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Erupted cumulate fragments in rhyolites from Lipari (Aeolian Islands)

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Contributions to Mineralogy and Petrology

Erupted cumulate fragments in rhyolites from Lipari (Aeolian Islands)

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Abstract:	<p>Over the last ~ 267 ky, the island of Lipari has erupted magmas with a range of compositions from basaltic andesites to rhyolites, with a notable compositional gap in the dacite field. Bulk geochemical and isotopic compositions of the volcanic succession, in conjunction with major and trace elemental compositions of minerals, indicate that the rhyolites were dominantly generated via crystal fractionation processes, with subordinate assimilation. Radiogenic (Sr, Nd, and Pb) and stable (O) isotopes independently suggest $\leq 30\%$ of crustal contamination with the majority of it occurring in mafic compositions, likely relatively deep in the system. Within the rhyolites, crystal-rich, K₂O-rich enclaves are commonly found. In contrast to previous interpretations, we suggest that these enclaves represent partial melting, remobilization and eruption of cumulate fragments left-over from rhyolite melt extraction. Cumulate melting and remobilization is supported by the presence of (1) resorbed, low-temperature minerals (biotite and sanidine), providing the potassic signature to these clasts, (2) reacted Fo-rich olivine, marking the presence of mafic recharge, (3) An₄₈₋₂₃ plagioclase, filling the gap in feldspar composition between the andesites and the rhyolites and (4) strong enrichment in Sr and Ba in plagioclase and sanidine, suggesting crystallization from a locally enriched melt. Based on Sr-melt partitioning, the high-Sr plagioclase would require ~ 2300 ppm Sr in the melt, a value</p>	

far in excess of Sr contents in Lipari and Vulcano magmas (50-1532 ppm) but consistent with melting of a feldspar-rich cumulate. Due to the presence of similar crystal-rich enclaves within the rhyolites from Vulcano, we propose that the eruption of remobilized cumulates associated with high-SiO₂ rhyolites may be a common process at the Aeolian volcanoes, as already attested for a variety of volcanic systems around the world.

Erupted cumulate fragments in rhyolites from Lipari (Aeolian Islands)

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Abstract

Over the last ~ 267 ky, the island of Lipari has erupted magmas with a range of compositions from basaltic andesites to rhyolites, with a notable compositional gap in the dacite field. Bulk geochemical and isotopic compositions of the volcanic succession, in conjunction with major and trace elemental compositions of minerals, indicate that the rhyolites were dominantly generated via crystal fractionation processes, with subordinate assimilation. Radiogenic (Sr, Nd, and Pb) and stable (O) isotopes independently suggest $\leq 30\%$ of crustal contamination with the majority of it occurring in mafic compositions, likely relatively deep in the system. Within the rhyolites, crystal-rich, K₂O-rich enclaves are commonly found. In contrast to previous interpretations, we suggest that these enclaves represent partial melting, remobilization and eruption of cumulate fragments left-over from rhyolite melt extraction. Cumulate melting and remobilization is supported by the presence of (1) resorbed, low-temperature minerals (biotite and sanidine), providing the potassic signature to these clasts, (2) reacted Fo-rich olivine, marking the presence of mafic recharge, (3) An₄₈₋₂₃ plagioclase, filling the gap in feldspar composition between the andesites and the rhyolites and (4) strong enrichment in Sr and Ba in plagioclase and sanidine, suggesting crystallization from a locally enriched melt. Based on Sr-melt partitioning, the high-Sr plagioclase would require ~ 2300 ppm Sr in the melt, a value far in excess of Sr contents in Lipari and Vulcano magmas (50-1532 ppm) but consistent with melting of a feldspar-rich cumulate. Due to the presence of similar crystal-rich enclaves within the rhyolites from Vulcano, we propose that the eruption of remobilized cumulates associated with high-SiO₂ rhyolites may be a common process at the Aeolian volcanoes, as already attested for a variety of volcanic systems around the world.

Keywords: Aeolian islands; Assimilation-Fractional Crystallization; cumulates; Lipari; rhyolite

Introduction

The genesis of intermediate to silicic magmas has been a topic of debate in igneous petrology for more than a century (Bunsen 1851; Harker 1909; Daly 1925; Bowen 1928). Broadly, the discussion revolves around whether evolved magmas are generated predominantly via processes associated with melting pre-existing lithologies (e.g., Price et al. 2005; Smith et al. 2006; Christiansen and McCurry 2008; Schmitt and Hulen 2008; Tamura et al. 2009; Simon et al. 2014) or whether they are predominantly made via fractionation with limited degrees of assimilation (e.g., Bacon and Druitt 1988; Halliday et al. 1991; Geist et al. 1995; Hildreth and Fierstein 2000; Lindsay et al. 2001; Haase et al. 2006; Peccerillo et al. 2007; Bachmann and Bergantz 2008). Volcanic suites with strongly bimodal erupted compositions ('Daly gaps') remain a key location for evaluating between fractionation-dominated and crustal melting-dominated scenarios (Brophy 1991; Francalanci et al. 1995; Thompson et al. 2001; Peccerillo et al. 2003; Szymanowski et al. 2015). Proponents of crustal melting interpret the lack of intermediate compositions to indicate that the evolved compositions are not generated via intermediates (e.g., Chayes 1963; Bohron and Reid 1997; Trua et al. 1999; Wiesmaier et al. 2012). Alternatively, recent models coupling fluid dynamics, geochemistry and thermal constraints indicate that compositional gaps may be generated via efficient crystal-liquid separation at relatively high crystal fractions (Brophy 1991; Mushkin et al. 2002; Dufek and Bachmann 2010). In such a scenario, highly evolved and crystal-poor caps can be produced by melt extraction from crystal-rich mushes of intermediate composition, which represent long-lived, thermally buffered 'rhyolite nurseries' (Bachmann and Bergantz 2004; Hildreth 2004; Cooper and Wilson 2014). Finding cumulate lithologies with a "silicic" signature would strongly support the claim that melt extraction from shallow evolved mushes is commonplace. In plutonic lithologies, such cumulate signatures are not rare (e.g., Bachl et al. 2001; Deering and Bachmann 2010; Gelman et al. 2014; Lee and Morton 2015), but it is clearly more challenging to find them in volcanic units, in which crystal-rich lithologies should not be eruptible, although some have been described from large ignimbrites (e.g., Bacon and Druitt 1988; Deering 2009; Deering et al. 2011b; Bachmann et al. 2014; Ellis et al. 2014).

In this context, the volcanic succession of the island of Lipari, characterized by the occurrence of both calc-alkaline rhyolites with crystal-rich potassic enclaves and cordierite-bearing dacites with a clear crustal isotopic signature, represents a key location to investigate the contribution of the different processes involved in the genesis of silicic magmas. In this paper, we combine information from a variety of techniques including petrography, bulk rock major, trace and isotopic data with mineral compositions and geochemical models, in order to fully characterize silicic magmas on Lipari, including the remarkable crystal-rich enclaves that occur in several of the crystal-poor rhyolitic units.

Geological background and magmatic history of Lipari

The Aeolian archipelago (active since ~1.3 Ma) is a volcanic arc constructed upon the ~ 15-25 km-thick continental crust of the Calabro-Peloritano basement and located in the Southern Tyrrhenian Sea at the boundary between the African and the Eurasian plates (Pirromallo and Morelli 2003). The seven islands and several seamounts comprising the arc exhibit compositions spanning from arc tholeiites (in the seamounts) to potassic rocks. Calc-alkaline (CA) and high-K calc-alkaline (HKCA) rocks are the most abundant throughout the arc, and dominate the volcanic succession at Alicudi, Filicudi, Salina, Lipari and Panarea. Shoshonites are spatially restricted to Vulcano and Stromboli (and subordinately to Panarea), where potassic rocks characterize the mature stages of volcanic activity (Peccerillo et al. 2013 and references therein). High-SiO₂ rhyolitic magmas (SiO₂>70%) are only present at Salina, Lipari, Vulcano and subordinately at Panarea.

Lipari, the largest of the Aeolian Islands, together with Salina and Vulcano, forms a NNW-SSE-oriented volcanic belt developed along the Tindari-Letojanni strike-slip fault system (Mazzuoli et al. 1995; Ventura et al. 1999). This regional-scale structure and the associated N-S and E-W extensional faults, are deemed to have strongly influenced volcanic activity in this sector of the archipelago (e.g. De Astis et al. 2003; Peccerillo et al. 2013) potentially favouring the formation of mid-crustal reservoirs conducive to rhyolite generation (Gioncada et al. 2003). Subaerial Lipari has grown since ~267 ka through nine successive eruptive epochs (=EE), interrupted by periods of volcanic quiescence (Fig. 1) (see Forni et al. 2013 for a full description).

Pyroclastic rocks and lavas on Lipari vary from CA and HKCA basaltic andesites to rhyolites with a notable compositional gap between the dacites and the rhyolites (Fig. 2). CA and HKCA basaltic andesites (with subordinate andesites and dacites) were dominant through the EE 1-4 (267-114 ka) when volcanic activity was linked to a series of monogenic and polygenic volcanic centers mainly located in the western and central sectors of the island (Figs. 1 and 2). HKCA andesites and minor dacites (including the cordierite-bearing lavas=CBL), related to the activity of M. S. Angelo and M. Chirica stratocones, became ubiquitous during the EE 5-6 (105-81 ka) (Figs. 1 and 2). Rhyolitic products were erupted during the EE 7-9 (from 43 ka to AD 1220), when volcanic activity resumed after a 40 ky-long period of quiescence (Figs. 1 and 2). HKCA andesitic to dacitic grey and banded pumices occur within the M. Guardia pyroclastic sequence (EE 8) (De Rosa et al. 2003). Latitic, trachytic and HKCA andesitic crystal-rich enclaves have been frequently recognized within the rhyolitic lavas of EE 7-9 (Gioncada et al. 2003; Gioncada et al. 2005; Davì et al. 2010; Forni et al. 2013) (Fig. 2). Notably, bulk-rock grey and banded pumices and crystal-rich enclaves-bearing lavas must be considered as mixed/mingled samples rather than pure compositions (Fig. 2).

CA and HKCA basaltic andesites and HKCA andesites (EE 1-4 and 6) contain plagioclase, clinopyroxene and minor olivine, with orthopyroxene, Ti-magnetite and apatite only occurring in the andesitic rocks. CBL (EE 5) are highly porphyritic rocks (up to 50% crystals including plagioclase,

clinopyroxene, orthopyroxene, cordierite, garnet and minor ilmenite, andalusite, spinel, biotite and sillimanite along with a large quantity of metapelitic and gabbroic xenoliths; Barker 1987), being interpreted as deriving from mixing between a basaltic-andesitic magma and a high-K peraluminous rhyolitic melt of anatectic origin (Di Martino et al. 2011). CA rhyolites (EE 7-9) instead are crystal-poor (<5% crystals) and contain K-feldspar, plagioclase and sporadic hornblende and biotite with zircon and apatite present as accessory minerals. Within several CA rhyolitic units two different types of dark-coloured porphyritic enclaves (~15-35% crystals) have been reported: (1) andesitic and dacitic to latitic enclaves (low-K enclaves) containing plagioclase, clinopyroxene, olivine, apatite and magnetite (Gioncada et al. 2003; Gioncada et al. 2005; Forni et al. 2013) and (2) latitic to trachytic enclaves (high-K enclaves) with a mineralogy of clinopyroxene, plagioclase, sanidine, olivine, biotite and magnetite with accessory apatite and zircon (Davì et al. 2009; Davì et al. 2010) (Fig.3). The pyroclastics of the M.Guardia sequence contain grey pumices with the same mineralogy as the enclaves in the lavas both as single clasts and within banded clasts with the typical crystal-poor white pumice (De Rosa et al. 2003).

Methods

Bulk rock and groundmass major and trace element compositions for the crystal-rich enclaves sampled within the Rocche Rosse and Pomiciazzo lava coulee (high-K enclaves; EE 9; Fig.1) were determined by XRF and ICPMS at ETH Zurich. A set of samples representative of the basaltic andesites to rhyolites and crystal-rich enclaves (EE 1-9) were analysed for mineral chemistry via field emission electron microprobe at the Centre for Experimental Mineralogy, Petrology and Geochemistry (Uppsala University, Sweden). The operating conditions were as follows: 10 nA beam current, 15 kV accelerating voltage, counting times of 10 s on peaks and 5 s on background positions. The beam size was focused (1 μm) for the analyses of pyroxene, olivine, biotite and oxides and enlarged to 5 μm for feldspars in order to minimize alkali migration.

Trace element concentrations in pyroxenes, plagioclases, biotite and olivine were determined by LA-ICPMS using a 193 nm Resonetics ArF excimer laser coupled to a Thermo Element XR ICP mass spectrometer at the Institute of Geochemistry and Petrology, ETH Zurich. Analytical conditions were identical to those described in Szymanowski et al. (2015) and data reduction employed the MATLAB-based program SILLS (Guillong et al. 2008) to determine concentrations from transient signals. USGS reference glass GSD-1G was used as secondary standard to monitor the accuracy of the instrument. The precision of a single spot analysis is difficult to quantify, but replicate analysis of a homogeneous mineral or glass gives precisions for element concentrations \gg LOD (limit of detection) better than 5% of the value.

Bulk rock Sr, Nd and Pb isotope ratios were determined at CNR-Istituto di Geoscienze e Georisorse (Pisa, Italy) by TIMS techniques using a Finnigan MAT 262 multicollector mass

spectrometer running in dynamic (Sr and Nd) and static (Pb) mode. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios have been normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. During the measurement period, the mean measured value of $^{87}\text{Sr}/^{86}\text{Sr}$ for the NIST–SRM 987 standard was 0.710242 ± 0.000013 (2SD, N = 25), while the mean $^{143}\text{Nd}/^{144}\text{Nd}$ for the La Jolla standard was 0.511847 ± 0.000008 (2SD, N = 25). The JNdi-1 standard (Tanaka et al. 2000) was also analyzed and yielded a $^{143}\text{Nd}/^{144}\text{Nd}$ value of 0.512100 ± 0.000010 (2SD, N = 25). The total procedure blanks, 0.5 ng and 0.1 ng of Sr and Nd, respectively, were negligible for the analyzed samples. Pb was extracted by chromatographic ion exchange in Dowex 1 anion resin, using standard HBr and HCl elution procedures. Replicate analyses of Pb standard NIST–SRM 981 indicated that Pb isotope ratios are accurate to within 0.025% (2SD) per mass unit, after applying mass discrimination corrections of $0.14 \pm 0.01\%$ per mass unit relative to the reference composition from Todt et al. (1993). The Pb blank varies from about 0.4 to 0.6 ng, and no blank correction was made.

Oxygen isotope compositions of mineral separates were obtained at CNR-Istituto di Geoscienze e Georisorse (Pisa, Italy) by laser fluorination technique (Sharp 1990) reacting the samples under F_2 gas atmosphere. Purified oxygen gas was directly transferred into a Finnigan Delta Plus Mass Spectrometer via 13A zeolite molecular sieve for isotopic ratio determinations (Sharp et al. 1995). All the data are given following the standard δ notation. During the course of analysis, an in-house laboratory standard was used (QMS quartz; $\delta^{18}\text{O} = 14.05\text{‰}$), yielding an average $\delta^{18}\text{O} = 14.08\text{‰}$, $\sigma = 0.14\text{‰}$. NBS28 standard ($\delta^{18}\text{O} = 9.60\text{‰}$) gave an average values of $\delta^{18}\text{O} = 9.50\text{‰}$, $\sigma = 0.18\text{‰}$. The full data set is provided in Online Resources 1-6.

Results

Bulk geochemistry

A complete set of bulk rock major and trace element data has been presented and discussed in Forni et al. (2013) and our study does not attempt to duplicate it. Rather, we have expanded the bulk geochemical dataset by focusing our analyses on the bulk and groundmass compositions of crystal-rich enclaves (Gioncada et al. 2003; Gioncada et al. 2005; Davì et al. 2009; Davì et al. 2010). Below we briefly summarise the salient points of Lipari bulk geochemistry.

- Major and trace element compositions (Fig. 4) describe a liquid line of descent with the early evolution dominated by fractionation of ferromagnesian phases (olivine and clinopyroxene) indicated by curved trace elemental patterns (e.g. MgO, CaO and Ni; Fig. 4 a, c and e). Crystallization in the andesites is dominated by plagioclase, orthopyroxene, spinel, and apatite driving compositions to higher SiO_2 with concurrent decreases in Sr, P and Ti (Fig. 4 f and d).

- Following a gap of ~ 40 ka, and leaping the dacitic compositional field, Lipari rhyolites show depletions in the Sr, Ba, Zr, Nb and Rb due to the crystallization of sanidine, zircon and biotite, consistent with their appearance in the rhyolites (Fig. 4 f-h). Normalized rare earth element (REE) patterns in the rhyolites reveal a marked negative Eu anomaly indicative of extensive plagioclase fractionation (Fig. 5).
- Crystal-rich enclaves present in the rhyolites span a range of compositions from HKCA andesites to trachyte (56-70 wt.% SiO₂; Fig. 2) and are all considerably more potassic than typical Lipari magmas (Gioncada et al. 2003; Gioncada et al. 2005; Davì et al. 2010). Additionally some of the high-K enclaves are notably enriched in P₂O₅ (Fig. 4d) and a variety of trace elements including Ba, Sr, Zr (Fig. 4 f-h) and Pb compared to typical Lipari volcanics.

Mineral chemistry

Feldspar

Euhedral to subhedral, normally or oscillatory zoned plagioclase phenocrysts are ubiquitous in samples of all compositions from Lipari but dominate in the andesites. Their compositions range from bytownite to labradorite in basaltic andesites (An₅₈₋₈₅) and andesites (An₄₈₋₈₃), whereas rhyolites only contain oligoclase (An₁₈₋₂₃) (Fig. 6a). Sanidine phenocrysts (Or₅₈₋₆₈) have been recognized in the high-SiO₂ andesites and rhyolites (Fig. 6a). Sr and Ba in feldspars increase with decreasing An content from the basaltic andesites to the andesites but drastically decrease in the rhyolites, indicating a switch from incompatible to compatible behavior.

The plagioclase within the crystal-rich enclaves covers a wide range from bytownite to oligoclase (An₂₁₋₇₇) including compositions of An₄₈₋₂₃ that are not represented in the other erupted rocks from Lipari (Fig. 6b). Two types of plagioclase have been recognized: type1 is andesine to oligoclase (An₂₁₋₃₈), low-Sr, (124-687 ppm) plagioclase forming single crystals or sanidine cores (anti-rapakivi structure) within the low-K and high-K enclaves respectively (Fig. 3a); type2 is bytownite to andesine (An₄₁₋₇₆), high-Sr (2816-4785 ppm) plagioclase generally forming euhedral crystals and containing cpx and melt inclusions (Fig. 3b).

K-feldspars (Or₅₅₋₆₇) are present as crystals in the high-K enclaves (Fig. 6b) and as microlites in the groundmass of the low-K enclaves (Gioncada et al. 2005; Davì et al. 2010). Ba content in the enclave sanidines (428-2394 ppm) increases towards crystal rims and is much higher than the one observed in the sanidines from the rhyolites (69-153 ppm).

Pyroxene

Euhedral to subhedral, complexly zoned clinopyroxene phenocrysts are variably present in the basaltic andesites and andesites, being augitic in composition (Fig. 7a). Pigeonitic clinopyroxene

and enstatitic orthopyroxene are abundant in the andesitic rocks (Fig. 7a) and are frequently overgrown by clinopyroxene rims (augite). The REE content in pyroxenes increases from the basaltic andesites to the andesites along with a progressively more negative Eu anomaly (Fig. 8).

The crystal-rich enclaves contain augitic to diopsidic clinopyroxene (Fig. 7b). Two types of crystals have been recognized: type1 is highly resorbed (Fig. 3c), augitic in composition and shows very deep Eu anomaly (Fig. 8); type2 is augitic to diopsidic, euhedral, oscillatory zoned and contains oxides and apatite inclusions (Fig. 3d). Type2 clinopyroxene has higher Eu/Eu* compared to type1 and is enriched in CaO, Al₂O₃, Sr and LREE/HREE compared to type1 and to the clinopyroxene from the basaltic andesites and andesites (Fig. 8).

Olivine

Olivine phenocrysts (Fo₇₇₋₇₉) are scarce in the basaltic andesites and andesites and frequently altered to iddingsite. In the crystal-rich enclaves olivines occur both as partially resorbed crystals very rich in apatite inclusions, frequently forming mineral clots together with apatite and oxides (type1; Fo₆₅₋₆₆; Ni=131-146 ppm; Fig. 3e) and euhedral crystals (type2; Fo₇₄₋₇₉; Ni=442-714 ppm).

Biotite

Biotite occurs as crystals (up to 0.6 mm) in the high-K enclaves and sporadically in the rhyolites. In the enclaves, biotite phenocrysts (Mg#=44-65) show normal zoning and embayed rims (Fig. 3f). Moreover they are enriched in a variety of trace elements such as Ba (up to 4866 ppm), Nb (up to 30 ppm) and Rb (up to 350 ppm). Tiny flakes of biotite have been recognized in the groundmass of the low-K enclaves.

Fe-Ti oxides

Ti-magnetite (Usp₂₆₋₅₀) occurs as microcrysts in the groundmass of basaltic andesites and as phenocrysts in the andesites, where oxides frequently form aggregates together with clinopyroxene and plagioclase. In the crystal-rich enclaves, Fe-Ti oxides (Usp₁₄₋₂₂) are abundant and occur both in the groundmass and as phenocrysts, often associated with, or included into clinopyroxene and olivine crystals (Fig. 3 c-e).

Crystallization temperatures

Plagioclase cores and bulk-rock compositions, as representative of the early stage of nucleation, have been used to estimate the crystallization temperatures using the plagioclase-liquid thermometer of Putirka (2008). The equilibrium condition between mineral and melt compositions has been determined by means of test for equilibrium based on plagioclase-liquid exchange reactions (Putirka 2008). These calculations reveal that plagioclase phenocrysts within the basaltic andesites, andesites and rhyolites are mostly in equilibrium with the bulk-rock (Fig. 9a). Conversely the

plagioclase contained into the low-K and high-K enclaves plots predominantly out of the calculated bulk-rock equilibrium range, towards lower values of $100 \cdot \text{Ca}/(\text{Ca}+\text{Na})$ (i.e. the plagioclase is more evolved than it should be based on the whole rock composition) (Fig. 9a). Plagioclase-liquid equilibrium pairs yield crystallization temperatures in the range of $\sim 1237\text{-}1077$ °C for the basaltic andesites, $\sim 1158\text{-}988$ °C for the andesites and $\sim 884\text{-}738$ °C for the rhyolites (Fig. 9b). Rhyolite temperatures are compatible with the temperature ranges obtained using a variety of different geothermometers (De Rosa et al. 2003; Gioncada et al. 2003; Poli et al. 2008) (Fig. 9b).

Isotope geochemistry

Radiogenic isotopes (Sr, Nd and Pb)

Sr and Nd isotope ratios of Lipari magmas range from 0.70442 to 0.70578 and from 0.51251 to 0.51271 respectively, partially overlapping with the isotopic data from the literature and the isotopic compositions of all the other Aeolian Islands, except for Alicudi, for which the most primitive isotopic compositions of the whole arc have been reported (Peccerillo et al. 2004) (Fig. 10a). The basaltic andesites of EE 2 cover almost the entire range of isotopic compositions. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{144}\text{Nd}/^{143}\text{Nd}$ ratios increase and decrease respectively with SiO_2 from the basaltic andesites to the andesites (EE 1-6), but remain fairly constant from the andesites to the rhyolites (EE 7-9) (Fig. 11). The highest Sr and lowest Nd isotopic ratios are reported for the CBL (EE 5) and one sample from EE 2, which contains granulitic xenoliths. Coherently, in the CBL the presence of cordierite and garnet, along with andalusite and considerable amounts of crustal xenoliths containing sillimanite, reflects a large crustal component involved in petrogenesis (Di Martino et al. 2011). Moreover, the frequent occurrence of metamorphic xenoliths (granulites, metapelites and quartzites) within the basaltic andesites and andesites from Lipari indicate that crustal assimilation was very efficient during the petrogenesis of basaltic andesites and andesites at lower and mid-crustal levels (25-15 km; Di Martino et al. 2010). Accordingly, the geophysical images are consistent with the crust beneath the Aeolian arc being composed of metapelites and granulites (Ventura et al. 1999; Peccerillo et al. 2006; Ventura 2013) akin to the Calabro-Peloritano basement rocks (Caggianelli et al. 1991). Moreover, the Sr and Nd isotopic composition of cordierite (Sr=0.71023 and Nd=0.51200), which is deemed to crystallize from the anatectic melt involved in the petrogenesis of the CBL (Di Martino et al. 2011), plots within the field of crustal rocks (Fig. 10a). Pb isotope ratios span narrow ranges ($^{206}\text{Pb}/^{204}\text{Pb}= 18.273\text{-}19.556$; $^{207}\text{Pb}/^{204}\text{Pb}=15.655\text{-}15.725$; $^{208}\text{Pb}/^{204}\text{Pb}=38.318\text{-}40.047$) and partially overlapping the fields of Vulcano and Salina.

Stable isotopes (O)

Oxygen isotopes from minerals on Lipari span a wide range in $\delta^{18}\text{O}$ from 6.2 ‰ for an olivine in a basaltic andesite to 14.4 ‰ for a garnet from one of the cordierite-bearing andesites (Fig.11a).

Oxygen isotopic compositions of euhedral clinopyroxene phenocrysts, which occur throughout the magmatic series on Lipari, vary from $\delta^{18}\text{O}$ of 6.3 to 7.9 ‰. Olivine phenocrysts within the basaltic andesites and andesites return $\delta^{18}\text{O}$ values of 6.2 and 6.8 ‰ indicating high temperature O isotopic equilibrium with the coexisting clinopyroxene ($\Delta_{\text{cpx-ol}}=0.5$ ‰ Valley et al. 2003). Cordierite and garnet within the CBL return very high $\delta^{18}\text{O}$ (13.4 ‰ and 14.4 ‰, respectively) indicative of the presence of a high- $\delta^{18}\text{O}$ crustal component in the region. Assuming a cpx-melt fractionation of 0.7 ‰ (diopside-andesitic melt at 1050°C; Valley et al. 2003), the calculated oxygen isotope composition of the equilibrium melts ($\delta^{18}\text{O}=7-8.6$ ‰) are higher than the ranges obtained by closed-system fractional crystallization of mantle-equilibrated melts and the calculated oxygen isotope compositions of equilibrium melts from Alicudi and Filicudi (Peccerillo et al. 2004; Santo and Peccerillo 2008) (Fig. 11a).

These observations, together with the weak correlation between $\delta^{18}\text{O}$ and bulk-rock SiO_2 confirm that crustal contamination processes were effective during early stages of magmatic evolution with little increase in assimilation at compositions more evolved than basaltic andesite (Fig. 11a). A positive correlation between oxygen isotope compositions and $^{87}\text{Sr}/^{86}\text{Sr}$ supports magma evolution dominated by AFC processes (= Fractional Crystallization coupled with crustal Assimilation; Francis et al. 1980; Taylor 1980).

Geochemical Modelling

In order to account for the generation of the whole compositional range observed for the Lipari magmas, geochemical models based on major element and isotopic compositions have been tested. The sample LIP 105 (EE2) has been chosen as parent in the models because of its primitive geochemical and isotopic signatures (Ni=68 ppm; Cr=170 ppm; $^{87}\text{Sr}/^{86}\text{Sr}=0.70442$; $^{143}\text{Nd}/^{144}\text{Nd}=0.51271$). According to the inferred nature of the basement rocks (Ventura et al. 1999; Peccerillo et al. 2006; Di Martino et al. 2011; Ventura 2013) and depth of magma storage (De Rosa et al. 2003; Di Martino et al. 2010), the average bulk-rock and isotopic composition of the metapelites and granulites from the Calabrian Basement (Caggianelli et al. 1991) has been considered as representative of the crustal contaminant. Least-square mass balance calculations have been carried out in two stages (step1, from basaltic andesites to andesites, Online Resource 7; step2, from andesites to rhyolites, Online Resource 8) using the OPTIMASBA code (Cabero et al. 2012) in order to account for the major changes in the crystallizing mineral assemblage occurring during magmatic evolution. Accordingly, average mineral and bulk-rock (i.e. parent and daughter magmas and contaminant) compositions have been used as input data in the model. The software calculates the relative proportions of crystallizing mineral phases producing the best fit between the simulated and the real derivative magma composition. During step1, crystallization of predominant Ca-rich plagioclase and pyroxenes with small amounts of olivine, Fe-Ti oxides and apatite, coupled with ~20% crustal

assimilation, drives the liquid from basaltic andesitic to andesitic compositions (Online Resource 7). During step2, plagioclase, clinopyroxene, biotite, Fe-Ti oxides, sanidine, olivine and apatite are the main crystallizing phases and smaller amounts of crustal contamination (~2%) are required to match the high-SiO₂ rhyolite daughter composition (Online Resource 8).

Further constraints can be placed on the degree of crustal assimilation required via AFC and mixing modelling using of the radiogenic and stable isotope data. The partition coefficients of Sr and Nd used in the AFC models (1.5 and 0.5 respectively) have been estimated according to Rollinson (1993) based on the relative proportions of crystallizing phases. AFC models show that the whole range of isotopic compositions displayed by the Lipari magmas can be obtained by fractional crystallization/assimilation ratios (R) between 0.1 and 0.3 (Fig. 10b). Unlike the Sr and Nd radiogenic isotope ratios, the O isotopic composition of the Calabrian Basement remains unknown. However, we can approximate it via use of the compositions of the CBL from EE 5. Both garnet-liquid and cordierite-liquid fractionation factors are poorly known, however given that globally siliclastic metamorphic rocks span the range of 4-21‰ (Bindeman 2008), using an average of the two measured values (13.4 ‰ and 14.4 ‰) seems a reasonable estimate. At high temperatures $\Delta(\text{Gt-zrc}) = 0.0 \pm 0.2$ ‰ (Valley et al. 1994) and zircon is typically a few per mil lower $\delta^{18}\text{O}$ than the co-existing melt (Bindeman 2008), so our inference of a crustal end member of 13.9 ‰ if anything has the potential assimilant 'too low' in $\delta^{18}\text{O}$ and thus errs on the side of over-estimating the proportion of assimilant required. Bearing this in mind, we can use mixing models to estimate how much of that component would be required to elevate the $\delta^{18}\text{O}$ of the magmas from a purely fractionation-derived value to the observed compositions (Fig. 11b). These simple bulk-mixing models confirm that between approximately 10 and 30% of the assumed crustal end-member would be required in petrogenesis. These values are in excellent agreement with those determined via mass balance modelling and radiogenic isotopic compositions above.

Discussion

Genesis of Lipari rhyolites

The occurrence of rhyolites during the most recent volcanism on Lipari (43 ka- AD 1220) and the presence of both a temporal and compositional gap prior to their appearance bear on how the rhyolites are made. A main role of crustal melting in the genesis of the rhyolites was invoked in the early studies of Crisci et al. (1991) and Esperanca et al. (1992), mainly supported by evidences of magma-wall rock interactions in some of the magmatic rocks older than 43-40 ka (in particular in the CBL; Barker 1987; Di Martino et al. 2011). However, in contrast to the CBL, the geochemical and isotopic signature of Lipari rhyolites strongly indicates that they were generated via fractional

crystallization of mantle-derived melts coupled with minor amounts of crustal contamination (Gioncada et al. 2003; Davì et al. 2010; this study). Based on the geochemical similarities between the rhyolites from Lipari and Vulcano and a similar influence of the regional tectonic setting on the volcanism of the two islands, Gioncada et al. (2003) and Davì et al. (2010) hypothesized that the two volcanic islands have shared the same plumbing system and magmatic evolution, at least during the last ~40 ka. In particular, the high-SiO₂ magmas on both islands have been deemed to derive via AFC from shoshonitic melts akin to the ones preserved as melt inclusions within some primitive olivine crystals at Vulcano (Gioncada et al. 1998; Davì et al. 2010). Such a scenario would be also supported by the finding of latitic enclaves within the rhyolites from Lipari, representing the intermediate products of this fractional crystallization path (Davì et al. 2009). Although we cannot exclude the existence of a connection between the plumbing systems of Lipari and Vulcano, such a connection appear unlikely for the shallow subvolcanic reservoirs (5-10 km depth range). Therefore we explore the possibility that the rhyolites from Lipari were derived, independently from those at Vulcano, via AFC from the intermediate compositions, with crystal-melt mechanics controlling the compositional gap (Dufek and Bachmann, 2010).

The rapid decrease of trace elements compatible into early crystallising phases (e.g., Ni, Cr; Fig. 4e) and the curved paths of elements which change the compatibility during differentiation (e.g. Sr, Ba, P as phases such as biotite, sanidine, plagioclase and apatite start crystallizing; Fig.4 d,f,g) strongly suggests that crystal fractionation is playing a key role throughout petrogenesis (Wade et al. 2005; Deering and Bachmann 2010; Lee and Bachmann 2014). This interpretation is supported by the marked Eu anomaly observed in the rhyolites (Fig. 5), which is indicative of extensive feldspar fractionation. Both the radiogenic and stable isotopes of Lipari volcanics indicate that assimilation of crustal material undoubtedly plays a role in petrogenesis especially during the evolution from the mafic to the intermediate compositions, where the proportion of crustal material involved is up to 30 %, but considerably decreases during the late stages, which led to the genesis of the rhyolitic melts (Fig.12). Mass balance calculations indicate that rhyolitic melts can be derived from the intermediate compositions via AFC processes after ~60% crystallization, coupled with ~2% crustal assimilation (Online Resource 8). Decreasing crustal contamination with ongoing magma evolution has been observed in a variety of magmatic suites (e.g., Cox and Devey 1987; Kerr et al. 1995; Peccerillo et al. 2004; Deering et al. 2011a) and is interpreted as the result of a less efficient magma-wall rocks interaction due to temperature drop and increase of magma viscosity as reservoirs are stored in relatively cold upper crust (see Depaolo 1981; Thompson et al. 2002; Dufek and Bergantz 2005).

Such a fractionation-dominated origin reflects a compositional evolution along a dominantly down-temperature trend (albeit with frequent periods of partial reheating following recharge into the mushy reservoirs) and yields relatively low-temperature silicic end-products. The suggestion that Lipari rhyolites have been “superheated” (Davì et al. 2010; Davì et al. 2011) appears untenable as magmatic temperature are relatively low, and crystals in equilibrium with the rhyolitic melt can be

found, albeit scarcely, within the Rocche Rosse lava flow (Gioncada et al. 2003; Gioncada et al. 2005; Davì et al. 2009; Davì et al. 2010). In particular, zircons can be found in Lipari rhyolites, and U/Pb dating of these crystals indicates that they grew within the rhyolitic melt in which they are hosted rather than being inherited from basement rocks (Poli et al. 2008). The presence of crystallising zircon within the Lipari rhyolites allows the use of zircon saturation thermometry of Watson and Harrison (1983) to be employed, which returns temperatures of 745-781 °C (n= 38). These typical rhyolitic temperatures are consistent with a number of other studies using other techniques on Lipari rhyolites, including two feldspar thermometry (760-770 °C, Gioncada et al. 2003), melt inclusion rehomogenisation (760 ± 10 °C, De Rosa et al. 2003), Ti-in-zircon thermometry (720-800 °C, Poli et al. 2008) and plagioclase-melt geothermometry (this study; Fig.9b). Additionally, the presence of biotite, a mineral with limited thermal stability, in some of the rhyolites, further suggests that temperatures were not unusually high.

Origin of crystal-rich enclaves on Lipari

The crystal-rich enclaves in the rhyolitic products of Lipari (EE 7-9) fall notably off the overall evolutionary trend described by Lipari magmas (Figs. 2 and 4). This geochemical characteristic was considered to be the result of the involvement of multiple magmas in the petrogenesis (Gioncada et al. 2003; Gioncada et al. 2005; Davì et al. 2009; Davì et al. 2010). In such a scenario, the enclaves were attributed to more alkaline magmas from the Vulcano suite, which mingled and co-erupted with magmas underneath Lipari and were co-erupted. Volumetrically, the proportion of enclaves within the rhyolitic host magma is small, typically less than 1% of the material erupted in the case of the Rocche Rosse lava flow (Davì et al. 2010). If the alkaline magma from Vulcano is present as a liquid in the upper crust beneath Lipari, the question of why more Vulcano-type magmas have not erupted from Lipari is hard to explain. This would also require that the subvolcanic reservoirs are linked, something unknown anywhere on Earth for such small arc volcanoes. Perhaps even more problematic for the Vulcano-Lipari connection is that while the Lipari crystal-rich enclaves compositionally resemble the alkaline magmas erupted on nearby Vulcano, a number of lines of evidence (including the presence of zircons in such alkaline magmas) suggest that their compositions do not reflect true magmatic liquids. In the following sections we explore the possibility that the crystal populations within the enclaves provide clues as to their origin.

Plagioclase

Simple partitioning calculations demonstrate that neither of the two types of crystals present in the enclaves (Fig. 3 a,b) are in equilibrium with the bulk-rock. Indeed, the low An (An₂₁₋₃₈) and Sr content (124-687 ppm) in type1 plagioclase are indicative of crystallization from a liquid

compositionally more evolved than the bulk-rock (Figs. 9a and 13). Additionally, textural features indicate that type1 plagioclase was partially resorbed and subsequently encountered a K₂O and Ba-rich melt from which the high-Ba sanidine rims grew (Fig. 3b). On the contrary, the euhedral type2 plagioclase (Sr= 2816-4785 ppm) rather crystallized from a Sr-rich melt (Figs. 3 and 13). The amount of Sr in the melt required to generate type2 plagioclases (836-2324 ppm), calculated using the experimentally-derived equation of Bindeman et al. (1998), is much higher than the actual Sr content in the bulk-rock enclaves (262-1116 ppm) and even in all the magmatic rocks from Vulcano (50-1532 ppm; Fig. 13). Additionally, the lack of correlation between Sr and An content suggests that crystallization of type2 plagioclase occurred under disequilibrium conditions (Blundy and Wood 1991).

Other minerals

- As for plagioclase, highly resorbed type1 clinopyroxene shows low Sr content (8-88 ppm, increasing towards the rim) and deep negative Eu anomaly (Eu/Eu*=0-0.1) indicative of crystallization from an evolved melt, which underwent intense feldspar fractionation (Fig. 8), while major and trace element composition of type2 euhedral clinopyroxene (Sr=94-278 ppm; Eu/Eu*=0.5-0.9) instead suggest that it crystallized from a Sr, CaO, Al₂O₃ and LREE- rich melt (Fig. 8).
- Similarly, type1 (Fo₆₅₋₆₆; Ni=131-146 ppm) and type2 (Fo₇₄₋₇₉; Ni=442-714 ppm) olivine crystals coexist within the crystal-rich enclaves. The occurrence of high-Fo euhedral olivine crystals suggests involvement of fresh mafic magma, whereas mineral aggregates containing low-Fo olivine (and abundant apatite, Fig. 3e) indicate accumulation of crystals under relatively reducing redox conditions and high H₂O content at relatively shallow depth (Berndt et al. 2005; Portnyagin et al. 2012). Removal of ~0.5% apatite during the fractionation step leading to the rhyolite formation (Online Resource 8) is enough to account for the significant drop of P₂O₅ in the bulk-rock (Fig. 4d).
- Biotite only occurs as individual crystals showing embayed rims (Fig. 3f) within the high-K enclaves, Trace element composition of biotite crystals is compatible with crystallization from a dacitic/rhyolitic melt, given the relatively high biotite/melt partition coefficients of Ba, Rb and Nb in these magmas (Rollinson 1993).
- Zircon crystals occur in some of the crystal-rich enclaves (Gioncada et al. 2005; Davì et al. 2009). Given the very low saturation temperature of zircon in K₂O-rich magmas (Watson and Harrison 1983), zircon crystallization is a strong argument against the crystal-rich enclaves representing clasts from alkaline melt.

The role of cumulates in rhyolite petrogenesis

The textural and compositional features of type1 crystals (plagioclase, clinopyroxene and olivine) and biotite indicate that they might represent the left-overs of the mineral phases which crystallized from an intermediate to evolved melt (i.e. dacite/rhyolite) during the fractionation step leading to the generation of the high-SiO₂ rhyolites. As discussed above, high-SiO₂ melts were produced via fractional crystallization, which requires a residue of crystals to accumulate somewhere in the crust. When the crystallinity of the system reaches a critical value (50-60% crystals; Bachmann and Bergantz 2004), efficient crystal-liquid separation might promote the formation of a crystal-poor cap of rhyolitic melt and residual cumulate showing specific geochemical signatures (Deering and Bachmann 2010).

In order to test the potential link between the crystal-rich enclaves and the crystal-poor rhyolites, we used trace element geochemical modelling based on the equations for the conservation of mass of a fractionating trace element (see Gelman et al. 2014 for a recent discussion on the derivation of these equations). Based on the variation diagrams of trace elements vs. SiO₂ (Fig. 4), we assumed that the starting compositions plot within the compositional gap (i.e. dacite) and used the modal proportions of fractionating phases derived from the mass balance calculations to estimate the bulk-rock partition coefficients. We tested the model using different starting compositions and a range of partition coefficients from Nash and Crecraft (1985) and Rollinson (1993) (see Online Resources 9). Additionally, according to the results of mass balance calculations and isotope geochemical modelling, small amounts of crustal contamination (i.e. 2%) were added to the system. When using the high partition coefficients (typical for silicic magmas), Sr, Ba and Rb concentrations predicted by the model for extracted liquids and cumulates match well the crystal-rich enclaves and rhyolites across a range of melt fractions (Fig. 14). Some of the scatter displayed by a few samples is induced by mixing/mingling textures (frequently observed in some of the rhyolites; De Rosa et al. 2003), and plot on straight-line trajectories (i.e. mixing lines) connecting the rhyolites and the crystal-rich enclaves compositions (Fig. 14).

The occurrence of euhedral crystals (type2 mineral phases) showing particularly enriched compositions and coexisting with type1 crystals within the enclaves, suggests the participation of locally enriched melts in petrogenesis. In particular, the presence of relatively primitive olivine crystals (type2), akin to ones from Lipari basaltic andesites, indicates crystallization from a mafic magma. Indeed, mafic recharge is required to provide the heat necessary to remobilize the cumulate material (which is likely uneruptible with crystallinities >50% when unmelted) by lowering its crystallinity via thermal remobilization (Deering et al. 2011b). This process, preferentially affecting the relatively low temperature crystallizing phases (feldspars and biotite), might have promoted the generation of melts (potentially interacting with the recharge magma) locally enriched in Sr, Ba, K₂O,

CaO, Al₂O₃ and LREE, from which type2 plagioclase and clinopyroxene and Ba-rich sanidines crystallized under disequilibrium conditions. Such a scenario would explain the coexistence within the crystal-rich enclaves of texturally and compositionally different mineral phases, showing clear evidences of disequilibrium. An alternative interpretation implies that all the type2 mineral phases may have been carried by the recharge magma rather than crystallizing from a mush-derived melt. Although we cannot exclude this possibility, we consider it extremely unlikely due to the very high Sr content in type2 plagioclases, which excludes crystallization from melts akin to any of the magmas erupted on Lipari and Vulcano.

Crystal-rich enclaves on Vulcano

Crystal-rich enclaves (~20-30% crystals), showing the same composition and mineral assemblage of Lipari high-K enclaves and containing Sr-rich plagioclases, have been recognized within the Pietre Cotte rhyolitic lava flow on Vulcano (Perugini et al. 2007; Piochi et al. 2009). Similarly to what has been suggested for Lipari (Gioncada et al. 2005; Davì et al. 2009), the Pietre Cotte enclaves have been considered to represent a latitic to trachytic magma interacting with a crystal-poor rhyolitic melt and being co-erupted (Perugini et al. 2007; Piochi et al. 2009). However, crystallization of such Sr-rich plagioclases (4406-5233 ppm; Piochi et al. 2009) from latitic to trachytic magma would require extremely high plagioclase/melt partition coefficients for Sr or extremely high Sr contents in the melt. Notably, the Sr plagioclase/melt partition coefficient calculated for the Pietre Cotte enclaves (9.92; Piochi et al. 2009) is significantly higher than the partition coefficient reported in the literature for the latitic (1.3; Villemant 1988) and trachytic melts (0.94-6.6; Nagasawa 1973; Villemant 1988). Hence, we suggest that such enclaves in Vulcano are also remobilized cumulates, as have been reported from other locations (Crater lake, Bacon and Druitt, 1988; Ammonia Tanks Tuff, Deering et al., 2011b; Carpenter Ridge Tuff, Bachmann et al., 2014; Snake River Plain, Ellis et al., 2014).

Concluding remarks

Following a hiatus of ~40 ka, the youngest eruptive epoch of Lipari is characterised by the appearance of crystal-poor rhyolites (42 ka-AD 1230; Gioncada et al. 2003; Forni et al. 2013). The eruption of rhyolites acts to bookend a compositional suite of rocks characterised by a marked lack of dacitic compositions. The mineralogy, geochemistry and isotopic ratios all support fractional crystallization with limited assimilation of crustal material as the main drive behind the generation of the calc-alkaline rhyolites of Lipari. Such a conclusion is particularly well illustrated in Lipari, as one can compare such AFC-derived silicic magmas with others (the CBL dacitic lavas) clearly involving high crustal components (50-60%; Di Martino et al. 2011). As expected on thermal grounds (e.g.,

Thompson et al. 2002), the majority of the crustal assimilation occurs during the early stages of petrogenesis with relatively steep increases in isotopic crustal tracers within the basaltic andesites and andesites, and then little further increase from andesites to rhyolites.

The crystal-poor nature and low eruptive temperatures of the Lipari rhyolites is testament to the efficiency of crystal-melt separation in silicic systems, not of superheating by mafic recharge. The presence of crystal-rich enclaves with cumulate signatures in the rhyolites directly provides the complementary crystal residues. Mass balance modelling indicate that such rhyolites can be generated by extraction of interstitial melts from originally intermediate compositions after ~ 60 % crystallization. These results are in agreement with fluid dynamics models, which suggest crystallinities of 50-70% crystals as the most suitable window for melt extraction and generation of compositional gaps (Dufek and Bachmann 2010). Such an association between crystal-poor rhyolites and crystal-rich, remobilized cumulates is increasingly recognized in many places around the world (e.g. Bachmann and Bergantz 2004; Shane et al. 2005; Bacon et al. 2007; Lipman 2007; Deering et al. 2011b; Deering et al. 2011c; Ellis et al. 2014) and strongly argues for the ubiquity of the process.

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Figure captions

Fig. 1 Geological sketch map of Lipari Island (modified after Forni et al. 2013) showing the main stages of its eruptive history (eruptive epochs=EE) and the high-K crystal-rich enclaves sampling localities (red stars). Ages and compositions of Lipari volcanics are reported in the legend (BA=basaltic andesites; AND=andesites; D=low-SiO₂ dacites; RHY=rhyolites)

Fig. 2 K₂O vs. SiO₂ classification diagram (Peccerillo and Taylor 1976) and frequency distribution histogram of bulk-rock SiO₂ for the volcanic rocks from Lipari. Data from De Rosa et al. (2003); Gioncada et al. (2003); Davì et al. (2010); Forni et al. (2013), and this study. EE=eruptive epoch

Fig. 3 Backscattered electron images of the main mineral phases hosted within the low- and high-K crystal-rich enclaves. Type1 (a) and type2 (b) plagioclases, type1 (c) and type2 (d) clinopyroxene, type1 olivine (e) and biotite (f). See text for further information

Fig. 4 Selected major and trace element vs. SiO₂ variation diagrams for the Lipari volcanics and enclaves. Data from De Rosa et al. (2003); Gioncada et al. (2003); Davì et al. (2010); Forni et al. (2013) and this study. EE=eruptive epoch

Fig. 5 Chondrite-normalized (Sun and McDonough 1989) REE patterns of Lipari volcanics and enclaves (average values for each eruptive epoch=EE). Data from Gioncada et al. (2003); Davì et al. (2010); Forni et al. (2013) and this study

Fig. 6 Composition of feldspar from Lipari basaltic andesites (=BA), andesites (=AND), rhyolites (=RHY) (a) and low- and high-K enclaves (=E) (b). In the enclaves, type1 and type2 plagioclase are distinguished (see text for further information). EE=eruptive epoch

Fig. 7 Composition of pyroxene from Lipari basaltic andesites (=BA), andesites (=AND) (a) and low- and high-K enclaves (=E) (b). EE=eruptive epoch

Fig. 8 Sr vs. Eu/Eu* diagram of clinopyroxenes from Lipari basaltic andesitic (=BA) to andesitic (=AND) volcanics and crystal-rich enclaves (=E) (type1 and type2 clinopyroxene are distinguished; see text for further information). EE=eruptive epoch

Fig. 9 a Range of plagioclase composition in the basaltic andesites (=BA), andesites (=AND), low- and high-K enclaves (=E) and rhyolites (=RHY) from Lipari. In the enclaves type1 and type2 plagioclase are distinguished (see text for further information). Grey fields indicate the ranges of plagioclase compositions in equilibrium with the bulk-rock calculated using the K_D (Ab-An)=0.16-0.38 for the BA, AND and high-K E ($T > 1050^\circ\text{C}$) and K_D (Ab-An)=0.05-0.15 for the low-K E and RHY ($T < 1050^\circ\text{C}$) (Putirka 2008). **b** plagioclase melt geothermometry based on equation 24a of Putirka (2008) for the plagioclase-liquid equilibrium pairs. Temperature ranges obtained for the Lipari rhyolites using other methods are shown for comparison

Fig. 10 a Nd vs. Sr isotope ratios of Lipari volcanics (BA=basaltic andesites; AND=andesites; CBL=cordierite-bearing lavas and RHY=rhyolites) compared to the literature isotope data of Lipari and the other Aeolian Islands (data from De Astis et al. 2013; Forni et al. 2013; Francalanci et al. 2013; Lucchi et al. 2013a; Lucchi et al. 2013b; Lucchi et al. 2013c; Lucchi et al. 2013d). The isotopic composition of the main lithologies of the Calabrian Basement (data from Caggianelli et al. 1991) and the cordierite from the CBL (from Di Martino et al. 2011) are shown as representative of the crustal rocks composition. **b** AFC geochemical modelling obtained for different assimilation/fractional crystallization ratios (R). Tick marks and numbers along the lines refer to the liquid fraction (f). See text for further information. EE=eruptive epoch

Fig. 11 a $\delta^{18}\text{O}$ of minerals and calculated equilibrium melt vs. SiO_2 variations in Lipari volcanics (BA=basaltic andesites; AND=andesites; CBL=cordierite-bearing lavas) and enclaves (=E). Oxygen isotope compositions of melts in equilibrium with clinopyroxene from Alicudi and Filicudi as representative of primitive isotopic compositions are reported for comparison (Santo and Peccerillo 2008; Peccerillo et al. 2013). **b** Simple bulk-mixing model showing the amount of crustal component to be added to a primitive magma ($\delta^{18}\text{O} = 6$) to match the range of oxygen isotope compositions observed in Lipari volcanics. $\delta^{18}\text{O}$ of the crustal component has been estimated based on the oxygen isotope compositions of cordierite and garnet from the CBL. See text for further information

Fig. 12 $^{87}\text{Sr}/^{86}\text{Sr}$ (a) and $^{143}\text{Nd}/^{144}\text{Nd}$ (b) vs. SiO_2 diagrams showing increasing and decreasing Sr and Nd isotope ratios from the basaltic andesites (=BA) to the andesites (=AND) and CBL (=cordierite-bearing lavas) and nearly constant values of the isotope ratios from the andesites to the rhyolites (=RHY). EE=eruptive epoch

Fig. 13 Range of Sr contents in plagioclase (symbols as in Fig.6) and bulk-rock (full bars) basaltic andesites (=BA), andesites (=AND), low- and high-K enclaves (=E) from Lipari. Calculated Sr content in the melt in equilibrium with the plagioclases (empty bars) and plagioclase/melt partition coefficients (= Kd (Sr)) have been obtained using the equation of Bindeman et al. (1998) within the range of $0.4 < X_{\text{An}} < 0.8$ of plagioclase compositions. Sr vs. An% variations in plagioclases are shown in the inset (in the enclaves type1 and type2 plagioclase are distinguished; see text for further information)

Fig. 14 Trace element modelling obtained using Rb vs. Ba (a) and Rb vs. Sr (b) diagrams. Symbols and numbers along the lines indicate melt fraction (f). Partition coefficients have been estimated based on the mineral proportions obtained from mass balance calculations. High and low partition coefficients have been tested (see Online Resource 9). See text for further information

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Figure 1

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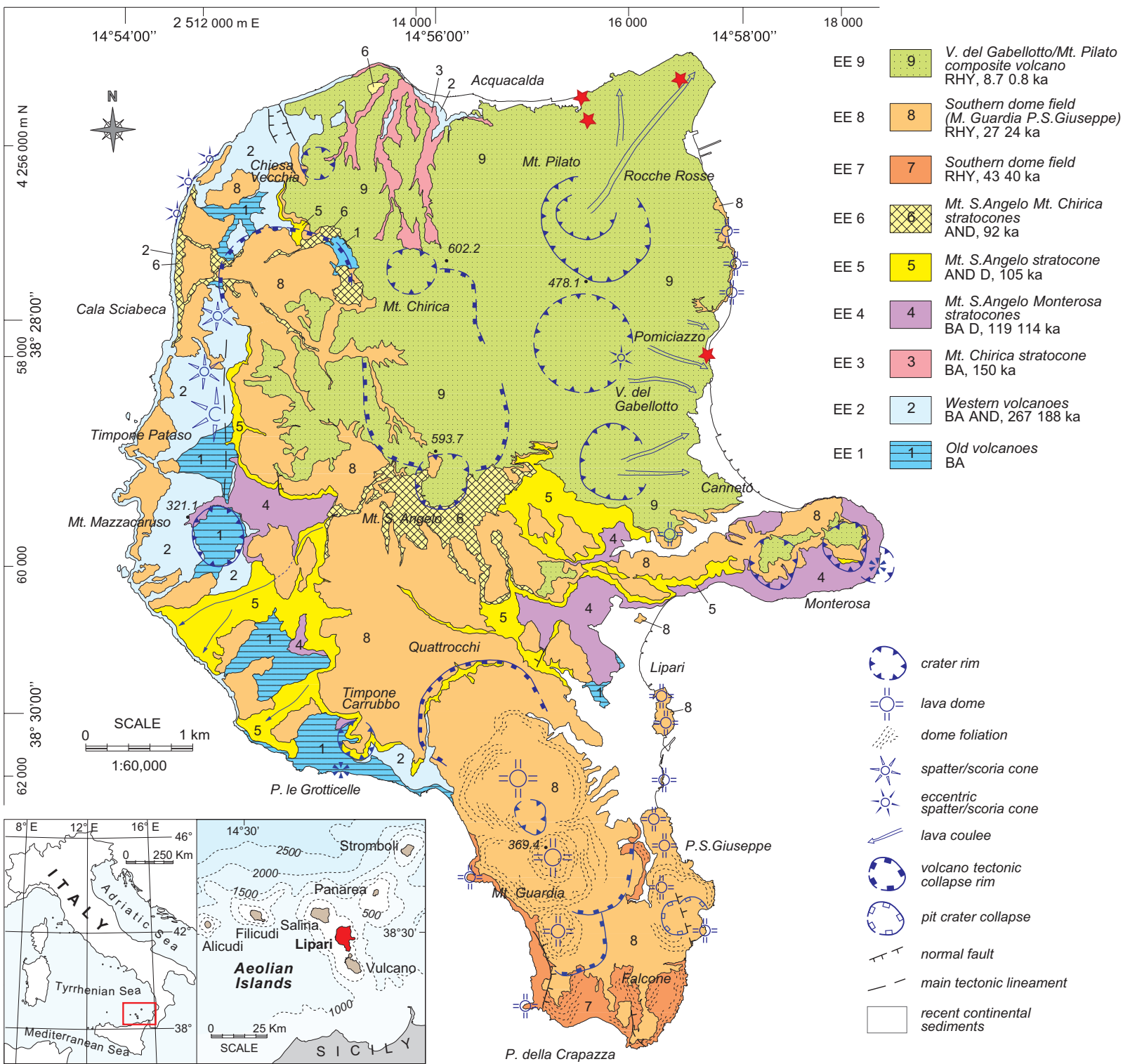


Figure 2
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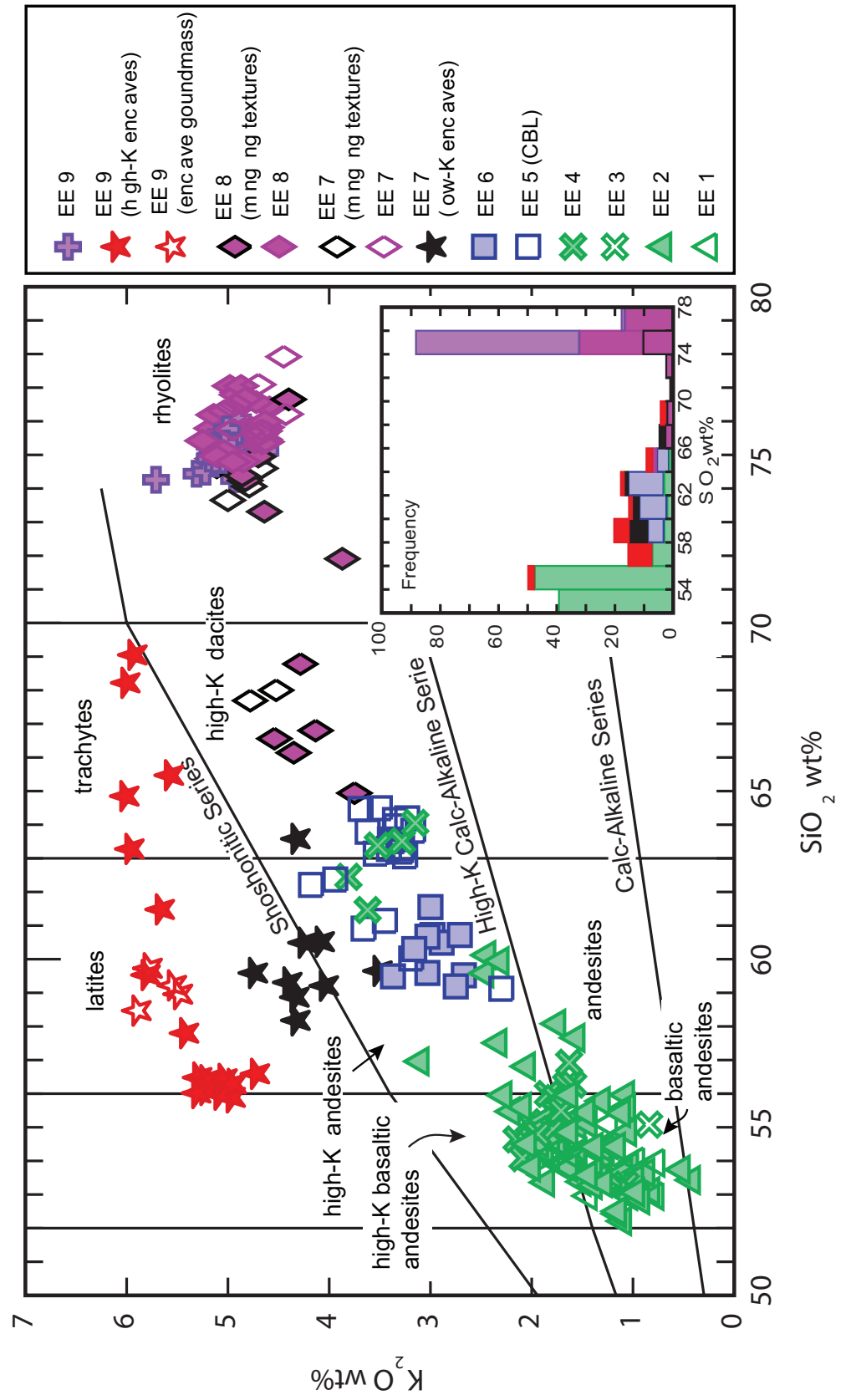


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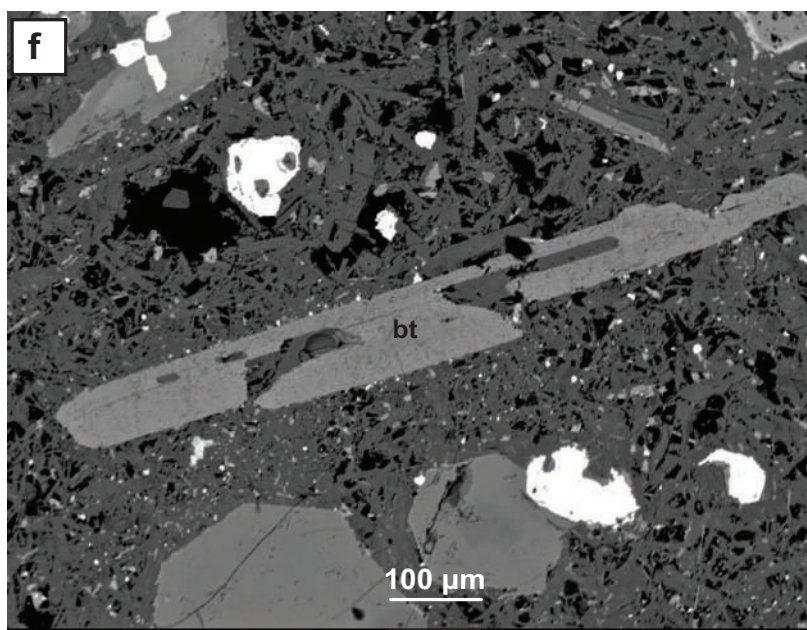
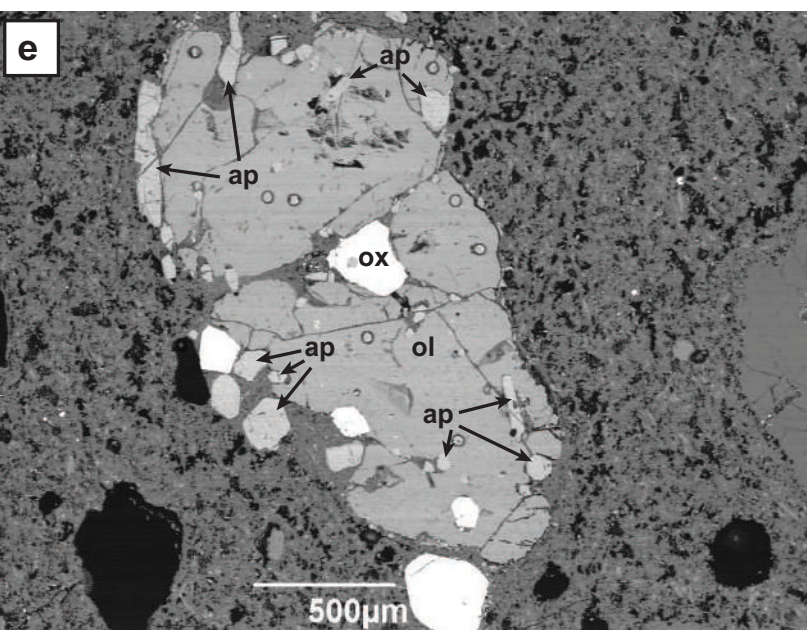
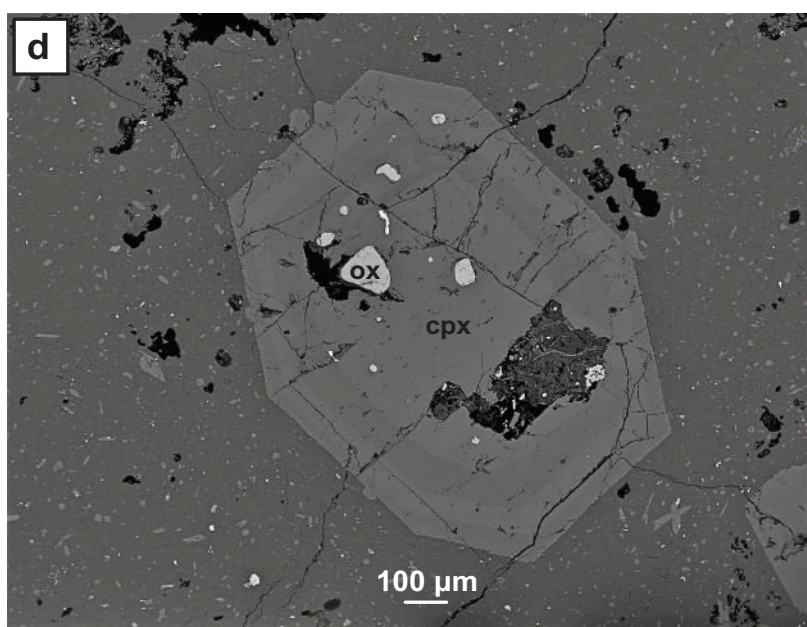
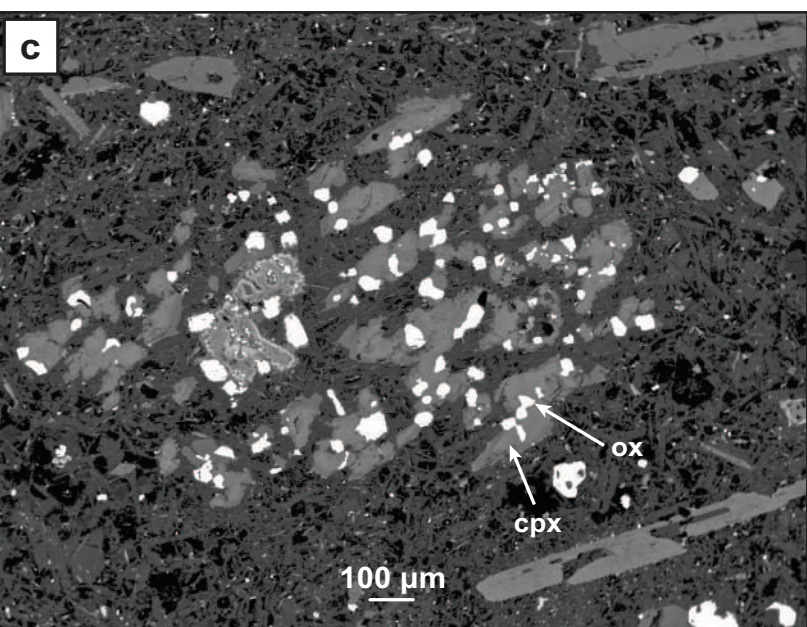
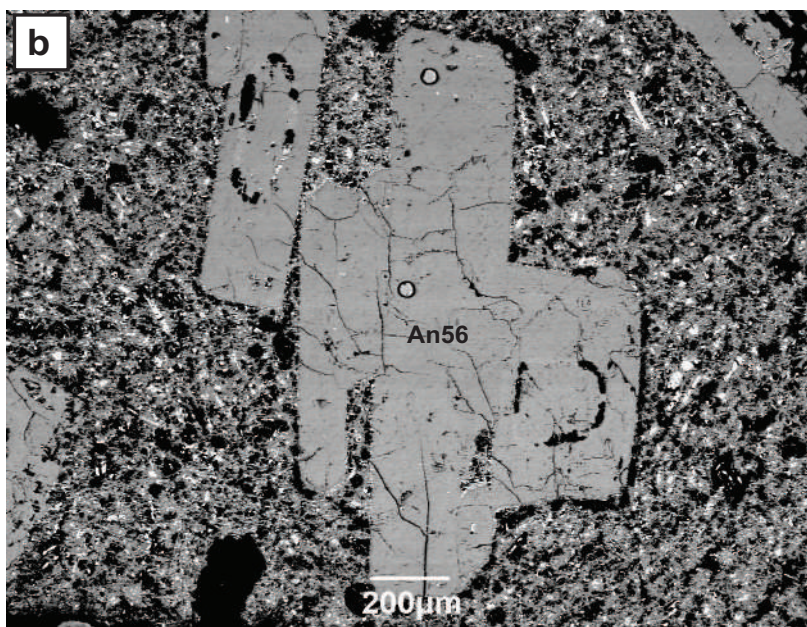
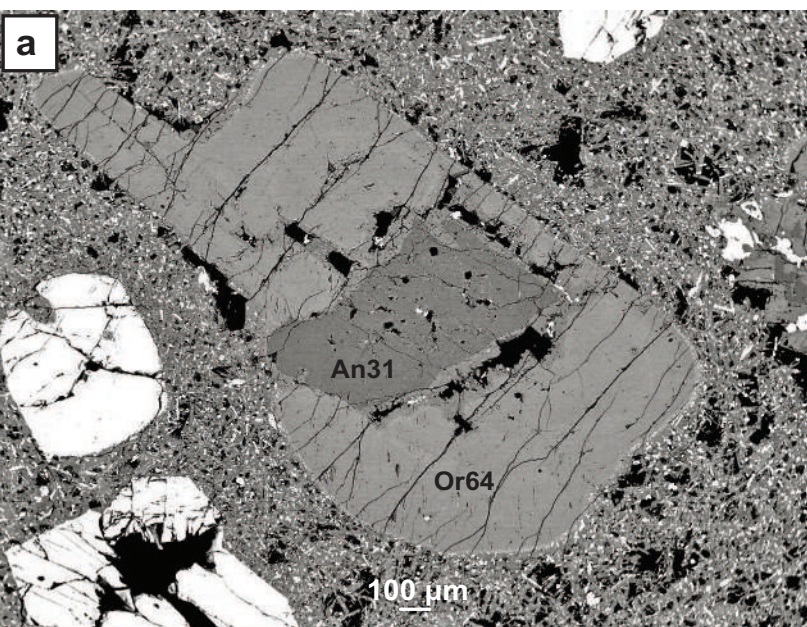


Figure 4

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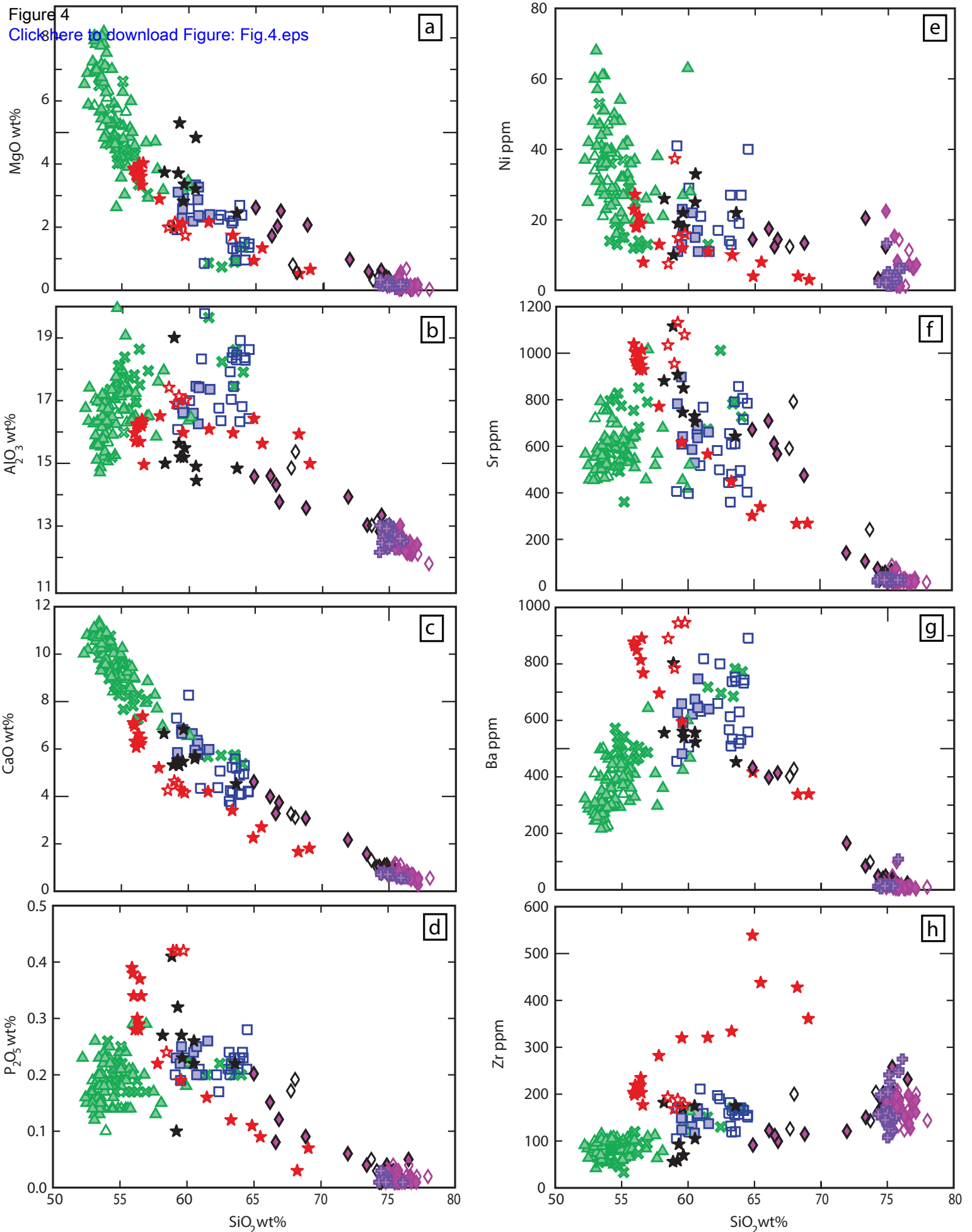


Figure 6
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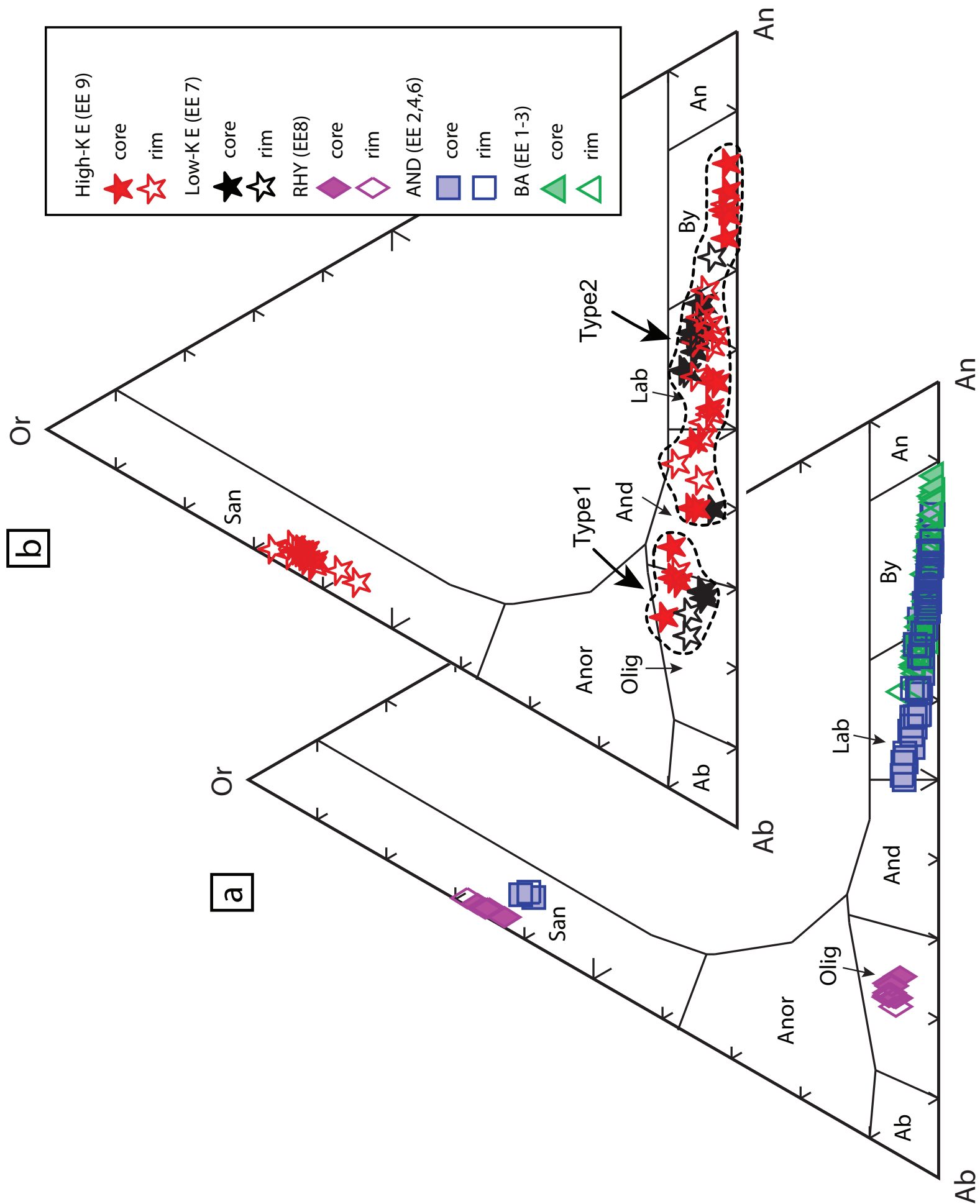


Figure 7
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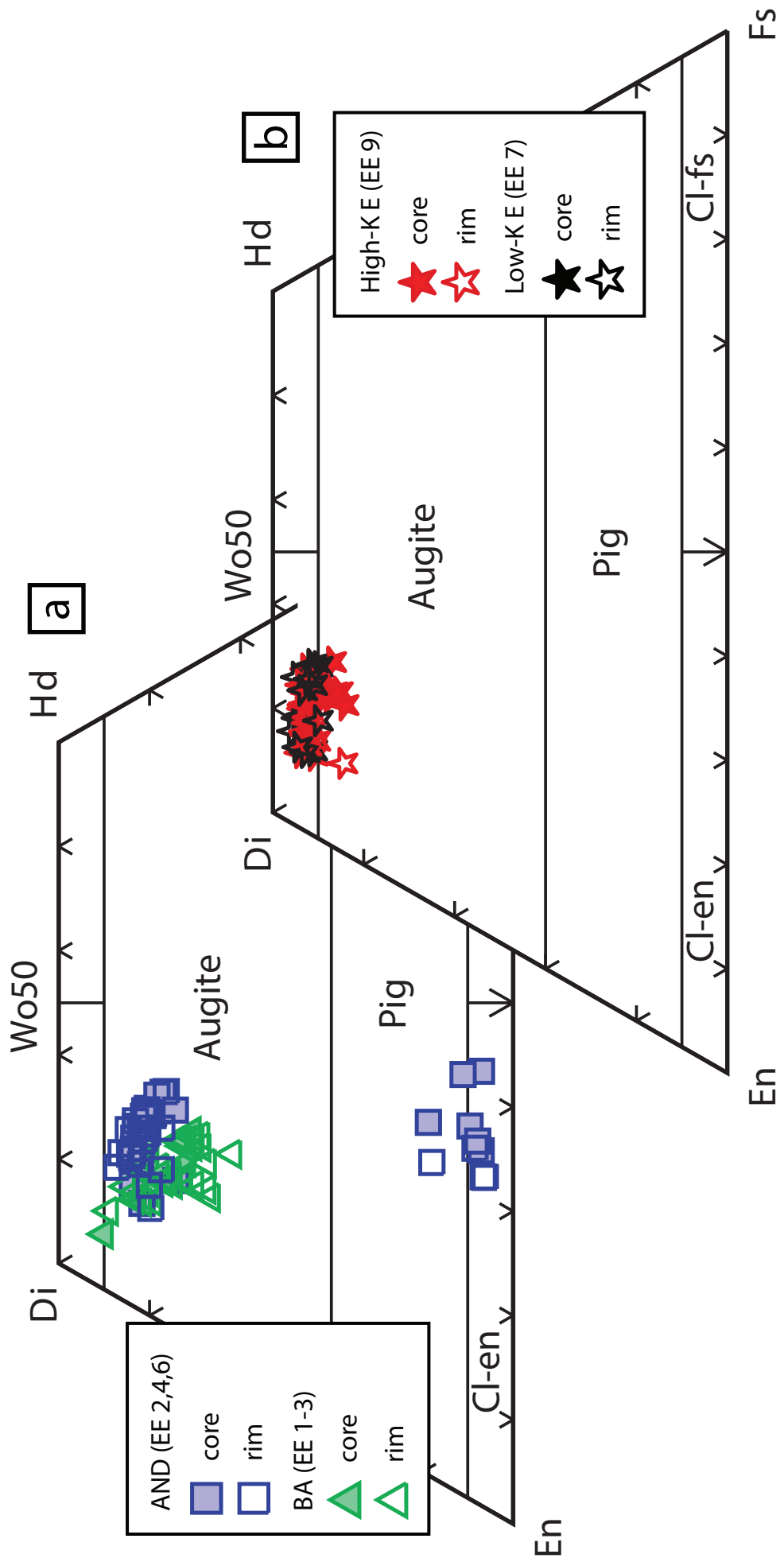


Figure 8

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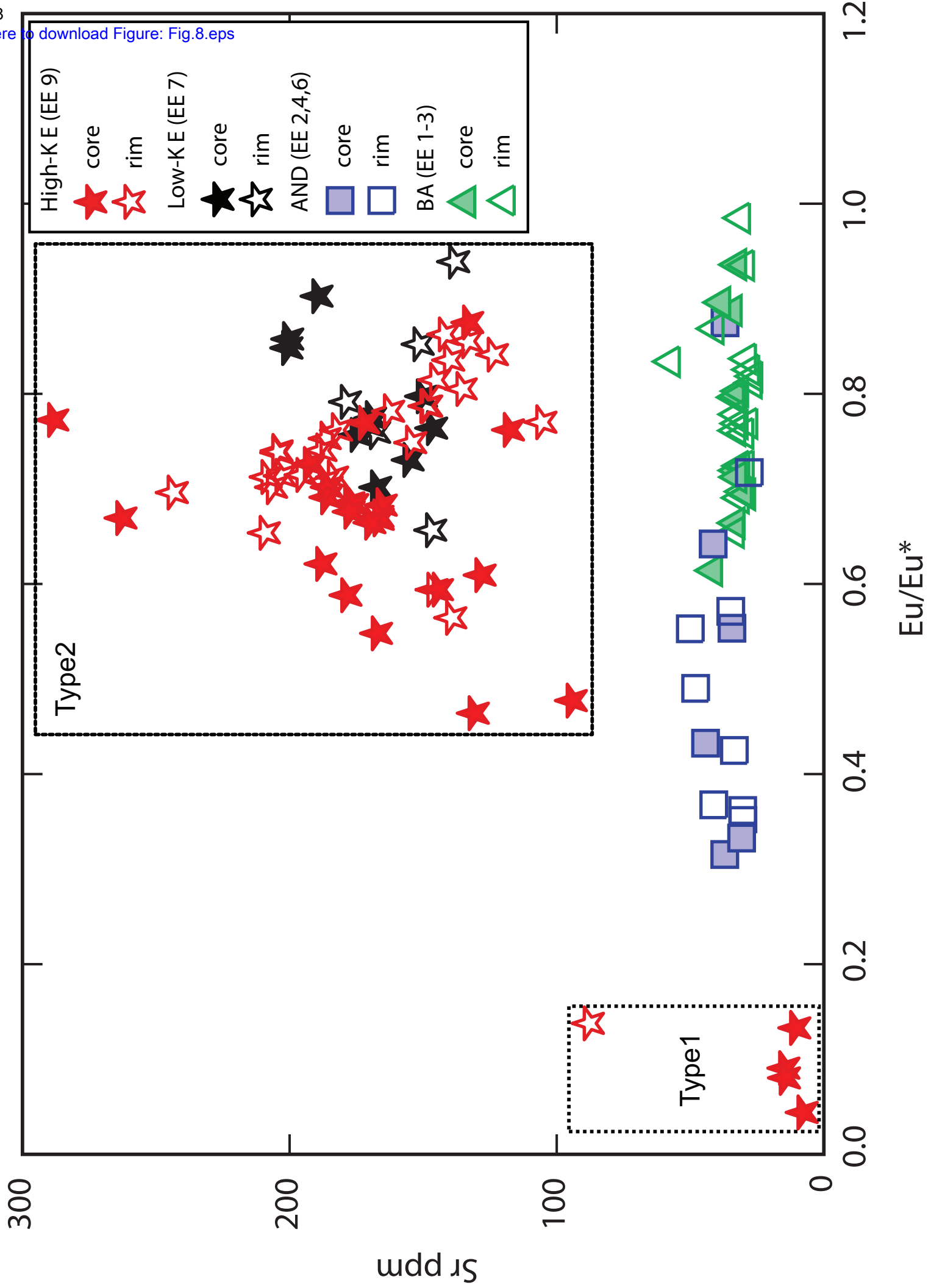


Figure 9

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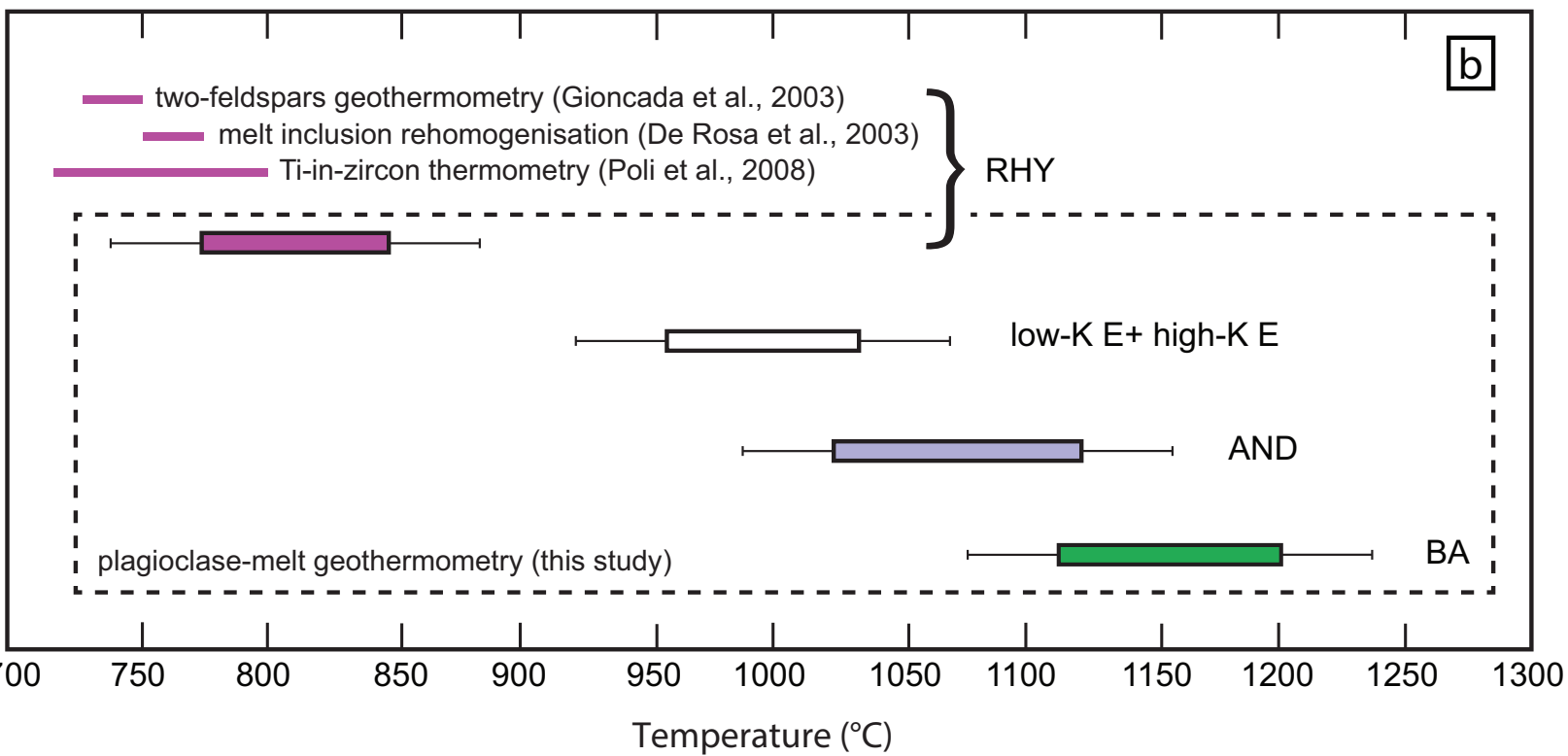
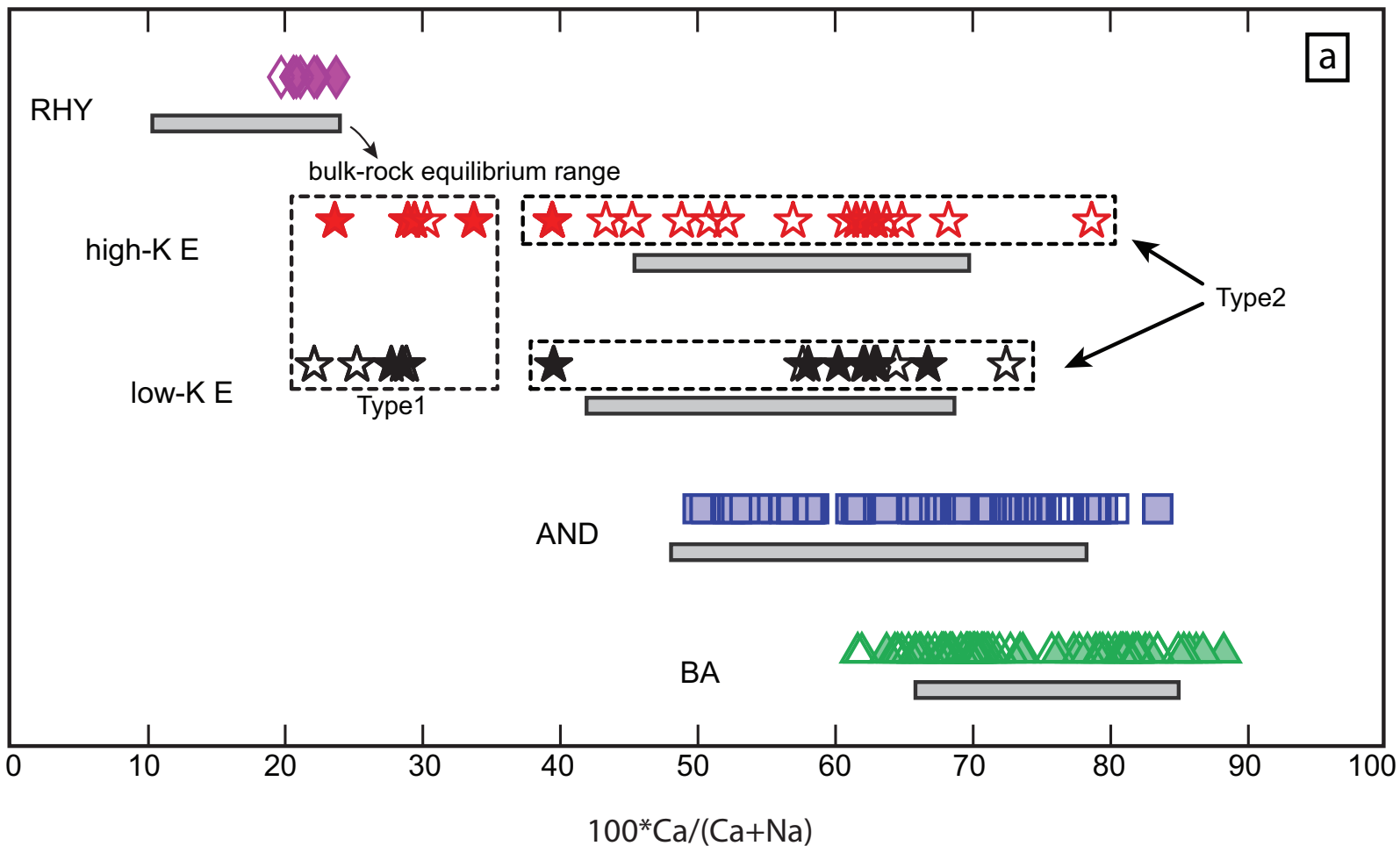


Figure 10

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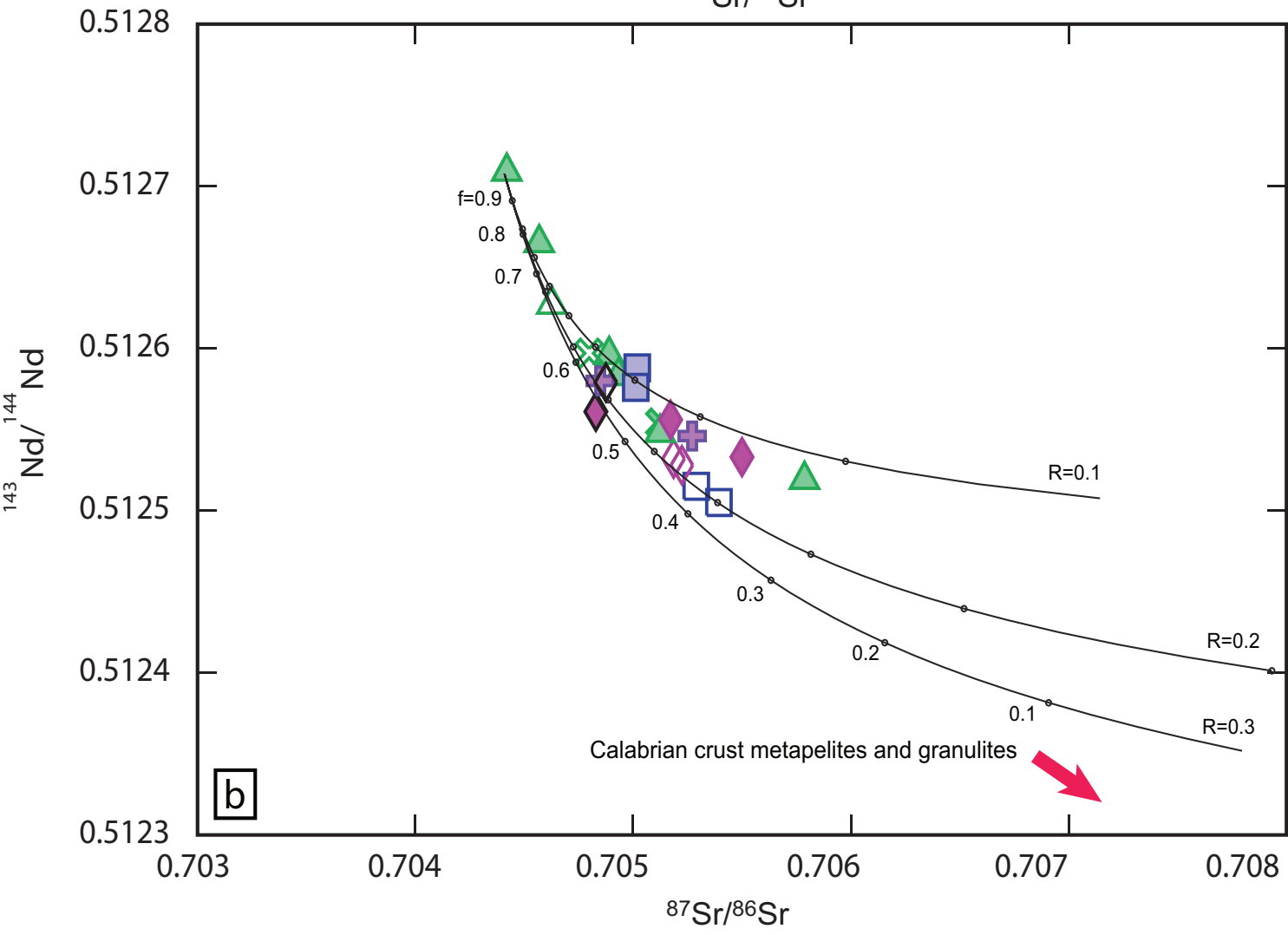
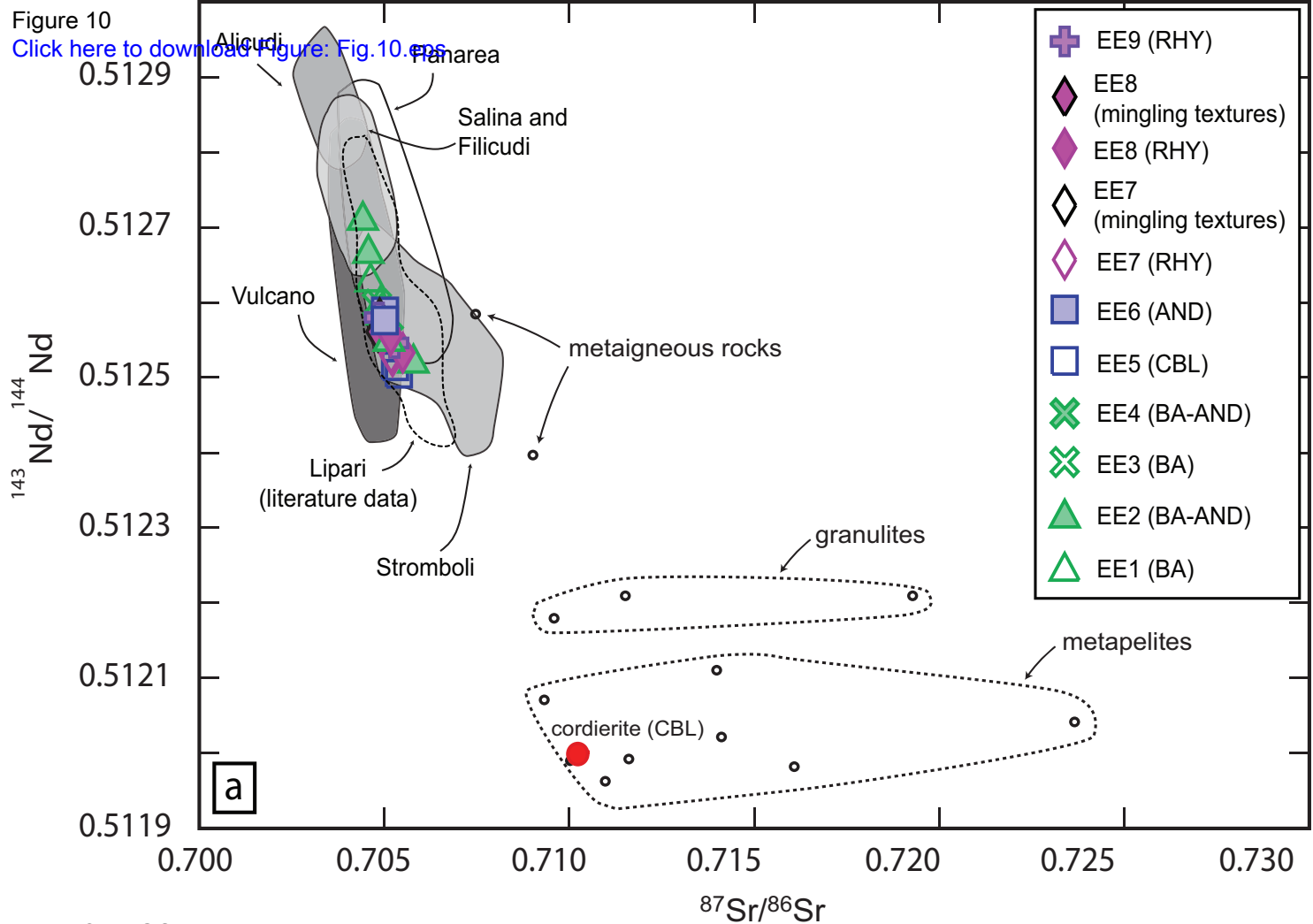


Figure 15

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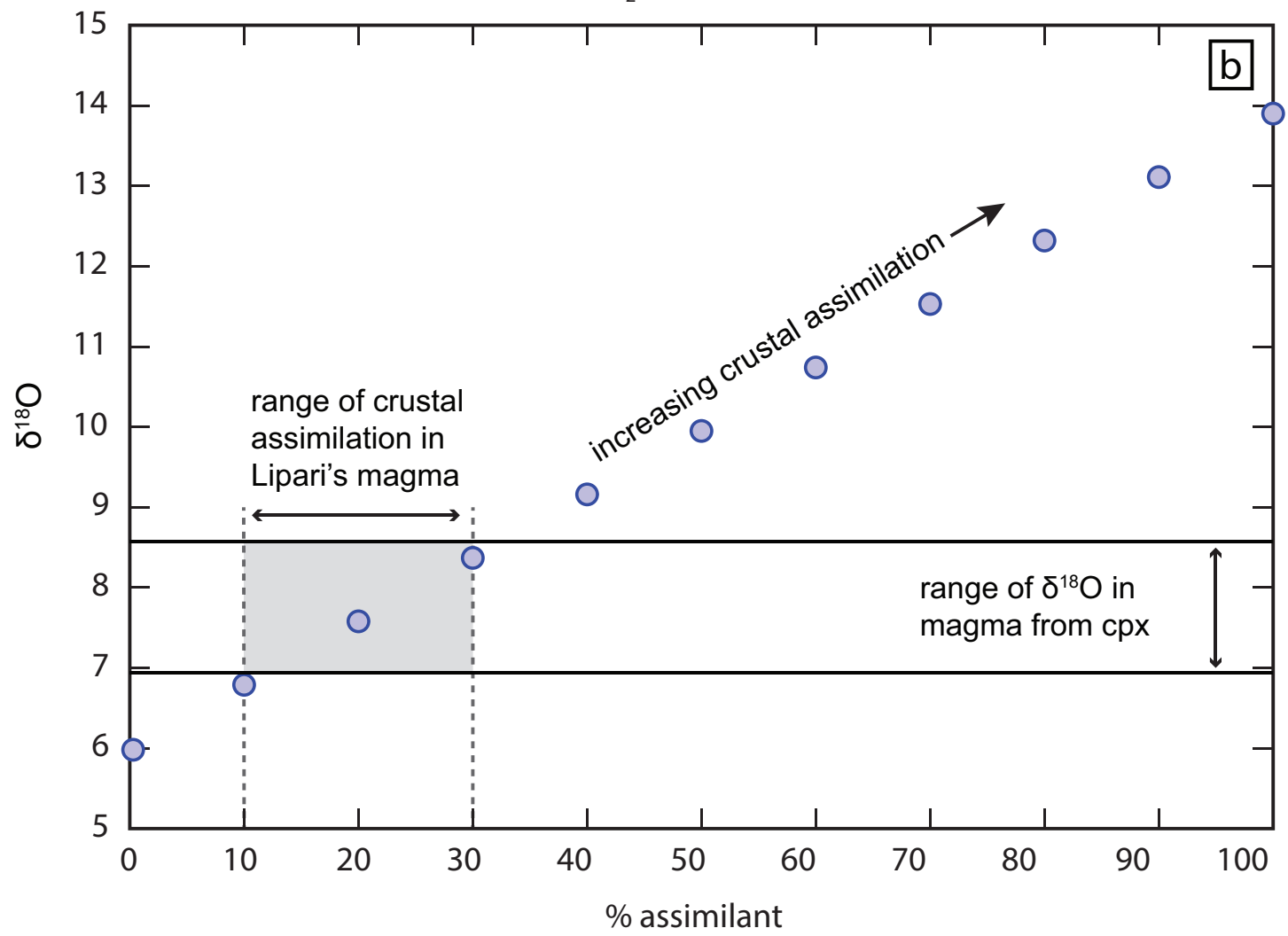
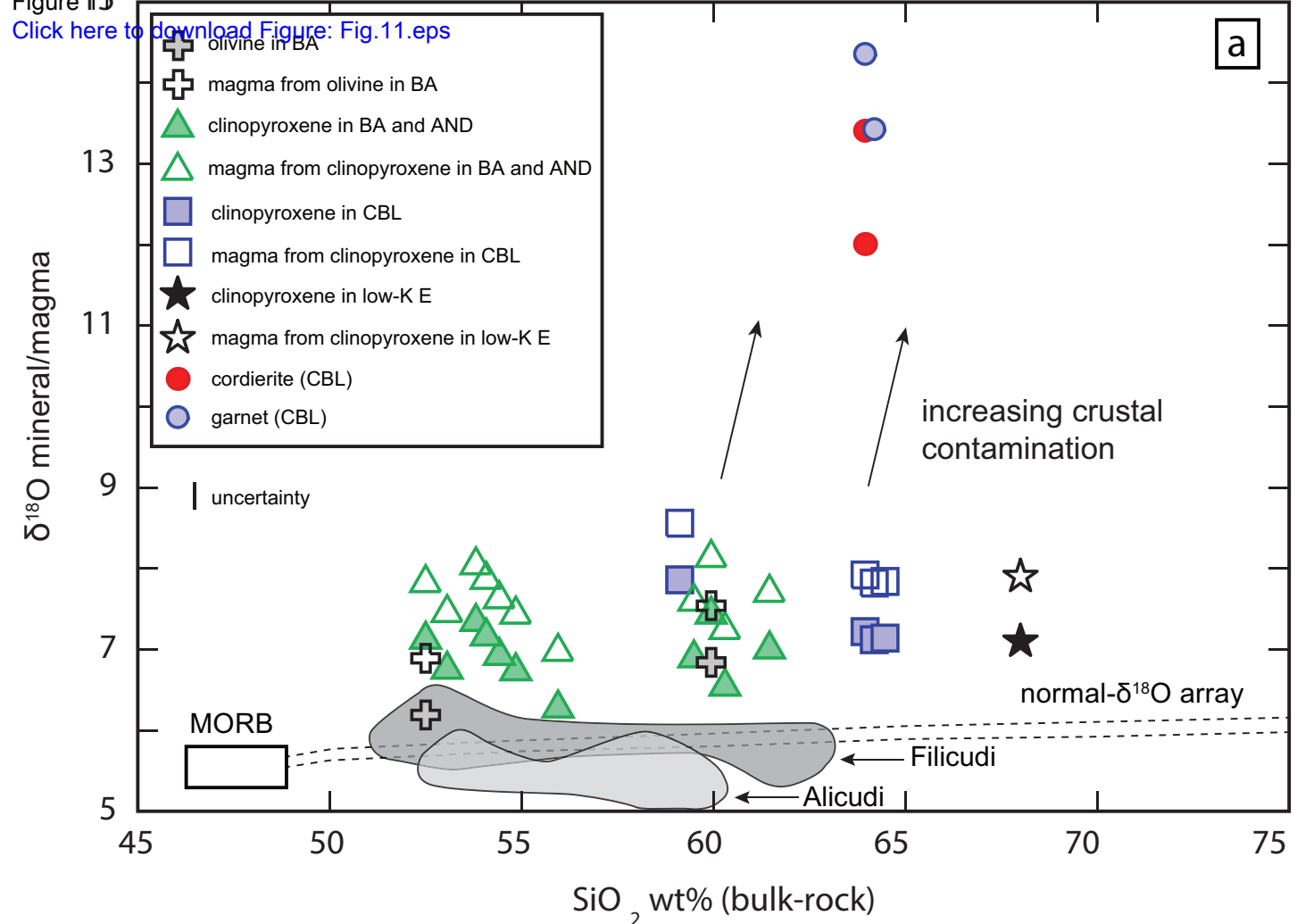


Figure 12
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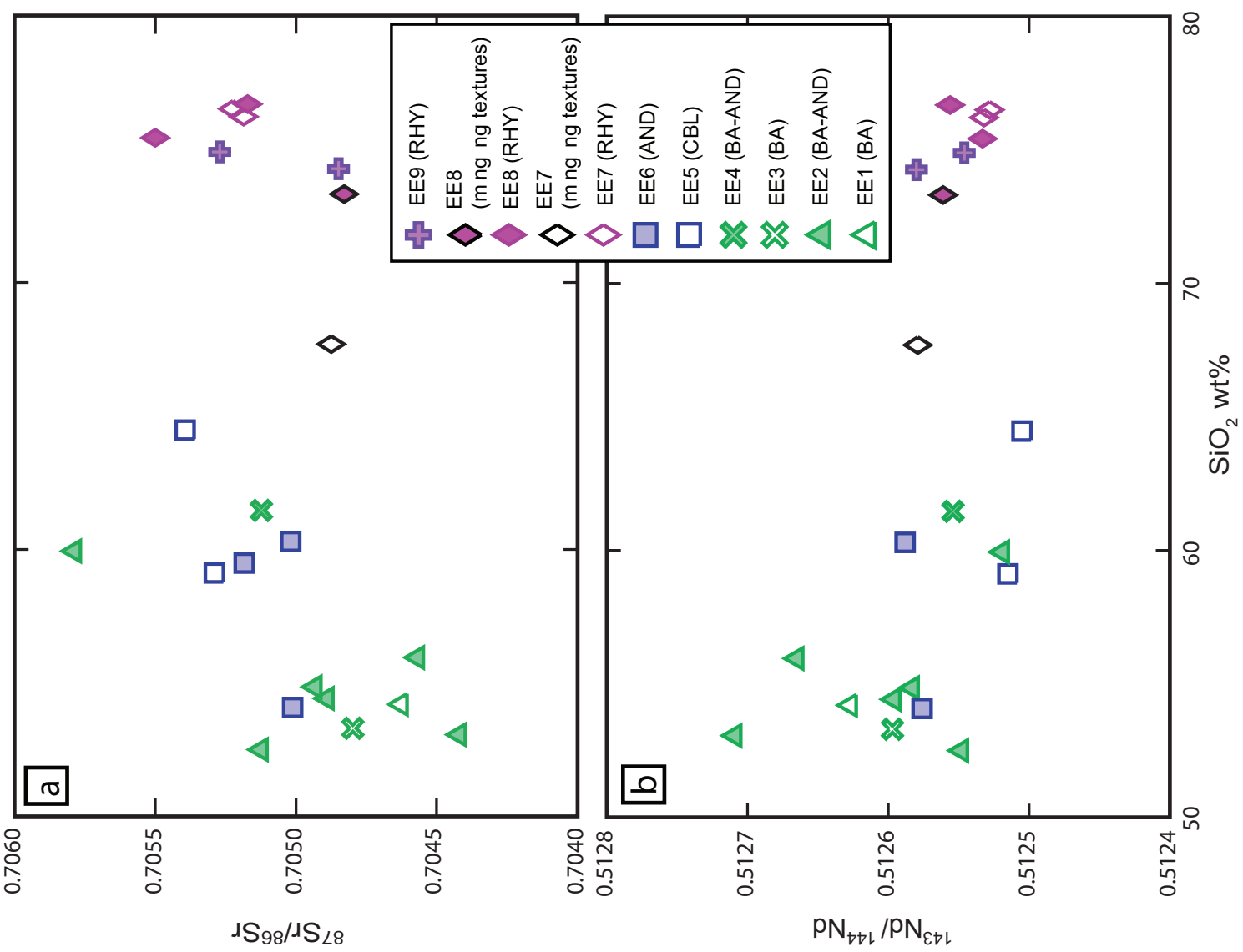


Figure 13

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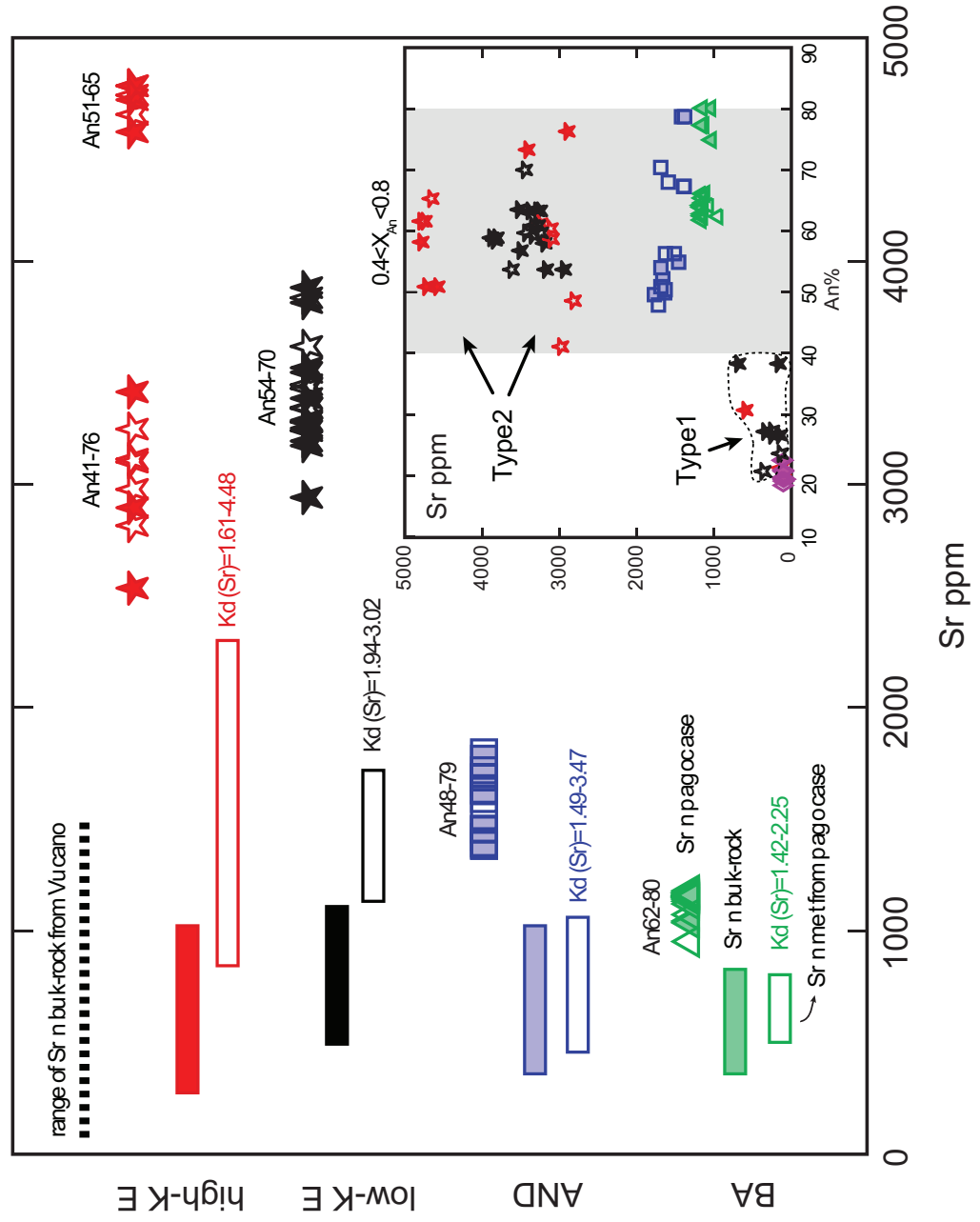
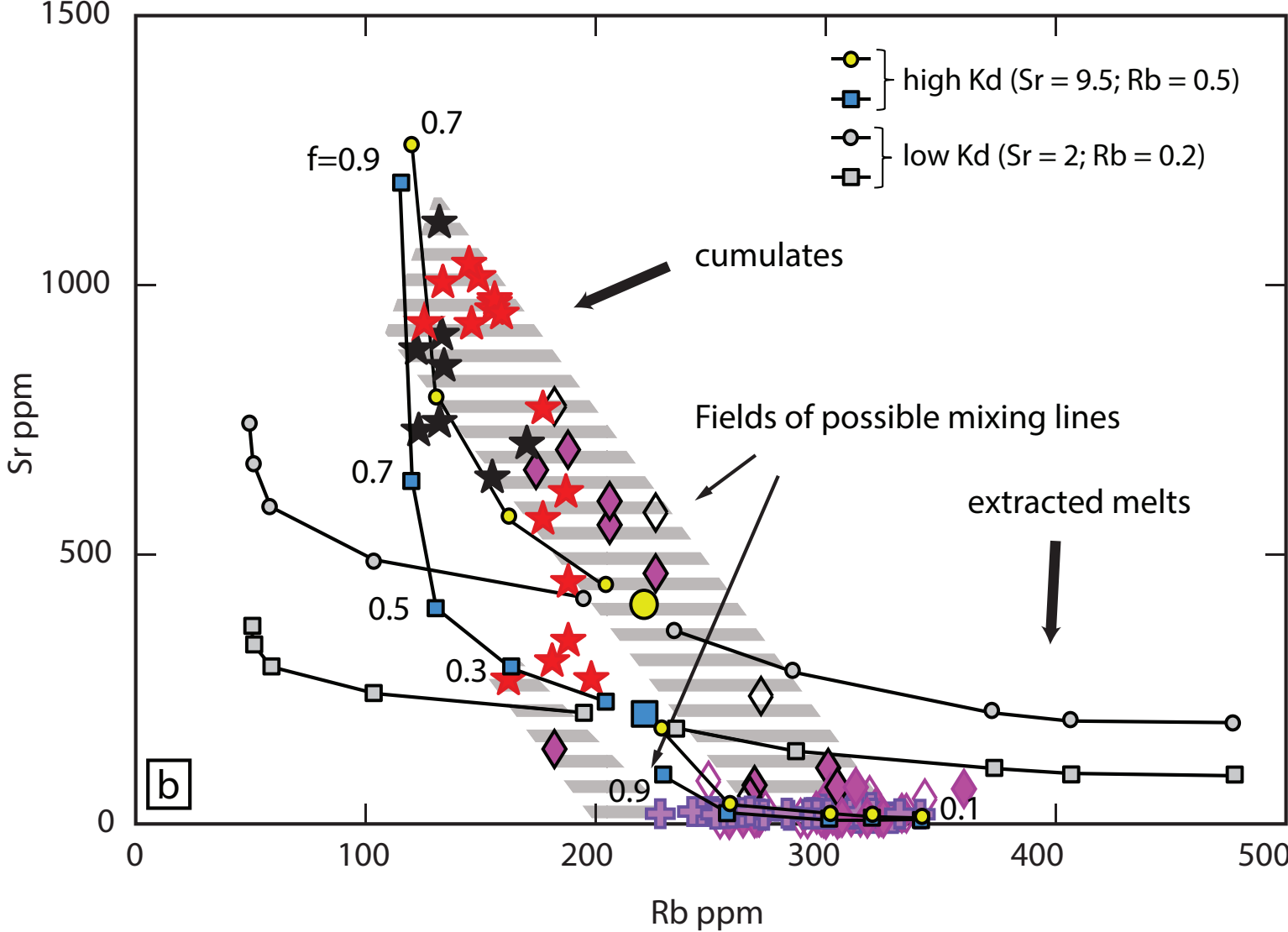
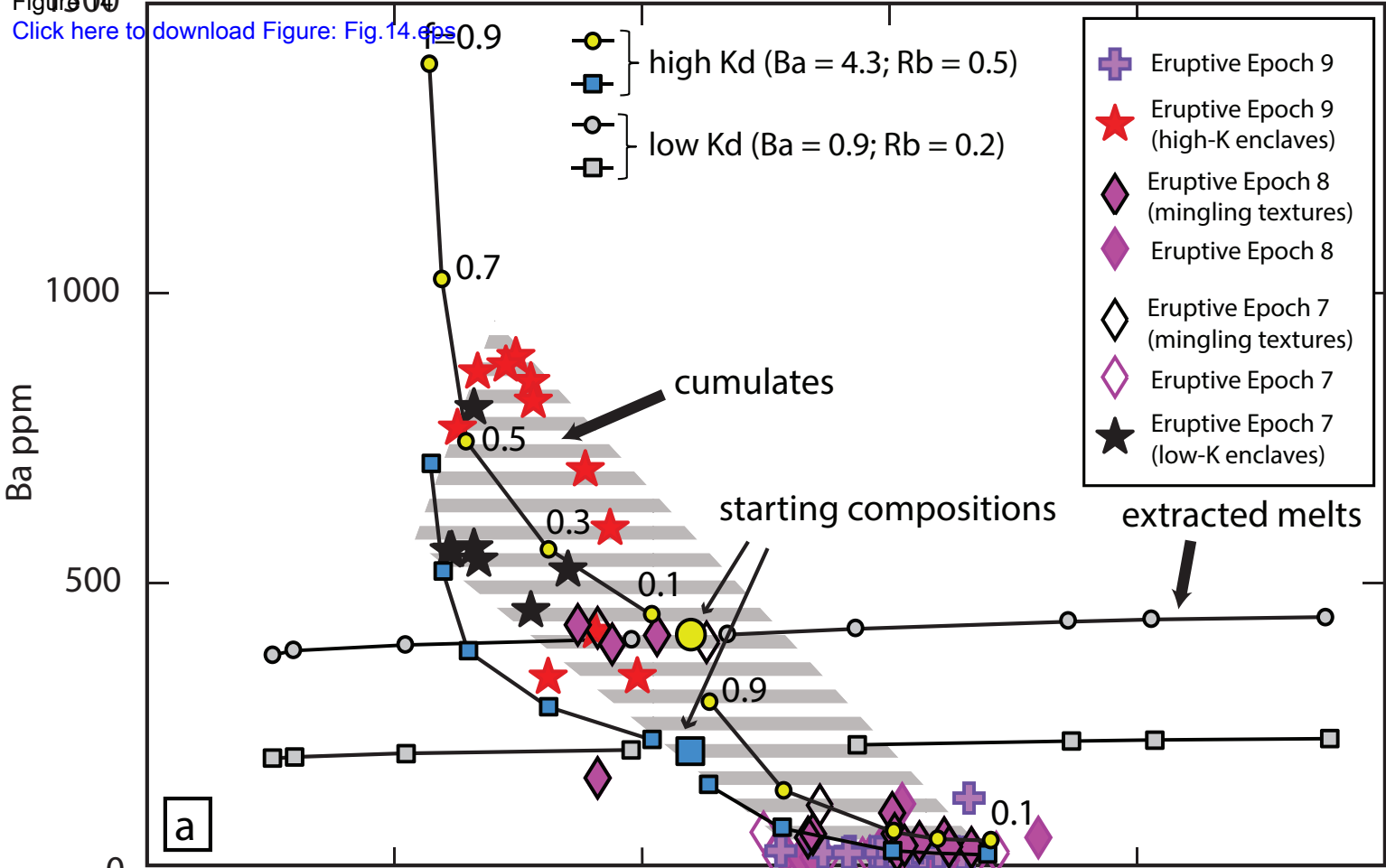


Figure 14

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