

Persistent Geochemical Zonation (“Striping”) within the Galápagos Mantle Plume

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Highlights:

- There are temporal variations in the degree of isotopic enrichment associated with different ‘stripes’ in the Galápagos Mantle Plume.
- Shallow mantle processes control the average depth of melting and the degree of geochemical enrichment in the resulting basalts.
- Northward migration of the Galápagos Spreading center above the plume center has modified the temporal trends in lava chemistry in predictable ways.
- Therefore, there is no need to invoke temporal variations in plume chemistry and source proportions over the last ~15 Myrs in the Galápagos.

Abstract

Some hotspot tracks, such as those formed by the Hawai’i and Galápagos mantle plumes, exhibit long-lived cross-track isotopic zonation, thought to reflect the streaking out of heterogeneous material in the plume conduit during upwelling. In lavas associated with the Galápagos mantle plume, three geochemical domains, present for at least 15 Myr, have been identified: northern, southern and central. The most extreme isotopic enrichments are observed in the northern domain of the Cocos Ridge at ~15 Ma, and in the southern domain of the Galápagos Archipelago at the present day. Owing to the northward migration of the Galápagos Spreading Centre above the plume since ~20 Ma, this relationship suggests that geochemical enrichment in the Galápagos basalts is always greatest above the region of the plume furthest from the nearby mid-ocean ridge. Here we demonstrate that the temporal variations in geochemical enrichment associated

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36 with Galápagos plume stripes likely reflect "shallow mantle control" associated with differences
37 in the mean depth of melting. We conduct forward melting models of a mixed peridotite-
38 pyroxenite mantle to calculate the isotopic composition of the resulting melts formed under two
39 mantle flow regimes. Our results demonstrate that variations in the average pressure of melt
40 generation, due to the influence of the nearby ridge axis, may explain the range of isotopic
41 compositions across ~15 Ma of Galápagos plume-related volcanism. The patterns of isotopic
42 zonation observed along the hotspot track strongly confirm the paradigm of persistent plume
43 striping, with variations in the degree of geochemical enrichment modulated by shallow mantle
44 processes.

45

46 **Introduction**

47 Mantle plumes and associated ocean-island basalts (OIBs) provide important insights into the
48 composition, dynamics, and heterogeneity of the Earth's convecting mantle (Hofmann, 1997;
49 Jackson et al., 2008; Weis et al., 2023; White et al., 1993; White and Hofmann, 1982). Variations
50 in the isotopic compositions of OIBs suggest contributions from deep (primordial?) mantle
51 reservoirs, recycled crust and re-fertilized lithospheric mantle, indicating a complex and
52 heterogeneous structure derived from the lower mantle (Cabral et al., 2013; Homrighausen et al.,
53 2020; Jackson et al., 2008; Williams et al., 2015; Zindler and Hart, 1986). These components are
54 entrained into rising mantle plumes and, depending on their viscosity, are likely stretched into
55 thin streaks of material that melt along with the plume matrix at shallow pressures, giving rise to
56 a range of isotopic compositions measured at the surface (Farnetani et al., 2012, 2018; Harpp et
57 al., 2014a; Hoernle et al., 2000; Kerr and Mériaux, 2004; Lohmann et al., 2009).

58

59 Spatial zonation in the composition of mantle plume-related basaltic lavas suggest the presence
60 of distinct geochemical domains or plume ‘stripes’, characterized by the presence of spatially-
61 variable chemical and lithological heterogeneities (Abouchami et al., 2005; Chauvel et al., 2012;
62 Gleeson et al., 2021; Harpp et al., 2014a; Harpp and Weis, 2020; Hoernle et al., 2000, 2015;
63 Huang et al., 2011; Payne et al., 2013; Weis et al., 2011). Classic examples are the Galápagos
64 (Hoernle et al., 2000; Werner et al., 2003), where three compositionally enriched ‘domains’ have
65 been observed, and Hawai’i, where basalts associated with the Loa trend display enriched
66 isotopic compositions relative to the Kea trend (Abouchami et al., 2005; Harpp and Weis, 2020;
67 Tatsumoto, 1978). The presence of compositional striping is now recognized as a common
68 feature of many plume structures worldwide, including the Marquesas, Easter, Samoa, and
69 Tristan-Gough Island/seamount chains, and may be connected to seismically identified
70 heterogeneities in the Earth’s lower mantle (Gleeson et al., 2021; Harpp et al., 2014b;
71 Homrighausen et al., 2023, 2020; Rohde et al., 2013; Weis et al., 2023)

72

73 Laboratory tank experiments and numerical models of thermal or thermo-chemical plumes
74 arising from the base of the mantle provide support for plume zonation/stripping (Farnetani et al.,
75 2012; Farnetani and Hofmann, 2009; Farnetani and Samuel, 2005; Kerr and Mériaux, 2004;
76 Lohmann et al., 2009). For example, Farnetani and Hofmann (2009) show that geochemically
77 distinct material from the lowermost mantle, once entrained into a plume conduit, gets stretched
78 into long filaments. This process leads to long-lasting spatial zonation of the plume conduit,
79 which affects the spatio-temporal isotopic variability of erupted lavas.

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81 Within individual plume stripes the isotopic composition of basalts is not constant, but instead
82 varies as a function of position above the plume conduit and orientation of plume motion relative
83 to each chemical domain (Farnetani et al., 2012; Hanano et al., 2010). What is less certain,
84 however, is whether adjustments in plume chemistry (that is, changes in the composition of
85 plume heterogeneities or proportions of different components) are also required to explain
86 temporal variations within an individual plume stripe. If variations in mantle plume composition
87 through time can be identified, this could have significant implications for the size, persistence,
88 and/or distribution of streaked heterogeneities within plume structures, and could also help
89 identify the “end-member” stripe compositions.

90

91 In this study, we examine potential changes in the composition of plume stripes associated with
92 the Galápagos mantle plume. Specifically, we compare the composition of basalts associated
93 with the Galápagos Archipelago (formed during the last 3 Ma) to those erupted at ~15 Ma along
94 the northern Galápagos plume track (Cocos Ridge; Fig. 1). We use numerical mantle melting
95 models, namely the Python3 package pyMelt (Matthews et al., 2022), to investigate the origin of
96 temporal variations in basalt chemistry. Our analysis focuses on the northern isotopic domain,
97 identified by Hoernle et al. (2000) and Werner et al. (2003), and examines whether differences in
98 the mean melting depth could cause the variations in geochemical enrichment associated with the
99 northern domain basalts over the last ~15 Ma. Our results demonstrate that, despite substantial
100 variations in the degree of isotopic enrichment, the ~15 Ma record of Galápagos plume-related
101 volcanism can be straightforwardly explained by variations in the mean melting pressure,
102 without having to invoke variations in plume composition and/or the proportion of different
103 plume components.

104

105 **Geological Background**

106 Basalts erupted along the Galápagos hotspot track contain melt contributions from at least three
107 isotopically-enriched mantle reservoirs, together with a depleted plume component (Geist et al.,
108 1988; Gibson et al., 2012; Gleeson et al., 2021, 2020; Harpp and White, 2001; Hoernle et al.,
109 2000; White et al., 1993). Importantly, the three isotopically-enriched regions are geographically
110 separate and spatially-distinct (Hoernle et al., 2000; White et al., 1993). In the northern
111 Galápagos, high $^{208}\text{Pb}/^{206}\text{Pb}$ ratios – found in basalts from the islands of Pinta, Wolf, Darwin, and
112 surrounding seamounts – characterize the “northern domain” (Fig. 2). In the western Galápagos,
113 Sr, Pb, and Nd isotope ratios, similar to the globally defined FOZO or ‘C’ plume component
114 (Hanan and Graham, 1996), together with elevated $^3\text{He}/^4\text{He}$ ratios define the “central domain”
115 (Kurz and Geist, 1999). Finally, in the southern Galápagos, radiogenic Pb isotope ratios (e.g.,
116 $^{206}\text{Pb}/^{204}\text{Pb} > 19.5$) associated with the island of Floreana and, to a lesser extent, the volcanoes of
117 Sierra Negra and Cerro Azul on Isabela define the “southern domain” (Harpp et al., 2014a;
118 Harpp and White, 2001; Hoernle et al., 2000).

119

120 Volcanism in the Galápagos Archipelago is the product of a near-ridge (“off-axis”) mantle plume
121 (see Fig. 1). The plume center is currently located about 150 – 250 km south of the Galápagos
122 Spreading Center (GSC); a mid-ocean ridge spreading center separating the Cocos and Nazca
123 tectonic plates (Canales et al., 2002; Gibson et al., 2015; Gleeson and Gibson, 2021; Harpp et al.,
124 2003; Mittal and Richards, 2017; Mittelstaedt et al., 2014, 2012; Schilling et al., 2003, 1982).
125 However, relative motion between the plume and ridge has led to substantial changes in the
126 plume-ridge geometry through time.

127

128 Volcanism above the mantle plume and plume-ridge interaction have led to the formation of two
129 separate plume tracks, the NE-SW orientated Cocos Ridge on the Cocos plate, and the E-W
130 orientated Carnegie Ridge on the Nazca plate (Harpp et al., 2005; Hoernle et al., 2000; Werner et
131 al., 2003). Between them, these aseismic ridges preserve a ~20 Myr record of plume volcanism
132 and plume-ridge interaction, tracking the evolving tectonic setting of the Galápagos mantle
133 plume and GSC during this period (Fig. 1). Furthermore, the relative volume of these ridges, and
134 the age of plume-related basalts relative to the underlying oceanic crust formed at the ridge axis,
135 provide important insights into the relative motion of the GSC and Galápagos mantle plume over
136 the last ~20 Myrs (Werner et al. 2003).

137

138 The E-W (and originally NE-SW) orientation of the GSC combined with the eastward motion of
139 the Nazca plate and north-east motion of the Cocos plate has resulted in a net northward
140 migration of the GSC, relative to the Galápagos mantle plume, since ~20 Ma (Mittelstaedt et al.,
141 2012; Werner et al. 2003). This northward progression, however, is complicated by a series of
142 ridge jumps that reflect the capture of the GSC by the mantle plume during its northward
143 migration. Existing plate reconstructions indicate that, at ~19.5 Ma, the GSC was likely anchored
144 to the northern portion of the Galápagos plume (Werner et al. 2003), leading to heightened
145 volcanism on the Nazca tectonic plate and increased volume of the Carnegie Ridge and
146 associated Malpelo Ridge. Around ~14.5-15 Ma, the orientation and location of plate separation
147 changed significantly, moving the center of plate separation to the southern margin of the mantle
148 plume (Fig. 3; Werner et al. 2003). Until 11-12Ma volcanism continued with the plume
149 positioned to the north of the spreading center, leading to a reduced volume of the Carnegie

150 Ridge and heightened activity along the Cocos Ridge (this is evidenced by the notable saddle in
151 the Carnegie Ridge in bathymetric maps – see Orellana-Rovirosa and Richards (2018), for an
152 analysis of how this process has modulated the complementary volumes of volcanism along the
153 two ridges in time). The new plume-ridge orientation resulted in an increased volume to the
154 northern Cocos Ridge – now located near the Costa Rica coastline – and the emergence of a
155 paleo-archipelago in this region at around 13 – 14.5 Ma (Werner et al., 2003, 1999). Notably,
156 basalts erupted as part of the northern Cocos Ridge are older than the underlying oceanic crust
157 (15 – 19 Ma with increasing age to the north, derived from magnetic stripes; Barchhausen et al.,
158 2001; Werner et al., 2003), supporting the interpretation that these basalts formed from plume-
159 related volcanism north of the spreading axis.

160

161 After ~11 Ma, likely due to the passage of a ridge offset across the plume center (Werner et al.
162 2003), the axis of plate spreading migrated northward relative to the Galápagos mantle plume.
163 Ultimately, this resulted in a situation not dissimilar to present-day Iceland, with the mantle
164 plume located almost directly beneath the ridge axis. The northward motion of the GSC relative
165 to the mantle plume caused enhanced melting south of the ridge axis and renewed growth of the
166 Carnegie Ridge on the Nazca plate. Since ~ 5Ma, the GSC has migrated to the north of the
167 Galápagos plume, with the distance between the plume and ridge controlled by a series of south-
168 ward ridge jumps (Mittelstaedt et al. 2012). Ultimately, this has led to further broadening of the
169 Carnegie Ridge and the construction of the present-day Galápagos Platform and volcanic islands
170 (Fig. 2).

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172 In this study, we focus our analyses and models onto two of the time periods outlined above,
173 reflecting in large part the availability of geochemical data as per Hoernle et al (2000) and
174 Werner et al. (2003). First, we focus on basalts along the Cocos Ridge that erupted at ~13 – 14.5
175 Ma. These samples, collected from the northernmost portion of the Cocos Ridge near Costa
176 Rica-Panama coastline, are found on oceanic crust that is ~2 – 5 Ma older than the plume-related
177 lavas and formed when the GSC was located near the southern margin of the upwelling plume
178 stem. Second, we consider the isotopic composition of basalts found in the present-day
179 Galápagos Archipelago, with the center of the upwelling plume stem >100 km south of the GSC.
180 These two suites of plume-related lavas represent an interesting point of comparison and allow
181 us to investigate how the position of the mantle plume relative to the GSC influences the
182 enrichment of erupted plume-related material.

183

184 Importantly, spatially distinct northern, central, and southern isotopic domains – linked to those
185 observed in the Galápagos Archipelago – have been identified in the northern Cocos Ridge lavas
186 by Hoernle et al. (2000). As a result, it has been proposed that the spatial distribution of the three
187 isotopically enriched components in the Galápagos mantle plume has been stable for at least ~15
188 Myrs. However, significant variations in the degree of isotopic enrichment of basalts within each
189 domain do exist, which raises uncertainties regarding the stability and longevity of plume
190 striping. Most notably, the basalts assigned to the northern plume domain along the Cocos Ridge
191 have Sr, Nd, and Pb isotope signatures that are outside the range of values measured in any <3
192 Ma basalt from the Galápagos Archipelago (but retain the high $^{208}\text{Pb}/^{206}\text{Pb}$ characteristics of the
193 northern domain; Hoernle et al., 2000). In fact, these isotopic compositions lie outside the range
194 that could be achieved by mixing previously estimated end members in the Galápagos plume

195 (Harpp and White, 2001; Hoernle et al., 2000). In contrast, the Pb and Sr isotope compositions of
196 basalts from the Galápagos island of Floreana (southern domain) are more radiogenic than those
197 measured in basalts assigned to the southern domain on the Cocos Ridge (Harpp et al., 2014a;
198 Hoernle et al., 2000). In this study, we provide a detailed analysis of variable geochemical
199 enrichment associated with the different stripes of the Galápagos mantle plume, investigating
200 whether the changes in chemical enrichment over 15 Ma require modifications in the plume
201 source (e.g., changing plume chemistry and/or proportions), or instead result from physical
202 variations in melting processes. In particular, we focus on the greater isotopic enrichment of
203 northern domain lavas, relative to the southern domain, from the Cocos Ridge at ~15 Ma and
204 *vice versa* for the present-day Galápagos-ridge system where the southern domain is more
205 enriched relative to the northern domain.

206

207 **Methods and Modeling**

208

209 The primary objective of this study is to evaluate how variations in melting parameters (e.g., the
210 average depth of melting) influence the isotopic composition of the resulting magmas. To do so,
211 we use the Python package pyMelt, which uses empirical parameterizations of the melting
212 behaviour for various mantle lithologies to simulate partial melting of a multi-component mantle
213 (Matthews et al., 2022). pyMelt assumes that the different sources are in thermal equilibrium but
214 are chemically isolated, this is a reasonable assumption since thermal diffusivity is much faster
215 than chemical diffusivity. For each lithology, the trace element content of the melt is calculated
216 by pyMelt via built-in partition coefficients taken from Gibson and Geist (2010), and
217 assumptions about the mineral proportions in the different source lithologies.

218

219 All models presented in this study assume a simple 2-component mantle, with an isotopically
220 depleted lherzolite (representing the plume “matrix”) and an isotopically enriched pyroxenite
221 (representing streaks of recycled crust and/or refertilized lithosphere; Hoernle et al., 2000;
222 Sobolev et al., 2007; Sun et al., 2020). The melting behavior of these two source components is
223 modeled in pyMelt using the melting parameterizations of the KLB-1 lherzolite and KG1
224 pyroxenite from Matthews et al. (2021) (Fig. 4). The potential temperature of the Galápagos
225 mantle plume (T_p) was set at 1400-1420 °C, with values between 1420 and 1500°C also tested.
226 The pressure of melt cessation, here representing the base of the lithosphere, was set between 1.3
227 – 1.9 GPa, consistent with geochemistry-constrained estimates of lithospheric thickness beneath
228 the Galápagos Archipelago (Gibson and Geist, 2010).

229

230 Here we examine the extent to which variations in the melting process affect the chemistry of
231 erupted basalts. First, we consider the chemistry of melts produced during transfer of material
232 from the plume stem towards the ridge axis on the near-ridge margin of the upwelling plume
233 stem (Fig. 5, blue arrows). In this scenario we assume that the upwelling velocity of plume
234 material passing beneath the near-ridge region (i.e., in the northern Galápagos) is constant,
235 controlled by the upwelling of material into the region created by the spreading plates (i.e.,
236 passive upwelling in near-ridge settings). This is supported by the geodynamic modelling of Ito
237 and Bianco (2014), whose results demonstrate that across a ~100 km depth interval there is little
238 to no change in the upwelling velocity of material during lateral transport to the ridge axis.
239 Throughout the rest of the paper, this scenario is referred to as “Model 1” or the “Ridge Flow
240 Model”.

241

242 Our second scenario considers the flow of mantle material on the side of the plume conduit that
243 is furthest from the nearby spreading center (i.e., beneath Volcan Cerro Azul and Isla Floreana at
244 the present day; Fig. 5, orange arrows). Here, gravitationally driven flow and plate separation do
245 not contribute to the upwelling of plume material. Instead, there is a strong depth dependence to
246 the upwelling velocity within the plume margin as the velocity vectors change from nearly
247 vertical at the base of the melting region to approximately horizontal at the base of the
248 mechanical lithosphere (Fig. 5). This scenario is described in detail by Ito and Mahoney (2005),
249 and we utilize their model of depth dependent lateral flow rates to estimate variations in the
250 relative melt production rate with pressure. Throughout the rest of the paper, we refer to this
251 scenario as “Model 2” or the “Plume Flow Model”. Models 1 and 2 are shown graphically in Fig.
252 5, which was adapted from a 3D geodynamic model of plume-ridge interaction by Ito and Bianco
253 (2014).

254

255 To calculate the isotopic and trace element composition of the accumulated melts we first need to
256 make some assumptions regarding the trace element and isotopic compositions of the different
257 source regions. The isotopic composition of the depleted lherzolite component in the Galápagos
258 mantle plume is taken from Harpp and White (2001), with the trace element chemistry set to that
259 of the Workman and Hart (2005) depleted MORB mantle (DMM). The rare earth element (REE)
260 composition of the enriched, pyroxenitic component in the northern Galápagos plume domain
261 was set as a 50:50 mixture of the Stracke et al. (2003) estimate for subducted basaltic crust and
262 the DMM. The Sr and Pb contents of this component were treated independently, with the
263 “enrichment factor” of each element (i.e., $[\text{Sr}]^{\text{pyroxenite}}/[\text{Sr}]^{\text{lherzolite}}$) iteratively adjusted to improve

264 the fit between the models and observations. Finally, because we focus on the northern
265 Galápagos domain, the isotopic signature of the pyroxenite component was set to the
266 composition of the most enriched value measured in basalts from the northern domain of the
267 Cocos Ridge (Hoernle et al., 2000; Werner et al., 2003).

268
269 To examine the spread of compositions that may be produced by mantle melting in the two
270 models shown in Figure 5, we used two approaches to simulate the variability in erupted basalt
271 chemistry. First, we use a Monte Carlo approach to sample uniform prior distributions for the
272 plume potential temperature (1400-1420°C), the pressure of melt cessation (1.3–1.5 GPa, Model
273 1; 1.4-1.9 GPa, Model 2), and the source proportions (pyroxenite fraction, $X_{px}=0.12 - 0.13$;
274 bounds iteratively adjusted to best fit the natural data). We also run a separate series of models to
275 examine how slight variations in these core variables influence the composition of the mean
276 accumulated melt phase, with the source proportion of pyroxenite having the dominant control
277 on the predicted isotopic composition of melts produced by each Model (Fig. 6). Second, we
278 simulate incomplete homogenization of mantle melts following the approach of Rudge et al.
279 (2013), using a Dirichlet mixing function to examine random mixing of mantle melts (with the
280 probability of sampling each melt determined by proportion of melt produced at each depth; Fig.
281 5). The extent of homogenization in these calculations is described by a single parameter N ,
282 where $N = 1$ indicates complete isolation (no mixing) of all melts produced in the melting region,
283 whereas $N = \infty$ represents complete homogenization of all mantle melts (see Supplementary
284 Material).

285

286 **Results**

287 Variations in isotopic enrichment between basalts from the Galápagos Archipelago and the ~13 –
288 15 Ma segment of the Cocos Ridge can be summarized by two main differences: (i) greater
289 isotopic enrichment of the northern domain along the Cocos Ridge, and (ii) greater enrichment of
290 the southern domain in the modern-day Galápagos Archipelago (Harpp et al., 2014, 2005;
291 Hoernle et al., 2000; Werner et al., 2003). The temporal changes in isotopic compositions of
292 erupted basalts could, in theory, have several possible origins, the simplest of which would be
293 variations in the composition of the mantle source (either through changing proportions of
294 mantle sources or variations in the composition of mantle end members; Fig. 6b). This
295 explanation appears to be supported by the extreme isotopic compositions observed in the
296 northern domain of the Cocos Ridge (Hoernle et al., 2000), where Sr-Pb-Nd isotope values plot
297 outside the field of compositions enclosed by previously proposed mantle end-member
298 compositions associated with the Galápagos plume (Harpp and Weis, 2020; Harpp and White,
299 2001). However, the 15 Ma northern domain Cocos Ridge lavas are characterized by the same
300 high $^{208}\text{Pb}/^{206}\text{Pb}$ signature that is typical of <3 Ma basalts from the islands of Wolf, Darwin,
301 Pinta, surrounding seamounts, and western GSC (Figs. 2 & 3). It is difficult to explain how this
302 signature is maintained if the composition of the northern domain mantle source components
303 vary through time, barring appeal to coincidence. Instead, this implies a stable “enriched” mantle
304 component in the northern Galápagos across this 15Ma record of Galápagos plume volcanism,
305 with a source isotopic signature that is more enriched than that proposed for the “Wolf-Darwin”
306 component of Harpp and White (2001). Compositional variations in basaltic lavas must,
307 therefore, originate through temporal fluctuations in mantle source proportions, or “shallow
308 mantle” processes such as the mean pressure of melting. Here we consider the second option,
309 tracking melt compositions formed during flow towards a nearby spreading center (“Model 1”)

310 and beneath a thicker lithospheric ‘lid’ on the far side of the plume stem (“Model 2”). For these
311 calculations we focus solely on the northern isotopic plume domain, due to ongoing uncertainties
312 in the lithology and origin of the southern plume component (Gleeson et al., 2021; Vidito et al.,
313 2013), but emphasize that our results are applicable to any scenario where an isotopically
314 enriched, more fertile mantle component starts to melt at greater pressures than a depleted
315 peridotite matrix during adiabatic decompression.

316

317 For each model we ran 200 simulations, randomly varying the mantle potential temperature,
318 lithospheric thickness, and source pyroxenite component as described above. In every simulation
319 we ‘sampled’ the magmas formed in the melting region 200 times using a Dirichlet Mixing
320 Model, following the method outlined in Rudge et al. (2013), to estimate the composition and
321 compositional variation of erupted magmas (N=2000). Ultimately this approach generated
322 40,000 hypothetical compositions for magmas formed in each model. The results shown in
323 Figure 7 demonstrate a significant shift in the average pressure of melting from ~2.3 GPa in
324 Model 1 to 3.15 GPa in Model 2, driven by differences in the modelled rates of melt formation at
325 each depth with variations in lithospheric thickness having a secondary influence (Fig. 6, 7). The
326 compositions generated in Model 2 consistently display more radiogenic Sr and Pb isotope
327 signatures and less radiogenic Nd isotope signatures compared to Model 1 (Fig. 7).

328

329 Isotope plots demonstrate that the Model 1 melts have isotopic compositions that closely
330 resemble basalts from the island of Pinta and surrounding submarine ridge (Fig. 7). The results
331 from Model 2 simulations stretch to more isotopically enriched compositions, overlapping with
332 the range of values measured in northern domain basalts from the Cocos Ridge. Notably, the

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333 compositions predicted by Model 2 simulations are outside the range of radiogenic isotope
334 values measured anywhere in the Galápagos Archipelago. As a result, our simulations
335 conclusively show that the isotopic composition of basalts from the northern Galápagos island of
336 Pinta and the northern domain of the Cocos Ridge can be recreated by the two melting models
337 considered in this study. Furthermore, owing to the greater contribution of high-pressure melts in
338 Model 2 – where lower degrees of melting in the presence of garnet occurs – significant trace
339 element enrichment, relative to Model 1, is observed in our simulations. Because of the large
340 uncertainty in the trace element content of the different mantle sources we don't try to directly
341 recreate the trace element profiles of the Galápagos and Cocos Ridge lavas, but instead consider
342 the trends in trace element enrichment/depletion between the two models. Model 2 melts display
343 a ~1.3 times enrichment in the $[Dy/Yb]_n$ ratio (where n indicates normalization to the primitive
344 mantle of Sun and McDonough, 1989), consistent with the enrichment observed in the Cocos
345 Ridge northern domain lavas relative to Pinta. These middle-to-heavy REE ratios are
346 predominantly controlled by the presence of garnet in the mantle source and, therefore, track
347 variations in the mean depth of melting consistent with the hypothesis presented here. Our
348 models also demonstrate significant enrichment in the light REE in the Model 2 melts (~1.4 –
349 1.5X). This enrichment is slightly less than the 2X enrichment observed between the Cocos
350 Ridge northern domain and Pinta lavas, which might originate through the absence of hydrous
351 melting regimes in our models and or slight errors in the source trace element contents.
352 Nevertheless, the differences between Model 1 and Model 2 melts recreate both the isotopic and
353 broad-scale trace element differences between the Cocos Ridge northern domain and Pinta lavas.
354

355 In summary, our results demonstrate that the composition of basalts from both the island of Pinta
356 (northern Galápagos Archipelago) and the northern plume domain of the Cocos Ridge can be
357 reproduced without having to invoke variations in mantle source chemistry, proportions, or
358 potential temperature (Fig. 6). Instead, we find that variations in the mean melting depth – driven
359 by predictable variations in mantle flow rates on either side of the Galápagos plume stem –
360 control the geochemical enrichment of individual plume stripes. Furthermore, the spread of data
361 at each location can be reproduced by accounting for minor stochastic variations in source
362 proportions, melting depths and mantle potential temperatures, as well as incomplete mixing and
363 homogenization of the melts produced (Fig. 7). To best fit the Galápagos Archipelago and Cocos
364 Ridge data, we propose that the northern domain in the Galápagos plume contains ~12.5 %
365 pyroxenite streaks in a depleted peridotite matrix. We also find that the northern enriched
366 pyroxenite source is likely represented by the isotopic composition of the most enriched northern
367 domain basalt from the Cocos Ridge. We stress that these values are non-unique solutions, as
368 different assumptions regarding the source lithology/volatile content etc. will likely influence the
369 estimated source composition. The actual end member composition of course could be more
370 extreme, lying on the northern domain array at more radiogenic Sr and Pb and less radiogenic Nd
371 isotopic composition. Nevertheless, we emphasize that the geochemical variations within the
372 Galápagos plume stripes can successfully be modelled without temporal changes in plume
373 chemistry/proportions.

374

375 **Discussion**

376 The presence of spatial zonation of isotopic domains in mantle plumes has been documented at
377 several locations worldwide, with variations in the composition of basalts erupted in each zone

378 linked to the distribution of heterogeneities within the plume structure and/or the influence of
379 plume dynamics on melt production and/or extraction (e.g., Chauvel et al., 2012; Farnetani et al.,
380 2012; Gleeson et al., 2021; Lohmann et al., 2009; Weis et al., 2011). Here we have shown that,
381 despite large and conspicuous variations in basalt chemistry, the end member compositions of
382 plume stripes within the Galápagos Archipelago are stable over timescales of at least ~15 Myrs.
383 We find that variations in the isotopic ratio of basalts originating from the northern component of
384 the Galápagos plume result from temporal changes in mean depth of melting, related to the
385 south-to-north motion of the GSC across the plume. Our results confirm the presence of
386 persistent plume striping in the Galápagos Archipelago and provide new constraints on the end-
387 member isotopic compositions of enriched plume components. Furthermore, our results indicate
388 that the northern domain is broadly homogeneous on timescales up to ~15 Myrs.

389

390 Our results do not include a full dynamical simulation of plume-ridge interaction in the
391 Galápagos region, so that melting in the Galápagos mantle plume could, in reality, deviate from
392 either “Model” proposed here. Nevertheless, our models demonstrate the importance of
393 considering melting parameters and mantle dynamics when inverting OIB geochemistry for
394 mantle geochemical signatures.

395

396 The presence of compositional striping in mantle plumes has been identified at many locations
397 worldwide (e.g., Abouchami et al., 2005; Chauvel et al., 2012; Harpp and Weis, 2020; Hoernle et
398 al., 2000; 2015; Huang et al., 2011; Rohde et al., 2013; Weis et al., 2011), with many plume
399 systems showing temporal trends in the degree of geochemical enrichment associated with
400 individual plume stripes (Chauvel et al., 2012; Homrighausen et al., 2019). Here we have shown

401 that variable enrichment within plume stripes of the Galápagos Archipelago can be explained by
402 shallow mantle processes modulating the mean depth of melting and, therefore, the degree of
403 chemical enrichment. Following on from this analysis we suggest that compositional variations
404 of lavas associated with stripes at other plume systems worldwide should be critically examined
405 in a similar manner. This could reveal where, and when, chemical variations are explained
406 through modulation of melt chemistry by shallow mantle processes and where temporal
407 variations in plume chemistry and/or source proportions are required. As a result, further analysis
408 of temporal variations in plume stripe chemical enrichment could help to quantify the typical
409 length scales of streaked heterogeneities, originating in the Earth's deep mantle, in upwelling
410 mantle plumes.

411
412 We suggest that this work should start with other near-ridge plume systems, such as the Walvis-
413 Ridge – Tristan-Gough plume track, where compositional stripes have been observed
414 (Homrighausen et al., 2019). Expansion of this analysis to intra-plate plume systems, however, is
415 also encouraged. For example, the Marquesas Islands in French Polynesia are split into two
416 geochemical tracks named the “Fatu Hiva” and “Ua Huka” groups (Huang et al., 2011; Chauvel
417 et al., 2012). Within each track there is distinct variability in erupted compositions and clear
418 temporal patterns over ~5 Myr in the overall isotopic enrichment of erupted basalts. Temporal
419 variations in lithospheric thickness and/or plume dynamics – such as in the Galápagos – are
420 unlikely to explain these trends owing to the greater lithospheric age and distance from any plate
421 boundary. Instead, subtle variations in other melting parameters, including mantle potential
422 temperatures, might explain these isotopic changes without having to invoke differences in the
423 composition and/or source proportions of the mantle plume. While we show variations in plume

424 T_p have a moderate influence on the isotopic composition of basalts erupted in the Galápagos
425 (Fig. 6), we suggest that variations in plume temperatures (and other melting parameters) might
426 have a far greater influence in systems where the melt fraction of isotopically depleted plume
427 components is low (typically <2%). For example, the Marquesas Islands are characterized by a
428 relatively low erupted flux (1 – 2 orders of magnitude lower than Hawai'i) and alkali-rich
429 basalts, indicating that small variations in melting processes could have a strong influence on the
430 isotopic composition of erupted basalts in this region (Chauvel et al., 2012).

431

432 **Conclusions**

433

434 Observed variations in the intensity of isotopic zonation of Galápagos basalts over the last 15 Ma
435 can be accounted for by variations in mean depth of melting during the northward migration of
436 the GSC above the Galápagos plume stem. The fact that these observations can be explained so
437 straightforwardly provides powerful verification of the elegant theoretical concept of mantle
438 plume geochemical striping (zonation), demonstrates the time-persistence of these stripes over at
439 least the past 15 Ma, allows us to constrain the end-member geochemical signatures of these
440 stripes, and therefore suggests that similar variations can be mapped and modeled along other
441 hotspot tracks, perhaps especially those that are near mid-ocean ridges, where temporal
442 variations in lithospheric thickness may have the most effect. This should ultimately lead to a
443 more refined definition of the nature of the true “end member” geochemical source regions for
444 mantle plumes, and their interaction with ambient mantle as they ascend from the core-mantle
445 boundary and undergo partial melting beneath the lithosphere.

446

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452

453 **Data and Code availability**

454 All code used for the production of models and figures used in this study is currently available
455 on GitHub (https://github.com/gleesonm1/GleesonEtAl_PlumeStriping), and is archived on
456 Zenodo <https://doi.org/10.5281/zenodo.18355128>. Data used in this study is taken from (Allan
457 and Simkin, 2000; Bow and Geist, 1992; Geist et al., 2006, 2002; Gibson et al., 2012; Gibson
458 and Geist, 2010; Harpp et al., 2003; Harpp and Weis, 2020; Harpp and White, 2001; Hoernle et
459 al., 2000; Kurz and Geist, 1999; McBirney and Williams, 1969; Saal et al., 2007; Standish et al.,
460 1998; Teasdale et al., 2005; White et al., 1993)

461

462 **CRedit authorship contribution statement**

463 **Matthew Gleeson:** Conceptualization, Writing – original draft, Formal analysis, Methodology;
464 **Mark Richards:** Conceptualization, Writing – review & editing; **Cinzia Farnetani:** Writing –
465 review & editing; **Kaj Hoernle:** Writing – review & editing; **Sally Gibson:** Writing – review &
466 editing.

467

468 **References**

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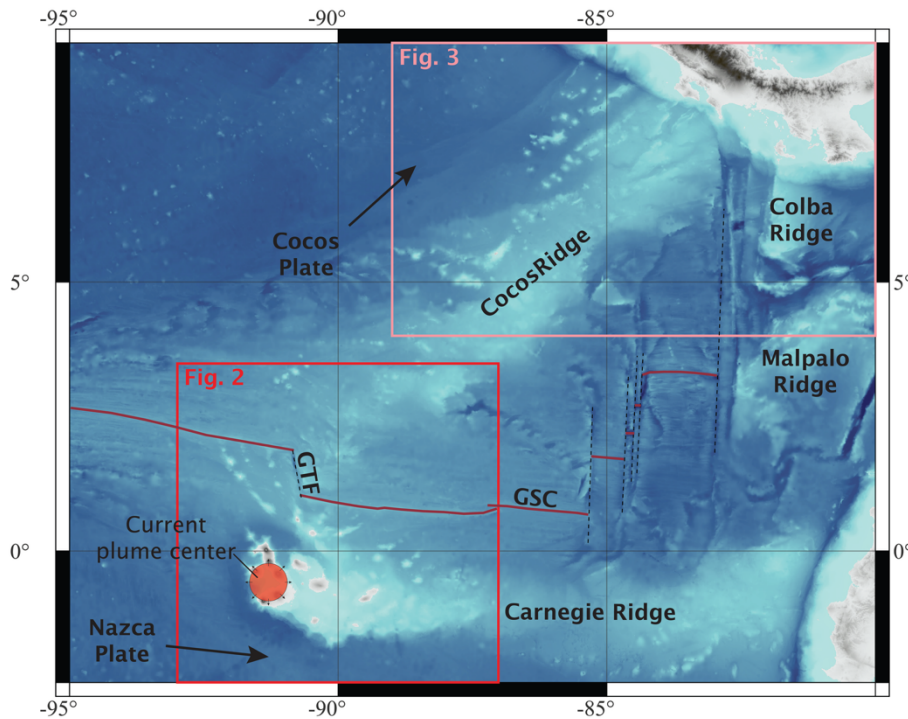


Figure 1 – Overview map of the tectonic setting of the Galápagos Archipelago and Galápagos Spreading Center (GSC). The Galápagos mantle plume is currently located under the islands of Isabela and Fernandina in the western Galápagos, around 100-200 km south of the GSC. Volcanic activity driven by plume upwelling has generated the W-E Carnegie Ridge and the SW-NE Cocos Ridge on the Nazca and Cocos tectonic plates, respectively. Combined these ridges preserve a ~20 Myr record of plume related volcanism. The position of the Galápagos Transform Fault (GTF) separating the western and eastern GSC is also shown. Bathymetric data from GEBCO (General Bathymetric Chart of the Oceans). Figure generated in QGIS using the World Geodetic Survey (1984) Geographic Coordinate System with an equirectangular projection system

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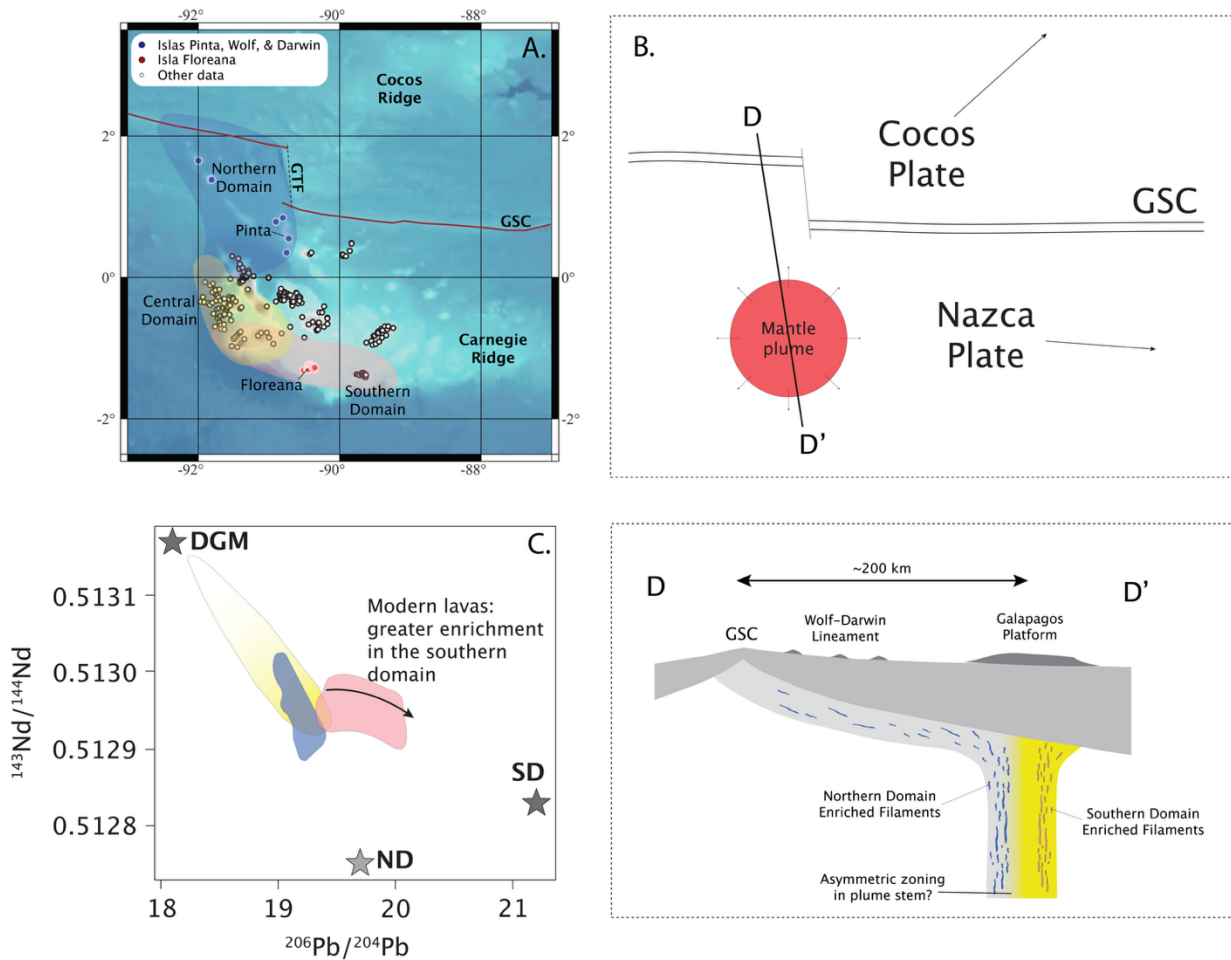


Figure 2 – Tectonic setting and geochemical signatures of the Galápagos Archipelago. **A.** Lavas associated with the Galápagos plume show evidence for at least three spatially distinct enriched plume components. These are used to define the northern domain, central domain, and southern domain. **B.** Schematic representation of the current plume-ridge geometry, with the mantle plume located approximately 100-200 km south of the ridge axis. **C.** Fields representing the range of isotopic composition of basalts from the three domains, the data indicates a high degree of enrichment in basalts associated with the southern domain (Isla Floreana). **D.** Schematic cross section of the current plume-ridge geometry, with the ridge located to the north of the center of plume upwelling. This encourages flow of northern plume material towards the ridge and a greater mean depth of melting on the southern margin of the plume stem. The ‘northern domain’ plume component (**ND**) represents the new composition/end-member proposed in this study. The composition of the Southern Domain (**SD**) plume component and Depleted Galápagos Mantle (**DGM**) are taken from Harpp and White (2001).

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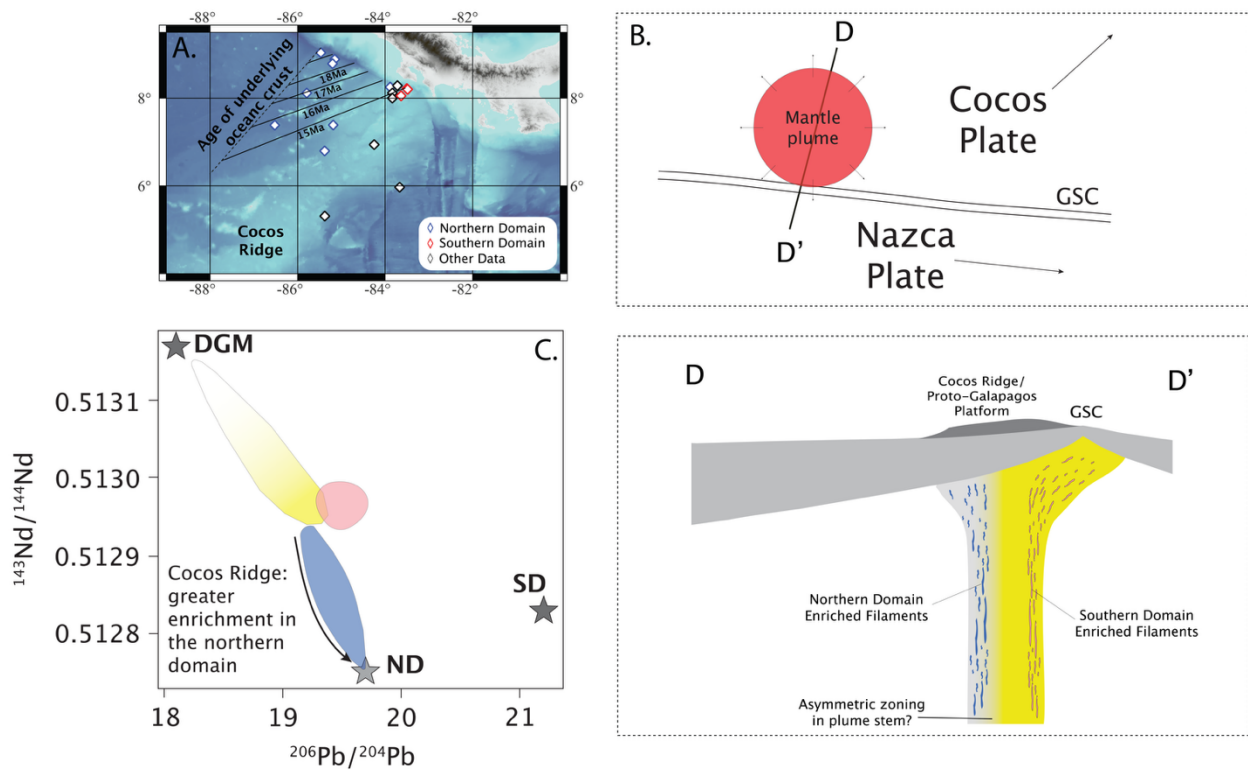


Figure 3 – Tectonic setting and geochemical signatures of the northern Cocos Ridge (lava ages from 13 – 14.5 Ma). **A.** Erupted lavas in the northern Cocos Ridge display evidence for spatial zonation in plume composition, with distinct northern, central, and southern domains consistent with the geochemical enrichment in the modern-day Galápagos Archipelago (Hoernle et al. 2000). Plume related basalts of the northern Cocos Ridge are significantly younger than the underlying oceanic crust (Werner et al. 1999). Age of the underlying oceanic crust is taken from Barckhausen et al. (2001). **B.** Schematic representation of the plume-ridge geometry proposed by Werner et al. (2003) at ~12 – 14.5 Ma. The mantle plume is proposed to lie to the north of the ridge axis, with the southern margin of the plume potentially anchored on the ridge. **C.** Fields representing the range of isotopic composition of basalts from the northern Cocos Ridge that have been linked to the three enriched domains, the data indicates a high degree of enrichment in basalts associated with the northern domain. **D.** Schematic cross section of the plume-ridge geometry at ~12-14.5 Ma, with the ridge located to the south of the center of plume upwelling. This encourages flow of southern plume material towards the ridge and a greater mean depth of melting on the northern margin of the plume stem. The ‘northern domain’ plume component (**ND**) represents the new composition/end-member proposed in this study. The composition of the Southern Domain (**SD**) plume component and Depleted Galápagos Mantle (**DGM**) are taken from Harpp and White (2001).

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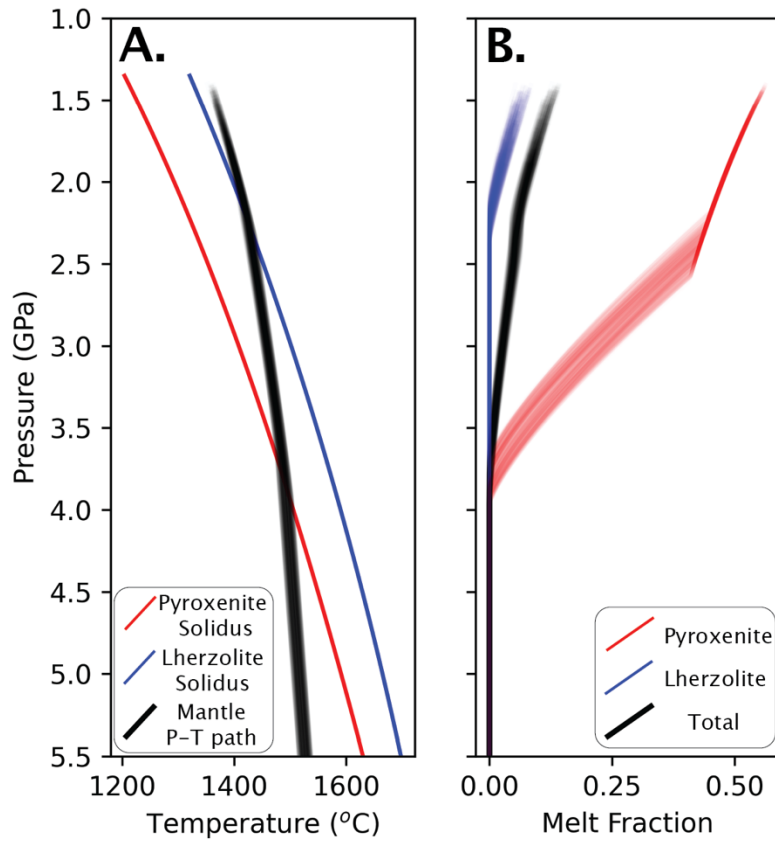
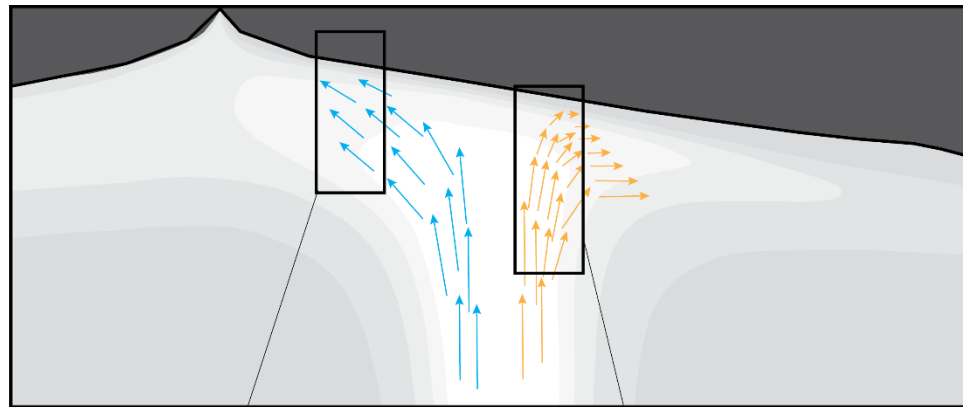


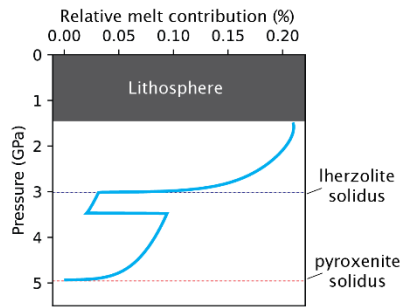
Figure 4 – Pressure vs temperature (A.) and melt fraction (B.) for the mantle melting models used in this study. 200 models are shown with the mantle potential temperature set between 1400 and 1420 °C and the proportion of pyroxenite in the mantle source between 0.12 and 0.13.

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Model 1: “Ridge Flow Model”

- Lateral deflection of plume material towards ridge
- Upwelling aided by plate separation – greater contribution of low-pressure melts.



Model 2: “Plume Flow Model”

- Plume flow controlled by lithospheric ‘lid’
- Melting in the plume stem – buoyancy driven upwelling causes stronger weighting to high-pressure melts.

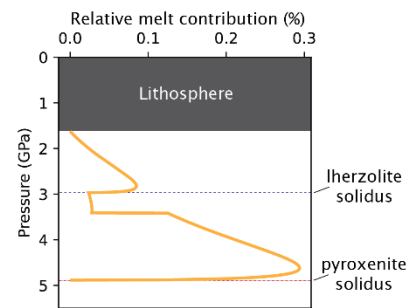


Figure 5 – Schematic illustrations of the 2 different models considered in this study. Greyscale variations in the top diagram indicate temperature variations in the mantle with hotter temperatures indicated by lighter colors (adapted from Ito and Bianco, 2014). In Model 1, we assume plume material is flowing ‘upslope’ along the base of the mechanical lithosphere, with the upwelling velocity primarily controlled by passive upwelling during plate separation. This ‘ridge flow’ results in a relatively constant upwelling velocity across the melting region beneath the islands of Pinta, Wolf, and Darwin in the northern Galápagos. In Model 2, representing the far side of the mantle plume to off-ridge island formation, mantle upwelling rates are strongly pressure dependent as the upwelling, buoyant plume mantle upwells beneath the rigid lithospheric lid (with little ‘upslope’ flow towards a spreading center possible).

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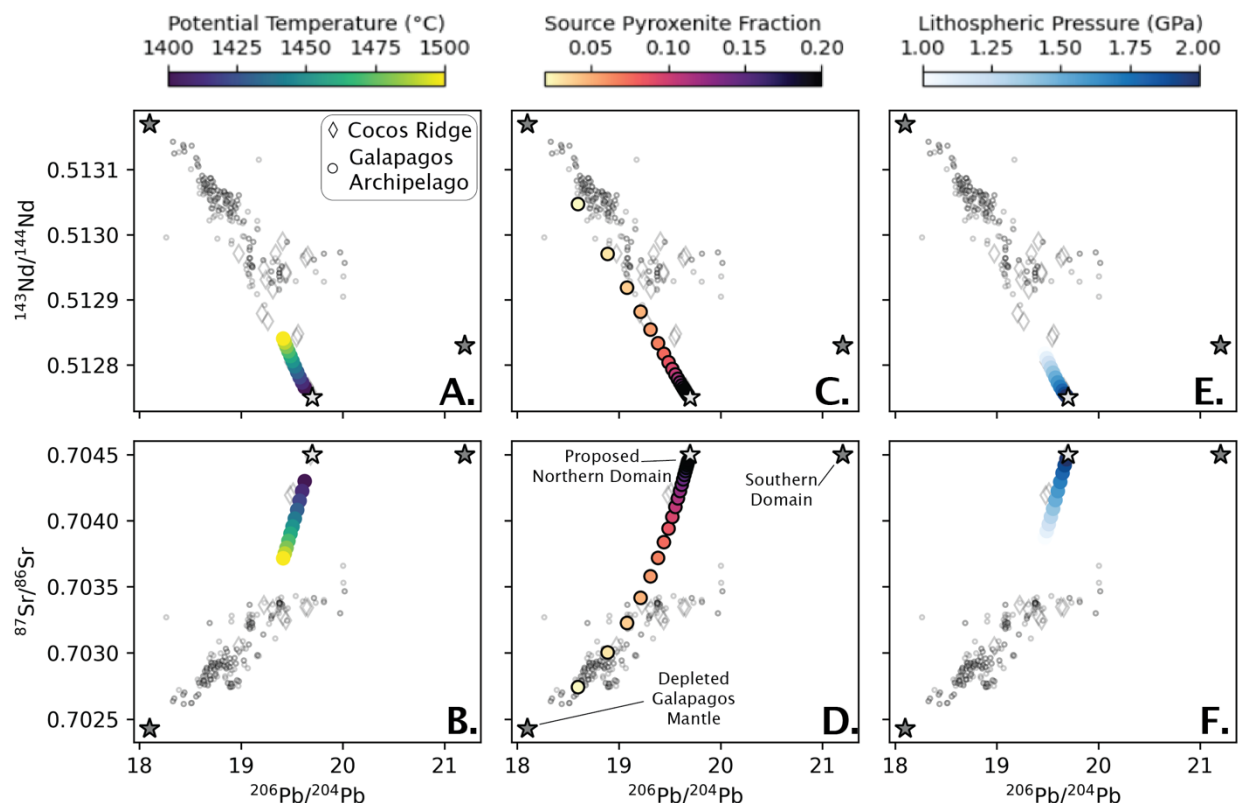


Figure 6 – Isotopic composition of basalts from the Galápagos Archipelago and the Cocos Ridge against compositions predicted in Model 2 with individual parameters varying independently. **A. & B.** Influence of mantle potential temperature. **C. & D.** Influence of source pyroxenite fraction. **E. & F.** Influence of variations in lithospheric pressure. Geochemical data from the Galápagos Archipelago and Cocos Ridge taken from: (Allan and Simkin, 2000; Bow and Geist, 1992; Geist et al., 2005, 2002, 2006; Gibson et al., 2012; Gibson and Geist, 2010; Harpp et al., 2003; Harpp and Weis, 2020; Harpp and White, 2001; Hoernle et al., 2000; Kurz and Geist, 1999; McBirney and Williams, 1969; Naumann et al., 2002; Saal et al., 2007; Standish et al., 1998; Teasdale et al., 2005; White et al., 1993).

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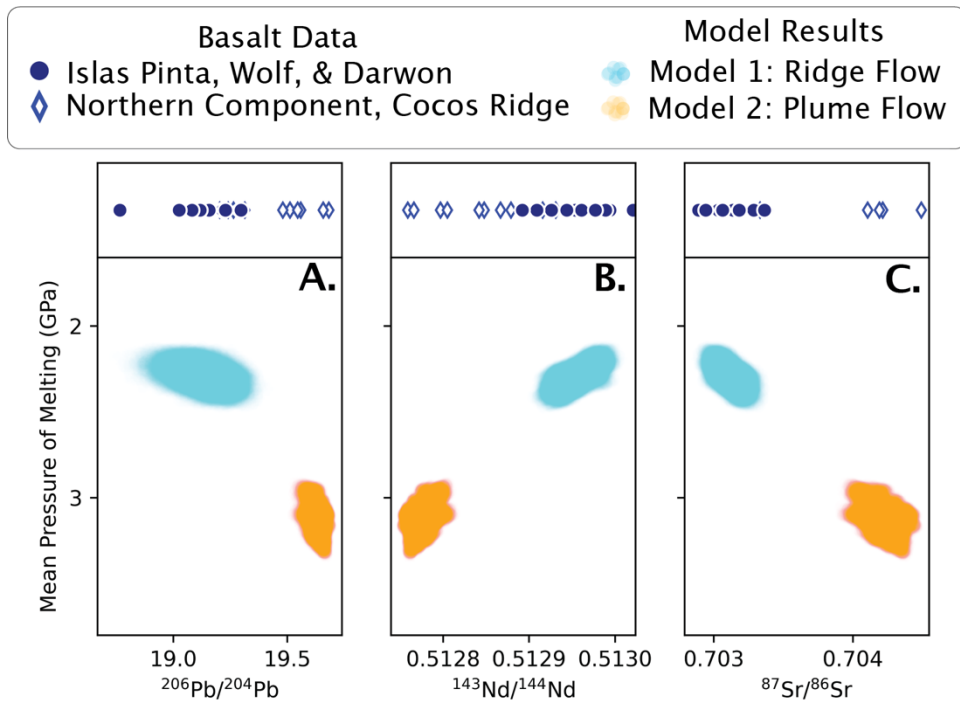


Figure 7 – Isotopic compositions predicted for melting Models 1 & 2 for **A.** $^{206}\text{Pb}/^{204}\text{Pb}$, **B.** $^{143}\text{Nd}/^{144}\text{Nd}$, and **C.** $^{87}\text{Sr}/^{86}\text{Sr}$, with Model 2 simulations typically displaying more enriched signatures overlapping with the compositions observed in the northern domain of the Cocos Ridge.

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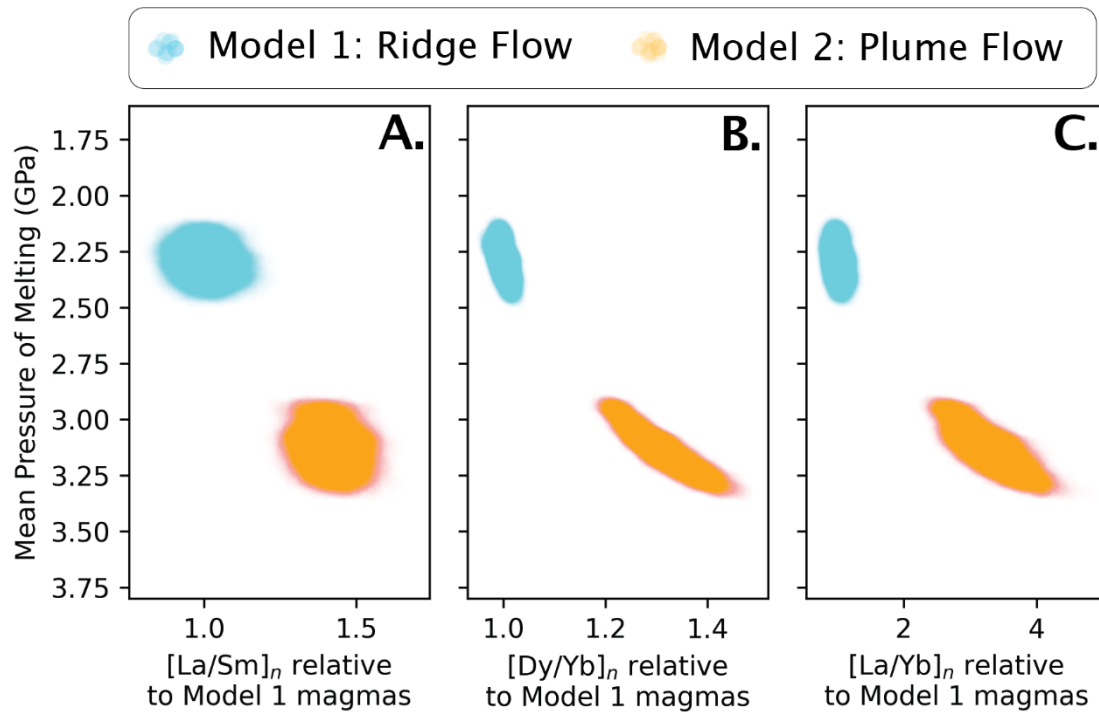


Figure 8 – Trace element compositions predicted for melting Models 1 & 2. **A.** $[La/Sm]_n$ tracks the overall enrichment of the mantle sources and is sensitive to the mean melt fraction in each model (lower for Model 2). **B.** $[Dy/Yb]_n$ tracks the influence of garnet in the mantle residue and, therefore, is sensitive to the mean pressure of melting. **C.** $[La/Yb]_n$ represents a combination of the other two parameters. The subscript 'n' indicates normalization to the primitive mantle composition of Sun and McDonough (1989).

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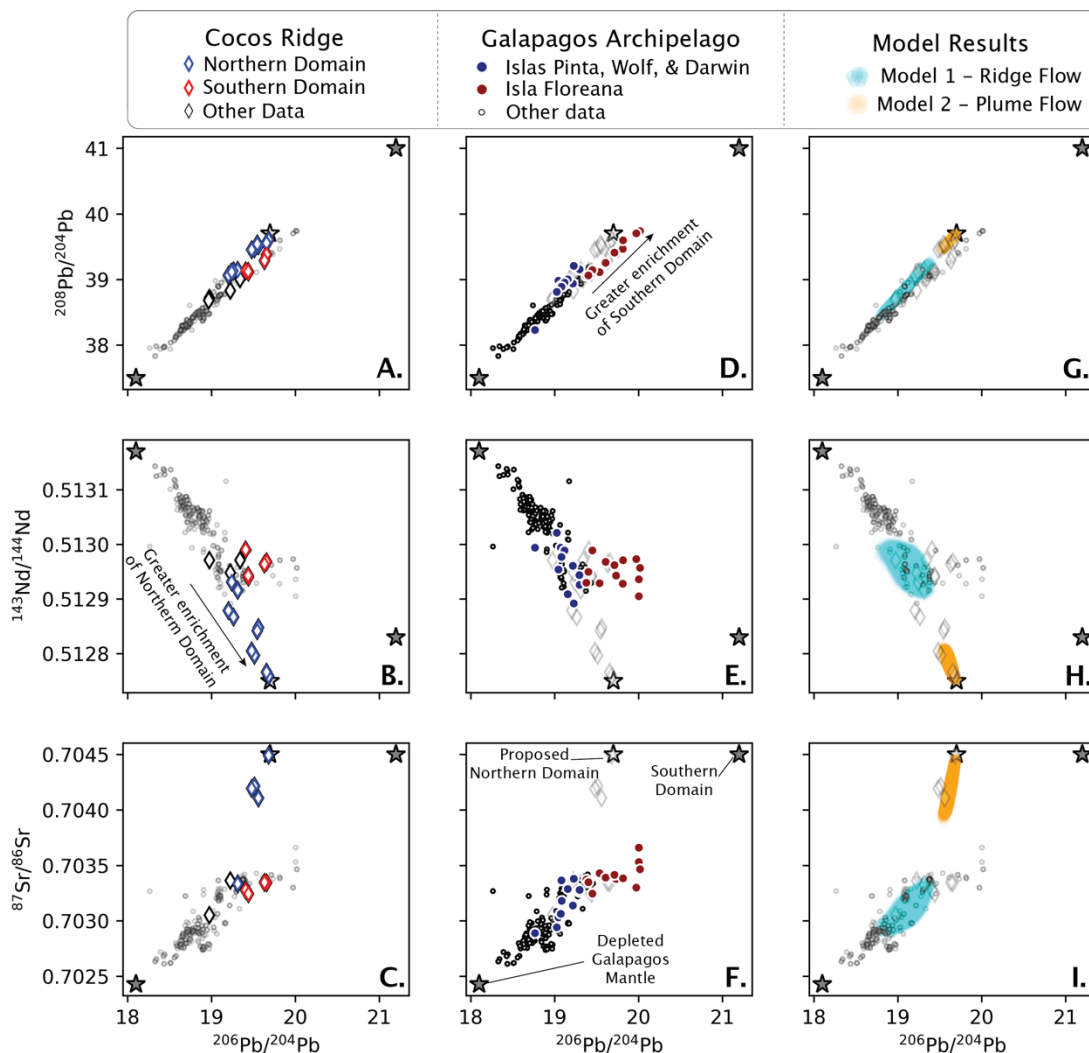


Figure 9 – Isotopic compositions of basalts from the Cocos Ridge (A. – C.), the present-day Galápagos Archipelago (D. – F.), and magmas predicted by Model 1 & Model 2 mantle melting simulations (G. – I.). In all panels the composition of the Depleted Galápagos Mantle and southern (“FLO”) component estimated by Harpp and White (2001) are indicated by the grey stars. The new composition proposed for the northern isotopic domain in the Galápagos mantle plume is also indicated by the lighter grey star in all panels. In panels A. – C. data from the Cocos Ridge displays a strong geochemical enrichment in basalts associated with the northern plume component. In contrast, panels D. – F. demonstrate that the strongest enrichment in the present-day Galápagos Archipelago is found in the southern isotopic domain (Isla Floreana). G. – I. Model 2 simulations produce more enriched isotopic compositions, overlapping with the enriched basalts collected along the northern margin of the Cocos Ridge (Hoernle et al. 2000).

Supplementary Material: Persistent Geochemical Zonation (“Striping”) within the Galápagos Mantle Plume

Matthew Gleeson^{1*}, Mark Richards¹, Cinzia Farnetani², Kaj Hoernle³, and Sally Gibson⁴

1. Melting behavior

Adiabatic decompression calculations are performed using the Python3 tool ‘pyMelt’ (Matthews et al. 2022) and parameterizations for the melting of the KLB-1 lherzolite and the KG1 pyroxenite from Matthews et al. (2021). Full details of the melting calculations and parameters used can be found in Matthews et al. (2021) and Katz et al. (2003), but are summarized briefly here.

Melting equations for both lithologies follow the form:

$$F = \begin{cases} T < T_{cpx-out}: \left(\frac{T - T_{solidus}}{T_{liquidus}^{lherz} - T_{solidus}} \right)^{\beta_1} \\ T > T_{cpx-out}: F_{cpx-out} + (1 - F_{cpx-out}) \left(\frac{T - T_{cpx-out}}{T_{liquidus} - T_{cpx-out}} \right)^{\beta_2} \end{cases}$$

where:

$$T_{solidus} = A_1 \ln(P + A_2) + A_3 P + A_4$$

$$T_{liquidus} = B_1 \ln(P + B_2) + B_3 P + B_4$$

$$T_{liquidus}^{lherz} = C_1 \ln(P + C_2) + C_3 P + C_4$$

$T_{solidus}$ and $T_{liquidus}$ represent empirical parameterizations of experimental and/or thermodynamic predictions of the melting behaviour for each lithology. $T_{liquidus}^{lherz}$ is the theoretical temperature at which the system would be entirely molten if clinopyroxene-present melting continued beyond $F_{cpx-out}$, the melt fraction at which clinopyroxene is exhausted. $F_{cpx-out}$ is determined by:

$$F_{cpx-out} = \frac{M_{cpx}}{R_{cpx}(P)}$$

where:

$$R_{cpx}(P) = r_0 + r_1 P$$

2. Partition coefficients and source compositions

In this study we used trace element partition coefficients from Gibson and Geist (2010) and based our source compositions on prior estimates for the trace element content of the depleted MORB mantle (Workman and Hart, 2005) and subducted igneous crust (Stracke et al. 2003). The parameters used in our models are summarized in the tables below (Table S.1 & S.2):

Table S.1 – partition coefficients used in mantle melting models to perform trace element calculations

Element	Olivine	Clinopyroxene	Orthopyroxene	Garnet	Spinel
Pb	0.003	0.012	0.009	0.005	0.0
Sr	0.00004	0.091	0.0007	0.0007	0.0
La	0.0005	0.049	0.0031	0.001	0.01
Ce	0.0005	0.08	0.004	0.005	0.01
Pr	0.0008	0.126	0.0048	0.014	0.01
Nd	0.00042	0.178	0.012	0.052	0.01
Sm	0.0011	0.293	0.02	0.25	0.01
Eu	0.0016	0.335	0.013	0.496	0.01
Gd	0.0011	0.35	0.013	0.848	0.01
Tb	0.0015	0.403	0.019	1.477	0.01
Dy	0.0027	0.4	0.011	2.2	0.01
Ho	0.0016	0.427	0.0065	3.315	0.01
Er	0.013	0.42	0.045	4.4	0.01
Yb	0.02	0.4	0.08	6.6	0.01
Lu	0.02	0.376	0.12	7.1	0.01

Table S.2 – source trace element contents (all values reported in parts per million) used in this study.

Element	Depleted MORB Mantle (DMM) Workman and Hart (2005)	Subducted Igneous Crust (SIC) Stracke et al. (2003)	KG1 estimated composition (50:50 DMM:SIC excluding Sr & Pb)
Pb	0.018	0.09	0.18
Sr	7.664	81	26.824
La	0.192	1.68	0.926
Ce	0.55	5.89	3.220
Pr	0.107	n/a	n/a
Nd	0.581	7.45	4.0155
Sm	0.239	2.69	1.4645
Eu	0.096	1.04	0.568
Gd	0.358	4.03	2.194
Tb	0.07	n/a	n/a
Dy	0.505	5.01	2.7575
Ho	0.115	n/a	n/a
Er	0.348	3.13	1.739
Yb	0.365	2.99	1.678
Lu	0.058	0.45	0.254

3. Determining the relative melt contribution in Ridge Flow and Plume Flow models

Our study demonstrates that variations in the mean depth of melting can influence the composition of erupted basalts, without having to invoke changes in the source proportions. In the Galápagos variations in mean melting depth are controlled by differences in the plume dynamics in the melting region, resulting in an offset in the relative melt contribution from each pressure.

For Model 1 (“Ridge Flow”) we assume a constant upwelling velocity of mantle through the melting region (primarily driven by passive upwelling in response to plate separation). Consequently, calculating the relative melt contribution from each depth/pressure interval to the final, homogenized melt composition is trivial:

$$w_{lith}^P = F_{inc_{lith}}^P * X_{lith}$$

where w_{lith}^P represents the relative contribution of melts derived from lithology ‘*lith*’ at pressure ‘*P*’. Similarly, $F_{inc_{lith}}^P$ is the incremental melt fraction of this lithology at the specified pressure and X_{lith} is the mass fraction of the specified lithology in the mantle source (prior to the initiation of melting). Importantly, w values are then normalized so that sum of the w values across each pressure interval and all different lithologies is equal to 1.

Constraining the depth- and lithology-dependent melt contribution factors is, however, more complex for the Model 2 (“Plume Flow”) calculations. In this scenario there is a depth dependence to the upwelling velocity but, instead of calculating this directly, we chose to use the parameterization of Ito and Mahonney (2005) for the horizontal velocity of material leaving the melting region to determine the relative contribution of melts from each depth. The Ito and Mahonney (2005) parameterization states that the horizontal velocity of material leaving the melting region in a mantle plume setting is:

$$U(z) = 2 \left(\frac{z}{H} \right) - \left(\frac{z}{H} \right)^2$$

Where z represents the depth of the region of interest (below the lithosphere) and H represents the maximum depth of the region of horizontal plume flow. By tracing the mantle components exiting the melting region at different depths we can calculate the depth-dependent melt contribution in regions away from the influence of a nearby spreading center. In this case we take H to also represent the base of the melting region to reduce the number of parameters that need to be fitted in our modelling. In addition, if the mantle has a constant average density across the melting region we can replace $\frac{z}{H}$ in the equation above with P' , where $P' = \frac{P - P_{min}}{P_{max} - P_{min}}$ (P_{max} represents the max pressure of melting and P_{min} represents the pressure of melt termination).

$$U(P') = 2P' - P'^2$$

The relative melt contributions from each depth in the melting region can then be calculated:

$$w_{lith}^P = \sum_0^{P'} (U(P') * F_{inc_{lith}}^P * X_{lith})$$

Ultimately this approach provides very similar depth-dependent melt contribution curves as assuming an exponential decrease in upwelling velocity of mantle material through the melting region with decreasing pressure (and 0 upwelling velocity at the base of the lithosphere).

4. Influence of the mixing parameter N

Incomplete homogenization and mixing of mantle melts from different lithologies and pressure intervals is simulated using a Dirichlet mixing function following the methods outlined by Rudge et al. (2013). We take this approach to try and simulate the natural variation in lava compositions associated with the northern domain of the Galápagos plume. The influence of varying the mixing parameter N is shown in Fig. S.1.

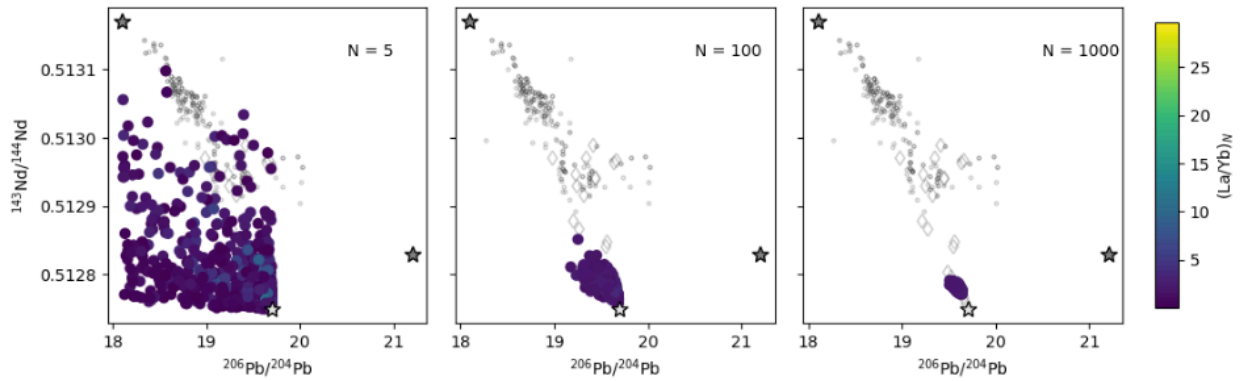


Fig S.1 – Influence of varying the mixing parameter “N” on the composition of basalts sampled from the melting region. With increasing values of N the compositions converge on the ‘true’ average melt composition.

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