

Linking Zircon Crystallization to Magmatic Processes in Basalt-Dominated Lava Fields: The Eastern Snake River Plain–Craters of the Moon Testbed

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Throughout the Quaternary, the Eastern Snake River Plain (ESRP) has been the focal zone of bimodal volcanism which filled an elongated downwarp within the northwestern North American continental lithosphere. In the wake of the southwest–northeast trending Yellowstone hotspot track and its silicic caldera volcanism, massive outpourings of olivine tholeiite lava occurred within the ESRP, but several prominent rhyolite domes and intermediate lava flows also erupted in its center. Along its northern margin, the Craters of the Moon volcanic field (COM) is also dominated by Middle Pleistocene–Holocene mafic lavas, but they comprise comparatively Fe-rich and Mg-poor trachybasalts and three evolved basaltic trachyandesite to trachydacite flows emplaced during its youngest eruptive episode. The occurrence of zircon crystals in these young lavas and rhyolite domes provides a unique opportunity to evaluate zircon crystallization in well-preserved and accessible basalt-dominated lava fields as analogs to large igneous provinces in the geologic past. Zircon crystallization ages determined by U–Pb and U–Th dating methods were paired with crystal-scale trace element and isotopic ($\delta^{18}\text{O}$, ϵ_{Hf}) data. Additionally, the eruption age of the evolved COM Highway flow was determined by combined (U–Th)/He and U–Th dating. U–Pb data indicate crystallization ages between 1540 ± 10 ka in the oldest and 335 ± 3 ka (2se) in the youngest ESRP rhyolite domes which are consistent with reported K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ eruption ages (1400 ± 28 – 327 ± 3.2 ka). Similarly, U–Th ages averaging $6.2^{+2.6}_{-2.5}$ ka compared to (U–Th)/He ages of 2.05 ± 0.56 ka (2se) for evolved COM lava also indicate crystallization with preeruptive residence over several millennia. Values for $\delta^{18}\text{O}$ (ESRP: $+1.89$ to $+4.32$; COM: $+5.13$ to $+6.03$) and ϵ_{Hf} (ESRP: -6.4 to -1.6 ; COM: -11.9 to -8.9) suggest that ESRP and COM magmas originated from a similar mantle source, but evolved via divergent assimilation and fractional crystallization paths. These paths for ESRP and COM magmas depend on the abundance of hydrothermally altered felsic rocks in the crust, which strongly influence ESRP magmas on the axis, but less so the COM magmas at the margins of the downwarp. Zircon crystallization postdates assimilation and is facilitated by progressive Zr enrichment due to extensive fractional crystallization. Parallel whole rock and zircon trace element (e.g. Zr/Hf, Eu/Eu*) fractional crystallization trends at nearly identical degrees of remaining liquid support autocrystic zircon crystallization and argue against crystal recycling from largely solidified crystal mushes. In a global comparison to volcanic rocks from large igneous provinces, such high-Zr mafic-intermediate lavas are scarce. Thus, autocrystic zircon in mafic lavas are clearly exceptional and should be scrutinized against potential contamination using trace element or isotopic data. Unless Zr abundances are exceptionally high, zircon saturation in mafic magmas only appears achievable in interstitial melt pockets in slowly crystallizing sills or intrusions.

Key words: Craters of the Moon; Yellowstone; O–Hf isotopes; AFC modelling; radiometric dating

INTRODUCTION

Zircon has become an indispensable tool to reconstruct thermochemical conditions of magmas, providing unique insights into the processes of magma origins, storage, and solidification, especially for felsic magmas prone to zircon saturation. Even in basalt-dominated environments such as mid-ocean ridges and oceanic islands zircon is present either in gabbroic to plagiogranitic rocks or rare evolved lavas (Grimes *et al.*, 2007, 2011; Schmitt *et al.*, 2011; Carley *et al.*, 2014). However, the occasional presence of zircon in rapidly cooled basaltic lavas, remains enigmatic, especially as such lavas are widespread surface constituents of Earth,

Moon, and the terrestrial planets. Explanations range from in situ crystallization to extraneous origins through natural or laboratory contamination (e.g. Bea *et al.*, 2022; Bjerga *et al.*, 2022; Rojas-Agramonte *et al.*, 2022). The search for datable zircon in mafic rocks is relevant for oceanic magmatism, continental rifts, but in particular for large igneous provinces (LIPs), which are characterized by the emplacement of large volumes of predominantly mafic magma over short durations in either continental or oceanic settings with pervasive and large-scale impacts on regional to global environments (e.g. Ernst, 2014). Reconstructing these impacts from LIP events throughout geological time requires reliable geochronological constraints. In some cases, this

has been successfully achieved by dating zircon crystals in intercalated evolved rocks (e.g. Schoene *et al.*, 2015; O'Connor *et al.*, 2022), or in slowly cooled sills or dikes (e.g. Gaynor *et al.*, 2022), although these are typically rare or poorly exposed. Where zircon has been reported from LIP mafic volcanic rocks and their metamorphic equivalents, its provenance often remains controversial. Besides crystallization in evolved melts, either in isolated pockets or in larger reservoirs, other interpretations include autocrystic crystallization (Schulz *et al.*, 2006; Zhu *et al.*, 2016; Hartmann *et al.*, 2019), sediment contamination in the crust (e.g. Shumlyanskyy *et al.*, 2016) or the mantle source (Xu *et al.*, 2018), or local contamination followed by zircon resorption and regrowth (Gaynor *et al.*, 2022, 2023; Davies *et al.*, 2021). Baddeleyite, which can crystallize in silica-undersaturated melts typical for LIPs, has been harnessed as a viable dating alternative (Davies *et al.*, 2017; Greber *et al.*, 2020), but often baddeleyite crystals are small, its age accuracy can be compromised by Pb loss and, in comparison to zircon, its petrochronological significance is much less established (Schaltegger & Davies, 2017).

Zircon saturation experiments (e.g. Watson & Harrison, 1983; Watson, 1996; Gervasoni *et al.*, 2016; Zhang & Xu, 2016; Borisov & Aranovich, 2019; Shao *et al.*, 2019; Crisp & Berry, 2022) predict crystallization in evolved melts under conditions met in slowly crystallizing basaltic magmas forming intrusive rocks (e.g. Hodych *et al.*, 2004; Bea *et al.*, 2006; Grimes *et al.*, 2011; Zheng *et al.*, 2011; Shumlyanskyy *et al.*, 2016). These studies also invoke rapid dissolution of xenocrystic zircon in contact with hot, mafic melts (within <1 day; Borisova *et al.*, 2020; cf. Xu *et al.*, 2018). Some studies emphasized local saturation in restricted highly evolved melt pockets where zircon precipitation is feasible (e.g. Bea *et al.*, 2022). Yet, reliable criteria for distinguishing between xenocrystic and autocrystic zircon in mafic hosts are lacking, which limits the petrochronological potential of zircon in such rocks.

To investigate zircon crystallization in the context of mafic magmatism, we turned to the Eastern Snake River Plain (ESRP) as a young analog for ancient LIPs (Fig. 1). Because of its Quaternary age, the ESRP allows reconstruction of zircon crystallization at high temporal fidelity. Although not a LIP per se, the ESRP is in direct lineage with the Columbia River flood basalt event. In the ESRP, southwest–northeast migration of the Yellowstone hotspot track has triggered massive outpourings of basaltic lavas that erupted from large shield volcanoes in the wake of preceding silicic caldera volcanism (Pierce & Morgan, 1992; Smith & Braile, 1994). Basaltic volcanism is predominant in the ESRP, which during the Holocene alone witnessed formation of at least eight lava fields that cover ~13% of its surface area. Among these volcanic fields, Craters of the Moon (COM) is the largest, which, in conjunction with the Kings Bowl and Wapi, has produced more than 60 eruptions over the past 15 000 years (Kuntz *et al.*, 1986; Kuntz, 1992; Kuntz *et al.*, 2007). Additionally, in the middle ESRP, the basaltic strata were penetrated and intruded by several Pleistocene evolved lavas and domes. Although some of these domes form prominent landmarks in the ESRP, their collective volume is only a small fraction of the total mafic output (McCurry *et al.*, 2008).

To cover the full spectrum of zircon in intermediate–silicic rocks in a predominantly mafic volcanic field, this study reports U–Th–Pb ages, O and Hf isotopes, and trace element compositions for zircon from all exposed ESRP domes. Similarly, zircon was investigated from three of the most recent COM trachyandesite lavas flows, and from a xenolith collected in the area as a potential crustal contaminant. Zircon crystallization ages in combination with new (U–Th)/He as well as published K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ erup-

tion ages underscore autocrystic zircon formation shortly before the eruption, whereas isotopic and trace element petrochronology links the zircon cargo in the ESRP and COM lavas to a liquid line of descent controlled by extensive fractional crystallization which postdates crustal assimilation.

Initiation and evolution of ESRP-COM magmatism

The ESRP is a topographic depression of 400-km length and 50- to 100-km width located in southern Idaho, USA (Fig. 1). In conjunction with the Columbia River Basalt province and caldera-related deposits of the Yellowstone hotspot track, this system represents one of the world's premiere examples of a large-volume continental igneous province (e.g. Hildreth *et al.*, 1991; Christiansen, 2001; Nash *et al.*, 2006; Pierce & Morgan, 2009; Coble & Mahood, 2012; Colón *et al.*, 2018). Volcanism in the area is extensive in both space and time (Fig. 1a): it started with the Miocene McDermitt caldera overlapping northern Nevada and southeastern Oregon (c. 16.4 Ma) and then progressively moved eastward forming in sequence the volcanic fields of Juniper Mountain/Owyhee Humboldt (c. 14–13 Ma), Bruneau–Jarbridge (c. 12.7–10.5 Ma), and Twin Falls (c. 11–6 Ma), followed by composite calderas of Picabo (c. 10–6 Ma), Heise (c. 8–4 Ma), and ultimately Yellowstone as the current volcanic focus (Kuntz, 1992; Stout *et al.*, 1994; Morgan & McIntosh, 2005; Shervais *et al.*, 2006; Shervais & Hanan, 2008; Watts *et al.*, 2011; Henry *et al.*, 2017; Colón *et al.*, 2018; Anders *et al.*, 2019). Subsequently and overlying the older Yellowstone-type rhyolitic ignimbrites and lava flows, the Pliocene–Quaternary ESRP (Fig. 1b) is the site of extensive, predominantly olivine tholeiitic volcanism covering a large swath of southern Idaho (e.g. Leeman, 1982a; Kuntz, 1992; Reid, 1995; Hanan *et al.*, 1997; Geist *et al.*, 2002; Hughes *et al.*, 2002a; Hughes *et al.*, 2002b; Shervais *et al.*, 2006). The youngest center in the ESRP is the Craters of the Moon volcanic field (COM) with morphologically extremely well-preserved volcanic shields, cinder and spatter cones. COM lavas were emplaced at the northern end of the Great Rift (Fig. 1c), a 2- to 8-km wide and up to 85-km-long belt of dike-fed extensional fissures that follow the same NW–SE orientation as regional normal faults of the Basin and Range Province (Leeman *et al.*, 1976; Kuntz *et al.*, 1986; Kuntz, 1992; Chadwick *et al.*, 2007). The COM covers ~1600 km², has a total volume of about 30 km³, and is considered the largest Holocene lava field in the USA (Kuntz, 1992; Stout *et al.*, 1994).

The crustal underpinnings of Yellowstone hotspot and subsequent ESRP volcanism (Fig. 1) include Archean rocks that belong to the western margin of the Wyoming Province (e.g. Chamberlain *et al.*, 2003) as demonstrated by crustal xenolith populations and chemical contamination signatures in the volcanic rocks (Leeman, 1982b; Menzies *et al.*, 1984; Kuntz, 1992). The Cretaceous Idaho Batholith underlies the westernmost parts of the Snake River Plain (e.g. Hyndman, 1983; Gaschnig *et al.*, 2011). In addition, the regional upper crust comprises Paleozoic–Mesozoic sedimentary strata, as well as shallow intrusions and volcanic deposits of Eocene age that are subsumed under the Challis magmatic episode (Fisher & Johnson, 1995a; Fisher & Johnson, 1995b). Eocene–Miocene to recent Basin and Range block faulting subsequently extended and dissected the crust, although faults are largely buried by younger volcanic rocks in the ESRP (e.g. Faulds *et al.*, 1998). Drill cores in the center of the ESRP (e.g. INEL-1; Fig. 1) and O-isotopic compositions of caldera and post-caldera units also indicate pervasive hydrothermal alteration of intra-caldera fill for the Miocene–Pleistocene volcanism (Borroughs *et al.*, 2012; Drew *et al.*, 2013; Ellis *et al.*, 2013). Remelting and/or assimilation of

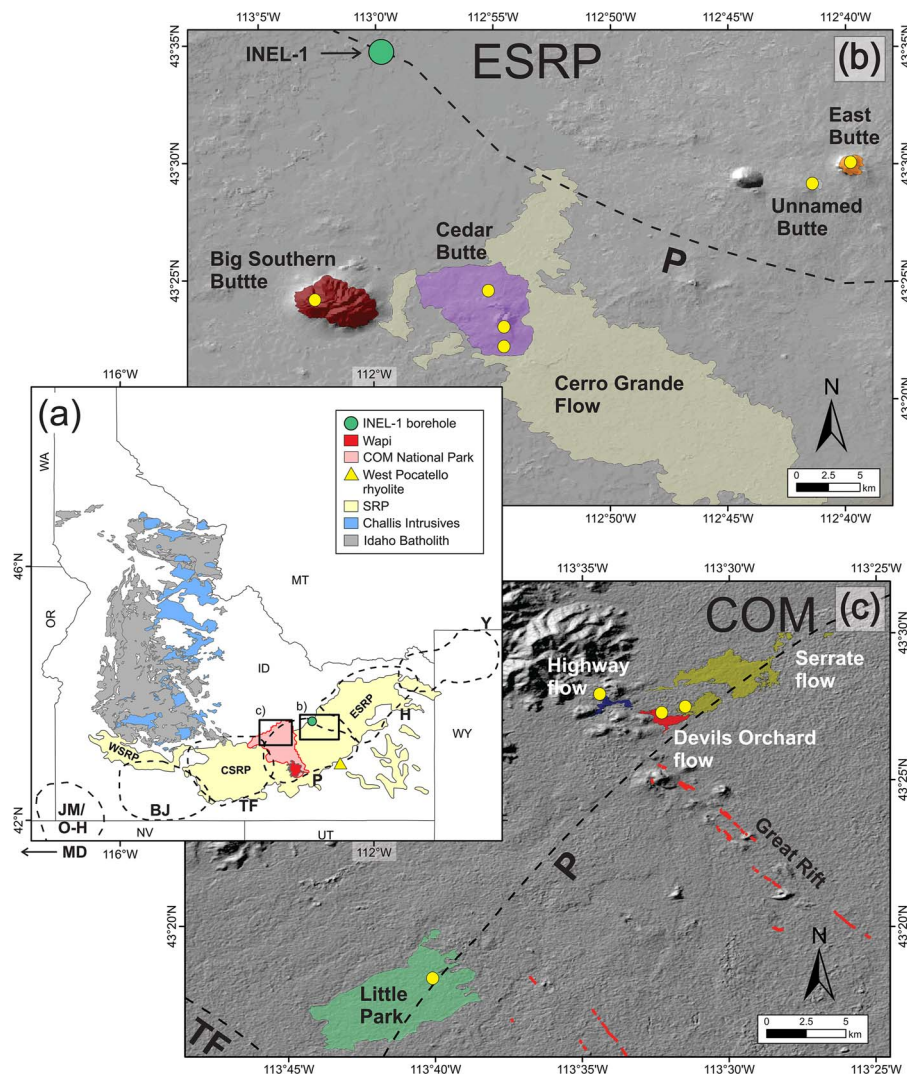


Fig. 1. Snake River Plain overview map with subdivisions of western, central, and eastern domains (WSRP, CSRP, and ESRP, respectively) as well as the Craters of the Moon (COM) volcanic field (delineated by the national monument boundaries) and additional geological features mentioned in the text (a). Dashed lines encompass time transgressive caldera fields of the Mid-Miocene to Quaternary Yellowstone hotspot track (MD = McDermitt, JM/O-H = Juniper Mountain/Owyhee Humboldt, BJ = Bruneau-Jarbridge, TW = Twin Falls, P = Picabo, H = Heise, and Y = Yellowstone) after [McCurry & Rodgers \(2009\)](#). Extent of Idaho Batholith and Challis Intrusives shown after [Gaschnig et al. \(2011\)](#). Sample locations for (b) ESRP Quaternary rhyolite domes and (c) COM basaltic trachyandesite intermediate flows. The Pleistocene-Holocene Cerro Grande basalt flow is plotted for comparison, and the basaltic Little Park lava field indicates sampling location for a COM xenolith. Caldera boundaries are shown as dashed lines (see panel a) along with other geological features mentioned in the text. Digital elevation model uses ASTER GDEM data (<https://gdemdl.aster.jspacesystems.or.jp>) and the geological units are based on [Kuntz et al. \(1994, 2007\)](#).

such rocks produces a characteristic low- $\delta^{18}\text{O}$ magma signature for many of the rhyolites (e.g. [Drew et al., 2013](#); [Bindeman & Simakin, 2014](#); [Colón et al., 2018](#)).

ESRP-COM magma evolution and the presence of zircon

Pliocene-Holocene post-hotspot volcanism in the ESRP is dominantly mafic, with lava flow outpourings that formed large shield volcanoes collectively reaching thicknesses of 1.7 to 3.1 km ([Anders & Sleep, 1992](#); [Whitehead, 1992](#); [Rodgers et al., 2002](#); [Chadwick et al., 2023](#)). Near the axis of the ESRP, evolved volcanism produced several endogenous domes (Unnamed Butte, East Butte, Cedar Butte, Big Southern Butte; [Fig. 1a](#)) and cryptodomes that created surface uplift but remained concealed in the subsurface (e.g. Middle Butte, Buckskin Dome, and Ferry Butte) between c. 1.4 and 0.3 Ma without any apparent time-transgressive relation ([Kuntz & Dalrymple, 1979](#); [Spear, 1979](#); [Spear & King, 1982](#);

[McCurry et al., 1999](#); [McCurry et al., 2008](#)). Rhyolite predominates in these domes, except for Cedar Butte flows, drilled samples from Unnamed Butte, and enclaves from East Butte which are intermediate in composition ([McCurry et al., 2008](#)). Although individual domes stand out prominently by up to ~700 m from the ESRP lava plains and reach impressive volumes of ~10 km³ in the case of Big Southern Butte, the collective volume of evolved lavas and domes of ~105 km³ is subordinate to the estimated 12 300 km³ of mafic lava flows ([McCurry et al., 2008](#)). Zircon in ESRP rhyolites has been petrographically identified as an accessory mineral (e.g. [McCurry et al., 1999](#); [Ganske, 2006](#)); its presence in ESRP intermediate rocks, however, has been elusive. No detailed petrochronological characterization of ESRP zircon existed prior to this study. Whole rock Zr abundances record a shift from initially increasing toward decreasing abundances with increasing SiO₂, and thus zircon saturation has been indirectly inferred to begin at 58 to 61 wt.% SiO₂ ([McCurry et al., 2008](#);

Fig. 2a). The exception are some intermediate compositions from a drillhole underneath Unnamed Butte in the ESRP, where a direct mixing trend between basaltic and rhyolitic endmembers without Zr enrichment is evident (Fig. 2a).

COM eruptive periods are classified according to their stratigraphic age ranging from 'H' at c. 15 ka to 'A' at c. 2.5 ka (Kuntz *et al.*, 2007). During the most recent period A, eight lava flows erupted from vents on the northern end of the Great Rift (Fig. 1b) mapped as the Broken Top, Blue Dragon, Trench Mortar Flat, North Crater, Big Craters, Serrate, Devils Orchard, and Highway flows (Stearns, 1928; Kuntz *et al.*, 1986; Kuntz *et al.*, 2007). Among these, Serrate, Devils Orchard, and Highway flows share the same appearance as jagged blocky and aa flows of evolved, trachyan-desitic composition. In contrast to the ESRP mafic lavas, which contain only rarely xenoliths, these are comparatively abundant in COM lavas. Xenoliths comprise two major types, felsic granulites and vesicular rhyolites (e.g. Leeman *et al.*, 1976; Stout *et al.*, 1994). The granulitic type is sourced in Archean basement (e.g. Menzies *et al.*, 1984; Leeman *et al.*, 1985), whereas the rhyolitic type has been assigned to Tertiary volcanic rocks present in the upper crust (Putirka *et al.*, 2009). Evolved COM lavas of period A also contain accessory zircon, which is absent in the more primitive basaltic lavas that erupted during the earlier periods (Kuntz *et al.*, 1986; Stout *et al.*, 1994). On one hand, zircon in COM lavas hosted in glomerocrysts has been interpreted as indicative for a xenocrystic origin (Stout *et al.*, 1994), but on the other, autocrystic zircon has been predicted from decreasing Zr abundances with increasing SiO₂ at ~59 wt.% (Leeman *et al.*, 1976; Fig. 2a).

MATERIALS AND METHODS

For the ESRP evolved centers, three samples from Cedar Butte, and one sample each from Unnamed Butte, East Butte, and Big Southern Butte were collected for this study (Table 1; Fig. 1b). For COM evolved lavas, one sample each from Devils Orchard and Highway flows, as well as two samples from Serrate flow were acquired within the Craters of the Moon National Monument (Fig. 1c; Table 1). Several cm-sized granulitic xenoliths were sampled from Serrate flow, but because of their small size, these were only used for petrographic inspection. A large (~12 cm in diameter) xenolith fragment from the Idaho State University collection was, therefore, included, which was derived from Little Park hawaiite flow on the south-southeast side of the COM (Fig. 1c).

Samples were crushed using a hammer or a hydraulic press, and the fine fraction was sieved to <250 µm. All powders were passed through hydrodynamic density separation yielding heavy minerals concentrates. For COM, the ilmenite-rich concentrates were further processed in a Frantz magnetic separator to enrich the nonmagnetic minerals including zircon. Zircon crystals were then hand-picked under a binocular microscope and mounted in epoxy together with zircon reference materials AS3 (Paces & Miller, 1993), 91 500 (Wiedenbeck *et al.*, 2004) and Temora 2 (Black *et al.*, 2004). Epoxy mounts were ground with silicon carbide abrasives to expose grain interiors and then polished using diamond paste (1 µm).

For scanning electron microscope (SEM) analysis using the Heidelberg University LEO 440, epoxy mounts were coated with a 20 nm Au layer. Images were generated using a backscattered electron (BSE) detector and a Gatan Mini-CL detector at beam conditions of 20 keV and 2.0–4.0 nA, and 10 keV and 15 nA, respectively, to visualize internal zircon textures. After that, the Au coat was gently removed, and mounts were ultrasonically

cleaned with aqueous decontaminants and rinsed multiple times with deionized water and methanol. They were then coated with a ~50 nm Au layer prior to secondary ionization mass spectrometry (SIMS) analysis.

U–Th–Pb geochronology and trace elements

For disequilibrium dating, zircon interiors were analyzed using the CAMECA ims 1300-HR³ SIMS at the John de Laeter Centre at Curtin University, Australia. A ¹⁶O[−] primary beam of ~40 nA in Gaussian mode was focused to an ~25 µm diameter spot, and secondary ions were extracted at 10 kV and steered through a 400 µm contrast aperture, a 4000-µm field aperture, and a 50 eV bandpass energy slit in circular mode. The secondary column was tuned to a mass resolution $m/\Delta m \sim 5000$ (250 µm multicollector slit width). Secondary ion intensities of ^{244.03}, ²³⁰Th¹⁶O, ²³²Th¹⁶O, and ²³⁸U¹⁶O were collected simultaneously in L2 and L1 electron multipliers (EMs) for the 244.03 background and ²³⁰Th¹⁶O peak, whereas Faraday cups with 10¹² Ω resistors were used in C and FC2 positions for the high-abundance peaks ²³²Th¹⁶O and ²³⁸U¹⁶O for 10 s per cycle. The static magnetic field was maintained using a nuclear magnetic resonance probe. ⁹⁰Zr⁹⁰Zr¹⁶O₄ and ⁹⁰Zr⁹²Zr¹⁶O₄ were measured by electrostatic peak switching using the dsp2-x deflector for 1 s per cycle, followed by another dsp2-x switch to measure off-peak intensities as background/baseline on 244.3, 246.3, 248.3, and 254.3 for 5 s per cycle. Each analysis started with a 30 s pre-sputtering during which the beam was rastered over a 10 × 10 µm area to remove surface contaminants, followed by automatic centering of the transfer deflectors. Secondary ion intensities were integrated over 20 cycles and corrected for dead-time and FC baseline using the in-house ZIPSv. 3.1.1. data reduction software. Backgrounds measured on 244.03 and 246.3 were averaged and subtracted from the ²³⁰Th¹⁶O peak. The U–Th relative sensitivity factor (RSF) was calibrated from linear regression in ²⁰⁸Pb/²⁰⁶Pb vs. ²³²ThO/²³⁸UO space (Reid *et al.*, 1997), and the accuracy of measured (²³⁰Th)/(²³⁸U) (activities indicated by parentheses) was confirmed by repeatedly analyzing secular equilibrium zircon reference AS3 (1099.1 Ma; Paces & Miller Jr., 1993), which yielded a weighted average (²³⁰Th)/(²³⁸U) = 1.008 ± 0.016 (2σ uncertainties throughout, unless otherwise indicated; mean square of weighted deviates MSWD = 1.09; n = 11). 91 500 reference zircon (81.2 µg/g U; Wiedenbeck *et al.*, 1995) was used to calibrate ²³⁸UO/⁹⁰Zr⁹⁰Zr¹⁶O₄ to infer U abundances. All analyses on unknowns where the signal of 246.03 was not at least 10% higher than the 244.03 background were omitted.

For U–Pb geochronology, crystals were analyzed using a 12–18 nA ¹⁶O[−] primary beam with a spot diameter of ~20 µm, with secondary ions being transmitted through contrast and field apertures of 400 and 3000 µm, respectively, and an energy bandpass of 50 eV in rectangular mode. For ²³²Th and ²³⁸U, a –20 eV offset was applied, but not for the other species (⁹⁴Zr₂O, ²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb, ²³⁸UO, and ²³⁸UO₂). Secondary ions were detected using the axial electron multiplier in peak hopping mode for 10 cycles after 20 s of pre-sputtering with a 10 × 10 µm raster. Replicate analyses of AS3 (COM), Temora2 (ESRP) and 91 500 (COM and ESRP) zircon references were used to establish U–Pb, U–Th, and U–Zr RSFs. For the U–Pb RSF, this was achieved by applying a linear regression in ²³⁸UO₂/²³⁸U vs. ²⁰⁶Pb/²³⁸U space; U–Th RSF values were determined from regression in ²⁰⁸Pb/²⁰⁶Pb vs. ²³²Th/²³⁸U, and for the U–Zr RSF values an 81.2 µg/g U abundance of 91 500 zircon along with stoichiometric Zr was used (Wiedenbeck *et al.*, 1995). Zircon reference 91 500 analyzed as a control yielded a ²⁰⁶Pb/²³⁸U age of 1066 ± 13 Ma (MSWD = 0.17, n = 21) relative to the reported

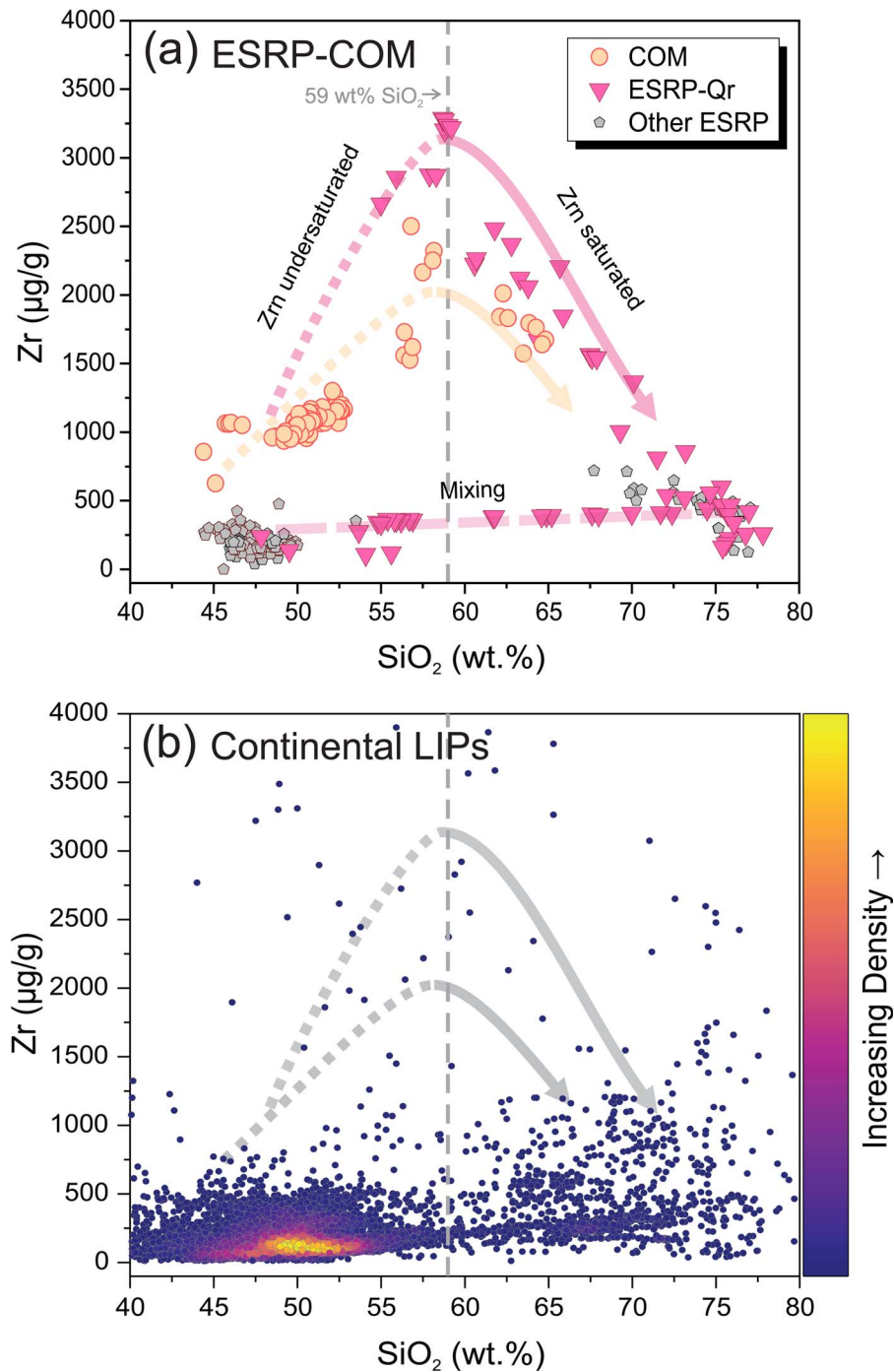


Fig. 2. Whole rock trace element Zr abundances against SiO_2 for Quaternary (Qr) volcanic rocks of the ESRP and COM systems (a). Short-dashed lines indicate magmatic Zr enrichment due to the incompatible behavior of Zr, whereas solid lines designate Zr depletion after the melt reaches zircon saturation at ~ 59 SiO_2 wt.% (McCurry *et al.*, 2008; Chadwick *et al.*, 2023). Long-dashed line represents magma mixing inferred from drill samples of Unnamed Butte (McCurry *et al.*, 2008). To compare whole rock compositions for EPSR and COM with volcanic rocks from large igneous provinces (LIPs), data for continental flood basalt provinces are plotted (b) from pre-compiled data for North American Igneous Province, Siberian Traps, Deccan, Central Atlantic Magmatic Province, Etendeka, Karoo and Ferrar, Paraná, and Etendeka (DIGIS Team, 2021). Gray arrows representing schematic Zr-evolution trends as in panel a.

age of 1065 Ma (Wiedenbeck *et al.*, 1995). Common Pb was corrected using ^{207}Pb , and all $^{206}\text{Pb}/^{238}\text{U}$ ages were adjusted for initial disequilibrium in ^{230}Th , which was estimated using measured Th/U in zircon divided by whole-rock Th/U for the corresponding sample from Ford (2005) and Ganske (2006) as stated in Table S1. All radiometric ages are calculated as weighted averages and uncertainties are reported at 95% confidence, accounting for

excess scatter by multiplying with the square root of the MSWD if >1 , and the corresponding percentile of the Student t distribution for the appropriate degree of freedom (Mahon, 1996). Systematic uncertainties (e.g. from decay constants) for such young zircons are negligible at the stated error levels; practically, this means that multiple levels of age error (i.e. without and with decay constant uncertainty; e.g. Condon *et al.*, 2024) result in the same

Table 1: Summary table of average U–Th and U–Pb ages and isotopic compositions of the zircons from ESRP and COM in this study

Unit	Eruption age (ka)	SiO ₂ (wt.%)	Zr (μg/g)	Sample name	Coordinates (WGS 84)		Zircon crystallization age		δ ¹⁸ O	ε _{Hf} ^z
					Latitude °N	Longitude °W	(ka or Ma)	MSWD		
ESRP										
Big Southern Butte	327 ± 3.2 ^a	76.4 ^d	309 ^d	BSB	43.40311	−113.04247	335 ± 5	0.56	+3.67 ± 0.37	−4.7 ± 0.9
Cedar Butte (Qpc)	400 ± 28 ^a	67.5 ^d	1,563 ^d	CBD	43.38417	−112.90831	426 ± 8	1.14	+3.60 ± 0.48	−4.4 ± 1.5
Cedar Butte (Qr1 flow)		67.6 ^d	1,551 ^d	CBF	43.37106	−112.90831	434 ± 15	1.73	+3.78 ± 0.45	−3.4 ± 2.0
Cedar Butte (Qb5 flow)		61.8 ^d	2,483 ^d	CBF	43.40978	−112.91944	459 ± 37	0.94	+4.01 ± 0.28	−3.9 ± 1.8
East Butte	580 ± 90 ^b	75.4 ^d	465 ^d	EB	43.50100	−112.66256	697 ± 10	0.71	+2.68 ± 0.41	−4.8 ± 1.6
Unnamed Butte	1400 ± 28 ^b	62.7 ^d	580 ^d	UNB	43.48661	−112.68986	1534 ± 22	0.81	+2.21 ± 0.24	−5.0 ± 1.2
COM										
Serrate flow (A)	2.4 ± 0.3 ^c	58.7 ^e	1,677 ^e	SF	43.45798	−113.52443	6.24 ^{+2.61} _{−2.55}	1.32	+5.41 ± 0.48	−10.5 ± 1.4
Devils Orchard flow (A)		58.1 ^e	1,850 ^e	DOF	43.45492	−113.53540			+5.68 ± 0.35	−10.6 ± 1.5
Highway flow (A)		60.6 ^e	1,600 ^e	HF	43.46518	−113.57297			+5.54 ± 0.31	−10.3 ± 1.6
Little Park (D)	-	-	-	Xenolith	43.30207	−113.66669	3239 ± 15 ^f	4.8	+5.82 ± 0.47	−75.8 ± 2.0

For comparison, eruptions ages and whole rock compositional data were compiled from the literature. Uncertainties for age data are stated at 95% confidence; isotopic ratios are stated with 2se uncertainties (see Supplementary Tables S1–S6 for individual zircon data). Map unit designations are stated in parentheses.

^a⁴⁰Ar/³⁹Ar age from Turrin *et al.* (2023). ^bK–Ar age from Kuntz *et al.* (1994). ^c¹⁴C age from Big Craters cogenetic flow from Kuntz *et al.* (2007). ^dAverage from McCurry *et al.* (2008). ^eAverage from Chadwick *et al.* (2023). ^fU–Pb age in Ma in italics. ^gInitial value for average zircon crystallization age, except for xenolith which is stated for a Holocene age.

numerical value, and hence only one error level is stated for simplicity. Similarly, the comparison between different age methods (e.g. U–Pb and K–Ar) theoretically should account for systematic uncertainties resulting from the respective decay constants or references used, but in practice our calculations show that these effects are smaller than the rounded uncertainties stated. For the same reason, we also cite published values as given in the literature.

Trace elements were obtained following methods described in Schmitt *et al.* (2017), applying a ~15 nA ¹⁶O[−] primary beam in Gaussian mode that targeted the same craters previously analyzed for U–Th–Pb geochronology; no extra coating was needed for this step. Energy filtered (~100 eV) secondary ions were detected at low (m/Δm = 2000) mass resolution. Molecular interferences of light rare earth element (LREE) oxides on atomic ion species of the middle and heavy rare earth elements (MREE, HREE) were subtracted for ¹⁵⁶Gd⁺ (¹⁴⁰CeO⁺), ¹⁵⁸Gd⁺ (¹⁴²CeO⁺, ¹⁴²NdO⁺), ¹⁵⁹Tb⁺ (¹⁴³NdO⁺), ¹⁶⁵Ho⁺ (¹⁴⁹SmO⁺), ¹⁶⁹Tm⁺ (¹⁵³EuO⁺), ¹⁷²Yb⁺ (¹⁵⁶GdO⁺), and ¹⁷⁵Lu⁺ (¹⁵⁹TbO⁺) using correction factors for oxide and hydride species determined on doped glasses. Trace element RSFs were calibrated on NIST SRM 610 (Pearce *et al.*, 1997), and accuracy was monitored by analyzing 91 500 reference zircons (Wiedenbeck *et al.*, 2004). All data on unknowns were screened for anomalies in the light rare earth element index (LREE-I = Dy/Nd + Dy/Sm), discarding all values <30 (Bell *et al.*, 2016, 2019), which are interpreted to indicate overlap onto non-zircon phases (e.g. apatite, glass, or oxide inclusions). Titanium abundances in zircon were translated into Ti-in-zircon crystallization temperatures (TiZr) based on the equation in Ferry & Watson (2007). For ESRP, a value of α_{TiO₂} = 0.57 was calculated from magnetite–ilmenite equilibrium exchange (Ghiorso & Evans, 2008) from Unnamed Butte using compositions reported in McCurry *et al.* (2008). Because most ESRP lavas contain quartz, α_{SiO₂} = 1 is applied, acknowledging that for Cedar Butte intermediate lavas this is likely an overestimate (e.g. by ~35°C for α_{SiO₂} = 0.72). For COM evolved flows, magnetite–ilmenite compositions from Stout *et al.* (1994) yielded α_{TiO₂} = 0.55, and α_{SiO₂} = 0.72 assuming melt equilibrium with orthopyroxene (Beattie, 1993). For completeness, COM xenolith zircon temperatures were also calculated assuming unity values of α_{TiO₂} and α_{SiO₂} in the absence of other constraints

(Fig. S3). Analytical uncertainties translate into temperature uncertainties of <10°C based on the external reproducibility of Ti analysis of 91 500 zircon reference (4.7 ± 0.3 μg/g; 1 s; n = 11). Melt oxygen fugacity (fO₂) is expressed relative to the fayalite–magnetite–quartz buffer (ΔFMQ) after Loucks *et al.* (2020). The full SIMS petrochronological data set, including values for the secondary zircon references, is presented in the Supplement (Tables S1–S3).

Zircon (U–Th)/He dating

After the U–Th disequilibrium dating, zircon from Highway flow were dated by (U–Th)/He methods in the Western Australia ThermoChronology Hub (WATCH) Facility at the John de Laeter Centre (Curtin University), following the methods described in (Danišik *et al.*, 2012; Danišik *et al.*, 2017; Danišik *et al.*, 2020). This sample was selected as it yielded comparatively abundant zircon in heavy mineral separates. Representative zircon crystals were individually photographed and measured for physical dimensions with the purpose of calculating their dimensions to further correct the ages for alpha ejection (Ft-correction; Farley *et al.*, 1996). Our initial experiments on He extraction in the investigated zircon revealed extremely low He contents in single zircon crystals stemming from their young eruption age. Therefore, composite samples with multiple zircon grains were analyzed to increase the He signal. Accordingly, multigrain aliquots (n = 33, 50, and 44) were prepared by selecting zircon grains of similar size and shape to minimize variation in Ft correction factors. The selected grains were then transferred and packed into niobium (Nb) microtubes and loaded into an Alphachron II instrument for He extraction. ⁴He together with other gases were extracted at ~1250°C under ultra-high vacuum using a diode laser, cleaned on Ti–Zr getters, and spiked with 99.9% pure ³He gas. The volume of ⁴He was measured by isotope dilution on a QMG 220 M1 Pfeiffer Prisma Plus mass spectrometer. A ‘re-extract’ was run after each analysis to verify complete outgassing of the crystal. Helium gas signals were corrected for blank, determined by analyzing empty Nb microtubes interspersed between the unknowns using the same gas extraction procedure. After the He measurements, Nb microtubes containing the crystals were retrieved from the Alphachron, spiked with ²³⁵U and ²³⁰Th, and dissolved in Parr acid digestions

vessels in two cycles of HF, HNO₃ (cycle 1), and HCl acids (cycle 2), following the procedures described in Evans *et al.* (2005). Sample, blank, and spiked standard solutions were then diluted with Milli-Q water and analyzed by isotope dilution for ²³⁸U and ²³²Th, and by external calibration for ¹⁴⁷Sm on an Agilent 7700 ICP-MS. The total analytical uncertainty (TAU) was calculated as a square root of sum of squares of uncertainty on He and weighted uncertainties on U, Th, and Sm measurements. The zircon (U–Th)/He ages were corrected for alpha ejection (Ft-correction) after Farley *et al.* (1996), whereby homogenous distributions of U, Th, and Sm were assumed for the crystals, and average Ft correction factor was calculated from Ft factors of individual grains in each aliquot. Fish Canyon Tuff zircon with a (U–Th)/He age of 28.3 ± 1.3 Ma (Reiners, 2005) was used as reference standard yielding a mean (U–Th)/He age of 28.5 ± 1.8 Ma (n = 3). The Ft-corrected (U–Th)/He ages were corrected for disequilibrium using the MCHecalc software (Schmitt *et al.*, 2010) where the D₂₃₀ parameter was calculated after dividing the Th/U ratios from the composite samples by the Highway flow whole-rock value (Chadwick *et al.*, 2023), whereas D₂₃₁ was assumed as 3.0 based on Pa/U zircon–rhyolite melt partitioning ratio (Schmitt, 2007). The full data set, including values for the secondary zircon reference, is presented in the supplement (Table S4).

Oxygen and hafnium isotopes

Oxygen isotope (¹⁶O and ¹⁸O) analysis of zircon was performed in two sessions using the CAMECA ims 1280-HR at the Heidelberg Ion Probe (HIP) laboratory. Primary beam intensities were 1.0 to 1.6 nA (Cs⁺, critically illuminated, 10 μm raster) and negative secondary ions of ¹⁶O and ¹⁸O were simultaneously detected using two FCs in the multicollection array with electrometers equipped with a 10¹⁰ Ω resistor for ¹⁶O and a 10¹² Ω resistor for ¹⁸O measurements. Each analysis comprised 20 seconds of presputtering (12 μm raster), automatic centering routines for the secondary beam, and 75 seconds of data collection (15 cycles of 5-second integration time). The FC baselines were recorded during the presputtering. For ¹⁶O, the corresponding baseline was subtracted from raw counts, whereas for ¹⁸O a model was fitted to all individual FC baseline measurements per day, and interpolated values were used for correction. Zircon reference materials 91500 (δ¹⁸O = +9.86‰; Wiedenbeck *et al.*, 2004) and AS3 (secondary standard, δ¹⁸O = +5.34‰; Trail *et al.*, 2007) were intermittently analyzed to determine instrumental mass fractionation factors and to monitor and correct instrumental drift. For drift correction, 10 to 15 unknowns were bracketed by eight analyses of 91500 or AS3 as primary references; the average repeatability (1s) for them was 0.17, (n = 48) and 0.16 (n = 75) for the two sessions (Table S5). Analytical uncertainties were calculated by propagating the internal error of each analysis and the standard error of the corresponding bracket in quadrature. Accuracy was monitored from replicate analyses of secondary reference zircons, and results were always within <0.4‰ of the reported values. All δ¹⁸O values are relative to VSMOW (Baertschi, 1976).

For Hf isotope measurements, a ThermoScientific NEPTUNE Plus multicollector (MC)–ICP–MS system at FIERCE, Goethe University Frankfurt, coupled to a RESOLUTION 193 nm ArF Excimer laser system with a S155 two-volume ablation cell was used. The analytical method including data processing was as described in Gerdes & Zeh (2006, 2009). Laser spots with lateral dimension of ~33 μm in diameter were set on top of SIMS analysis pits. Each grain was ablated for 38 seconds using a fluence of 3.5 J cm^{−2} and a frequency of 5.2 Hz, which resulted in an ablation rate of 0.6 μm s^{−1}, or ~23 μm total depth. An exponential law

with ¹⁷⁹Hf/¹⁷⁷Hf value of 0.7325 was used for instrumental mass bias corrections. For Yb isotopes, instrumental mass bias was corrected using the Hf mass bias of the individual integration step multiplied by a daily βHf/βYb offset factor (see Gerdes & Zeh, 2009). This correction factor is very reproducible for over 20 years with a value of 1.080. All data were adjusted relative to the JMC475 ¹⁷⁶Hf/¹⁷⁷Hf ratio = 0.282160 and quoted uncertainties are quadratic additions of the within-run precision of each analysis and the reproducibility of JMC475 (0.0028% 2se, n = 8). Initial ¹⁷⁶Hf/¹⁷⁷Hf and εHf were calculated using the U–Th–Pb and U–Pb ages determined by SIMS, and the following CHUR parameters: ¹⁷⁶Lu/¹⁷⁷Hf = 0.0336, and ¹⁷⁶Hf/¹⁷⁷Hf = 0.282785 (Bouvier *et al.*, 2008). Secondary zircon reference GJ-1 (¹⁷⁶Yb/¹⁷⁷Hf of <0.008) and Temora (¹⁷⁶Yb/¹⁷⁷Hf of 0.02–0.07) analyzed in the same analytical session show a repeatability of 0.6 and 0.8 εHf units (2s) and agree well with published values (Table S6).

RESULTS

Zircon crystallization and eruption ages

Most of the zircon crystals from ESRP rhyolite domes and flows have typical dimensions of >100 μm in length (up to 320 μm) and occur as individual crystals in the matrix (Fig. 3a). By contrast, COM lava flow zircon crystals are generally smaller and between 60 and 250 μm in length and they are also frequently ingrown with other mineral phases, especially with olivine and ilmenite (Fig. 3b–d). Zircon crystal morphology in both sample suites is mostly euhedral and dominated by prismatic forms. Internal structures include glass and apatite inclusions, with CL imaging frequently showing coarse oscillatory and minor sector zonation, as well as dark interiors for ESRP and COM (Fig. 4a–b). COM xenolith zircons are up to ~250 μm long and have rounded edges. Internally, they display fine oscillatory zonation with truncated inner domains (Fig. 4c).

ESRP zircon crystallization ages typically predate the corresponding eruption ages based on published K–Ar and ⁴⁰Ar/³⁹Ar results (Table 1 and Fig. 5a–f; Kuntz *et al.*, 1994; Turrin *et al.*, 2023). The oldest dome, Unnamed Butte, yielded a ²⁰⁶Pb/²³⁸U age of 1534 ± 22 ka (MSWD = 0.81, n = 38) while the published K–Ar age is 1400 ± 28 ka (Kuntz *et al.*, 1994). After a c. 800 ka hiatus, the next youngest zircon crystallization episode occurred underneath East Butte at 697 ± 10 ka (MSWD = 0.71, n = 22) with a K–Ar date of 580 ± 90 ka recording the eruption. Following another hiatus of c. 250 ka, the Cedar Butte dome and one of its stratigraphically lowest flows (silicic trachyte Qr1; McCurry *et al.*, 2008) record the same zircon crystallization episode at 426 ± 8 (MSWD = 1.14, n = 21) and 434 ± 15 ka (MSWD = 1.73, n = 22), respectively. A second flow from Cedar Butte (mafic trachyte Qb5) yielded similar zircon crystallization ages averaging 459 ± 37 ka (MSWD = 0.94, n = 13), but uncertainties are elevated because of comparatively low U concentrations in these zircons (see next section). Collectively, Cedar Butte zircon crystallization ages are close to the K–Ar and ⁴⁰Ar/³⁹Ar ages of 400 ± 19 ka and 400 ± 28 ka, respectively (Kuntz *et al.*, 1994; Turrin *et al.*, 2023). The youngest zircon crystallization ages were determined for Big Southern Butte dome at 335 ± 5 ka (MSWD = 0.56, n = 37), whereas the inferred eruption age based on K–Ar and ⁴⁰Ar/³⁹Ar geochronology is only slightly younger at 309 ± 10 ka and 327 ± 3.2 ka, respectively (Kuntz *et al.*, 1994; Turrin *et al.*, 2023). The differences between U–Pb zircon crystallization ages and eruption ages (Δt) translate into zircon residence times that are generally brief (between 8 ± 5 ka for Big Southern Butte and 134 ± 47 ka for Unnamed Butte). Even when considering decay constant uncertainties for the different dating

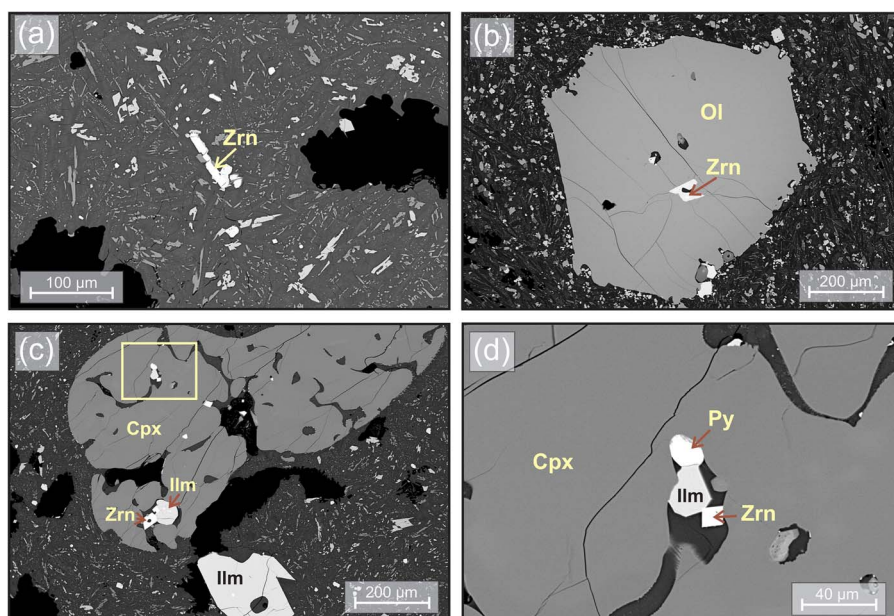


Fig. 3. Backscattered electron images for COM thin sections: (a) zircon crystal in matrix; (b) zircon crystal included in olivine; (c) two zircon crystals associated with ilmenite and pyrite partially included in clinopyroxene; (d) enlarged image of zircon included in clinopyroxene.

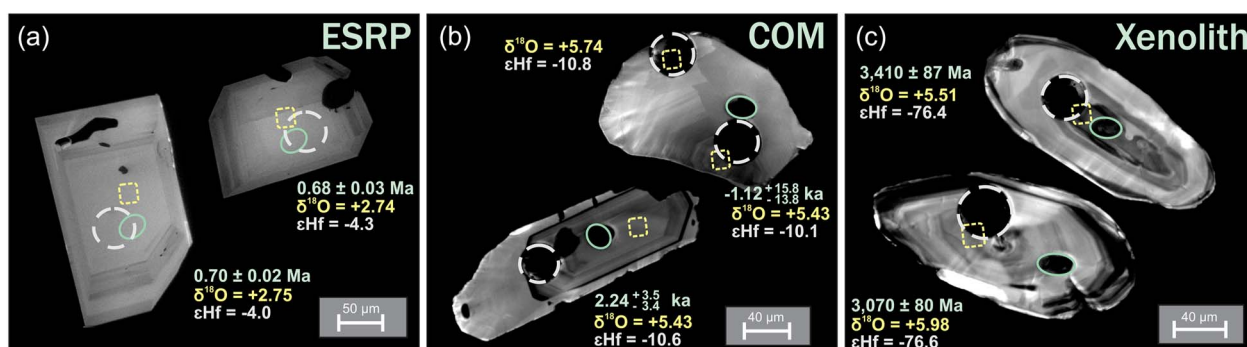


Fig. 4. Cathodoluminescence images of zircon from ESRP and COM samples: (a) East Butte zircons SRP_EB_113 (left) and SRP_EB_111 (right), (b) Devils Orchard flow zircon COM_DOF_49 (upper) and Highway flow zircon COM_HF_14 (lower), and (c) xenolith zircon from Little Park flow COM_MHK_01_61 (upper) and COM_MHK_01_100 (lower). Note oscillatory and minor sector zonation in a and b, as well as rounded crystal boundaries in c. Analytical spots and corresponding zircon ages and isotopic values are matched by color; trace elements (not shown) were analyzed in the same spot as for dating.

methods, these values only change insignificantly (i.e. when using Renne et al., 2010, vs. the original Steiger & Jäger, 1977, values, zircon residence times are between 23 ± 13 ka for Big Southern Butte and 121 ± 46 ka for Unnamed Butte).

Zircon crystals extracted from Craters of the Moon evolved lavas (Highway, Devils Orchard, and Serrate flows) fall on the same near-horizontal trend in the $(^{230}\text{Th})/(^{232}\text{Th})$ vs. $(^{238}\text{U})/(^{232}\text{Th})$ diagram (Fig. 6a), signifying very recent zircon crystallization. By fixing an isochron through a melt composition as represented by whole-rock data for Highway flow (Reid, 1995) which we used also for (U–Th)/He dating, a model age of $6.2^{+2.6}_{-2.5}$ ka (MSWD = 1.30; $n = 23$) is obtained. Only four crystals (two from Devils Orchard flow and two from Highway flow) lie on the equiline, in agreement with five reconnaissance rim analyses on xenolith zircons, thus indicating zircon crystallization ages >375 ka. For zircon in secular equilibrium from the Devils Orchard flow, we subsequently determined concordant $^{206}\text{Pb}/^{238}\text{U}$ ages of 23.2 ± 2.0 Ma and 1390 ± 132 Ma, respectively. In order to quantify the difference between zircon crystallization and eruption ages for COM, (U–Th)/He zircon dating was performed on three composite zircon samples from previously undated Highway flow. The resulting

(U–Th)/He age of 2.05 ± 0.56 ka ($n = 3$; Goodness of fit = 0.48) agrees with the inferred eruption age for Highway flow based on stratigraphic correlation with the most recent flows from period A for which ^{14}C ages indicate emplacement at c. 2.4 ka (e.g. Kuntz et al., 2007). This translates into a nominal $\Delta t = 4.1 \pm 2.7$ ka, implying that zircon residence for the evolved COM lavas was on the millennial scale. Zircon from COM xenolith MHK-01 is mostly discordant (Fig. 6b), with discordia pointing to, with one exception, an Archean crystallization age that is defined by a 3239 ± 15 Ma upper concordia intercept of an errorchron defined by the data (MSWD = 4.8; $n = 18$). The lower intercept at 155 ± 117 Ma indicates Cenozoic Pb-loss.

Zircon trace elements

Uranium concentrations in the ESRP zircon crystals range from 15 µg/g (Qb5 flow from Cedar Butte) to 2970 µg/g (Big Southern Butte). As typical for igneous zircon (e.g. Grimes et al., 2007), they also correlate with Y, HREE (e.g. Yb), Hf, and Th concentrations, with a large compositional range covered by Cedar Butte intermediate lava flows, whereas concentrations for individual domes are higher and largely homogeneous (e.g. Fig. S1). Uranium

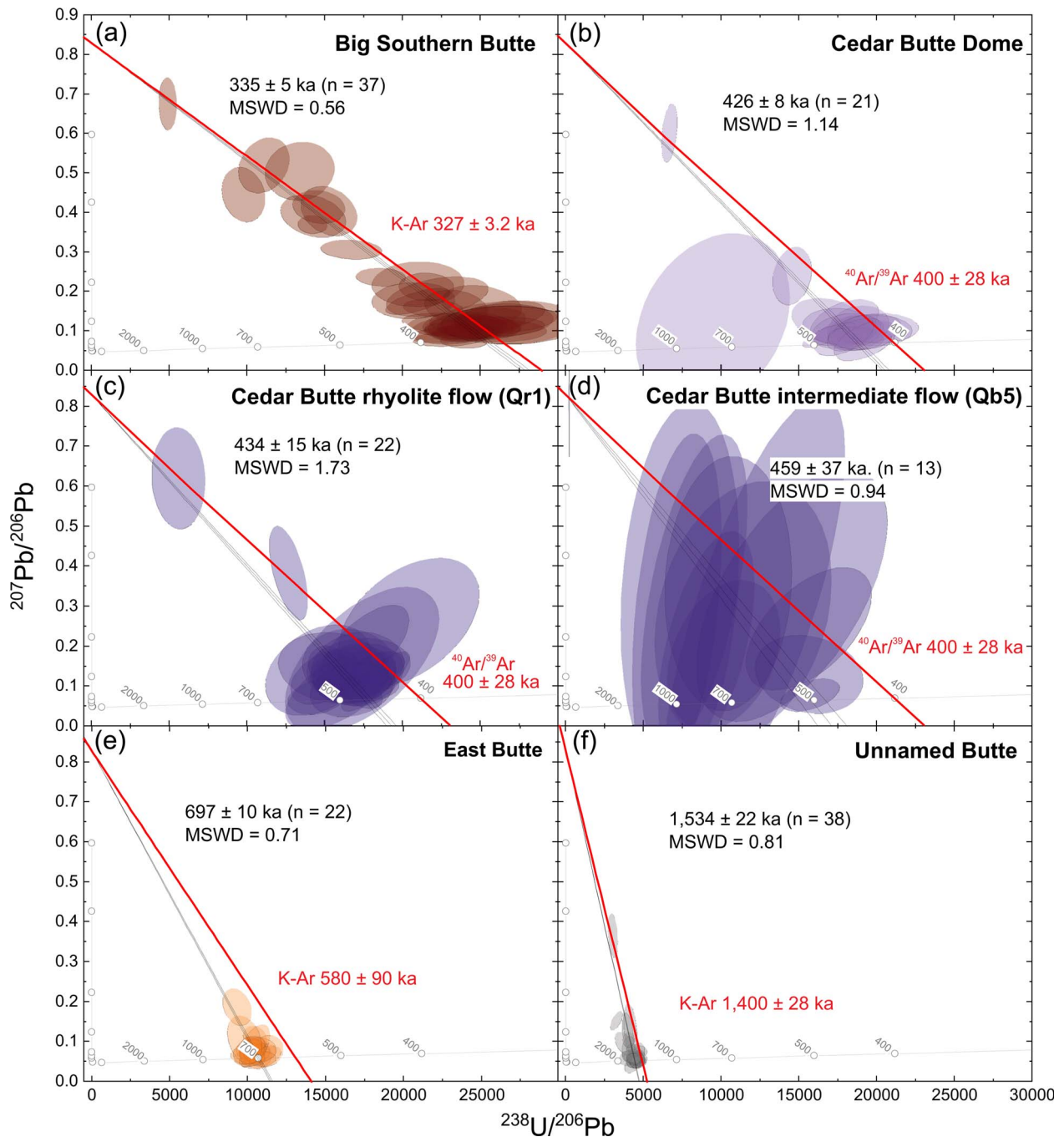


Fig. 5. Zircon $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$ data (2se) for ESRP rhyolite domes and intermediate flows plotted uncorrected for common Pb. Individual samples are plotted for Big Southern Butte (a), Cedar Butte Dome (b), Cedar Butte rhyolite flow (c), Cedar Butte intermediate flow (d), East Butte (e), and Unnamed Butte (f). Data define linear arrays between common Pb (y-axis intercept) and radiogenic Pb. Sample ages were calculated as weighted averages of $^{206}\text{Pb}/^{238}\text{U}$ ages which were individually adjusted for initial disequilibrium of ^{230}Th (see text). Concordia is plotted for disequilibrium of ^{230}Th and ^{231}Pa at the time of crystallization using initial activity ratios for the intermediate daughter and parent isotopes of 0.2 and 2.2, respectively. Thick (red) line shows model isochrons for eruption ages (Table 1).

concentrations in COM zircon start at a similarly low limit of 11 $\mu\text{g/g}$, but only reach a maximum value of 512 $\mu\text{g/g}$; other trace elements including Y, Yb, Hf, and Th track this enrichment trend (Fig. S1). Xenolith zircon U concentrations range from 225 to 1450 $\mu\text{g/g}$, with typical correlations between U and these trace elements (Fig. S1). U/Yb ratios are used to discriminate petrogenetic origins of zircon, especially between continental and oceanic crustal sources with $\text{U/Yb} > 0.1$ and < 0.1 , respectively (Grimes *et al.*, 2007). Zircon from the ESRP domes plots near $\text{U/Yb} = 1$,

whereas the least enriched zircon from the Cedar Butte and COM intermediate flows are close to the $\text{U/Yb} = 0.1$ line (Fig. S2). Xenolith and xenocrystic zircons, by contrast, typically display $\text{U/Yb} > 1$ (Fig. S2).

All ESRP and COM zircon REE patterns are enriched in HREE relative to MREE and LREE (Fig. 7a–h) with typical magmatic slopes as expressed by Gd/Yb being on average between 0.07 and 0.1, respectively. They also display negative Eu and positive Ce anomalies which primarily reflect plagioclase fractionation and

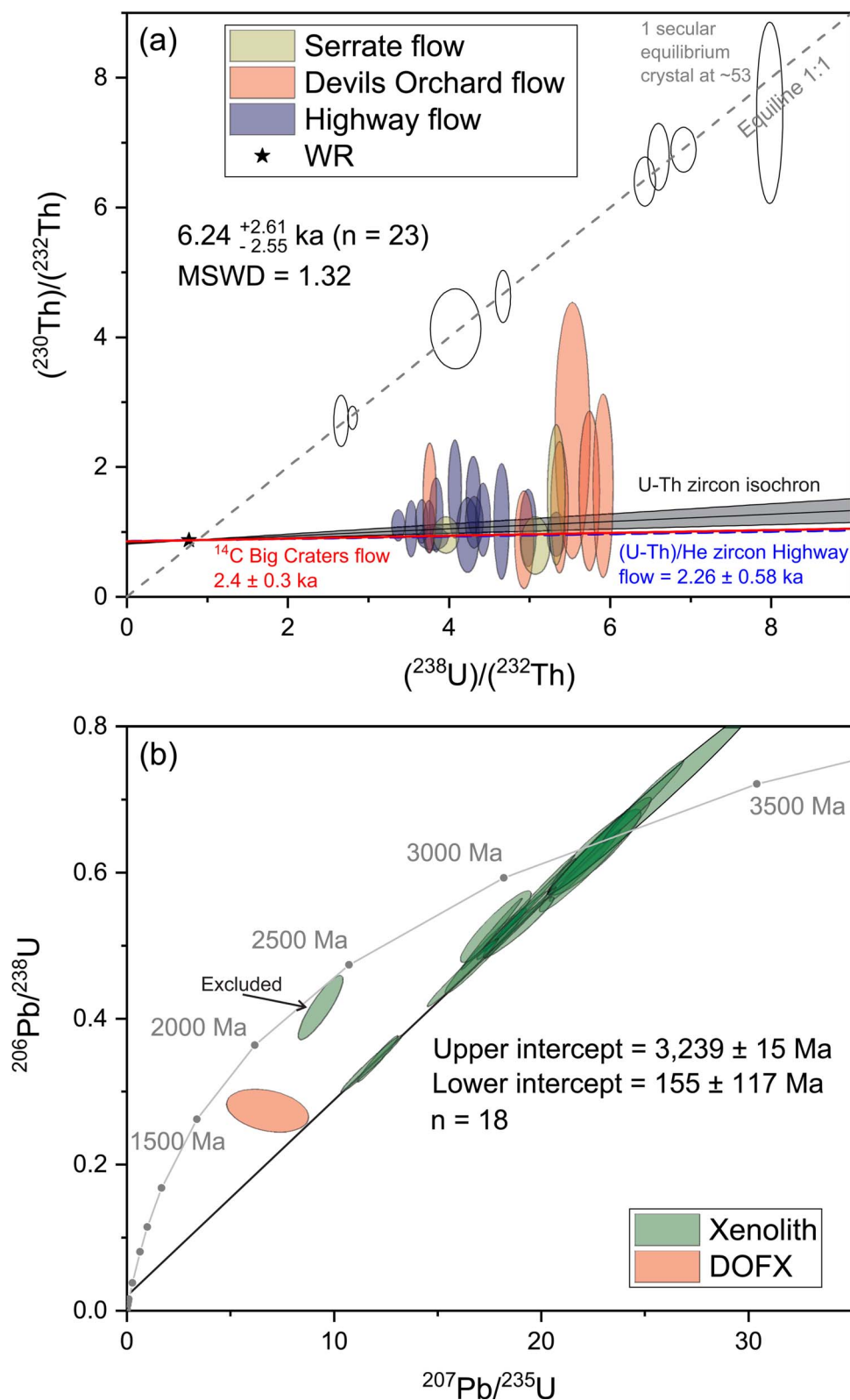


Fig. 6. (a) Zircon $(^{230}\text{Th})/(^{232}\text{Th})$ vs. $(^{238}\text{U})/(^{232}\text{Th})$ isochron diagram for COM intermediate lavas. Whole rock composition of Highway flow (Reid, 1995) is used to anchor the isochron. Model isochron based on the average (U-Th)/He zircon age is plotted for comparison. Open ellipses overlapping with the equiline are interpreted as secular equilibrium xenocrysts, in part confirmed by U-Pb analysis. (b) Zircon $^{206}\text{Pb}/^{238}\text{U}$ vs. $^{207}\text{Pb}/^{235}\text{U}$ data from the xenolith and two secular equilibrium crystals from the Devils Orchard flow. Data is corrected for common Pb using measured ^{204}Pb . Lower and upper concordia intercept ages are determined using Isoplot R (Vermeesch, 2018); error ellipses and ages at 2se.

preferential incorporation of Ce^{4+} into zircon controlled by $f\text{O}_2$, respectively (Trail et al., 2012; Loucks et al., 2020; Trail et al., 2024). Cedar Butte intermediate lavas display a wide range of

Eu/Eu^* from 0.03 to 0.74, again similar to COM zircon with a range of 0.14 to 0.63. Zircon from the rhyolitic ESRP domes are individually homogeneous, with Eu/Eu^* decreasing in the

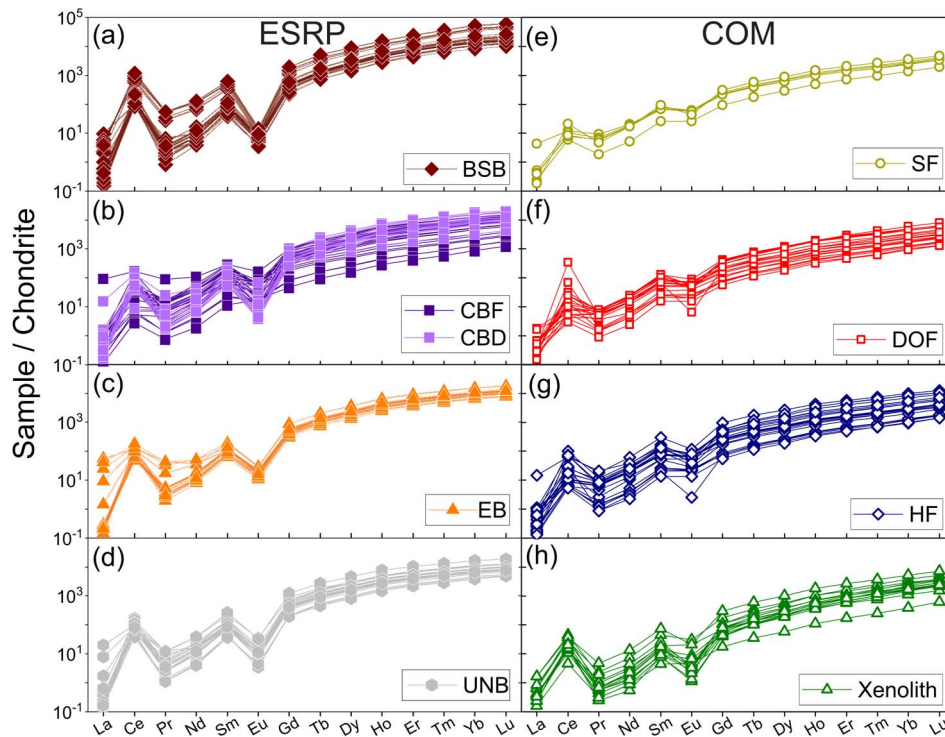


Fig. 7. Chondrite normalized REE abundances for ESRP (a–d), COM (e–g) and xenolith (h) zircon crystals. All data were screened for contamination by non-zircon phases using the LREE-I > 30 filter (see text). Unit abbreviations from Table 1. Chondrite values from Anders & Grevesse (1989).

order (average $\pm 1\sigma$): East Butte (0.079 ± 0.010), Unnamed Butte (0.052 ± 0.011), Big Southern Butte (0.037 ± 0.003), and Cedar Butte dome (0.058 ± 0.088). Xenolith zircon REE patterns are comparatively steep (average $Gd/Yb = 0.05$), and Eu/Eu^* ranges between 0.06 and 0.58.

Translating Ti concentrations into Ti/Zr values indicates average temperatures for zircon crystallization from 733°C (Unnamed Butte) to 683°C (Big Southern Butte). In comparison, higher, but variable Ti/Zr values are determined for ESRP Cedar Butte (695–928°C; Fig. 8a) and COM lava flows (710–919°C; Fig. 8c). Ti/Zr values positively correlate with Eu/Eu^* and Zr/Hf (Fig. 8b and d), indicating that variable Ti concentrations in zircon primarily record magmatic temperature variations rather than variable $aTiO_2$. Changes in temperature correlate with oxidation state (ΔFMQ) only in the intermediate lavas of ESRP and COM. They start out at high temperature below FMQ, consistent with low- fO_2 inferred for ESRP magma differentiation (Whitaker et al., 2008), and reach high ($\Delta FMQ = +2$ – $+3$), but constant values for rhyolitic zircon (Fig. 9a and b). By contrast, the prominent decrease in Th/U and Gd/Yb for the rhyolitic ESRP lavas suggests coprecipitation of apatite preferentially incorporating Th and MREE (Fig. 9c and e) while this is negligible for COM zircon (Fig. 9d and f). Xenolith Ti/Zr values are nominally between 710 and 838°C (Fig. S3).

Zircon O- and Hf-isotopes

ESRP zircon $\delta^{18}O$ values overall vary between +1.99 and +4.32 (Table S5; Fig. 10a), significantly below typical mantle zircon values ($+5.3 \pm 0.3$; Valley et al., 2005). Although zircon $\delta^{18}O$ averages for individual ESRP domes and lavas are distinct, individual units display homogeneous $\delta^{18}O$ within analytical uncertainty (Table 1; Fig. 10a). The lowest average values are present in Unnamed Butte ($+2.21 \pm 0.24$; 2s), followed by East Butte ($+2.68 \pm 0.41$). In comparison, Big Southern Butte with an average of $+3.67 \pm 0.37$ and Cedar Butte flows and dome ($+3.85 \pm 0.45$ and $+3.67 \pm 0.37$,

respectively) are isotopically slightly heavier. By contrast, COM zircon $\delta^{18}O$ values are $+5.54 \pm 0.31$ (Highway flow), $+5.68 \pm 0.65$ (Devils Orchard flow), and $+5.41 \pm 0.48$ (Serrate flow), all closely overlapping mantle zircon compositions, without any significant difference between units. Zircon crystals from the COM granulate xenolith have an average $\delta^{18}O$ value of $+5.82 \pm 0.47$, which is slightly elevated relative to mantle values (Fig. 10b).

Zircon ϵ_{Hf} values for ESRP vary between -6.4 and -1.3 (Table S6; Fig. 10c). Whereas $\delta^{18}O$ values between different ESRP domes and flows are variable, their Hf-isotopic composition is remarkably homogeneous (Table 1; Fig. 10c). In stark contrast, COM lava flow zircon ϵ_{Hf} values are more negative (-12.0 to -8.9 ; Table S6; Fig. 10d). Precambrian zircons from the xenolith display the lowest ϵ_{Hf} values between -77.1 and -73.2 for present day. Initial values calculated back to the concordia upper intercept age are between -4.1 and 0.3 , near or slightly below the chondritic value. In all cases, zircon ϵ_{Hf} values are lower than regional mafic rocks representing potential parental magmas with average $\epsilon_{Hf} = +0.6$ (Stefano et al., 2019).

DISCUSSION

Autocrystic and xenocrystic provenance of ESRP and COM zircons

ESRP and COM are volcanic fields dominated by mafic lava flows which based on their high temperatures and low Zr abundances were far from zircon saturation. One key difference between the ESRP with its prominent—albeit volumetrically subordinate—rhyolite domes is the complete lack of such highly evolved magma compositions in the erupted COM lavas. Cedar Buttes lavas together with enclaves in East Butte and some Unnamed Butte borehole samples in the ESRP are intermediate in composition with similar SiO_2 as evolved COM lavas (Fig. 2a). With differentiation, Zr increases during the zircon-undersaturated; at

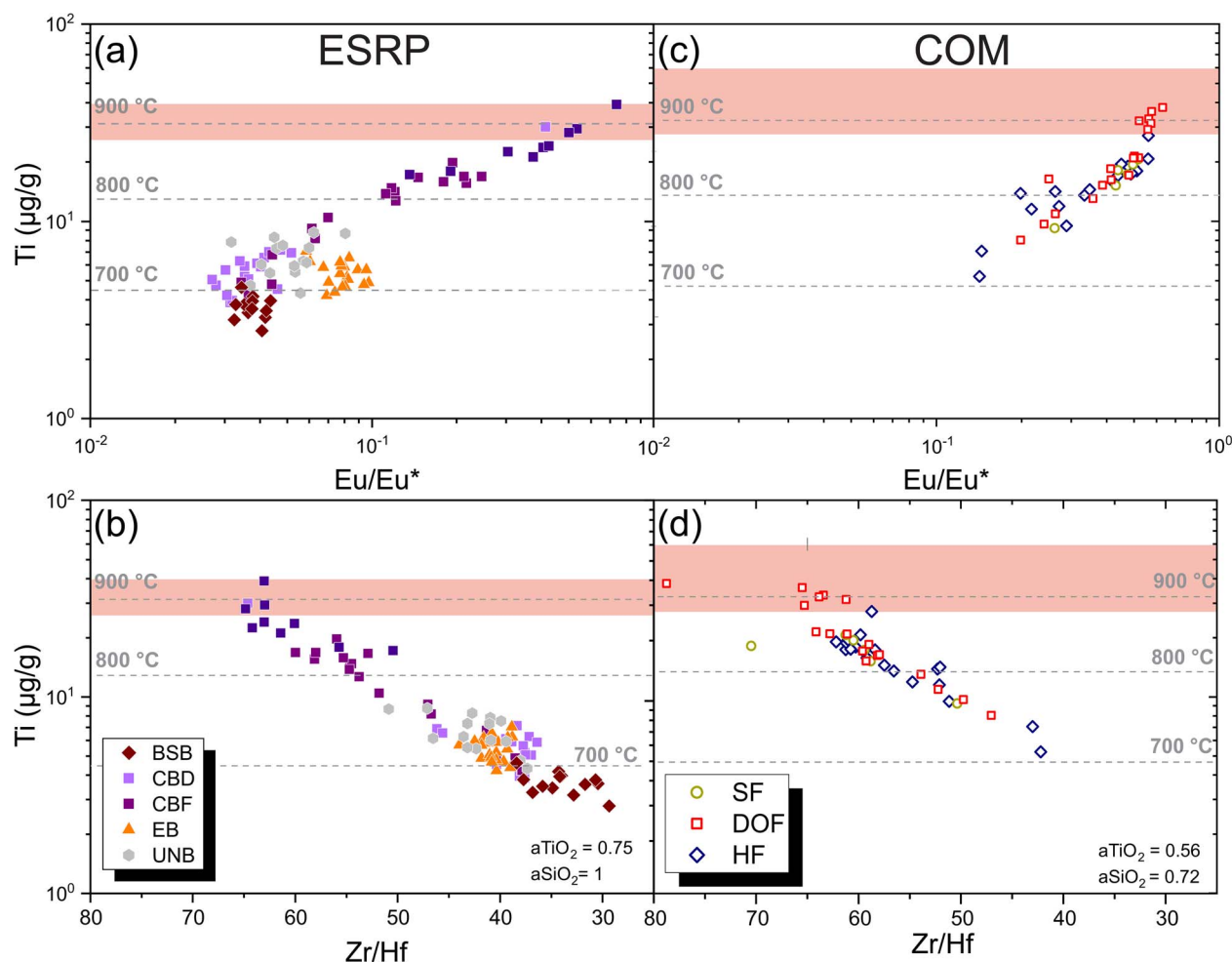


Fig. 8. Zircon Ti abundances with corresponding Ti-in-zircon temperatures (dash-dot lines) calculated for ESRP (a–b) and COM (c–d) samples against Eu/Eu* and Zr/Hf. Average zircon saturation temperatures for corresponding whole rock compositions (fields) were calculated after [Crisp and Berry et al. \(2022\)](#). Unit abbreviations from [Table 1](#).

SiO₂ concentrations of ~59 wt.% Zr depletion kicks in, which is indicative of zircon saturation ([Fig. 2](#)). Hence, autocrystic zircon is expected for ESRP rhyolites, and our geochronological data confirm this, as no xenocrysts were detected among 140 zircon crystals dated, and all U–Pb crystallization ages are close to the Middle–Late Pleistocene eruption ages inferred from K–Ar and ⁴⁰Ar/³⁹Ar dating. For the COM intermediate lavas with SiO₂ between 52.3 and 64.6 wt.% ([Chadwick et al., 2023](#)), the presence of autocrystic zircon is less clear cut in the Zr vs. SiO₂ diagram because Zr averages across the ~59 wt.% SiO₂ boundary are subequal ([Fig. 2](#)). [Kuntz \(1992\)](#) reported zircon in granulitic xenoliths within COM lavas, whereas [Stout et al. \(1994\)](#) were first to note the presence of accessory zircon in COM lava flows. They described it as inclusions in plagioclase megacrysts and glomerocrysts comprising olivine, clinopyroxene, magnetite, and apatite. Based on textural considerations, [Stout et al. \(1994\)](#) concluded that the presence of zircon points to crustal contamination, and/or crystallization from melts unusually enriched in Zr, but neither the xenolith nor melt sources for these zircon crystals were identified.

Based on U–Th zircon crystallization ages, we classify most COM zircon crystals as Holocene. U–Th disequilibrium dating of magmatic accessory minerals is based on the decay of ²³⁰Th with a shorter half-life compared with its indirect long-lived parent

²³⁸U, and it is applicable to the time until the system reaches secular equilibrium (i.e. <375 ka; [Cheng et al., 2000](#)). Diffusion experiments for the relevant parent–daughter elements U and Th ([Cherniak & Watson, 2003](#)) indicate negligible mobility even at magmatic conditions (i.e. ~930 °C; [Putirka et al., 2009](#)), and U–Th zircon ages, therefore, reliably date crystallization in the magma. By contrast, He diffusion in zircon is fast (e.g. [Reiners et al., 2004](#); [Cherniak et al., 2009](#); [Gautheron et al., 2020](#)), and although zircon crystallization predates the eruption, the (U–Th)/He zircon ages record eruptive cooling at 2.26 ± 0.29 ka. The overlapping uncertainties between U–Th and (U–Th)/He imply zircon crystallization immediately prior to eruption. Moreover, the (U–Th)/He zircon eruption age overlaps within uncertainty with ¹⁴C ages for mafic lavas of period A ([Kuntz et al., 2007](#)), confirming that the evolved COM flows are part of the youngest eruptive episode of the field.

Only four zircons in secular equilibrium were found in Highway and Devils Orchard flows. For two of these, U–Pb ages indicate Oligocene–Miocene and Proterozoic ages. Trace elements and isotopic compositions ($\delta^{18}\text{O}$, ϵ_{Hf}) for one of the undated zircons resembles the Archean xenolith population. Although xenocrysts are rare in COM lavas, these data confirm that Archean country rock contaminated COM lavas, but also that other lithologies are present at depth, including hotspot-related felsic rocks (e.g. [Putirka et al., 2009](#)).

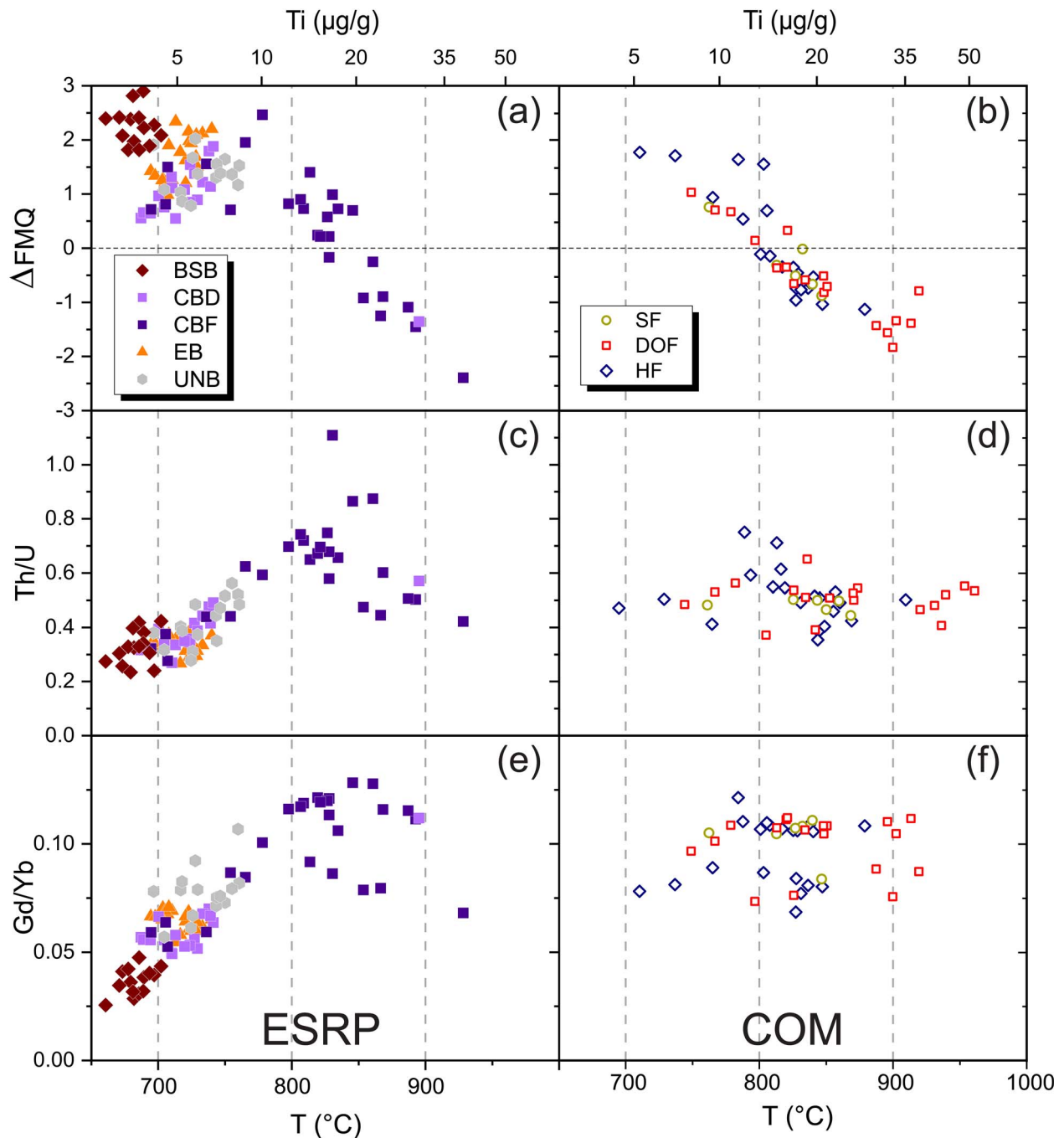


Fig. 9. Zircon Ti abundances and Ti-in-zircon temperatures calculated for ESRP and COM zircon crystals against (a–b) ΔFMQ calculated after Loucks *et al.* (2020), Th/U (c–d), and Gd/Yb (e–f). Unit abbreviations from Table 1.

Boundary constraints for magma evolution from previous whole-rock based models

Before evaluating new zircon petrochronological data, we briefly review existing models for magma evolution in the ESRP–COM systems. Mafic endmembers in the ESRP comprise primitive olivine tholeiite magmas which generally experienced only minor fractional crystallization and negligible assimilation of Archean basement (e.g. Leeman, 1982a). COM lavas are generally more evolved than the ESRP olivine tholeiites and the most mafic endmembers differ compositionally from the bulk of the ESRP olivine tholeiites (Christiansen & McCurry, 2008). Nonetheless, intermediate to evolved magmas in the ESRP and the COM follow a common evolutionary trend (Christiansen & McCurry, 2008). Starting

from the most primitive intermediate composition of Cedar Butte, a low-Si rhyolite is reached after 77% fractional crystallization (McCurry *et al.*, 2008). For COM, the compositional range of the youngest eruptive period A, which includes the most evolved lavas erupted from this field, indicates fractional crystallization of 50% between less and more evolved lavas flows as represented by Big Craters (51.5 wt.% SiO_2) and Highway Flow (63.5 wt.% SiO_2) (Leeman *et al.*, 1976). Minor covariation between isotopic ratios (e.g. $^{87}\text{Sr}/^{86}\text{Sr}$; $^{143}\text{Nd}/^{144}\text{Nd}$) and fractionation indices (e.g. SiO_2) for this suite of lavas also requires crustal assimilation concomitant with fractional crystallization (AFC; Chadwick *et al.*, 2023; Putirka *et al.*, 2009). COM parental liquids can be linked to ESRP parent magmas with small addition (~ 1 –2%) of Archean

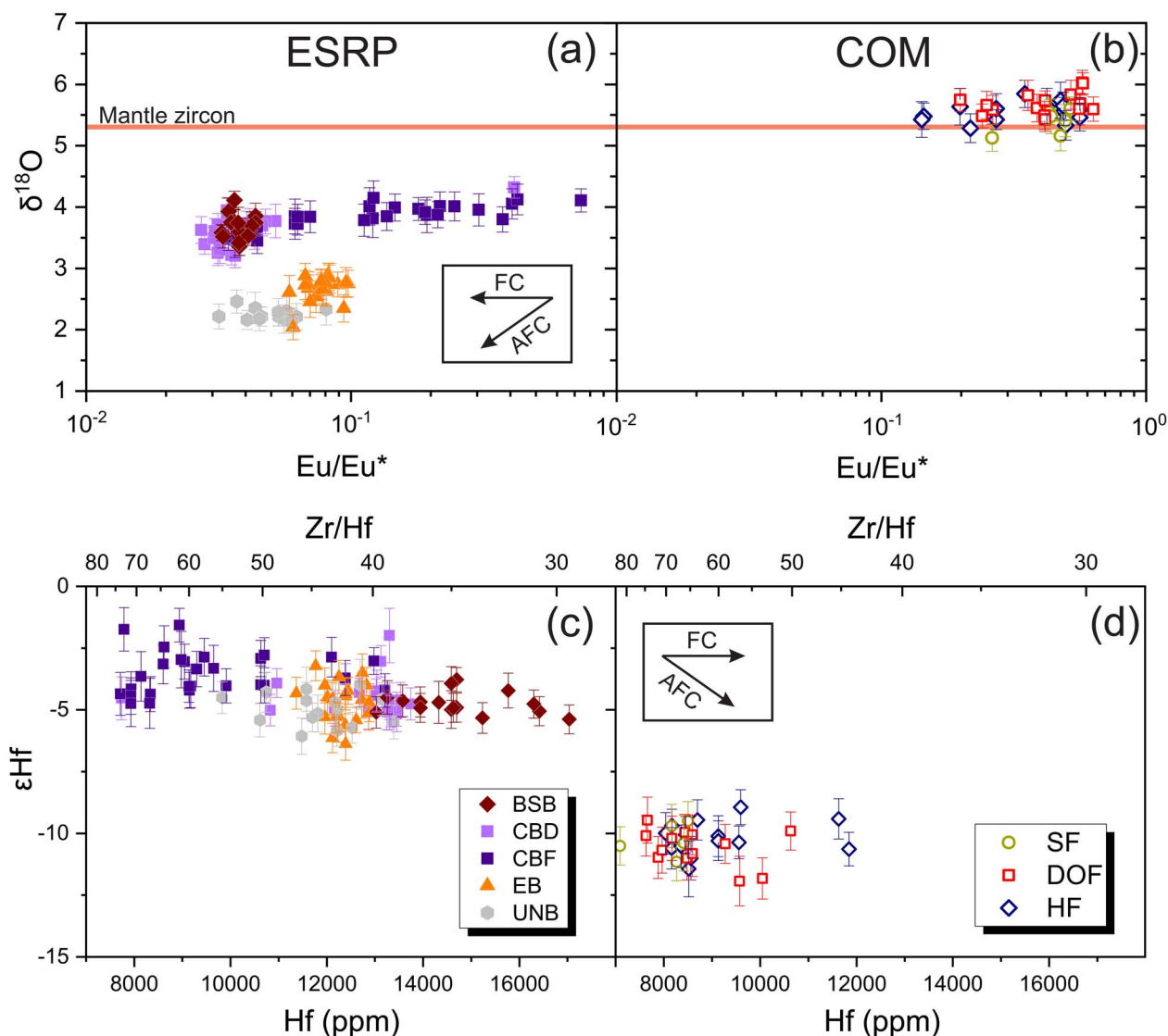


Fig. 10. Zircon O isotopes vs. Eu/Eu^* (a–b), and zircon Hf isotopes vs. Hf (c–d) for ESRP and COM systems (2se). Arrows indicate schematic trends for fractional crystallization (FC) and assimilation coupled with fractional crystallization (AFC). Unit abbreviations from Table 1.

crust during a first assimilation stage (Leeman *et al.*, 1976; Kuntz, 1992; Putirka *et al.*, 2009; Chadwick *et al.*, 2023). A second stage of assimilation involved addition of 15% to 20% felsic wallrocks that are the mid-crustal equivalents to upper-crust derived rhyolitic xenoliths in some COM lavas (Putirka *et al.*, 2009). Even higher degrees of assimilation at 31% were modeled for a rhyolite crustal endmember equivalent to rocks from the Miocene Picabo volcanic field that erupted as a precursor to modern Yellowstone (Chadwick *et al.*, 2023). The evolved COM lavas are inferred to reflect protracted magma storage and differentiation in middle crustal reservoirs at 14 to 17 km (Putirka *et al.*, 2009).

High and low temperature zircon in basalt dominated systems

Having established the predominantly autocrystic origins of zircon in ESRP and COM rocks, it is possible to compare the thermochemical conditions at which they crystallized with constraints from whole rock data. The isotopic invariance of zircon in individual units indicates that assimilation was negligible during zircon crystallization, and we, therefore, interpret zircon trace elements as primarily controlled by

fractional crystallization of the host magma. For ESRP, the average of the Cedar Buttes lavas (Table 2) was selected as a zircon-undersaturated parental magma (McCurry *et al.*, 2008), and for COM, the starting composition was the most mafic rock sample of the three investigated lava flows (Devils Orchard flow, $\text{SiO}_2 \approx 52$ wt.%; Table 2; Chadwick *et al.*, 2023). As differentiation indicators, we modeled trace element ratios (Eu/Eu^* , Zr/Hf) because absolute abundances in zircon are often highly variable at the sub- μm -scale (e.g. Hofmann *et al.*, 2009). By varying F (fraction of remaining liquid), Eu/Eu^* in the remaining melt was obtained using bulk distribution coefficients for Sm, Eu, and Gd calculated based on the mineral assembly required to reach evolved compositions from the respective parental magmas (Table 2). Literature partition coefficients (Table 2) were compiled matching the trachyandesite and basaltic trachyandesite starting compositions for ESRP and COM, respectively.

Modeled melt Eu/Eu^* was transformed into zircon Eu/Eu^* using the partition coefficient $(\text{Eu}/\text{Eu}^*)_D$ derived from Trail *et al.* (2024):

$$\left(\frac{\text{Eu}}{\text{Eu}^*}\right)_D = \frac{1}{1 + 10^{0.30 \pm 0.04 - (0.27 \pm 0.03) \times \Delta \text{FMQ}}} \quad (1)$$

Table 2: Summary of modelling parameters for fractional crystallization (FC)

Mineral phase	Abundance (wt.%)	ESRP			Concentration (μg/g)	
		Sm	Eu	Gd	Zr	Hf
Ol	9.8	0.028 ^a	0.017 ^a	0.016 ^a		
Pl	48.1	0.168 ^b	4.2 ^b	0.24 ^b		
Cpx	17.5	1.890 ^c	1.69 ^c	0.78 ^d		
Mt	20.2	0.024 ^e	2.52 ^e	0.018 ^e		
Ap	4.4	31.4 ^f	25.2 ^f	21.7 ^f		
Total	100					
D*		1.80	3.94	1.21		
Parental magma		29.2	8.80	26.6	2373	49.6
Mineral phase						
	Abundance (wt.%)	COM			Concentration (μg/g)	
		Sm	Eu	Gd	Zr	Hf
Ol	16.7	0.028 ^a	0.017 ^a	0.016 ^a		
Pl	66.6	0.146 ^c	2.35 ^c	0.044 ^c		
Cpx						
Mt	11.9	0.08 ^g	0.07 ^g	0.3 ^g		
Ap	4.8	14.6 ^h	9.6 ^h	15.8 ^h		
Total	100					
D*		1.80	3.94	1.21		
Parental magma		26.01	7.65	24.04	1153	23.4

^a Dunn & Sen (1994), ^b Dudas *et al.* (1971), ^c Nagasawa (1973), ^d Larsen (1979), ^e Lemarchand *et al.* (1987), ^f Luhr *et al.* (1984), ^g Mahood & Stimac (1990), ^h Paster *et al.* (1974).

where the parameter ΔFMQ as the logarithmic difference in fO_2 to the FMQ buffer was derived from the zircon oxygeobarometer of Loucks *et al.* (2020) (Fig. 10). Zr/Hf in zircon was calculated based on the Zr–Hf geothermometer of Aranovich & Bortnikov (2018):

$$T = \frac{1531}{\ln K_d + 0.883} \quad (2)$$

where the parameter K_d is a ratio of the respective partition coefficients between melt (M) and zircon (S) as:

$$K_d = \left(X_{\text{Zr}}^{\text{S}} \times X_{\text{Hf}}^{\text{M}} \right) / \left(X_{\text{Zr}}^{\text{M}} \times X_{\text{Hf}}^{\text{S}} \right) \quad (3)$$

(X representing concentrations of Zr and Hf in μg/g). Following Aranovich & Bortnikov (2018), T is calculated as a function of fractional crystallization via:

$$f = 1 + k(T - T_0) \quad (4)$$

where k refers to the inverse of the difference between the initial temperature of zircon saturation (Fig. 8; T_0 , which was varied between 900°C and 1150°C) and the end of zircon crystallization (here: 650°C). The f value in Equation 4 strictly represents the remaining melt fraction after zircon crystallization. Because our model is intended to only capture the melt evolution during zircon saturation, we can equate f with F , and use the f -dependent model temperature in Equation 4 to calculate ΔFMQ in Equation 1.

Curves for zircon Eu/Eu* and Zr/Hf in ESRP and COM from the respective fractional crystallization scenarios parallel the principal trends displayed by the data. For ESRP, zircon trace elements are matched by F values ranging from just under 1 to ~0.25 (Fig. 11a), whereas a narrower range between $F = \sim 0.95$ and ~0.40 covers the data for COM (Fig. 11b). The bulk (~77%) of ESRP zircon plots near the lower end of the modeled trend, consistent with the rhyolitic host rock. Only Cedar Butte intermediate flow

zircon records crystallization at $F > 0.5$. The bulk (~87%) of COM zircon is consistent with near-saturation crystallization at $F > 0.5$, whereas only ~13% indicate crystallization from residual melts at $F < 0.5$. Although not directly indicating temperature, the modeled temperatures also agree with the observed TiZr distribution for ESRP and COM, where most ESRP zircon crystals indicate low temperature crystallization, whereas TiZr for intermediate Cedar Butte and COM lavas is closer to the respective zircon saturation temperatures derived from their whole-rock compositions (Fig. 8).

For ESRP, zircon trace elements and TiZr values faithfully record melt evolution by fractional crystallization as previously modeled based on whole rock compositions. These models indicate that ESRP rhyolites are residual melts after significant crystallization ($F = 0.23$; McCurry *et al.*, 2008), remarkably close to the value obtained from modelling zircon trace elements ($F = 0.25$; Fig. 11a). For COM, whole rock fractional crystallization was previously modeled for the compositional range displayed by evolved COM period A magmas at $F = 0.49$; notably, this is only a fraction of the total compositional range at COM that requires $F = \sim 0.3$ to cover the entire range between $\text{SiO}_2 = \sim 44$ and ~64 wt.% (Leeman *et al.*, 1976). The value of $F = 0.49$ closely corresponds to the modeled compositional range for the majority of COM zircons that crystallized at $F > 0.5$, and only a small portion of zircon also records a crystal content of up to 60% ($F = 0.4$) that is not captured by bulk lava compositions (Fig. 11).

Zircon as a recorder of crustal assimilation in ESRP and COM parental magmas

Different AFC models have been explored to explain the isotopic (e.g. Sr, Nd, Pb, Th) and compositional covariation of igneous rocks in the ESRP (e.g. Leeman & Manton, 1971; Leeman *et al.*, 1976; Stout *et al.*, 1994; Reid, 1995; McCurry *et al.*, 2008; Putirka *et al.*, 2009; Chadwick *et al.*, 2023). As an outcome of these models, a similar primitive magmatic parent has been invoked for ESRP and COM (Putirka *et al.*, 2009), whereas crustal assimilation produced different isotopic evolutionary trends. The abundance of Archean xenoliths in COM basaltic–intermediate lavas also

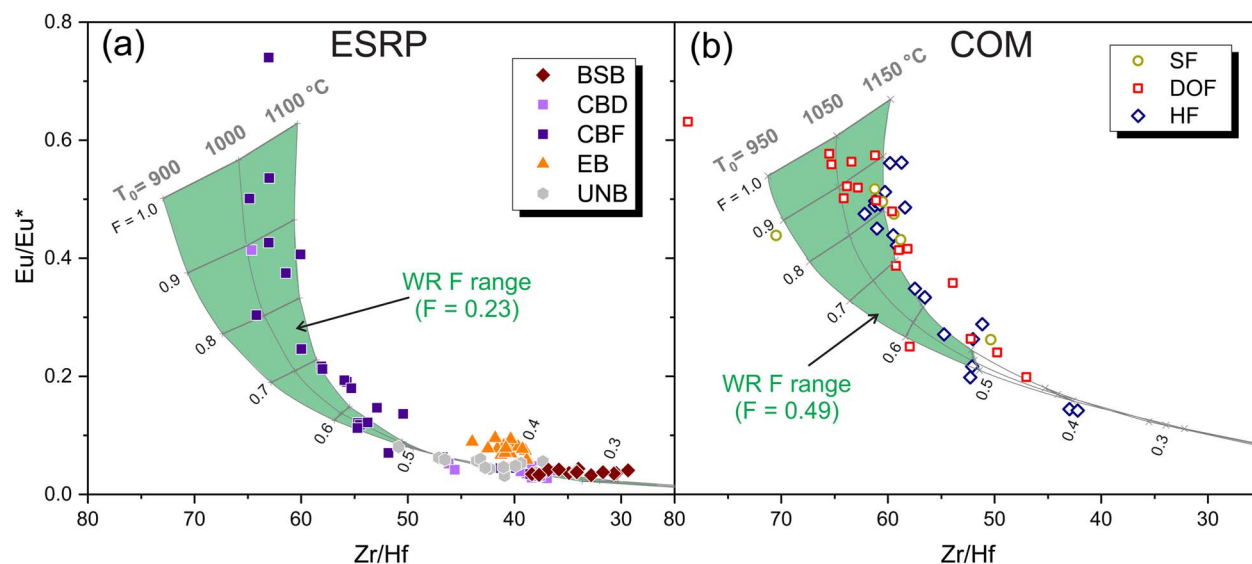


Fig. 11. Fractional crystallization models for zircon Eu/Eu^* vs. Zr/Hf compositions for ESRP (a) and COM lavas (b; see text). Shaded (green) fields represent the range of F values (fraction of remaining liquid) of the whole rock-based models from the literature ($F = 0.23$ for ESRP; [McCurry et al., 2008](#); $F = 0.49$ for COM; [Leeman et al., 1976](#)). Unit abbreviations from [Table 1](#).

implies a strong role for assimilation of such a crustal component for COM magmas, and likely less so for ESRP.

Zircon isotopic ratios ($\delta^{18}\text{O}$ and ϵ_{Hf}) are invariant with indicators for temperature and/or differentiation (e.g. Ti/Zr , Eu/Eu^* , and Zr/Hf) ([Fig. 8](#)). This implies that crustal assimilation, which clearly shifted host magma compositions relative to uncontaminated mantle values, predates zircon crystallization. Furthermore, trace element as well as O and Hf isotopic differences also exist between ESRP and COM zircon ([Figs 9 and 10](#)), and we, therefore, explore AFC models independently for both magma systems ([Table 3](#)). For our AFC modelling ([Table 3](#)), we use equations from [DePaolo \(1981\)](#) to reproduce the Hf and O isotopic trends in zircon, acknowledging that this approach ignores energy conservation constraints as well as recharge, and may only incompletely capture complex multi-stage wallrock assimilation upon magma ascent (e.g. [Huppert et al., 1985](#); [Bohrson & Spera, 2001](#); [Spera & Bohrson, 2001](#)). Nonetheless, this approach has reasonably reproduced the observed whole rock data trends for COM, thus facilitating a direct comparison between published whole rock and new zircon-based AFC models. For the same reason, we also adopted AFC parameters (e.g. rate of assimilation over rate of crystallization $r = 0.8$) from previous models ([Putirka et al., 2009](#)). We acknowledge that zircon only covers the last stages of a protracted differentiation path, and hence only retroactively records the previous isotopic evolution of the mafic parent magma.

For the ESRP, assimilation of rocks with sub-mantle $\delta^{18}\text{O}$ is evident ([Table 3](#); [Fig. 12a](#)). Suitable low- $\delta^{18}\text{O}$ rocks are present within the Yellowstone hotspot track (e.g. in the Picabo volcanic field; [Drew et al., 2013](#)), and as regional Eocene volcanic rocks and their subvolcanic intrusive equivalents of the Challis group ([Kuntz et al., 1986](#); [Criss et al., 1991](#)). Assimilation of Archean crust has been suggested to play a minimal role ($<7\%$) for ESRP rhyolite magmas ([McCurry et al., 2008](#)), which is supported by Hf-isotopes in ESRP zircon, which would allow only for assimilation of 0.3% of Archean rocks with $\epsilon_{\text{Hf}} = -44$ ([Vervoort & Patchett, 1996](#)). Assimilation of plutonic equivalents of altered Picabo ignimbrites (e.g. West Pocatello rhyolite) and Challis altered rocks would match the observed $\delta^{18}\text{O}$ and ϵ_{Hf} values in ESRP zircon after $\sim 26\%$ and $\sim 8\%$ of fractional crystallization with concomitant assimilation

([Fig. 12a](#)). These comparatively low values of F (i.e. 0.74 and 0.92) require that the ESRP magmas continued to fractionate significantly after their isotopic composition was fixed during earlier AFC stages ([Fig. 12a](#)). This agrees with invariant zircon $\delta^{18}\text{O}$ and ϵ_{Hf} within each of the ESRP units, independent of the apparent amount of assimilation experienced by their assumed parental magmas.

In contrast to ESRP zircon with sub-mantle $\delta^{18}\text{O}$ compositions, COM evolved lavas contain zircon with $\delta^{18}\text{O}$ slightly above mantle values ([Fig. 10b](#)). The lower ϵ_{Hf} coupled with higher $\delta^{18}\text{O}$ in COM zircon relative to ESRP requires a two-stage AFC evolution ([Putirka et al., 2009](#)): the first stage involves assimilation of $\sim 1\%$ of Archean crust, which is followed by a second stage which mainly increases $\delta^{18}\text{O}$ while maintaining nearly constant ϵ_{Hf} ([Fig. 12b](#)). Possible candidates for the second stage of assimilation include unaltered intracaldera wallrocks from the Tertiary hotspot (e.g. similar to Tuff of Arbon Valley from the Picabo volcanic field; [Drew et al., 2013](#)) and late Cenozoic felsic plutons from the Idaho Batholith ([King et al., 2007](#); [Gaschnig et al., 2011](#)). Corresponding F values of 0.81 and 0.86 are again larger than the overall range in F recorded in primitive to intermediate COM lava compositions (0.3; e.g. [Leeman et al., 1976](#)) ([Fig. 12](#)). Moreover, as with ESRP lavas, zircon $\delta^{18}\text{O}$ and ϵ_{Hf} values are homogeneous in the COM lavas, indicating that AFC must predate the magmatic differentiation recorded by zircon ([Fig. 10](#)). Comparatively high degrees of oxidation that are recorded in low-T zircon ([Fig. 9](#)), however, indicate that oxidized fluids of surface origin may have influenced the wallrocks of the shallow magma reservoirs where differentiation occurred (e.g. [Colon et al., 2018](#)).

Magma evolution in the ESRP and COM volcanic fields from the zircon perspective

Magmatic evolution in the ESRP is the product of a complex differentiation history that ultimately started with an inferred picrite parent ($\text{MgO} > 10\%$; [McCurry et al., 2008](#); [Putirka et al., 2009](#)). Ascending from the mantle, these parental magmas ponded in the lower crust (>20 km) until fractional crystallization to MgO concentrations of <10 wt.% caused their density to decrease, facilitating further ascent within the crust ([Putirka et al., 2009](#)).

Table 3: Summary of modelling parameters for assimilation coupled with fractional crystallization (AFC)

D r	1 0.8 ^a		0.1 0.8 ^a	
	$\delta^{18}\text{O}$	O (wt.%)	ϵHf	Hf ($\mu\text{g/g}$)
Mafic parent (MP)	+5.71 ^b	50	+0.6 ^c	3 ^b
Crustal endmembers				
Archean crust (AC)	+7.4 ^d	50	-44 ^e	9 ^e
Challis volcanics (C)	-3.6 ^f	50	-12.5 ^g	9 ^g
West Pocatello rhyolite (WP)	+1.41 ^h	50	-7.5 ^h	6.5 ⁱ
Tuff of Arbon Valley (P)	+7.9 ^h	50	-28 ^h	1.4 ^h
Idaho Batholith (IB)	+9.06 ^j	50	-18.6 ^k	2.9 ^k

^a Putirka et al. (2009). ^b Hart et al. (1999). ^c average Stefano et al. (2019). ^d Watts et al. (2011). ^e Vervoort and Patchett (1996). ^f Criss et al. (1991). ^g Gaschnig et al. (2011). ^h Drew et al. (2013). ⁱ For felsic volcanics from Condie (1993). ^j Atlanta lobe average from Gaschnig et al. (2011). ^k Atlanta lobe average from King et al. (2007).

In addition to effusive eruption of the resulting olivine–tholeiites in discrete rift zones, these magmas also formed an extensive sill complex (e.g. Sparlin et al., 1982; Peng & Humphreys, 1998) aided by low density silicic intrusions and their Yellowstone-type eruptive equivalents that predominate along the axis of the ESRP (Fig. 13). At the ESRP margins, preexisting Archean crust is preferentially preserved and, therefore, exerts a stronger influence on COM magmas relative to those erupted along the axis of the downwarp, where the crust was strongly reworked during passing of the Yellowstone hotspot (McCurry et al., 2008; Putirka et al., 2009). While these magmas ponded in the middle to upper crust, strong Fe-enrichment and a concomitant increase in density hindered eruption until they became sufficiently evolved (Christiansen & McCurry, 2008), possibly further aided by assimilation of mid-crustal felsic rocks (e.g. Putirka et al., 2009; Chadwick et al., 2023). Ultimately, evolved magmas comprising basaltic trachyandesite–rhyolite formed as the end products of differentiation in shallow reservoirs where assimilation was largely inhibited (McCurry et al., 2008).

Zircon $\delta^{18}\text{O}$ and ϵHf isotopes are consistent with two independent branches of differentiation from a similar mantle source, where assimilation of Archean crust is negligible for the ESRP, but required to explain COM isotopic compositions (Fig. 12). Distinct $\delta^{18}\text{O}$ trends also point to differential assimilation of mid–upper crustal rocks: ESRP parental magmas assimilated hydrothermally altered low- $\delta^{18}\text{O}$ rocks prevailing along the axis of the Yellowstone track (Fig. 12a), whereas for COM only high- $\delta^{18}\text{O}$ crustal mid-crustal endmembers are detected. These could either represent unaltered intrusions at the margins of the Yellowstone hotspot track, or preexisting early Cenozoic plutons (Fig. 12). Isotopic clustering of zircon from individual domes and lava flows without correlation between isotopic values and host magma composition indicates that zircon crystallization postdates these different histories of crustal assimilation (Fig. 13). Assimilation and fractional crystallization of the ESRP and COM parental magmas led to progressive enrichment of Zr (Fig. 2), but at the stage of reaching zircon saturation, assimilation within the upper crust had largely ceased.

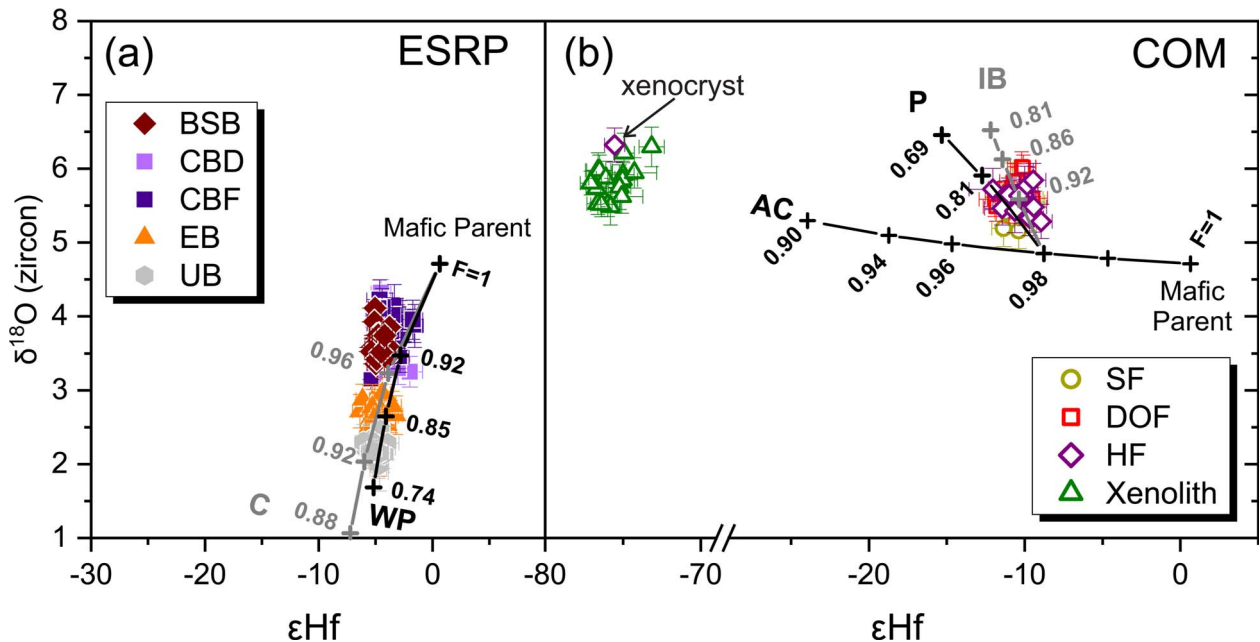


Fig. 12. Covariation diagram of O vs. Hf isotopes with AFC trends modeled for a common parental magma ($\delta^{18}\text{O} = +5.71$, $\epsilon\text{Hf} = +0.6$) and different crustal endmembers. Numbers indicate the fraction of remaining melt (F). Crustal endmembers for ESRP (a) are Challis group (C) and hydrothermally altered West Pocatello rhyolite (PR). For COM (b), two-stage AFC initially uses an Archean crust (AC) endmember, and subsequently unaltered Picabo caldera rhyolite (P) and Idaho Batholith (IB) granite as endmembers. For both systems, an $r = 0.8$ value (ratio of assimilation to fractionation; Table 2) was used based on Putirka et al. (2009). Unit abbreviations from Table 1.

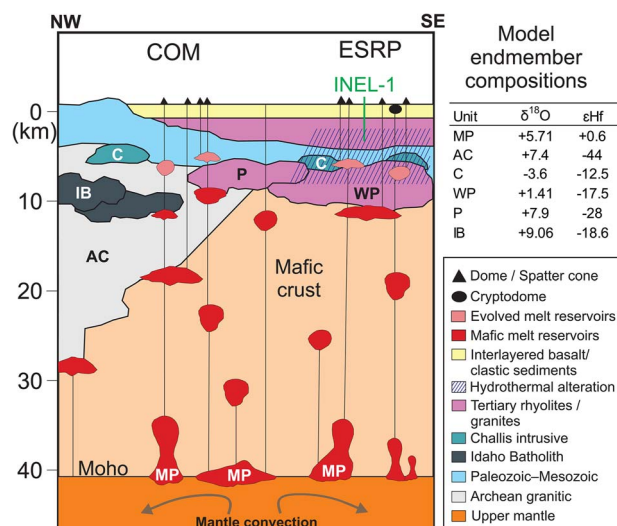


Fig. 13. Simplified conceptual model for COM and ESRP Quaternary magma genesis, modified after [McCurry et al. \(2008\)](#). Picrite or high-Mg olivine tholeiite magmas generated at the Moho (MP) experienced extensive fractional crystallization during ascent through the crust. For COM magmas, this ascent was accompanied by mid-crustal assimilation of an Archean component (AC), followed by additional minor assimilation of rocks from either the Idaho Batholith (IB) or unaltered Tertiary intracaldera rocks (P) at the margins of the hot spot track. In contrast, ESRP magmas assimilated significant amounts of altered Challis volcanic (C) rocks, Tertiary intracaldera rocks, or their plutonic equivalents at mid to upper crustal levels (WP). Zircon crystallized during subsequent magma residence which lasted for few 1000 years in the case of the COM evolved lavas, and not more than c. 100 kyrs for ESRP rhyolites.

For the ESRP domes, we observe a systematic relationship between the zircon $\delta^{18}\text{O}$ isotopic composition and geographical location, as well as their age: the eastern domes (East Butte; Unnamed Butte) are older and have lower $\delta^{18}\text{O}$, whereas the western domes (Big Southern Butte; Cedar Butte) are younger and have higher $\delta^{18}\text{O}$ (Table 1). Although the eastern domes overlie the inferred Arbon Valley Tuff caldera margin (e.g. [Anders et al., 2019](#)), where hydrothermal alteration is reasonably expected to be intense, their zircon cargo displays comparatively high $\delta^{18}\text{O}$. Magma residence durations as derived from the difference between crystallization and eruption ages (Δt) are more protracted for the eastern domes, and briefer for the western domes. This may indicate that the magma system producing the eastern domes was potentially larger and longer-lived, resulting in more efficient crustal assimilation compared to the source magma for the western domes. The most evolved lava flows at COM erupted after a hiatus of c. 1 ka at the beginning of eruptive period A ([Kuntz et al., 1986](#)). Period A magmas formed in two stratified but physically interconnected bodies where the younger and more mafic magmas from below triggered evacuation of the shallower, more evolved body ([Chadwick et al., 2023](#)). The brief zircon residence time of c. 4.3 ka is consistent with the evolved period A magmas being leftovers from earlier eruptive periods, indicating magma differentiation within several millennia at most.

Zircon in basalt-dominated magma systems

High-temperature zircon crystallization for COM and ESRP basaltic trachyandesite and trachyandesite magmas near their

respective zircon saturation temperatures of $\sim 900^\circ\text{C}$ was facilitated by the high Zr contents of their host melts (Figs 2b and 8). The strong enrichment of Zr in these magmas was the consequence of multiple fractional crystallization episodes that for the ESRP and COM systems reached extreme levels ($857 \mu\text{g/g}$ at 44.4 wt.% SiO_2) compared to a global compilation of LIP volcanic rocks (median $148 \mu\text{g/g}$ at 50.0 wt.%; Fig. 2b). Even the most primitive COM lavas have comparatively low MgO contents (4.9 wt.%; [Leeman et al., 1976](#)), which requires prior magmatic differentiation. The high Zr melt abundances that trigger zircon saturation are then produced by further fractional crystallization, which in the case of COM are estimated at $F = 0.3$ (e.g. [Leeman et al., 1976](#)). Furthermore, zircon crystallization at high temperatures can be caused by local saturation in crystal boundary layers ([Bacon, 1989](#)) or confined pores after extensive solidification and fractionation of initially low-Zr mafic magmas ([Bea et al., 2022](#)). Local saturation in COM magmas is texturally evident by zircon being commonly associated with olivine, clinopyroxene, ilmenite, and pyrite, sometimes in confined melt pockets (Fig. 3). However, zircon is also present in the matrix (Fig. 3a), and its composition is consistent with crystallization of an evolving, high-temperature melt as indicated by the fractional crystallization model involving the same mineral phases (olivine, plagioclase, clinopyroxene, magnetite, and apatite) as observed in the COM lavas. Typical magmatic oscillatory CL zonation and absence of inherited cores further support this interpretation (Fig. 4). The consistent range of F values recorded by zircon and whole-rock compositions argues against zircon recycling from a crystal-rich mush, although a small proportion of zircon in COM lavas indicates slightly more differentiated and cooler source melts than represented by the bulk rock (Fig. 11). This is consistent with a layered architecture proposed for the COM magma system, where shallow, evolved melt strata were flushed by the arrival of more mafic magma from below ([Chadwick et al., 2023](#)).

Despite the unequivocal autocrystic nature of most ESRP and COM zircons, and the clear correlation between zircon presence in lavas and the compositional turnover from Zr enrichment to depletion in the whole rock data, the comparison between Fig. 2a and b implies that this is a globally rare phenomenon because high-Zr intermediate and rhyolitic magmas are exceedingly scarce in LIPs. Nonetheless, the ESRP-COM systems serve as valuable testbeds for constraining the conditions under which zircon saturation can occur in mafic magmas typical for LIPs. The scarcity of low-MgO, high-FeO, and high-Zr magmas in the terrestrial volcanic record could imply that they often remained unerupted and solidified at depth. This notion is supported by frequent cryptodome formation in the ESRP ([McCurry et al., 2008](#)) as well as evidence for eruptions of evolved magmas requiring external triggering such as mobilization due to mafic recharge ([Chadwick et al., 2023](#)). Another important observation is that trace elements and isotopic compositions of ESRP and COM zircons indicate a lack of contamination during zircon crystallization, albeit their host magmas experienced significant crustal contamination during their earlier stages of differentiation. This contrasts with localized contamination of mafic sills intruded into basin sediments, which formed zircon that is isotopically heterogeneous ([Davies et al., 2021](#); [Gaynor et al., 2022](#); [Gaynor et al., 2023](#)), whereas ESRP and COM zircons are homogeneous within each unit. This comparison underscores that a petrochronological approach integrating age information with trace element and isotopic constraints is essential for resolving autocrystic vs. xenocrystic zircon origins in LIP environments.

CONCLUSIONS

The study of intermediate-silicic lavas from the ESRP and COM volcanic fields unravels how magmatic processes imprint on zircon, and how zircon compositions reflect magma evolution in basalt-dominated systems. Autocrystic zircon crystallization is indicated by U–Pb and U–Th dating that closely aligns with the Middle–Late Pleistocene and Holocene eruption ages for ESRP and COM lavas, respectively, with only rare xenocrystic zircon detected. The progressive enrichment in melt Zr due to extensive fractional crystallization is a prerequisite for zircon formation in these systems. Once zircon saturation is achieved in both ESRP and COM melts, their trace element composition parallels magmatic differentiation via fractional crystallization where ~40% and ~25% residual melts represent the most evolved lavas in the respective fields. ESRP and COM zircons also record different pathways of crustal interaction of a common mantle-derived parental magma. For the ESRP, assimilation of Archean crust is negligible, but low- $\delta^{18}\text{O}$ rocks generated by hydrothermal alteration of felsic volcanics of Yellowstone-type calderas were likely assimilated to a significant degree. By contrast, COM parental magmas clearly interacted with Archean crust, and in a second stage with unaltered felsic crustal components at the margins of the Yellowstone hotspot track.

The ESRP–COM systems serve as a unique analog for zircon behavior in LIP settings where basaltic magmas predominate. The results for the ESRP–COM zircons indicate that Zr enrichment sufficient to saturate zircon at high temperatures requires extensive fractional crystallization of mantle-derived magmas in a multi-level lithosphere-scale architecture. While magma conditions amenable to crystallize high-T zircon might be reached in other LIPs, the overall scarcity of mafic–intermediate high-Zr volcanic rock compositions in the global data base (comprising data from the North American Igneous Province, Siberian Traps, Deccan, Central Atlantic Magmatic Province, Emeishan, Karoo and Ferrar, Paraná, and Etendeka LIPs; DIGIS Team, 2021) suggests special circumstances for the formation or eruption of such melts. Zircon in mafic volcanic rocks from ancient LIPs thus should be carefully vetted against potential xenocrystic or contamination origins.

SUPPLEMENTARY DATA

Supplementary data are available at *Journal of Petrology* online.

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DATA AVAILABILITY

Analytical data for this study are available in the online supplementary material, and in EarthChem Library, at <https://doi.org/10.60520/IEDA/113491>.

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