underneath the southern part of the intrusion; these magmas rose into the uppermost part of the upper zone and the roof zone to form the two-pyroxene andesite dikes. Arrival of these porphyritic magmas at the surface could have formed thick lava flows (Fig. 9C).

In this model, emplacement of basaltic andesite in the central WCb resulted in convection and homogenization of the upper zone magmas (Figs. 9C–9E). There are numerous examples of dacitic eruptions triggered by the arrival of mafic magma at the base of silicic systems (e.g., Bachmann et al., 2002; Pallister et al., 1992). If the dacitic roof dikes are appropriate examples of the state of the upper zone at the time of eruption, then the percentage of crystals, ~40–50%, is similar to crystal proportions in homogeneous dacitic ignimbrite deposits described throughout the southwestern United States (Hildreth, 1981; Bachman et al., 2002; Christiansen, 2005). The size of the WCb upper zone is not comparable to these large ignimbrite deposits, which represent 1000–5000 km3 of erupted magma, but it could easily have fed eruptions as large as or larger than the 7000 ka culminating eruption of Crater Lake (Bacon and Druitt, 1988) (Figs. 9E, 9F). As volcanism waned, the upper zone magma differentiated by fractional crystallization, developing a felsic cap that could have erupted as rhyolitic domes, capable of producing minor pyroclastic flows.

CONCLUSIONS

On the basis of analogies with modern arc volcanoes, such as in the Cascade Range, the WCb corresponds to the type of magmatic reservoir expected to be beneath typical arc-related composite volcanoes. The first magmatism to occur was mafic to intermediate and represented emplacement of multiple batches of magma in the middle crust, magma batches that underwent very limited homogenization. As a result, individual magma batches can be identified on the basis of augite compositional variation. The central zone of the pluton corresponds to a complex mixing and transition zone, where magmas from the lower zone interacted with the upper zone magmas. It is also a part of the intrusion where replenishment of mafic magmas occurred. The mafic magmas provided heat (and possibly fluids) to the upper part of the system, but interaction of the mafic magmas with the host dactic (upper zone) magmas was limited. The upper zone crystallized from convecting, relatively homogeneous, dactitic magma. Once convection stopped, fractional crystallization occurred by melt percolation through the mush pile, resulting in a broad, nonsystematic upward zoning of the unit with more mafic, partly cumulative rocks in structurally lower parts of the zone and progressively more evolved granodiorite and granite toward the roof. Roof dikes of dacitic composition indicate that the upper zone magmas were eruptible and escaped the underlying magma chamber. Unlike many models for incremental assembly of plutos, we can demonstrate that fractional crystallization was a viable process in the middle to upper parts of the crust and that evolved volcanic rocks need not acquire all their chemical characteristics in the lower crust.

This study demonstrates that mineral trace element chemistry provides an important complement to geochronology in order to better understand the history and pace of batholith emplacement. In particular, the mineral compositions provide a tool to map the extent of ancient magma reservoirs, identify magma batches, and understand their chemical and physical evolution and interactions.

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REFERENCES CITED

Pluton assembly using trace elements in augite and hornblende


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