Time scale for the development of thickened crust in the Cretaceous North Cascades magmatic arc, Washington, and relationship to Cretaceous flare-up magmatism

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ABSTRACT

Heavy rare earth element (HREE)–depleted trace-element patterns are a relatively common feature of granitoids within mature Cordilleran continental margin arcs (e.g., Sierra Nevada batholith, Coast Mountains batholith, North Cascades, Peninsular Ranges batholith). This depletion is commonly interpreted to indicate the presence of garnet during granitoid melt formation, which requires thick arc crust (>40 km) to achieve the necessary pressure conditions to stabilize garnet in the lower crust. This work focused on understanding the evolution of thickened crust in an ancient continental arc using whole-rock geochemical data, high-precision chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb zircon geochronology, and in situ zircon Hf and O isotopic data from three contemporaneous Cretaceous plutons in the North Cascades of Washington State, USA.

New data show that, over time, magmas from three North Cascades plutons became more felsic and more HREE-depleted but lacked a significant change in zircon δ18O. The gradual depletion in HREEs through time is interpreted as evidence for a progressively thickening crust in the North Cascades arc that only became thick enough to stabilize garnet between 90 Ma and 87 Ma. This time frame also coincides with the end of a major period of contraction and plutonism in the region, suggesting a link between thick crust and the end of a major magmatic flare-up. The absence of appreciable change in zircon δ18O values during this time suggests that thickening may have been the result of crustal shortening within the arc, causing migration of the magma source region to below the garnet stability threshold.

INTRODUCTION

Many arc rocks show strongly depleted heavy rare earth element (HREE) trace-element patterns, which are often interpreted to indicate the presence of garnet ± amphibole during melt generation. Garnet may be a residue left behind following partial melting in the deep crust (e.g., Kelemen et al., 2003; Saleebey et al., 2003; Lee et al., 2006), or it may form during fractional crystallization (Green, 1972; Green and Ringwood, 1968; Müntener et al., 2001; Alonso-Perez and Müntener, 2009). Both of these processes require that the lower crust to reach sufficiently high-pressure conditions for garnet to be stable during magma genesis (~1.2 GPa; Alonso-Perez and Müntener, 2009). The nature and timing of the growth of garnet in the lower crust of arcs are still relatively poorly understood but have been studied in the Andean arc (Mamani et al., 2010), where a correlation was found between the development of a garnet signature and crustal thickening. Additional work by Chapman et al. (2015), Chiaradia (2015), and Profeta et al. (2015) found distinct correlations between geochemical signatures (elevated Sr/Y and La/Yb) and crustal thickness; however, the age resolution of crustal thickening was relatively coarse. The relative timing and rates of crustal thickening are important to establish because they provide constraints on mechanisms of crustal thickening (Karlstrom et al., 2014; Chin et al., 2015). This timing is particularly relevant in the North America Cordillera, where crustal thickening has been invoked as the primary trigger for the Cretaceous magmatic flare-ups (Ducaea, 2001; DeCelles et al., 2009; Girardi et al., 2012).

Understanding the triggering mechanisms of flare-up magmatism requires a fuller picture of the timing of physical and chemical changes that occur within an arc as the crust thickens, which in turn requires detailed geochemical and isotopic work on sufficiently well-dated igneous rocks. Several studies of exposed and tilted oceanic arcs suggest that garnet as a fractionating phase in the lower crust dominates the HREE budget, while amphibole exerts primary control on the middle rare earth elements (MREEs; Greene et al., 2006; Jagoutz et al., 2009). These minerals, along with plagioclase, affect the rare earth element (REE) profiles for magmas produced in, or interacting with, the deep crust. Although ideal for focused research, such exposures of complete arc sections are rare. Structural, geochronologic, and geochemical studies of arc granitoids are often limited to a single crustal level within an arc magma system, providing an incomplete window into arc processes. In particular, exposures...
of deep crustal (i.e., >25 km) plutons are uncommon, making it difficult to understand the connection (if any) between deep and shallow crustal magmatism (e.g., Miller and Snoke, 2009, and references therein). In general, plutonic systems are not well dated, typically relying on one or two samples to represent hundreds or thousands of square kilometers. This temporal aspect is critical to establishing the rate of crustal thickening and the changing composition of the lower crust, which are, in turn, critical to understanding the detailed petrogenesis of arc magmas, and especially the triggers for magmatic flare-ups in continental arcs.

The North Cascades of Washington State offer an exceptional field laboratory in which to study magmatism in an ancient continental arc because exposures of igneous rocks extend from the deep crust (>30 km) to the surface. These exposures offer unique snapshots into the workings of a classic Cordilleran magmatic arc at various depths. Several Late Cretaceous plutons in the North Cascades, which intruded during a magmatic flare-up and represent different crustal levels, have been very well dated using isotope dilution–thermal ionization mass spectrometry (ID-TIMS) U-Pb zircon geochronology (Matzel et al., 2006) and chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS; Shea, 2014; Shea et al., 2016; Chan et al., 2017). The high-precision geochronology from these plutons provides a temporal link for new geochemical and isotopic data and published structural information. CA-ID-TIMS is commonly considered the “gold standard” for geochronology and regularly achieves precisions of <0.1%. Here, we used high-precision geochronology along with geochemical and isotopic data to investigate the time scale for the development of a garnet-bearing lower crust in a Cretaceous continental arc near the end of a magmatic flare-up.

GEOLeC SETTING

The crystalline core of the North Cascades (Cascades core) is located at the southern end of the 1500-km-long Coast Plutonic Complex, which stretches from southeast Alaska to northern Washington (Rubin et al., 1990). The core is bounded by the Eocene Straight Creek fault to the west, the Late Cretaceous (–Eocene Ross Lake fault zone (composed of the Hozameen, Foggy Dew, and Ross Lake faults, and the Gabriel Peak fault zone) to the east, and the Cretaceous Windy Pass thrust to the south (Fig. 1; Misch, 1966; Miller, 1985). Early Cretaceous magmatism in the North Cascades region occurred largely to the east of the Cascades core in the Okanogan Range batholith (Hurlow and Nelson, 1993) and in the Chelan Complex, an assemblage of deep-crustal arc rocks with uncertain relation to younger magmatism in the Cascades core (Hopson and Martinson, 1994; Dessimoz et al., 2012). The Cascades core records evidence for mid-Cretaceous crustal shortening, metamorphism, and plutonism followed by Eocene transtension and magmatism (Misch, 1966; Tabor et al., 1989; Hauerger et al., 1991; Miller et al., 2009, 2016; Eddy et al., 2016). These two periods represent episodes of significant magmatic additions to the Cascades core (magmatic flare-ups), when large amounts of new igneous material were added to the region (Miller et al., 2009; Paterson et al., 2011).

This study focused on three contemporaneous Cretaceous plutons in the Cascades core that were intruded during the mid-Cretaceous (96–89 Ma) magmatic flare-up: (1) the shallow-crustal Black Peak intrusive complex, (2) the mid-crustal Seven-Fingered Jack pluton, and (3) the deep-crustal Tenpkeep pluton (Fig. 1). These plutons comprise a variety of shapes, from thinly sheeted to broadly elliptical (Miller et al., 2009), and were generally intruded into previously accreted oceanic and arc terranes (Tabor et al., 1989). The plutons were intruded at a range of paleodepths (<5 km to >35 km; Brown and Walker, 1993; Dawes, 1993; Miller and Paterson, 2001; Miller et al., 2009), offering the ability to investigate contemporaneous magmatism at different crustal levels.

Geology of the Black Peak Intrusive Complex

The Black Peak intrusive complex (Shea et al., 2016) is an elliptical-shaped body emplaced along the eastern edge of the Cascades core (Fig. 1). Two dextral faults form the eastern and western margins of the region: the North Creek fault on the east and the Gabriel Peak fault zone on the west, both of which are considered part of the Ross Lake fault zone (Fig. 2A; Miller and Bowring, 1990; Miller, 1994). Al-in-hornblende barometry on samples from the Black Peak intrusive complex indicates that the eastern margin of the complex crystallized at 0.1–0.3 GPa, which is compatible with the low metamorphic grade of host rocks on the east into which the complex intrudes (Fig. 1; D.L. Whitney and R.B. Miller, personal data from four samples).

Rocks of the Black Peak intrusive complex are dominated by tonalite and granodiorite, but diorite, gabbro, and granite are also present (Adams, 1964; Miller, 1994). Intrusive relationships and geochronology suggest a temporal progression from older mafic intrusions to younger felsic intrusions (Adams, 1964; Dragovich et al., 1997; Shea et al., 2016). The Black Peak intrusive complex is subdivided into four units based on textural and compositional characteristics. Listed from oldest to youngest, along with their zircon U-Pb CA-TIMS dates and 2σ analytical uncertainty, these units are the Crescent Mountain unit (91.755 ± 0.040 Ma to 91.18 ± 0.12 Ma), the Stiletto Mountain unit (90.345 ± 0.030 Ma to 88.81 ± 0.30 Ma), the Reynolds Peak unit (88.467 ± 0.031 Ma to 87.765 ± 0.072 Ma), and the War Creek unit (87.474 ± 0.052 Ma to 86.862 ± 0.06 Ma; Fig. 2A; Shea et al., 2016). Isotopic data from the Black Peak intrusive complex (Matzel et al., 2008; Shea et al., 2016) have a very limited range: εNd(t) whole-rock values are between +5.14 and +6.3 and zircon δ18O values are between 6% and 7%. Both isotopic data sets indicate contributions from isotopically juvenile sources such as subarc asthenospheric mantle and/or isotopically immature arc crust (e.g., the Napeequa/Twisp Valley schist, metamorphosed Methow basin sediments; Shea et al., 2016).

Geology of the Seven-Fingered Jack Pluton

The Seven-Fingered Jack pluton is an elongate intrusion emplaced east of the Entiat fault (Figs. 1 and 2B). The pluton intrudes the Napeequa unit (Napeequa schist of Cater and Crowder, 1967), which is in part Triassic in age (Sauer et al., 2017), and the Triassic Dumbrell orthogneiss on its western margin (Fig. 2B; Cater and Crowder, 1967; Cater and Wright, 1967; Tabor et al., 1989). On its eastern margin, the Seven-Fingered Jack pluton intrudes the Dumbrell orthogneiss and Triassic Holden assemblage (Fig. 2B; Cater and Crowder, 1967; Cater and Wright, 1967; Tabor et al., 1989). The southern margin of the Seven-Fingered Jack pluton is intruded by the Entiat pluton along a poorly mapped contact (Fig. 2B). Al-in-hornblende barometry from several locations in the Seven-Fingered Jack pluton suggests pressures of 0.6–0.7 GPa, indicating crystallization depths of 20–25 km (Dawes, 1993). These data are consistent with pressures of 0.7 GPa (Garnet-Aluminosilicate-Silica-Plagioclase (GASP) at 650 °C from a sample of metapelitic Napeequa unit collected adjacent to the Seven-Fingered Jack pluton (Valley et al., 2003). Extensive field work by Paterson and Miller (1998) and Miller and Paterson (2001) indicated that the Seven-Fingered Jack pluton is composed of multiple sheets of compositionally variable rock.

The Seven-Fingered Jack pluton is divided into three units, all of which are dominated by tonalite (Fig. 2B; Miller and Paterson, 2001; Matzel, 2004; Shea, 2014): Blue Creek unit (ca. 91.7 Ma to 90.5 Ma), the Cougar Mountain unit (ca. 90 Ma), and the Whistling Pig unit (ca. 88 Ma). Directly east of the Seven-Fingered Jack pluton, there is the Kelly Mountain pluton (ca. 78 Ma), an assemblage of tonalite, diorite, gabbro,
Figure 1. Crystalline core of the North Cascades. Numbers in plutons are crystallization ages. BP—Black Peak intrusive complex; BR—Bearcat Ridge orthogneiss; CH—Chaval pluton; CS—Chiwaukum schist; DF—Dirtyface pluton; GPTZ—Gabriel Peak tectonic zone; HP—High Pass pluton; MC—Marble Creek pluton; NCF—North Creek fault; NQ—Napeequa unit; RC—Ruby Creek plutonic belt; RRC—Railroad Creek pluton; RLF—Ross Lake fault; RLFZ—Ross Lake fault zone; RP—Riddle Peaks pluton; SC—Sloan Creek plutons; SFJ—Seven-Fingered Jack pluton; SM—Sulfur Mountain pluton; TP—Tenpeak pluton; TVS—Twisp Valley schist; WPT—Wind Pass thrust; WRG—Wenatchee Ridge gneiss. Regional fold traces and major highways are shown. Time periods: C—Cenozoic; J/Jur—Jurassic; K—Cretaceous; P—Permian; Pg—Paleogene; Tr—Triassic.
Figure 2. (A) Geologic map of the Black Peak intrusive complex (Shea et al., 2016). (B) Geologic map of the Seven-Fingered Jack pluton. (Continued on following page.)
and hornblende that is interfingered with the Seven-Fingered Jack pluton (Dustin, 2015). Tonalites in the Kelly Mountain pluton are impossible to differentiate from those in the Seven-Fingered Jack pluton in the field; because of this, we grouped rocks from these plutons together as the “Seven-Fingered Jack pluton.”

Geology of the Tenpeak Pluton

The ca. 92–89 Ma Tenpeak pluton (Cater, 1982; Miller and Paterson, 1999; Matzel et al., 2006; Chan et al., 2017; Miller et al., 2018) is an irregularly shaped, elongate body that also intrudes the Napeequa unit (Figs. 1 and 2C). The southern margin of the Tenpeak pluton intrudes the Chiwaukum Schist; however, this contact is in part obscured by the NE-dipping, reverse-slip White River shear zone (Fig. 2C; VanDiver, 1967; Magloughlin, 1986). Al-in-hornblende barometry (Brown and Walker, 1993; Dawes, 1993; Miller et al., 2009) and the presence of migmatic epidote (Zen, 1988) imply crystallization pressures around 0.7–1.0 GPa, suggesting an emplacement depth of ~20–35 km.

Geochronological and isotopic work by Dawes (1993), DeBari et al. (1998), Matzel et al. (2008), Chan et al. (2017), and Miller et al. (2018) suggests that rocks in the Tenpeak pluton were derived from mixing between mantle-derived mafic magma and crustally derived felsic magma (Miller et al., 2018). Noncumulate rocks of the Tenpeak pluton span a range of mantle-derived mafic magma and crustally derived felsic magma (Miller et al., 2009). Stars on A and B represent sample locations. Stars on C represent geochronologic sample locations (Shea, 2014; Chan et al., 2017).

ANALYTICAL METHODS

Whole-rock samples collected as part of this study were analyzed for major and trace elements at the Washington State University Geo-Analytical laboratory using a ThermoARL Advant’XP sequential X-ray fluorescence spectrometer (XRF) and Agilent 7700 inductively coupled plasma–mass spectrometer (ICP-MS). A detailed description of the sample preparation and XRF analysis following a single low-dilution Li-tetraborate fused bead procedure was described by Johnson et al. (1999). The samples analyzed using ICP-MS followed procedures described by Lichte et al. (1987), Jarvis (1988), and Longerich et al. (1990). The precision for REE and other trace elements was 5% and 10%, respectively.

Oxygen isotopes in zircon were measured from a subset of samples from the Seven-Fingered Jack intrusive complex and Tenpeak pluton. Zircons selected for in situ oxygen (O) and hafnium (Hf) isotopic analysis were mounted in epoxy, polished to approximately half their thickness, and imaged using a cathodoluminescence (CL) detector on a scanning electron microscope. The imaging for the zircon crystals used for the O analyses was conducted at the University of California–Los Angeles.
were analyzed using the UCLA CAMECA ims 1270 high-resolution ion microprobe. All analyses were conducted during a single session; settings and conditions were similar to those described in the supplementary data in Schmitt (2006). Our data are reported relative to the R33 internal standard (5.55±0.04%; Valley, 2003).

Zircon Hf isotope ratios were measured at the MUN Micro-Analysis Facility. We followed the methods of Fisher et al. (2011), with the exception that N2 was added to the Ar carrier gas for increased sensitivity. The analyses were done using a Thermo Finnigan multicollector (MC) ICP-MS interfaced to a Geolas Pro 193 nm Ar-F excimer laser, operating at 10 Hz, 5 J/cm², and with a 50-μm-diameter spot size. The analyses on the Cascades zircon crystals were interspersed with secondary reference zircon samples Plešovice, MUN/Zirc 2 and MUN/Zirc 4, and R-33, with measured ratios agreeing (within 1 standard deviation [SD]) with published solution (S)-MC-ICPMS isotopic compositions (see Fisher et al., 2011). The cup configuration, analytical methodology, and data reduction protocol were described in more detail by Fisher et al. (2011). Initial εHf values were calculated using a chondritic uniform reservoir (CHUR) value of [187Hf/187Hf]0 = 0.282785, as reported by Bouvier et al. (2008).

RESULTS

Major- and Trace-Element Chemistry

Whole-rock major- and trace-element data are summarized here for: 14 samples from the Black Peak intrusive complex (11 new analyses and 3 analyses from Chan, 2012); 16 samples from the Seven-Fingered Jack pluton (2 new analyses, 6 analyses from Dustin, 2015, and 8 analyses from Elkins, 2015); and 75 samples of the Tenpeak pluton (19 analyses from Chan et al., 2017, and 56 analyses from Miller et al., 2018). The major- and trace-element results and the coordinates for sample locations are given in the GSA Data Repository Item (Table S1).

Samples of igneous rocks from all three intrusions are classified as calc-alkaline (Fig. S1 [footnote 1]; Miyashiro, 1974) and fall within the diorite-tonalite-granodiorite-granite suite using the classification of Cox et al. (1979) (Fig. S2 [footnote 1]). They are metaluminous to slightly peraluminous with a range in Al2O3 (13–21 wt%), with one outlier at 8 wt%; Fig. S3 [footnote 1]; Shand, 1943). SiO2 varies from 45.6 to 73.8 wt%, and all three intrusions show decreasing TiO2, FeO, MgO, and CaO and increasing K2O with increasing SiO2 (Fig. 3). Al2O3 and P2O5 show considerable scatter below 55 wt% SiO2 and a general decreasing trend above 55 wt% SiO2.

Trace-element data from samples of the Black Peak intrusive complex, Seven-Fingered Jack pluton, and Tenpeak pluton have primitive mantle–normalized trace-element patterns characterized by light rare earth element (LREE) enrichment and HREE depletions (Fig. 4). Most samples lack a significant positive or negative Eu anomaly. Samples have an overall enrichment in large ion lithophile elements (LILs), depletion of Nb and Ta relative to U and La, and pronounced positive Pb anomalies.

Black Peak Intrusive Complex

Although the majority of the Black Peak intrusive complex is composed of tonalite and granodiorite, the samples studied span a considerable range in composition, and SiO2 varies accordingly between 51 and 72 wt%. MgO varies between 1 and 6 wt%. Black Peak samples have a more limited range of Al2O3 (15–17 wt%) than the other two plutons. Harker diagrams of the major oxides (e.g., MgO, CaO, FeO*) form a linear trend, although K2O shows appreciable scatter (Fig. 3).

Trace elements from the Black Peak intrusive complex are characteristic of arc rocks and show enrichment in LILs and depletion in high field strength elements (HFSEs; Fig. 4A). Black Peak samples display moderate to steep REE patterns ([La/Yb]N = 5.2–21.3; Figs. 5A and 6A). Dy/Yb is high (1.25 < [Dy/Yb]N < 2; Fig. 6B). [La/Yb]N and [Dy/Yb]N increase with SiO2 (Figs. 6A and 6B). LREE/MREE is generally high (1.2 < [La/ Nd]N < 2.5) and broadly increases with SiO2 (Fig. S4 [footnote 1]). MREE/HREE varies from moderate to high (1.5 < [Gd/Yb]N < 3.5) and increases with SiO2 (Fig. S5 [footnote 1]). Samples are characterized by small Eu anomalies that range from positive to negative (Fig. 5A). Sr/Y generally increases with SiO2 and varies between ~20 and 180 (Fig. 6C).

Seven-Fingered Jack Pluton

Similar to Black Peak intrusive complex, most of the Seven-Fingered Jack pluton is tonalitic. However, a wide range of compositions was observed; SiO2 varies from 50 to 70 wt%, and MgO varies from 1 to 6 wt%. Al2O3 ranges from 16 to 21 wt%, although the majority of analyses are between 16 and 18 wt%. Harker diagrams of the major oxides (e.g., MgO, CaO, FeO*) show a linear trend, but TiO2, Na2O, K2O, and P2O5 show considerable scatter (Fig. 3).

Also, like the Black Peak intrusive complex, rocks from the Seven-Fingered Jack pluton are enriched in LILs and depleted in HFSEs (Fig. 4B). REE patterns are moderate to steep ([La/Yb]N = 2.6–9.7; Figs. 5B and 6A). Dy/Yb is moderate (1.4 < [Dy/Yb]N < 1.5; Fig. 6B). Both [La/Yb]N and [Dy/Yb]N decrease with SiO2 (Figs. 6A and 6B). LREE/MREE is generally moderate (0.9 < [La/Nd]N < 2; Fig. S4 [footnote 1]) and has no clear relationship with SiO2. MREE/HREE varies from moderate to high (1.8 < [Gd/Yb]N < 2.9; Fig. S5 [footnote 1]) and has no clear relationship with SiO2. Some samples have slightly negative Eu anomalies, while others have none (Fig. 5B). Unlike data from the Black Peak intrusive complex and the Tenpeak pluton, Sr/Y data from the Seven-Fingered Jack pluton show no clear relationship with SiO2 and vary between ~20 and 40 (Fig. 6C).

Tenpeak Pluton

Samples from the Tenpeak pluton span a wide compositional range: from 50 to 73 wt% SiO2 and 1–8 wt% MgO. Al2O3 ranges from 14 to 20 wt%. Harker diagrams show broadly linear trends for all the major oxides (e.g., MgO, CaO, FeO*), excepting Al2O3, which has considerable scatter (Fig. 3; Miller et al., 2018).

As summarized by Miller et al. (2018), trace elements in the Tenpeak pluton are enriched in LILs and depleted in HFSEs (Fig. 4C). REE patterns are generally moderate to steep (2.6 < [La/Yb]N < 24.9), with some very steep outliers ([La/Yb]N > 30; Figs. 5C and 6A). Dy/Yb is moderate to high (1.3 < [Dy/Yb]N < 2.4; Fig. 6). [La/Yb]N and [Dy/Yb]N increase with SiO2 (Figs. 6A and 6B). LREE/MREE is moderate to high (0.8 < [La/Nd]N < 4.7; Fig. S4 [footnote 1]) and generally increases with increasing SiO2. MREE/HREE varies from moderate to very high (1.5 < [Gd/Yb]N < 5.2; Fig. S5 [footnote 1]) and broadly increases with SiO2. Eu anomalies are generally absent or slightly positive (Fig. 5C). Sr/Y generally increases with SiO2 and varies between ~20 and 100 (Fig. 6C).

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1 GSA Data Repository Item 2018356, Supplementary text and figures, and data tables, is available at http://www.geosociety.org/datarepository/2018, or on request from editing@geosociety.org.
Oxygen and Hafnium Isotope Data

New in situ zircon oxygen isotopic data are presented here from six samples of the Tenpeak pluton and two samples of the Seven-Fingered Jack pluton, and these are compared to five published analyses from the Black Peak intrusive complex (Shea et al., 2016). New in situ zircon Hf isotopic data are presented from four samples of the Black Peak intrusive complex.

Zircon oxygen isotope data were obtained from representative samples of each of the three plutons (Fig. 7). Multiple (6–12) zircon crystals were analyzed for each sample, and some zircons were analyzed in more than one location. All data are presented in Figure 7 and the Data Repository (Table S1 [footnote 1]). Analyses of zircons from five samples from the Black Peak intrusive complex are summarized from Shea et al. (2016). Average zircon δ¹⁸O values from Black Peak intrusive complex samples range from 6.5‰ (MAF) to 7.0‰ (PX10–86). Average zircon δ¹⁸O values from the two samples of the Seven-Fingered Jack pluton are 6.4‰ and 7.7‰. Average zircon δ¹⁸O values from six samples of the Tenpeak pluton range from 7.3‰ (PX10–243A) to 8.2‰ (IC2), although most samples have values between 7.5‰ and 8‰ (Fig. 7). Uncertainties (1 SD) for all data points are 0.55‰.

Zircon Hf isotopes were obtained from representative zircons from four samples of the Black Peak intrusive complex. Multiple (4–7) zircon crystals were analyzed for each sample, and some zircons were analyzed in more than one location (where it was clear in the CL images that there was no inherited core). All data are presented in Figure S6 and the Data Repository (Table S1 [footnote 1]). Average initial ε⁵⁷Hf values from these samples range from ε⁵⁷Hf = 9.26 (PX10–236) to ε⁵⁷Hf = 8.11 (PX10–221), although individual spot analyses vary between ε⁵⁷Hf = 10.10 and ε⁵⁷Hf = 7.04 (Fig. S6 [footnote 1]). No Hf isotopic data were obtained from the Seven-Fingered Jack or Tenpeak plutons.

DISCUSSION

The Black Peak, Seven-Fingered Jack, and Tenpeak intrusions provide three views of contemporaneous magmatism in the same arc. Each
Figure 4. Primitive mantle–normalized (Sun and McDonough, 1989) multi-trace-element diagram of the studied samples. (A) Black Peak intrusive complex. (B) Seven-Fingered Jack pluton. (C) Tenpeak pluton.

Figure 5. Primitive mantle–normalized (Sun and McDonough, 1989) rare earth element (REE) diagrams of the studied samples. (A) Black Peak intrusive complex. (B) Seven-Fingered Jack pluton. (C) Tenpeak pluton.
intrusion was emplaced at a different crustal level, ranging from relatively shallow (Black Peak intrusive complex), to intermediate (Seven-Fingered Jack pluton), to deep (Tenpeak pluton). Major elements in these plutons show a broad trend toward more-evolved compositions (e.g., higher silica, lower FeO*, lower MgO) in progressively shallower plutons (Fig. 3). All three plutons generally become more felsic and homogeneous with time (Fig. 8). Because these plutons are not linked spatially (they are separated by ~60 km across strike; Fig. 1), it is impossible to know whether they were connected at depth to a broader melt source region. However, their major-element geochemistry is suggestive of a connected melt source region for all three plutons (Fig. 3).

REE Patterns in North Cascades Plutons

The most-striking feature in our geochemical data set is a significant depletion in HREE in samples from the Black Peak intrusive complex and the Tenpeak pluton (Figs. 5A and 5C). Samples from the Seven-Fingered Jack pluton show a moderate HREE depletion that is less pronounced than in the other intrusions (Fig. 5B). HREE depletion is commonly associated with the presence of amphibole and/or garnet in the lower crust (e.g., Mamani et al., 2010; Profeta et al., 2015). Amphibole and/or garnet in the deep crust may form as a restite caused by partial melting in the deep crust (e.g., Hildreth and Moorbatch, 1988; Hanchar et al., 1994) or as a cumulate from fractional crystallization (e.g., Müntener et al., 2001; Alonso-Perez and Müntener, 2009). In either case, the lower crust of the arc must reach high-pressure conditions in order for garnet to stabilize (~1.2 GPa; Alonso-Perez and Müntener, 2009). It is possible to determine the presence of garnet in the melt region by using trace-element geochemistry: Magmatic equilibration with garnet during partial melting or fractional crystallization generally increases both La/Yb and Dy/Yb versus SiO₂, while equilibration with amphibole generally increases La/Yb versus SiO₂ and decreases Dy/Yb versus SiO₂ (Macpherson et al., 2006; Davidson et al., 2007). Note also, that for Black Peak and most Tenpeak samples, the lack of an Eu anomaly is also consistent with no appreciable role for plagioclase in either fractionation or melting.

Both the Black Peak intrusive complex and the Tenpeak pluton show a positive correlation between La/Ybₙ and SiO₂, but the Seven-Fingered Jack pluton has considerable scatter (Fig. 6A). The Black Peak intrusive complex shows a positive correlation between Dy/Ybₙ and SiO₂ for the entire data set; however, close inspection suggests a negative correlation between Dy/Ybₙ and SiO₂ until ~66 wt% SiO₂, at which point the correlation becomes distinctly positive (Fig. 6B). The Tenpeak pluton has a more scattered but overall positive correlation between Dy/Ybₙ and SiO₂ (Fig. 6B). No obvious correlation between Dy/Ybₙ and SiO₂ is observed for the Seven-Fingered Jack pluton (Fig. 6B).

The La/Ybₙ and Dy/Ybₙ versus SiO₂ data from the Black Peak intrusive complex suggest that samples with <66 wt% SiO₂ may have solidified from magmas that equilibrated with amphibole (although this conclusion is based on a limited data set), and that more-evolved samples (>66 wt% SiO₂) likely solidified from magmas that equilibrated with garnet (Fig. 6A). La/Ybₙ and Dy/Ybₙ versus SiO₂ data from the Tenpeak pluton are less definitive, but they also indicate that magmas equilibrated with garnet during fractionation and/or partial melting (Miller et al., 2018). In contrast, REE data from the Seven-Fingered Jack pluton suggest equilibration of magmas with hornblende.

High-precision CA-ID-TIMS geochronology on the same samples of the Black Peak intrusive complex that were used for whole-rock analyses allows us to also track these geochemical shifts temporally, indicating that the garnet signature in the complex appeared after 90 Ma (Fig. 9). Although we do not have geochronologic data for the entire Tenpeak
pluton sample suite, high-precision CA-ID-TIMS geochronology on a modest number of samples suggests that this signature evolved similarly (Fig. 9), and thus we infer that at the arc scale, the garnet signature becomes more pronounced after 90 Ma.

**Isotopic Evidence for Crustal Melt**

Zircon Hf isotopic data provide additional insight into the nature of crustal thickening in the North Cascades; because the mantle is isotopically different from crustal rocks, these data allow an investigation into magma source material (crustal vs. mantle input). The $\delta^{18}$O values for zircons from the Tenpeak pluton show no obvious trend between ca. 92 Ma to ca. 89 Ma (Fig. 7). One sample from the Seven-Fingered Jack pluton was analyzed for $\delta^{18}$O in zircon and has a value similar to that found in the Tenpeak pluton (Fig. 7). The $\delta^{18}$O values for zircons from the Black Peak intrusive complex all overlap each other within uncertainty, although post–90 Ma zircons show a slight shift to higher average $\delta^{18}$O. Hafnium isotopic analyses on zircons from the Black Peak intrusive complex all overlap within uncertainty but show a $\epsilon_{Hf}$ unit negative shift in the average $\epsilon_{Hf}$ (from an average of approximately +9.2 to an average of +8.2) after 90 Ma (Fig. S6 [footnote 1]). These values require a source that was relatively immature, but they are less than expected for pristine Late Cretaceous depleted mantle (+12 to +16; Vervoort and Blichert-Toft, 1999). There is a fairly striking contrast in zircon $\delta^{18}$O values between the Tenpeak and Black Peak intrusions, with the Tenpeak rocks being isotopically heavier in O isotopic composition. Zircon $\delta^{18}$O in all samples and zircon $\epsilon_{Hf}$ from the Black Peak intrusive complex indicate that magmas in the three intrusions are not “pure” mantle-derived melts. Using simple binary mixing models based on oxygen isotopes, it is possible to estimate the relative contributions from crustal and mantle components, assuming $\delta^{18}$O in zircon approximates the $\delta^{18}$O value in the magma from which it crystallized. Using a value of 15‰, which is near the upper range of values for the Chiwaukum Schist (host rock to the southwestern Tenpeak pluton; Anderson et al., 2012) and similar to typical altered oceanic crust values (Lackey et al., 2005), and a canonical mantle $\delta^{18}$O of 5.3‰ (Valley, 2003), we estimate between 20% (PX10–243A) and 30% (TP30) crustal component in the Tenpeak pluton and between 10% (MAF-1) and 20% (PX10–86) in the Black Peak intrusive complex (Shea et al., 2016). We note, however, that these are conservative estimates, because the assimilant could be isotopically lighter, and the host rocks of these intrusions are generally immature accreted oceanic and arc rocks.

The contrast in zircon $\delta^{18}$O values between the Black Peak and Tenpeak intrusions indicates either that melt sources for the Black Peak intrusive complex and Tenpeak pluton were isotopically different, or that isotopically similar primary arc magmas assimilated isotopically distinct contaminants. The principal exposed host rocks into which both plutonic complexes were intruded are compositionally similar (Fig. 1; Chiwaukum Schist/Napeequa unit and the correlative Twisp Valley schist), suggesting the latter explanation is less likely. Although part of the Black Peak intrusive complex intruded into green-schist-facies metamorphosed sediments of the Methow basin along its eastern margin, there is limited field evidence for assimilation of these host rocks into the intruding pluton at the level of emplacement. It is unlikely that the small volume of sediments assimilated along this contact significantly contributed to the isotopic signature of the entire complex. The isotopic differences may reflect (1) isotopic domains in the melt sources that varied within the arc depending on the composition and lithology of the deep crust, or (2) more effective assimilation and partial melting of host rock in the deep crust (Tenpeak pluton) than at shallow levels (Black Peak intrusive complex).

Although there is a slight shift in zircon Hf and O isotopic values with time in the Black Peak samples as noted above, the data set as a whole for both the Tenpeak and Black Peak rocks suggests that the contribution from crustal sources shifted little over the lifetime of both plutons. Perhaps most important for this study is that the zircon isotopic data show that magmatic differentiation involved the crust throughout the lifetime of these intrusions. No zircons analyzed had $\delta^{18}$O and $\epsilon_{Hf}$ values indicative of pristine mantle ($\delta^{18}$O ≈ 5.3‰; Valley et al., 2003; $\epsilon_{Hf}$ ≈ +12–16; Vervoort and Blichert-Toft, 1999), and thus the shift in trace-element patterns with time must reflect changes resulting from processes in the deep crust—remelting of arc crust or deep-crustal assimilation accompanying fractional crystallization.

Figure 7. High-precision chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb zircon age (Ma) vs. zircon $\delta^{18}$O for samples from the Black Peak intrusive complex (circles), the Seven-Fingered Jack pluton (triangles), and the Tenpeak pluton (diamonds). Each suite of symbols represents a distinct sample. Lighter symbols represent $\delta^{18}$O for a single spot on a zircon; darker symbols with a black border represent the average $\delta^{18}$O for each sample. 1σ errors for individual spot analyses are <0.1‰. Dashed line indicates mantle value (5.3‰; Valley, 2003). Asterisk indicates sample where one outlier at 10‰ was not included in the average $\delta^{18}$O calculation. Zircon geochronology is from Shea (2014), Shea et al. (2016), and Chan et al. (2017). Typical uncertainty of sample ages is less than 100 k.y.
Figure 8. Variation diagrams of high-precision chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb zircon age (Ma) vs. whole-rock major elements: (A) SiO₂, (B) TiO₂, (C) Al₂O₃, (D) FeO* (E) MgO, (F) CaO, (G) Na₂O, (H) K₂O, and (I) P₂O₅. Samples from the Black Peak intrusive complex and Seven-Fingered Jack pluton were individually dated. Samples from the Tenpeak pluton are grouped based on the age of the unit from which they were collected. Symbols are described in legend. Zircon geochronology is from Shea (2014), Shea et al. (2016), and Chan et al. (2017). Typical uncertainty of sample ages is less than 100 k.y.
Evidence for Late, Thick Crust in the North Cascades Arc

Several recent studies have correlated crustal thickness in magmatic arcs with trace-element ratios; in particular, increasing Sr/Y in arc rocks has been linked to an increase in crustal thickness in the North American Cordillera during the Mesozoic (Schwartz et al., 2014; Chapman et al., 2015; Profeta et al., 2015), in Mesozoic Fiordland intrusions (Schwartz et al., 2017), in the Central Andes during the Cenozoic (Profeta et al., 2015), and in a global data set of Pliocene–Quaternary arcs (Chiaradia, 2015).

Sr/Y values in our data set are positively correlated with SiO$_2$ (Fig. 6C) and with time (Fig. 9C) but are negatively correlated with Y (Fig. S7 [footnote 1]). We interpret these correlations and the shift in REE patterns (Fig. 5) and key REE ratios (La/Yb$_N$ and Dy/Yb$_N$; Figs. 6B and 6C) as evidence that the arc matured and thickened during the time interval over which these intrusions were emplaced, and that the deep crust eventually reached pressures that stabilized garnet (Müntener et al., 2001; Alonso-Perez and Müntener, 2009).

Crustal thickening can be caused by underthrusting of the retroarc lithosphere (e.g., DeCelles et al., 2009), imbrication and shortening of the forearc (e.g., Sauer et al., 2017), intra-arc crustal shortening and magmatic thickening (e.g., Jagoutz and Schmidt, 2013; Cao et al., 2016), and/or reactivation (e.g., Hacker et al., 2011). Although a combination of these processes may have operated in the North Cascades, evidence suggests that crustal thickening was dominated by regional NE-SW horizontal shortening (e.g., Misch, 1966; Brandon et al., 1988; Miller et al., 2009). This evidence includes thermobarometry of host rocks for a number of 96–90 Ma plutons that indicate pressure increases of up to 0.5 GPa (Brown and Walker, 1993; Miller et al., 1993; Whitney et al., 1999) along with contact aureoles showing evidence for andalusite overprinted by kyanite. Geothermobarometry also indicates that supracrustal rocks were buried to depths of 25–40 km in many parts of the arc (Whitney et al., 1999; Valley et al., 2003).

Regional-scale shortening in the North Cascades has been linked to the collision of the Insular superterrane with the Intermontane superrteraine (Rubin et al., 1990) between ca. 100 Ma and 90 Ma, contemporaneous with the mid-Cretaceous magmatic flare-up in the region between 96 and 89 Ma (Miller et al., 2009). This shortening was expressed at shallow levels by SW-directed thrusting and at deep levels by ductile folding, cleavage development, SW-directed reverse shear, and subhorizontal orogen-parallel stretching (Miller et al., 2006). Recent U-Pb and Hf isotope work on detrital zircons from the metasedimentary host rocks of the Tenpeak, Seven-Fingered Jack, and Black Peak intrusions (Sauer et al., 2017) suggested that thickening involved imbrication and deep thrusting of these sequences in the arc at approximately the same time as emplacement of the intrusions. The development of a depleted HREE signature and the pronounced shift in Sr/Y and La/Yb$_N$ ratios in the Black Peak intrusive complex and the Tenpeak pluton can be interpreted as evidence for the stabilization of garnet in the lower crust of the arc beginning around 90 Ma, near the end of both the magmatic flare-up and a major period of shortening in the North Cascades (Fig. 10). Post–89 Ma intrusions in the
Black Peak intrusive complex have progressively more depleted HREE signatures, suggesting further development of garnet in the lower crust, even as magmatic activity waned (Fig. 10).

Garnet in the lower crust may be the product of fractional crystallization of mantle-derived melts where garnet is stable (e.g., Müntener et al., 2001; Alonso-Perez and Müntener, 2009) or may be due to the burial and subsequent partial melting of preexisting (possibly coenogenetic) lower-crustal rocks in the garnet stability field. These data support a model wherein a major episode of horizontal crustal shortening (between ca. 100 Ma and 90 Ma) and a contemporaneous magmatic flare-up (96–89 Ma) caused crustal thickening that shifted the melt source to pressures high enough to stabilize garnet. The lack of pronounced variation in zircon \( \delta^{18}O \) and \( e_{Hf} \) with time suggests that the isotopic composition of magma did not change significantly while the crust was thickening. In each pluton, magma produced during the flare-up (pre–90 Ma) is not appreciably different isotopically from magma produced after the end of the flare-up (post–90 Ma), although the distinctly different zircon \( \delta^{18}O \) values between the Black Peak and Tenpeak samples may reflect isotopic provinciality in melt source regions of the two intrusions. The absence of any marked change in zircon \( \delta^{18}O \) and \( e_{Hf} \) suggests that intra-arc shortening and magmatic addition may have been responsible for crustal thickening (e.g., Jagoutz and Schmidt, 2013; Cao et al., 2016; Sauer et al., 2017). This thickening caused the magma source region to migrate to depths below the stability threshold for garnet, resulting in the production of melts with depleted HREE signatures. The appearance of the garnet signature shortly (<1 m.y.) before the cessation of the mid-Cretaceous flare-up in the North Cascades requires that shortening was concurrent with the flare-up, reaching maximum crustal thickness (i.e., garnet stability) toward the terminal stages of the flare-up (Fig. 10). Karlstrom et al. (2014) and Chin et al. (2015) elucidated how thickening may result in the cessation of arc flare-up magmatism, rather than serving as a trigger (e.g., DeCelles et al., 2009), and it is intriguing to postulate, given the timing constraints presented here, that thickening into the garnet stability field may have ultimately led to the cessation of the Cretaceous flare-up in the North Cascades around 89 Ma.

Based on work in the Early Cretaceous Chelan Complex of the North Cascades, Dessimoz et al. (2012) suggested that hornblende fractionation played an important role in the production of peraluminous, high-Sr/Y tonalitic melts in the arc. However, existing U-Pb geochronologic data suggest that the Chelan Complex represents an earlier period of magmatism (possibly much earlier) that preceded the peak magmatic flare up (96–89 Ma) in the North Cascades (Mattinson, 1972; Hopson and Mattinson, 1994; Mattinson et al., 2014; Miller et al., 2009). In general, our data are not inconsistent with an earlier role for hornblende fractionation, but we maintain that our geochronologic and geochemical data clearly document a role for garnet fractionation that was coincident with intra-arc shortening and the flare-up in the North Cascades.

Data from younger (post–flare-up) intrusions in the North Cascades, including a small data set from the 72 Ma Cardinal Peak pluton (Parent, 1999), suggest that this garnet signature (and by implication, thicker crust) may have persisted for over 15 m.y. Late Cretaceous (100–80 Ma) rocks of the Coast plutonic complex in central British Columbia also show a similar garnet signature (Girardi et al., 2012) that is broadly synchronous with the major flare-up of magmatism there, suggesting that thickening affected a large area of the arc.

CONCLUSIONS

Data from several Cretaceous plutons support the origin of the North Cascades in a continental magmatic arc built on juvenile arc lithosphere and show a general trend toward more-evolved magmas with time and with decreasing depth. Elevated La/Yb\(_n\), Dy/Yb\(_n\), and Sr/Y ratios in intrusions younger than 90 Ma indicate the presence of garnet in the lower crust following ~10 m.y. of regional shortening, crustal thickening, and magmatism associated with the collision of the Insular and Intermontane superterranes. The lack of evidence for any significant change in \( \delta^{18}O \) values in the Black Peak intrusive complex or the Tenpeak pluton indicates that this was not caused by melting of deeply buried crustal material, but instead it was related to crustal thickening, which was likely due to regional shortening and magmatism. This thickening may have contributed...
to the waning and eventual cessation of the Cretaceous flare-up in the North Cascades around 89 Ma.

ACKNOWLEDGMENTS

Samuel Bowring’s impact on this paper is unseen but felt in many ways. We are grateful for his long-standing enthusiasm and support of research in the North Cascades. Careful and thoughtful reviews by Jay Campbell, Dan Condie, and an anonymous reviewer significantly improved the manuscript. This paper benefited from conversations with Seth Burgess, Frederick Frey, Tim Grove, Oli Jagoutz, and Erin Todd. Analytical assistance from Jahan Ramezani, Nilanjan Chatterjee, Frank Dudás, Rita Economos, Axel Schmitz, and Rebecca Lam is gratefully acknowledged. Support is gratefully acknowledged from the Methow Valley Ranger District, Entiat Ranger District, and the North Cascade National Park for access. This research was supported by National Science Foundation grants EAR-0948338 to S. Bowring, EAR-0948685 to J. Miller and R. Miller, and EAR-1119358 to R. Miller.

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