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1 **Melting below Mid Ocean Ridges: thermal effects of pyroxenites in the peridotitic**
2 **mantle**

3

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13 *After travelling in the Earth's interior for up to billions of years, recycled material once*
14 *injected at subduction zones can reach a subridge melting region as pyroxenite dispersed*
15 *in the host peridotitic mantle. We studied genetically related crustal basalts and mantle*
16 *peridotites sampled along an uplifted lithospheric section created at a segment of the Mid*
17 *Atlantic Ridge through a time interval of 26 Ma. The arrival of low-solidus material into*
18 *the melting region forces the elemental and isotopic imprint of the residual peridotites and*
19 *of the basalts to diverge with time. We show that a pyroxenite-bearing source entering the*
20 *subridge melting region induces undercooling of the host peridotitic mantle, due to*
21 *subtraction of latent heat by melting of the low-T solidus pyroxenite. Mantle undercooling*
22 *in turn lowers the thermal boundary layer leading to a deeper cessation of melting. A*
23 *consequence is to decrease the total amount of extracted melt, hence magmatic crustal*
24 *thickness. The degree of melting undergone by a homogeneous peridotitic mantle is higher*
25 *than the degree of melting of the same peridotite but veined by pyroxenites. This effect,*
26 *thermodynamically predicted for a marble-cake type peridotite-pyroxenite mixed source,*
27 *implies incomplete homogenisation of recycled material in the convective mantle.*

28

30 Mantle rising beneath the 60,000 km long Mid Oceanic Ridge system contains, as in a
31 slow-motion movie, a record of ancient upwelling and melting events and of interaction
32 with subduction or hot spot-derived components. It is difficult to reconstruct temporal
33 records of these ancient events due to lack of suitable samples; however, we were given
34 the opportunity to explore the temporal evolution of the oceanic lithosphere composition
35 and structure at 11°N along the Mid Atlantic Ridge (MAR) where an uplifted > 300 km
36 long sliver of lithosphere exposes a basal mantle peridotite unit, lower crustal gabbros, a
37 dyke complex and erupted basalts¹⁻³. This lithospheric section (Vema Lithospheric Section
38 or VLS) was generated at an 80 km long segment of the MAR (EMAR segment,
39 Supplementary Fig. S1) during a 26 Ma time interval^{1,2,4-6}. Both crustal basalts and their
40 mantle peridotite parents have been densely sampled at the VLS along a seafloor spreading
41 flow line⁴ allowing comparisons of their isotopic and elemental composition throughout
42 the 26 Ma time interval^{1,2,4-6} (Fig. 1).

43 Surprisingly, temporal variations of mantle degree of melting estimated from basalt Na_8 ^{7,8}
44 anti-correlate with the degree of melting derived from spinel Cr# of the peridotites^{9,10},
45 although the two curves converge to a common value in the youngest 3 Ma stretch of the
46 VLS (Fig. 1). Older, isotopically enriched basalts display the lower Na_8 values of the entire
47 VLS, suggesting they were generated by a higher degree of melting of their mantle source;
48 in contrast, the genetically associated mantle peridotites record a relatively low extent of
49 melting, in agreement with a thinner crust recorded by geophysical data¹. This anti-
50 correlation contrasts with what is inferred to be the “normal” signature of partial melting at
51 mid-ocean ridges.

52 We offer a solution to this conundrum by suggesting that a subridge variably veined
53 mantle hosts chemically enriched, fertile, low-T melting components, i.e. pyroxenites.
54 Thermodynamic-based studies predict dramatic effects when pyroxenites are present in the
55 mantle source and partially melt along a decompressive path¹¹⁻¹⁵. Low-T solidus
56 components lower the extent of melting of the host peridotite due to subtraction of latent
57 heat of fusion¹³⁻¹⁵. Pyroxenites melt preferentially, generating isotopically enriched, low
58 Na_8 melts and cooling the host mantle peridotites, thereby lowering the degree of melting
59 of the peridotite mantle in proportion to its pyroxenite content¹³. In this work we account
60 for variable extents of the melting column by assuming that different pyroxenitic contents

61 result in variable extent of undercooling before the ambient peridotite starts melting. This
62 situation mimics varying the potential temperature of the mantle resulting in changes in the
63 length of the melting column.

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66 Temporal variability of crustal thickness and mantle degree of melting

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68 Basalt Nd, Sr and Pb isotopes vary coherently along the 26 Ma-long VLS section,
69 showing a decrease in isotopic enrichment towards younger ages that hints at temporal
70 variations of composition and thermal state of the rising mantle⁸. We represent these
71 variations as discrete steps (Fig. 1a), defining three time domains: 0-2 Ma, 2-13 Ma and
72 13-26 Ma (Supplementary Tables S1-S3).

73 The basalts major element composition also varies with age, the older samples being
74 poorer in sodium than the younger ones. Na_8 , an inverse proxy of the degree of melting
75 experienced by the mantle column^{7,8} (Supplementary Table S2), is on average lower in the
76 VLS older basalts; thus, they were apparently produced by a degree of melting higher than
77 the younger basalts (Supplementary Fig. S1a). However, Na_8 has been defined for a
78 homogeneous lherzolitic source⁸. A heterogeneous source generates complex melt mixing
79 patterns depending on the relative extent of melting, homogenisation and enrichment of
80 each component¹⁶⁻¹⁸. For this reason we adopt here the expression “apparent” degree of
81 melting.

82 The proxy equivalent (but reverse) to basalt Na_8 is in peridotites the Cr#
83 ($Cr\# = Cr/(Cr+Al)$) of spinels and pyroxenes^{9,10}. During the last 26 Ma, peridotite spinel
84 Cr# increased along the VLS on average from 22 to 37 (Supplementary Fig S1c)
85 suggesting that the amount of magma delivered at ridge axis increased through time, in
86 agreement with gravity profiles running along spreading-flow-lines (Supplementary Fig.
87 S1c), revealing that crustal thickness increased from 4.8 ± 0.2 km in the 22-27 Ma interval
88 to 5.4 ± 0.2 km between 0-5 Ma^{1,4}.

89 Based on calibrations of Warren, 2016¹⁹, the degree of melting (F_{max}) of the VLS
90 peridotites increased from 8.0 to 14.2 F% toward younger crustal ages (average 10.8, Fig.
91 2). MORB glasses Na_8 increased with time along the VLS from 2.6 to 3.0 on average
92 (Supplementary Fig. S1a). F_{max} in the basalts can be estimated according to⁷ and compared
93 to the F_{max} of the mantle peridotites. Comparing temporal sections of mantle residua and of
94 their melt products must take into account a time-delay in the emplacement of the mantle

95 peridotites in the oceanic crust. According to^{1,4} we corrected the crustal ages of the mantle
96 rocks relative to that of basalts by a relative time lag of 2.2 Ma (see online Methods: Age
97 Correction).

98 The calculated apparent F_{\max} of MORB glasses vary little during the 26 Ma-long VLS
99 stretch (Fig. 1b) with values ranging from 18.0 to 15.8 F% (ave. 16.7), significantly higher
100 than those estimated from mantle residual peridotites (Fig. 1b). A striking feature of the
101 Na_8 degree of melting curve is the decrease of the apparent F_{\max} through time that
102 countertrends with both the associated mantle peridotites degree of melting curve and the
103 gravity-inferred crustal thickness (Fig. 1b; Supplementary Fig. S1c).

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105 Significance of the decoupling

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107 The degrees of melting estimated from the mantle peridotites and from the basalts can
108 be generated by variations of mantle potential temperature, mantle composition and
109 spreading rate. Changes in mantle temperature or source fertility will result in coherent
110 changes of the degrees of melting estimated from basalts and from peridotites. Similarly,
111 changes in spreading rate cannot decouple the behaviour of residual mantle and extracted
112 basalts, because, in a passive upwelling scenario, decreasing the spreading rate lowers the
113 thermal state of the entire melting region, and vice versa²⁰. During the last 26 Ma the half
114 spreading rate at the EMAR segment decreased from 17.2 mm/a (Chron 6) to 16.9 mm/a
115 (Chron 5) to present-day 13.6 mm/a^{21,22} (Supplementary Fig. S1b). Such decrease of
116 spreading rate toward younger ages should lower the mantle degree of melting by about
117 1%²⁰; thus, the increase in degree of melting recorded by the VLS mantle peridotites must
118 be caused by processes other than changes in spreading rate, mantle temperature or
119 fertility.

120 We consider now a heterogeneous mantle source. Thermodynamic modelling of
121 melting a two-component mantle source predicts that when a fertile heterogeneity, i.e.
122 pyroxenite, starts melting, the temperature of the whole mantle parcel is lowered due to the
123 latent heat of melting¹⁵. If the heterogeneity is less than a few kilometres, some heat is
124 transferred from the peridotite into the melting heterogeneity increasing its melt
125 productivity while cooling the surrounding mantle¹³⁻¹⁵. Accordingly, the vertical interval
126 where only pyroxenites undergo melting represents an undercooling region whose extent is
127 proportional to the amount of pyroxenites (Fig. 2). It follows that, for a given P-T
128 decompression path, the degree of melting undergone by a homogeneous peridotitic

129 mantle is higher than the degree of melting of the same peridotite but veined by
130 pyroxenites.

131 A lithologically homogeneous mantle source, resulting in coherent estimates of the
132 degree of melting between peridotites and basalt proxies, is approximated in the younger
133 (< 5 Ma) portion of the VLS where both basalt- and peridotite-derived F% converge
134 toward a common value. We assume that the present-day subridge mantle (Vema
135 Unveined Mantle or VUM) contains negligible amounts of pyroxenites, not sufficient to
136 perturb thermally the melting process of the host mantle peridotite. Thus, for this region
137 the Na₈-derived degree of melting is in line with its original interpretation^{7,8,23}.

138 In contrast, in a veined mantle scenario, degrees of melting estimated from peridotites
139 and from basalts differ strongly. While the Cr# records the true F_{max} of the ambient
140 peridotite, the pooled melts aggregate the compositional signal of both (low F) peridotitic
141 and (high F) pyroxenitic melts¹⁸. We propose that the low degree of melting of the older
142 portion of the VLS peridotites is due to heat consumption during preferential melting of a
143 pyroxenitic component at nearly constant mantle potential temperature. Along the VLS we
144 have decreasing quantities of pyroxenites injected into the melting region, with consequent
145 decrease of the undercooling effect and expansion of the anhydrous peridotite melting
146 region.

147 We tested this hypothesis by modelling the decompressive adiabatic melting of a mixed
148 source based on the experimental-parameterized algorithm Melt-PX²⁴. Mantle potential
149 temperatures have been constrained using the passive flow temperature field model of
150 Bonatti et al., (2003)¹ giving a mantle T_p = 1350 °C. For this temperature and a lherzolithic
151 source containing 15% clinopyroxene, Melt-PX calculations overestimate crustal thickness
152 and mantle degrees of melting observed at the EMAR segment (Fig. 3). These calculations
153 assume that melting ceases at the base of the crust, a boundary condition acceptable for
154 high mantle T_p settings as in fast spreading ridges or hotspots^{24,25}. They represent the
155 model maximum allowed thickness at a given thermal setting (Fig. 3). Mantle flow
156 models^{1,26} predict the end of melting to occur at P_f ≈ 0.7 GPa, well below the base of the
157 crust. This condition applies to low-spreading ridges due to heat conduction to the surface
158 resulting in a deep transition from the conductive to the convective thermal region^{20,26–28}.

159 As low-melting component we adopted the silica deficient pyroxenite M7-16²⁹ for
160 reasons defined in the next section. We observe that an increase in the fraction X_p of
161 pyroxenite in the mantle is paralleled by a decrease of degree of melting of the host

162 peridotite (F_r) (Fig. 3) depending also on the final pressure of melting (varying in the
163 range 0.3-1.1 GPa, Fig. 3).

164 Integration of the melt productivity of the host peridotite and of the pyroxenite along an
165 adiabatic path, weighted by their relative abundance, allows estimating magmatic crustal
166 thickness. Lambart et al., (2016)²⁴ show that the total magmatic productivity increases
167 proportionally to the amount of pyroxenite in the source. This observation apparently is in
168 contrast with the increase toward younger ages along the VLS of gravity-inferred crustal
169 thickness, paralleled by a decrease of the estimated amount of pyroxenites dispersed within
170 the mantle (Fig. 2, Supplementary Fig. S1c). This contradiction is solved considering that
171 increasing the amount of low-melting lithologies enhances the undercooling of the host
172 peridotitic mantle due to heat diffusion into the melting pyroxenite¹³. Undercooling
173 estimated by Melt-PX calculations can reach up to 40 °C for adiabatic melting under the
174 assumed conditions. Reduced undercooling due to a decrease of the mantle pyroxenite
175 fraction during melting results in shallowing the final pressure of melting from ≈ 1.1 to 0.7
176 GPa going from the oldest to the youngest VLS sectors (Fig. 3, right panel).

177 These observations and numerical experiments reveal that in slow-spreading ridges, at
178 constant mantle potential temperature, the arrival of a pulse of pyroxenites in the mantle
179 source region will cause a reduction of crustal thickness. The countertrend of crustal
180 thickness and degree of mantle melting with spreading rate along the VLS is a positive test
181 of our hypothesis.

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184 Composition of the low-solidus component and effects on Na8

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186 The effect of pyroxenite-derived melts on the final composition of the pooled MORB
187 depends upon the nature of the heterogeneities and on their dilution in the peridotite-
188 derived melt^{17,30-32}. In mid-ocean ridge settings most pyroxenites produce melts with
189 major-element composition similar to those derived from peridotite³³. Hence, little
190 reactivity is expected when pyroxenite-derived melts mix with mantle peridotite-derived
191 melts³⁴. We thus explore the possible composition of pyroxenites dispersed in the VLS
192 mantle by considering mixing of silica-deficient (SD) and silica-enriched (SE) pyroxenite-
193 derived melts with peridotite-derived melts. In a Na₂O versus MgO diagram, young VLS

194 lavas plot at higher Na₂O content than the older ones (Supplementary Fig. S2). We assume
195 that the youngest VLS Na-rich lavas, i.e. those showing a degree of melting similar to that
196 of the associated peridotites, derive from melting of the VUM source defining a peridotite
197 primitive melt composition that matches the VUM average Na₈ and F% (Supplementary
198 Table S2). We then calculated mixing lines between the VUM primitive melt and
199 pyroxenitic-derived melts obtained experimentally at variable degrees of melting from
200 different sources (Supplementary Fig. S2-S3). Melt mixing affects the estimated Na₈, and
201 consequently the estimated apparent degree of melting, depending on the absolute Na
202 content and Mg/Na ratio of the added melt fraction (Supplementary Fig. S3). Inferred Na₈
203 and apparent F of the mixed compositions show that SE-derived melts do not reproduce
204 the observed variability for reasonable fractions and degrees of melting of the pyroxenitic-
205 derived components (Supplementary Fig. S3). Among the SD pyroxenites only those
206 having high Mg/Na ratios, as M7-16, match the VLS observed variability. It is worth
207 noting that the mixed melts matching the VLS data are those obtained at high F (F=65%),
208 a value close to those predicted by thermodynamics¹³⁻¹⁵. Based on these calculation the
209 VLS variability can be approximated by linear mixing of a VUM primitive melt with ca.
210 30% M7-16 type pyroxenite-derived melt.

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213 Size and nature of mantle domains

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215 An important result of thermodynamic modelling^{13,14} is that the undercooling inferred
216 from F% decoupling occurs only for a composite source where a lower-T solidus
217 component is finely dispersed in a high-T-solidus host to ensure efficient heat diffusion.
218 Adopting the calibration of¹³ for tabular heterogeneities limits their size to be < 1 km, a
219 dimension close to those estimated from seismic scatter (<10 km)³⁵. Cryptic stripes of
220 comparable size (4 km) have been described at the SEIR³⁶ and modelled based on Nd-Hf
221 MORB variability by³⁷. The compositional trend observed along the VLS suggests a
222 decreasing decoupling of the measured parameters. The lateral extension of the melting
223 region reaches 300 km in the spreading direction, ca. 70 km along axis²⁶. A single large
224 heterogeneity in the older mantle section would have measured tens of km in width,
225 excluding efficient heat diffusion within the heterogeneity itself. We deduce that the older
226 mantle section contained a cluster of small, tabular, low-T solidus components as in Figure
227 2, decreasing in time from ca. 15% to 0% of the volume.

228 VLS basalts $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{208}\text{Pb}/^{206}\text{Pb}$ ratios duplicate one of the mixing trends
229 recognized in South Atlantic basalt suites^{38,39} (Supplementary Fig. S4), suggesting that
230 those endmembers are ubiquitous in the sub-Atlantic mantle.

231 VLS glasses $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary in a restricted range (0.51295-
232 0.51317 and 0.70258-0.70351 respectively, Fig. 1, S5, Online Methods); in contrast,
233 mantle peridotite clinopyroxenes display a large scatter ($^{143}\text{Nd}/^{144}\text{Nd}$: 0.512024-0.513616;
234 $^{87}\text{Sr}/^{86}\text{Sr}$: 0.702220-0.705508, age corrected values). This different compositional scatter
235 can be generated by chromatographic dispersion in the melting column⁴⁰ or through time
236 by reactive melt infiltration and veining^{41,42}.

237 The VLS oldest basalts show the highest apparent degree of melting and the most
238 enriched Sr-Nd-Pb isotopic signature (Supplementary Fig. S4-S5). We suggest that their
239 compositions reflect a larger contribution of pyroxenite-derived melts. As a result, the
240 peridotites of the older domain are less affected by decompression melting, and record
241 lower degrees of melting, possibly preserving their original DMM isotopic fingerprint. In
242 contrast, the VLS younger basalts received a negligible contribution of pyroxenite-derived
243 melts. Unfortunately, we do not have enough data on the mantle peridotites of this young
244 VLS stretch because they are still buried below the sea floor; however the few available
245 samples indicate that this parcel of peridotitic mantle underwent high degree of melting.

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249 Distribution of pyroxenites along Mid Ocean Ridges

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251 We attempt now to extend our findings to other portions of the mid ocean ridge system
252 and interpret the chemistry of genetically related basalt/peridotite in terms of the
253 proportion of pyroxenite dispersed in the mantle source. Only two stretches of the global
254 MOR system have basalt-peridotite pairs sampled densely enough to allow first order
255 observations: the northern Mid Atlantic and the Southwest Indian Ridges^{19,43}. In both cases
256 along axis mantle peridotites record an extent of melting systematically lower than that
257 recorded by the associated basalts (Fig. 4).

258 In light of our findings we propose that pyroxenites are widely distributed in subridge
259 mantle sources proportionally to $\Delta F_{\pi}^{\beta} = F_{\beta}^{app} - F_{\pi}$ the difference between the degree of
260 melting derived from basalts and that derived from peridotites (see Online Methods). This

261 interpretation, if correct, should be confirmed by a correlation between the extent of ΔF_{π}^{β}
262 and chemical indicators of the presence of pyroxenite in the source, e.g. the isotopic ratios
263 of radiogenic elements, expected to be enriched in recycled materials. A broad negative
264 correlation appears between the measured ΔF_{π}^{β} and the basalt $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (Fig. 5,
265 Supplementary Table S4), suggesting the dependence of the Nd isotopic composition on
266 ΔF_{π}^{β} ($r^2=0.53$) and revealing similar enriching mechanisms in the two ridge systems.

267 Spreading rate and mantle potential temperature may not be the leading factors
268 affecting the composition of the extracted basalts and of the residual mantle: the relative
269 proportion of pyroxenites versus peridotites in the mantle source maybe more important. A
270 major implication is that at constant temperature a pulse of low-melting pyroxenite
271 entering the melting region may not lead to a pulse of magmatism because the increased
272 undercooling of the mantle shrinks the peridotitic melting region contrasting the increase
273 of instantaneous pyroxenite melt production.

274 Our results show that low-T melting heterogeneities dispersed in the mantle source
275 affect not only the composition of the extracted basaltic melts, but also the total extent of
276 melting, the volume of extracted melts, and consequently crustal thickness.

277

278

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289

290 Author contributions

291 D.B. performed the modelling. A.C. analysed the samples. D.B. and A.C. processed the
292 geochemical data and wrote jointly the paper. E.B. provided the opportunity and support
293 for sea-expeditions and work. All the authors discussed the results and the interpretations.

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- 396

398 **Figure captions**

399

400 **Figure 1:** Figure 1. a) Temporal variation of the VLS basaltic glasses $^{143}\text{Nd}/^{144}\text{Nd}$
 401 isotopic ratios. Solid circles define three domains (average Nd isotopic ratio of each sector
 402 is indicated). Dashed line is average of oldest sector without one enriched sample.

403 b) Temporal variation of the degree of melting as inferred by mantle peridotite
 404 residues¹⁹ and erupted MORBs⁷. Each point represents a dredge average. Thick bold lines
 405 are linear regressions; thin lines show 95% confidence bands. The age of mantle rocks is
 406 corrected for the time lag between their arrival at the seafloor and the arrival of the basaltic
 407 melts they produced according to ^{1,4} (see Methods).

408

409

410 **Figure 2:** Interpretative sketch of the upwelling mantle column below the Vema
 411 Lithospheric Section. At constant mantle potential temperature the presence of pyroxenites
 412 causes a contraction of the melting region. Total degree of decompressive adiabatic
 413 melting is computed based on Melt-PX algorithm²⁴. In the older VLS sectors, the onset of
 414 melting of the mantle peridotite is delayed and its degree of melting reduced ($F_{\text{max}}=8$). The
 415 associated pyroxenite melts more ($F_{\text{max}}=53$) and contributes to the higher apparent degree
 416 of melting in the pooled MORBs. Undercooling of the mantle peridotite causes deepening
 417 of the end of the melting column estimated to shallow by 0.4 GPa from the older to the
 418 younger sector.

419

420

421 **Figure 3:** Melt-PX²⁴ numerical experiments for adiabatic melting of a two-component
 422 mantle source: lherzolite plus SD pyroxenite (M7-16²⁹). The thick black line represents the
 423 model limit value for melting ceasing at the base of the crust.

424 Left Panel: the degree of melting of the host peridotite is lowered by adding up to 40%
 425 pyroxenite in the source. Variable extents of the melting column are computed assuming P_f
 426 from 1.1 to 0.3 GPa. Red dashed lines correspond to the VLS average melting interval
 427 from old ($F_{\text{max}}=8.0$) to young ($F_{\text{max}}=14.2$) sectors.

428 Panel B: Variation of the aggregated crustal thickness as function of pyroxenite fraction
429 in the source and final depth of melting. Average values for the VLS extremes are plotted
430 as red symbols and 1σ standard deviation

431

432 **Figure 4:** Variation of the degree of melting estimated from mantle peridotite (blue
433 dots) and associated basalts (red dots) along the Southwest Indian Ridge (upper) and the
434 Mid Atlantic Ridge from the Equator to the Azores hotspot region (lower). Degrees of
435 melting are calculated based on ^{19,23,43} on data compiled from PetDB
436 (www.earthchem.org/petdb). Variations of the amount of low-T melting heterogeneities
437 (pyroxenite) in the source result in larger differences in the estimated degree of melting
438 (ΔF_r^β), a proxy for the along axis pyroxenite vol% content of the source.

439

440 **Figure 5:** Difference of degree of melting estimated from genetically related basalts and
441 peridotites (ΔF) versus the Nd isotopic composition of basalt. The VLS sectors are plotted
442 as large red circles; delta F increases from sector 1 to 3 (young to old). Black diamonds
443 represent SWIR, blue circles MAR. Regression lines are calculated for the whole
444 population (red solid line), the SWIR (black solid line) and the MAR (blue solid line).
445 Data and parameters in supplementary Table S4; 95% confidence bands are plotted in
446 Supplementary Fig. S6. Our interpretation is that the ΔF between basalt and peridotite is a
447 proxy of the amount of low-T solidus pyroxenites in the source.

448

449

450 **METHODS**

451

452 **Analytical Methods**

453 Major elements

454 Major elements on mineral phases were collected with the electron probe (Cameca SX100)
455 at the American Museum of Natural History (NY) using 15kV acceleration voltage, 20nA
456 beam, a 10 μ m diameter beam and 30 s counting times. Sodium, potassium and chlorine
457 were run under different conditions to attain a higher precision and monitor their mobility,
458 with a 5nA beam and counting times of 80 s. A subset of samples has been analysed with a
459 Cameca X-Five microprobe at the CAMPARIS micro-analytical center (University of
460 Paris VI), following procedures detailed in ⁴⁴. A number of primary mineral standards
461 were used, as well as the MORB JDF-D2 standard.

462 Isotope ratios

463 For isotopic determinations 50 to 250 mg of basaltic glass and clinopyroxene separate
464 were prepared by grinding, sieving and handpicking under a binocular microscope. Glass
465 chips were leached in 8N HNO₃. Clinopyroxenes were treated with three leachates to
466 eliminate the effects of seawater alteration ^{5,45}. Pb was separated using AG1-X8 anion
467 resin, Sr was separated using Eichrom Sr resin and Nd was separated in a two-column
468 procedure using Eichrom TRU-spec resin to separate the rare-earth elements, followed by
469 α -hydroxy isobutyric acid. Isotopes were measured on a VG Sector 54 multicollector mass
470 spectrometer housed at the Lamont Doherty Earth Observatory of Columbia University. Sr
471 and Nd isotopes were measured in multidynamic mode. The mass fractionation corrections
472 were based on ⁸⁶Sr/⁸⁸Sr = 0.1194 and ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219. Over the period of analytical
473 work, repeated analyses yielded a ⁸⁷Sr/⁸⁶Sr ratio of 0.710271 \pm 0.000015 for the NBS-987
474 Sr standard (2 σ external reproducibility, n>22) and a ¹⁴³Nd/¹⁴⁴Nd ratio of 0.512096 \pm
475 0.000023 for the JNdi-1 (2 σ external reproducibility, n>38). Total blanks for Sr and Nd did
476 not exceed 80pg. Pb data on basaltic glasses were collected in static mode, using the
477 double spike technique with the calibrated 207/204 spike. Replicate analyses of the Pb
478 isotope standard NBS981 gave an average of 16.9317 \pm 0.0022 and 15.4912 \pm 0.0027 and
479 36.7060 \pm 0.0066 for ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb, respectively. These measured
480 Pb isotope ratios were corrected to the values defined by ⁴⁶ of 16.9356, 15.4891, and
481 36.7006, respectively, for NBS 981. Reproducibility for NBS981 is 130, 174, and 181 ppm
482 (2 σ , N=47), for ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb ratios, respectively. Pb blanks
483 measured were below 100pg and thus negligible relative to the amount of sample analysed.

484

485 **Estimate of the ΔF_{π}^{β} and ¹⁴³Nd/¹⁴⁴Nd of the associated basalts**

486 Figure 4 of the main text reports the correlation between ΔF_{π}^{β} , as the differential in degree
487 of melting between basalt and peridotites, and the Nd isotopic composition of basalts of
488 several portions of the Mid Ocean Ridge system where both basalts and peridotites have

489 been sampled. Among all the explored ridge segments, only few localities report basalts
490 and mantle peridotites sampled in the same dredge haul or site. Basaltic rocks are generally
491 more abundant and present a much denser lateral sampling than mantle peridotites. Basalt
492 chemistry and isotopic ratios appear to vary systematically over variable length scale
493 defining domains in which they show little variability or monotonic changes along the
494 ridge axis (e.g. ^{43, 47-48} among others). Domains characterized by reduced variability of the
495 chemical character of the basalts are usually bounded by major transform faults, which
496 also act as thermal barriers^{47,49-51}. Therefore, in Figure 4 (main text), we have chosen to
497 integrate the dataset of the sites where peridotites and basalts were sampled together with
498 sites where only mantle rocks were recovered, but for which is possible to infer the
499 isotopic composition of the associated basalts from the regional variability. As a
500 conservative approach, we only considered those domains in which lateral isotopic and
501 compositional variability is very low or described by simple monotonic trends.
502 The degree of melting of mantle-peridotites can be affected by Cr# fluctuations due to
503 melt/rock interaction that modify spinel⁹ and pyroxene compositions⁵². Therefore we
504 filtered spinel composition and applied a threshold of TiO₂<0.15 wt% as discussed in ^{2,19}.
505 The supplementary Table S4, reports the inferred values showing the measured and
506 calculated value for each basalt-peridotite couple. The “regional regression” data are
507 calculated considering sets of neighboring peridotites and basalts, whereas the “local
508 average” set of data refers to basalts and peridotites sampled from the same site. $\pm\sigma$ on
509 ΔF_{π}^{β} are estimated by error propagation.

510

511 **Age correction**

512 Basalts from the VLS were erupted over a time range of 26 Ma. We applied, hence, an age
513 correction for radiogenic ingrowth since the closure of the system to allow comparisons of
514 their initial isotopic composition, which is controlled only by the source. The correction
515 time should be calculated since the closure of the system, represented by the moment of
516 separation between the source and the melt at depth, assuming there was no significant
517 melt-rock interaction thereafter. This is attested by mantle residual rocks from the VLS
518 being all equilibrated in the spinel field showing no late interaction with melts. The only
519 report of plagioclase-equilibrated mantle rocks concerns the strain-driven formation of
520 plagioclase in fertile lherzolites during mylonitization⁵³. We identify the melt-source
521 separation at the estimated upper limit of melting as inferred by modelling with Melt-PX
522 (see main text).

523 Based on the estimation of melting a composite source and varying the position of the end
524 of the melting column between 1.2 and 0.3 GPa, it appears that the end of the melting
525 column can be constrained based on the correlation with the measured crustal thickness
526 and degree of melting of the residual mantle. For the time stretch relative to the VLS the
527 end of the melting column can be set at 1.1 GPa for the older domain and 0.7 GPa for the
528 younger domain. Based on Africa-South America Euler vectors of ²¹ and the geomagnetic
529 time scale of ²², the spreading rate at the EMAR axis can be estimated through time. As
530 shown in Figure 1 of the main text, the spreading rate decreased steadily in the last 30 Ma.
531 The absolute value decreases from 17.2 to 16.9 mm/y at Chron 6 (half spreading rate) and

532 from 16.9 to 13.6 mm/y at Chron 5. Crustal ages can be calculated accordingly during this
533 time stretch assuming basalts erupted in the axial region.

534 Times of extraction for the mantle residue and for the basalts are, however, very different.
535 Ascent rates for the basaltic liquids are estimated in the range of m/y⁵⁴⁻⁵⁶. Setting the
536 melt/source separation at 1.1 and 0.7 GPa and upwelling rates in the range 1-5 m/y⁵⁶ gives
537 upwelling times ranging 7-36 ka for the older VLS domain and 5-23 ka for the younger
538 VLS domain. Such delays are unimportant in the age correction of long-time decay
539 systems as those here discussed (Sm-Nd; Rb-Sr; U-Pb); however, we considered this
540 contribution for the total age correction when discussing age corrected values in Figs 1B
541 and S1.

542 Upwelling of the mantle lasts a significant amount of time and can sensibly modify the
543 isotopic relationships. The time necessary for a mantle parcel to join the crust from the end
544 of melting depth is 2.1 Ma for the older VLS domain and 1.7 Ma for the younger VLS
545 domain.

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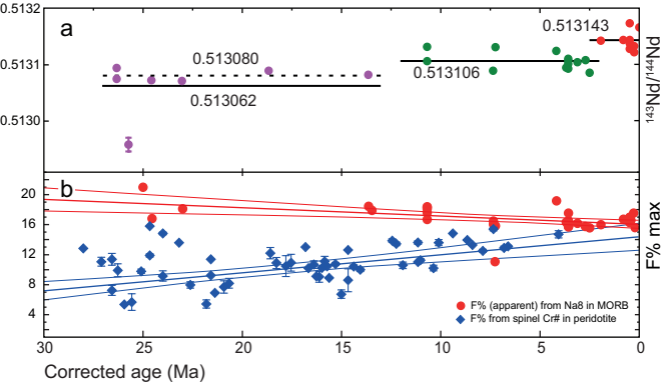


Fig. 1

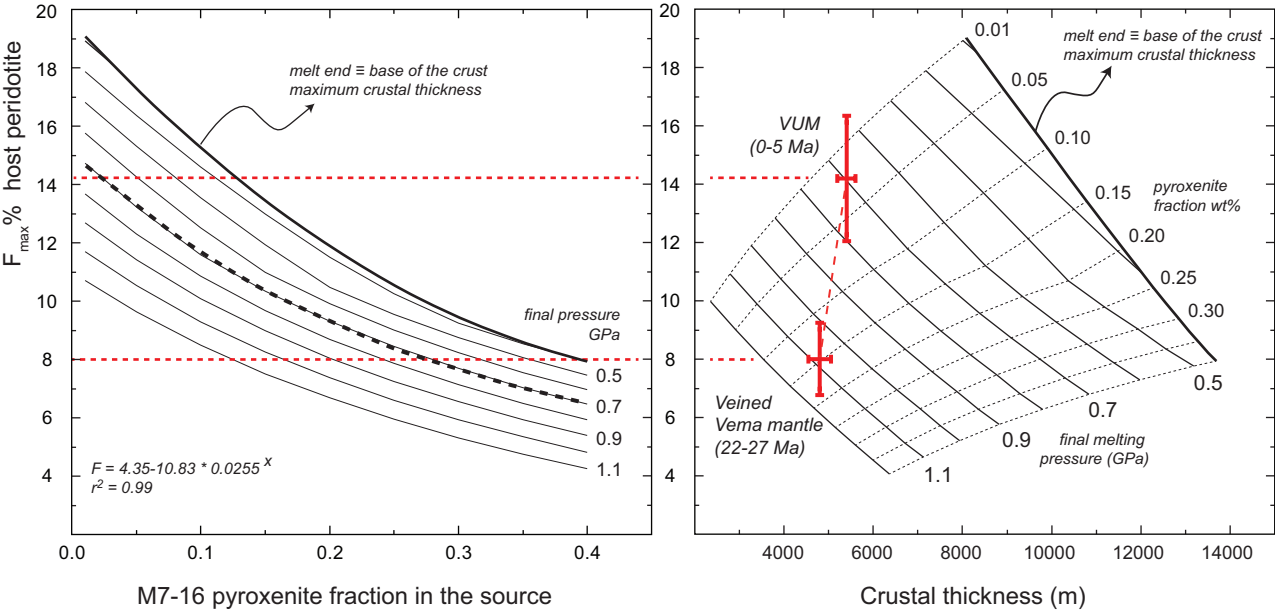


Figure 3_last

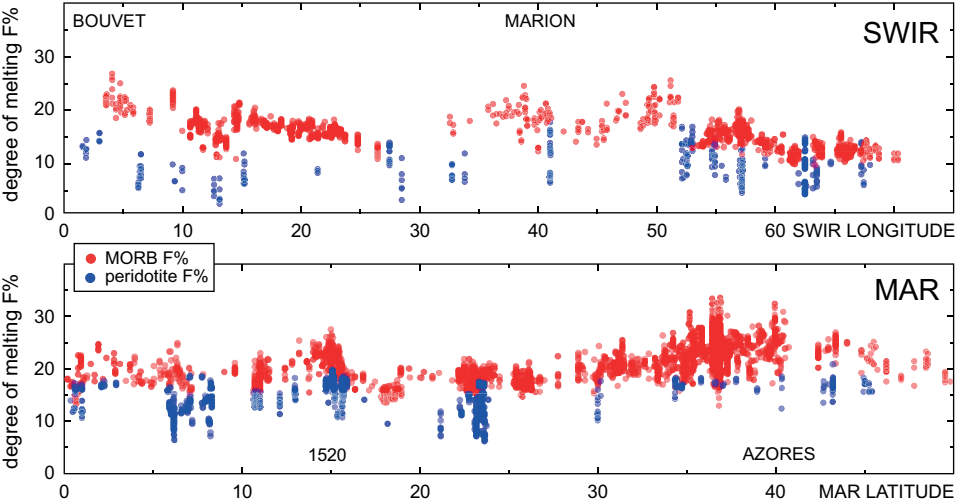


Figure 4-last

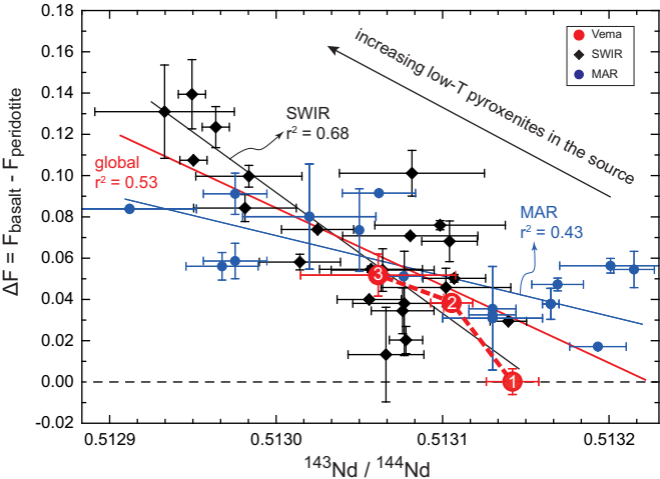


Figure 5 - last