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Oxygen isotope ratios in zircon and garnet: A record of assimilation and fractional crystallization in the Dinkey Dome peraluminous granite, Sierra Nevada, CA

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Abstract

The 119 Ma Dinkey Dome pluton in the central Sierra Nevada Batholith is a peraluminous granite and contains magmatic garnet and zircon that are complexly zoned with respect to oxygen isotope ratios. Intracrystalline SIMS analysis tests the relative importance of magmatic differentiation processes vs. partial melting of metasedimentary rocks. Whereas $\delta^{18}O$ values of bulk zircon concentrates are uniform across the entire pluton (7.7‰ VSMOW), zircon crystals are zoned in $\delta^{18}O$ by up to 1.8‰, and when compared to late garnet, show evidence of changing magma chemistry during multiple interactions of the magma with wall rock during crustal transit. The evolution from an early high-$\delta^{18}O$ magma [$\delta^{18}O$(WR) = 9.8‰] towards lower values is shown by high $\delta^{18}O$ zircon cores (7.8‰) and lower $\delta^{18}O$ rims (6.8‰). Garnets from the northwest side of the pluton show a final increase in $\delta^{18}O$ with rims reaching 8.1‰. In situ REE measurements show zircon is magmatic and grew before garnets. Additionally, $\delta^{18}O$ in garnets from the western side of the pluton are consistently higher (Ave = 7.3‰) relative to the west (Ave = 5.9‰).

These $\delta^{18}O$ variations in zircon and garnet record different stages of assimilation and fractional crystallization whereby an initially high $\delta^{18}O$ magma partially melted low $\delta^{18}O$ wallrock and was subsequently contaminated near the current level of emplacement by higher $\delta^{18}O$ melts. Collectively, the comparison of $\delta^{18}O$ zoning in garnet and zircon shows how a peraluminous pluton can be constructed from multiple batches of variably contaminated melts, especially in early stages of arc magmatism where magmas encounter significant heterogeneity of wall-rock assemblages. Collectively, peraluminous magmas in the Sierran arc are limited to small < 100 km^2 plutons that are intimately associated with metasedimentary wallrocks and often surrounded by later and larger, metaluminous tonalite and granodiorite plutons. The general

associations suggest that early stage arc magmas sample crustal heterogeneities in small melt batches, but that with progressive invigoration of the arc, such compositions are more effectively blended with mantle melts in source regions. Thus, peraluminous magmas provide important details of the nascent Sierran arc and pre-batholithic crustal structure.

Keywords: peraluminous granite, garnet, zircon, Sierra Nevada, oxygen isotopes, REE, SIMS

Introduction

The petrogenesis of peraluminous granites is a longstanding question (Clemens and Wall 1981; Patiño Douce and Johnston 1991; Frost et al. 2001; Villaros et al. 2009; Lackey et al. 2011). Peraluminous composition in granitoid rocks is defined by molar proportions of Al_2O_3 in excess of combined CaO, Na_2O, and K_2O: $Al_2O_3/(CaO+K_2O+Na_2O) > 1$, or an aluminum saturation index (ASI > 1; Zen 1988). Peraluminous granitoids typically form from a high proportion of melts of aluminous crustal or sedimentary source rocks (Chappell and White 1974; Scaillet et al. 2016). Alternative mechanisms can explain the generation of weakly peraluminous compositions in granitoid rocks (e.g. fractional crystallization, crustal anatexis and vapor phase transfer; Zen 1988).

Isotope tracers can discriminate between source, magmatic differentiation, and contamination characteristics of peraluminous magmas. Early work showed correlated $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ as indicative of crustal melting (O'Neil and Chappell 1977; Halliday et al. 1981). Oxygen isotope ratios are affected by assimilation of crustal rocks, which have different $\delta^{18}O$ values than mantle-derived magmas (e.g., Taylor and Sheppard 1986; Valley et al. 2005). In cases where crustal melts are produced from young source rocks, radiogenic isotopes are not sensitive, and oxygen isotopes are typically the most sensitive isotopic tracer (e.g., Valley 2003; Lackey et al. 2011; Jeon et al. 2012).

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Oxygen isotopes can be measured in retentive zoned minerals to record information about magma evolution (e.g., Valley 2003; Bindeman 2008; Lackey et al. 2011). Self-diffusion rates of oxygen in garnet and zircon are among the slowest in common minerals (Coughlan 1990; Wright et al. 1995; Watson and Cherniak 1997; Vielzeuf et al. 2005; Page et al. 2007a, 2010; Bowman et al. 2011), and crystallization of both minerals in peraluminous granites allows them to be used in tandem to record a more complete time history than would be provided by a single mineral (Lackey et al. 2011). The $\delta^{18}O$ values of zircon and garnet are quenched upon crystallization and growth zoning provides a record of magmatic evolution (King and Valley 2001; Valley 2003; Lackey et al. 2006).

Zircon and other accessory minerals also record the rare earth element (REE) compositions of felsic magmas during their growth (Sawka and Chappell 1988; Hoskin et al. 2000; Hoskin and Schaltegger 2003). Rare earths are incorporated in zircon by coupled substitution mechanisms (Speer 1982; Hinton and Upton 1991; Halden et al. 1993; Hoskin and Ireland 2000; Finch et al. 2001; Hoskin and Schaltegger 2003).

In this study, we employ secondary ion mass spectrometry (SIMS) to measure oxygen isotope ratios ($\delta^{18}O$) and trace element compositions, including Y + REEs. The SIMS method provided accurate and precise measurements of intracrystalline zoning at high spatial resolution (ca. 10 μm) in zircon and garnet crystals collected throughout the Dinkey Dome granite (Fig. 1). The zoning measured within these crystals is useful to contextualize contamination, assimilation, and/or high temperature alteration processes during growth of both zircon and garnet.

The resulting data constrain models for the origin and contamination of silicic melts in the Sierra Nevada batholith. The processes that formed this and other granites senso stricto in the Sierra are critical to understand the relative contribution of preexisting crust in the Sierran arc and evaluate the different processes that affected the composition of final magmas. Thus, in situ analysis of refractory magmatic minerals helps in discriminating the relative amounts of crustal contribution by: crystal fractionation and partial melting of mafic magmas, where no crustal contribution is required in the production of felsic magmas (Ratajeski et al. 2001, 2005; Wenner and Coleman 2004); contamination, assimilation and fractional crystallization whereby magmas become silicic (Lackey et al., 2005, 2006, 2008; Nelson et al. 2013); and wholesale melting of crustal material by deep heat sources with no contribution of mafic material (Holden et al. 1987).

Geology

The Sierra Nevada Batholith

The voluminous Cretaceous Sierra Nevada batholith, California (Fig. 1) consists mainly of tonalite to granodiorite plutons to depths of ~35 km (Saleeby et al. 2003), with more mafic diorite and refractory gabbroic residues continuing to ca. 45 km (Fliedner et al. 2000). Gabbro complexes and mafic enclaves are a common but volumetrically small part of the batholith and have been targeted to study the mass balance of mantle and crustal melt inputs to produce the intermediate, granodiorite compositions that are the bulk of the batholith (Dorais et al. 1990; Coleman et al. 2004; Wenner and Coleman 2004). Other studies have examined the sub-arc mantle and residual mafic root of the batholith, sampled as pyroxenite, garnet-clinopyroxenite, and lherzolite xenoliths in Cenozoic volcanic rocks (Moore and Dodge 1980; Ducea 2001; Lee et al. 2006; Chin et al. 2014). Such studies provide additional information on mantle controls of magmatic heat budgets and mafic magma flux, revealing in detail that multi-stage crystallization and re-melting episodes are required to build granodioritic crust that complements the major element (e.g., Mg) and isotopic compositions of xenoliths. In addition, experimental studies show that high silica melts can be produced from re-melting of Sierran gabbros (Sisson et al. 2005; Ratajeski et al. 2005)

Despite considerable attention to magmatic origins recorded in mafic to ultramafic rocks, few studies have focused on potential high-silica melts; $\delta^{18}O$ studies of granodiorite and tonalite suites require at least 15-30% input of melts from supracrustal sources, thus partial melting of gabbros is not the sole source of potential high-silica end-member melts. Thus, direct studies of rocks with relatively undiluted high-silica crustal melts are important, but only a handful have been undertaken: Wenner and Coleman (2004) studied several granites in a regional survey of both mafic and felsic plutons in the Sierra; Zeng et al. (2005) examined partial melting in a lower crustal migmatite complex in the Southern Sierra Nevada; Lackey et al. (2006) studied regional and pluton-scale patterns of $\delta^{18}O$ of peraluminous granites in the Sierra. These three studies found evidence of crustal melting in the granitic plutons and migmatites, highlighting the importance of such melts as a factor in the isotopic variability in many Sierran granodiorites, hence, added motivation to study the Dinkey Dome granite.

The Dinkey Dome granite

The garnet, two-mica Dinkey Dome granite is a relatively small (~30 km^2) pluton surrounded by granodiorite plutons (e.g., Dinkey Creek Granodiorite) and other granites of the Shaver Intrusive Suite (Figs. 1 and 2; Bateman 1992; Lackey et al. 2006). With a U-Pb zircon age of 119 Ma (Frazer et al. 2008), the Dinkey Dome pluton is coeval with the oldest members of the Fine Gold Intrusive Suite to the west, and significantly older than other members of the Shaver Intrusive Suite (Frazer et al. 2008). Thus, the Dinkey Dome represents a case of magmatism that was anomalously inboard of the broad magmatic “locus” in the Sierra at the time it was emplaced, and a departure from the broad trend of eastward-younging Cretaceous intrusive suites in the Sierran Arc (Chen and Moore 1982; Memeti et al. 2010; Davis et al. 2012; Lackey et al. 2012; Ardill et al. 2018; Chapman and Ducea 2019). Initial $^{87}Sr/^{86}Sr$ ratios of the Dinkey Dome granite are 0.7065 (Kistler and Peterman 1973). Metamorphic wallrocks consist of quartzite, mica schist, biotite hornfels, and marble (Fig. 2) (Bateman and Wones 1972). The chemistry of the pluton and aluminous minerals (garnet, muscovite, Al_2SiO_5) contained therein are typical for peraluminous granites. Whole rock geochemical analyses of the Dinkey Dome show that aluminum saturation indices (ASI) are peraluminous (west side ASI = 1.02; east side ASI = 1.07; Lackey et al. 2006). Average garnet compositions in the Dinkey Dome granite are $Alm_{72.4}Sps_{19.5}Pyp_{5.8}Grs_{2.3}$ on the west and $Alm_{78.6}Sps_{19.1}Pyp_{0.9}Grs_{1.4}$ on the east. There are no garnet-bearing metamorphic wallrocks in contact with the Dinkey Dome, consistent with the magmatic origin of garnet (Lackey et al. 2011). Shallow crystallization is inferred by scattered miarolitic cavities in the east side of the pluton that are inferred to indicate a pressure of <1 kbar (Wones et al. 1969). The preservation of coarse, euhedral books of muscovite, suggests some of the muscovite formed at depth, and was preserved during ascent and final crystallization of the magma. The Dinkey Creek Granodiorite, which is ~8 million years younger than the Dinkey Dome granite (Frazer et al. 2008), engulfs the Dinkey Dome pluton and its pendant. Al-in-hornblende pressure estimates of the Dinkey Creek Granodiorite are 4.0 ±0.4 kbar, from 10 widely distributed samples collected by Ague and Brimhall (1988a) and recalculated by Tobisch et al. (1993). The pressures derived from hornblende in the Dinkey Creek Granodiorite imply considerable difference of depth, although, hornblende may just record early, deeper magmatic conditions which is plausible given increasing evidence of hornblende populations showing a continuum of magma conditions (Barnes et al. 2017). Nevertheless, the pressure/age differential implies 10 km of burial of the Dinkey Dome granite and its pendant rocks in 8 million years.

Construction of younger batholith rocks with higher apparent pressures around pendants and older plutons that are relatively shallow is seen elsewhere in the Sierra Nevada. For instance, recent oxygen isotope analysis of skarn garnets in the Mineral King pendant (85 km SE) shows paleo-hydrothermal systems were infiltrated by meteoric water at ca. 135 Ma (Ryan-Davis et al. 2019) and again at ca. 109 Ma (D'Errico et al. 2012), but that younger (98 Ma) voluminous granodiorite plutons surround the pendant and these hydrothermal systems record apparent emplacement pressures of ca. 3 kbar (Ague and Brimhall 1988b), deeper than the possible brittle-ductile transition that would permit extensive meteoric water circulation. Similarly, volcanic rocks in the Ritter Range pendant (60 km N) have steep, down-dip stretching lineations and are found adjacent to slightly younger plutonic rocks (Tobisch et al. 2000). This juxtaposition of younger shallow plutons against older, higher pressure rocks, suggests that the older rocks were engulfed by later magmas as the batholith is built around them through interplays of bulk-arc thickening, structural shortening, or a density driven settling (e.g., Glazner and Miller, 1997).

Petrography. The Dinkey Dome granite contains quartz, plagioclase, K-feldspar, zircon, biotite and muscovite, and commonly has garnet, perthite, sericitized feldspar, granophyre and myrmekite. Accessory phases include zircon, monazite, apatite, magnetite and ilmenite. Andalusite and sillimanite show scattered occurrence on the eastern side of the pluton (Guy, 1980). Molybdenite and uraninite have been reported on the eastern side as well (Lackey et al., 2006). Andalusite has textural traits indicative of magmatic crystallization, including uniformly sized and distributed, euhedral to subhedral grains free of carbonaceous chiastolite inclusions that are typically associated with metamorphic andalusite (Clarke et al. 2005). Where observed, fibrolitic sillimanite is uniformly distributed in the granite and does not appear to form at the expense of andalusite or vice versa. The concentration of fibrolite toward the interior of the pluton was interpreted to record high temperatures in the interior of that domain that may have lasted longer. Guy (1980) also noted fine-grained, secondary muscovite replacing andalusite and sillimanite, likely forming under subsolidus conditions (Guy, 1980), unlike early, phenocrystic muscovite; thus, muscovite and aluminosilicates in the pluton likely integrate varying P-T conditions as magma ascended, was contaminated, and crystallized. Analyses of $\delta^{18}O$ (And) in two samples showed values consistent with magmatic crystallization; values of 8.5‰ are ~2‰ lower than pluton $\delta^{18}O$ (WR) values whereas adjacent wallrock $\delta^{18}O$ values are 3-4‰ higher than andalusite and inconsistent with high temperature equilibrium. Values of $\delta^{18}O$ of fibrolite from one sample are similar to co-existing andalusite, a result consistent with crystallization of the two minerals from the same magma (Lackey et al. 2006).

Garnet and Zircon: occurrence, morphology and internal structures. Magmatic garnet occurs throughout the Dinkey Dome and is found in all the samples collected in this study and by Lackey et al. (2006). Unlike other peraluminous plutons in the Sierra Nevada where garnet is concentrated near contacts, magmatic garnet in the Dinkey Dome occurs throughout the entire pluton. These crystals are generally subhedral to euhedral and range in size from 200-2000 µm. The pluton contains both pink and red garnet. Garnet in the west side of the pluton is darker (red) vs. lighter (pink) in the east side. BSE imaging reveals subtle concentric oscillatory growth-zoning within these grains. Garnet grains aren't generally inclusion-rich and the distribution of inclusions from grain-to-grain is not uniform throughout the samples. Garnet crystals contain inclusions of plagioclase, K-feldspar, muscovite, biotite, quartz, monazite, apatite, ilmenite and zircon (Fig. 3).

Zircon in the Dinkey Dome granite occurs as euhedral crystals that range in size from 20 to 300 µm but are generally ~100 µm long and 25-50 µm wide. In some cases, zircon crystals occur as inclusions within garnet (Figs. 3c and d). Conversely, some zircon grains contain inclusions of quartz, K-feldspar, apatite, plagioclase, biotite, ilmenite, and magnetite that have been identified by EDS. Ortiz (2010) examined 50 zircons by SEM in a polished grain mount from sample 1S79 and found apatite inclusions in 9 zircon grains, K-feldspar in 8, quartz in 4, biotite in 2, and Fe-Ti oxide in 1. BSE and CL imaging reveals oscillatory zoning; convolute zoning is present in some zircon cores. Representative textures are seen in Figure 4.

Previous isotopic work

Lackey et al. (2006) analyzed $\delta^{18}O$ by laser fluorination of garnet (Grt), zircon (Zrn), quartz (Qz), andalusite (And) and whole rock powders (WR) in the Dinkey Dome pluton (Fig. 2) as part of a regional study of peraluminous granitoid plutons in the Sierra Nevada. Their study produced several key results. First, values of $\delta^{18}O$ are elevated: $\delta^{18}O$(WR) = 9.6–10.4‰ VSMOW, $\delta^{18}O$(Qz)= 10.6–11.3‰, $\delta^{18}O$(And)= 8.4–8.5‰, $\delta^{18}O$(Zrn)= 7.0–7.8‰, and $\delta^{18}O$(Grt)= 6.7–7.4‰. The surrounding Kings Sequence metasedimentary rocks (marbles, hornfels, and quartzites) of the Dinkey Creek pendant have $\delta^{18}O$(WR) of 9.5–11.6‰. Other peraluminous granites in the Sierra Nevada Batholith also have high $\delta^{18}O$(Zrn) >7.5‰, while metaluminous granitic rocks near the Dinkey Dome have average zircon $\delta^{18}O$ values that range from 6.5–7.5‰ (Lackey et al. 2006, 2008).

Values of $\delta^{18}O$ for zircon, quartz and whole rock are unimodal across the entire pluton, however, $\delta^{18}O$ values of magmatic garnet are bimodal, decreasing by ~0.6‰ on the east side of the central metasedimentary septum within the Dinkey Dome (Fig. 2). High $\delta^{18}O$ values and equilibrium fractionations of garnet and zircon on the west side of the pluton ($\Delta^{18}O$(Grt-Zrn) = 0.06 ±0.13‰), indicate that prior to the crystallization of both minerals, the magma was elevated in $\delta^{18}O$. On the eastern side, $\delta^{18}O$ values of garnet are lower and not equilibrated with zircon

($\Delta^{18}O$(Grt-Zrn) = –0.6 ±0.13‰), recording a change in magmatic $\delta^{18}O$ synchronous with crystallization. These differences in fractionation are small but distinct. The lower $\delta^{18}O$ values in garnet that formed later than zircons, seen as inclusions in garnet (Figs. 3c and d), suggest that low $\delta^{18}O$ material was assimilated after the crystallization of zircon and before crystallization of garnet. These results are interpreted to indicate that the Dinkey Dome granitic magmas evolved through contamination by low $\delta^{18}O$ material, however no low $\delta^{18}O$ country rocks are exposed in the Dinkey Creek Pendant. Thus, partial melting of such a low $\delta^{18}O$ contaminant would be required to have occurred deeper in the crust (Lackey et al. 2006).

Methods

Sample preparation and imaging

Zircon and garnet mineral separates of ten samples from Lackey et al. (2006) (samples 1S51 to 1S82) were handpicked and cast in 25-mm diameter round epoxy mounts along with the Kim-5 zircon (Valley 2003) and UWG-2 garnet (Valley et al. 1995) standards, ground to the level of best mineral exposure, polished to a smooth, flat, low-relief surface, and carbon coated prior to imaging. Secondary Electron (SE), Backscattered Electron (BSE) and Cathodoluminescence (CL) images were obtained for each grain, and Energy Dispersive X-ray Spectrometry (EDS) was conducted using the UW-Madison, Dept. of Geoscience Hitachi S-3400N SEM (Scanning Electron Microscope). Based on the images obtained, approximately 10 zircon grains (that display distinctive rims and cores) and five garnet grains were chosen for in situ analysis from each sample. The SEM images were also used to locate the positions for SIMS analyses. Carbon coats were then removed, and the mounts were coated with gold for SIMS analysis.

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Twenty-five-mm round thin sections were made of samples collected during this study (samples 10DD02-10DD19, Table 1, Supplementary Table A) (Fig. 2) and top mounted with UWQ-1 quartz standard (Kelly et al. 2007) and UWG-2 in their centers. Isotopic analysis of minerals in thin section rather than in grain mounts permits detailed descriptions of zircon and garnet that are in known petrographic relation to each other. Imaging prior to SIMS analysis was conducted in the same manner as described above.

Major and minor element analyses of garnet by electron microprobe

Determination of composition and testing for compositional zoning preceded every SIMS $\delta^{18}O$ analysis of garnet. Major and minor element analyses of garnet were obtained using the UW-Madison, Dept. of Geoscience CAMECA SX51 electron microprobe by wavelength dispersive spectrometry. Eight elements were analyzed: Si, Al, Fe, Mg, Mn, Ca, Ti, and Cr (Supplementary Table B). The operating conditions were accelerating potential of 15KeV, 40° takeoff angle, and a fixed focused beam at 20 nA. Counting time for all elements was 10 seconds on-peak and 10 seconds off-peak. LIF, PET, and TAP analyzer crystals were used to acquire Kα X-ray intensities for Mn, Fe and Cr; Ca and Ti; and Al, Si and Mg, respectively. Crystalline standards were used: Minas Gerais rutile for Ti; U.W. synthetic fayalite for Fe; synthetic tephroite for Mn; USNM 143968 Kakanui pyrope for Mg; $Andradite_{99}$-Rota (Hungary) for Ca; synthetic Cr_2O_3 for Cr; and HU $Almandine_{56}$ for Al and Si.

Laser Fluorination analysis of $\delta^{18}O$

In order to assess any correlation of chemical composition with $\delta^{18}O$, individual garnet grains (~1.5-2.0 mg) from 10 samples previously studied by Lackey et al. (2006) were analyzed in the UW-Madison, Dept. of Geoscience Stable Isotope Laboratory by laser fluorination using BrF_5 as the reagent. A dual-inlet gas-source Finnigan/MAT 251 mass spectrometer was used to

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measure isotope ratios. Standardization was done using UWG-2 ($\delta^{18}O$ = 5.80 ‰ VSMOW), which provides high precision and accuracy in laser analyses (Valley et al. 1995). Values of $\delta^{18}O$ of whole rock powders (~2mg) of Dinkey Dome granite and Dinkey Creek sedimentary rocks were analyzed by laser fluorination using an airlock sample chamber (Spicuzza et al. 1998) (Supplementary Table E).

SIMS analysis of $\delta^{18}O$ in garnet and zircon

Oxygen isotope ratios were measured at the WiscSIMS Laboratory, Department of Geoscience, UW-Madison with a CAMECA ims-1280 large-radius multicollector ion microprobe/SIMS (Kita et al. 2009; Valley and Kita 2009). Oxygen isotopes were analyzed using a 2.0–2.2 nA primary Cs^{+} beam accelerated by 10 kV (impact energy = 20 kV) and focused on sample surface with ~10–12 µm spot diameter. Secondary ^{16}O and $^{18}O^{-}$ ions were measured by two Faraday cup detectors simultaneously. Zircon standard KIM-5 ($\delta^{18}O$ = 5.09 ‰ VSMOW; Valley 2003) and garnet standard UWG-2 ($\delta^{18}O$ = 5.80 ‰ VSMOW; Valley et al. 1995) were mounted in the center of each sample and used as running standards to bracket unknown sample analyses. Four consecutive measurements of the standard were made before and after every set of 10 sample analyses. Additional standardization and calibration of garnet standards was performed to account for the compositional effects on instrumental bias as described previously (Page et al. 2010, Russell et al. 2013, Kitajima et al. 2016). Typically, two analysis spots were made on each zircon (core and rim), and approximately 3-8 spots (rim to rim) on each garnet.

Cracks, inclusions, radiation-damaged zircon domains and other features that can compromise an analysis were avoided by secondary electron (SE), backscattered electron (BSE), and cathodoluminescence (CL) imaging of minerals before in situ analysis. In addition, all SIMS pits were imaged post-analysis using BSE and SE, and pits that hit cracks or contain mineral inclusions are culled from the final dataset (Supplementary Tables C-D).

SIMS analysis of REEs in zircon

Zircon grains were analyzed by SIMS for trace elements, including rare earth elements (REEs) in the WiscSIMS Laboratory at UW-Madison with a CAMECA ims-1280. The following elements were analyzed: Li, Si, P, Ca, Ti, V, Fe, Y, La, Ce, Pr, Nd, Sm, Eu, Tb, Gd, Dy, Ho, Er, Tm, Yb, Lu, Hf, Th, and U (Supplementary Table F). Similar conditions as Page et al. (2007b) were used: impact energy of 23 kV, 4 nA O^- ion beam shaped to a diameter of 25 μm on the sample surface and a secondary ion accelerating voltage of 10 kV. For trace element analysis the configuration of the secondary ion optics was optimized for high transmission (Kita et al. 2009). A single electron multiplier, field aperture of 4000 μm, MRP of 3000, and secondary beam energy offset of 40 V were used and allow resolution of the selected REE peaks from the interfering REE oxides. Measured counts for each element were normalized to ^{30}Si. During the analysis session, NIST-610 glass was used as a running standard. To estimate the matrix effects on relative sensitivity factor (RSF) between zircon and NIST 610, the zircon standards 91500 (Y, REE, Hf, Th and U: Wiedenbeck et al. 2004) and Xinjiang (Li: Ushikubo et al. 2008) were analyzed at the beginning of the trace element session. For Ti concentration, we used the correction factor on RSF between zircon and NIST-610 reported by Fu et al. (2008). No correction for matrix effect was applied on P, Ca, V and Fe because their concentrations in the 91500 zircon are unknown. Counting times were adjusted for NIST-610 because of the difference in REE composition in comparison to natural zircons (Page et al. 2007b) (Supplementary Table F).

Results

Garnet composition by EPMA

Garnet grains from the Dinkey Dome pluton are almandine-spessartine-rich with minor pyrope and grossular (X_{Alm}= 0.60-0.86; X_{Sps} = 0.11-0.27; X_{Pyp} = 0.01-0.07; X_{Grs} = 0.02-0.06) **(**Fig. 5a). Most of the garnets analyzed in this study slightly increase in spessartine and decrease in almandine at the rims. Internal cation zoning is generally subtle, however crystals 1S52-02, 1S52-04, 1S80-04, and 10DD07b-02 show bell-shaped rim-to-rim profiles. Compositionally, garnet is similar to garnet from other Sierran granitoids (Guy and Wones 1980, Calk and Dodge 1986, Ague and Brimhall 1988a, Liggett 1990; Lackey et al. 2006). Values of X_{Grs} are higher in western side of the pluton, suggesting slightly higher crystallization pressures. Crystals from the eastern side of the pluton have less pyrope and are generally more almandine-rich (Fig. 5a), which can explain the difference in color east to west.

Oxygen isotope ratios by Laser Fluorination

Laser fluorination analyses of oxygen isotope ratios in garnet from a west to east traverse (A-A', Fig. 2 and 6) were conducted to assess variations between pink garnet (lower X_{Alm}) and red garnet. Only one of the samples analyzed (1S51) shows a variation in $\delta^{18}O$ between red (7.25 ±0.23‰ 2SD) and pink garnet (6.80 ±0.23‰); the other samples showing different color grains (1S77, 1S79, and 1S82) show no variation in $\delta^{18}O$.

Individual garnets from samples 1S51, 1S52, 1S53, 1S77, 1S79, 1S80, and 1S81 were handpicked and analyzed by both laser fluorination and SIMS (Supplementary Table E). Values of $\delta^{18}O$ obtained by laser fluorination for these grains average 6.90 ±0.18‰ (1SD) for the eastern side and 7.63 ±0.17‰ to the west. These values are similar to those obtained Lackey et al. (2006), who also reported lower $\delta^{18}O$ for garnet from the eastern part of the pluton.

Whole rock analyses where also made of granite and metasediment samples 10DD-02 through 10DD-22 (Table 1). The granite $\delta^{18}O$(WR) values range from 9.0 to 10.5‰, and the metasedimentary rocks (biotite hornfels and quartzite) range in $\delta^{18}O$(WR) from 11.7 to 12.8‰.

Oxygen isotope ratios by SIMS

Garnet. Garnets from the western side of the pluton analyzed by SIMS resemble the laser fluorination analysis (within uncertainty). In contrast, garnets from the eastern side of the pluton show consistently lower values relative to laser fluorination analysis, with a difference of $\delta^{18}O$ values ranging from 0.6 to 1.5‰ (Fig. 6b) (Table 1). These differences likely result from quartz inclusions within garnet crystals that are higher in $\delta^{18}O$ and were unavoidably analyzed by laser fluorination. The SIMS analyses avoid inclusions that are plainly visible in polished surfaces and thus SIMS values of $\delta^{18}O$ are not affected by inclusions.

Values of $\delta^{18}O$ in epoxy-mounted garnet grains from western side of the pluton are higher (average $\delta^{18}O$(Grt) = 7.4 ±0.2‰) than eastern side $\delta^{18}O$(Grt) values of 6.3 ±0.2‰, and show no significant core to rim zoning in $\delta^{18}O$. Average values of $\delta^{18}O$ measured from garnets selected in thin section show a similar trend with higher values on the western side (6.9 ±0.3‰) and lower values on the eastern side (5.2 ±0.3‰). However, unlike garnet hand-picked from mineral separates, garnets in thin section show variation in $\delta^{18}O$ from rims to cores (Fig.7). The core to rim variation is more prominent on the larger (>1 mm) garnets from the northwestern side of the pluton. The zoning of the eastern-side garnets is more subtle and less common (Table 1).

This difference likely results due to analysis of larger subhedral garnets in thin section rather than the smaller equant garnets that were selected from mineral separates. It is also possible that the low $\delta^{18}O$ garnets are more delicate and were destroyed by the disk mill during sample processing. The $\delta^{18}O$ in garnets from thin sections on the east side is very low (ave. = 5.2

±0.3‰). This pattern, combined with the observation of large miarolitic cavities in the east side of the pluton suggests the possibility of the garnet rims growing into the sub-solidus realms, with some non-magmatic water infiltrating the system. Studies of Cretaceous skarns in the south-central Sierra show that garnet growing in shallow hydrothermal systems may record multiple episodes of fluid flow and low-$\delta^{18}O$ domains record incursions of meteoric water at different times, including waning stages of garnet growth (e.g., D'Errico et al. 2012; Ryan-Davis et al. 2019). Therefore, the Dinkey Dome garnet may record some surface water infiltration on the East side.

Zircon. SIMS analyses of rims and cores of individual zircon grains from 10 Dinkey Dome samples along the A-A' traverse (Fig. 2, 1S51-1S82) show constant $\delta^{18}O$ values for the cores: 7.8 ±0.3‰ on the east side and 7.7 ±0.3‰ on the west side. The rims of the zircons have consistently lower $\delta^{18}O$ values that average 6.7 ±0.3‰ on the east side and 6.9 ±0.3‰ on the west side (Fig. 6a).

SIMS $\delta^{18}O$ values of zircon in thin sections from the eastern side average 7.2 ±0.2‰ (Fig. 6a, Table 1). The average $\delta^{18}O$ in each zircon from the western part of the pluton is 7.6 ±0.2‰. These values are consistent with SIMS data for zircon cores (7.7 to 7.8‰) that dominate the mass of each zircon. Zircons from some samples (10DD-02a-b, 10DD-05a, 10DD-16c, 10DD-17, and 10DD-19c) did not have rims that were distinguishable by CL.

Trace elements in zircon

Trace element compositions in cores and rims of grains from the Dinkey Dome granite are summarized in chondrite-normalized REE diagrams (Fig. 8). The REE data are consistent with igneous zircon from continental crust (Belousova et al. 1998; Hoskin and Ireland 2000;

Belousova et al. 2002, Grimes et al. 2007) and show HREE enrichment, a positive Ce anomaly and a negative Eu anomaly (Fig. 8).

All zircon data plot within the 'magmatic' field in REE discriminant diagrams: $(Sm/La)_N$ vs. La (ppm) and Ce/Ce* $((Ce)_N/\sqrt{((La)_N(Pr)_N})$ vs. $(Sm/La)_N$ (Figs. 9a and b). None of the cores or rims have REE compositions similar to hydrothermal zircon (Hoskin 2005). Although Chondrite normalized REE patterns are similar in cores and rims of grains (Fig. 8), $(Sm/La)_N$ vs. La (ppm) and Ce/Ce* vs. $(Sm/La)_N$ are clearly bimodal, with rims having slightly flatter LREEs and being higher in [La] and lower $(Sm/La)_N$, (Figs. 9a and 9b).

Discussion

Causes of $\delta^{18}O$ zoning

The in situ measurements of $\delta^{18}O$ and trace elements from magmatic garnet and zoned zircon grains reveal a more complex magmatic history of assimilation and fractional crystallization for the Dinkey Dome granite than was resolved by bulk-mineral analysis. High-$\delta^{18}O$ zircon cores crystallized from an initially high-$\delta^{18}O$ magma derived by melting of a high-$\delta^{18}O$ source deeper in the crust (Figs. 10 and 11). The zircon cores average 7.7‰, indicating $\delta^{18}O$(magma) values of 9.4‰ [~69 wt. % SiO_2; $\Delta^{18}O$(WR-Zrc) ≈ 0.0612 (wt.% SiO_2) – 2.5‰ (Lackey et al. 2008)]. The inclusions of zircon in garnet and the steep positive slope of HREEs in zircon indicate that the majority of garnet grew after zircon. Lower $\delta^{18}O$ values in the rims of zircon (ave. 6.8‰) and throughout most garnets show that a lower $\delta^{18}O$ contaminant, possibly hydrothermally altered rocks, contributed some low-$\delta^{18}O$ melt into parts of the magma at depth. Quartz from the Dinkey Dome and whole rocks (including feldspars) do not record such low $\delta^{18}O$ values (Lackey et al. 2006), however absence of low-$\delta^{18}O$ values in quartz and feldspar could arise from exchange of oxygen isotopes between these minerals with igneous fluids

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contributed from younger, more voluminous magmas (e.g., the 101 Ma Dinkey Creek Granodiorite) that engulfed the Dinkey Dome pluton and its aureole. The assimilation and fractional crystallization history of Dinkey Dome magma thus appears preferentially preserved in zircon and garnet due to the minerals' slower diffusion rates relative to quartz and feldspar.

It is significant that no low $\delta^{18}O$ (< 5‰) rocks are identified in the pendant immediately adjacent to the Dinkey Dome pluton. Thus, the lower $\delta^{18}O$ domains in garnet and zircon point to this stage of melting and contamination of the magma at depths greater than final crystallization depths. The wallrock in the Sierran arc is heterogeneous by nature, containing domains of Triassic and Jurassic hydrothermally altered volcanic wallrocks with relatively low $\delta^{18}O$ (e.g., Peck and Van Kooten 1983; D'Errico et al. 2012; Ryan-Davis et al. 2019). Such metavolcanic and metasedimentary wallrocks are the most likely source of a low-$\delta^{18}O$ assimilation signature. Evidence of melting is found where migmatite complexes are developed in metavolcanics rocks at mid- to lower-crustal level pendants in the southern Sierra Nevada (Saleeby et al. 2003). Nevertheless, thermal budgets of peraluminous magma should limit significant melting and assimilation of wallrock at emplacement levels of the Dinkey Dome pluton, or mixing of magmas, thus the low-$\delta^{18}O$ assimilation may be restricted to a thin veneer of the eastern half of the pluton, where garnets record the lowest $\delta^{18}O$ values. The eastern part is also at a higher elevation, and thus it is possible the isotopic signatures are restricted to a thin copula (Fig. 11), and/or that the isotopic signatures of the eastern side are correlative to rocks eroded away from the western side. Discrete zoning is shown by the heterogeneous nature of the Dinkey Dome pluton; the northwestern samples show a rim-ward shift to higher $\delta^{18}O$ in some garnet rims suggesting that as some of the magma was produced and transported, it encountered high-$\delta^{18}O$ rock and was locally contaminated (Fig. 11).

Episodic contamination of a peraluminous magma

The findings from this work also raise the question of how, given their small size and limited thermal budgets, high silica magmas may episodically interact with wallrocks in arc crust. This process is partially illustrated at the pluton scale by the Hall Canyon pluton in the Panamint Range, California (Mahood et al. 1996). Here, a roof zone with pegmatitic and aplitic domains is enriched in peraluminous minerals (garnet, muscovite) compared to lower in the pluton, however the entire pluton is peraluminous. The authors invoke in situ fractionation of the magma in the upper roof zone with additional melts also percolating up into the roof zone from the lower reaches of the pluton. In such a scenario, fractionating melt increases peraluminosity (such as seen in the eastern Dinkey Dome pluton) and promotes additional growth of peraluminous minerals (e.g. garnet, sillimanite, and andalusite). Because fractionating peraluminous magmas sees increased concentrations of water and incompatible elements, this fractionation would counteract the tendency of cooling and crystallization to impede distribution of new melt into the "mushy" roof zone of the pluton (e.g., Scaillet et al. 2000). A similar process might have occurred in the eastern Dinkey Dome pluton whereby a more highly fractionated roof zone continued to receive melts and consequently achieved the isotopic heterogeneity seen in crystal-scale $\delta^{18}O$ zoning that was not recorded in the western domain. Sustained melt and fluid percolation in the east side also could have allowed low-$\delta^{18}O$ values to be recorded in some generations of garnet that continued to grow as peraluminosity increased. In contrast, the west side of the pluton crystallized deeper and relatively earlier and thus did not record incorporation of the low-$\delta^{18}O$ material and thus was more homogeneous in its composition. Thus, the east and west sides of the Dinkey Dome pluton behaved as two magma batches, separated by a septum of metasediments, that accumulated, fractionated, and ultimately crystallized.

Timing and preservation of crustal melts in arcs

Given that the Dinkey Dome pluton was emplaced into its pendant rocks earlier than other plutons in the Shaver Intrusive suite, it would have been emplaced inboard of the main locus of magmatism in the arc, and likely encountered a thicker, more heterogeneous crustal column and interactions with that crust superimposed additional contamination on the magma. The Dinkey Dome is not the only example in the region. The Grant Grove peraluminous granite, which shares similarities with Dinkey Dome, like being isolated by pendant rocks and surrounded by younger metaluminous plutons, shows higher $\delta^{18}O$ in its margin indicative of localized contamination (Lackey et al. 2006). That localized "veneer" of later contamination is manifested by additional growth of garnet and aluminosilicates at the margin of the pluton where it intruded and partially melted schists near emplacement levels.

Although of Jurassic age, peraluminous plutons that intrude the Julian Schist in the Peninsular Ranges batholith in southern California are comparable (Shaw et al. 2003). These granites are relatively small compared the younger (Cretaceous) tonalite and granodiorite plutons that surround them, comprising most of the Peninsular Ranges batholith. Some of the Jurassic plutons are directly associated with migmatitic zones in the Julian Schist and have elevated Sr_i (>0.71) and $\delta^{18}O$ (16-20‰), values that overlap with the schist itself indicating it was the source of the melts that produced the peraluminous plutons. Though of much greater age difference than the Dinkey Dome and younger plutons that surround it and associated pendant rocks, the progression from early, small peraluminous plutons to larger metaluminous plutons is the same. Unlike the Peninsular Ranges example, the Dinkey Dome does not have evidence of a localized migmatite complex, consistent with its final shallow (low pressure) emplacement, although there is evidence produced of migmatite complexes in lower crustal exposures in the Sierra (Zeng et al. 2005). In addition, the peraluminous plutons of the Peninsular Ranges contain abundant, Proterozoic zircon cores and crystals (Shaw et al. 2003), likely because melts from which they crystallized were saturated in zirconium and unable to dissolve grains inherited from their metasedimentary sources (e.g., Miller et al. 2003). The Dinkey Dome granite contains few inherited cores (Fig. 4). It follows is that the Dinkey Dome magmas were potentially derived from hotter or inheritance-poor sources (Miller et al. 2003) and presumably are farther separated from said source(s) that might be analogous to those in migmatite complexes in the southern Sierra (Zeng et al. 2005).

Overall, a theme emerges from the Mesozoic arc segments in California, wherein small volume peraluminous melts are the primary archive of melts of crustal "character," but that these melts are restricted to earlier magmatism. It follows that because these antecedent magmas form in early stages of magmatism, as heat content in the arc is ramping, they may have sustained mobility, which allows them to preserve more chemical heterogenity. The increased vigor of arc magmatism also means that such crustal melt expression is muted in younger arc plutons. These later stages are typified by periods of high-flux magmatism from more organized sources that efficiently homogenize magmas with greater proportions of mantle melts (Lackey et al. 2012).

Conclusions and Implications

Contrasting records of isotopic heterogeneity in zircon and garnet crystals within different domains of the Dinkey Dome pluton exemplify how: 1) Sierran felsic granitoids originate though a variety of processes and are not restricted to end member models; and 2) early-stage plutons record crustal melting in the nascent stages of arc magmatism. The evolution of early-stage Sierran granitoids reflects intermediate processes whereby magmas became more felsic due to a combination of processes such as contamination, assimilation and fractional crystallization at different stages (Lackey et al., 2005, 2006, 2008; Nelson et al. 2013). The requirement for an enriched-mantle model whereby granites are derived solely by several stages of partial melting and fractionation of these enriched-mantle sources, with no crustal input (Coleman and Glazner 1997; Ratajeski et al. 2001, 2005), need not be so restrictive given the high oxygen isotopic values recorded in cores of zircon grains ($\delta^{18}O$ =7.7 to 7.8‰) in the Dinkey Dome pluton. Instead, before an arc organizes a Melting-Assimilation-Storage-and Homogenization (MASH) system, hot zone (Annen et al. 2006), or similar magmatic source region capable of homogenizing unusual melt compositions, peraluminous plutonic "harbingers" like Dinkey Dome sample the pre-batholithic crustal structure. This crustal melt sampling period is brief and rare, but may be echoed in later stages arc stages when tectonic adjustments to the arc reposition crustal rocks into sites of melting (e.g., DeCelles et al. 2009), but also created conditions that are favorable to preserve such melts in structurally and thermally isolated regions of arcs such as in shear zones and metamorphic wallrock complexes. As a case in point, the shut-down of the Sierran arc is accompanied by a structural disruption of the arc and introduction of fertile schists into the subduction channel that are expressed as late stage, small-volume, high $\delta^{18}O$ peraluminous melts (Chapman et al. 2013). Other intervals, such as during a major re-organization within the arc at 105 Ma saw an increase in $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ consistent with a crustal melting episode (Holland et al. 2013). Thus, peraluminous plutons in arcs, though small and temporally restricted, record important aspects of the greater tectono-magmatic feedback systems and can potentially be used to identify cryptic crustal end members that become greatly diluted in more vigorous stages of arc magmatism produce the typical voluminous granodiorite.

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List of figure captions

Figure 1. Generalized map of the Central Sierra Nevada region, showing the location of peraluminous plutons and the study area, Dinkey Dome granite. Initial $^{87}Sr/^{86}Sr$=0.706 and PA/NA= Panthalassan/North American Break from Kistler (1990). The high $\delta^{18}O$(Zrn) (6.5-7.5‰) and low $\delta^{18}O$(Zrn) (5.5-6.5‰) domains of the central Sierra Nevada intrusives are modified from Lackey et al. (2006, 2008).

Figure 2. Geologic map of the Dinkey Dome with sample sites from this study and Lackey et al. (2006). Map after Bateman and Wones (1972).

Figure 3. Garnet and associated minerals in the Dinkey Dome granite. Sample 10DD02a. Images a and b were taken under transmitted light (PPL= Plain polarized light, XPL= Cross polarized light). Image c is a backscattered electron image (BSE) and shows quartz, muscovite, apatite, ilmenite and zircon, included in a typical garnet from the Dinkey Dome pluton. Values of $\delta^{18}O$ for this garnet are shown in Fig. 7.

Figure 4. Cathodoluminescence (CL) images of zircon grains from grain mounts 1S51-1S82 (Transect A-A' Fig. 2). Circles represent SIMS spots and numbers represent oxygen isotope ratios ($\delta^{18}O$). SIMS spots are 10 μm; scale bars are 50μm.

Figure 5. Cation composition of garnet phenocrysts from grain mounts and thin sections. (a) Almandine–spessartine–pyrope ternary plot includes garnet from both sides of the Dinkey Dome, showing that east side garnet is closer to the almandine-spessartine binary than garnet from the western half. (b) Representative rim-to-rim zoning profiles through the cores of garnets from samples 10DD06b-05 (west) and 10DD19c-02 (east). Background information for data presented in this figure can be found in supplementary data table B.

Figure 6. Oxygen isotope ratios in zircon (a) and garnet (b), and (c) Δ18O (Grt_{Ave}-Zrc_{Ave}) from a traverse of the Dinkey Dome pluton (A-A' in Fig. 2) measured by ion microprobe (this study) and laser fluorination data of Lackey et al. (2006). Data represents zircon and garnet from grain mounts (1S51-1S82) and thin sections (10DD-02-10DD-19). Samples on x-axis are spaced according to relative distance of their field localities. See figure 2 for sample localities. Background information for data presented in this figure can be found in supplementary data tables A, C-D.

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Figure 7. Example of a rim-rim traverse of oxygen analyses for a single garnet (10DD-02a-13, BSE image in thin section, see Fig. 3) measured by SIMS. Note the image is rotated ~90° from figure 3.

Figure 8. Chondrite normalized REE patterns for zircons from four samples measured by SIMS (a. 1S82, b. 1S53, c. 1S79, d. 1S77) from the Dinkey Dome (1S82 and 1S79 from the west side, 1S53 and 1S77 from the east side). Filled symbols are from cores and open symbols are rims. Background information for data presented in this figure can be found in supplementary data Table F.

Figure 9. (a) $(Sm/La)_N$ vs. La (ppm) in rims and cores of zircon in the Dinkey Dome, from the same four samples from figure 8 (1S82, 1S53, 1S79, 1S77). (b) Ce/Ce* vs. $(Sm/La)_N$ in rims and cores from zircon in the Dinkey Dome, from the same four samples as in (a). Magmatic and hydrothermal fields from Hoskin (2005) and Grimes et al. (2007). Background information for data presented in this figure can be found in supplementary data Table E.

Figure 10. Evolution of $\delta^{18}O$ of zircon and garnet in the Dinkey Dome pluton recording assimilation and fractional crystallization through time for the (a) western and (b) eastern sides of the pluton.

Figure 11. Model of the genesis of the Dinkey Dome pluton from the earliest stage (I) increments of melt transiting wallrocks of different $\delta^{18}O$ in the presence of early stage zircon, to (IV) final stage crystallization of garnet in the composite pluton.

Table 1. Oxygen isotope data summary.

Sample	Lithology	δ^{18}O WR ‰ VSMOW (laser)	δ^{18}O Zrn ‰ VSMOW (laser)	δ^{18}O Zrn ‰ VSMOW (SIMS)	δ^{18}O Grt ‰ VSMOW (laser)	δ^{18}O Grt_{Ave} ‰ VSMOW (SIMS)	Δ^{18}O (Grt_{Ave}-Zrc_{Ave})
1S51	granite	9.71*	7.76*	C=7.6, R=7.3	6.9	6.2	-1.4
1S52	granite	9.80*	7.51*	C=7.8, R=6.7	7.4	5.9	-1.3
1S53	granite	9.57*	7.53*	C=7.9, R=6.4	6.9	5.9	-1.4
1S54	granite	9.90*	7.81*	C=8.0, R=6.8		6.4	-1.0
1S58	granite	9.90*	7.77*	C=7.5, R=7.0		7.4	0.1
1S77	granite	9.81*	7.63*	C=7.6, R=6.5	6.9	7.0	-0.4
1S79	granite	9.96*	7.67*	C=7.6, R=6.8	7.7	7.5	0.2
1S80	granite	10.30*	7.72*	C=8.1, R=6.3	7.9	7.3	0.1
1S81	granite	9.79*	7.76*	C=7.7, R=7.1	7.2	7.4	0.0
1S82	granite	9.79*	7.73*	C=7.8, R=7.3	8.0	7.6	-0.1
10DD-02	granite	10.47		7.9**		7.1	-0.8
10DD-05	granite	10.14		7.4**		6.7	-0.7
10DD-06	granite	9.65					
10DD-07	granite	9.41					
10DD-08	granite	9.48					
10DD-10	bt hornfel	12.72					
10DD-15	quartzite	11.82					
10DD-16	granite	9.35		7.6**		5.7	-1.9
10DD-17	granite	9.12		6.9**		5.1	-2.0
10DD-18	granite	8.96					
10DD-19	granite	9.06		7.0**		5.0	-2.0
10DD-20	enclave	8.25					
10DD-21	granite	9.17					
10DD-22	granite	9.06					

*Lackey et al. (2006) oxygen isotope analyses by Laser Fluorination.
**SIMS analyses on grains too small to distinguish core vs. rim.
C=Core, R=Rim; Zrn=Zircon, Grt=Garnet, WR=Whole Rock.
Samples 1S51 - 1S82 are grain mounts; 10DD-02 - 10DD-22 are thin sections.

Supplementary Data Tables:

Supplementary Table A: Sample locations and δ^{18}O summary.

Supplementary Table B: EPMA data.

Supplementary Table C: Zircon SIMS δ^{18}O data.

Supplementary Table D: Garnet SIMS δ^{18}O data.

Supplementary Table E: Garnet Laser Fluorination δ^{18}O data.

Supplementary Table F: Zircon REE data.

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Fig. 1

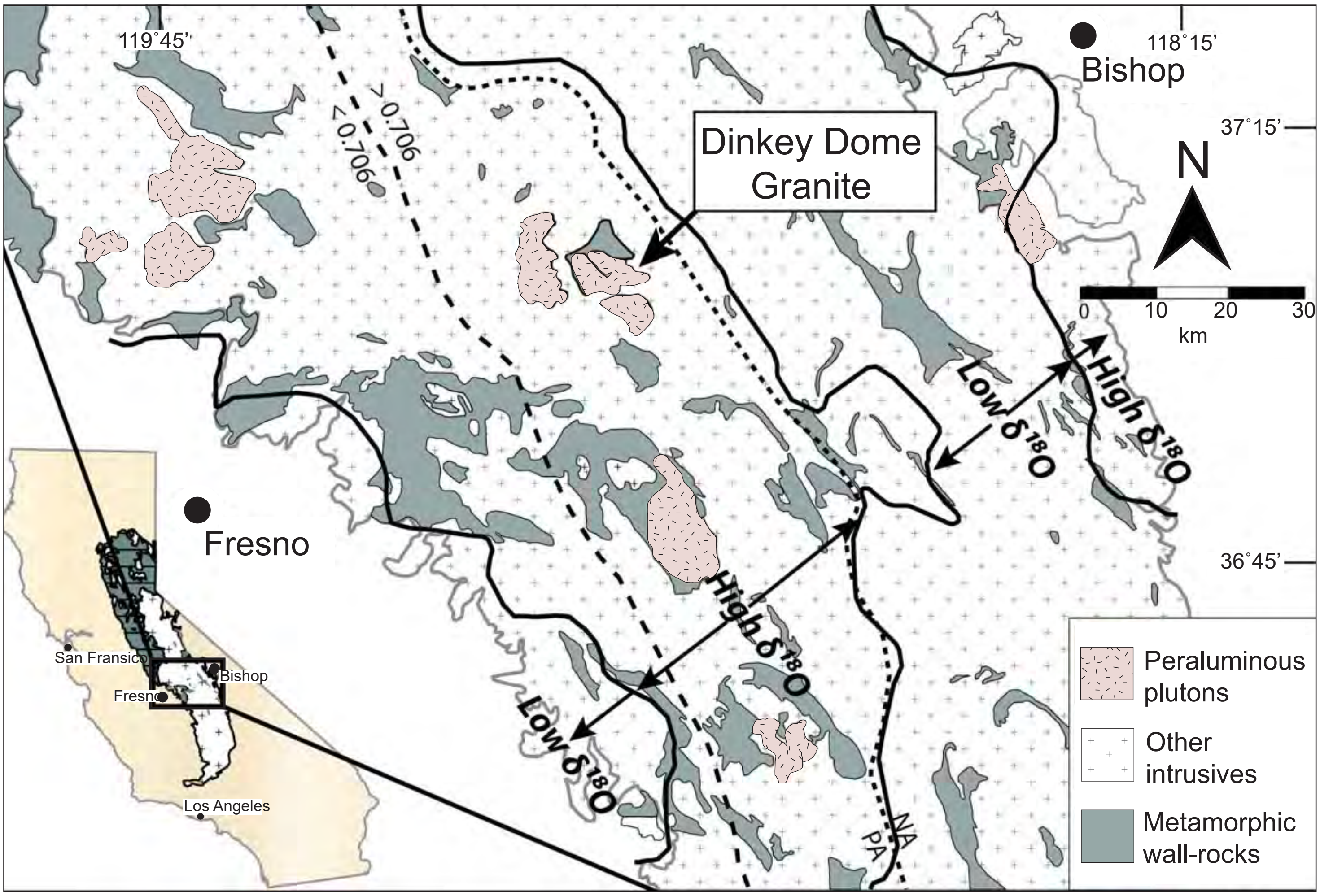

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Fig. 2

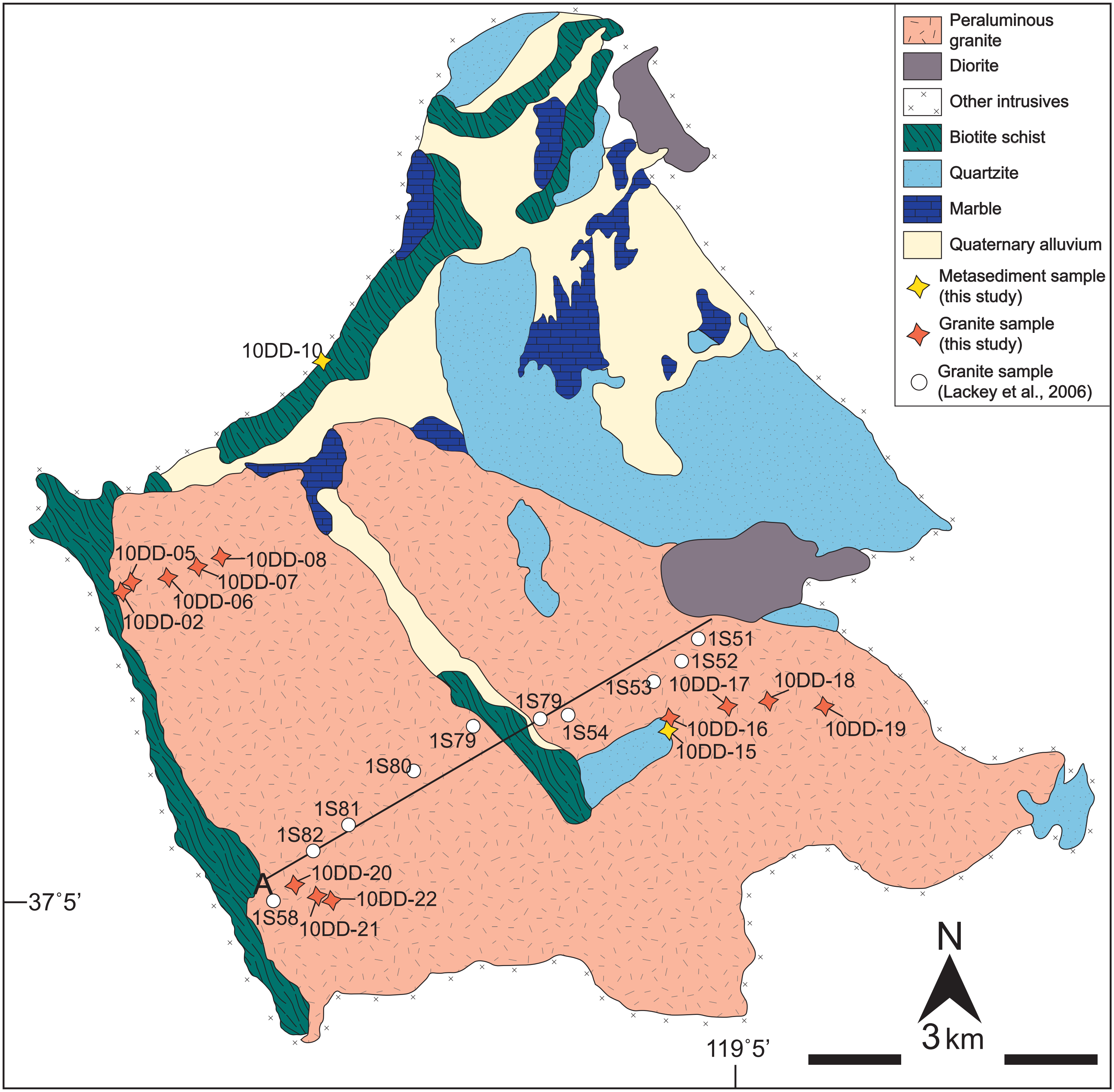
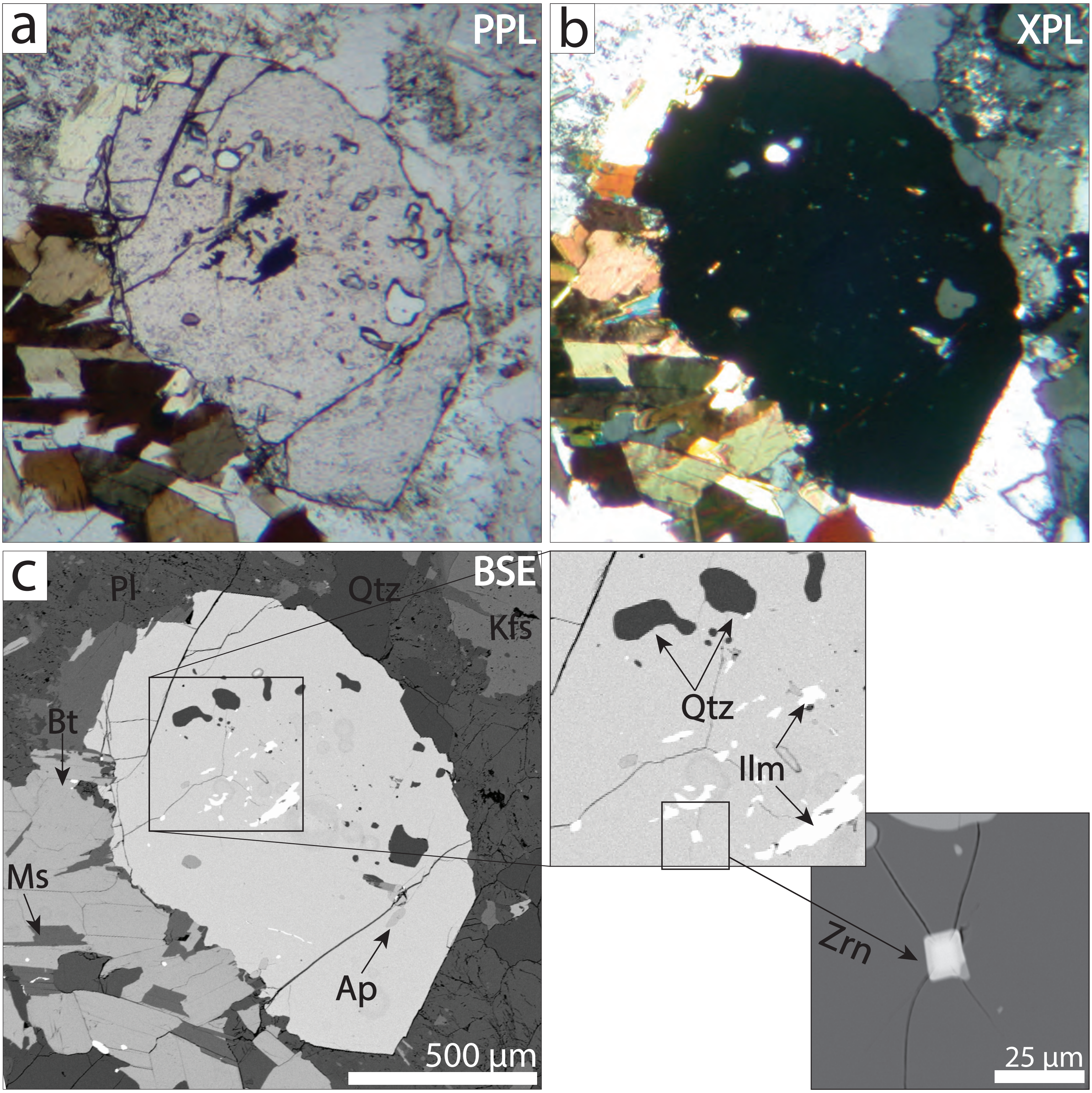


Fig. 3



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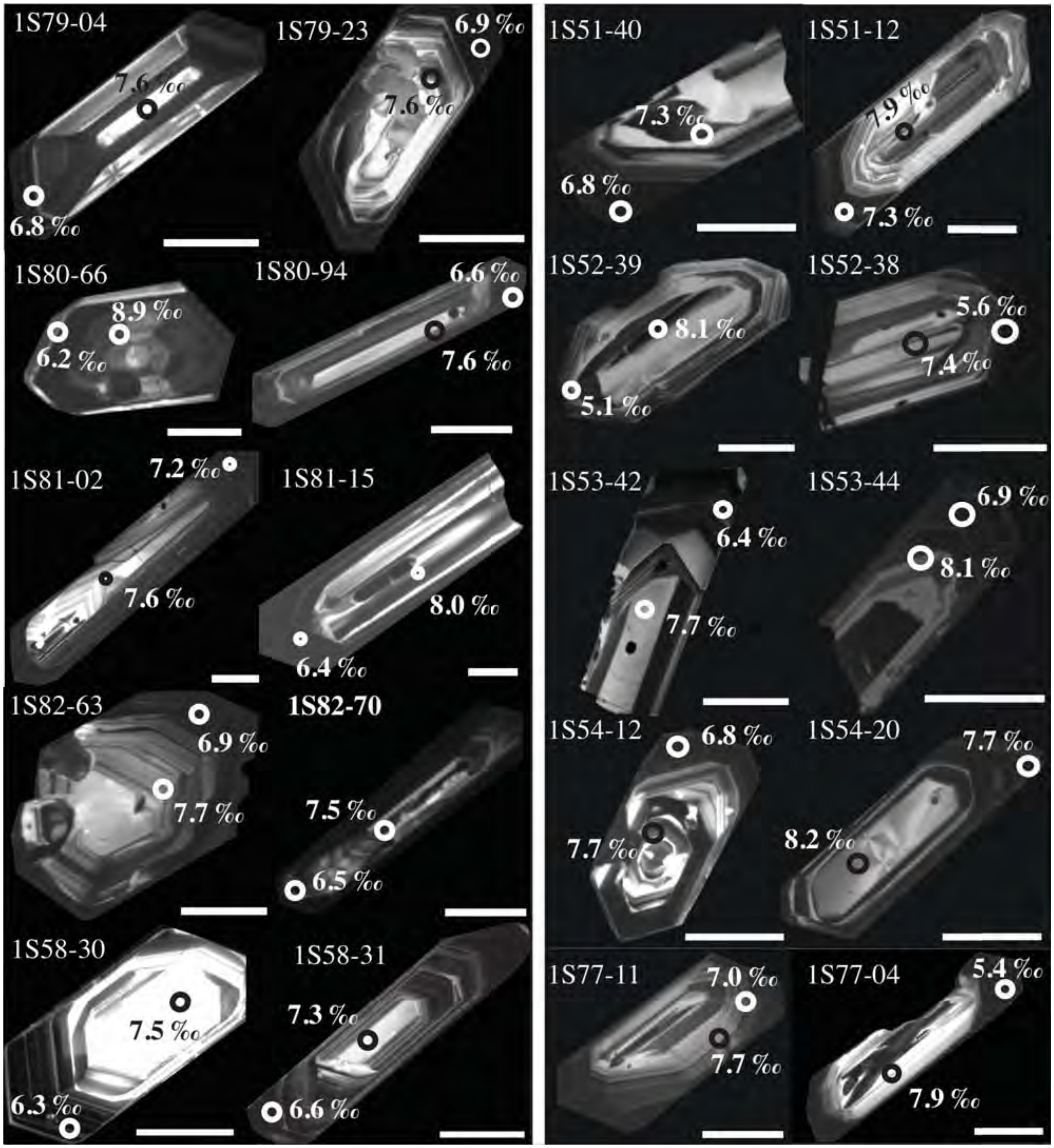
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1S80-94
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1S52-38
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Fig. 5

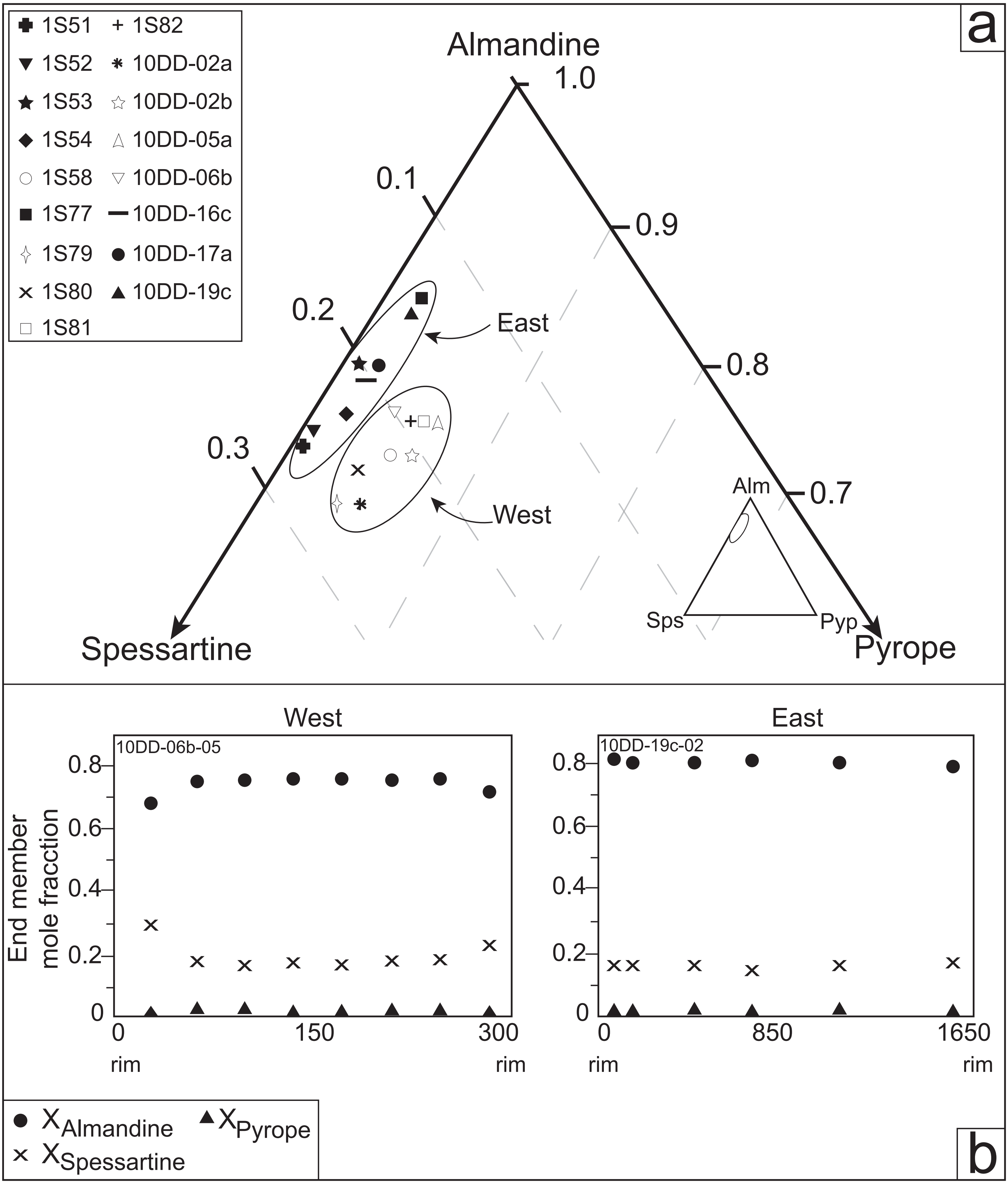
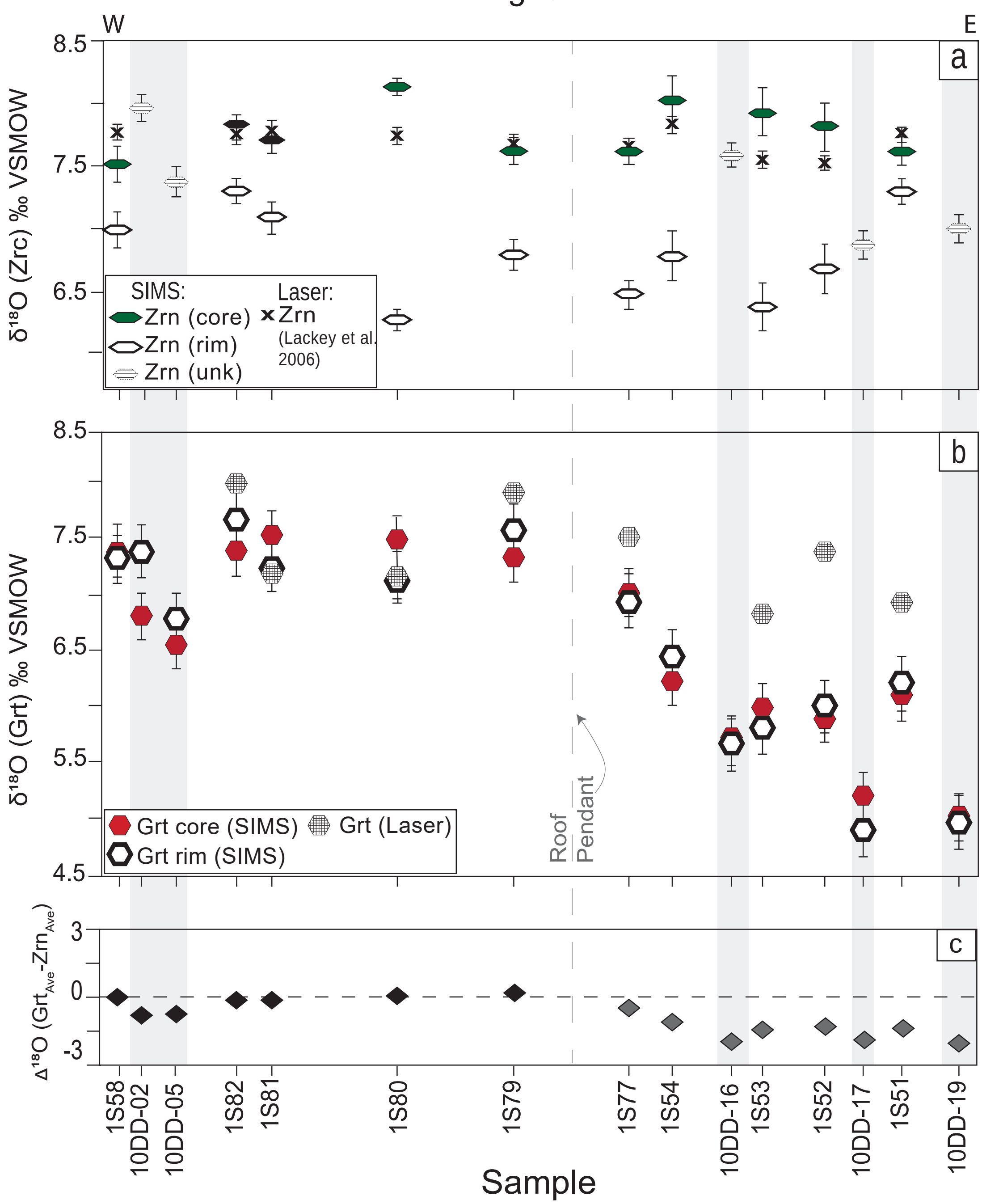


Fig. 6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δ¹⁸O (Zrc) ‰ VSMOW
δ¹⁸O (Grt) ‰ VSMOW
Δ¹⁸O (Grt_Ave-Zrn_Ave)
SIMS:
Zrn (core)
Zrn (rim)
Zrn (unk)
Laser:
Zrn
(Lackey et al.
2006)
Grt core (SIMS)
Grt (Laser)
Grt rim (SIMS)
Roof
Pendant
1S58
10DD-02
10DD-05
1S82
1S81
1S80
1S79
1S77
1S54
10DD-16
1S53
1S52
10DD-17
1S51
10DD-19
Sample

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Fig. 7

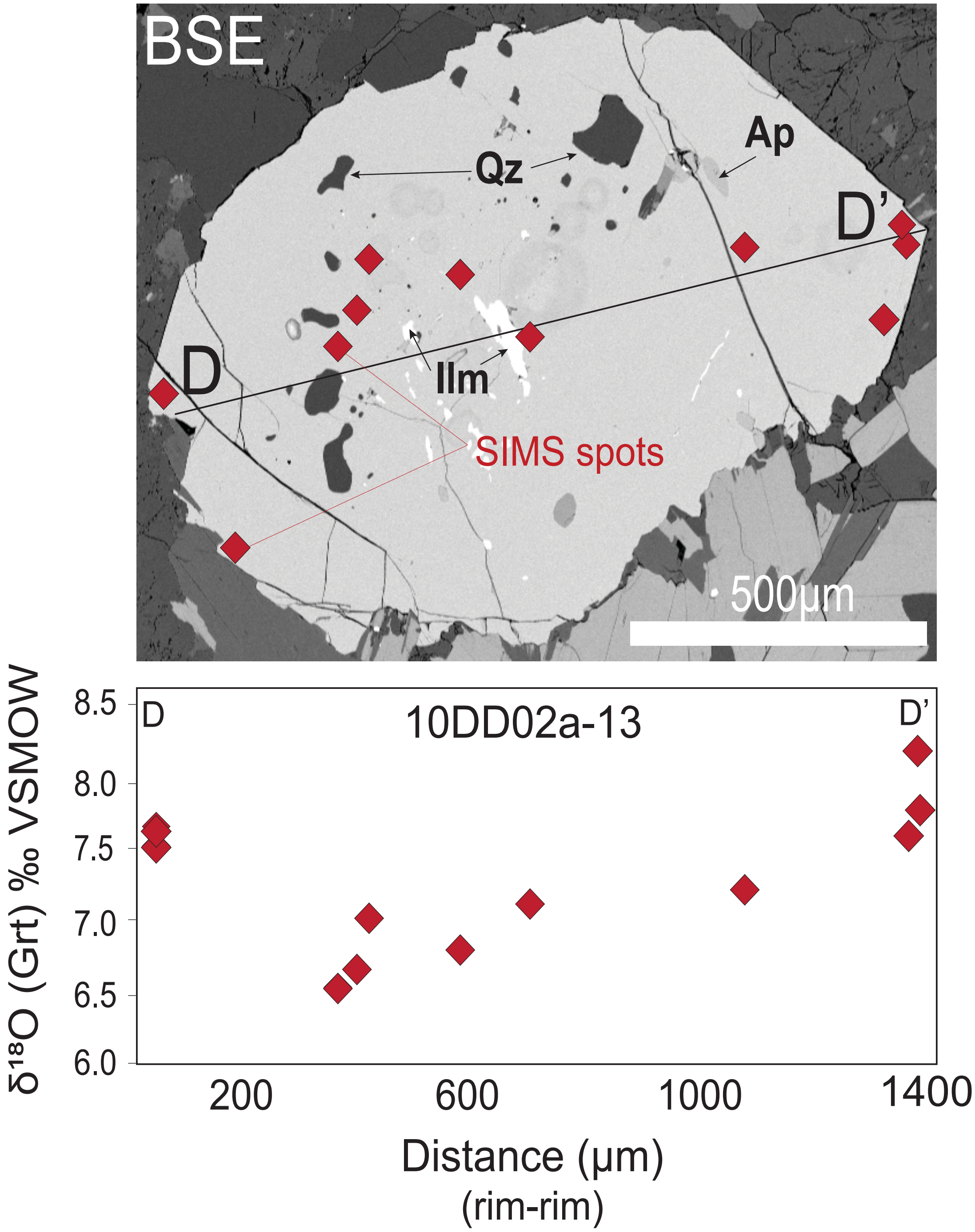

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Fig. 8

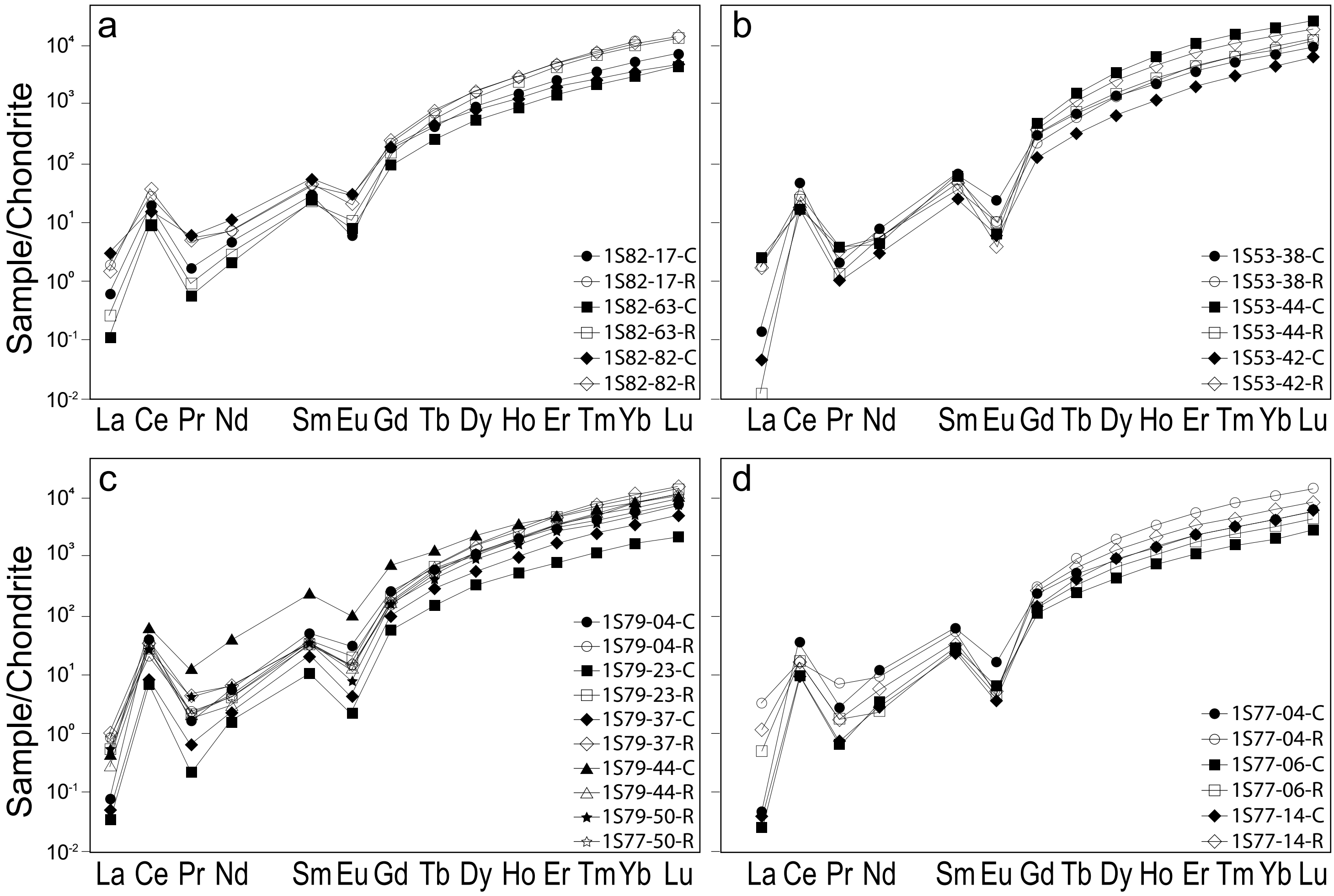

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Fig. 9

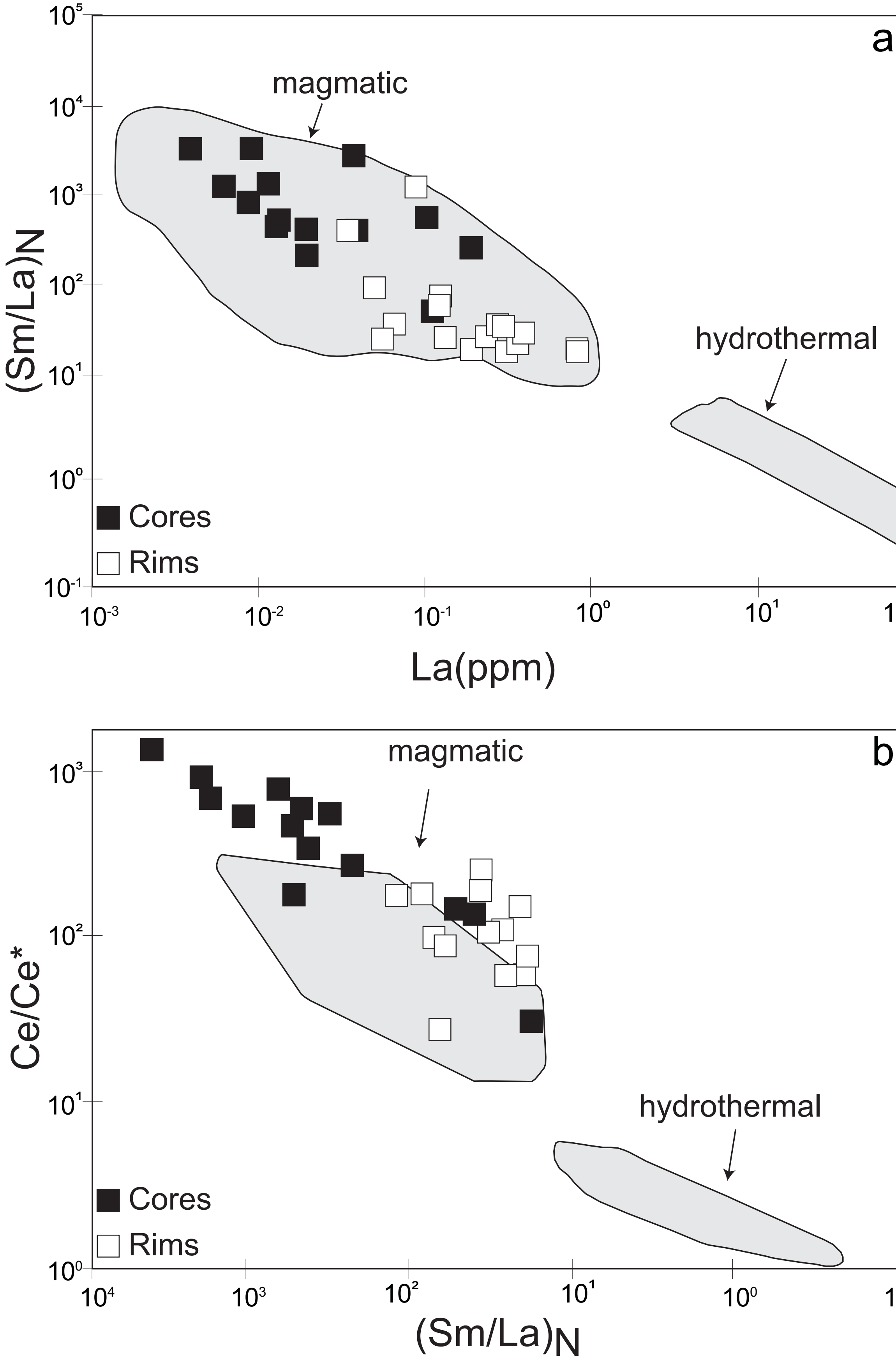

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Fig. 10

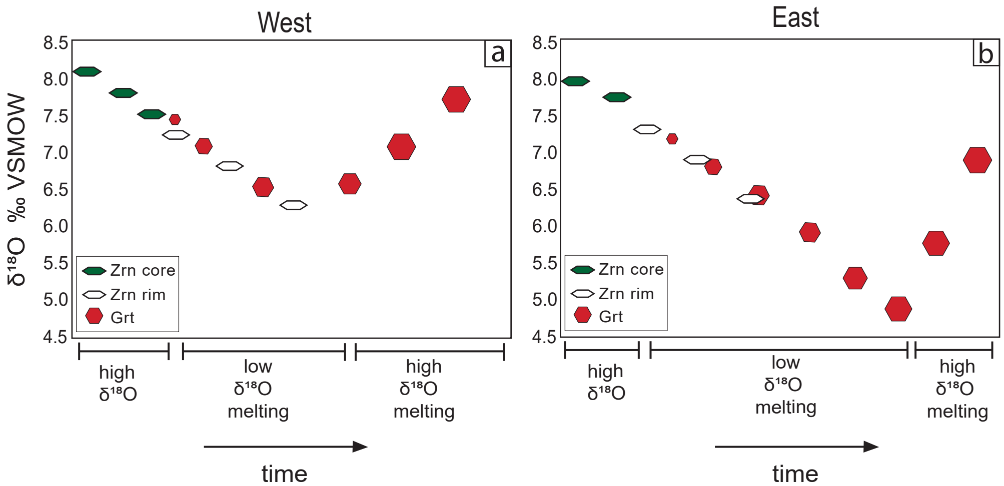

Fig. 11
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