Magmatism as a response to exhumation of the Priest River complex, northern Idaho: constraints from zircon U-Pb geochronology and Hf isotopes

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ABSTRACT

Zircon and monazite U-Pb geochronology and zircon Hf isotopes place constraints on the temporal and source relationships between crustal anatexis, magmatism, and exhumation of the Priest River metamorphic core complex, northern Idaho. Granitoids that intruded the migmatitic, pelitic Hauser Lake gneiss include the $<76.5 \pm 0.1$ Ma Spokane granite, 50.13 ± 0.02 Ma Silver Point quartz monzonite, *c.* 47.9 Ma Wrencoe granodiorite, $<46.4 \pm 1.8$ Ma Rathdrum granite, and a $<49.8 \pm 0.4$ Ma leucocratic dike. Cretaceous magmatism preceded the *c.* 64 Ma peak metamorphism (recorded by monazite) of the Hauser Lake gneiss, whereas discrete pulses of Eocene magmatic activity post-date the onset of exhumation by 10 Ma. The relative timing of pluton emplacement in the Priest River complex indicates that it was primarily a response to decompression rather than a cause. The mylonitized Silver Point and undeformed Wrencoe plutons bracket the end of a rapid phase of exhumation to *c.* 50-48 Ma. Zircon $\epsilon_{\text{Hf}(i)}$ values and Lu-Hf isotope evolution indicate that the Silver Point and Wrencoe plutons crystallized from homogeneous magmas sourced from Archean–Proterozoic basement orthogneisses, whereas the Spokane granite and two leucocratic units appear to have been produced by partial melting of the Hauser Lake gneiss. Comparison of the Priest River complex with other deeply exhumed northern Cordilleran complexes indicates variability in the timing and, therefore, relative influences of partial melting and magmatism on the initiation of exhumation, which must be accounted for in numerical models of metamorphic core complex formation and evolution.

Keywords: Priest River complex; exhumation; magmatism; zircon; CA-TIMS; Hf isotopes

1. Introduction

Partial melting and magmatism are spatially and temporally associated with continental metamorphic core complexes in the North American Cordillera (e.g., Armstrong, 1982; Davis and Coney, 1979). The resulting decrease in crustal viscosity and strength due to the presence of melt may initiate and/or enhance core complex exhumation during gravitational collapse of thickened crust (e.g., Armstrong, 1982; Armstrong and Ward, 1991; Coney and Harms, 1984; Foster and Fanning, 1997; Gans, 1989; House et al., 2002; Lister and Baldwin, 1993; Vanderhaeghe and Teyssier, 2001). In order to assess the influence of partial melting and magmatism on exhumation of core complexes, the timing of these processes must be well constrained. Variation in the volume and timing of partial melting and magmatism among Cordilleran metamorphic core complexes suggests that the influence of each on exhumation may vary (e.g., Armstrong, 1982; Whitney et al., 2013). Such differences must be either allowed or accounted for in numerical models of metamorphic core complex formation and evolution.

Core complexes within the northern North American Cordillera typically exhibit greater degrees of partial melting and were exhumed during the Paleocene and Eocene, whereas complexes of the southern belt display lesser degrees of partial melting and were exhumed during the Oligocene and/or Miocene (Fig. 1; e.g., Armstrong, 1982; Coney, 1980; Coney & Harms, 1984; Rey et al., 2009; Whitney et al., 2013). The northern belt core complexes are also similar in position relative to the Sevier thrust front, craton margin ($^{87}\text{Sr}/^{86}\text{Sr} = 0.706$), and the compressional North American plate margin. This study focuses on the northern belt complexes as examples of metamorphic core

complexes with histories of partial melting and magmatism in a far-field compressional tectonic setting.

The degree of partial melting in core complexes within the northern Cordillera varies from the diatexite (melt-dominated) migmatites of the Shuswap complex gneiss domes (e.g., Gordon et al., 2008; Norlander et al., 2002; Vanderhaeghe et al. 1999), to the metatexite (lower melt fraction) migmatites of the Priest River complex (e.g., Stevens et al., 2015), to migmatite-absent complexes, such as the Clearwater complex (e.g., Lang and Rice, 1985) (Fig. 1). Similarly, magmatism varies in character, volume, and timing. Leucogranite in the Thor-Odin dome within the Shuswap complex crystallized during exhumation (Fig. 1; Vanderhaeghe et al., 1999). In the Priest River complex, granitoid intrusions formed either well prior to peak metamorphism or following the onset of extension (syn- to post-kinematic) (Fig. 1; Archibald et al., 1984; Bickford et al., 1985; Doughty and Price, 1999; Harms and Price, 1992; Stevens et al., 2015; Whitehouse et al., 1992).

The Priest River complex (PRC) is located at the southern end of the Omineca crystalline belt, the eastern metamorphic core of the Cordillera; the Shuswap complex is also located within the Omineca Belt (Fig. 1; Parrish et al., 1988). The PRC, Shuswap, Clearwater, and Bitterroot complexes were exhumed from similar depths (~7–11 kbar, or ~23–33 km) (Fig. 1; Doughty et al., 2007; Foster et al., 2001; House et al., 1997; Norlander et al., 2002; Stevens et al., 2015). The PRC, along with the Shuswap and Bitterroot complexes, is among the larger complexes in the northern belt (Fig. 1); however, the Shuswap complex contains melt-dominated, vertically exhumed gneiss domes (e.g., Gordon et al., 2008; Norlander et al., 2002), and the Bitterroot complex is

dominated by the Idaho batholith and behaved as a gneiss dome (e.g., Chase et al., 1983; Foster and Fanning, 1997; Hyndman, 1980). The PRC's shared characteristics with other northern belt complexes, as well as its lesser degree of partial melting and late pluton emplacement, suggest that it is a representative northern belt complex that will provide constraints on the relative influences of crustal anatexis and plutonism on exhumation.

In order to characterize the relationships between partial melting, magmatism, and exhumation in the PRC and other northern belt core complexes, two hypothetical end member scenarios are presented here. In the first scenario, partial melting and/or pluton emplacement occur during prograde metamorphism and crustal thickening, and thermal weakening initiates core complex exhumation. In the second scenario, partial melting and/or pluton emplacement are responses to decompression during exhumation and retrograde metamorphism. While melt may aid continued exhumation, it does not have a role in initiating exhumation in this scenario. The Valhalla dome of the Shuswap complex (Fig. 1) serves as an example of the first end member scenario, as leucogranite crystallized during exhumation, immediately following peak metamorphism (Gordon et al., 2008; Spear and Parrish, 1996). The Clearwater complex (Fig. 1) represents the latter scenario, as intrusion of Eocene (*c.* 52–46 Ma) granites followed Cretaceous–Tertiary prograde metamorphism and exhumation (Burmester et al., 2004; Doughty et al., 2007; Gaschnig et al., 2010; Marvin et al., 1984).

Evaluation of the scenarios presented above requires precise constraints on the timing of prograde metamorphism, exhumation, crustal melting, and magmatism. Constraints on the timing of the first three events in the PRC are based on monazite and xenotime petrochronology (Stevens et al., 2015); however, previous geochronological

constraints on granitoids in the PRC lack the precision to determine the timing of emplacement relative to metamorphism and exhumation. This study analyzes zircon from metapelitic rocks and granitoids for U-Pb geochronology, trace element geochemistry, and Lu-Hf isotopes using laser-ablation inductively coupled plasma mass spectrometry (LA-ICPMS), chemical abrasion thermal ionization mass spectrometry (CA-TIMS), and laser-ablation split-stream multi-collector inductively coupled plasma mass spectrometry (LASS-MC-ICPMS). These new data improve the geochronological constraints on the emplacement of granitoids in the PRC footwall and clarify the temporal relationships between crustal melting, magmatism, and exhumation of the PRC, allowing for comparison with other core complexes in the northern belt of the North American Cordillera. These results thus have implications for the significance of crustal melting and magmatism in the formation of metamorphic core complexes.

2. Background

The majority of the PRC is located within the northern Idaho panhandle (Fig. 2). The footwall of the PRC was exhumed by the Purcell Trench fault to the east and the Newport fault to the northwest (Fig. 2). The footwall contains the Hauser Lake gneiss, which is a migmatitic, mylonitic, pelitic schist and gneiss that is considered the metamorphosed equivalent of the Prichard Formation, the lowermost unit of the Mesoproterozoic Belt Supergroup (Cressman, 1989; Doughty and Chamberlain, 2008; Lewis et al., 2010; Sears et al., 1998). The Hauser Lake gneiss was intruded by Cretaceous and Eocene granitoids (Fig. 2), as was the low-grade Belt Supergroup in the PRC hanging walls.

93 The pressure–temperature (P – T) conditions and timing of metamorphism of the
94 Hauser Lake gneiss were reported by Stevens et al. (2015). Monazite and xenotime Th-
95 Pb dating indicates metamorphism spanned *c.* 74–54 Ma, with peak metamorphism at *c.*
96 68–64 Ma. Mineral chemistry and isochemical phase diagrams (pseudosections) indicate
97 that the peak metamorphic assemblage (garnet + sillimanite + biotite $\pm$ muscovite +
98 plagioclase + K-feldspar $\pm$ rutile $\pm$ ilmenite + quartz) records peak P – T conditions of
99 ~9.6–10.3 kbar and ~790°C. Near isothermal decompression of the PRC began *c.* 60–57
100 Ma and melt crystallization occurred *c.* 54–44 Ma at mid-crustal levels.
101 Thermochronological (K-Ar, $^{40}\text{Ar}/^{39}\text{Ar}$) data are consistent with this interval for the
102 exhumation of the PRC footwall, given cooling through muscovite and biotite closure
103 temperatures (~350 °C) at *c.* 50–43 Ma (Doughty and Price, 1999; Miller and Engels,
104 1975).

105 Cretaceous igneous intrusions in the PRC footwall include the Selkirk igneous
106 complex (Kaniksu batholith) and the Spokane granite (Fig. 2). The Selkirk igneous
107 complex records U-Pb zircon TIMS multi-grain fraction lower intercept dates of $93.8 \pm$
108 0.9 Ma and 93.4 ± 0.9 Ma, although there are U-Pb zircon LA-ICPMS dates as old as *c.*
109 120 Ma (Archibald et al., 1984; Gaschnig et al., 2012). The Spokane granite records U-
110 Pb zircon multi-grain fraction lower intercept dates of 92 ± 5 Ma for a muscovite-bearing
111 phase and 72 ± 5 Ma for a biotite-bearing phase (Bickford et al., 1985).

112 Eocene granitoids that intruded the Hauser Lake gneiss include the Silver Point
113 quartz monzonite, the Wrencoe granodiorite, and the Rathdrum granite (Fig. 2). U-Pb
114 zircon multi-grain fraction lower intercept dates for the Wrencoe and Silver Point
115 plutons, which have been interpreted as a continuous intrusive body (Doughty and Price,

1999), are 52.1 ± 1.2 Ma and 50.1 ± 6.3 Ma, respectively (Whitehouse et al., 1992). The Rathdrum granite records a zircon U-Pb lower intercept of 52 ± 10 Ma (Bickford et al., 1985). The Silver Point pluton is mylonitized along the trace of the Newport fault (Doughty and Price, 1999; Harms and Price, 1992). The Rathdrum granite is also mylonitized; both plutons are interpreted as late synextensional (Doughty and Price, 1999; Rhodes and Hyndman, 1984).

The Hauser Lake gneiss is migmatitic, with visual estimates of 25-40% *in situ* (crystallized in place) and/or in-source (migrated from formation site, crystallized in source layer) leucosome in outcrops (Sawyer, 2008; Stevens et al., 2015). Textures observed at thin section to outcrop scale are consistent with patch metatexite migmatite and stromatic metatexite migmatite (Stevens et al., 2015). Leucocratic lenses and layers are primarily concordant with the foliated Hauser Lake gneiss; discordant leucocratic dikes occur more rarely (Stevens et al., 2015).

Archean crystalline basement is exposed within the core of an antiform near the center of the PRC (Fig. 2; Doughty et al., 1998). The Pend Oreille gneiss, interpreted to be an orthogneiss, has a zircon U-Pb multi-grain fraction upper intercept date of $2651 \pm 21/-19$ Ma (Doughty et al., 1998). The Pend Oreille gneiss is mantled by the Gold Cup Mountain quartzite, which is overlain by the Hauser Lake gneiss (Doughty et al., 1998). The Gold Cup Mountain quartzite has been correlated with the Neihart quartzite, which underlies the Prichard Formation (Doughty et al., 1998). The Laclede augen gneiss has a zircon U-Pb multi-grain fraction upper intercept date of 1576 ± 13 Ma, and is tectonically interleaved with the Hauser Lake gneiss (Evans and Fischer, 1986; Doughty et al., 1998). Recently recognized exposures of Neoproterozoic amphibolite, Paleoproterozoic orthogneiss,

and the unconformably overlying Razorback quartzite at the southern end of the PRC (Fig. 2) are correlated with the Pend Oreille gneiss and Gold Cup Mountain quartzite (Buddington et al., 2014; Vervoort et al., 2016; Wang et al., 2014). Zircon Hf isotope compositions suggest that the Neoarchean amphibolite was derived from a mantle source, whereas the Paleoproterozoic basement formed from a mixture of evolved continental crust and depleted mantle (Wang, 2015). The Sm-Nd and Pb isotope compositions and U-Pb zircon upper intercept dates suggest a Neoarchean or Paleoproterozoic cratonic lower crustal source for the Silver Point and Wrencoe plutons (Whitehouse et al., 1992). The PRC basement may be continuous with either the Hearne province or the Medicine Hat block to the east, or may represent an independent block of Archean-Proterozoic crust, but is unlikely to be related to the Wyoming craton, which lacks the 1.86 Ga signature of orthogneisses in the Priest River complex (Doughty et al., 1998; Foster et al., 2006; Vervoort et al., 2016; Wang et al., 2015; Whitehouse et al., 1992).

3. Methods

3.1 Sample Selection

Nine samples from the PRC footwall were analyzed (Fig. 2; Table 1), including six igneous samples and three samples of Hauser Lake gneiss. Samples from the Spokane granite, Silver Point quartz monzonite, Wrencoe granodiorite, and Rathdrum granite were dated because existing geochronological data lack the precision to address this study's goals. The sampled Spokane granite contains muscovite and biotite, and is weakly foliated. The Silver Point and Wrencoe granitoids are biotite-bearing with K-feldspar

phenocrysts; the Silver Point quartz monzonite is locally mylonitized. The sample of Rathdrum granite is from the small Spirit Lake pluton (fine-grained, biotite-bearing, and nonfoliated granite correlated with the Rathdrum granite). An undeformed, biotite-bearing leucocratic dike that crosscuts the Hauser Lake gneiss is included in order to constrain the timing of deformation. Hauser Lake gneiss samples were selected to further constrain the partial melting history. These samples contain the peak metamorphic assemblage garnet + sillimanite + biotite + plagioclase + K-feldspar $\pm$ rutile $\pm$ ilmenite + quartz (Stevens et al., 2015). Samples show evidence of prograde kyanite pseudomorphed by sillimanite, and of retrograde muscovite (Stevens et al., 2015).

3.2 Methods Overview

A detailed description of the methods used in this study is provided in the supplement. Samples were crushed, milled, sieved to $<500\ \mu\text{m}$, and washed. Zircon and monazite were separated using standard magnetic and heavy liquid techniques. Most samples were processed using methylene iodide (MeI), whereas Silver Point and Wrencoeur pluton samples were processed using lithium polytungstate (LST), with sufficient yield. Zircon grains were hand-picked with a preference for clear, euhedral grains in various colors and sizes from the igneous samples; grains from metapelitic samples were euhedral or subrounded. Both zircon and monazite were separated from the leucocratic dike sample; monazite grains were pale yellow to pink, and varied from small, transparent, inclusion-free, lentil-shaped grains, to larger fractured grains with inclusions.

Zircon grains were annealed at 900°C for 60 hours in a muffle furnace in quartz beakers at Boise State University. Grains and zircon standards were mounted in epoxy and polished until grain centers were exposed. Monazite grains were also mounted in epoxy at Boise State University. Zircon and monazite in these mounts were imaged at the University of Montana using cathodoluminescence (CL) imaging for zircon and backscattered electron (BSE) imaging for monazite via a Tescan Vega 3 scanning electron microscope. Images were used to characterize zircon and monazite zoning and guide placement of *in situ* analyses (Figs 3, S1–S3).

The Hf isotope composition and concurrent U-Pb date of zircon were determined via *in situ* laser ablation split-stream multi-collector inductively coupled plasma mass spectrometry (LASS-MC-ICPMS) at the Radiogenic Isotope and Geochronology Laboratory (RIGL) at Washington State University (Tables S1, S2). The LASS technique is used here to detect inherited Precambrian domains when targeting young (i.e., Eocene) growth domains by monitoring the degree of U-Pb concordance (Fisher et al., 2014a, b). In the case of samples of the Hauser Lake gneiss or inherited cores in younger granites, the U-Pb date and Hf isotope composition are not reported for analyses that are more than 10% discordant. The ^{207}Pb -corrected $^{206}\text{Pb}/^{238}\text{U}$ dates, determined using Isoplot 3.0 (Ludwig, 2003), are reported for analyses of younger zircons (Table S2).

LASS analysis was followed by *in situ* laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) analysis for zircon U-Pb geochronology and trace elements to further characterize zircon chemistry at Boise State University (Table S3). An attempt was made to analyze zircon grains using both the LASS-MC-ICPMS and LA-ICPMS methods, given interest in the grains and available space for analysis; otherwise,

the two methods were used to characterize as many grains as possible. Finally, selected grains were plucked from epoxy mounts for chemical abrasion thermal ionization mass spectrometry (CA-TIMS, zircon) or isotope dilution thermal ionization mass spectrometry (ID-TIMS, monazite) for U-Pb geochronology at Boise State University (Table S4). CA-TIMS analysis was used to determine high precision dates and interpret igneous crystallization ages. Weighted mean dates were calculated from equivalent dates using Isoplot 3.0 (Ludwig, 2003). Errors for weighted mean dates and dates from individual grains are given at 2 sigma.

Samples from the Silver Point and Wrencoe plutons were not analyzed via LA-ICPMS in favor of more precise CA-TIMS analysis. The Spokane, Silver Point, and Wrencoe granitoids and the leucocratic dike were selected for CA-TIMS analysis based on LASS-MC-ICPMS and LA-ICPMS ages and lack of zircon inheritance. Where analysis of whole zircon grains with distinct cores, interpreted as detrital or xenocrystic, was undesirable, zircon tips were broken off and analyzed (Fig. 3, Table S4).

4. Results

The results are summarized by sample, including descriptions of zircon and monazite textures, ranges of U-Pb dates yielded from zircon cores and rims from LASS-ICPMS and LA-ICPMS analysis, and U-Pb dates from CA-TIMS (zircon) and ID-TIMS (monazite) analysis. Dates presented are determined from $^{206}\text{Pb}/^{238}\text{U}$ (< 1.0 Ga) or $^{207}\text{Pb}/^{206}\text{Pb}$ (> 1.0 Ga) ratios. Epsilon Hf values for the zircons from most granitoids are calculated using the igneous crystallization age (or maximum crystallization age) interpreted from CA-TIMS or LA-ICPMS U-Pb zircon dates rather than the LASS

determinations as the former gives the best determination of the Hf formation age of the magmatic zircon. In the case of inherited zircon cores (e.g., the Hauser Lake gneiss and the Rathdrum granite), $\epsilon_{\text{Hf}(i)}$ were calculated using corresponding LASS-MC-ICPMS dates, which provide the best determination of the age of the inherited components. For this reason, interpretations regarding granitoid crystallization ages are included in this section and then summarized in the discussion.

4.1. Hauser Lake Gneiss

4.1.2. PR-11-033B

Zircon has distinct, dark detrital cores and CL bright rims with complex oscillatory zoning representing igneous growth (Fig. S1a). Zircon yield was low; only six small (65–140 μm) grains were mounted. Small grain size and the presence of detrital cores prevented Hf isotope analysis. Analysis of four zircon cores yielded dates of 1781 ± 41 to 1277 ± 61 Ma ($n = 4$) (Fig. 4a; Table S3).

4.1.3. PR-10-055

Zircon has distinct cores and rims with oscillatory zoning, interpreted as detrital cores and igneous rims (Fig. S1b). Analysis of zircon cores from 14 grains yielded Proterozoic (2371 ± 22 to 1563 ± 26 Ma, $n = 12$) and Cretaceous-Paleocene (66 ± 1 to 60.7 ± 3.2 , $n = 4$) dates (Figs 4a, 5a, 6a; Tables S1, S3). Dates from rims of eight zircon grains vary similarly with values of 2489 ± 45 to 1454 ± 34 Ma ($n = 5$) and 71 ± 1 to 61.3

± 1.9 Ma ($n = 5$) (Figs 4a, 5a, 6a; Tables S1, S3). Initial epsilon Hf values vary from -16.3 calculated at 66 Ma to +10.1 at 1839 Ma ($n = 11$) (Fig. 6b, Table S1).

4.1.4. PR-11-040

Zircon has distinct, detrital cores and rims with simple or complex oscillatory, igneous growth zoning (Fig. S1c). Zircon analyses from both cores and rims of fifteen grains yielded Proterozoic dates of 1800 ± 27 to 1473 ± 26 Ma ($n = 15$), and a single Cretaceous date of 97.0 ± 4.7 Ma from the rim of a separate grain (Figs 4a, 5a, 6a; Tables S1, S3). Epsilon Hf values vary from -0.7 calculated at 1800 Ma to +7.7 at 1508 Ma (Fig. 6b; Table S1).

4.2. Granitoids

4.2.1. Spokane Granite (PR-11-038)

Zircon has distinct cores with patchy or oscillatory zoning and oscillatory-zoned rims, suggesting igneous growth on xenocrystic cores (Figs 3a, S2a). Tips of grains yielded CA-TIMS dates of 85.57 ± 0.07 to 76.49 ± 0.06 Ma ($n = 7$) (Fig. 7a; Table S4); another grain contains a considerably older component (165.67 ± 0.15 Ma). Agreement between the youngest LA-ICPMS and CA-TIMS dates suggests a maximum crystallization age of 76.5 Ma. Values for $\epsilon_{\text{Hf}(77 \text{ Ma})}$ range from -21.2 to -13.0 (Fig. 6; Table S2).

4.2.2. Silver Point Quartz Monzonite (PR-11-010)

Zircon has sector zoning (Figs 3b, S2b); some grains have distinct, xenocrystic cores. The youngest four CA-TIMS dates from whole single grains have a weighted mean of 50.13 ± 0.02 Ma (MSWD 2.4; probability of fit = 0.06) (Fig. 7b, Table S4), which is interpreted as the igneous crystallization age. One other date is slightly older at 50.52 ± 0.04 Ma and two others have highly discordant dates that indicate the presence of inherited components. Epsilon Hf values calculated at 50 Ma range from -25.3 to -21.5 (mean = -23.7; 2SD = 2.1; $n = 22$) (Fig. 6; Table S2).

4.2.3. Wrencoe Granodiorite (PR-11-011)

Zircon has sector and oscillatory zoning, with distinct (xenocrystic) cores in some grains (Figs 3c, S2c). Seven whole single zircon grains yielded CA-TIMS dates between 48.17 ± 0.04 and 47.85 ± 0.03 Ma ($n = 7$) (Fig. 7c; Table S4), with six dates spread over 0.2 Ma between 48.04 ± 0.04 to 47.85 ± 0.03 Ma. Due to the narrow age range of the majority of analyzed grains, we interpret the igneous crystallization age as close to 47.9 Ma. Epsilon Hf values at 48 Ma range from -27.4 to -24.2 (mean = -26.0; 2SD = 1.8; $n = 22$) (Fig. 6; Table S2).

4.2.4. Rathdrum Granite (PR-10-043, PR-11-005)

Most zircons have distinct (xenocrystic) cores with oscillatory-zoned mantles on some grains, and thin, CL bright exterior rims (Fig. S3a,b). Analysis of zircon cores in nineteen grains yielded Archean-Proterozoic U-Pb dates of 2672 ± 34 to 1519 ± 38 Ma ($n = 26$) and Cretaceous-Eocene dates of 79 ± 2 to 46.4 ± 2.0 Ma ($n = 9$) (Figs 4b, 5c, 6a; Tables S1, S3). Analysis of zircon rims from seven grains yielded similar results between

2649 $\pm$ 45 and 45 $\pm$ 1 Ma ($n = 9$). The two youngest LA-ICPMS dates are equivalent, with a weighted mean date of 46.4 $\pm$ 1.8 Ma (MSWD = 0.001, probability of fit = 0.97) interpreted as a maximum igneous crystallization age (Fig. 5c). Epsilon Hf values range from -30.0 calculated at 79 Ma to +2.1 at 2632 Ma, while one core analysis gives an $\epsilon_{\text{Hf}(i)}$ value of +2.4 at 1519 Ma (Fig. 6b; Table S1).

4.2.5. *Leucocratic Dike (PR-11-014)*

Zircon has distinct, complexly zoned (xenocrystic) cores and oscillatory-zoned rims that are CL bright (interior rim) and dark (exterior rim) (Figs 3d, S3c). Analyses of zircon cores from five grains yielded U-Pb dates of 72.8 $\pm$ 4.1 to 66.1 $\pm$ 1.4 Ma ($n = 5$) (Tables S2, S3). Zircon rims from twenty-five grains yielded U-Pb dates of 78.5 $\pm$ 1.8 to 40.7 $\pm$ 1.2 Ma (Tables S2, S3). The twelve oldest LA-ICPMS dates, which are mostly from CL bright interior rims, yielded a weighted mean date of 69.4 $\pm$ 1.5 Ma (MSWD = 1.2; probability of fit = 0.29) (Fig. 5d). The youngest four dates, all from CL dark exterior rims, yielded a weighted mean date of 51.8 $\pm$ 1.5 Ma (MSWD = 2.0; probability of fit = 0.12) (Fig. 5d). While all zircon has low Th/U (0.00–0.05), older zircon has low total REE contents (500–2800 ppm) compared with younger zircon (2600–7400 ppm) (Table S3). CA-TIMS dates from zircon tips are 65.94 $\pm$ 0.06 to 49.92 $\pm$ 0.04 Ma ($n = 8$) (Fig. 7d; Table S4), and likely represent mixtures of the range of dates in the interior and exterior rims, as captured by LA-ICPMS analysis. Monazite grains have either patchy zoning or xenocrystic cores with subtle, oscillatory rims interpreted as igneous in origin (Fig. 3e). Monazite fragments yielded ID-TIMS $^{207}\text{Pb}/^{235}\text{U}$ dates of 57.56 $\pm$ 0.09 Ma to 49.79 $\pm$ 0.41 Ma ($n = 7$) (Fig. 7d; Table S4). The youngest date is interpreted as the

maximum igneous crystallization age, with all older dates reflecting the presence of inherited components. Values for $\epsilon_{\text{Hf}}(50 \text{ Ma})$ range widely from -16.8 to -2.7 (Fig. 6; Table S2), also indicating incorporation of inherited components in some of these analyses.

5. Discussion

Zircon U-Pb geochronology provides constraints on zircon inheritance and timing of melt crystallization and pluton emplacement. Zircon trace element geochemistry from LA-ICPMS only permits identification of zircon populations from the leucocratic dike. Zircon U-Pb geochronology from CA-TIMS provides the first high-precision date for the Silver Point quartz monzonite, and improves crystallization dates for the Spokane granite, Wrencoë granodiorite, and a leucocratic dike that intrudes the Hauser Lake gneiss. Zircon Hf isotope analysis using LASS-MC-ICPMS allows determination and comparison of magma source characteristics with the PRC basement and Hauser Lake gneiss for the analyzed granitoids. These data create a detailed timeline for the sequence of events in the PRC, clarify the relationship between partial melting, magmatism, and exhumation in the PRC, and permit comparison of the timing of these processes between northern belt core complexes.

5.1. Detrital and Metamorphic Zircon Ages

Archean–Proterozoic zircon dates from Hauser Lake gneiss samples and the Rathdrum granite are consistent with detrital zircon dates from the Prichard Formation and previously published Archean-Proterozoic zircon dates for the Hauser Lake gneiss (Fig. 4c; Doughty & Chamberlain, 2008; Lewis et al., 2010; Wang, 2015). This indicates

that the Prichard Formation is the likely protolith of the Hauser Lake gneiss samples in this study, and suggests zircon in the Rathdrum granite is inherited from the Hauser Lake gneiss or the Prichard Formation.

Cretaceous–Tertiary zircon and monazite dates from the Hauser Lake gneiss (*c.* 66–61 Ma), Rathdrum granite (*c.* 67–46 Ma), and leucocratic dike (*c.* 78–41 Ma) are synchronous with $^{208}\text{Pb}/^{232}\text{Th}$ monazite dates recording prograde to peak (*c.* 68–64 Ma) metamorphism, isothermal decompression and retrograde metamorphism, and melt crystallization (*c.* 54–44 Ma) in the Hauser Lake gneiss (Stevens et al., 2015).

5.2. Crystallization Ages

Zircon CA-TIMS and LA-ICPMS dates constrain the age of igneous crystallization of granitoids within the PRC footwall. The Silver Point quartz monzonite is the only sample with a precise interpreted crystallization age because it is the only unit without a significant inherited zircon component or a prolonged crystallization history. Scattered dates from the other samples give maximum crystallization ages based on the youngest dates. A summary of interpreted crystallization ages are: $< 76.5 \pm 0.1$ Ma for the Spokane granite, 50.13 ± 0.02 for the Silver Point quartz monzonite, *c.* 47.9 for the Wrencoë pluton, $< 46.4 \pm 1.8$ Ma for the Rathdrum granite, and $< 49.8 \pm 0.4$ for the leucocratic dike.

5.3. Magmatic Sources

Analyses of detrital zircon from the Hauser Lake gneiss and inherited cores from the Rathdrum granite have both positive and negative $\epsilon_{\text{Hf}(i)}$ values ranging from -10 to

+10. These results are used to constrain the Lu-Hf evolution of the potential source materials to the granitoids in this study (Fig. 6b,c). The Wrencoe and Silver Point plutons have homogeneous $\epsilon_{\text{Hf}(i)}$ values, each with a range of fewer than four ϵ_{Hf} units, suggesting crystallization from relatively homogeneous magmas and a lack of inherited components (Fig. 6c). The spread of $\epsilon_{\text{Hf}(i)}$ values for analyses from the Spokane granite and leucocratic dike (Fig. 6c) can be attributed to the incorporation of inherited zircon components in the analyses. The Hf isotope compositions for zircon from the granitoids are consistent with derivation from Archean–Proterozoic sources, possibly the PRC basement and/or the Hauser Lake gneiss, which they intrude (Fig. 6b).

Modeling the Lu-Hf isotope evolution of the PRC basement and detrital zircon of the Hauser Lake gneiss offers constraints on the source materials from which Eocene granitoids may have formed. The Hf isotope compositions, when calculated at 50 Ma would be approximately –22 to –8 for the Hauser Lake gneiss, and –44 to –18 for Neoarchean amphibolite and Paleoproterozoic orthogneisses (Wang, 2015). The Spokane granite has $\epsilon_{\text{Hf}(77 \text{ Ma})}$ values of –21 to –13, and the leucocratic dike has a mean $\epsilon_{\text{Hf}(50 \text{ Ma})}$ of approximately –10. These values overlap with those projected from a Hauser Lake gneiss source, but are more radiogenic than the Hf isotope compositions calculated at 50 Ma for Archean or Paleoproterozoic basement lithologies. Conversely, the Silver Point and Wrencoe plutons and the Rathdrum granite have $\epsilon_{\text{Hf}(i)}$ values that more closely overlap the Lu-Hf isotope evolution of Archean and Paleoproterozoic basement lithologies (Fig. 6c). Constraints on the source of the Rathdrum granite are more difficult to obtain from the limited LASS-MC-ICPMS data; however, the Rathdrum granite contains abundant inherited zircon that include *c.* 2.6 Ga and 1.8 Ga components, consistent with 1) detrital

populations in the Hauser Lake gneiss, and 2) the ages and $\varepsilon_{\text{Hf}(i)}$ values of basement amphibolites and orthogneisses (Doughty et al., 1998; Doughty and Chamberlain, 2008; Lewis et al., 2010; Vervoort et al., 2016; Wang, 2015). More extensive LA-ICPMS analysis of inherited cores in the Rathdrum granite yield a range of ages not presently observed in the basement orthogneisses, strongly suggesting the Hauser Lake gneiss as its source.

The $\varepsilon_{\text{Hf}(i)}$ values for zircons in the Spokane granite and leucocratic dike suggest that partial melting of Hauser Lake gneiss could serve as a viable source. The less radiogenic values for the zircons from the Silver Point and Wrencoe plutons, and possibly the Rathdrum granite, suggest that these plutons may have had greater input from Paleoproterozoic, and perhaps Archean, components (Fig. 6b, c). This distinction suggests that the *c.* 50–48 Ma plutons (Silver Point, Wrencoe) were derived from partial melting of PRC basement sources, whereas Cretaceous–Eocene leucocratic material was sourced from the Hauser Lake gneiss, indicating contributions from an additional magma source either during or following exhumation of the Priest River complex. More deeply sourced late synkinematic to post-kinematic plutons suggest that exhumation permitted formation and/or emplacement of these melts.

The zircon Hf isotope data are consistent with previously published data and interpretations that suggest igneous intrusions in the PRC footwall derive, at least in part, from crustal melting of the metamorphosed Belt Supergroup, or its protoliths, and that the more unradiogenic Silver Point and Wrencoe plutons have a Neoarchean or Paleoproterozoic cratonic lower crustal source (Whitehouse et al., 1992). These data support the presence of Archean basement in the PRC, but do not discriminate between

the Hearne province, Medicine Hat block, or an unknown Archean province, due to the lack of constraints on the nature of the basement rocks throughout this region (Foster et al., 2012; Vervoort et al., 2016)

5.4. Constraints on Exhumation

The crystallization ages for the Silver Point and Wrencoë plutons are distinguishable, indicating that these plutons are not continuous, but rather represent discrete pulses of crustal melting during the exhumation of the PRC. These crystallization ages also bracket the timing of deformation and exhumation in the PRC, indicating that extension along the Newport fault, which crosscuts the mylonitized Silver Point pluton, and/or the Purcell Trench fault ceased by *c.* 47.9 Ma, when the undeformed Wrencoë pluton was emplaced. Crystallization of the cross cutting leucocratic dike at *c.* 49.8 Ma is consistent with this interpretation. Mica cooling ages from K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology indicate that the Hauser Lake gneiss reached mid-crustal levels by *c.* 50–43 Ma (Doughty and Price, 1999; Miller and Engels, 1975). These Eocene plutons were emplaced at mid-crustal levels (~12 km) following near-isothermal decompression to ~5.0–4.4 kbar (Stevens et al., 2015).

5.5. Evolution of the PRC

The data presented above, in combination with previously published work, permit construction of the most detailed timeline yet available for metamorphism, crustal anatexis, magmatism, and exhumation in the PRC (Fig. 8). The Spokane granite was emplaced and crystallized during Sevier crustal thickening and prograde metamorphism,

prior to the *c.* 64 Ma peak metamorphism of the Hauser Lake gneiss (Stevens et al., 2015). The majority of partial melting, ~14–24 volume % melt, was produced during prograde metamorphism (Stevens et al., 2015). By *c.* 60–57 Ma the exhumation of the PRC footwall had been initiated (Stevens et al., 2015). Initially, exhumation followed a near-isothermal *P–T* path, decompressing to 5.5–4.0 kbar before cooling and crystallization of melt *c.* 54 Ma; partial melting of the crust during retrograde metamorphism was limited to a few volume percent (Stevens et al., 2015). The Silver Point pluton was emplaced at mid-crustal levels at *c.* 50 Ma along the Newport fault. Exhumation continued along the Newport fault, shearing and mylonitizing the Silver Point pluton in the process. The *c.* 47.9 Ma Wrencoë pluton was emplaced shortly after the cessation of extension along the Newport fault, along with the *c.* 46 Ma Rathdrum granite. By *c.* 44–43 Ma, the PRC footwall cooled through mica closure temperatures (~300°C) and remaining crustal melt had crystallized (Miller and Engels, 1975; Stevens et al., 2015).

Examining the relationship between the crystallization of igneous units relative to the timing of metamorphism and exhumation of the Hauser Lake gneiss permits evaluation of hypothetical exhumation scenarios. The Selkirk igneous complex crystallized prior to the Cretaceous–Tertiary metamorphism of the Hauser Lake gneiss, whereas crystallization of the Spokane granite occurred during early prograde metamorphism. The Silver Point, Wrencoë, and Rathdrum plutons were emplaced at mid-crustal levels following the majority of near-isothermal decompression of the PRC. This time frame is consistent with a scenario in which magma is formed and/or emplaced as a response to exhumation, rather than triggering exhumation. It is possible, however, that

459 emplacement of the Silver Point pluton along the Newport fault may have eased
460 displacement during the final phases of extension and exhumation.

461 These data do not fully resolve the role of prograde crustal anatexis in the
462 exhumation of the PRC. The PRC contains the Spokane dome, a gneiss dome that
463 exhibits neither the relatively high melt volume, the extensively interconnected melt
464 network, nor the near-vertical foliation patterns suggestive of diapiric rise, as seen in the
465 Valhalla or Thor-Odin domes (Fig. 2; Gordon et al., 2008; Norlander et al., 2002;
466 Whitney et al., 2004). We interpret the Spokane dome to have formed during *c.* 60–48
467 Ma exhumation of the PRC. Pseudosection modeling indicates that ~14–24 volume %
468 partial melt was produced during prograde metamorphism in the PRC, prior to the onset
469 of exhumation at *c.* 60 Ma (Stevens et al., 2015). The melt connectivity transition, when
470 the melt phase is interconnected, is achieved at melt volumes of ~7%, which reduces rock
471 strength by ~90% (Rosenberg and Handy, 2005). It is thus expected that partial melt
472 weakened the crust within the PRC prior to exhumation; however, crustal weakening
473 associated with the solid-to-liquid transition that occurs at ~40 volume % partial melt was
474 not achieved in the PRC (Rosenberg and Handy, 2005). The solid-to-liquid transition is
475 more likely associated with diatexite migmatite, such as that of the Shuswap gneiss
476 domes. It is suggested that the resulting increase in magma buoyancy at this transition
477 may cause the gravitational instability that leads to gneiss dome formation
478 (Vanderhaeghe, 2009). This suggests that partial melting in the PRC was sufficient to
479 weaken the crust and potentially allow for crustal flow, but was not a sufficient melt
480 volume to trigger exhumation.

5.6. *Exhumation in Northern Belt Complexes*

Available data and interpretations regarding the timing of metamorphism, peak metamorphism, pluton emplacement, onset of exhumation, leucosome crystallization, and cooling for several deeply exhumed northern belt complexes and gneiss domes have been compiled (Table S5 and references therein) and summarized in Figure 9. Data for the Bitterroot complex include timing of metamorphism (*c.* 64 Ma, 7–8 kbar, 700 °C; House et al., 1997; Foster and Fanning, 1997), crystallization ages for early, main, and synextensional plutons, and thermochronological data (Fig. 9, Table S5 and references therein). These data demonstrate that peak metamorphism occurred during the main phase of plutonism, followed by slow cooling. Rapid exhumation (mylonitization, and doming) occurred *c.* 48 Ma (Foster et al., 2001; Foster and Raza, 2002; House & Hodges, 1994; House et al., 1997), whereas synextensional pluton crystallization ages range from *c.* 52 Ma to *c.* 47 Ma (e.g., Foster et al., 2001; Foster et al., 2007; House et al., 2002). The apparent record of a lengthy cooling history for the Bitterroot complex, in comparison to those of other complexes, is an artifact of abundant low-temperature thermochronology data for the Bitterroot complex.

Data for the Cretaceous–Tertiary evolution of the Clearwater complex south of the PRC include the timing of latest metamorphism, magmatism, and cooling (thermochronology) (Fig. 9; Table S5). Evidence for Cretaceous-Eocene crustal anatexis in the exhumed units of the Clearwater complex has not been described. Peak metamorphism (~8–11 kbar, ~700 °C, *c.* 64 Ma; Doughty et al., 2007; Grover et al., 1992) was followed by exhumation *c.* 59 Ma (Doughty et al., 2007). Intrusion of the

Roundtop, Beaver Creek, and Bungalow plutons occurred during exhumation and cooling (Burmester et al., 2004; Gaschnig et al., 2010; Marvin et al., 1984).

Data for the Valhalla dome in the Shuswap complex include the timing of metamorphism, crystallization of prograde leucogranite and later equivalents, and cooling. Peak metamorphism (*c.* 61 Ma, ~8 kbar, > 800 °C; Schaub et al., 2002; Spear & Parrish, 1996) was immediately followed by the crystallization of leucosome (Fig. 9; Table S5). Several phases of the Ladybird leucogranite record crystallization ages of *c.* 61 Ma to *c.* 47 Ma (Parrish et al., 1988; Sevigny & Parrish, 1993). Exhumation is interpreted to have occurred as late as *c.* 51 Ma, based on zircon rim ages obtained through depth profiling (Gordon et al., 2009). The pattern of crystallization and cooling in the Thor-Odin dome is similar to that of the Valhalla dome (Fig. 9; Table S5). Data from the Thor-Odin dome are limited primarily to the timing of crystallization of leucosome and the Ladybird leucogranite, and cooling based on thermochronological data. Crystallization of the Ladybird leucogranite at *c.* 60 Ma (Vanderhaeghe et al., 1999) was followed by crystallization of leucosome, exhumation at *c.* 56 Ma, and cooling.

In the Bitterroot complex and Valhalla and Thor-Odin domes, crustal anatexis and pluton emplacement immediately preceded exhumation (Fig. 9). Additionally, significant crustal anatexis (diatexite migmatite) is observed within the lower units of the Valhalla and Thor-Odin domes (e.g., Gordon et al., 2008; Vanderhaeghe et al., 1999). This suggests that these complexes are examples of an end member scenario in which partial melting and magmatism occur during prograde metamorphism, resulting in thermal weakening and core complex exhumation. Thermal weakening followed by exhumation is consistent with models developed by researchers working in the Valhalla and Thor-

Odin domes who hypothesize that significant crustal anatexis led to diapirism within the middle crust, initiating low-angle rolling hinge detachment zones along which displacement was enhanced by syntectonic leucogranite laccoliths (Gordon et al., 2008; Norlander et al., 2002; Vanderhaeghe et al., 1999). Foster et al. (2001) suggest that Tertiary extension was focused in the Bitterroot core complex due to the large volume of magma likely present at the time.

Exhumation of the PRC was preceded by crustal anatexis (metatexite migmatite), but pluton emplacement at exposed levels followed the onset of exhumation by ~10 Ma (Figs 8, 9). Neither crustal anatexis nor plutonism preceded the exhumation of the Clearwater complex; pluton emplacement followed the onset of exhumation by ~7 Ma (Fig. 9; Table S5). This indicates that the PRC and Clearwater are examples of the second hypothetical scenario, in which pluton emplacement is primarily a response to decompression. Although exhumation in the PRC may have been influenced by partial melting, triggers other than partial melting must have been involved in the exhumation of the Clearwater complex.

5.7 Modeling Core Complex Exhumation

The varied relationships between crustal anatexis, pluton emplacement, and exhumation among the northern belt complexes indicate that exhumation triggers vary significantly. Individualized models for core complex development need not take these variations into account; however, attempts to develop generalized models, including numerical models, for the initiation and evolution of metamorphic core complexes must account for them. The two-dimensional numerical models of Tírel et al. (2008), Rey et al.

(2009a, b), and Bendick and Baldwin (2009) detail extension and exhumation of thickened continental crust during the development of core complexes, and are applicable to the development of core complexes in the northern North American Cordillera. Although the models vary in configuration, boundary conditions, and model output data, each produces a geologically plausible scenario for northern belt core complexes, while allowing for variation in model parameters (e.g., thermal conditions, extension rate, melt fraction).

An important comparison of these models is how each allows for the assessment of the hypothetical end member scenarios regarding melting and magmatism as either causes or consequences of core complex exhumation. The model of Tírel et al. (2008) imposes a pre-existing density-viscosity anomaly, such as a granitic pluton, to localize a detachment fault. Rey et al. (2009b) impose a crustal detachment to localize strain, although the authors address the possibility of a point anomaly, such as an intrusion. The Rey et al. (2009b) model incorporates partial melting, but model experiments suggest that partial melting is secondary to strain rate in determination of the form of the resultant complex. The Bendick and Baldwin (2009) model imposes a decrease in effective viscosity, preferring either crustal anatexis or pluton emplacement as the cause.

For all three numerical models described above, melting and/or magmatism (granitic intrusion, crustal detachment or intrusion, or decreased effective viscosity) figure in the initiation of exhumation of metamorphic core complexes. Furthermore, all three models permit, but do not require, near-isothermal decompression, allowing for decompression melting and pluton emplacement during retrograde metamorphism. Thus, all three models potentially apply to the exhumation of core complexes in the northern

Cordillera, with the possible exception of the Clearwater complex. Given the apparent absence of pre- or early synextensional magmatism in both the Priest River and Clearwater complexes, and the lack of Cretaceous-Eocene partial melting in the Clearwater complex, the Bendick and Baldwin (2009) model is preferred for its ability to address the variability in northern Cordilleran metamorphic core complexes.

6. Conclusions

Analysis of zircon from the Hauser Lake gneiss and the granitoids that intrude it provides constraints on the temporal relationships between metamorphism, crustal anatexis, magmatism, and exhumation in the PRC. The Spokane granite completed crystallization by 76.5 Ma, likely prior to or during early prograde metamorphism of the Hauser Lake gneiss. Distinct crystallization ages for the Silver Point and Wrencoë plutons indicate multiple pulses of crustal melting during exhumation, and constrain exhumation and deformation in the PRC footwall to no later than *c.* 50-48 Ma. The Silver Point and Wrencoë plutons were likely derived from PRC basement lithologies, with significant input from Paleoproterozoic orthogneisses, whereas partial melting of the Hauser Lake gneiss served as a likely source for the Spokane granite and leucocratic dike. Late pluton emplacement at exposed levels relative to the initiation of exhumation at *c.* 60 Ma suggests magmatism in the PRC is a response to, rather than a trigger of, exhumation.

The timing of magmatism relative to exhumation is variable for deeply exhumed core complexes in the northern North American Cordillera, indicating that the influence of partial melting and magmatism is variable among core complexes. Numerical models

for the exhumation of metamorphic core complexes are consistent with this observation, suggesting that melting and magmatism are influential in exhumation, but allowing for variability between core complexes.

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FIGURES

Fig. 1. Metamorphic core complexes of the northern North American Cordillera. (a) The northern, central, and southern belts of complexes in the North American Cordillera. (b) Core complexes in the northern belt of the North American Cordillera, including gneiss domes and complexes within the Shuswap complex. FC – Frenchman Cap; GF – Grand Forks; K- Kettle; O – Okanagan; SRP – Snake River Plain; T-O – Thor-Odin; V – Valhalla. (Modified from Stevens et al., 2015, and references therein)

Fig. 2. Simplified geologic map of the footwall of the Priest River complex. Sample locations for this study are shown. Cross section lines A-A' and B-B' are referenced in Fig. 9. (Modified from Lewis et al., 2012; Stevens et al., 2015; Wang, 2015)

Fig. 3. Zircon cathodoluminescence and monazite backscattered electron images for grains analyzed by CA-TIMS or ID-TIMS. Zircon analyzed by CA-TIMS for (a) Spokane granite, (b) Silver Point pluton, (c) Wrencoë pluton, and (d) the leucocratic dike cross cutting the HLG. (e) Monazite analyzed by ID-TIMS for the leucocratic dike within the HLG. Dashed lines show approximate boundary of zircon tips broken off for analysis; otherwise, whole zircon grains were analyzed. Monazite grains were broken into two fragments that were analyzed separately. The $^{206}\text{Pb}/^{238}\text{U}$ dates are given. Circles represent locations of LASS-MC-ICPMS analyses (Figs S2, S3).

Fig. 4. Archean–Proterozoic LA-ICPMS dates. (a) Concordia diagram for the Hauser Lake gneiss. (b) Concordia diagram for the Rathdrum granite. Error ellipses are 2σ and

are shaded by sample. (c) Histogram comparing metamorphic and detrital zircon $^{207}\text{Pb}/^{206}\text{Pb}$ LA-ICPMS dates from the Hauser Lake gneiss and Prichard Formation. Dates from the Prichard formation, the interpreted protolith of the Hauser Lake gneiss, are from quartzites (Lewis et al., 2010). Dates from the Hauser Lake gneiss from previous studies are from quartzites and gneisses (Doughty and Chamberlain, 2008; Lewis et al., 2010).

Fig. 5. Ranked $^{206}\text{Pb}/^{238}\text{U}$ date plots for Cretaceous–Tertiary LA-ICPMS zircon analyses. (a) Hauser Lake gneiss. (b) Spokane granite. (c) Rathdrum granite. (d) Leucocratic dike. Box heights are 2σ .

Fig. 6. Plots of $\varepsilon_{\text{Hf}(i)}$ vs. age for zircon analyses and a Concordia diagram for Archean–Proterozoic LASS-MC-ICPMS analyses. (a) Concordia diagram for Archean–Proterozoic LASS-MC-ICPMS analyses from the Hauser Lake gneiss and Rathdrum granite. (b) Plot of $\varepsilon_{\text{Hf}(i)}$ vs. age for all analyses from this study. Plot shows the evolution of Archean–Proterozoic basement (dark field) and Hauser Lake gneiss (light field), with crustal $\text{Lu}/\text{Hf} = 0.015$ (Goodge and Vervoort, 2006). (c) Detail of plot (a) for $\varepsilon_{\text{Hf}(i)}$ for analyses with $^{206}\text{Pb}/^{238}\text{U}$ dates < 100 Ma. Most error bars are smaller than, or approximately the same size as, the symbols representing the data.

Fig. 7. Zircon and monazite U-Pb dates TIMS dates. (a) Spokane granite. (b) Silver Point quartz monzonite. (c) Wrencoë granodiorite. (d) Leucocratic dike. Box heights are 2σ .

Fig. 8. Schematic cartoon depicting the evolution of the PRC. See Fig. 2 for the locations of A-A' and B-B'. A timeline (not to scale) shows the timing of metamorphism, magmatism, exhumation, and crystallization relative to the depicted stages. a) Emplacement of the Spokane granite at *c.* 76.5 Ma. (b) Peak metamorphism and crustal anatexis at *c.* 64 Ma. (c) The initiation of exhumation, *c.* 60 Ma. (d) Exhumation and emplacement of the Silver Point pluton, *c.* 50 Ma. (e) The final stages of exhumation and the emplacement of the Wrencoë pluton, *c.* 48 Ma. (f) The present day, shown with basement. NF – Newport Fault; PTF – Purcell Trench Fault; R – Rathdrum granite; S – Spokane granite; SD – Spokane Dome; SP – Silver Point pluton; W – Wrencoë pluton. Pluton locations depict depth of emplacement and do not imply a magma source.

Fig. 9. Comparison of the timing of metamorphism, magmatism, cooling, crystallization, and exhumation for deeply exhumed northern belt core complexes. Data is compiled from sources listed in Table S5.

TABLES

Table 1. Samples included in this study.

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(a)

(b)

Figure 2 - For Print

Figure 2 - For Web

(a) Spokane granite (PR-11-038)

(b) Silver Point quartz monzonite (PR-11-010)

(c) Wrencoee granodiorite (PR-11-011)

(d) Leucocratic dike (PR-11-014, zircon)

(e) Leucocratic dike (PR-11-014, monazite)

(a) Spokane granite (PR-11-038)

(b) Silver Point quartz monzonite (PR-11-010)

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(e) Leucocratic dike (PR-11-014, monazite)

Figure 4.

Figure 4 - For Web

(a) Hauser Lake gneiss, Archean–Proterozoic dates

(b) Rathdrum granite, Archean–Proterozoic dates

(c) Histogram for detrital dates from the Prichard Formation and Hauser Lake gneiss

Figure 5

(a) Hauser Lake gneiss

(b) Spokane granite (PR-11-038)

(c) Rathdrum granite (PR-11-005)

(d) Leucocratic dike (PR-11-014)

(a) Hauser Lake gneiss

(b) Spokane granite (PR-11-038)

(c) Rathdrum granite (PR-11-005)

(d) Leucocratic dike (PR-11-014)

Figure 6

Figure 6 - For Print

Figure 6 - For Web

Figure 7

Figure 7 - For Web

(a)

(b)

(c)

(d)

Figure 8 - For Print

Figure 9 - For Print

Figure 9 - For Web

Table 1

Table 1. Samples included in this study.

Sample	Unit	Location		Analyses		
		Latitude (°N)	Longitude (°W)	LASS-MC-ICPMS	LA-ICPMS	TIMS
PR-11-033B	Hauser Lake gneiss	48.0157	117.0057	(z)	z	
PR-10-055	Hauser Lake gneiss	48.3102	116.8040	z	z	
PR-11-040	Hauser Lake gneiss	48.5289	116.7557	z	z	
PR-11-038	Spokane granite	47.9197	117.1101	z	z	z
PR-11-010	Silver Point quartz monzonite	48.1191	117.2732	z		z
PR-11-011	Wrencoe granodiorite	48.2529	116.6116	z		z
PR-10-043	Rathdrum granite	47.9433	116.9238	z	z	
PR-11-005	Rathdrum granite	47.9598	116.9332	z	z	
PR-11-014	Leucocratic dike	48.1964	116.7199	z	z	z, m

Analyses: LASS-MC-ICPMS for zircon U-Pb geochronology and Hf isotopes; LA-ICPMS for zircon U-Pb geochronology and trace element geochemistry; TIMS for U-Pb geochronology.

Abbreviations: z = zircon; m = monazite. Abbreviations are given in () where analysis was unsuccessful.