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2 **Is there a global carbonate layer in the oceanic mantle?**  
3

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22  
23  
24 **Key Points:**

- 25
- 26 • CO<sub>2</sub> release during subduction and its transfer to the mantle initiates compression melting  
at the lithosphere-asthenosphere boundary.
  - 27 • Dissolution-precipitation driven melt porous flow results in migration of carbonatite melt  
28 away from the mantle wedge.
  - 29 • Steady supply of carbonated material by subduction may build up a global carbonatite  
30 melt layer at the top of the oceanic asthenosphere.  
31

32 **Abstract**

33 Previous modeling of carbonate subduction by high-pressure experimentation has allowed to  
34 propose scenarios for bulk carbon return to the mantle, but the detailed transfer mechanisms have  
35 seldom been studied. We monitored carbonate – silicate reactions by combining high-pressure  
36 experiments and synchrotron-based x-ray diffraction. Carbonates break down at moderate  
37 pressure and high temperature and CO<sub>2</sub> is trapped at grain boundaries. Further isothermal  
38 compression yields melting, which may control continuous carbon introduction, first in the  
39 mantle wedge, and next, away from the wedge. Carbon presence has been discussed in a variety  
40 of magmatic contexts, under the oceanic lithosphere (hotspots, petit spots, fossil ridges). We  
41 suggest the presence of a global carbon-rich layer under the oceanic lithosphere that is steadily  
42 fed by subduction processes. This layer can be the source of mechanical weakening of the  
43 lithosphere-asthenosphere boundary under the oceans. Therefore, carbon-induced compression  
44 melting may be a key mechanism of modern-style plate tectonics.

45

46 **Plain Language Summary**

47 Plate tectonics, the mechanism by which rigid plates migrate atop the Earth's mantle, necessitate  
48 mechanical decoupling between the plates (lithosphere) and the underlying mantle  
49 (asthenosphere). The presence of a melt layer acting as a lubricating agent has been discussed on  
50 the basis of geophysical evidence, such as provided by seismic and electrical conductivity data  
51 analysis. In this study, we show that this layer can be fueled by the melting of carbonate-rich  
52 material carried down by oceanic plates returning to the mantle at subduction zones. This process  
53 is continuous because the Earth is an active planet with permanent seismic and volcanic activity.  
54 It may have been going on for billions of years and may have been fundamental for the initiation  
55 of plate tectonics.

56

## 57 **1 Introduction**

58 Continuous volcanic activity causes the escape of volatiles from the mantle. This  
59 degassing process is partly counterbalanced by the recycling of volatiles at subduction zones,  
60 including water, carbon dioxide, and noble gases (Poli & Schmidt, 1995; Zhang & Zindler, 1993;  
61 Sleep et al., 2012; Jackson et al., 2013). Volatile compounds are introduced in the mantle with  
62 altered oceanic crust at subduction zones.

63 Previous studies showed that breakdown of water- and hydroxyl-bearing minerals  
64 releases important quantities of water-rich fluids at depths corresponding to fore-arc to back-arc  
65 regions (Schmidt & Poli, 1995). However, the importance of CO<sub>2</sub>-rich fluids is increasingly  
66 recognized and CO<sub>2</sub> may be as important as H<sub>2</sub>O in petrogenetic processes (Molina & Poli, 2000;  
67 Gorman et al., 2006). CO<sub>2</sub>-rich fluid released from the slab may be introduced into the mantle  
68 wedge, via a complex process that could involve crack propagation, as discussed by Brenan &  
69 Watson (1988).

70 At depths corresponding to pressures lower than 2.5 GPa (ca. 75 km), the addition of CO<sub>2</sub>  
71 is not expected to cause mantle melting because the the pressure is too low for carbon to initiate  
72 melting (the carbonate ledge, as defined by Eggler, 1987). In addition, Watson et al. (1990)  
73 showed that H<sub>2</sub>O-CO<sub>2</sub> mixed fluids have very poor wetting properties, even at low CO<sub>2</sub> content.  
74 Therefore, if present, CO<sub>2</sub>-rich fluids do not have a tendency to form an interconnected network  
75 and cannot migrate by porous flow. This absence of mobility may be the cause of the common  
76 occurrence of CO<sub>2</sub> fluid inclusions in mantle minerals (Roedder, 1965, and the recent review by  
77 Frezzotti and Thouret, 2014). By remaining locked at grain boundaries, CO<sub>2</sub>-rich fluids create  
78 CO<sub>2</sub>-impregnated domains confined at the slab/mantle wedge interface, resulting in the  
79 formation of CO<sub>2</sub>-enriched peridotite bodies as discussed by Scambelluri et al. (2016). Because  
80 of the viscous coupling of the plunging slab and the wedge, this modified mantle region may be  
81 entrained to greater depths. The burial of this oceanic plate-borne carbonated material is near  
82 isothermal because in this region, the isotherms are sub-parallel to the slab mantle interface  
83 (Syracuse et al., 2010).

84 Here we explore the outcome of having CO<sub>2</sub>-rich fluids at the slab-mantle interface, by  
85 using high-pressure in-situ experiments combined with synchrotron radiation. Compared to  
86 quench studies, our approach allows for testing a global scenario of carbon delivery and reaction  
87 at mantle conditions by reproducing pressure-temperature path of subduction, coupled with  
88 continuous x-ray monitoring. A major point that is tackled is whether, once formed, CO<sub>2</sub>-rich  
89 fluids escape the system or remain and act as fluxing agents at high pressure. As will be  
90 discussed, the possibility for the dissemination of carbon-bearing material at the lithosphere-  
91 asthenosphere boundary depends on the answer to this question.

92

## 93 **2 Materials and Methods**

94 Experiments were performed using the 2000 ton multianvil press operated on ID06  
95 beamline of the European Synchrotron Radiation Facility, following the approach used in  
96 Hammouda et al. (2014). The details on the experimental and analytical methods are provided in  
97 the supporting-information file. We used a synthetic model system with almost purely magnesian  
98 endmember minerals. Compared to natural systems, this experimental strategy results in shifting  
99 all reaction temperatures by about 250 °C toward higher values, without changing the overall

100 arrangement of the phase diagram (Hammouda & Keshav, 2015). Two experimental runs are  
101 reported. The first one was designed to investigate CO<sub>2</sub>-fluid release, whereas the second one  
102 explored further reactions involving the fluid. The latter pressure-temperature path is illustrated  
103 in Figure 1 and the corresponding x-ray diffraction patterns are shown in Figure 2.

104 A mixture of [enstatite + dolomite] was first compressed and heated, up to 2.25 GPa and  
105 977°C (Figure 1). Enstatite and dolomite diffraction lines are visible on the first pattern of Figure  
106 2. Further heating to 1100°C at about the same pressure results in dolomite diffraction line  
107 disappearing, while diopside lines start growing. If present, forsterite lines are faint. They are  
108 visible only after further heating to 1150°C. This combination of diffraction lines growth  
109 (diopside and forsterite) and collapse (dolomite) attests for crossing of the following  
110 decarbonation reaction

111

112 Enstatite + Dolomite = Diopside + Forsterite + CO<sub>2</sub>.

113

114 The texture resulting from decarbonation, with isolated pores (previously filled with CO<sub>2</sub>  
115 fluid) is illustrated in Figure 3A, which presents the result of the experiment that was interrupted  
116 at this stage.

117 Heating up to over 1300°C results in no major change in the diffraction pattern. Further  
118 compression with the temperature kept around 1330°C yields no change up to 2.71 GPa.  
119 However, while increasing pressure from 2.71 to 3.02 GPa, diopside and forsterite diffraction  
120 lines collapse, and enstatite lines start growing (Figure 2), whereas dolomite diffraction lines  
121 were not observed. This is an indication that the melting reaction has been crossed (Figure 1). In  
122 this experiment, the melt presence is difficult to demonstrate on the sole basis of diffraction  
123 patterns. However, after the experiment was quenched, the resulting assemblage was studied by  
124 scanning electron microscopy. The corresponding texture (Figure 3B) shows evidence for  
125 melting. In this experiment the melting of a silicate + CO<sub>2</sub> assemblage has been directly  
126 monitored and demonstrated to occur solely as an effect of compression at constant temperature.  
127 The crossing of the solidus boundary defines the carbonate ledge (Eggler, 1978; 1987). Melting  
128 is a result of fluid CO<sub>2</sub> remaining trapped in the solid silicate assemblage.

129

### 130 **3 Discussion**

131 In our experimental simulation, the carbonatitic melt is produced at conditions  
132 corresponding to a CO<sub>2</sub>-impregnated mantle wedge crossing the carbonate ledge. Therefore, the  
133 ledge acts as a melting curve at about 80 km depth, in the context of subduction zones. This is a  
134 situation wherein silicate + CO<sub>2</sub> melting occurs because of compression, at nearly isothermal  
135 conditions. Usually, melting in the mantle is associated with decompression in upwelling  
136 plumes, either at mid-ocean ridges or at hot spots, because in most cases, melting is accompanied  
137 by volume expansion. Otherwise, melting can be caused by the introduction of a fluxing agent,

138 such as H<sub>2</sub>O in the case of arc magmatism. Here we illustrate an occurrence of melting  
139 associated with downwelling, because of the peculiarities of the silicate + CO<sub>2</sub> phase diagram.

140 The excellent wetting properties of carbonatitic melts (dihedral angles < 30 °, Hunter &  
141 McKenzie, 1989; Watson et al., 1990; Hammouda & Laporte, 2000) and their very low viscosity  
142 (Kono et al., 2014) at mantle conditions allow for their very efficient percolation into the  
143 surrounding melt-free mantle, provided they are not reduced to graphite (Stagno et al., 2013).  
144 Therefore, at about 80 km depth, a region of carbonated melt-impregnated mantle is formed near  
145 the slab (Figure 4). This modified mantle may be entrained to greater depths because of the  
146 coupling between the plate and the mantle. However, the melt may also spread horizontally,  
147 because carbonatitic melts have a spontaneous tendency to infiltrate the melt-free mantle by  
148 surface-energy minimization. This migration, which is driven by dissolution-precipitation, is  
149 faster than mantle flow and can operate even against gravity (Hammouda & Laporte, 2000).  
150 Decoupling between melt and mantle movement is thus expected, contrary to the CO<sub>2</sub> fluid-  
151 mantle situation, and a carbonatite-rich region extending far from the slab can develop in the  
152 mantle wedge and further away from the slab, at about 80 km depth. This carbonatite-rich  
153 domain is limited toward shallow levels by the ledge. It is limited toward greater depth (about  
154 150 km) by carbon reduction (Stagno et al., 2013; Hammouda & Keshav, 2015).

155 On the basis of thermodynamic modeling, previous studies (Gorman et al., 2006) have  
156 shown that high CO<sub>2</sub> contents should be expected in fluids produced in a hot subduction  
157 environment, which is precisely the case of the Cascadia area. Here, we suggest that highly  
158 mobile carbonatitic melts, originating from the encounter of released CO<sub>2</sub>-rich fluids with the  
159 mantle wedge peridotites are the cause of the observed high conductivity of the Cascadia  
160 subduction zone. At a larger scale, this mechanism could be responsible for delivering carbon-  
161 rich material at the base of the lithosphere, creating a shallow mantle carbon reservoir (Sleep,  
162 2003; Kelemen & Manning). If reheated, this metasomatized mantle region could be the source  
163 of kimberlite-carbonatite magmatism such as observed in North America along a line parallel to  
164 the western paleo-subduction margin of the North American plate (Duke et al., 2014) or along  
165 the North-China craton (Chen et al., 2016). After spreading at the bottom of the lithosphere, the  
166 CO<sub>2</sub>-bearing melts may be involved in the source of petit spots, as recently discussed by  
167 Machida et al. (2017). It may also be anticipated that carbonated material that would escape  
168 remobilization at petit spots could be buried during a later subduction event. There, carbon-  
169 enriched material is located at the back of the subducting slab (Figure 4). It is therefore separated  
170 from the hot mantle wedge by the thickness of the subducting lithosphere. This two-stage  
171 process is a much more efficient way of introducing carbon to great depth because the thermal  
172 regime is cooler than when carbon is carried on top of the plate.

173 The amount and composition of the fluid released by the altered oceanic crust are highly  
174 dependent on the subduction pressure-temperature regime. The recent thermo-mechanical  
175 models (Syracuse et al., 2010) point toward paths hotter than earlier estimates mainly because of  
176 the coupling between the slab and the mantle wedge. With the new models, it is possible that  
177 even cool subduction paths also allow for the release of carbon-bearing fluids. Earlier  
178 experimental studies have shown that the depth of the carbonate ledge depends only slightly on  
179 the CO<sub>2</sub>/H<sub>2</sub>O ratio (Hammouda & Keshav, 2015). Therefore, the presence of a layer of  
180 carbonatite-impregnated mantle wedge may be a general feature of subduction zones, assuming

181 that carbonated material (altered crust or sediments) is carried by the plate, which is a likely  
182 situation.

183

#### 184 **4 Conclusions**

185 The release of CO<sub>2</sub> during the subduction of altered oceanic crust and its transfer to the  
186 mantle wedge initiates compression melting at the lithosphere-asthenosphere boundary. The melt  
187 formed can migrate away from the mantle wedge by porous flow driven by dissolution-  
188 precipitation at grain boundaries. Because subduction zones are continuously active on the scale  
189 of the whole planet, the steady supply of carbonated material may build up a global carbonatite  
190 melt layer at the top of the asthenosphere.

191 Recent petrological and geophysical investigations lend support to the presence of a  
192 carbon-rich melt layer at the oceanic lithosphere/asthenosphere boundary, away from mid-ocean  
193 ridge axes. Petrological evidence comes from the low-degree melts in spreading centers (Zhang  
194 et al., 2017), from petit spot volcanism (Machida et al., 2017) and from metasomatism in the  
195 source of hot spots (Hauri et al., 1993). Further evidence of melt presence at the lithosphere-  
196 asthenosphere boundary arises from seismic data (Schmerr, 2012) and these melt physical  
197 properties strongly suggest a carbonate-rich composition (Sifré et al., 2014). Our results show  
198 that melt trapped at this depth may originate, not only from upwelling regions (Liu et al., 2020)  
199 but also from down-going oceanic plates. This possibility is further substantiated by electrical  
200 conductivity anomalies observed in the Cascadia subduction zone (Soyer & Unsworth, 2006). So  
201 far, serpentinized mantle, caused by water-rich fluid circulation, has been invoked, but serpentine  
202 presence cannot explain high conductivity (Guo et al., 2011).

203 Carbonate-based sediments are present in Archean terranes, and 3.7 Ga old stromatoliths  
204 have been reported from the Isua region in Greenland (Nutman et al., 2016). Recycling of  
205 material of crustal origin in modern type subduction started at least about 3.2 billion years ago  
206 (Næraa et al., 2012). Given that the hot subduction regime was rather the rule for the young  
207 Earth, it can be anticipated that the process of sub-lithospheric impregnation by carbonate melts  
208 has been active very early in the planet history. It may have aided mechanical decoupling at the  
209 lithosphere-asthenosphere boundary, triggering modern-style plate tectonics.

210

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222

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- 324
- 325

326

327 **Figure 1.** Pressure – temperature path followed during the compression melting experiment.  
328 Phase relationships in the synthetic CaO-MgO-SiO<sub>2</sub>+CO<sub>2</sub> system (solid red lines) drawn after  
329 (Eggler, 1978). Univariant reactions are labeled according to the list displayed in the top left-  
330 hand corner. Diamond symbols represent P – T conditions where phases have been identified,  
331 using in-situ X-ray diffraction. Squares with numbers inside represent P – T conditions where  
332 mineral transformations have been observed. Phase-in indicates that new diffraction peaks have  
333 been observed, corresponding to the phase in question (except for melt). Arrows indicates  
334 decrease or increase in diffraction line intensities of the corresponding phases. Melting is  
335 deduced from these variations, as discussed in the main text, and confirmed by imaging of  
336 quenched textures (see Fig.3). Major phase fields are indicated by underlined and italicized  
337 labels. Abbreviations are: dol = dolomite; en = enstatite; di = diopside; fo = forsterite; V = vapor  
338 (CO<sub>2</sub>); L = carbonate melt.  
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341 **Figure 2.** X-ray diffraction patterns recorded during the experiment. Units of x-ray intensity on  
342 the count abscissa are arbitrary (a.u.) and patterns are offset vertically for clarity. Pressure and  
343 temperature conditions are given for each pattern. Diagnostic diffraction lines for minerals are  
344 indicated with arrows. Each pattern corresponds to a diamond symbol of Figure 1. Highlighted  
345 patterns (in red) are evidence for decarbonation and melting.

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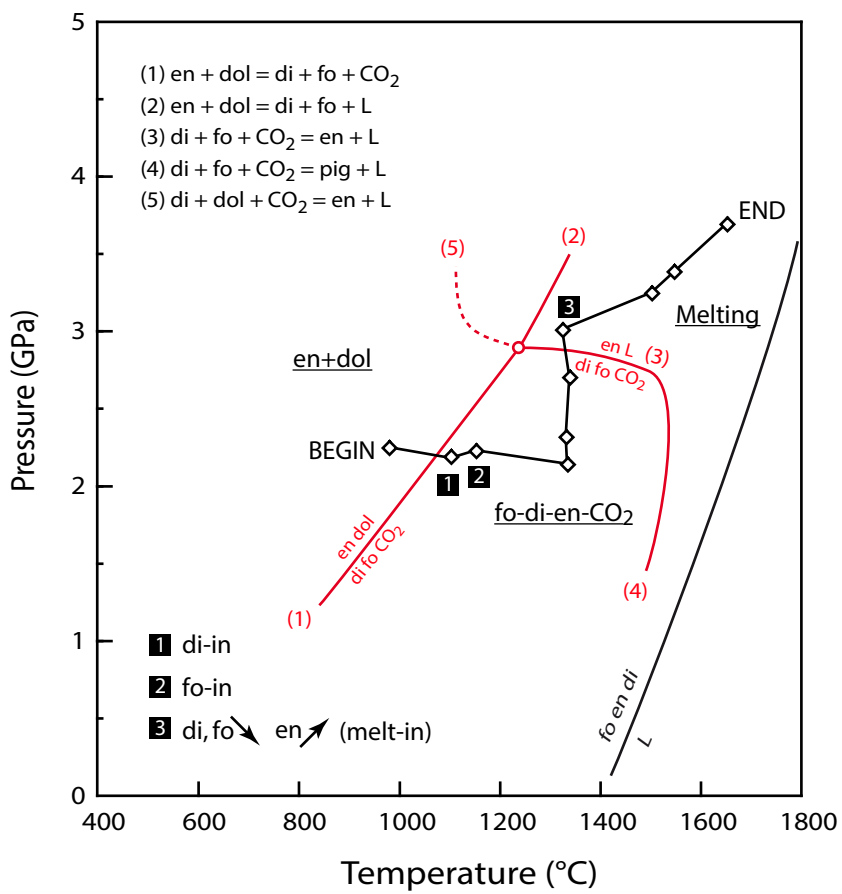
349 **Figure 3.** Scanning electron images of the quenched experiment illustrating the processes  
350 involved in compression melting. (A) Texture resulting from the decarbonation reaction dolomite  
351 + enstatite = diopside + forsterite + vapor. The presence of vapor is attested by the vacuoles  
352 (arrows) surrounded by rims consisting of an intergrowth of diopside and forsterite. (B) Texture  
353 resulting from the melting reaction diopside + forsterite + CO<sub>2</sub> = enstatite + melt, producing a  
354 carbonatitic liquid. The melt is disseminated at grain boundaries, throughout the sample.  
355 Abbreviations are the same as those used in Figure 1, except for Carbo-Melt=quenched melt of  
356 carbonatitic composition.

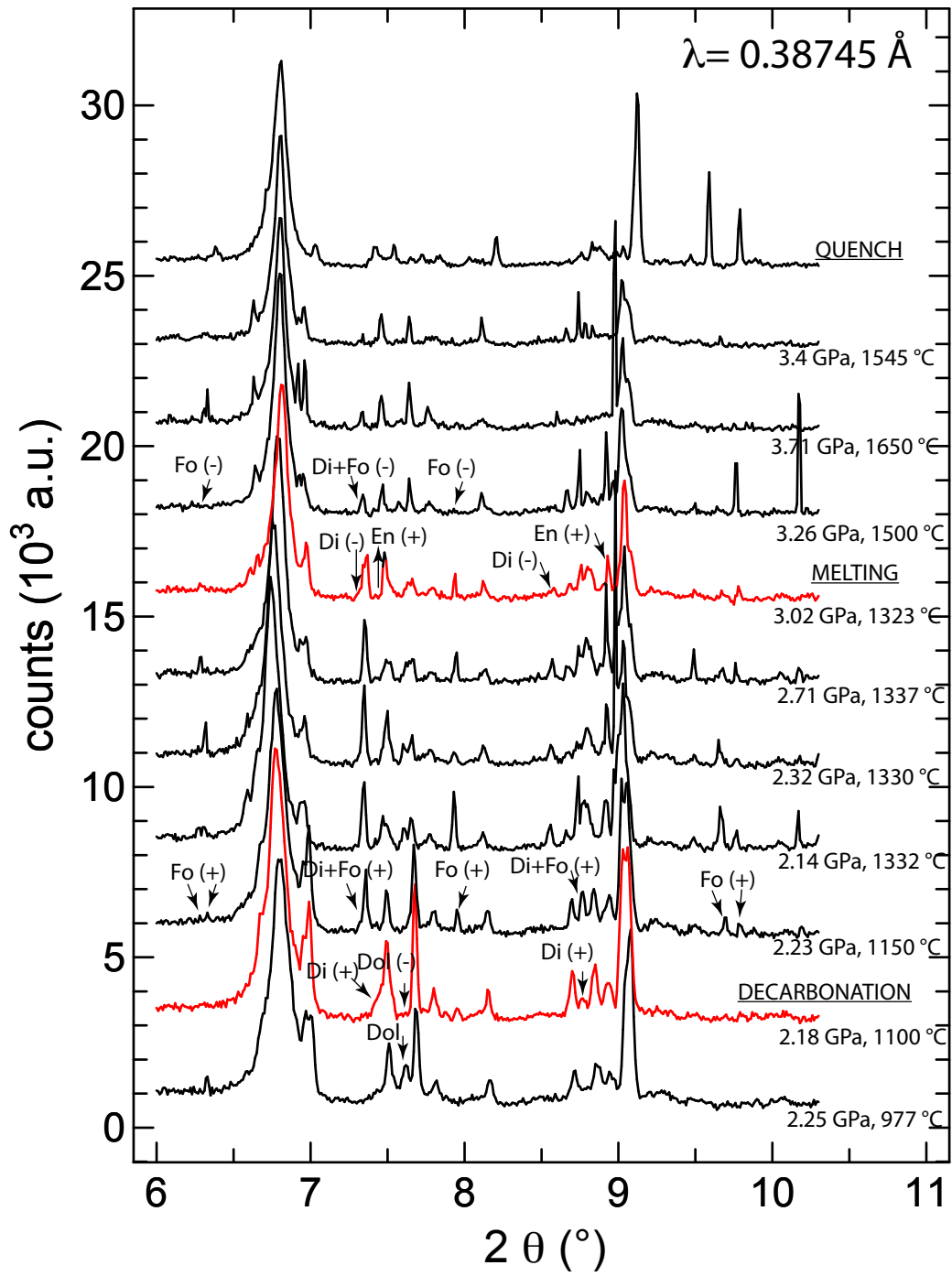
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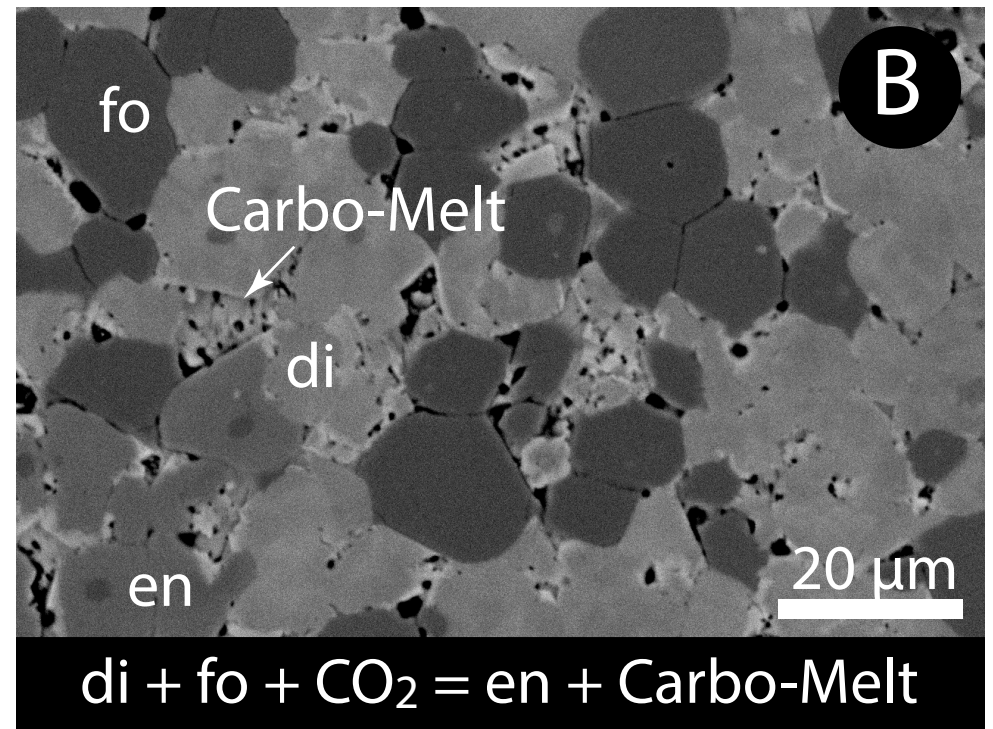
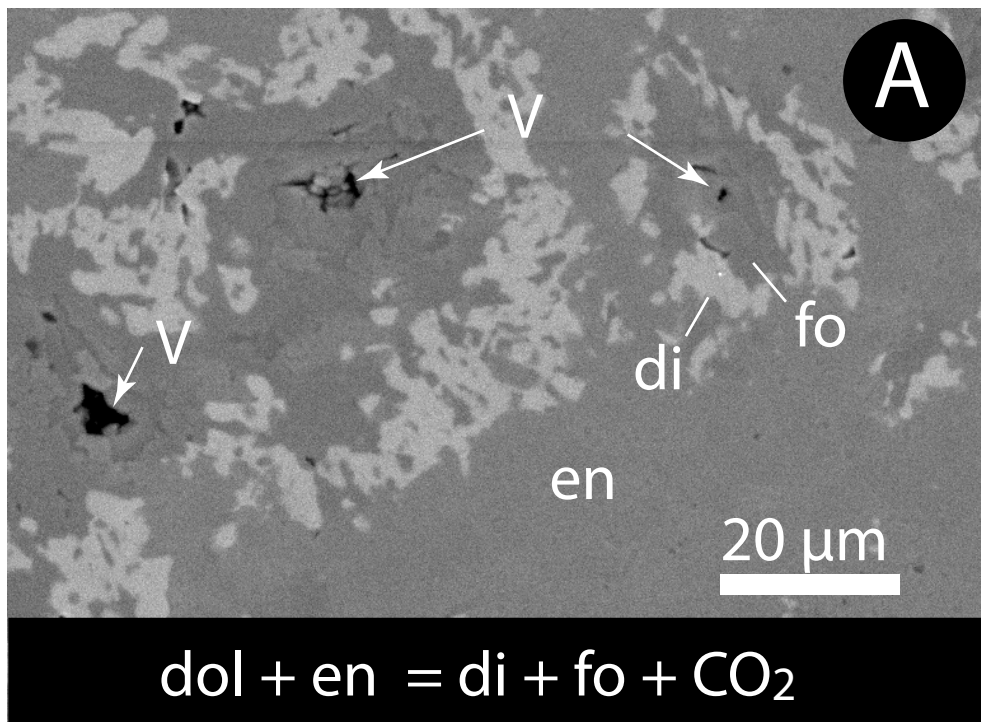
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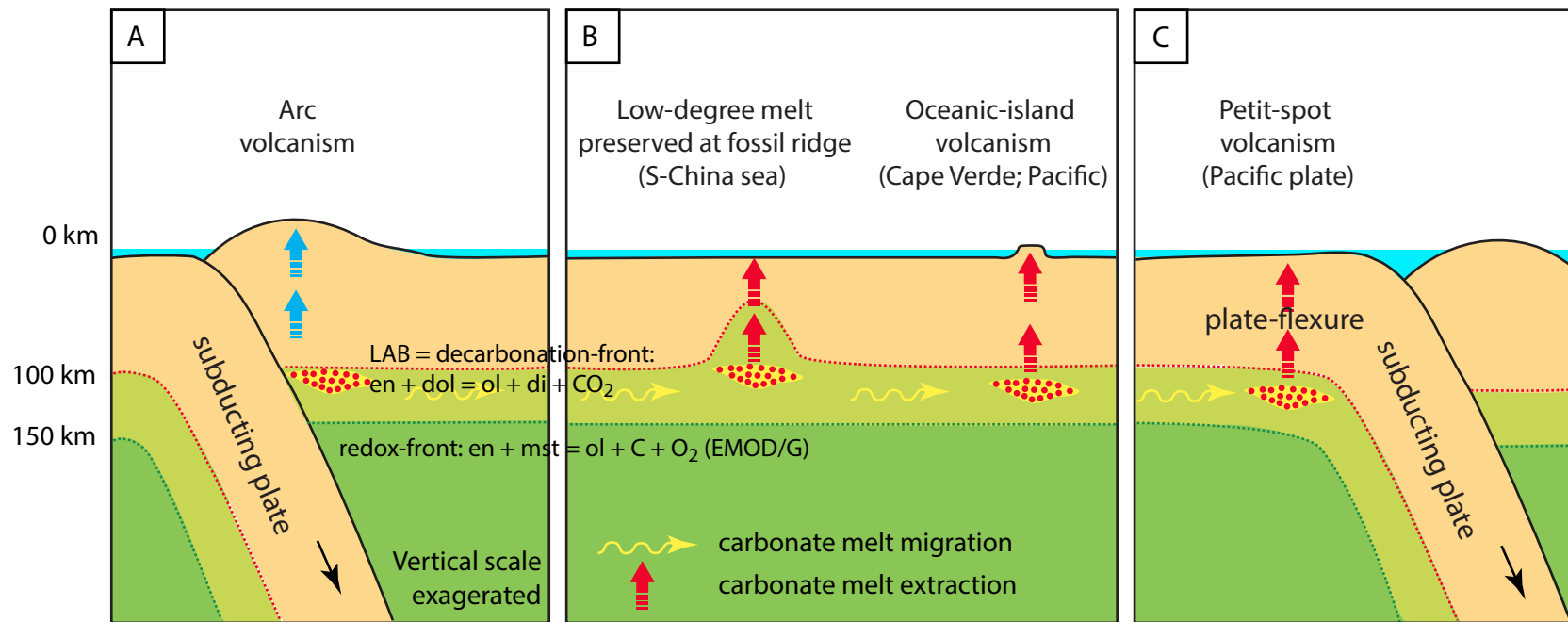
360 **Figure 4.** Summary of the proposed mechanism for the introduction and dissemination of  
361 carbonate melts. (A) Carbonate melts produced by compression melting in front of the  
362 subducting slab occur at depth between 100 and 150 km (yellow patch beneath the arc region).  
363 The region for stable carbonatite melt (in light green) is bounded by the Lithosphere-  
364 Asthenosphere Boundary, where decarbonation occurs (LAB, dotted red line), on top, and the  
365 redox front controlled by the EMOD/G equilibrium, below which the stable form of carbon is  
366 graphite/diamond (dotted green line). Yellow arrows illustrate carbonatite migration away from  
367 the subduction zone. Other yellow patches indicate region of carbonatite implication in  
368 magmatism, such as spreading centers and hot spots (B), and petit spots (C). Vertical red thick  
369 arrows represent oceanic magmatism where carbonate melts participate and may be extracted. In  
370 panel (C), note that carbonate melt present at the back of the subducting plate can be introduced  
371 to greater depth. Vertical blue thick arrows represent arc magmatism, with implication of water-  
372 dominated fluids. Abbreviations are: dol = dolomite; en = enstatite; di = diopside; ol = olivine,  
373 mst = magnesite.











## **Refueling the mantle carbon by compression melting**

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### **Supplementary material: Methods**

#### **High-pressure experiment design**

Experiments were performed using the 2000 ton multianvil press operated on ID06 beam line of the European Synchrotron Radiation Facility, following the approach used in Hammouda et al. (2014). The difference from quench experiments is that several data points were collected during the same experiment. The conditions investigated are shown on Figure 1 (main text). As pressure and temperature were varied, the phases present in the experiment were identified using X-ray diffraction (see below).

In order to introduce CO<sub>2</sub> and have it react with mantle mineral assemblage a mixture of natural dolomite (DNY, Hammouda et al. (2014)) and synthetic enstatite was used. Heating of dolomite + enstatite at pressure below 2.7 GPa produces CO<sub>2</sub> vapor phase, according to the reaction decarbonation reaction:



This reaction has been studied, together with other important reactions in the peridotite + CO<sub>2</sub> synthetic system (Eggler, 1978), as illustrated in Figure 1 (main text). Formation of

diopside and forsterite by the decarbonation reaction yielded phases that are involved in peridotite + CO<sub>2</sub> melting (see Figure 1 of main text).

The starting mixtures were loaded in graphite capsules. The samples were compressed using 18-mm edge length Cr-doped MgO octahedra as pressure transmitting media that were squeezed with 11 mm edge length tungsten carbide anvils (18/11 assembly). Heating was performed using a stepped graphite furnace. The sample temperature was determined using a W-Re (5-26) thermocouple that was separated from the graphite capsule by a thin MgO disk. No correction was applied for the effect of pressure on the e.m.f. Pressure was determined using the cell parameters of the graphite capsule material and the equation of state of Hanfland et al. (1989). Error on temperature and pressure are considered to be about 50 °C and 0.2 GPa, respectively. No water was added to the starting material. The experiment is therefore considered to be nominally anhydrous, although we cannot exclude the presence of small quantities of adsorbed water in the starting material.

### **X-ray data collection**

We used in situ X-ray diffraction for phase identification at high pressure and high temperature. Monochromatic X-ray diffraction patterns ( $\lambda=0.38745$  Å) were collected with a Tl:NaI Bicon scintillator detector, scanning on a large-radius pseudorotation over a  $2\theta$  angular range 6-11°, with step size 0.01°. Two sets of adjustable slits were used to define the volume of the diffracting sample, and remove diffraction signal from the up- and downstream gaskets and furnace. Angular calibration of the instrument was performed before the experiment against NIST standard LaB<sub>6</sub> SRM660a. X-ray diffraction patterns were processed by full profile LeBail refinements with the Jana2006 software package (Petricek et al., 2006).

### **Chemical analysis of the quenched phases**

After quenching, the phase assemblage was characterized using a JEOL JSM-5910LV scanning electron microscope operating at 15 kV. Quantitative analyses of the quenched phases were performed using a CAMECA SX100 electron microprobe. Accelerating voltage was 15 kV and probe current on the Faraday cup was 15 nA. We used focused beam for minerals and defocused beam for quenched melts. In the latter case, beam diameter depended on the size of quenched melt pockets (usually, a few micrometers), and beam damage is difficult to avoid. The following standards were used: wollastonite for Ca and Si, olivine for Mg, magnetite for Fe. Carbon was determined by stoichiometry for the carbonates, and it was estimated by difference to 100% on analysis total for the quenched melt, and by stoichiometry, assuming that all cations (except Si) are bound to carbonate. As can be seen in **Table ST1**, the agreement between of the two methods is satisfactory (within 5% relative), given the difficulty in analyzing small quenched carbonate melt pools.

Electron microprobe analyses of quenched phases are displayed in **Table ST1**. Forsterite has end-member composition, while enstatite contains a slight amount of calcium (about 0.1 atom per formula unit, on the basis of 6 oxygen atoms). Two clinopyroxene types are found. One is close to diopside composition, while the other is of pigeonite type. We note that the diopside-like pyroxene is deficient in Ca, with 1.33 Mg and 0.74 Ca atoms per formula unit. This deficit can be attributed either to Ca partitioning in the melt (melt has  $\text{Ca}/(\text{Ca}+\text{Mg}) > 1$ , see **Table ST1**) or in the vapor phase, as suggested by Martin and Hammouda (2011). Melt composition varies among the different analytical spots. Variations are found for  $\text{SiO}_2$  content and for  $\text{Ca} / (\text{Ca}+\text{Mg})$  atomic ratios ( $\text{Ca}\#$ ). Given the difficulty encountered while analyzing the quenched melt, it is not clear whether this observation reflects true compositional zoning, contamination by adjacent silicate minerals during analysis, or modification during quenching. In the absence of large separated quenched melt pools, it is not possible to

conclude on the effect of quenching, in the manner discussed by Dalton and Presnall (1998). For the purpose of the present investigation, the key point is that melts are carbonatitic.

### **Supplementary references**

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## Supplementary Table

Table 1: Average phase composition obtained by electron microprobe analysis.

phase <sup>a</sup>	opx (2) <sup>b</sup>	cpx ? (1)	pig (2)	fo (2)	melt (8)
SiO <sub>2</sub>	60.0 (0.1) <sup>c</sup>	53.17	58.8 (0)	43.3 (0.6)	1.3 (0.7)
FeO	0.02 (0.01)	0.02	0.01 (0.01)	0.04 (0.01)	0.04 (0.02)
MgO	37.8 (0.6)	24.06	32.0 (0.5)	56.5 (0.2)	10.1 (1.3)
CaO	2.69 (0.27)	18.71	9.69 (0.50)	0.40 (0.04)	45.0 (1.8)
CO <sub>2</sub> by diff <sup>d</sup>					44.0 (0.8)
CO <sub>2</sub> by stoich <sup>e</sup>					46.3 (1.4)
Total	100.6	95.95	100.5	100.2	<sup>f</sup>
atoms per formula unit <sup>g</sup>					
Si	2.01 (0.01)	1.97	2.01 (0)	1.01 (0.01)	
Fe	0.0 (0)	0.00	0.0 (0)	0.0 (0)	
Mg	1.89 (0.03)	1.33	1.63 (0.02)	1.97 (0.02)	
Ca	0.10 (0.01)	0.74	0.35 (0.02)	0.01 (0.00)	
Ca# <sup>h</sup>	0.05 (0.01)	0.36	0.18 (0.01)	0.01 (0.00)	0.76 (3)

### Notes:

<sup>a</sup>phase abbreviations are: opx=orthopyroxene, cpx?=clinopyroxene not confirmed, pig=pigeonite (Ca-poor clinopyroxene), fo=forsterite, melt=quenched carbonatitic melt.

<sup>b</sup>number in brackets indicated the number of analyses performed for each phase.

<sup>c</sup>number in bracket next to composition indicates 1 standard deviation.

<sup>d</sup>CO<sub>2</sub> calculated by difference to 100%, using the sum of the other oxides.

<sup>e</sup>CO<sub>2</sub> calculated by stoichiometry, assuming that all cations (except Si) are bound to carbonate.

<sup>f</sup>no total is given because CO<sub>2</sub> is not directly analyzed.

<sup>g</sup>atoms per formula units are calculated assuming the following number of oxygen: pyroxene=6, forsterite=4.

<sup>h</sup>Ca# = Ca / (Ca+Fe+Mg) in atoms.