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4 **New constraints from Central Chile on the origins of enriched continental compositions in thick-crusted**
5 **arc magmas**

6 Penny E. Wieser^{1,2*}, Stephen J. Turner³, Tamsin A. Mather¹, David M. Pyle¹, Ivan P. Savov⁴ and Gabriel
7 Orozco⁵

8
9 ¹ Department of Earth Sciences, University of Oxford, South Parks Road, Oxford OX1 3AN, UK

10 ² Now at Department of Earth Sciences, University of Cambridge, Downing Street, Cambridge, UK, CB2 3EQ.

11 ³ Washington University in St. Louis, One Brookings Drive, St. Louis, MO 63130

12 ⁴ School of Earth and Environment, Institute of Geophysics & Tectonics, University of Leeds, Leeds LS2 9JT,
13 UK.

14 ⁵ Red Nacional de Vigilancia Volcánica, Servicio Nacional de Geología y Minería, Av. Santa María 0104,
15 Providencia, Chile.

16
17 * Corresponding author: pew26@cam.ac.uk. Phone: 07593670818

18 **Abstract**

19 Volcanics from thick-crusted continental arcs have elevated incompatible element abundances and
20 “enriched” radiogenic isotope ratios compared to volcanics erupted in island arcs and continental arcs with
21 thinner crust. The relative influence of the slab, mantle, and upper plate on this variability remains debated.
22 The Andean Southern Volcanic Zone (SVZ; 33-46° S) is an ideal setting to investigate the production of
23 enriched continental arc volcanics, because both crustal thickness and magma chemistry show coherent
24 along-strike variations. However, the scarcity of primitive volcanism in the thick-crusted northern SVZ has
25 hindered previous regional studies. To better address the origin of enriched continental compositions, this
26 study investigates the geochemistry (major and trace element abundances, ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd) of new
27 mafic samples from Don Casimiro and Maipo volcanoes (within the Diamante-Maipo Caldera Complex of the
28 northern SVZ). While evolved Diamante-Maipo samples show evidence for crustal assimilation, the trace
29 element and isotopic enrichment of the most mafic samples cannot result from crustal processing, as a
30 compilation of regional and global basement lithologies are not enriched in all of the necessary incompatible
31 trace elements. This reflects the fact that many of the elements which are enriched in the mafic Don
32 Casimiro-Maipo samples (e.g. Sr, Zr, P) are depleted by the various complex petrogenetic processes affecting
33 basement lithologies (e.g. fractionation of accessory phases). Subduction erosion models similarly fail to
34 account for the enriched isotopic and trace element signature of these samples. Instead, the enrichment of
35 the northern SVZ is most consistent with an enriched ambient mantle component (similar to EM1-type
36 ocean island basalts), superimposed on a northward decline in melt extent. A substantial, but nearly uniform
37 contribution of melts from subducting sediment and altered oceanic crust are also required at all latitudes.
38 The EM1-like enrichment may arise from recycling of metasomatized subcontinental lithospheric mantle (M-
39 SCLM), as the isotopic trajectory of primitive rear-arc monogenetic cones trend towards the composition of
40 SCLM melts sampled across South America. The wide spatial distribution of rear-arc centres with isotopic
41 data clearly demonstrates that the latitudinal variations in the amount of EM1-type enrichment observed in
42 arc-front samples are also present in the rear-arc. Rear-arc trace element systematics indicate that

43 significant but variable quantities of slab melts are also transported to the rear-arc mantle. Overall, we show
44 that a unified model incorporating variable mantle enrichment, slab additions, and melt extents can account
45 for along and across arc trends within the SVZ. As subduction zones have been the loci of continental crust
46 creation for over 2 Ga, the finding that mantle enrichment plays a dominant role in the production of
47 enriched continental compositions in the SVZ has important implications for our understanding of the
48 chemical evolution of the Earth. If ambient mantle enrichment is not taken into account, petrogenetic
49 models of evolved lavas may overestimate the role of crustal assimilation, which, in turn, may lead models of
50 continental crust growth exaggerating the amount of continental material that must be recycled back into
51 the mantle to satisfy mass balance.

52

53 **Key Words**

- 54 • Mantle heterogeneity
- 55 • Andean Southern Volcanic Zone

56 **Highlights**

- 57 • NSVZ isotopic and trace element enrichment indicate an EM1-like mantle source.
- 58 • EM1-like signatures originate from recycling of M-SCLM from nearby cratons.
- 59 • Slab melting, low mantle melt extents, and M-SCLM may enrich continental arc lavas
- 60 • The slab component supplied to the rear-arc and arc-front have similar compositions.
- 61 • Slab materials are not transported vertically to the wedge from their point of origin.

62 **1. Introduction**

63 Magmas which erupt from continental arc-front stratovolcanoes are compositionally distinct from
64 oceanic arc magmas. With the exception of volcanics from intra-arc rift zones (e.g. Conrey and Schmidt,
65 2004), these continental magmas exhibit elevated incompatible element abundances, steeper incompatible
66 trace element patterns (Fig. 1a), and often have isotopic signatures offset to higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower
67 $^{143}\text{Nd}/^{144}\text{Nd}$ than oceanic arc magmas (Leeman et al., 1983; Plank and Langmuir, 1988; Hildreth and
68 Moorbath, 1988; Turner and Langmuir, 2015a; Farner and Lee, 2017). Many of these compositional
69 characteristics correlate strongly with crustal thickness (Turner et al., 2015a), and have been used as proxies
70 to estimate changes in crustal thickness and rates of surface uplift through time (Profeta et al., 2015;
71 Chiaradia, 2015; Chapman et al., 2015; Scott et al., 2018). Some of the compositional offsets between thick
72 and thin-crust arcs can be accounted for by higher extents of crystal fractionation and crustal assimilation
73 in continental settings (e.g. Farner and Lee, 2017). However, other compositional differences persist in high
74 Mg# arc lavas (Fig. 1a) that have not been extensively overprinted in the crust (e.g. Plank and Langmuir,
75 1988; Turner and Langmuir, 2015a; Turner et al., 2017; Schmidt and Jagoutz, 2017), suggestive of a mantle
76 origin. Compositional differences among parental magmas from continental vs. oceanic arc settings have
77 been attributed to variations in slab temperature (Ruscitto et al., 2012; Turner and Langmuir, 2015b;
78 Schmidt and Jagoutz, 2017), different extents of mantle melting (e.g. Plank and Langmuir, 1988; Tormey et
79 al., 1991; Turner et al., 2016), subduction erosion (e.g. Stern, 1989), and heterogeneity of the “ambient”
80 mantle (the mantle prior to the addition of subducted materials, e.g. Hickey et al., 1986; Ewart and
81 Hawkesworth, 1987; Rogers and Hawkesworth, 1989; Hochstaedter et al., 2001; Pearce et al., 2007; Turner
82 et al., 2017). Furthering our understanding of the thermal structures of subducting plates, the growth of the
83 continental crust, and the petrogenesis of evolved arc magmas requires constraints on the relative influence
84 of each factor in producing the enriched compositions of continental arc magmas.

85

86 The Andean Southern Volcanic Zone (SVZ) is an excellent natural laboratory in which to investigate how the
87 thick crusts and lithospheres of active compressional plate margins affect the compositions of magmas
88 added to the continents. The SVZ (33-46° S) is one of four volcanically active segments within the Andean
89 Cordillera, where the Nazca Plate subducts under South America. It is commonly subdivided into three sub-
90 segments: the northern, transitional, and southern SVZ (NSVZ, TSVZ and SSVZ respectively; Dungan et al.,
91 2001; Fig. 2a). From the easily-accessed segments of the SSVZ to the NSVZ (42° S to 33° S), crustal thickness
92 increases from ~30 to 50 km (Fig. 2c), sub-arc slab depth increases from ~70 to 120 km (Tassara and
93 Echaurren, 2012), and the lower plate age increases from 16 to 37 Ma (Völker et al., 2011), while trench
94 sediment thickness decreases (Völker et al., 2013). In the NSVZ, seismic and heat flow data indicate that the
95 lithosphere thickness increases rapidly behind the arc, while behind the SSVZ, the lithosphere is thinner
96 (Tassara and Echaurren, 2012; Valdenegro et al., 2019). As a result, the mantle wedge in the NSVZ is colder
97 and deeper (Turner et al., 2016). There are also well-established geochemical gradients from the SSVZ to the
98 NSVZ (Tormey et al., 1991; Hildreth and Moorbath, 1988; Hickey et al., 2016, Fig. 2d-e). The trace element
99 signatures of mafic SSVZ volcanics overlap those of oceanic arcs (Fig. 1a). In contrast, the more mafic TSVZ
100 lavas have substantially elevated trace element abundances. The limited amount of trace element data from
101 mafic NSVZ lavas (Fig. 1a; Hickey et al., 1986) demonstrates that this segment is even more enriched than
102 the TSVZ, with compositions similar to the bulk continental crust and mafic to intermediate volcanics from
103 other thick-crust arc segments such as the Northern Cascades (Fig. 1a). Thus, the chemical variability
104 between the SSVZ and NSVZ is analogous to the chemical offsets between island arcs, continental arcs, and
105 the bulk continental crust.

106
107 While many studies have investigated the origin of along-strike compositional variability of the SVZ arc-front
108 volcanoes (Lopez-Escobar et al., 1977; Hickey et al., 1986; Hildreth and Moorbath, 1988; Tormey et al., 1991;
109 Jacques et al., 2014; Hickey et al., 2016), the role of enrichment and heterogeneity within the ambient
110 asthenospheric mantle (prior to the addition of subducted material) has received relatively little attention
111 (Hickey et al., 1986; Jacques et al., 2014). Though often overlooked, the extent to which the compositions of
112 continental arc volcanics are inherited from the ambient mantle has important implications for models of
113 crustal growth. If the trace element enrichment of continental arcs is inherited partially from the mantle,
114 and not exclusively from extensive crystal fractionation and crustal remelting, a substantially smaller amount
115 of material must be returned to the mantle to account for the continental mass balance (e.g. Sisson and
116 Keleman, 2018). The amount of crustal recycling, in turn, has direct relevance for our understanding of the
117 generation of large-scale mantle heterogeneity.

118
119 Heterogeneity of the upper mantle is commonly inferred from trace element and isotopic compositions of
120 mid-oceanic ridge basalts (MORBs) and ocean island basalts (OIBs; Zindler and Hart, 1986; Willbold and
121 Stracke, 2010). OIB isotopic variability can be generally be accounted for by a handful of isotopically distinct
122 end-member components, such as the “enriched mantle” end-members (EM1, EM2; e.g. Zindler and Hart,
123 1986). Isotopic end-members are often modelled as the products of recycled lithospheric mantle, sediment,
124 and ancient oceanic/continental crust (Willbold and Stracke, 2010). There is also evidence for heterogeneity
125 in the ambient mantle that feeds volcanic arcs, despite additional tectonic complexity. For example, a global
126 compilation of samples from rear-arc provinces, filtered to minimize slab contributions (Turner and
127 Langmuir, 2015b), form a linear array between Depleted MORB Mantle (DMM) and EM-1 type OIB
128 compositions (Fig. 3a). As corner flow transports the mantle feeding rear-arc volcanism over hundreds of
129 kilometres towards the arc-front (MacDougall et al., 2017), isotopic enrichment within rear-arc provinces
130 indicates that ambient sub-arc mantle compositions also vary substantially (e.g. Woodhead et al., 2012). The

131 Andean Southern Volcanic Zone (SVZ) rear arc exemplifies this enriched ambient mantle signature (Kay et al.,
132 2013; Jacques et al. 2013, 2014; Søger et al., 2015a; Turner et al. 2017), with more enriched $^{87}\text{Sr}/^{86}\text{Sr}$ and
133 $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than any other rear-arc province globally (Fig. 3a).

134
135 This study investigates the petrogenesis of mafic lava samples from the volcanoes Don Casimiro and Maipo
136 (within the Diamante-Maipo Caldera complex; 34°S , Fig. 2b) in order to determine the geochemical character
137 of high Mg# NSVZ magmas. These lavas from Don Casimiro and the basal portion of Maipo, which erupted
138 immediately after the catastrophic formation (~ 150 ka) of the Diamante Caldera (Orozco et al., 2015), are
139 among the most primitive ever sampled within the NSVZ (Fig. 4a; Hickey et al., 1986). Their primitive nature
140 may indicate that the storage and eruption of these melts preceded the re-organization and maturation of
141 crustal magma reservoirs beneath the Diamante caldera (analogous to post collapse lavas at Campi Flegrei;
142 Forni et al., 2018), or that ascending magmas utilized tension cracks formed during the collapse event
143 (Jacques et al., 2014). Regardless of the exact mechanism permitting the eruption of mafic lavas within the
144 thick-crustal NSVZ, these new samples provide an opportunity to rigorously characterise and explore the
145 origins of the prominent chemical differences between the SSVZ and the NSVZ. As the chemical variability
146 between the SSVZ and NSVZ is analogous to the chemical offsets between island arcs and continental arcs,
147 exploring these regional trends provides constraints on origins of magmatic enrichment within a thickened
148 continental subduction zones globally.

149
150 This study also incorporates samples erupted within the last 1 Myr from seven Argentinean monogenetic
151 cones from the Payenia Volcanic Province (34.33° to 35.38°S ; Folguera et al., 2009). The origin of this rear-arc
152 volcanism, located up to 350 km behind the NSVZ arc front, remains a topic of active research. Volcanism
153 initiated in the earliest Miocene, associated with a period of shallow slab subduction, and continued into the
154 Holocene (Kay and Copeland, 2006; Pallares et al., 2016). In the southernmost part of Payenia, it is generally
155 accepted that the slab steepened towards the present-day dip of 33° (Tassara and Echarren, 2012) by 5 Ma
156 (Kay et al., 2005). Folguera et al. (2009) suggest that volcanism propagated northwards following extensional
157 relaxation of the San Rafael block, while Gudnanson et al. (2012) attribute the northward propagation of
158 volcanism within the Payenia to a northward delay in slab rollback. Despite uncertainty in the chronology
159 and mode of magmagenesis of the SVZ rear-arc, the wide spatial distribution of volcanic products, along with
160 their relatively primitive chemistry, provide vital constraints on the composition of the ambient sub-arc
161 mantle beneath the SVZ (Jacques et al., 2013; Turner et al., 2017).

162 **2. Samples and methods**

163 Six samples were collected from the basal sequence of Maipo Volcano, the active stratovolcano at the centre
164 of the caldera, and nine from Don Casimiro Volcano, a small eroded stratovolcano located ~ 10 km SW of
165 Maipo with activity restricted to preglacial times (Charrier, 1979; Fig. 2b). $^{40}\text{Ar}/^{39}\text{Ar}$ dates indicate that early
166 activity at these centres was contemporaneous (Orozco et al., 2015). All lavas contain ubiquitous olivine and
167 clinopyroxene phenocrysts within a glassy groundmass. The more evolved samples also contain plagioclase,
168 oxides, and orthopyroxene (details in SIA2). Seven samples were collected from monogenetic scoria cones in
169 the northern rear-arc (34.3 to 35.4°S ; Fig. 2a). Olivine, clinopyroxene, and oxide phenocrysts dominate in
170 these scorias, with microcrysts of plagioclase and alkali feldspar, and minor apatite.

171
172 All 22 samples were powdered for whole-rock analysis with an agate ball mill at the University of Oxford.
173 Major element analysis was conducted using a PANalytical Axios Advanced X-ray fluorescence spectrometer
174 at the University of Leicester (Knott et al., 2016). Powders were digested in clean labs at the University of

175 Oxford for ICP-MS analysis. Ba and Sr were measured at a 1,200,000x dilution using a Thermo Element 2 ICP-
176 MS, while all other trace elements were measured at a 6000x dilution on a Perkin Elmer NexION 350D
177 quadrupole ICPMS (both at the University of Oxford). Four certified reference materials (CRMs; BCR-2,
178 BHVO-2, AGV-2, and W-2a) were digested alongside these samples. Data was reduced using calibration
179 curves generated from these 4 CRMs. Repeat solution analysis (n=19) and repeated digestions indicate an
180 analytical uncertainty of <3% (<5% for Cs), and calculated values for the CRMs yield concentrations within
181 $\pm 5\%$ of the preferred values (Appendix 3).

182

183 $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios were analysed for eleven samples at the University of Leeds.
184 Following Sr and Nd purification, samples were loaded onto outgassed W (Sr) and Re (Nd) filaments, and
185 analysed on a Thermo Scientific Triton-series multicollector mass spectrometer. NIST SRM-987, La Jolla and
186 BHVO-2 were analysed throughout the session to monitor accuracy and precision (Appendix 3).

187

188 Olivine grains were analysed for major elements on a Cameca SXFive FEG-electron microprobe at the
189 University of Oxford. St Johns Olivine and Fayalite were analysed as secondary standards to assess accuracy
190 and precision. LA-ICP-MS olivine analysis was conducted at the University of Cambridge on a 193ESI Laser
191 Ablation system coupled to a Nexion ICP-MS. Multiple secondary standards were run to monitor instrument
192 drift, and accuracy and precision. Additional analytical details are provided in SIA3.

193 3. Results

194 The new Don Casimiro-Maipo samples are primitive calc-alkaline basaltic andesites to andesites. The whole-
195 rock Mg#s (assuming $\text{Fe}^{3+}/\text{Fe}_T=0.3$; Holm et al., 2016) of Don Casimiro-Maipo samples mostly range from 60-
196 70, indicating that these samples have undergone only minor crystal fractionation from primary magmas
197 (Fig. 4a). There is little variation in SiO_2 (55.3–57.9 wt.%) within the sample set. Previous studies of the
198 Diamante-Maipo Caldera (Hickey et al., 1986; Futa and Stern, 1988, Sruoga et al., 2005; Holm et al., 2014)
199 have primarily sampled volcanics with significant europium anomalies ($\text{Eu}^* = [\text{Sm}_N \times \text{Gd}_N]^{0.5}$; Fig. 4b),
200 indicating fractionation of plagioclase or mixing between primitive and evolved magmas (e.g. Turner and
201 Langmuir, 2015a). In contrast, our higher Mg# samples have $\text{Eu}/\text{Eu}^* > 0.9$ (Fig. 4b). Olivine phenocryst
202 compositions range from $\text{Fo}_{83.6-75.7}$ at the arc-front, which lie below the olivine-whole rock equilibrium line,
203 indicating that some olivine accumulation has occurred. This demonstrates the importance of utilizing
204 individual mineral analysis in conjunction with whole rock data, as the highest Mg# Maipo lavas (Mg#~70)
205 appear to represent the addition of up to 7wt% olivine to more evolved melts rather than the most
206 “primitive” bulk compositions (Fig. 4a-b; Fig. SB2; Hickey et al., 2016).

207

208 Like most arc magmas, mafic Don Casimiro-Maipo lavas have high abundances of Cs, Rb, Ba, U, Th, Pb, and
209 Sr, and depletions in Nb and Ta relative to other elements of similar incompatibility during mantle melting
210 (Fig. 1b). Highly incompatible trace element abundances of these lavas are elevated relative to typical mafic
211 to intermediate samples from oceanic arcs and the SSVZ. Additionally, compared to other primitive SVZ
212 samples, Don Casimiro-Maipo lavas exhibit lower Cs/Rb, smaller Zr-Hf depletions ($\text{Hf}/\text{Sm} \sim 0.9$; Fig. SB3), and
213 significantly more enriched isotope ratios (higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$; Fig. 3b). Finally, the
214 isotopic compositions of the most primitive Don Casimiro samples from this study have lower $^{87}\text{Sr}/^{86}\text{Sr}$ and
215 higher $^{143}\text{Nd}/^{144}\text{Nd}$ than Maipo lavas and literature samples from throughout the Diamante-Maipo caldera
216 (Fig. 3b).

217

218 Rear-arc scoria samples are basalts-trachybasalts (Fig. 3a; Mg#s of 58–71; assuming $\text{Fe}^{3+}/\text{Fe}_T=0.15$). Olivine
219 core compositions have higher forsterite contents than Don Casimiro-Maipo samples ($\text{Fo}_{90.1-80.6}$ vs. $\text{Fo}_{83.6-75.7}$),
220 and lie close to the equilibrium field (Appendix 3; Fig. SB2a). Trace element patterns in the rear-arc show
221 prominent subduction signatures (e.g. Nb-Ta depletion, Pb enrichment, and Th/Nb up to 0.72), and
222 concentrations even higher than Don Casimiro-Maipo lavas (Fig. 1b). As noted by Holm et al. (2016), rear-arc
223 samples have slight depletions in the high field strength elements Zr and Hf, though such depletions are
224 comparable to those observed within the SVZ arc front (Hf/Sm \approx 0.6; Fig. SB2). $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$
225 ratios of the new rear-arc samples plot between Don Casimiro-Maipo and literature SVZ data, and are mostly
226 offset from the filtered rear-arc literature data to higher $^{87}\text{Sr}/^{86}\text{Sr}$ at a given $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 3b).

227 **4. Discussion**

228 The northward increase in crustal thickness in the SVZ theoretically provides the ideal setting to understand
229 the relationship between primitive magma compositions and crustal thickness. However, the scarcity of
230 geochemical data for mafic lavas in the NSVZ has hindered unbiased assessments of the composition of
231 parental magma compositions, and thus the origin of regional geochemical variations. The more primitive
232 nature of Don Casimiro-Maipo samples investigated in this study provide a novel opportunity to re-evaluate
233 the relative contributions from crustal processing, mantle melting, slab fluxes, subduction erosion, and
234 ambient mantle heterogeneity.

235

236 **4.1. Crustal processing at Don Casimiro-Maipo**

237 Hildreth and Moorbath (1988) suggested that mantle-derived magmas in the SVZ are processed in a melting,
238 assimilation, storage, and homogenization (MASH) zone in the lower crust. In their model, mantle melts
239 throughout the SVZ have similar compositions, and the enriched “baseline” composition of NSVZ magmas
240 results from increased MASH processing in the north, where the crust is thickest. A compilation of samples
241 from the entire Don Casimiro-Maipo caldera (this study and literature) indicates a role for crustal
242 assimilation or mixing during fractional crystallization (AFC) in the petrogenesis of more evolved lavas. For
243 example, the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of Diamante Caldera lavas correlate with MgO and Eu/Eu* (Fig.
244 SB1). These geochemical trajectories can be recreated with a simple model whereby the compositions of
245 sampled crustal lithologies (Lucassen et al. 2001; Lucassen et al., 2004) are mixed in variable proportions
246 with primitive Don Casimiro-Maipo lavas (Fig. 4c-d). The MASH model proposed that the isotopic and trace
247 element enrichment in even the most primitive NSVZ volcanics were caused by crustal assimilation and
248 fractionation of parental magmas that are similar to SSVZ primary magmas. However, the new primitive
249 NSVZ samples, when combined with an SVZ dataset extending from the SSVZ to the NSVZ, clearly
250 demonstrate that the enriched compositions of high Mg# NSVZ lavas cannot be produced by MASH.

251

252 To test the MASH hypothesis, the composition of SSVZ volcanics will be represented by an average
253 composition of mafic samples from Villarrica, which erupts high-Mg# lavas and has a similar trace element
254 and isotopic composition to the other volcanoes of the SSVZ (Fig. 1a, Fig. 3b; Turner et al., 2016). As an initial
255 demonstration, the crustal lithologies which recreate the geochemical trajectories *within* the Diamante-
256 Maipo Caldera were mixed with the composition of mafic Villarrica lavas. These mixing trends produce
257 rapidly decreasing K/Rb ratios prior to producing adequate enrichment in $^{87}\text{Sr}/^{86}\text{Sr}$, and none reach high
258 enough Rb/Y (Fig. 4c-d). Thus, assimilation of these lithologies cannot bridge the compositional gap between
259 the SSVZ and NSVZ.

260

261 To test whether *any* plausible Andean crustal assimilant can bridge the compositional gap between the NSVZ
262 (Don Casimiro-Maipo) and SSVZ (Villarrica), the compositions of 348 basement outcrops and basement
263 xenoliths from the SVZ and the surrounding area (22–46°S) were compiled (Lucassen et al. 2001; Lucassen et
264 al., 2004 and others; Appendix 5). As above, assimilation was modelled by mixing the average composition of
265 mafic Villarrica samples with each basement lithology (for mixing proportions between 0–100%). Only 14%
266 of potential assimilants produce mixed compositions within $^{87}\text{Sr}/^{86}\text{Sr} \pm 0.0001$ and $^{143}\text{Nd}/^{144}\text{Nd} \pm 0.00005$ of
267 the primitive Don Casimiro-Maipo average (Fig. 5a). Of these isotopically plausible mixed compositions, only
268 35% have Eu anomalies within the range of the primitive Don Casimiro-Maipo samples ($\text{Eu}/\text{Eu}^* > 0.9$). None
269 of this subset of mixed compositions have trace element signatures resembling Don Casimiro-Maipo (Fig.
270 5b). Incorporating crystal fractionation alongside these simple assimilation models also cannot account for
271 these compositional offsets, because the resulting compositions produce mismatches not only in overall
272 elemental abundances, but also in a variety of incompatible element ratios that are not fractionated by
273 early-crystallizing phases.

274
275 It remains possible that the required MASH assimilant has evaded sampling within the Southern Andes. To
276 address this possibility, we assess the chemical characteristics that a hypothetical assimilant must possess to
277 bridge the compositional gap between the NSVZ and SSVZ. Assuming a maximum of 20% assimilation, based
278 on the relatively high Mg#s of primitive Don Casimiro-Maipo lavas (assimilation drives cooling and crystal
279 fractionation; DePaolo, 1981), any suitable assimilant must have greater than ~450ppm Zr, ~1420ppm Sr,
280 and ~0.63wt% P_2O_5 . It is improbable that crustal lithologies will possess these characteristics, because these
281 elements typically become depleted, rather than enriched, during late stage crystal fractionation or crustal
282 melting (Turner and Langmuir, 2015a). To illustrate this point, a compilation of all continental granites,
283 diorites, syenites, and monzonites in the GEOROC database (N~3000 with trace element data) was searched
284 for suitable assimilants (considering only trace element abundances). Only 15 samples (0.5% of the entire
285 compilation) possessed sufficient concentrations of Zr, Sr, and P_2O_5 . Mixing of these lithologies with mafic
286 Villarrica lavas produces compositions with erratic trace element patterns, testament to the complex
287 petrological histories of these unusually enriched basement lithologies. None recreate the observed trace
288 element pattern of Don Casimiro-Maipo, particularly the negative Nb-Ta anomaly (Fig. 5c). Thus, although
289 crustal assimilation is common among more evolved samples from the NSVZ, the compositional offset
290 between mafic NSVZ and SSVZ lavas is unlikely to have been generated by assimilation in the lower crust.
291 The failure of the MASH model is most apparent when the full trace element signatures of regional crustal
292 assimilants are considered, along with fundamental considerations regarding the processes causing enriched
293 basement lithologies worldwide.

294
295 The geochemistry of rear-arc samples further reinforces the conclusion that the isotopic offsets between
296 NSVZ and SSVZ samples are not produced by MASH. The trace element and isotopic compositions of all the
297 rear-arc samples collected in this study indicate that the rear-arc mantle has been sporadically infiltrated by
298 slab melts. By filtering a compilation of all published data from SVZ rear-arc monogenetic cones to remove
299 samples with large slab additions, it becomes apparent that rear-arc $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ data plot along
300 the “mantle array” (Fig. 3a; Sørensen et al., 2013; 2015a-b; Sørensen and Holm, 2013; Kay et al., 2013; Jacques et
301 al., 2013; 2014; Holm et al., 2014; Turner et al. 2017). This trajectory is not consistent with assimilation of
302 available basement (e.g. Fig. 5b), because highly primitive samples are available at all points along the array.
303 Additionally, unlike the arc-front samples, many of the primitive rear-arc volcanics bear highly forsteritic
304 olivines, nearly in equilibrium with the mantle (Fig. SB2a). It is difficult to reconcile the presence of these
305 primitive olivines with models invoking large amounts of crustal assimilation (e.g. 70% assimilation; Sørensen

306 et al., 2013), as heat loss during assimilation would drive olivine compositions towards more evolved
307 forsterite contents. Additionally, rear-arc isotopic variability shows similar spatial trends to that observed at
308 the arc-front, despite substantial longitudinal offsets (Fig. 6a-b). Crustal contamination cannot explain this
309 similarity, as accretion and eastward migration of the arc-front since the Palaeozoic has produced a
310 longitudinally fragmented basement, with different basement domains striking parallel to the coast (Kay et
311 al., 2005). Assimilation of lithologies within these different basement domains would be expected to
312 produce uncorrelated rear-arc and arc-front isotope systematics. In contrast, mantle flow paths *are*
313 predicted to travel directly from the rear-arc to arc front, providing a coherent mechanism to account for the
314 synchronized latitudinal changes in arc-front and rear-arc volcanics. In fact, the coincidence of along-strike
315 rear-arc and arc-front geochemical trends is one of the strongest pieces of evidence that the enriched
316 compositions of NSVZ magmas are derived from the mantle rather than the crust. Additional consideration
317 will be given to this observation in section 4.5.

318

319 **4.2 Assessing the slab and mantle melting contributions to SVZ compositional variation**

320 As with other arcs, variable slab components and mantle melting processes may affect the chemistry of SVZ
321 volcanics. The NSVZ mantle wedge is likely to have a lower maximum extent of melting than the SSVZ,
322 because the thickened over-riding plate in the NSVZ forces high wedge temperatures to greater pressures
323 (Turner et al., 2016). This should lead to enrichment of incompatible elements in the NSVZ (e.g. Tormey et
324 al., 1986; Jacques et al., 2014; Turner et al., 2016). In addition to variability in mantle melt extent (F), the
325 trace element compositions of primary arc magmas are influenced by slab materials. For example, Jacques et
326 al. (2014) attribute compositional variability in the SSVZ to varying slab fluxes. To assess whether the
327 compositional differences between the NSVZ and SSVZ can be accounted for by varying F and/or slab
328 components, it is useful to consider the maximum solution space generated by varying these parameters in
329 tandem.

330

331 For this exercise, trace element and isotopic concentrations were calculated using an adapted version of the
332 mantle melting model of Turner and Langmuir (2015a; see SIA4). The compositions of subducted sediment
333 and AOC were taken from the average isotopic and trace element compositions of sediment core ODP1232
334 (Turner et al., 2017) and NMORB (Gale et al., 2013). As numerical models indicate that SVZ slab top
335 temperatures are well above the wet pelite and AOC solidi (Syracuse et al., 2010; Hermann and Spandler,
336 2007; Carter et al. 2015; Jacques et al., 2013; 2014), and only hydrous melt (as opposed to aqueous fluid)
337 eclogite partition coefficients produce successful models of trace element abundances (Turner et al., 2017),
338 it was assumed that slab melts (rather than fluids) were added to the mantle wedge. Slab melt compositions
339 were calculated using the aggregate fractional melting equation and partition coefficients consistent with
340 experiments (SIA4.1; Kessel et al., 2005; Hermann and Rubatto, 2009; Skora and Blundy, 2010). Slab melts
341 were then mixed with a depleted upper mantle composition assigned the trace element abundances of
342 Workman and Hart (2005), and the isotopic composition of the rear-arc sample 126171 (Baseline Andean
343 Mantle (BAM); Figs 3a, 6a-b; Sjøger et al., 2013). This isotopic composition defines the low Sr and high Nd
344 end-member of the rear-arc array (Fig. 3a) and has a trace element composition that appears free of slab
345 material (e.g. Th/Nb<0.06, Ce/Pb>25). Finally, the composition of mantle melts were estimated using the
346 batch melting equation, with experimentally-derived hydrous peridotite partition coefficients and melt
347 reaction coefficients (model details in SIA4.1).

348

349 Firstly, we consider the solution space for $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$, which is simplified by the fact that these
350 trajectories are not noticeably fractionated by variations in slab temperature, and are insensitive to varying

351 melt extents. The pink field on Fig. 7a shows the solution space produced by variations in slab temperature
352 and the proportions of slab melts. The isotopic composition of Villarrica (and most SSVZ) volcanics is
353 encompassed within this field. However, TSVZ and Don Casimiro samples plot significantly below this field.
354 As variable slab contributions predominantly drive wedge compositions to higher $^{87}\text{Sr}/^{86}\text{Sr}$ at near constant
355 or increasing $^{143}\text{Nd}/^{144}\text{Nd}$, another process is clearly needed to recreate the low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in TSVZ
356 and NSVZ volcanics.

357
358 A similar solution space for trace element ratios is shown in Fig. 7b. The pink area encompasses the space
359 made available by varying the proportions of a slab melts (with slab melting at ~ 800 C) and F within
360 reasonable bounds. $F=0.04$ was chosen as the lowest plausible mantle melting extent because F values <0.04
361 tend to produce alkali-basalt or silica-undersaturated major element compositions (e.g. Baasner et al., 2016).
362 Villarrica samples plot within this field, and a model fit can be found that reproduces the entire suite of
363 Villarrica incompatible trace elements and radiogenic isotope ratios (Figs. 7-8). While this model fit does not
364 represent a unique solution for Villarrica, it does demonstrate that a feasible quantitative solution exists that
365 is consistent with available experimental constraints. As in Fig. 7a, the compositions of most TSVZ and Don
366 Casimiro-Maipo lavas fall outside of this solution space. The addition of sediment and AOC melt to the
367 depleted mantle source can generate high La/Sm, but only at excessively high Sr/Nd (Fig. 7a). While reducing
368 F can increase La/Sm, even at $F=0.04$ there is no plausible trace element match to the Don Casimiro-Maipo
369 lavas within this parameter space.

370
371 Slab temperatures might also affect the trace element compositions of SVZ volcanics. There are systematic
372 changes in the slab age and depth beneath the SVZ arc front, both of which may lead to variability in the
373 temperature of the slab surface beneath the arc front (Syracuse et al., 2010). Certain aspects of the trace
374 element differences between Don Casimiro-Maipo and Villarrica are indeed suggestive of slab temperature
375 variations. Compared to Villarrica, Don Casimiro-Maipo has low Cs/Rb and high Hf/Sm ratios (Fig. 1b). These
376 element ratios are not substantially fractionated from each other during melting, and may indicate higher
377 slab temperatures in the NSVZ. For example, Cs is highly mobile in low temperature aqueous fluids, and may
378 have been disproportionally lost during early slab dehydration in the NSVZ, where the sub-arc slab is deeper
379 (Savov et al., 2007; Spandler et al., 2007). Additionally, a hotter slab temperature can destabilize zircon
380 during slab melting, which may lead to the less pronounced Zr-Hf depletions at Don Casimiro-Maipo (e.g.
381 Hirai et al., 2018). The blue field on Fig. 7b shows the expansion of the solution space for slab temperatures
382 hotter than 900°C (calculated using the maximum measured experimental mobility of light rare earth
383 elements in slab melts; see SIA4.1). While the solution space at very high slab temperatures may shift toward
384 higher La/Sm ratios at a given Sr/Nd ratio, the mafic NSVZ volcanic rocks remain well outside of the realm of
385 possible solutions. Overall, the trace element and isotopic differences between mafic SSVZ and NSVZ
386 volcanics cannot be recreated by variable slab additions and melt extents, even in combination. This
387 indicates that the sub-arc mantle within the SVZ must be variably enriched to account for the prominent
388 latitudinal variations in geochemistry.

389

390 **4.3 Is there geochemical evidence for subduction erosion?**

391 SVZ rear-arc volcanics form a linear array between MORB-like compositions and EM1-type OIBs in $^{87}\text{Sr}/^{86}\text{Sr}$
392 vs. $^{143}\text{Nd}/^{144}\text{Nd}$ space (Kay et al., 2013; S ager et al., 2015a; Fig. 3a), and both rear-arc and arc-front samples
393 are characterized by similar along-strike variability in these isotope ratios (Fig. 6a-b). As the isotopic offset
394 between the mafic NSVZ and SSVZ samples is inconsistent with crustal assimilation or variable slab additions,
395 an alternative mechanism is required to introduce an EM1-like contaminant to the mantle source of the SVZ.

396 The origins of EM1-type OIBs are debated, with suggestions including deep recycling of lower continental
397 crust (LCC), pelagic sediment and oceanic crust, or metasomatized subcontinental lithospheric mantle
398 (Willbold and Stracke, 2010). The first suggestion is particularly apt, as it has been argued that the NSVZ
399 endured extensive subduction erosion (the removal of upper plate material via abrasion and plucking by the
400 lower plate) throughout the Cenozoic (Stern, 1989; Stern, 2011). Although the composition of the Andean
401 LCC is not well constrained, an extensive compilation of regional basement compositions (Appendix 5)
402 constrains the compositional range of this reservoir. While these samples were collected from surface
403 outcrops, Miocene episodes of tectonic thickening and under-thrusting have resulted in similar lithologies
404 residing at deeper levels in the crust (Hildreth and Moorbath, 1991). Hence, this compilation can be used to
405 assess the subduction erosion hypothesis, regardless of whether the UCC, LCC or the entire crustal column
406 (Holm et al., 2014; 2016) on the leading edge of the South American plate undergoes recycling.

407
408 We assess the ability of subduction erosion to reproduce along-arc geochemical trends by mixing the
409 composition of basement lithologies (Appendix 5) into the estimated composition of the metasomatized
410 Villarrica mantle source. Melts of the resulting mixes were calculated using the model described above at
411 melt generation conditions indicated by thermal models for Don Casimiro-Maipo ($F=0.056$, $P=33\text{kbar}$; Turner
412 et al., 2016). Only 11% of the mixed compositions pass within $^{87}\text{Sr}/^{86}\text{Sr}\pm 0.0001$ and $^{143}\text{Nd}/^{144}\text{Nd}\pm 0.00005$ of
413 the most primitive Don Casimiro-Maipo samples (Fig. 9a). Only 15% of these isotopically plausible mixed
414 compositions have Eu anomalies within the range of the primitive Don Casimiro-Maipo samples
415 ($\text{Eu}/\text{Eu}^* > 0.9$), none of which recreate the concentrations of Y, Gd, Sm, Sr, Ba and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Fig. 9b-c).
416 This analysis demonstrates that no mixed composition with adequate trace element data plots within the
417 range of the Don Casimiro-Maipo for $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$, and trace element abundances.

418
419 While some previous studies of the NSVZ (e.g. Holm et al., 2014; 2016) have concluded that eroded
420 continental material is the most plausible candidate for the enriched mantle end-member in this region, our
421 analysis does not support this conclusion. This discrepancy results from the more primitive nature of the
422 newly sampled Don Casimiro-Maipo samples. The low Eu/Eu^* ratios seen in samples from other studies may
423 well result from plagioclase fractionation, magma mixing, or crustal assimilation, implying that this signature
424 is a secondary overprint, rather than a source characteristic.

425
426 As with crustal assimilation during magmatic ascent, it is possible that the appropriate eroded crustal
427 lithology has avoided sampling due to the restricted basement exposure in the SVZ. However, regional
428 considerations indicate that LCC recycling is unlikely to produce the observed EM1-like enrichment. Willbold
429 and Stracke (2010) present a model in which EM1-like isotopic compositions are produced by mixing LCC
430 starting materials into normal MORB mantle. However, this model assumes that the recycled LCC domains
431 have been extracted and isolated from the mantle for >4 Ga, while the Chilenia and Cuyania terranes
432 comprising the lower crust of the NSVZ-TSVZ are only of Grenvillian age ($\sim 1\text{Ga}$; Ramos, 2010). This relatively
433 young age causes typical Southern Andean basement to have $^{187}\text{Sr}/^{86}\text{Sr}$ ratios that are too high (relative to
434 $^{143}\text{Nd}/^{144}\text{Nd}$) to serve as an appropriate end-member (see mixed compositions in Fig. 9a).

435
436 Rear-arc geochemical variability is also inconsistent with the subduction erosion hypothesis, because rear-
437 arc isotopic enrichment persists at distances exceeding 600 km from the trench. It is unlikely that material
438 eroded at the trench could spread hundreds of kilometres towards the rear arc, in the opposite direction of
439 corner flow (MacDougall et al., 2017). While eroded material may instead be carried down with the
440 subducting slab and released alongside other slab inputs, in this case the enriched signature would co-vary

441 with other slab components. Instead, at 550 km from the trench, rear-arc monogenetic samples have Ce/Pb
442 ratios that range from 5 (similar to the arc front) to >20 (similar to MORBs; Fig. 10b), but retain $^{143}\text{Nd}/^{144}\text{Nd}$
443 ratios that are consistently enriched relative to DMM (Fig. 3a). Additionally, within a given latitude bracket,
444 rear-arc $^{143}\text{Nd}/^{144}\text{Nd}$ ratios have limited variability, and do not correlate significantly with Ce/Pb or other
445 indices of slab additions. Finally, isotopic enrichment in both the rear arc and arc front increases in a
446 reasonably symmetrical, coherent fashion on either side of a pronounced minima at 39° S (Fig. 6a-b; Turner
447 et al., 2017). Although it has been suggested that subduction erosion increases northwards (Stern, 1989),
448 this phenomenon cannot account for the increase in enrichment south of Villarrica. The spatial systematics
449 of rear-arc compositions are therefore also inconsistent with subduction erosion.

450
451 Overall, when the full array of trace element abundances and isotopic ratios are considered, subduction
452 erosion cannot recreate the notable geochemical trends in rear-arc and arc-front data. This may indicate
453 that the extent of subduction erosion has been over-estimated for this region. Erosion rates are calculated
454 based on the assumption that the distance between the trench and the arc front remained constant
455 throughout the Miocene (Stern, 1989). However, more recent geophysical models find that arc-front
456 migration relative to trench location might be a common process (e.g. Karlstrom et al., 2014). The suggestion
457 that the locus of subduction-related volcanism has shifted during this same period as a function of varying
458 slab angle also undermines a direct connection between subduction erosion and arc-front location.
459 Alternatively, it is possible that subduction erosion did indeed influence the mantle wedge composition at
460 some point in the Miocene, but corner flow towards the arc has since flushed the present-day mantle wedge
461 of eroded material. Regardless of the explanation, the plausible geochemical impacts of subduction erosion
462 are inconsistent with the along-strike enrichment trend of the SVZ. Clearly some other mechanism is
463 required to explain SVZ geochemical variability.

464 465 **4.4 EM1-type enrichment from a sub-continental lithospheric mantle source**

466 An alternative source for the enriched EM1-like mantle signature of the NSVZ is the recycling of
467 Metasomatised Subcontinental Lithospheric Mantle (M-SCLM). The infiltration of low degree mantle melts
468 ($F \sim 0.005$) into the base of the lithospheric mantle produces enriched incompatible element concentrations,
469 which rapidly evolve EM1-like isotopic compositions (McKenzie and O’Nions, 1995; Turner et al., 2017; Fig.
470 11). Unlike crustally-derived materials, M-SCLM provides a more homogenous end-member composition
471 that is capable of recreating the remarkably coherent geochemical trends observed in the SVZ (Rogers and
472 Hawkesworth, 1989; Turner et al, 2017). M-SCLM material may be stored for long periods in the lithospheric
473 mantle until it is returned to the asthenosphere by delamination or erosion driven by corner flow.

474
475 Field evidence supports the hypothesis that the SCLM surrounding the SVZ has EM1-like isotopic affinities.
476 Three geographically separated suites of isotopically enriched igneous rocks across South America have been
477 interpreted as direct melts of the enriched SCLM (see Fig. 2a):

- 478 1) Alkaline igneous rocks erupted near the edges of the Rio Apa-Luis Alves and Sao Francisco Cratons
479 (Gibson et al., 1995; 2005; Carlson et al., 1996).
- 480 2) Mesozoic metabasites from the Southern Rift (Lucassen et al., 2002).
- 481 3) Carboniferous granitoids from the Santo Domingo Complex of the Coastal Batholith (Parada et al.,
482 1999).

483 The $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ of samples from these regions plot as a linear extension of the SVZ rear-arc
484 array (Fig. 11) and cannot be accounted for by addition of variable slab components or assimilation of
485 available basement. The remarkable alignment of these lithologies in isotopic space, despite being

486 geographically separated by hundreds to thousands of kilometres, suggests that M-SCLM east and north of
487 the SVZ has relatively homogeneous EM1 affinities. The isotopic trajectory of these SCLM melts is consistent
488 with the expected isotopic evolution of SCLM enriched by low degree mantle melts (Turner et al., 2017; Fig.
489 11). In contrast, the field produced by addition of sampled basement lithologies to the mantle has higher
490 $^{87}\text{Sr}/^{86}\text{Sr}$ at a given $^{143}\text{Nd}/^{144}\text{Nd}$ than the array of SCLM-melts and rear-arc volcanics, and a significantly
491 broader trajectory (grey field, Fig. 11).

492
493 Rather than attempt to estimate the full trace element composition of very low-degree metasomatising
494 mantle melts, a proxy for the composition of mantle domains which has been enriched by M-SCLM melts
495 was estimated by inverting for the mantle source composition of Gough Island EM1-type OIB volcanics
496 (Turner et al., 2017), where enrichment has been specifically linked to deep recycling of SCLM (Gibson et al.,
497 2005; Boyet et al., 2019). Trace element and isotopic ratios of the inverted source composition were
498 subsequently adjusted within the limits of EM1-like ocean island basalts (Stracke et al. 2003; Willbold and
499 Stracke, 2010) to produce a generic EM1 source (SIA4.2). Including EM1-like mantle heterogeneity as an
500 additional parameter sufficiently expands the solution spaces on Fig. 7a-b (grey fields) to overlap the
501 composition of primitive Don Casimiro-Maipo lavas. Furthermore, a model fit incorporating this EM1-type
502 mantle component recreates the isotopic composition and entire suite of incompatible trace element
503 concentrations at Don Casimiro-Maipo (Fig. 7–8). As with the model fit for Villarrica, this does not represent
504 a unique solution, but does demonstrate that the observed geochemical enrichment at Don Casimiro-Maipo
505 is quantitatively consistent with all available constraints, unlike models relying upon MASH, LCC recycling, or
506 variable slab components.

507
508 The addition of M-SCLM to the mantle by erosion or delamination at the base of continental cratons behind
509 the volcanic arc also accounts for the spatial variability in rear-arc and arc-front geochemistry. Turner et al.
510 (2017) proposed two explanations for the gradual decline in isotopic enrichment towards a distinct minimum
511 at 39°S (Fig. 6a-b). Firstly, they note the presence of a Permian terrane suture at 39° S (Rapalini et al., 2010).
512 It is possible that M-SCLM was lost during rifting or subsequent collision between cratonic blocks (Fig. 12a).
513 Secondly, a vertical and horizontal slab tear has been inferred based on seismic imaging at ~39° S, (Pesicek et
514 al., 2012). If M-SCLM is entrained equally throughout the SVZ, upwelling of depleted Pacific mantle (which
515 has not interacted with M-SCLM) through these features could dilute the EM1 signature. The upwelling,
516 depleted mantle would then spread northwards and southwards, perhaps aided by toroidal flow (Zandt and
517 Humphreys, 2008), resulting in a gradual northward and southward increase in enrichment (Fig. 12b). This
518 upwelling material could also produce the magnetotellurically imaged "plume" in the rear-arc (Burd et al.,
519 2014). Geochemical signatures associated with upwelling through a slab window in the Kula Volcanic Field
520 (W. Anatolia, Turkey) are traced over >250km (Klaver et al., 2016), similar to the distances observed here.

521

522 **4.5 A unified model for the SVZ arc-front and rear-arc**

523

524 The preceding sections have shown that the first-order geochemical variability in the SVZ arc-front and rear-
525 arc is consistent with the mixing of EM1-type and isotopically depleted mantle end-members. This is similar
526 to the proposal of Jacques et al. (2013, 2014), though by including the influence of ambient mantle
527 enrichment on trace element abundances, our model does not require substantial variations in the
528 composition of slab components, or exceedingly low extents of melting to account for arc-front
529 compositions. In contrast to these unified models of arc-front and rear-arc geochemistry, other models
530 calling upon unique processes specific to particular regions of the rear-arc have been proposed. For example,

531 Søger et al. (2015a; b) invoke the presence of a distinct mantle component that is specific to Southern
532 Payenia, Søger et al. (2013, 2015a) propose that rear-arc magmas have assimilated up to 70% LCC to
533 accommodate certain trace element ratio trajectories in basalts from the Nevado volcanic field (~35-36°S),
534 and Holm et al. (2016) invoke enriched mantle and eroded crust to account for the geochemical variation
535 observed in the Northern Segment of the rear-arc (~34-35 °S). To assess whether these region-specific
536 observations can be accounted for by our unified model, it is important to consider the detailed
537 characteristics of across-arc geochemical trends and the influence of temporal variability in SVZ subduction
538 geometry.

539
540 The extent to which slab materials have infiltrated the rear-arc mantle can be investigated using trace
541 element ratios sensitive to the addition of slab melts (e.g. Th/Nb, Th/La, and Ce/Pb; Johnson and Plank,
542 1999; Plank, 2005; Fig. 10a). Remarkably, while these slab-sensitive ratios generally show declining slab
543 fluxes with increasing trench distance, some cones retain elevated Th/La and Th/Nb, and low Ce/Pb at nearly
544 600km from the trench (Fig. 10b). Variations in slab temperature should significantly fractionate these ratios.
545 If across-arc slab temperature variations are in accord with thermal models (Syracuse et al., 2010), then
546 Th/La in slab melts may increase by ~25% from the arc-front to the rear-arc, and Th/Nb by nearly an order of
547 magnitude (Hermann and Rubatto, 2009). This variability in the composition of slab melts should result in
548 rear-arc and arc-front lavas following distinct mixing trajectories. Yet, all rear-arc samples lie on mixing lines
549 between the modelled arc-front slab melt composition (e.g. Fig. 8) and an ambient mantle with varying
550 amounts of EM1-like enrichment (Fig. 10a). The high slab temperatures beneath rear-arc cones should also
551 destabilize zircon (e.g. Harai et al., 2018). However, rear-arc samples have Zr-Hf anomalies similar to TSVZ
552 and SSVZ arc-front volcanoes (Fig. SB3). These observations imply that the slab components reaching the
553 rear-arc and arc-front mantle sources were extracted from the slab at similar temperatures.

554
555 It is possible that the signal of relatively cold slab melts beneath the rear arc is a remnant from a period of
556 shallow subduction, and not related to the present-day slab geometry (Søger et al., 2013), though it is
557 difficult to reconcile the chronology of such a process. In Southern Payenia, it is generally accepted that the
558 slab steepened towards its current configuration at ~5 Ma (Kay et al., 2005). However, Gudnanson et al.
559 (2012) attribute the northward propagation of quaternary rear-arc volcanism to a significant delay in slab
560 rollback in this region, while Folguera et al. (2009) suggest that the recent volcanism behind the NSVZ is
561 instead due to extension collapse. In considering these competing theories, an essential observation is that
562 volcanism within the Diamante-Maipo caldera has been essentially uninterrupted since the Late Miocene
563 (Sruoga et al., 2005). This indicates that the NSVZ slab geometry is unlikely to have shifted substantially over
564 the past 5 Ma. To date, no physical model of late slab steepening has been proposed which satisfies
565 observations from both the arc-front and rear-arc. Because the rear-arc samples of this study were likely
566 erupted between 0.01–0.7 Ma (Folguera et al., 2009), corner flow over the 4-5 Myrs following slab
567 steepening should have led to ~330 km of wedge turnover prior to their eruption, effectively flushing any
568 Miocene slab inputs from the rear-arc asthenosphere.

569
570 Thus, the slab component observed in northern rear-arc cones was likely derived from the slab in its current
571 geometry, and some other mechanism is required to account for the uniform across-arc slab-component
572 compositions. The results of two-phase thermo-mechanical models provide an explanation for this
573 phenomenon. The mantle source at the arc front is likely influenced by slab materials that have been
574 extracted from the slab well beyond the arc front, and then deflected towards the trench along compaction
575 pressure gradients (Wilson et al., 2014). The rear-arc asthenosphere is plausibly influenced by slab melts

576 extracted at shallower depths and then transported away from the trench in zones of low porosity (Cerpa et
577 al., 2017; Fig. 14b). It also is possible that slab materials could be transported non-vertically by “mélange
578 diapirs”. However, Cruz-Urbe et al. (2018) found that melts of such diapirs have >50 wt. % SiO₂, >18 wt. %
579 Al₂O₃, and <8 wt. % MgO, while the most primitive SVZ rear-arc basalts have >10 wt. % MgO, <48 wt. % SiO₂,
580 and <15 wt. % Al₂O₃, as is typical of arc alkali basalts generated by low degrees of melting from a lherzolite
581 source (e.g. Baasner et al., 2016). The experimental melts also have positive Zr-Hf anomalies, unlike the
582 negative Zr-Hf anomalies of SVZ rear-arc lavas (Fig. 1b). Overall, the geochemical systematics of SVZ rear-arc
583 volcanics appear to be most consistent with a strong role for slab melting followed by advective transport,
584 compaction channelling, and generally non-vertical transport of slab-derived materials through the wedge,
585 which is broadly consistent with numerical models of two-phase flow (Cagnioncle et al., 2007; Wilson et al.,
586 2014; Cerpa et al., 2017).

587
588 Given the general viability of the unified SVZ model of ambient mantle, slab, and melting variation for both
589 the rear-arc and arc front, it is worthwhile to assess whether this model can also account for the
590 observations that motivate region-specific models. Here, we limit this discussion to trends (Figure 13b-d)
591 that have been interpreted as evidence for extensive lower crustal assimilation (up to 70%; Sjøager et al.,
592 2013; Sjøager et al., 2015a). While rear-arc compositions on these plots appear to trend toward the
593 compositions of certain lower crustal xenoliths (Kay et al., 1996), calculated mixing lines with these xenoliths
594 are strongly hyperbolic, and deviate substantially from the actual rear-arc compositions (Figure SB5).
595 “Adjusted” lower crust compositions have been proposed to mediate such discrepancies (Sjøager et al.,
596 2013), though the resulting compositions resemble neither generic models of the LCC (e.g. Rudnick and Gao,
597 2003), nor the actual compositions of Andean basement samples (Figure SB6). These observations, along
598 with the general difficulty of maintaining high Mg# bulk compositions while enduring such large amounts of
599 crustal assimilation, motivate a re-examination of the assimilation model in particular.

600
601 Our analysis suggests that the regional systematics identified by Sjøager et al. can be fully accounted for by
602 the variations expected from slab fluxes, mantle melting extents, and the ambient mantle. On Figure 13c,
603 both axes are controlled by mantle and slab components, while the y-axis (La/Sm) is also affected by
604 variations in F. The arc-front volcanics have a relatively uniform slab contribution, so follow a trajectory
605 determined by F and the ambient mantle (e.g. Turner et al., 2017). Many of the rear-arc samples overlap the
606 arc-front data, but when slab additions are diminished, the extent of mantle melting often drops to very low
607 values (Jacques et al., 2013), which drives up La/Sm. This is apparent, from the similarly high La/Sm
608 compositions of basalts from the Gough Island chain, which are also alkali basalts produced by low-F
609 melting. Similar systematics can account for variability on Figure 13d, though the divergence between arc-
610 front and rear-arc compositions is less apparent here, as both low-F melt generation and bulk slab additions
611 skew compositions similarly, toward very high Th/Sm. Thus, the variability on both 13c and 13d are easily
612 accommodated without a crustal assimilate.

613
614 Unlike Th/Sm and La/Sm, the highly-incompatible element ratios on Figure 13b are less affected by
615 variations in melt extent, and can be accounted for by variations in slab flux and mantle enrichment alone.
616 Ba/Th is particularly sensitive to small variations in the compositions of subducting slab materials and
617 temperatures of slab melting (e.g. Carter et al., 2015; Patino et al., 2000). This is evident from the high Ba/Th
618 mafic enclaves erupted at Longavi Volcano, which are thought to be representative of parental magmas
619 (Rodriguez et al., 2007). Rodriguez et al. demonstrate that these lavas reflect input of a particularly high
620 Ba/Th slab component to the mantle, and suggest this is caused by the proximal subduction of the Mocho

621 Fracture Zone (Figure 13a). Hydrothermal alteration, which was likely enhanced along the fracture zone,
622 produces elevated Ba abundances in both ocean crust and sediment (Kelley et al., 2003; Plank et al., 2013).
623 Alternatively, the fracture zone may introduce an additional high-Ba fluid component to the wedge.
624 Subducting fracture zones have inevitably swept along the arc over time, causing transient geochemical
625 impacts of such extremely sensitive slab tracers in local volcanics (Figure 13a). Given the large range in the
626 Ba/Th ratio of the slab component, there is no need to invoke extremely large quantities of lower crustal
627 assimilation (>70%) to account for the regional systematics. While geochemical variability, when considered
628 on a small enough scale, necessarily requires case-specific ad-hoc modelling, the regional trends of the SVZ
629 appear consistent with the relatively simple three parameter model developed here, which incorporates
630 processes seen in subduction zones worldwide.
631

632 **5. Conclusions**

633
634 The origin of trace element and isotopic enrichment in thick-crustal continental arcs has profound
635 implications for the formation of the continental crust and the petrogenesis of evolved magmas. The new
636 high Mg# samples characterized here provide valuable new constraints on the compositions of parental
637 magmas from the thick-crustal northern SVZ (NSVZ). These new analyses demonstrate that while evolved
638 NSVZ underwent crustal assimilation during fractional crystallization, the geochemical offsets between the
639 most primitive Southern SVZ (SSVZ) and NSVZ volcanics cannot be produced assimilation of any basement
640 lithologies either within the Southern Andes, or globally. Similarly, while variations in the extent of mantle
641 melting and subtle differences in slab additions may arise due to the thicker lithosphere and deeper slab in
642 the NSVZ, the solution space made available by these parameters does not encompass the trace element or
643 isotopic compositions of mafic Don-Casimiro Maipo samples. Only models incorporating an enriched
644 ambient mantle domain that is compositionally similar to the mantle source of EM1-type OIBs can recreate
645 the observed isotopic and trace element enrichment in mafic samples from Don Casimiro-Maipo. Coincident
646 arc-front and rear-arc regional variations in isotopic enrichment indicate that this signal arises from the
647 ambient mantle. A forward model based on an inversion for an EM1-like mantle source successfully
648 reproduces the isotopic composition and trace element composition of Don-Casimiro Maipo, demonstrating
649 that this interpretation is consistent with available experimental constraints.
650

651 The SVZ EM1 component is most plausibly generated by melt-metasomatism of sub-continental lithospheric
652 mantle, which is eventually returned to the asthenosphere by erosion at the base of cratonic blocks,
653 transported trench-ward by corner flow. An alternative scenario in which the EM1 signal results from
654 subduction erosion of continental material produces unsuitable isotopic systematics and erratic trace
655 element patterns. Additionally, the coherent latitudinal variations observed in arc-front and rear-arc lavas
656 are difficult to reconcile with subduction erosion of fragmented basement domains.
657

658 The isotopic trajectory of the rear-arc lavas is co-linear with other South American lavas that are thought to
659 be melts of metasomatized sub-continental lithospheric mantle (M-SCLM). This supports a model in which
660 the enriched mantle source is derived from the addition of M-SCLM material to the mantle wedge by erosion
661 or delamination from the cratonic lithosphere behind the arc. Unlike the longitudinally fragmented and
662 highly heterogeneous Andean basement, the composition of the M-SCLM appears to be relatively
663 homogenous across a large area of South America. Thus, unlike subduction erosion or crustal assimilation,
664 variable mantle enrichment due to mixing of an EM1-like NSVZ mantle source with a depleted mantle

665 component can account for along and across-arc geochemical variations. In addition to chemical variability
666 induced by variable M-SCLM addition, rear-arc cones receive highly variable slab inputs that are surprisingly
667 similar in composition to the arc front. This, along with variable melt extents, accounts for the offset
668 trajectories of rear-arc and arc-front lavas, and provides support for recent two-phase flow models
669 indicating non-vertical transport of slab materials (Cagnioncle et al., 2007; Wilson et al., 2014; Cerpa et al.,
670 2017).

671
672 Overall, we demonstrate that the enriched trace element and isotopic compositions of primitive lavas in the
673 thick-crustal NSVZ are produced by low-degree melting of an enriched ambient mantle that is
674 metasomatized by slab melts (Turner et al., 2017), rather than crystal fractionation and crustal assimilation.
675 The similarities between the trace element signatures of primitive NSVZ lavas and the bulk continental crust
676 (Fig. 1) implies that ambient mantle enrichments may play a major role in the origin of enriched continental
677 compositions globally. Reevaluation of enriched signatures in primitive lavas of other thick-crustal arcs (e.g.
678 Cascades, Mexico, Guatemala, Colombia) will allow assessment of the global importance of ambient mantle
679 enrichment. If similarities with the SVZ are found, models of continental crust generation and elemental
680 cycling within the silicate earth will require re-evaluation.

681 682 **Acknowledgments**

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686 isotopes in Leeds.

687
688

689 **Figure Captions**

690
691 Fig. 1– Incompatible trace element “spider diagrams”, normalized to D-MORB (Gale et al., 2013). a)
692 Literature data for Don Casimiro-Maipo (Hickey et al., 1986) has incompatible trace element abundances
693 comparable to bulk continental crust (Rudnick and Gao, 2003) and the Northern Cascades (Mt Ranier - Sisson
694 et al., 2014; Mt Hood; Baggerman et al. 2011). Don Casimiro-Maipo shows significantly more enriched trace
695 element abundances than the TSVZ, SSVZ, and the majority of island arcs (GEOROC compilation). Villarrica,
696 which has a similar trace element signature to other SSVZ centres (Llaima, Puyehue and Osorno shown here;
697 Turner et al. 2016), is used as a representative SSVZ end-member in figures 3, 4, 6, 8 and 9. b) Trace element
698 data collected in this study for 7 rear-arc cones, and the most primitive samples from Don Casimiro-Maipo.
699 The observed HREE depletion in Don Casimiro-Maipo samples indicates the presence of residual garnet
700 during mantle melting. Rear-arc samples show distinctive arc-like signatures, and even greater trace element
701 enrichment than Don Casimiro-Maipo samples.

702
703 Fig. 2-Geographical context and sample locations. a) Map of the SVZ, showing the segment boundaries
704 defined by Dungan et al. (2001). The locations of arc-front and rear-arc samples of this study are shown,
705 along with the locations of arc-front volcanoes passing the filters of Turner et al. (2016). The geographical
706 regions containing inferred M-SCLM melts (see Fig. 11) are overlain. Base map from GEOMAP APP. b) Sample
707 locations of Don Casimiro and Basal Maipo lavas within the Diamante-Maipo Caldera Complex. Imagery from
708 Google Earth. c) Moho depth (from the model of Tassara and Echaurren, 2012) vs. volcano latitudes

709 projected along the vector of convergence to the trench. d-e) Incompatible major and trace element 6-values
710 (Turner et al. 2016, and this study) increase northwards, mirroring changes in crustal thickness.

711

712 Fig. 3- a) SVZ rear-arc samples which are not dominated by slab inputs ($Th/Nb < 0.4$) plot along the “mantle
713 array”, between DMM and EM1-type ocean island basalts. They greatly extend the isotopic range observed
714 among average rear-arc compositions from other arcs (black dots; Turner et al., 2017). Literature values
715 from Don Casimiro-Maipo lie at the enriched end of the global rear-arc array (Hickey et al., 1986). The pink
716 and blue fields envelop all EM1 and EM2-type ocean island basalts from Stracke et al. (2003). The enriched
717 isotopic end-member used for quantitative modelling is taken from Pitcairn (57DS9; Stracke et al., 2003). The
718 Baseline Andean Mantle (BAM) isotopic end member is taken from an SVZ rear-arc sample with no trace of a
719 slab contribution (e.g. $Th/Nb < 0.06$, $Ce/Pb > 25$; Sørensen et al., 2013; sample 126171). The Nazca Plate
720 sediment composition is the average of ODP1232 (Turner et al., 2017). b) Don Casimiro and Maipo and rear-
721 arc data from this study, alongside literature data from Villarrica and the SVZ that pass filters indicating they
722 are not substantially affected by crustal contamination or mixing (Turner et al., 2016). The filtered rear-arc
723 samples shown in a) are also plotted, along with unfiltered data from the entire Diamante Caldera.

724

725 Fig. 4- a) Don Casimiro-Maipo samples from this study are high Mg# basaltic andesites. Circled samples have
726 major element characteristics consistent with olivine accumulation (Fig. SB2). Rear-arc samples are primitive
727 basalts-trachybasalts. Mg#s were calculated assuming $Fe^{3+}/Fe_T = 0.3$ for the arc-front (Holm et al., 2016), and
728 $Fe^{3+}/Fe_T = 0.15-0.3$ for the rear-arc (Sørensen et al., 2015b, cyan and blue triangles respectively). b) Eu
729 anomalies are negligible in the most primitive samples of this study (> 0.9), whereas the majority of literature
730 samples show substantial negative Eu anomalies. Eu/Eu^* was calculated with $Eu^* = (Sm_N \times Gd_N)^{0.5}$ or with
731 $Eu^{**} = (Sm_N^2 \times Tb_N)^{1/3}$ where no Gd data was reported. In both cases, element concentrations were
732 normalized to CI chondrite (McDonough and Sun, 1995). c-d) Mixing arrays between eight SVZ basement
733 samples and primitive Don Casimiro-Maipo samples encompass the chemical diversity within the Diamante
734 Caldera (from Lucassen et al. 2001; 84-31-1, 84-31-2, 84-42-1, 84-42-11, and Lucassen et al. 2004; 00-77, 00-
735 55, 00-58 and 00-05). However, mixing between the same 8 samples and Villarrica (grey lines) do not pass
736 through Don Casimiro-Maipo. Models originate from the primitive average of Don Casimiro-Maipo and
737 Villarrica samples. Error bars in c) show $\pm 1\sigma$ of the variation among primitive samples.

738

739 Fig. 5-Crustal assimilation models. a) Mixing lines produced between 348 crustal lithologies thought to be
740 representative of SVZ basement (compiled from GEOROC and other sources) and the average composition of
741 primitive Villarrica samples. Only a small proportion pass near Don Casimiro-Maipo. b) Trace element
742 patterns of mixed compositions that have isotopic ratios close to Don Casimiro-Maipo (red lines in a), as well
743 as $Eu/Eu^* > 0.9$. Mixtures providing a reasonable match to the La-Ce concentrations (cyan lines) have trace
744 element patterns that are completely distinct from that of Don Casimiro-Maipo. c) Mixed compositions for
745 the small subset of the ~3000 plutonic rocks from GEOROC that have sufficient concentrations of Zr, Sr and
746 P_2O_5 to recreate the enrichment of these elements at Don Casimiro-Maipo via assimilation. Mixed
747 compositions were calculated for the % mixing required to recreate Zr concentrations. All produce erratic
748 trace element patterns that are dissimilar to those observed at Don Casimiro-Maipo.

749

750 Fig. 6 - Across and along-arc variation in rear-arc isotopes. a-b) Rear-arc isotopic ratios form latitudinally
751 coherent trends, with statistically significant trend lines ($p < 0.005$). A pronounced minimum in $^{87}Sr/^{86}Sr$, and a
752 maximum in $^{143}Nd/^{144}Nd$ is observed at $\sim 39^\circ S$, with a reasonably linear and symmetric decrease to the north
753 and south that can be explained by variable EM1 enrichment (pink arrows). Scatter about this first order

754 trends within the rear-arc results from variable slab fluxes. The blue bar shows the bulk slab compositions
755 for a mix of 20% sediment melt, 80% AOC melt, incorporating the $\pm 1\sigma$ variation of the sediment
756 composition from ODP1232 (Turner et al., 2017). Rear-arc samples from this study fit within the regional
757 trends. Arc-front centres lie close to the rear-arc best fit line, showing that these spatial trends in
758 enrichment are also present at the arc-front. The deviation in $^{143}\text{Nd}/^{144}\text{Nd}$ between the rear-arc and the arc-
759 front south of 39° S likely results from the sparse sampling of rear-arc centres, making it difficult to resolve
760 regional trends from local variations in heterogeneity. This difference may also reflect much smaller slab
761 input to the rear-arc than arc-front

762

763 Fig. 7 – Trace element and isotopic solution space produced by varying the composition of slab melts, the
764 extent of melting, and the composition of the ambient mantle. a) The pink field represents the maximum
765 isotopic solution space available by the addition of varying slab melt compositions (up to 2% sediment melt,
766 10% AOC melt) to a depleted mantle (“baseline ambient mantle”; BAM), with a trace element composition
767 of DMM, and isotopic composition of Sample 126171 (Søager et al., 2013). Variable slab temperatures have
768 very little effect on the size of this isotopic solution space, so are not shown. The grey field represents the
769 additional solution space made available by the addition of an EM1-type mantle domain to the composition
770 of the baseline Andean mantle (BAM). b) The pink field represents the maximum solution space available by
771 varying slab additions as in a), in addition to variations in F between 0.04 and 0.3. The teal field represents
772 the additional solution space made available by an increase in slab temperature (with a change in the
773 partition coefficients of La, Sm, and Nd; see SIA4.1). as in a), the grey field represents variable enrichment by
774 an EM1-type mantle source. In both figures, it is clear that only the addition of an EM1-type mantle source
775 to the sub-arc mantle can recreate the composition at Don Casimiro-Maipo. The diamonds represent the
776 best model fits for Villarrica (F=0.1, 7.5% slab addition consisting of 17% sediment, with no enrichment) and
777 Don Casimiro-Maipo (F=0.056, 7.5% slab addition consisting of 23% sediment, with 80% EM1 enrichment).
778 The full trace element signatures of these models are shown in Fig. 8.

779

780 Fig. 8 – Trace element modelling results. The trace element composition of Villarrica can be well matched at
781 F=0.1 with addition of 7.5% slab melt consisting of 17% sediment. The extreme trace element enrichment at
782 Don Casimiro-Maipo is best recreated by addition of 80% enriched source, and slight changes in the
783 composition of the slab addition (23% AOC vs. 17% at Villarrica). The mantle melting model includes a
784 parameterization for changing garnet proportions based on melt depths ($G_t=0.0067 \cdot \text{GPa}+0.0017$); we
785 assume that melting at Don Casimiro-Maipo and Villarrica occurs at 3.3 GPa and 2.2 GPa respectively (Turner
786 et al., 2016). The solid black lines in both figures represents the best model fit for each centre at the
787 enrichment and proportions of AOC:SED mentioned above using low temperature slab partition coefficients.
788 This provides the best match for Villarrica. The dashed grey lines represent the model fit for each centre at
789 increased slab temperatures, assuming 70% loss of Cs in the forearc. This provides the best model fit for Don
790 Casimiro-Maipo. Model envelopes are produced by adding and subtracting 1σ of 33 measurements of
791 ODP1232 (Turner et al., 2017). Melting conditions were generated directly from the physical modelling
792 results of Turner et al. (2016).

793

794 Fig. 9– Modelling contamination of the mantle source by subduction erosion. a) Isotope systematics of
795 contamination of the Villarrica mantle source by subduction erosion. The isotopic composition of the
796 Villarrica mantle source was calculated from the Villarrica primitive average, and the trace element
797 composition of the Villarrica mantle source was calculated from the best model fit in Fig. 8. Only a small
798 percentage of mixing lines pass near Don Casimiro-Maipo. b-c) Trace element and isotopic signatures of the

799 mixed compositions that produce isotopic ratios close to Don Casimiro-Maipo (red lines in a), and have
800 $\text{Eu}/\text{Eu}^* > 0.9$. None of these lithologies can recreate Sr, Ba and $^{87}\text{Sr}/^{86}\text{Sr}$ systematics at Don Casimiro-Maipo.
801 Error bars show 1σ of primitive samples at Don Casimiro-Maipo and Villarrica based on 5 and 3 analyses
802 respectively.

803

804 Fig. 10- Variation in rear-arc slab inputs. a) Mixing models of slab-sensitive ratios between a melt of a rear-
805 arc mantle source ($F=0.02$, melting at 3.3 Gpa) that has not received slab inputs, and a melt of a rear-arc
806 mantle source at the same conditions that has received slab inputs (20% sediment, 80% AOC). The orange
807 wedge reflects progressive mixing of 10% slab (with $\text{ODP1232} \pm 1\sigma$) into a mantle source with 0% EM1
808 enrichment, and the green wedge reflects mixing of 17% slab (with $\text{ODP1232} \pm 1\sigma$) into a mantle source with
809 80% EM1 enrichment. Rear-arc trace element ratios follow these mixing lines remarkably closely, with the
810 small amount of scatter likely explained by variable conditions of melting along the rear-arc. b) Trace
811 element ratios sensitive to slab additions versus trench distance. MORB-like ratios (blue histogram; Gale et
812 al., 2013) start to appear at distances greater than 425 km. Remarkably high, arc-like ratios (black histogram
813 shows the range of these ratios at the SVZ arc-front) are seen up to 600 km from the trench, with large
814 scatter in ratios at large trench distances, contrary to the narrow range of observed isotopes.

815

816 Fig. 11 –Isotopic compositions of inferred SCLM melts within South America (approximate locations on Fig.
817 2a). These clearly extend the array that passes from DMM/DAM through the arc-front and rear-arc, to Don
818 Casimiro-Maipo. The field produced by contaminating the Villarrica mantle source with crustal lithologies
819 (Fig. 9a) is also overlain (grey). This follows a much shallower trajectory than SCLM melts, and is significantly
820 broader than the SCLM field. Mobile belt M-SCLM melts represent alkaline volcanics erupted on the edges of
821 the Sao Francisco craton and the Rio Apa-Luis Alves craton on the east coast of South America (Gibson et al.,
822 1995; 2005 and Carlson et al., 1996). Southern Rift melts were erupted just north of the NSVZ (Lucassen et
823 al., 2002). Carboniferous granitoids are from the Santo Domingo Complex of the Coastal Batholith (Parada et
824 al., 1999). All are suggested to have sampled M-SCLM. The green and blue lines show a recycling model in
825 which low degree melts are extracted from the depleted mantle (Salters and Stracke, 2004) and then aged
826 for up to 2 Ga using the parameters from Stracke et al. (2003; see Turner et al., 2017). A model with $F=0.003$ -
827 0.008 encapsulates the range of isotopic compositions observed in SCLM melts.

828

829 Fig. 12- Schematic diagrams showing possible explanations for the isotopic enrichment minimum, and its
830 relationship to tectonic features. a) A Permian suture zone corresponds with an absence of M-SCLM. b) A
831 vertical and horizontal slab tear at $\sim 39^\circ\text{S}$ may allow depleted Pacific MORB mantle to upwell, and dilute the
832 EM1-type enrichment in the sub-arc mantle located above the tear. As the proportion of depleted Pacific
833 MORB mantle declines northwards and southwards, isotopic signatures become more enriched. Upper plate
834 geometry adapted from box model of S ager et al. (2015b), pp. 1514.

835

836 Fig. 13 –A unified model for the SVZ can account for the different trajectories of arc-front and rear-arc
837 samples. a) Large variations in Ba/Th within arc-front and rear-arc olivines can be accounted for by variations
838 in slab additions. For example, mafic enclaves from Longavi volcano show Ba/Th ratios up to ~ 500 , which
839 have been attributed to the subduction of the Mocho fracture zone beneath this centre (Rodriguez et al.,
840 2007). b-d) Trace element and isotopic plots showing rear-arc and filtered SVZ arc-front samples
841 ($\text{Eu}/\text{Eu}^* < 0.9$), along with the composition of the two mantle components used in this study (BAM, EM1), and
842 the composition field of MORB (Gale et al., 2013). The composition of mafic enclaves from Longavi are
843 shown to demonstrate the amount of variability that can be generated from the subducting slab. Melting

844 trajectories from the model described in section 4.2 are shown in c and d) to demonstrate the fractionation
845 of certain trace element ratios by the low melt extents that likely dominate the rear-arc. The unit vector
846 directions from BAM generated by varying different components are shown. Søger et al. (2013; 2015a)
847 suggest that the offset trajectory of the SVZ arc-front and rear-arc in the cross plots shown in b-d are
848 produced by LCC assimilation. However, the variability made available by varying degrees of EM1
849 enrichment, variable slab fluxes and variable melt extents can easily account for the offsets between these
850 trajectories.

851

852 Fig. 14 – Schematic diagrams of rear-arc slab supply. a) Traditional model of slab supply, with buoyant rise of
853 slab fluids to the rear-arc from a slab melting at higher pressure and temperature than at the arc-front. The
854 composition of these fluids should change with increasing trench distance, as the depth to the slab
855 increases. Additionally, at large trench distances, very little to no slab supply would be expected, due to the
856 exhaustion of hydrous phases. b) Proposed model of this study, inspired by Cagnioncle et al. (2007), Wilson et
857 al. (2014) and Cerpa et al. (2017). Down dragging of fluids released at arc-front depths explains the similar
858 composition of arc-front and rear-arc slab supply.

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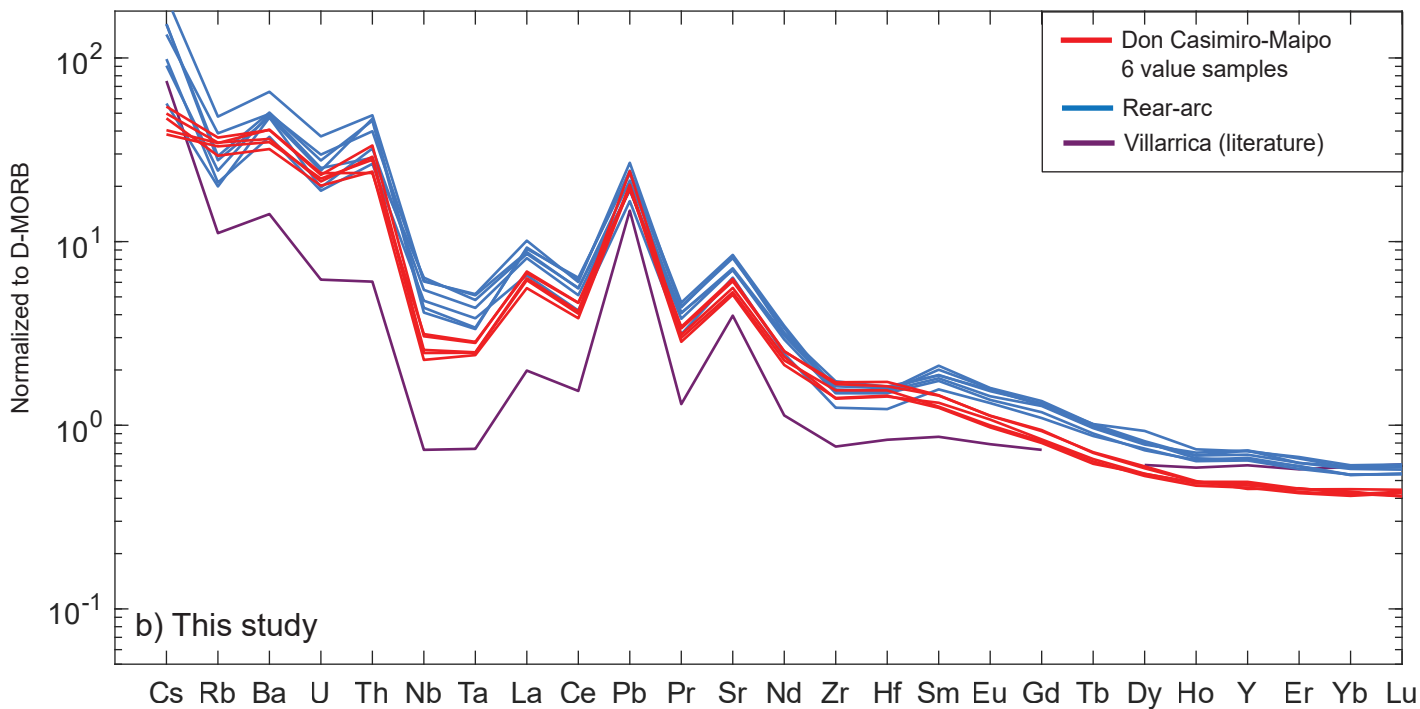
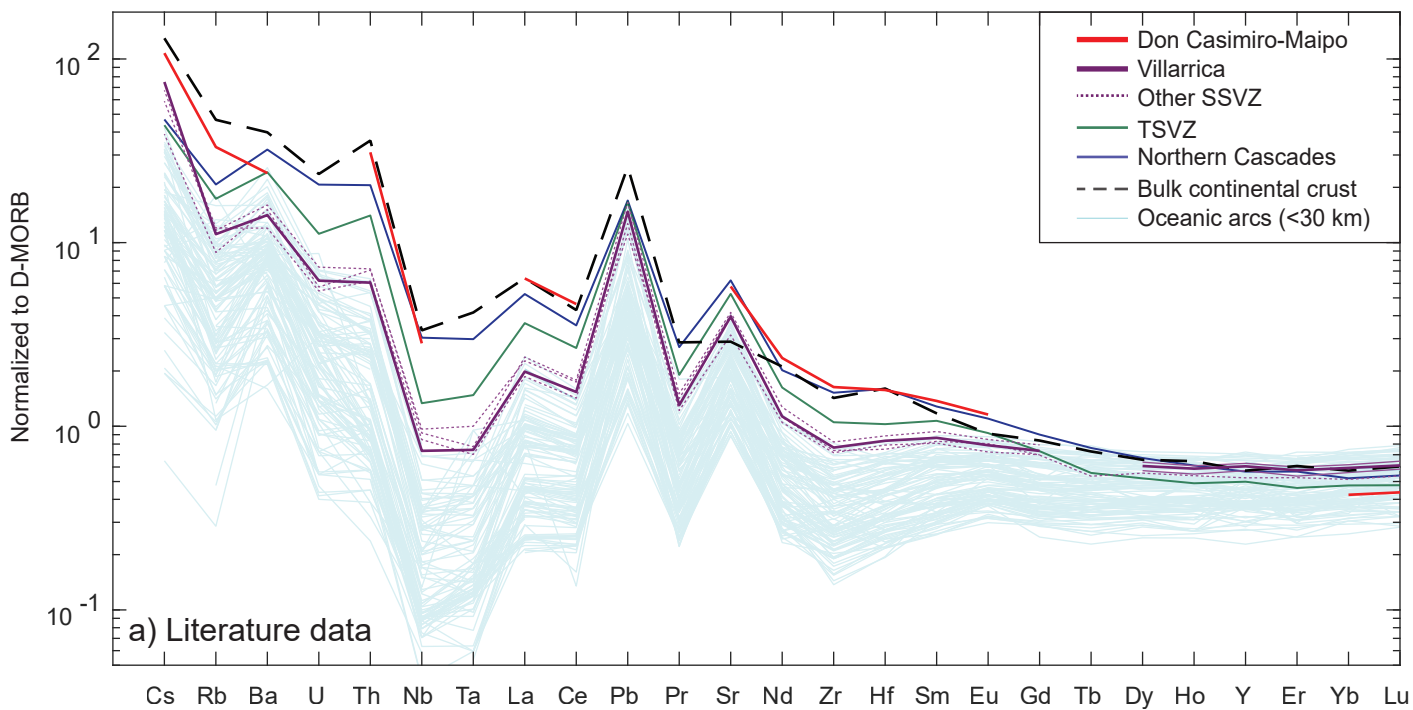
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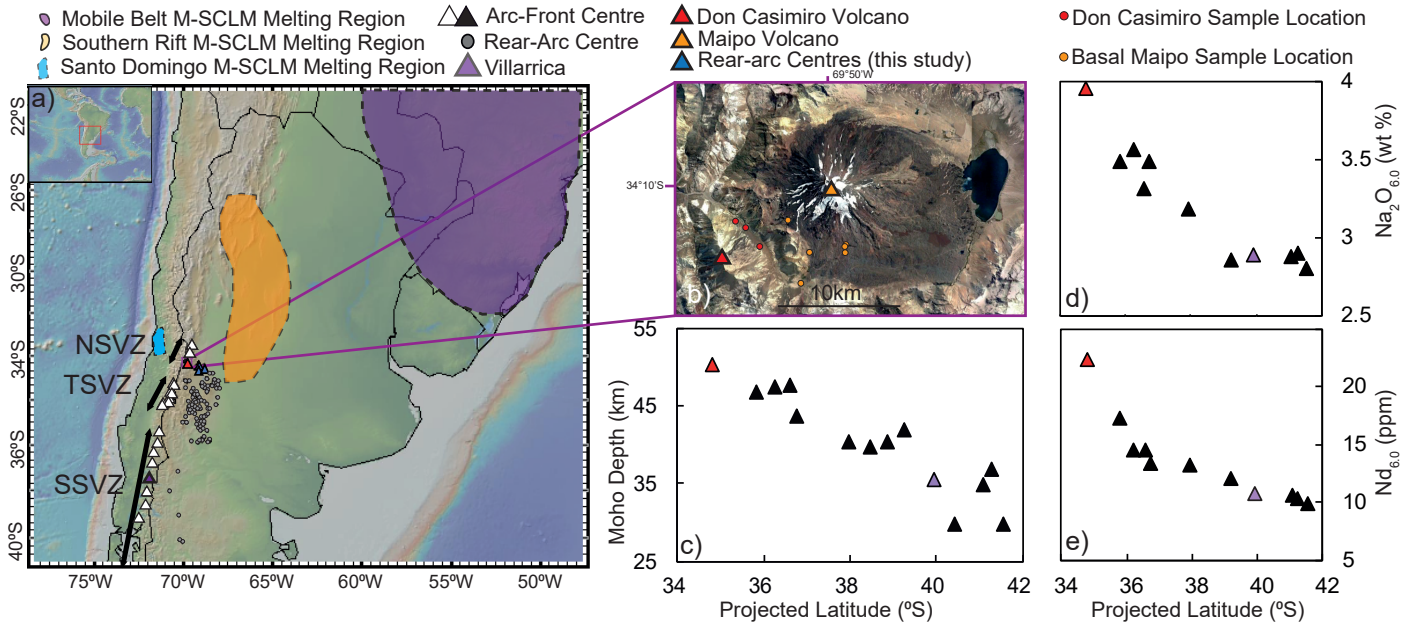
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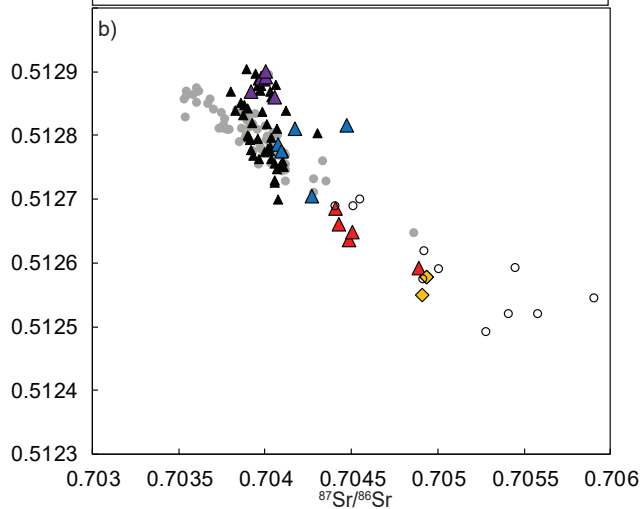
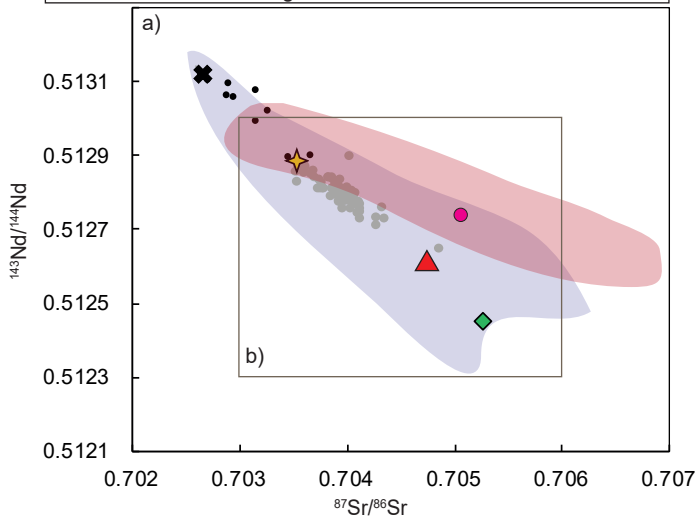
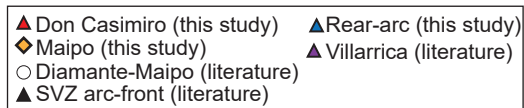
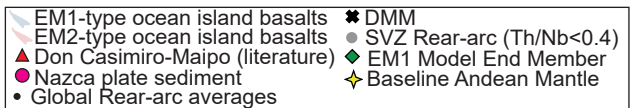
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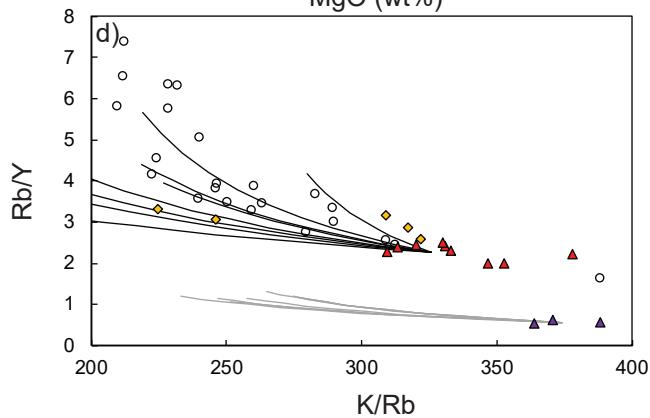
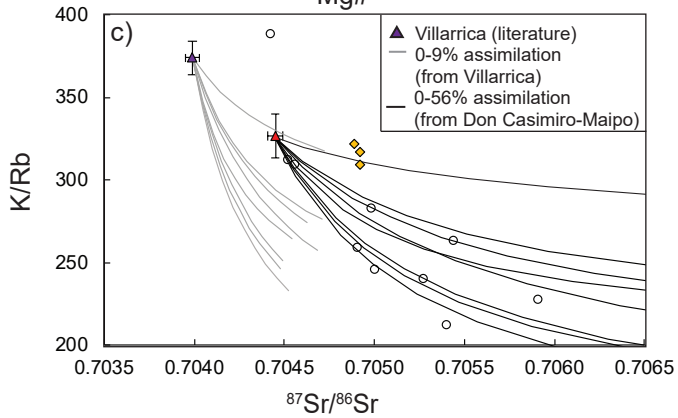
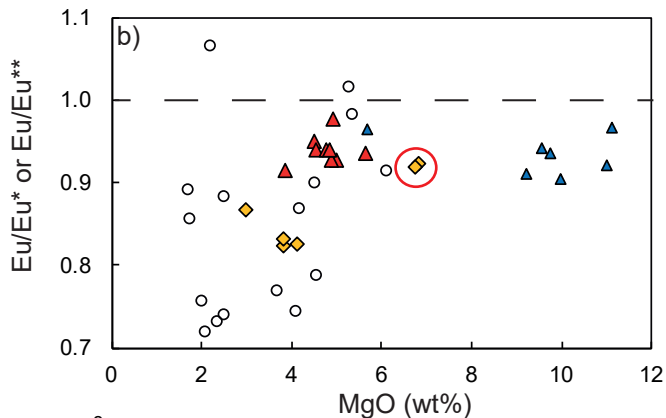
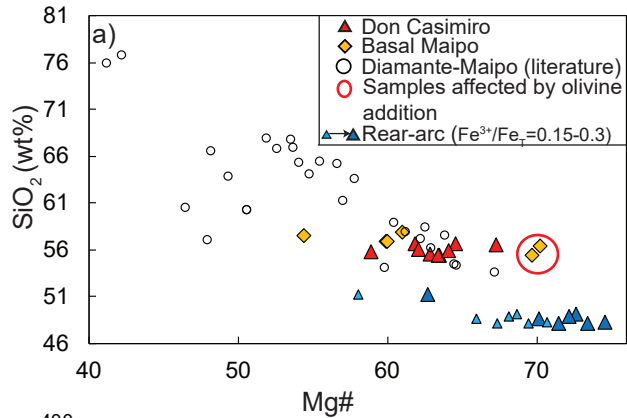
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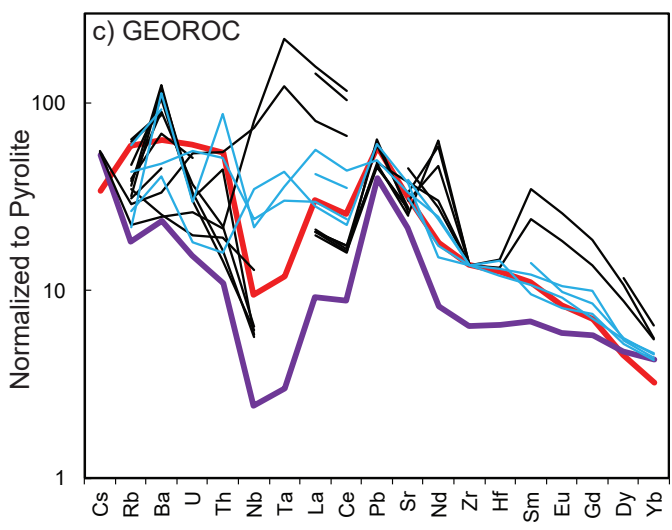
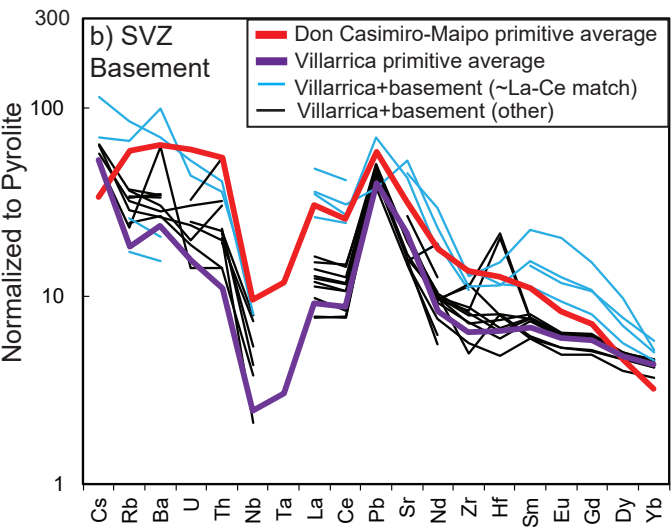
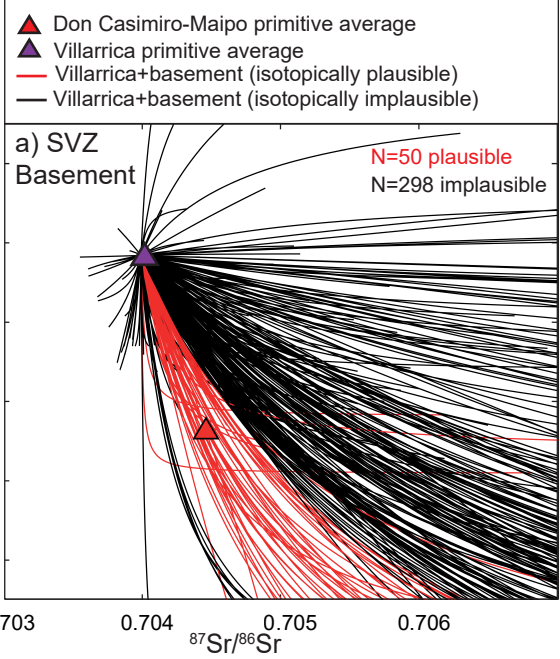
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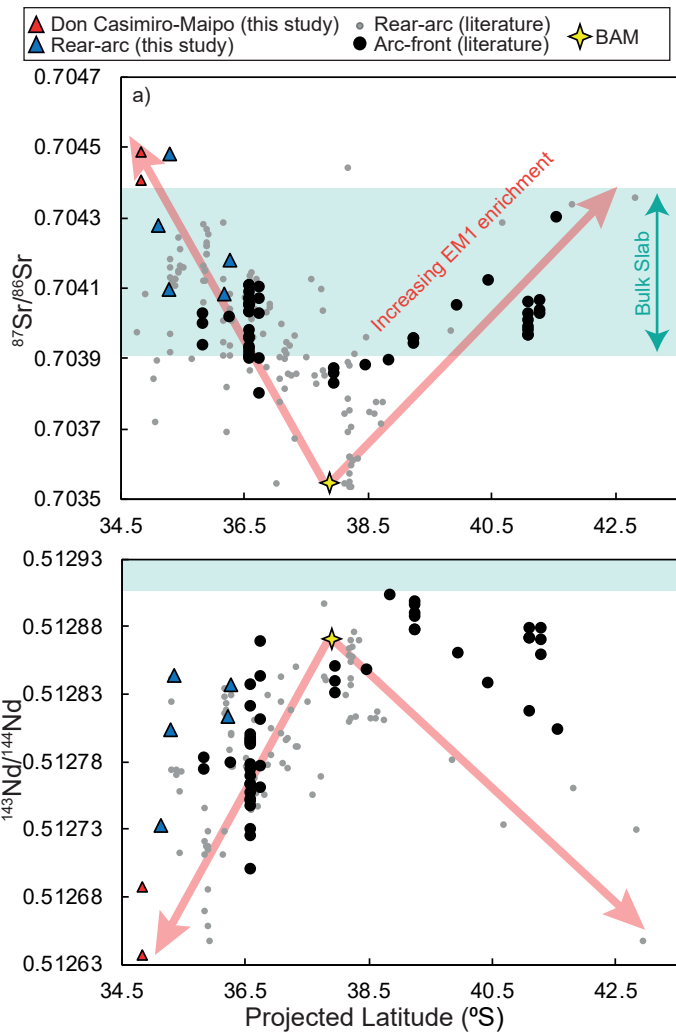


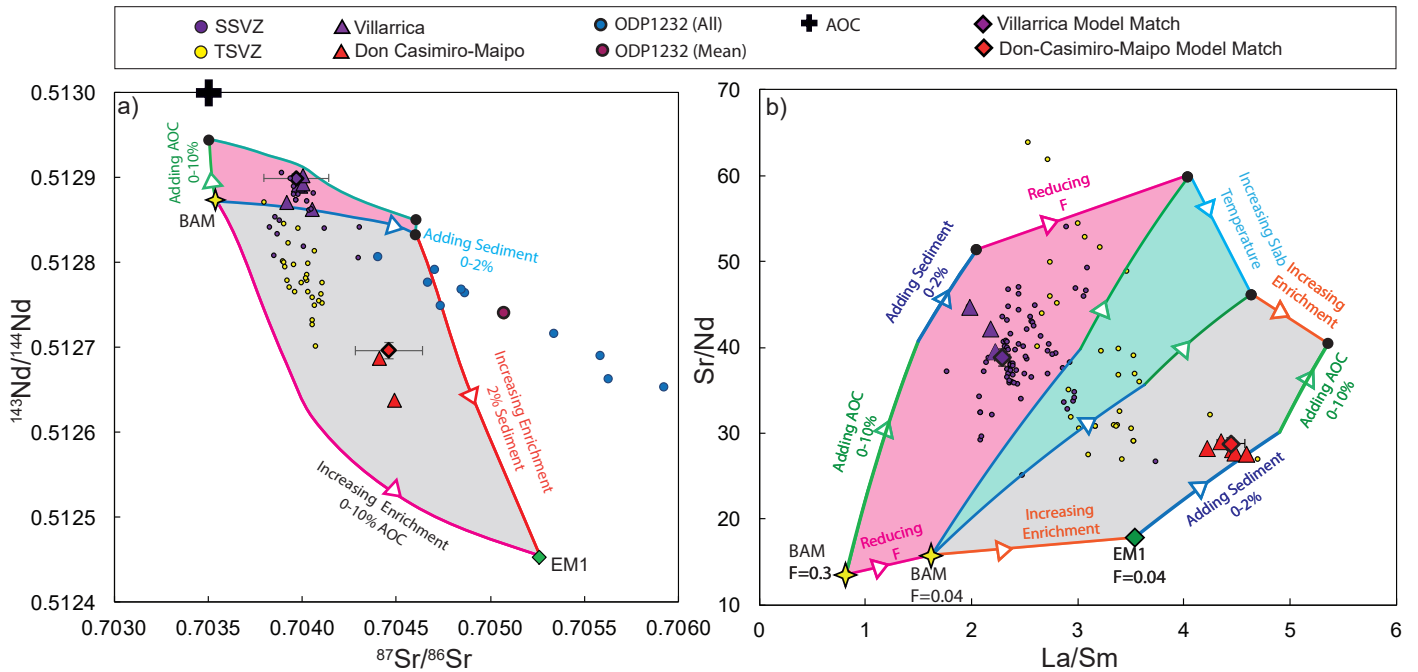




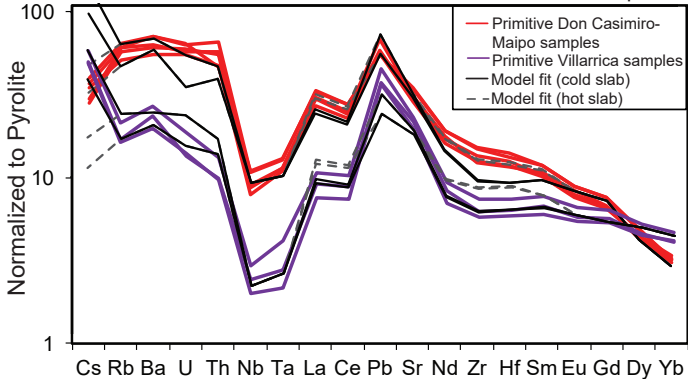




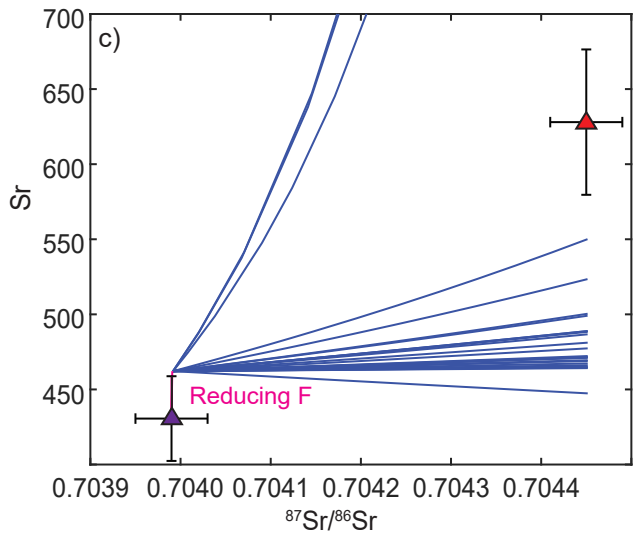
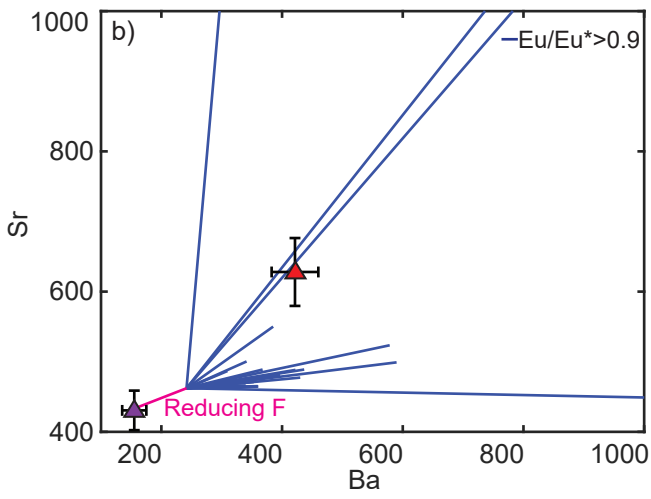
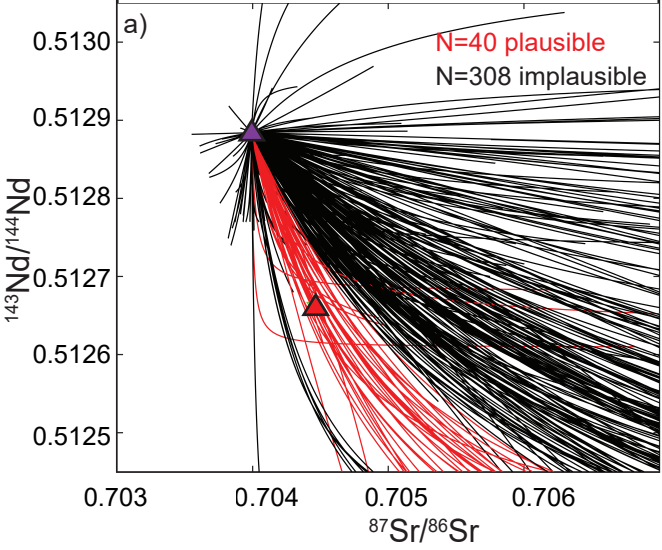


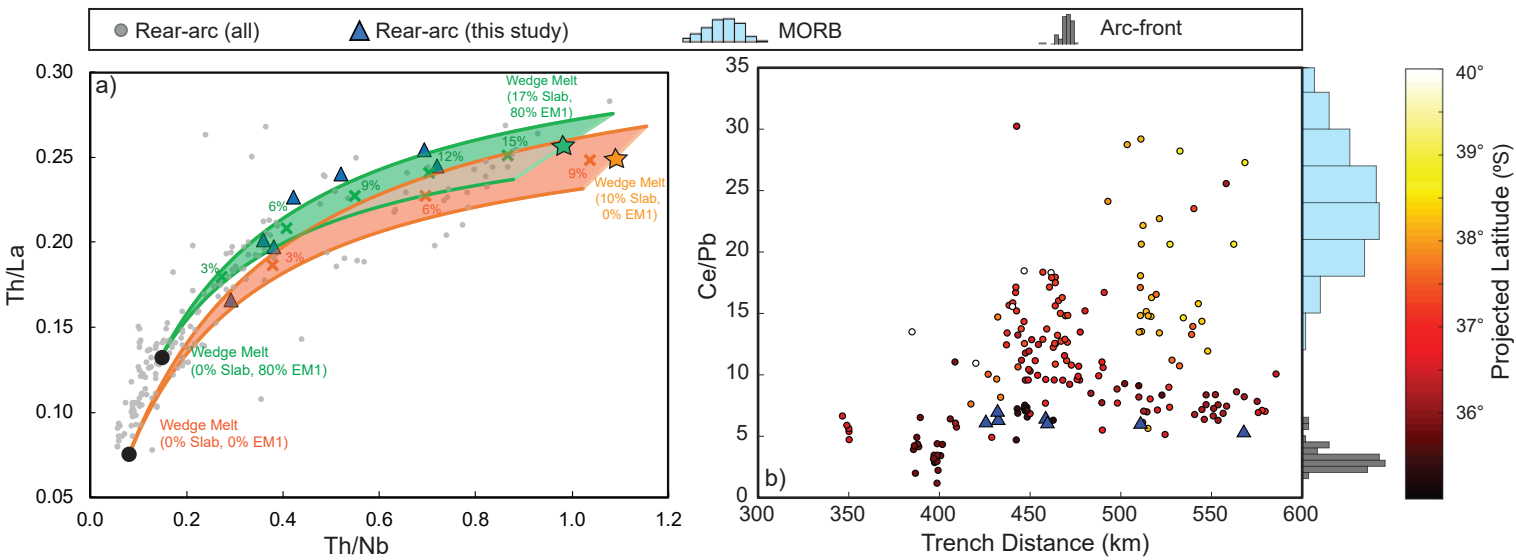


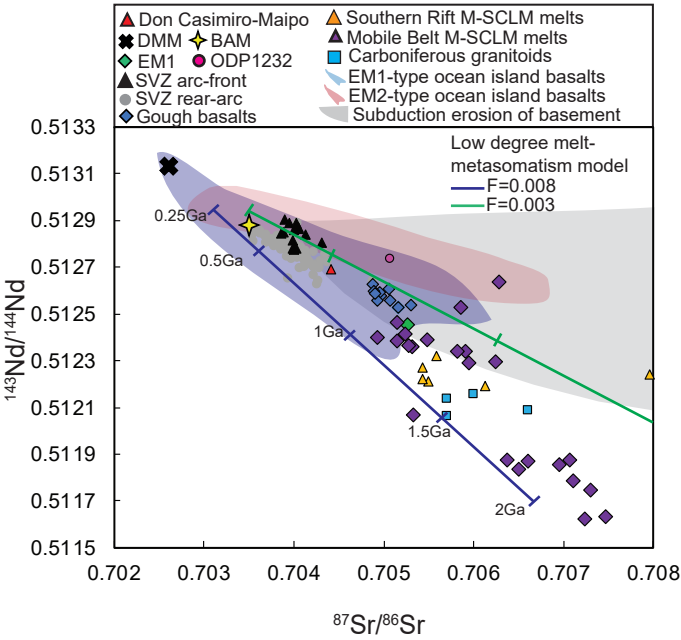
Data vs. model results for Villarrica and Don Casimiro-Maipo

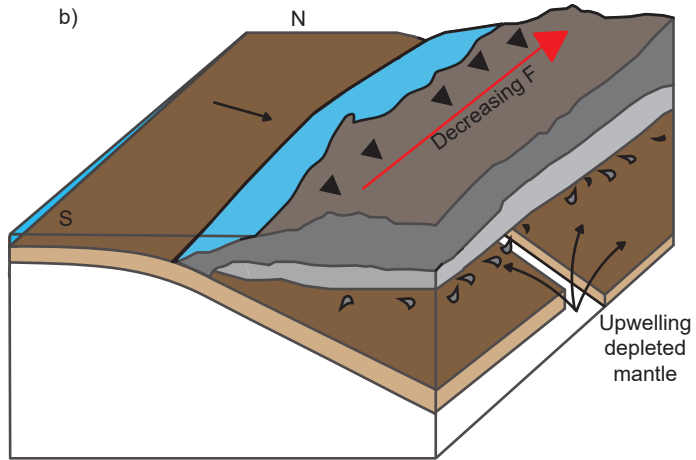
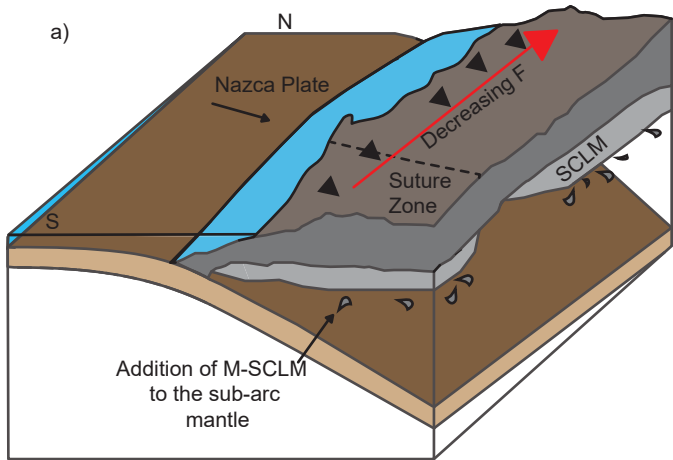


- ▲ Don Casimiro-Maipo primitive average
- ▲ Villarrica primitive average
- Isotopically plausible
- Isotopically implausible



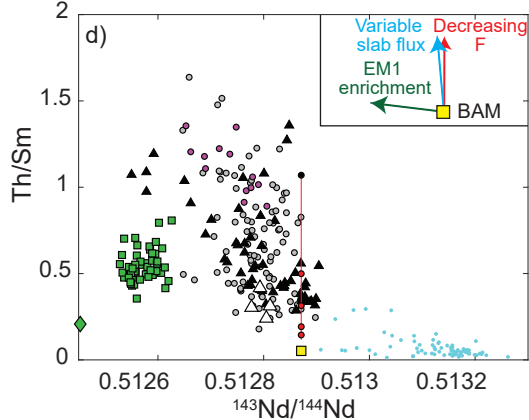
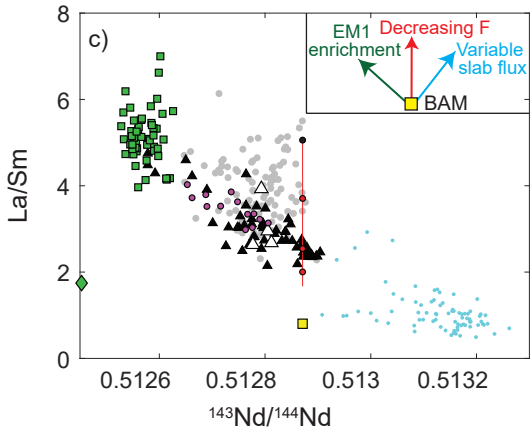
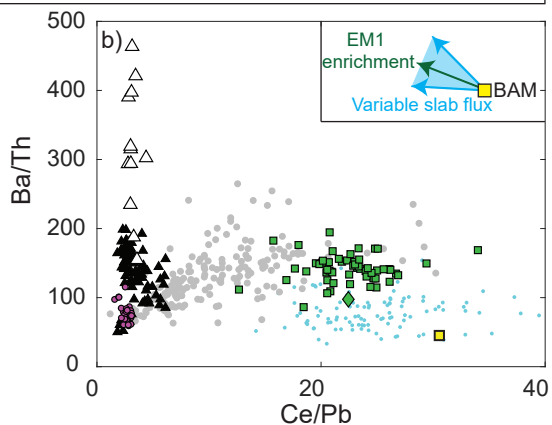
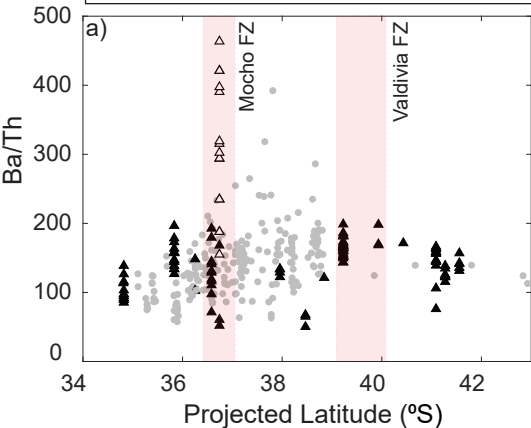
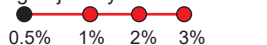


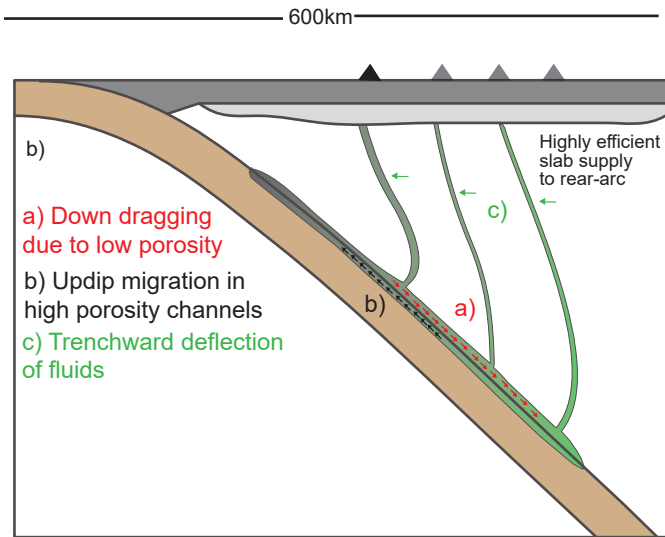
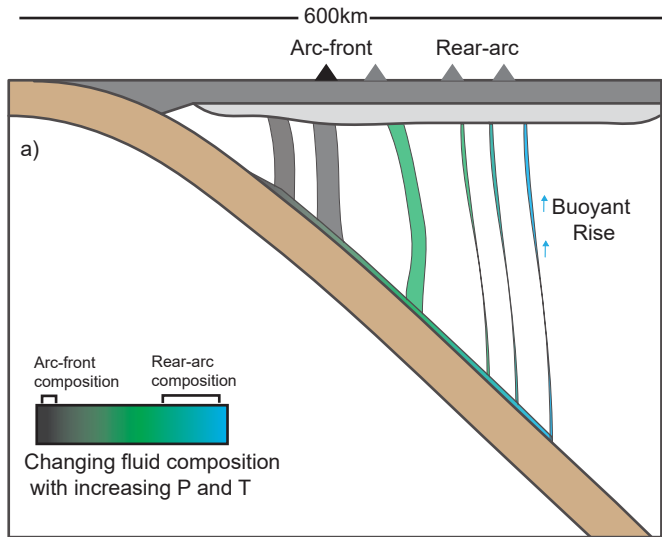






Melting trajectory from BAM





Appendix 1: Methods and Sample Details

SIA1 Data and Databases

1 Whole-rock and olivine analysis from this study is shown in Appendix 3. Compiled rear-arc
 2 and Diamante-Maipo literature data used in this study is in Appendix 4. The compilation
 3 of SVZ basement compositions is in Appendix 5.

4 SIA2 Sample Characteristics

5 Six 100 μ m sections were made for Don Casimiro-Maipo samples spanning a range of MgO
 6 contents. These were scanned using a 5MP slide scanner. Point counting was conducted in
 7 JMicroVision by overlaying a 1000 point grid on these scans. Microscope observations were
 8 made on any crystals that couldn't be identified with certainty in scans. Crystals smaller
 9 than ~ 0.05 mm were classified as matrix. Amorphous or crystalline material sufficiently
 10 distinct from the surrounding matrix that could not be identified as a mineral phase was
 classified as alteration.

Table SA1: Point counting results for Don Casimiro-Maipo samples (ordered by decreasing wt% MgO).

Sample	MgO (wt%)	Crystals (%)	Matrix (%)	Vesicles (%)	Alteration (%)
G0020315-1	6.82	11	89	0	0.1
G0020315-2	6.73	10	90	0	0.3
G0270215-1D	5.63	17	81	0	0.3
G0270215-1C	4.85	14	85	0	0.2
G0030315-4	3.87	11	72	16	0.0
G0030315-1	2.99	42	48	0	10

Table SA2: Percentage of each phenocryst phase in the crystal fraction. *The fine grained nature of these samples meant that phenocryst phases were harder to identify, so proportions are not as reliable as for other samples.

Sample	Olivine (%)	Clinopyroxene (%)	Plagioclase (%)	Oxide (%)	Orthopyroxene (%)
G0020315-1	66	34	0	0	0
G0020315-2	66	34	0	0	0
G0270215-1D*	6	15	46	33	0
G0270215-1C*	8	20	32	41	0
G0030315-4	7	5	89	0	0
G0030315-1	3	13	82	0	3

Table SA3: Description of Don Casimiro-Maipo samples in thin section.

Sample	Olivine	Plagioclase	Clinopyroxene	Matrix Textures
G0020315-1	Predominantly subhedral-anhedral 2-0.5mm grains, and smaller, <0.5mm anhedral grains. Both types form clots, and show association with clinopyroxene. Most olivines contain oxide inclusions.	No phenocrysts.	Subhedral-anhedral 2-0.5mm grains, often forming clots. Smaller, <0.5mm anhedral grains within the matrix. Some show breakdown in their cores.	Dark, fine grained matrix with white streaks of amorphous material. Needle-shaped matrix plagioclase aligns with these streaks. A few 1-2mm patches of alteration are associated with reacted olivine cores.
G0020315-2	Euhedral 1-3mm grains, subhedral 0.5-2mm clots associated with clinopyroxene, and smaller isolated <1mm anhedral grains.	No phenocrysts.	Subhedral 1-2mm grains, often associated with olivine clots. As above, some show broken down cores.	Dark, fine grained matrix, with white amorphous material in streaks. Small needle-shaped plagioclase shows alignment around large crystals. Abundant anhedral oxide grains.
G0270215-1D	Anhedral <0.2mm grains with abraded edges. These are associated with patches of small red amorphous material.	Blocky to needle-shaped plagioclase. No alignment.	Sub-anhedral, 0.3-1mm grains. These are more commonly associated with the red material than olivines.	Dark, fine grained matrix, with poorly aligned, needle-shaped plagioclase, and dispersed anhedral oxides. Patches of white, amorphous material.
G0270215-1C	Anhedral, <0.1mm grains.	Elongated needles, 20:1 aspect ratio. Aligned in matrix.	Sub-anhedral 0.3-1mm grains, associated with red amorphous material and oxide growth.	Dominated by small, aligned, needle-shaped plagioclase, with euhedral-anhedral opaque oxides.
G0030315-4	Anhedral, <1mm grains, with highly abraded edges, likely these were originally clots.	Large, 2mm+, blocky plagioclase, showing sieve textured cores (indicating rapid growth/cooling/heating).	Euhedral, 2-4mm grains, and smaller, <1mm highly abraded anhedral grains.	Highly vesicular texture. Matrix dominated by glass.
G0030315-1	A few subhedral, 1-2mm olivines. Some associated with large, 1-2mm cubic oxide phases.	Several populations of plagioclase. 1) blocky 2-4mm rectangular macrocrysts, with large fluid inclusions. 2) lath shaped 1-3mm grains, with inclusion-free cores. 3) needle shaped (<1mm) grains, with sieve textured cores.	Euhedral-subhedral, 1-2mm grains with oxide inclusions. Some are also associated with very large, 1-3mm oxide grains. Additionally, one 4mm clot of several grains and oxides is observed. In some places, clinopyroxene forms anhedral intergrowths with plagioclase, and orthopyroxene.	Extremely crystal-rich. Large areas of amorphous material within the slide are associated with large blocky plagioclase macrocrysts. Between phenocrysts, the matrix is dark and amorphous.

12 SIA3 Methodologies

13 SIA3.1 Whole Rock Chemistry

14 Sixteen samples from Don Casimiro-Maipo and seven rear-arc samples were prepared for whole-rock analysis
 15 at the University of Oxford. Arc-front lava samples were cleaned in an ultrasonic bath until the water ran
 16 clear. These were then split into 2-3cm chunks using a pneumatic rock splitter. Rear-arc samples were wet
 17 sieved, and unaltered clasts from the coarsest fraction (>2mm) were hand picked for whole rock analysis.
 18 Both sample types were crushed in a steel jaw crusher, and then powdered in an agate ball mill.

19
 20 **Major Element Analysis:** Major element analyses was conducted by using a PANalytical Axios Advanced
 21 X-ray fluorescence spectrometer at the University of Leicester, UK, following the methods of Knott et al. (2016).

22
 23 **Trace Element Analysis:** ~ 100 mg of each powder as well as reference materials BCR-2a, BHVO-2,
 24 W-2a and AGV-2, were digested in ~4ml of 7.5M HNO₃ and ~1ml of 24M HF. The samples were then
 25 evaporated and redissolved twice in ~4ml 7.5M HNO₃ to eradicate fluoride precipitates. At this stage, a
 26 sample split was made for isotopic analysis. Trace element analyses were conducted using a Perkin Elmer
 27 NexION 350D quadrupole ICP-MS for the majority of elements. An additional 1200x dilution and addition
 28 of an internal standard containing Rh, Re, and In was achieved using the ESI prepFAST system producing
 29 a 6000x solution in which to measure lower concentration elements. All prior dilutions were performed
 30 gravimetrically. Sample G0030315-4 was run every four samples, and the entire run was duplicated in reverse
 31 order to monitor drift. Drift correction and data reduction were carried out following Turner et al. (2013).
 32 Due to malfunctioning of the Prepfast system, a planned run at 60,000x dilution for higher concentration
 33 elements using the same method failed. Thus, Sr and Ba concentrations were measured using a Thermo
 34 Element 2 ICP-MS (also at the University of Oxford) at a 1,200,000x dilution (performed gravimetrically).

35 The digestion procedure, and sample homogeneity can be assessed by calculating the % difference between
 36 the measured element concentrations of two duplicates produced from the powders of this study, and two
 37 from an additional set of samples digested alongside these. Most measurements lie within 5%, suggesting
 38 that the digestion process, and samples, were reasonably homogeneous (table SA4).

39 Isotopic Analysis

40 Strontium (Sr) and neodymium (Nd) isotope analyses were performed on a Thermo Scientific Triton-series
 41 multicollector mass spectrometer at the School of Earth and Environment, University of Leeds. Eleven
 42 whole-rock powders were digested alongside those for trace element analysis, followed by subsequent drying
 43 and pick up in concentrated ultrapure HNO₃ and HCl acids. The final dried samples were diluted and
 44 centrifuged for 15 minutes at 2000 rpm. Sr and Nd were extracted by using Sr-Spec and TRU-spec resins and
 45 conventional ion-exchange chromatographic techniques. ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd ratios were normalized for
 46 mass fractionation to ⁸⁶Sr/⁸⁸Sr=0.1194 and ¹⁴⁶Nd/¹⁴⁴Nd=0.7219. Samples for Sr were loaded onto previously
 47 outgassed W filaments and samples for Nd were loaded onto previously outgassed Re filaments. Sr and Nd
 48 were analysed while the current was maintained between 3-5 V (for ⁸⁸Sr) and 0.4-0.8 V (for ¹⁴⁴Nd).

49 The average ⁸⁷Sr/⁸⁶Sr obtained from repeated measurements of NIST SRM-987 during the course of this
 50 study was 0.710271. Similarly, we measured ¹⁴³Nd/¹⁴⁴Nd=0.511837 for the LaJolla standard. Two samples
 51 (G0020315-2 and G0270215-1C) were analyzed for ¹⁴³Nd/¹⁴⁴Nd in a separate run, in which we measured
 52 ¹⁴³Nd/¹⁴⁴Nd=0.511850 for the LaJolla standard. We also successfully reproduced the published USGS
 53 standard BHVO-2 values for ⁸⁷Sr/⁸⁶Sr=0.703468 and ¹⁴³Nd/¹⁴⁴Nd=0.512995 (Weis et al., 2005). Details
 54 on the sample preparation and analytical protocols are described in Crummy et al. (2014).

55 SIA3.2 Correction of Isotopic data of this study and Literature 56 Data

57 Sr and Nd isotopes from the literature and this study were corrected to be in line with the analytical procedure
 58 used for SVZ data presented in Turner et al. (2016). Various standards were corrected to be equivalent
 59 to La Jolla=0.511858, NBS 987=0.710235 based on session averages.

Table SA4: % deviation of four duplicates ran during analysis on the "quad" (except starred elements, where % deviation is reported from the "element"). (4) represents a duplicate of G0030315-4, (1) represents a duplicate of MDZ1, 079a and B1 are samples measured in the same run, but for use in a study on the Main Ethiopian Rift. % deviation is defined as (Measurement 1-Measurement 2)/(Average of measurement 1 and 2) x 100.

	% difference (4)	% difference (1)	% difference (079)	% difference (B1)
Li	4.13	1.74	-0.97	0.35
Be	-4.52	60.2	-12.4	1.29
Sc	5.89	-27.0	-20.5	-0.72
V	2.02	-1.95	-0.65	0.15
Cr	3.02	0.41	1.14	-0.99
Ni	3.53	0.74	0.86	1.47
Cu	4.71	-3.94	-0.26	1.76
Rb	3.64	0.28	-9.66	0.30
Sr*	5.54	1.74	0.35	-5.67
Y	2.99	-3.08	-6.18	-0.94
Zr	2.12	-0.29	-0.42	0.32
Nb	3.09	0.21	-0.30	1.16
Cs	8.90	-5.49	-4.43	9.11
Ba*	6.08	3.17	3.65	-2.89
La	3.52	0.58	1.10	0.82
Ce	3.31	-1.09	0.94	0.70
Pr	4.57	0.71	2.05	0.80
Nd	3.29	3.63	-0.88	-0.53
Sm	0.00	2.17	1.35	2.23
Eu	2.21	2.20	1.55	-1.12
Gd	3.09	-2.92	-2.87	1.04
Tb	3.26	-2.17	2.29	0.91
Dy	3.83	-0.60	2.67	0.78
Ho	4.64	-2.73	-0.32	3.88
Er	2.99	4.01	-2.53	3.70
Tm	2.81	-5.67	-3.75	0.51
Yb	2.67	1.49	-0.23	1.69
Lu	1.28	1.41	-3.58	1.15
Hf	3.34	0.93	-0.87	1.40
Ta	-2.98	-0.57	1.60	-0.94
Pb	3.55	-2.14	-9.60	1.72
Th	3.48	-2.05	-17.6	-1.10
U	3.33	-0.09	0.56	-0.11

60 SIA3.3 Olivine Chemistry

61 Olivines were picked from the 1000-250 μ m grain fractions, and mounted in epoxy rings. These
62 mounts were polished and carbon coated for EMPA analysis. Analysis was conducted with
63 an accelerating voltage of 15KeV, a current of 100nA and a 10 μ m spot size. St John's Island
64 and Fayalite were run as secondary standards, with % recoveries reported in table SA5.

65 LA-ICP-MS analysis was conducted at the University of Cambridge. Three instrument
66 runs were performed over two days. The run conditions were 20Hz, 80 μ m, 9.5J/cm², 0.8L/min
67 of gas flow for He and Ar. Where possible, three spots were placed in the centre of the olivine,
68 with no overlap. For very small grains, only 2 spots were placed, as it is preferable to have
69 fewer spots nearer the centre than more spots towards the grain edges, due to the reasonably
70 high diffusivity of elements of interest to this study. To allow drift correction, 3 spots on each of

71 NISTSRM612, NISTSRM610, 06SGOL03 (inhouse olivine standard) and BD4074OL-7 (a San
 72 Carlos olivine) were run every 30 analyses. ^{29}Si was used as an internal standard, with values
 73 obtained from EMPA grain averages. Signals were selected using Glitter Processing software,
 74 with normalization to NISTSRM612 via a quadratic drift correction. This was effective at
 75 removing drift, as determined by observing the stability of NISTSRM610, 06SGOL03 and
 76 BD4074OL-7 following this correction. Multiple secondary standards were run to assess the
 77 accuracy of the measurement for elements of interest (NISTSRM610, NISTSRM614, BCR-2G).
 78 Measurement of inhouse olivines O6SGO103 and BD407OL-7 provided an excellent assessment
 79 of the success of the drift correction performed on glass standards. Elements of interest in this
 80 study drifted by less than 7%. Although Mn and Ni recovery on glass standards were good,
 81 these elements were recovered more poorly on BCR-2G. This, and the fact that previous
 82 studies in this area have used EMPA Ni and Mn data, resulted in us choosing to use EMPA
 83 data for consistency, as most of the peridotite-pyroxenite ratios involve normalization to Fe
 84 (which is reported from EMPA). Detection limits, and 1 sigma uncertainties calculated from
 85 Glitter for a representative sample (MDZ2) are shown in table SA6.

Table SA5: Peak count times, and calibration materials used. % recovery calculated below is defined as Measured Average/Published Average x 100. Where no published value exists for a standard, - is shown. Detection limits are reported from 9 spots on MDZ8. These are assumed to be representative of the analysis as a whole.

	Mn	Fe	Mg	Si	Ca	Ni
Peak count times (s)	20	30	20	20	70	70
Calibration material	Mn metal	Andradite	MgO	Albite	Wollastonite	Ni metal
% Rec Fayalite	97.7	94.2	-	99.6	97.0	-
% Rec St Johns Island Olivine	106.1	96.1	99.6	99.2	-	90.1
Detection limit (ppm)	173	197	173	98	74	137

Table SA6: Minimum (99% Confidence) detection limits (ppm), and 1 sigma uncertainties (ppm) for all spots on MDZ2, taken as representative for the analysis as a whole. The mass in brackets represents the isotope measured on the ICP-MS.

Element	MDL	1 σ uncertainty
Li(7)	0.115	0.151
Na(23)	1.121	6.92
Al(27)	0.318	39.3
P(31)	13.76	47.3
Ca(43)	133	192
Sc(45)	0.062	0.412
Ti(47)	0.551	18.4
V(51)	0.043	0.504
Cr(53)	0.578	56.0
Mn(55)	0.136	88.4
Co(59)	0.049	7.91
Ni(60)	0.171	191
Cu(65)	0.351	0.416
Zn(66)	0.373	10.1
Ga(69)	0.042	0.049
Ge(72)	0.370	0.208
Y(89)	0.037	0.025
Zr(90)	0.033	0.052
Nb(93)	0.035	0.021

Table SA7: % recovery of NISTSRM610 (average of 3 spots). (x,y) refers to when this secondary standard was run, x referring to the instrument run (1-3), and y to the order of measurement within a given run. These values were used to check the success of the quadratic drift correction. Across the three instrument runs, most elements drift by less than 10%.

Element	(1,1)	(1,2)	(1,3)	(1,4)	(1,5)	(2,1)	(2,2)	(2,3)	(2,4)	(3,1)	(3,2)	(3,3)	(3,4)	AVERAGE
Li(7)	98.0	105.5	103.7	106.7	105.9	107.3	104.3	101.1	103.6	108.0	99.6	96.8	98.2	103.0
Na(23)	98.2	105.7	107.0	106.1	105.9	105.9	104.9	101.4	103.9	105.0	103.0	101.1	102.1	103.9
Al(27)	102.3	105.5	105.9	106.2	107.8	108.3	104.2	104.9	107.3	105.6	107.2	103.5	104.9	105.7
P(31)	107.8	105.2	99.7	130.9	194.5	100.5	103.9	101.0	100.5	125.0	113.1	117.6	133.9	118.0
Ca(43)	95.9	103.7	100.6	106.7	107.3	105.2	102.5	99.2	102.2	105.2	100.1	99.5	104.5	102.5
Sc(45)	98.7	109.1	108.0	109.8	108.8	110.2	110.9	105.6	108.9	108.5	106.7	108.0	109.6	107.9
Ti(47)	115.2	120.8	116.9	120.0	119.0	125.1	128.7	124.5	125.5	125.3	125.9	122.8	125.4	122.7
V(51)	95.7	101.9	100.1	105.2	105.8	104.8	102.6	100.1	105.1	105.9	101.7	99.9	103.7	102.5
Cr(53)	103.4	107.7	109.4	113.6	119.5	120.7	115.9	112.1	115.1	114.1	111.4	112.7	116.7	113.2
Mn(55)	94.6	100.0	100.0	103.7	102.4	105.1	100.2	99.0	103.5	104.9	97.1	96.1	97.7	100.3
Co(59)	95.5	103.0	101.7	103.7	100.8	104.2	102.0	98.9	104.3	103.6	101.0	100.3	103.4	101.7
Ni(60)	92.1	100.9	99.8	103.7	99.9	105.6	100.1	98.2	100.0	104.3	101.8	100.5	102.2	100.7
Cu(65)	89.9	98.8	97.3	100.1	97.5	100.5	98.4	94.9	99.0	100.1	96.0	95.9	97.6	97.4
Zn(66)	98.5	104.1	102.3	108.8	110.2	111.1	104.4	100.1	106.6	109.9	106.4	104.1	102.9	105.3
Ga(69)	92.2	97.7	96.5	101.1	101.5	101.7	99.0	97.2	99.4	101.4	98.9	98.3	100.8	98.9
Ge(72)	81.5	86.7	85.7	89.1	88.8	92.4	89.0	85.6	86.6	92.0	87.5	87.8	89.8	87.9
Y(89)	93.1	102.1	99.8	102.8	102.1	104.0	102.9	100.5	100.3	102.5	101.4	101.8	103.8	101.3
Zr(90)	89.8	95.9	94.6	97.9	98.1	98.6	98.5	94.6	95.4	97.5	96.8	96.0	98.7	96.3
Nb(93)	92.8	99.9	98.9	102.4	101.7	102.2	102.2	98.1	100.2	102.9	99.9	99.2	101.1	100.1

Table SA8: % recovery of BCR-2G, using averages from 3 spots, and % recovery of NISTSRM614. The % recovery for NISTSRM614 is not shown for the first run. These were the first spots ran by the laser, and it has been shown that the system takes a few spots to warm up and settle after switching from solution mode to laser mode.

Element	(1,1)	(2,1)	(2,2)	(3,1)	(3,2)	BCR-2G Average	(2,1)	(3,1)	NISTSRM614 Average
Li(7)	100.2	104.5	101.7	105.8	94.6	101.4	84.4	103.7	94.1
Na(23)	96.2	99.1	97.0	101.4	96.9	98.1	105.5	103.1	104.3
Al(27)	103.8	103.4	99.9	104.6	103.3	103.0	106.6	106.3	106.4
P(31)	101.3	84.9	86.8	108.4	120.1	100.3	268.2	260.0	264.1
Ca(43)	99.0	97.9	98.9	101.1	99.1	99.2	106.2	103.7	105.0
Sc(45)	97.5	102.1	96.4	100.8	99.9	99.3	217.1	185.9	201.5
Ti(47)	107.3	108.4	105.2	110.9	106.3	107.6	122.1	101.3	111.7
V(51)	97.6	100.1	101.1	102.8	99.6	100.2	109.7	99.9	104.8
Cr(53)	93.2	104.4	92.9	97.1	101.1	97.7	94.8	113.0	103.9
Mn(55)	92.4	96.6	94.9	98.5	90.0	94.5	105.6	89.7	97.7
Co(59)	99.0	97.8	97.0	97.9	96.2	97.6	92.6	95.0	93.8
Ni(60)	89.4	86.9	83.8	90.2	86.5	87.4	107.9	71.8	89.8
Cu(65)	71.0	72.5	67.5	71.1	69.6	70.3	88.8	109.0	98.9
Zn(66)	121.5	134.0	127.4	134.9	120.7	127.7	91.4	89.2	90.3
Ga(69)	111.6	117.1	109.2	118.4	114.5	114.2	96.9	93.9	95.4
Ge(72)	99.6	112.9	95.6	101.1	102.4	102.3	86.3	110.8	98.5
Y(89)	87.3	90.0	81.9	90.6	89.1	87.8	102.0	99.4	100.7
Zr(90)	84.5	87.0	78.3	87.7	86.3	84.7	90.3	91.6	90.9
Nb(93)	88.6	91.6	86.6	91.0	90.2	89.6	93.7	99.3	96.5

86 SIA4 Mantle Melting Model Details

87 As discussed in the text, this study uses an updated version of the model of Turner and
88 Langmuir (2015b). Below, the justification of various inputs to the model are discussed. The
89 updated Matlab scripts ADD_SLAB.m, ADD_SLAB_HOTP.m and ModeMelt_Hydrous.m
90 have been uploaded into the supplementary information.

91 SIA4.1 Model Partition Coefficients

92 **Hydrous Partition Coefficients for Mantle Melting:** This study utilizes hydrous
93 partition coefficients and mantle modes from Turner and Langmuir (2015b; detail in their
94 supplement). However, we slightly updated the partitioning behaviour of several elements:
95

96 1) The bulk partition coefficient of Eu is obtained by a log extrapolation from those of Sm and
97 Gd:

$$98 \quad Kd_{Eu} = (Kd_{Sm} * Kd_{Gd})^{0.5}$$

99 correcting an error in Turner and Langmuir (2015b).

100 2) The partition coefficient of Hf is assumed to be equal to that of Sm.

101 3) The partition coefficient of Zr is approximated as:

$$102 \quad Kd_{Zr} = (Kd_{Nd} * Kd_{Sm})^{0.5}$$

103 **Hydrous Partition Coefficients for slab melting:** As above, this study uses slightly
 104 adapted partition coefficients for slab melting to those in Turner and Langmuir (2015b).
 105 These lie within the limits of experimental values (figure SA1). The partition coefficients for
 106 the LREE (La, Ce, Nd, and Sm) were reduced by 40% to reflect the hotter slab underneath
 107 Don Casimiro-Maipo (inline with experimental high temperature partitioning results; figure
 108 SA1). Additionally, the partition coefficients of Zr and Hf were reduced, and 70% of Cs
 109 was assumed to be lost to the forearc.

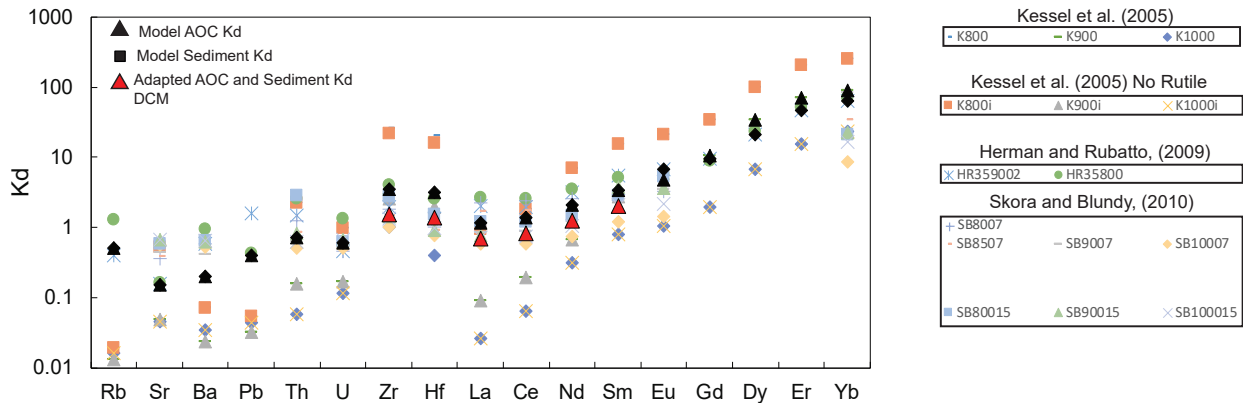


Figure SA1: Slab melting partition coefficients lie within experimental values for 800-1000°C.

110 SIA4.2 Model End-Members

111 **"Ambient Andean Mantle" (Depleted Mantle End-Member):** The isotopic composition
 112 of the depleted mantle end member was set to $^{87}\text{Sr}/^{86}\text{Sr}=0.70355$ and $^{143}\text{Nd}/^{144}\text{Nd}=0.512871$,
 113 the isotopic composition of sample 126171 of Soager et al. (2013) following the isotopic
 114 correction discussed above. This sample lies at the depleted end of the rear-arc array,
 115 and thus reflects the composition of the Andean mantle prior to the addition of EM1
 116 enrichment. For trace element modelling, the depleted mantle end-member is DMM of
 117 Workman and Hart (2005).

118 **AOC:** The subducting Nazca plate has Pacific affinities. Trace element concentrations
 119 were obtained from melting of N-MORB (Gale et al., 2013) at 876°C, $F=0.3$. The isotopic
 120 composition of AOC was set to $^{87}\text{Sr}/^{86}\text{Sr} = 0.7035$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.5130$.

121 **Sediment:** The trace element composition of the sediment end-member was obtained from 33
 122 samples of ODP1232 (c.f. Turner et al., 2017). To produce an envelope of model results for trace
 123 elements and isotopes, one standard deviation of measured samples were added and subtracted
 124 to the mean. Three outliers were discarded based on their anomalously high Sr isotope ratios.

125 **Enriched EM1 end-member:** Enrichment at Gough represents a mix of M-SCLM, and
 126 ambient mantle. The isotopic signature of recycled M-SCLM depends on the storage time
 127 (Turner et al., 2017). As several samples within this study have signatures more enriched
 128 than Gough, we use the isotopic composition of a sample from Pitcairn (57DS9; Stracke et
 129 al., 2003) as the EM1-like isotopic end member. This lies at the enriched end of the field

130 enclosing EM1-type ocean island basalts. The trace element concentration was obtained
 131 from the inversion of Turner et al. (2017), with a few changes:

132 1) the bulk Eu partition coefficient was set at:

$$133 \quad Kd_{Eu} = (Kd_{Sm} * Kd_{Gd})^{0.5}$$

134 correcting an error in Turner et al. (2017) where it was set at 0.

135 2) The Zr concentration in Gough primary melts was adjusted to the standard EM1 value
 136 $Zr/Hf = 45$.

137 3) Zr partition coefficients were extrapolated from Salters and Longhi (1999).

138 4) Nb and Ta partition coefficients were taken as $Kd_{La}/3$, based on ratios from Workman and
 139 Hart (2005).

Table SA9: Gough mantle source composition used in this study compared to that of Turner et al. (2017; T2017).

	Eu	Zr	Nb	Ta
This study	0.130	6.70	0.403	0.024
T2017	0.0234	11.47	0.472	0.028

140 SIA5 References

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Appendix 2

SIB1 Evidence for AFC processing

As mentioned briefly in the main text, when samples from the entire Diamante-Maipo caldera complex are considered, the correlations between $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ and indices for fractionation and assimilation (MgO and Eu/Eu^*) provide strong evidence for AFC (figure SB1)

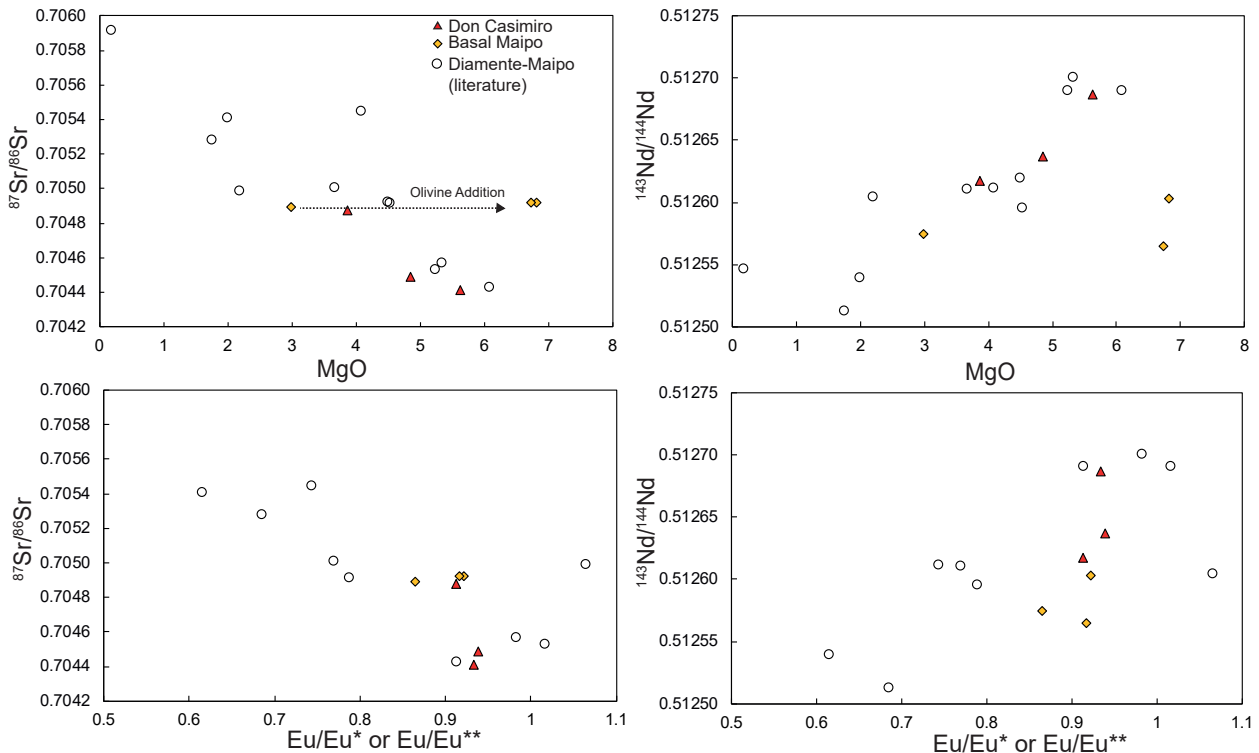


Figure SB1: Within the Diamante-Maipo Caldera, isotopic enrichment increases with assimilation and fractionation. The signal of olivine addition for the two most primitive Maipo samples is also clear.

SIB2 Evidence for Olivine Addition

EMPA analysis indicates that Don Casimiro-Maipo olivines lie significantly below the olivine-whole rock equilibrium line (Figure SB2a), indicating excess olivine accumulation. The two most primitive samples, which are from the basal section of Maipo volcano, can be related to the otherwise coherent liquid line of descent among the Diamante Caldera samples by the addition of 7 wt. % Fo 81 olivine (figure SB2b-d.). All samples have phenocrysts of olivine and clinopyroxene, so this trend is not consistent with fractionation of these two phases together. Olivine may have been accumulated from a mush pile produced by fractionation before clinopyroxene appears on the liquidus. This serves as an excellent example of the importance of utilizing individual mineral analyses in conjunction with whole-rock data when interpreting subtle geochemical variations within individual volcanic groups. While it is tempting to represent the parental composition of a volcanic group using the most

olivine-phyric samples (6-7vol% olivine, in this case), such samples can often have more complicated petrogenetic histories.

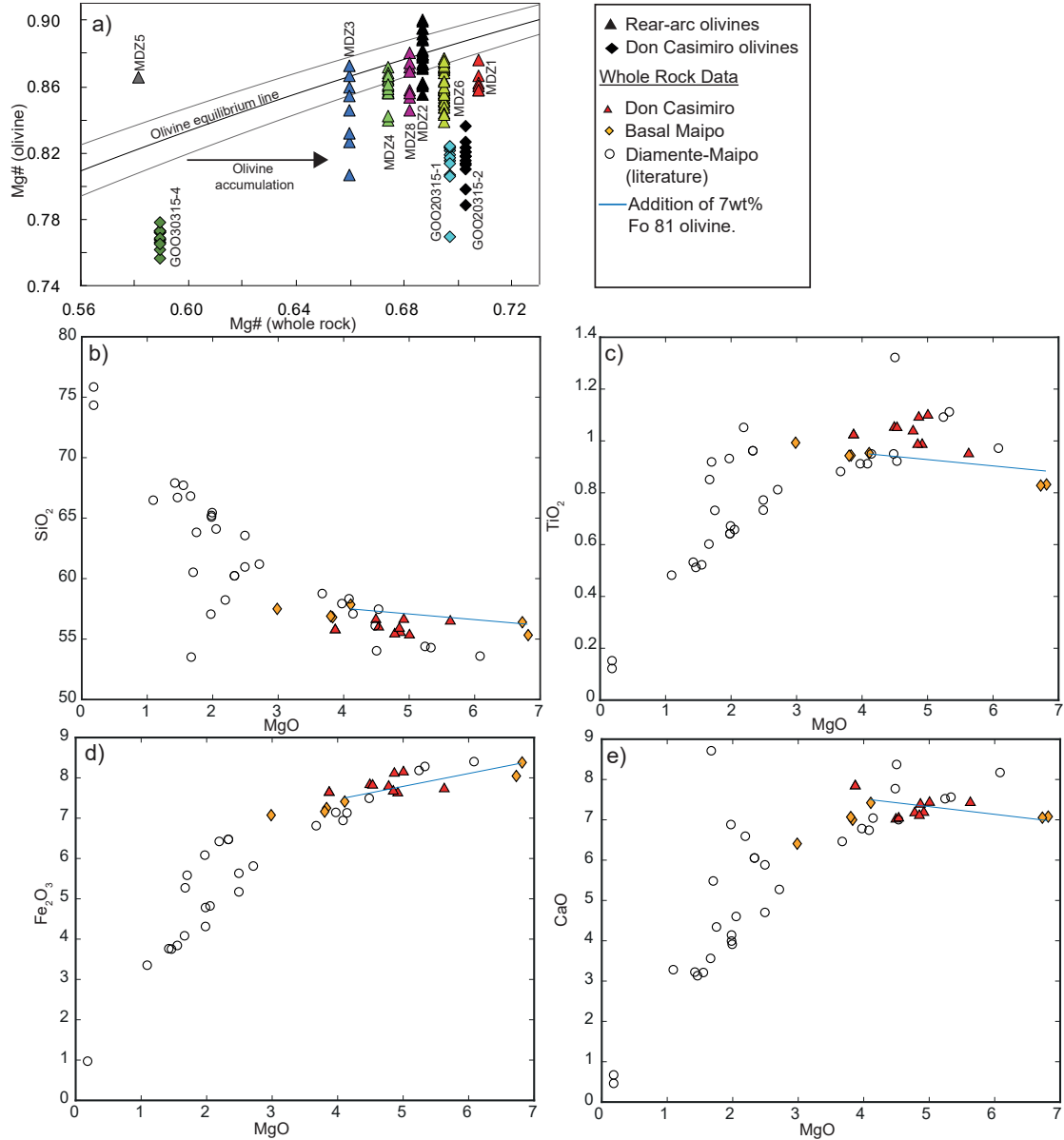


Figure SB2: a) Don Casimiro-Maipo samples show strong evidence for olivine addition, with olivine compositions plotting below an equilibrium line calculated using Roeder and Emslie (1970), $K_d=0.3\pm 0.03$, assuming whole-rock $Fe^{3+}/Fe_T=0.3$ (Holm et al., 2016). Most rear-arc olivines plot near to or below the equilibrium line, suggesting minor olivine addition (assuming whole-rock $Fe^{3+}/Fe_T=0.15$, Søger et al., 2015b). b-d) Addition of 7 wt% Fo 81 olivine from more evolved Maipo samples produces a good match to the most primitive samples, showing that their high MgO contents are spurious. Diamante-Maipo whole-rock data literature data used in this study is from Hickey et al. (1986), Futa and Stern, (1988), Sruoga et al. (2005), Holm et al. (2014) and Hickey et al. (2016).

SIB3 Changes in Slab Conditions Inferred from Zr-Hf Anomalies

Using Hf/Sm as a proxy, we find that rear-arc and most arc-front volcanics have a similar range of Zr-Hf values, but Don Casimiro-Maipo has a less pronounced Zr-Hf anomaly (figure SB3). The smaller Hf/Sm (and thus Zr-Hf anomaly) at Don Casimiro-Maipo may reflect the fact that the slab is slightly hotter, which causes residual slab zircon to become more soluble in the slab melt (Hirai et al., 2018).

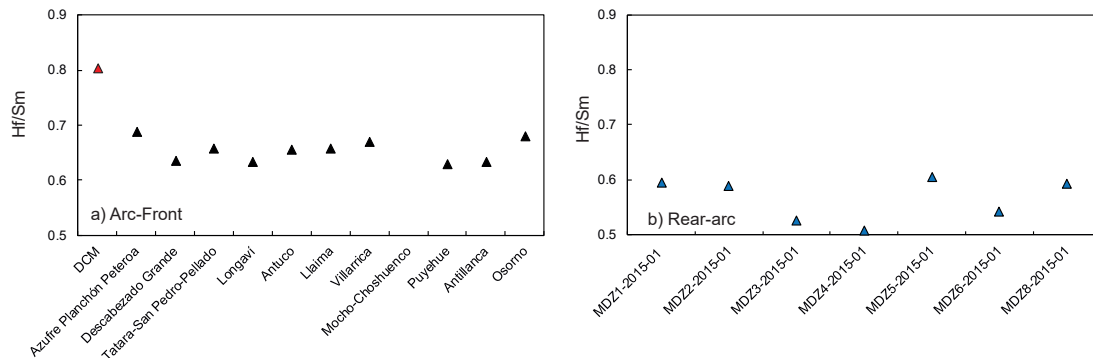


Figure SB3: a) Don Casimiro-Maipo has a significantly different Hf/Sm signature to the rest of the SVZ (data for other centers from Turner et al., 2016). b) Rear-arc centers of this study have similar Hf/Sm ratios to the SVZ arc-front as a whole.

SIB4 Assessing the evidence for LCC assimilation in the SVZ rear-arc

As discussed in section 4.5 of the main text, the offset trajectories of rear-arc and arc-front lavas in trace element and isotope space have been explained by various regional models. We show that these offsets can instead be accounted for a model incorporating variable amounts of EM1 enrichment, combined with variable slab fluxes and melt extents (figure 13 of the main text). This three-parameter model can also account for the offset trajectories in Nd-Hf isotope space (figure SB4). Arc-front and rear-arc lavas both spread along a mixing line between depleted and enriched mantle sources. Arc-front lavas are shifted to the right of the trajectory defined by rear-arc samples due to higher slab fluxes. Additionally, while rear-arc lavas appear to trend towards the composition of LCC xenoliths (Kay et al., 1996) in Ba/Th vs U or Ce/Pb space, calculated mixing lines are hyperbolic, and do not reproduce the trajectory of the rear-arc array (Figure SB5). To accommodate the mis-match between the LCC xenolith mixing model and the rear-arc data, Soager et al. (2013) instead adapt a model LCC composition that has been adjusted for several elements. These adjustments, however, diverge substantially from both global models of the LCC composition as well as the compositional field defined by the larger compilation of Andean basement lithologies (figure SB6).

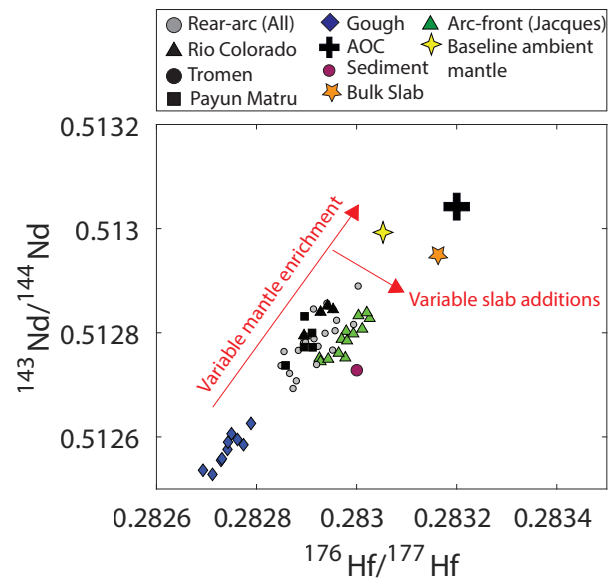


Figure SB4: The offset trends of rear-arc and arc-front lavas in Hf-Nd isotope space can be accounted for a unified model with variable EM1-type enrichment combined with variable slab inputs (see Turner et al., 2017).

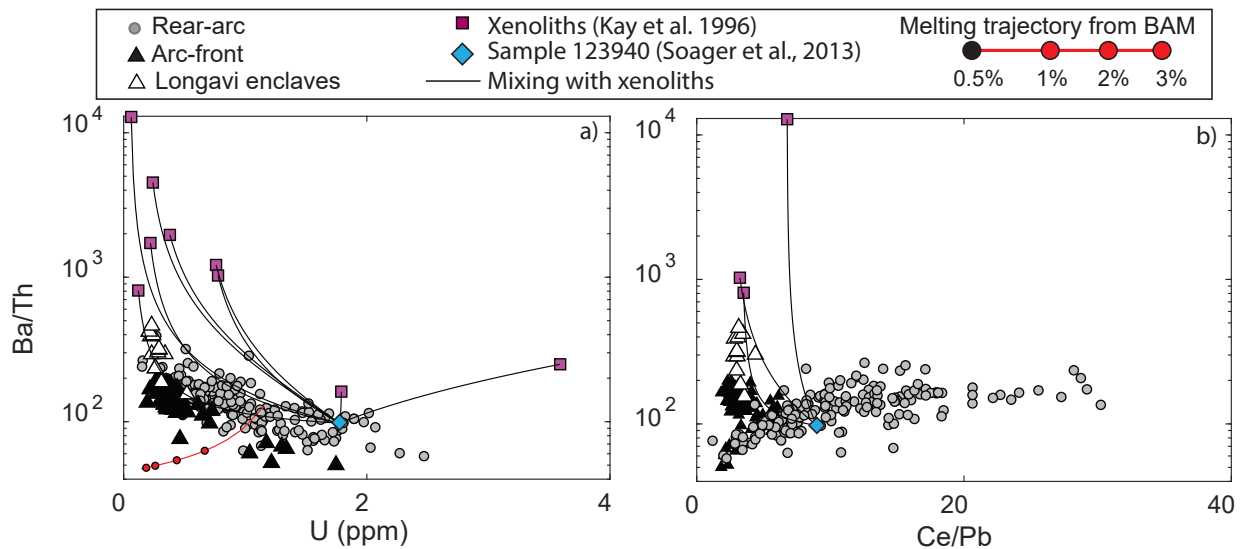


Figure SB5: Soager et al. (2013) suggest that the trajectory of rear-arc lavas trend towards the composition of xenoliths sampled by Kay et al. (1996) in Ba/Th vs. U and Ce/Pb space. However, calculated mixing trajectories between sample 123940 (used in the modelling of Soager et al., 2013) and these xenoliths fail to recreate the rear-arc trends on both plots.

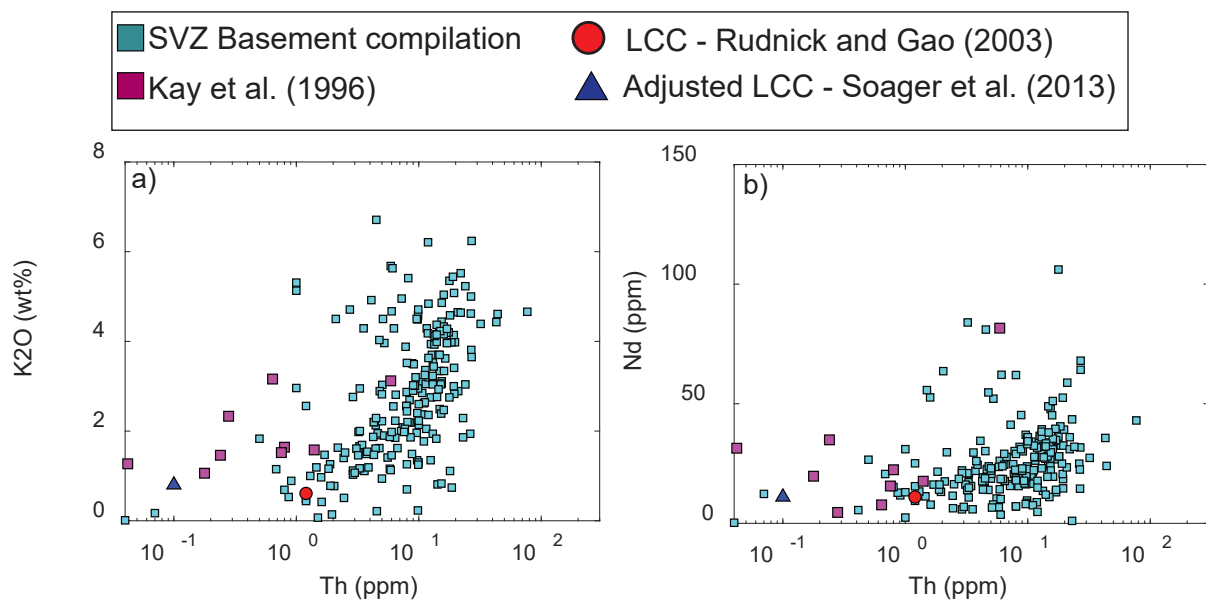


Figure SB6: Comparison of SVZ basement compositions (cyan squares) to the composition of the xenoliths of Kay et al., (1996), the LCC (Rudnick and Gao, 2003), and the basement composition used in the modelling of Soager et al. (2013). It is clear that neither the xenoliths nor the LCC composition of Rudnick and Gao (2003) is particularly representative of the composition of SVZ basement lithologies. Given that the LCC composition of Rudnick and Gao (2003) already deviates substantially from observed SVZ basement lithologies, the adjustments made to the global LCC composition by Soager et al. (2013) are difficult to justify.

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