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1 **Unravelling partial melt distribution in the oceanic low velocity zone**

2

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23 **Abstract**

24 The widespread low seismic velocity zone (LVZ) in the shallow oceanic mantle has long been debated
25 in terms of mantle melting. At LVZ depths, volatiles (CO₂ and H₂O) are present in minute amounts,
26 which implies mantle incipient melting down to below 1000°C with the production of minute amounts
27 of volatile-rich melt, well below 1 vol.%. However, melt compositions and distributions in the incipient
28 melting regime have only been inferred from experiments departing from actual mantle conditions.
29 Here, we experimentally reproduce incipient melting by re-equilibrating a naturally CO₂- and H₂O-
30 bearing mantle rock at mantle temperatures and pressure. By using cutting-edge microscopy
31 characterizations, we evidence that minute amounts of volatile-rich melts fully interconnect in mantle
32 rocks down to lithospheric temperatures, enabling thus the modification of geophysical signals from the
33 mantle. These findings and the correspondence of the domain of local, sharp drops in shear wave
34 velocity (Vs) with the domain of (CO₂+H₂O)-melting in the LVZ strongly supports that these
35 geophysical anomalies relate to mantle melting. Geophysical surveys image *in situ* the very low and
36 highly heterogeneous distribution of melt in the mantle generated by the very low and highly
37 heterogeneous distribution of volatiles probed by surficial geochemical surveys. The global-scale
38 geophysical signature of the LVZ appears mainly unaffected because the average background melt
39 fraction is very low, estimated at ~0.03-0.05 vol.% melt. However, enhanced geophysical signals arise
40 from sporadic, localized areas where melt fraction is increased, such as the ~0.2 vol.% melt estimated
41 for detecting sharp Vs drops using SS precursors. In-depth deciphering of the dynamics of melt and
42 volatiles in the LVZ calls for investigations on the seismic velocity, permeability and rheology of
43 partially molten mantle rocks covering the diversity of mantle melt compositions, fractions and
44 temperatures.

45

46 **Keywords**

47 Mantle incipient melting, Volatile-rich melt interconnection in mantle rocks, Low Velocity Zone,
48 Distribution and dynamics of volatiles and melt in the upper mantle

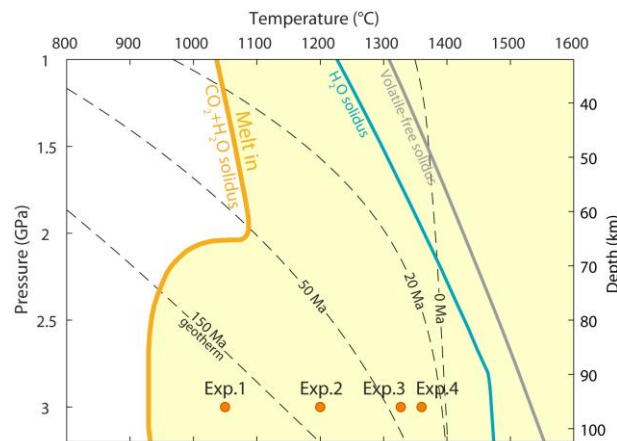
49 1. Introduction

50 Both CO₂ and H₂O are present in the upper mantle, though in minute amounts and heterogeneously
51 distributed (Le Voyer et al., 2017). Melting is thus expected to occur in a vast region extending from
52 young and hot to old and cold oceanic mantle (Fig. 1), and to generate volatile-rich melts with low and
53 heterogeneous volume fractions correlated to the low and heterogeneous volatile contents (Wallace and
54 Green, 1988; Hirano et al., 2006; Hirschmann, 2010; Dasgupta et al., 2013; Massuyeau et al., 2015;
55 Machida et al., 2017; Gaillard et al., 2019). To date, melting experiments on mantle rocks with very
56 small amounts of both CO₂ and H₂O (< 1000 wt. ppm) producing very small fractions (< 1 vol.%) of
57 (CO₂+H₂O)-rich incipient melts down to lithospheric temperatures have never been investigated. This
58 raises important issues. First, melt compositions and fractions produced in the incipient melting regime
59 may differ from that inferred from conventional experiments where CO₂ and H₂O are added in amounts
60 greatly exceeding those present in the mantle. Second, the intergranular distribution of small fractions
61 of incipient melts in mantle rocks is unknown. Those melts have to form interconnected networks to
62 enable significant modifications of mantle properties in the LVZ. Among those are seismic velocity
63 lowering (Faul et al., 2004; Takei and Holtzman, 2009a, 2009b; Rychert and Shearer, 2011; Schmerr,
64 2012; Chantel et al., 2016; Takei, 2017; Tharimena et al., 2017), electrical conductivity increase (Naif
65 et al., 2013; Sifré et al., 2014; Laumonier et al., 2017), and steady-state viscosity lowering (Hirth and
66 Kohlstedt 1995a, 1995b, 2003; Takei and Holtzman, 2009a, 2009b; Holtzman, 2016). Our knowledge
67 about melt interconnection in the upper mantle remained far too partial as it involves experiments at too
68 high temperatures, too large melt fractions, and in simplified petrological systems (Minarik and Watson,
69 1995; Faul et al., 2004; ten Grotenhuis et al., 2005; Yoshino et al., 2009, 2010; Zhu et al., 2011; Miller
70 et al., 2014; Mu and Faul, 2016, Laumonier et al., 2017). Unravelling melt interconnection at actual
71 mantle conditions require investigations down to lithospheric temperatures (~1000°C) and on minute
72 fractions (< 1 vol.%) of (CO₂+H₂O)-rich melts.

73 Here we report extensive investigations on experimental reproductions of actual mantle melting. Our
74 novel approach consists in re-equilibrating a volatile-bearing mantle xenolith at mantle-relevant
75 conditions (Fig. 1). This starting material naturally contains minute amounts of volatiles, i.e. 500 ± 50

76 wt.ppm CO₂ and 630 ± 210 wt.ppm H₂O. It was used to perform high pressure and temperature
77 experiments in the (CO₂+H₂O)-assisted melting domain, at conditions corresponding to ~100 km depth
78 beneath young and hot to old and cold oceanic lithospheres, where both volatile-free-melting and H₂O-
79 only-assisted melting are not possible (Fig. 1). We directly evidence that minute amounts of both CO₂
80 and H₂O do generate minute amounts of interconnected, (CO₂+H₂O)-rich melts in mantle rocks down
81 to lithospheric temperatures. The distribution of melt and volatile and its relationship with geophysical
82 signals in the LVZ are finally discussed.

83



84

85 **Fig. 1.** Incipient melting at shallow mantle temperatures and pressure. Present experiments investigate the domain of
86 (CO₂+H₂O)-assisted melting (Wallace and Green, 1988) which covers young and hot to old and cold oceanic mantle, down to
87 temperatures where both volatile-free-melting and H₂O-only-melting (200 wt.ppm) (Sarafian et al., 2017) are precluded. Mantle
88 thermal structure is G13R1350 model with 1350°C potential temperature from Grose and Afonso (2013).

89

90 2. Materials and methods

91 2.1 Starting material

92 The mantle xenolith is a spinel-peridotite sampled at Lanzarote (Canary Islands). A large piece of the
93 xenolith was crushed in agate mortar and the powder remained stored in desiccator prior to analysis and
94 experiments. Bulk rock composition was measured twice using inductively coupled plasma atomic
95 emission spectroscopy (ICP-AES) at LMV. CO₂ and H₂O contents were measured three times on

96 samples of 1 mg each using organic elemental analyser at LMV. Averages of these measurements, with
97 uncertainties given as twice the standard deviation of the mean (95% confidence interval), are reported
98 in [Table 1](#). Note that the low values of 630 ± 210 wt.ppm H₂O and 500 ± 50 wt.ppm CO₂ evidence the
99 xenolith did not experience significant weathering at the surface.

100 *2.2 High pressure and temperature experiments*

101 The xenolith powder was loaded in 3-mm diameter Au₈₀Pd₂₀ capsules and welded shut with no
102 additional dopant. High pressure and temperature experiments were performed at LMV using piston-
103 cylinder apparatus with conventional 1/2-inch assembly made of NaCl and pyrex glass as pressure
104 medium and thermal insulator, respectively. The temperature was generated by a graphite furnace and
105 monitored with a C-type thermocouple placed at about 0.5 mm above the capsule. Experiments started
106 with the pressurization of the assembly to 1 GPa followed by heating to 650°C at a rate of 60°C/min.
107 The assembly was then pressurized to 3 GPa, before heating to target temperatures: 1050, 1200, 1327
108 and 1360°C for Exp. 1, 2, 3 and 4, respectively ([Table 2](#)). Uncertainties on temperature and pressure are
109 within 20°C and 0.3 GPa, respectively. At the end of the experiment, power was switched off, resulting
110 in a cooling rate of ~100°C/s, and then slowly decompressed to room pressure. Recovered capsules were
111 sawed and one half was embedded in epoxy resin for mechanical polishing down to 1/4 μm using
112 diamond abrasives. Sawing and polishing were performed using ethanol as lubricant; no water was used
113 to avoid carbonate dissolution ([Wallace and Green, 1988](#)). The samples were prepared for electron
114 microscopy by coating the surfaces with a few tens nm layer of carbon using a carbon coater.

115 *2.3 Electron microscopy characterizations*

116 The characterization of the experiments was performed using the scanning electron microscope coupled
117 with focused Ga-ion beam (SEM-FIB) of CIMAP, equipped with an energy-dispersive X-ray
118 spectroscopy detector (EDS) and a scanning transmission electron microscopy detector.

119 Melt fraction, grain size and grain boundary wetness, i.e. the fraction of grain surface wetted by melt,
120 were determined on images acquired at the centre of the capsules. Acquisitions were performed at 5 kV
121 and 0.05-3.2 nA in backscattered electron mode using high definition, 3072×2048- or 6144×4096-pixel

122 images with pixel sizes of 8 to 34 nm (Fig. 2). Combining these resolutions together with large enough
123 fields of view mostly required image mappings of the regions of interest. Because of a poorer quality of
124 mechanical polishing, the images of Exp. 1 were acquired on two large ion-beam polished sections
125 extracted perpendicular to sample surface (Fig. 2). Mineralogical assemblage was identified from EDS
126 maps at 10-15 kV and 3.2-13 nA, using Aztec EDS acquisition and data processing software by Oxford
127 Instruments (Fig. 2, Table 2).

128 High magnification images of melt areas were acquired in transmission mode at 30 kV and 0.1 nA on
129 100-200 nm-thick, electron-transparent ion-thinned sections extracted perpendicular to the surface at
130 the centre of the capsules (Fig. 3). The chemical composition of the melt in the samples was based on
131 the EDS mapping of the electron-transparent thin sections in transmission mode at 30 kV and 0.8-3.2
132 nA (Figs. 3-4; Figs. S1-S2). This was necessary since using conventional SEM-EDS mapping of bulk
133 sample surface, (i) carbon signals from the carbon layer deposited onto sample surface would have
134 interfered with those from melt, and (ii) signals from surrounding grains would have interfered with
135 those from melt because of too large X-ray generation volumes. As there are only a few melt occurrences
136 per thin section, 28 thin sections were extracted over the four samples in order to improve statistics.

137 These high magnification images were also used to measure the apparent contact angles at the junctions
138 of melt with the various mineral phases of the samples (Fig. 5). The contact angles reported in Table 2
139 are the medians of these measurements (Laporte and Watson, 1995), and associated uncertainties
140 correspond to twice the standard deviation of the mean (95% confidence intervals).

141 Olivine iron number at the vicinity of capsule walls was investigated from EDS maps at 15 kV and 6.4-
142 13 nA. The maps were decomposed into series of 30 μm -wide rectangles with long side (170-260 μm)
143 parallel to capsule wall. The profiles reported in Fig. 5 correspond to the olivine iron number in each
144 rectangle, after calibrating the EDS signals on olivine standard.

145 The 3D distribution of melt was investigated by excavating small volumes in the centre of capsules
146 using serial FIB sectioning-electron imaging (see [Cocoo et al., 2013](#) for further details on the technique).
147 We used 30 kV and 0.43-2.5 nA Ga-ions for producing series of cross sections every 27 to 67 nm, and

148 2 kV and 0.05 nA secondary electrons for imaging each cross section with 3072×2048-pixel images and
149 pixel sizes of 4.5 to 14 nm. Each volume involved 98 to 172 images (Fig. 5, Videos S1-S4).

150 Segmentation of the images was performed using the hand tracing tools of Fiji software (Schindelin et
151 al., 2012). This was especially required for precise contouring of melt areas whose distribution is mostly
152 sub-micrometric. Melt fraction, grain size and grain boundary wetness were calculated as in Mu and
153 Faul (2016). Corresponding uncertainties are given as twice the standard deviation of the mean (95%
154 confidence intervals) (Table 2). 3D images of melt were generated from the stacks of binary images
155 produced by serial FIB sectioning-electron imaging using the 3D viewer plugin of Fiji (Fig. 5, Videos
156 S1-S4). Melt interconnection is defined as the proportion of melt residing in the largest interconnected
157 network of melt.

158

159 **Table 1.** Chemical composition of starting material. Concentrations are given as wt.%. Uncertainties correspond to 95%
160 confidence intervals.

SiO ₂	42.8 ± 0.2
TiO ₂	0.013 ± 0.017
Al ₂ O ₃	0.80 ± 0.37
Cr ₂ O ₃	0.33 ± 0.24
FeO _t	7.58 ± 0.04
NiO	0.36 ± 0.16
MnO	0.120 ± 0.003
MgO	47.2 ± 0.9
CaO	0.71 ± 0.17
K ₂ O	0.050 ± 0.097
Na ₂ O	0.040 ± 0.010
CO ₂	0.050 ± 0.005
H ₂ O	0.063 ± 0.021

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165 **Table 2.** Summary of melting experiments. Uncertainties correspond to 95% confidence intervals. Ol = olivine, Opx =
 166 orthopyroxene, Cpx = clinopyroxene, Spl = spinel, Grt = garnet, Phl = phlogopite.

Sample	Exp. 1	Exp. 2	Exp. 3	Exp. 4
T (°C)	1050	1200	1327	1360
P (GPa)	3	3	3	3
Run duration (h)	192	192	96	63.5
Assemblage	Ol+Opx+Cpx+Grt+Phl+Melt	Ol+Opx+Cpx+Grt+Melt	Ol+Opx+Cpx+Grt+Melt	Ol+Opx+Cpx+Melt
Melt fraction (vol.%)	0.11 ± 0.05	0.16 ± 0.02	0.15 ± 0.02	1.5 ± 0.5
Melt interconnection (%)	~100	~100	~100	~100
Contact angle (°)	32.5 ± 6.0	30.7 ± 4.0	36.3 ± 4.0	26.3 ± 4.5
Gain boundary wetness (%)	4.4 ± 1.4	5.4 ± 1.0	9.6 ± 2.1	20.9 ± 5.0
Grain size (µm)	4.9 ± 0.9	4.6 ± 0.9	6.4 ± 1.0	10.3 ± 1.8
Melt chemistry (wt.%)				
SiO ₂	3.3 ± 2.4	9.6 ± 4.4	9.8 ± 4.8	45.2 ± 9.5
Al ₂ O ₃	0.69 ± 0.49	1.4 ± 0.7	1.5 ± 1.0	11.0 ± 4.0
FeO	4.5 ± 1.9	3.1 ± 1.2	2.4 ± 0.9	6.0 ± 2.3
MgO	29.2 ± 9.3	19.1 ± 6.3	18.2 ± 6.0	12.9 ± 4.6
CaO	11.8 ± 6.8	20.4 ± 6.6	17.6 ± 5.9	9.9 ± 3.6
K ₂ O	1.7 ± 3.0	1.9 ± 2.1	2.6 ± 2.8	2.0 ± 1.1
Na ₂ O	0.60 ± 1.12	0.58 ± 0.82	1.0 ± 1.1	3.3 ± 1.3
CO ₂	36.9 ± 9.8	34.5 ± 8.6	34.6 ± 8.4	7.0 ± 3.6
H ₂ O	11.4 ± 8.0	9.4 ± 3.2	12.4 ± 3.8	2.6 ± 0.9

167

168 3. Results

169 3.1 Phase assemblage, melt chemistry and melt volume fraction

170 All samples contain small amounts of volatile-rich melt after experiments. Assemblages are olivine +
 171 orthopyroxene + clinopyroxene + garnet (up to Exp. 3 at 1327°C) + phlogopite (up to Exp. 1 at 1050°C)
 172 + melt (Fig. 2, Table 2). Melt does not quench as a glass at the end of the experiments (e.g. Wallace and
 173 Green, 1988), but results in quenched nano-crystals with 5-30 vol.% of pores most likely filled with a
 174 K₂O-Na₂O-CO₂-H₂O-bearing fluid phase that was removed during sample preparation (Fig. 3, Figs. S1-
 175 S2). Melt crystallization sequence during cooling of the experiments is (i) garnet between 1360 and
 176 1327°C, (ii) phlogopite between 1200 and 1050°C, and (iii) carbonate (+fluid) below 1050°C. This is
 177 consistent with the evolution of phase assemblage; starting from minute amount of carbonatitic melt at

178 1050°C, additional melting of phlogopite occurs between 1050 and 1200°C, and melting of garnet
179 occurs between 1327 and 1360°C.

180 The melt compositions reported in Fig. 4 and Table 2 combine the composition measured on quenched
181 nano-crystals and that inferred for the fluid in pores (see calculation details in Supplementary Material).
182 The rather large uncertainties partly result from the fact that melt distribution is mostly sub-micrometric,
183 requiring unconventional characterizations in transmission electron mode on electron-transparent cross
184 sections (Fig. 3, Fig. S1-S2). From 1050°C to 1327°C, 0.11-0.15 vol.% of carbonated melt is produced,
185 with CO₂, H₂O and SiO₂ concentrations of about 35, 10 and 3-10 wt.%, respectively. A transition occurs
186 before 1360°C where melt fraction shifts to 1.5 vol.% and melt composition becomes mainly silicated,
187 with CO₂, H₂O and SiO₂ concentrations of about 7, 3 and 45 wt.%, respectively. These melt
188 compositions, and their evolution with temperature, are in line with previous volatile-doped experiments
189 with much higher melt fractions (Fig. 4).

190 3.2 Melt distribution and interconnection

191 Melt phase is virtually fully interconnected in all the samples (Fig. 5, Videos S1-S4, Table 2). (i) High-
192 resolution 3D reconstructions evidence melt mainly forms tubule networks along grain edges, and tends
193 to spread along grain boundaries in the sample with the highest melt fraction (Fig. 5). Grain boundary
194 wetness range from about 5-10% in samples with 0.1-0.15 vol.% melt (Exp. 1, 2 and 3) to about 20% in
195 the sample with 1.5 vol.% melt (Exp. 4) (Table 2). (ii) Measurements of the apparent contact angles at
196 the junctions of melt with the various mineral phases yield median values ranging around 30° in all
197 samples (Fig. 5, Table 2), theoretically implying full melt interconnection (von Bargen and Waff, 1986).
198 (iii) Long range interconnection of melt is confirmed by a minor loss of iron over a few hundreds of
199 micrometres from olivine crystals to capsule walls, though made of Au-Pd (iron depletion reaches 5-
200 10% at the interface with capsule walls) (Fig. 5). This loss occurred via diffusion in the melt network
201 since iron diffusion in melt-free aggregate would have been limited to much shorter distances (Fig. 5,
202 Supplementary Material). It should be recalled that melt analyses, as well as the other measurements,
203 were performed in the centre of the 3 mm diameter capsules and thus were virtually unaffected by such

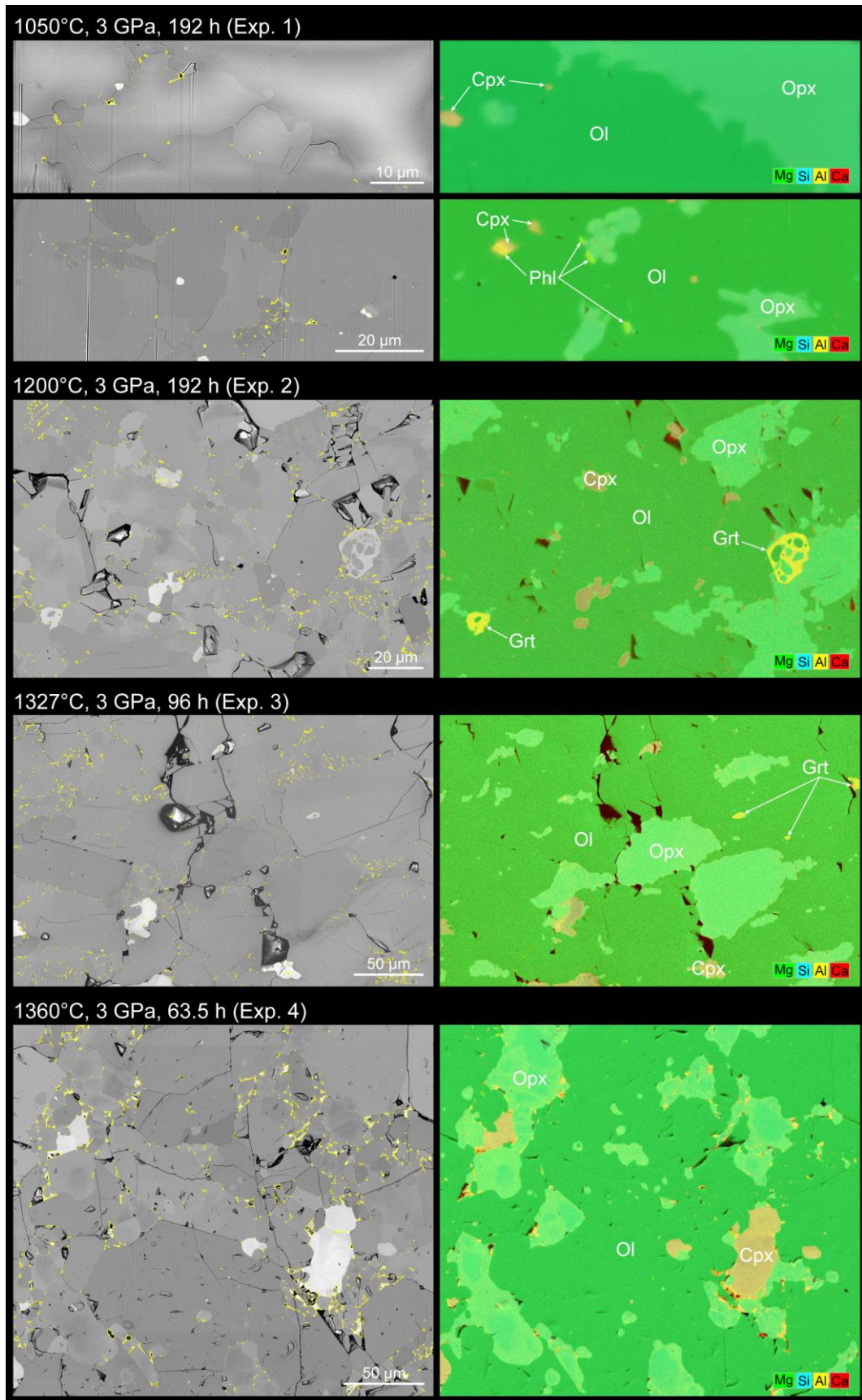
204 a minor iron loss restricted to the vicinity of capsule walls. Indeed, no significant differences can be
205 found between the iron melt contents in our samples and those from literature (Fig. 4).

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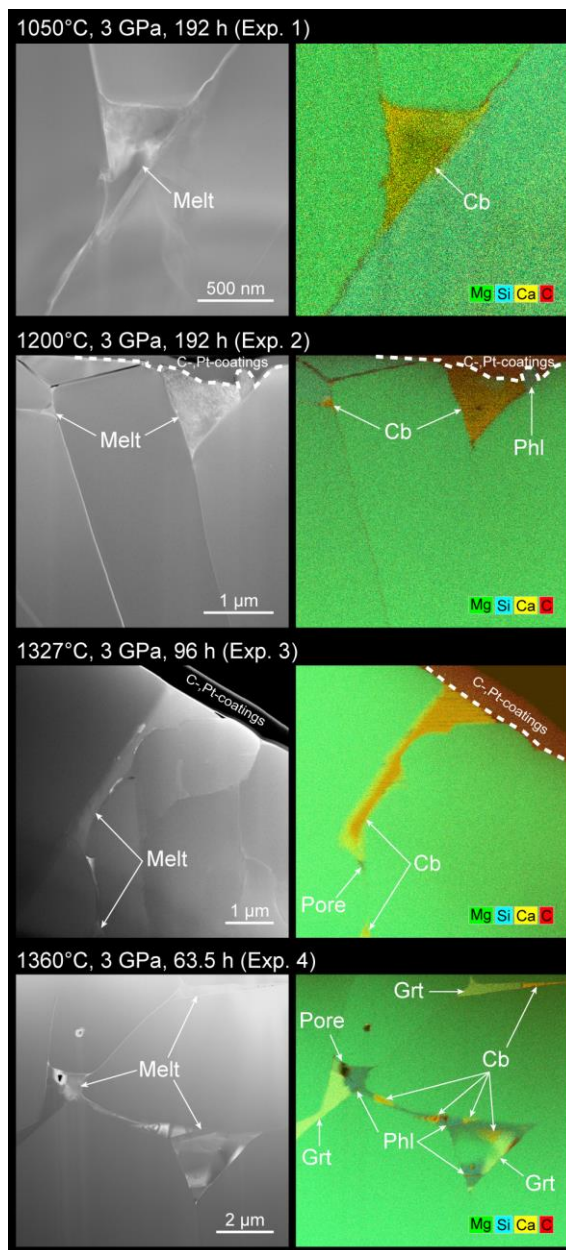
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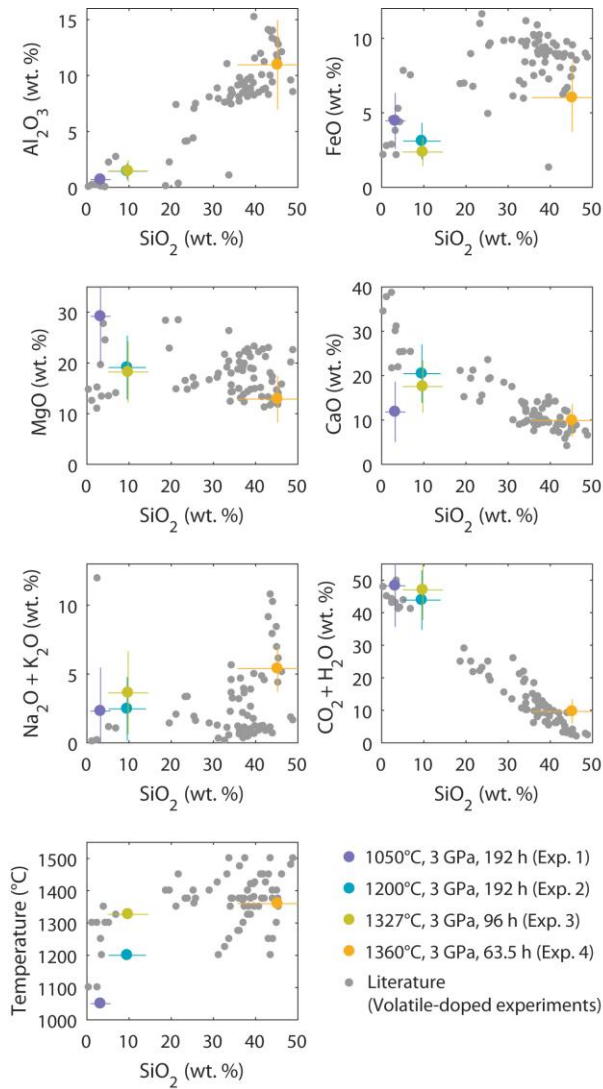
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211 **Fig. 2.** Broad image and chemical mappings of a (CO₂+H₂O)-bearing mantle rock experimentally re-equilibrated at shallow
 212 mantle temperatures and pressure. Left panels, backscattered electron image mappings on which melt contours are reported in

213 yellow. Right panels, corresponding chemical mappings reported as superposition of magnesium, silicon, aluminium and
 214 calcium EDS maps. Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene, Grt = garnet, Phl = phlogopite.
 215



216
 217 **Fig. 3.** High magnification imaging and chemical mappings of incipient melts in a (CO₂+H₂O)-bearing mantle rock
 218 experimentally re-equilibrated at shallow mantle temperatures and pressure. Left panels, transmission electron imaging of
 219 electron-transparent thin sections extracted perpendicular to sample surfaces. Right panels, corresponding chemical mappings
 220 reported as superposition of magnesium, silicon, calcium and carbon EDS maps. Note that the melt does not quench as glass
 221 at the end of the experiments but results in quenched nano-crystals and fluid-filled pores (Cb = carbonate, Phl = phlogopite,
 222 Grt = garnet).



223

224 **Fig. 4.** Composition of incipient melts in a (CO₂+H₂O)-bearing mantle rock experimentally re-equilibrated at shallow mantle
 225 temperatures and pressure. Literature data regroup experiments producing volatile-bearing melts (CO₂ and/or H₂O) at
 226 comparable pressures (2.5-3.5 GPa) and temperatures (up to 1500°C), in at least CaO-MgO-Al₂O₃-SiO₂-FeO systems and
 227 where melts equilibrate with at least olivine and orthopyroxene. See [Table S1](#) for literature data selection.

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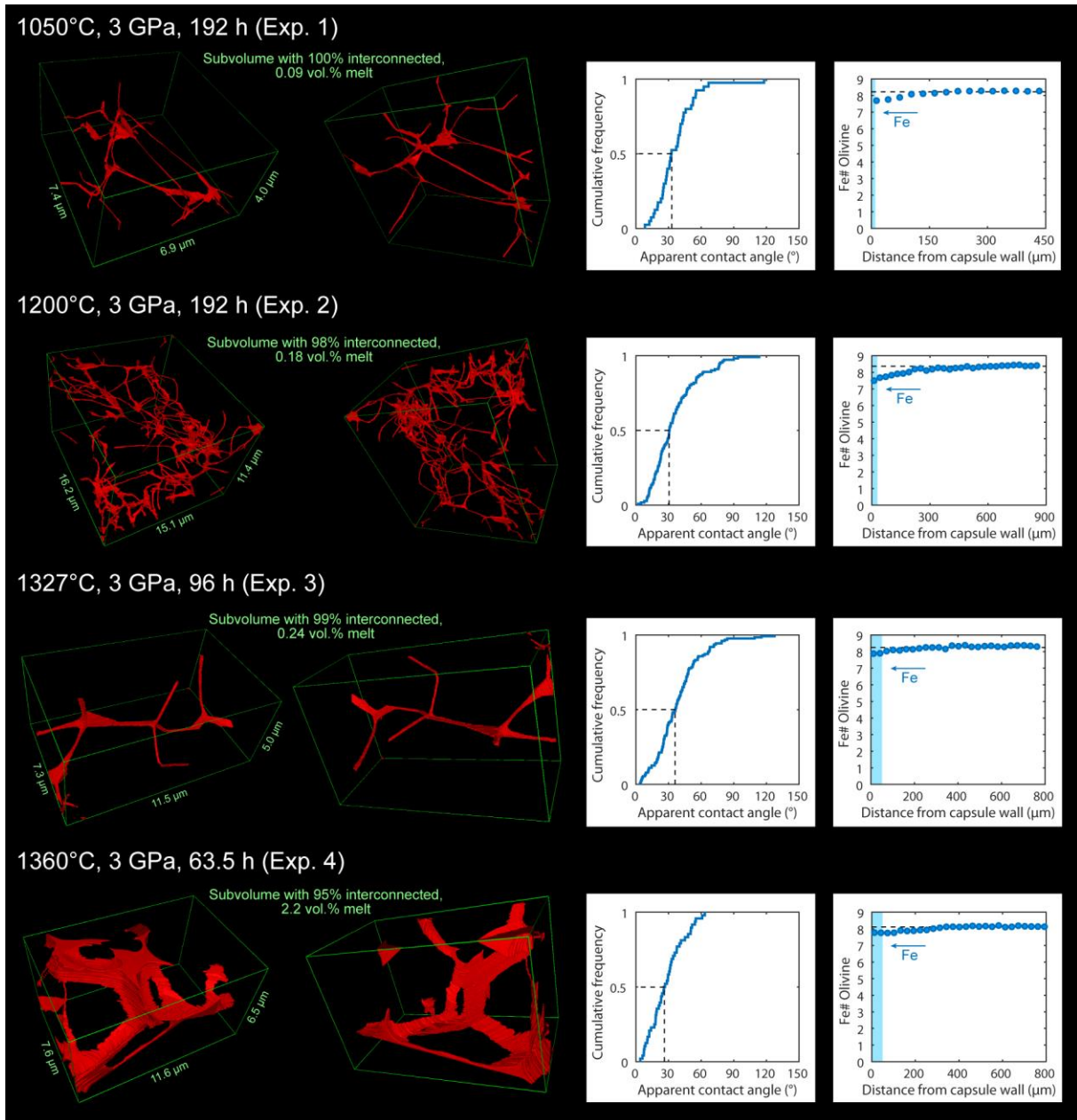
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235 **Fig. 5.** Interconnection of incipient melts in a (CO₂+H₂O)-bearing mantle rock experimentally re-equilibrated at shallow mantle
 236 temperatures and pressure. (CO₂+H₂O)-rich melts virtually fully interconnect in the intergranular medium of all samples, from
 237 1.5 vol.% melt at 1360°C (Exp. 4) down to 0.11 vol.% melt at 1050°C (Exp. 1). Left panels, high-resolution 3D reconstructions
 238 reveal virtually fully interconnected melt networks (rotations of the volumes are available as [Videos S1-S4](#)). Middle right
 239 panels, cumulative distributions of apparent contact angles at the junctions of incipient melts with the various mineral phases
 240 of the samples. Median contact angles are about 30° in all samples, i.e. much less than the 60° threshold below which
 241 interconnection occurs at any melt fraction. Right panels, long range interconnection is evidenced by a minor iron loss at the
 242 vicinity of Au-Pd capsule walls. For comparison, the much shorter iron diffusion lengths that would have occurred in melt-free
 243 aggregates are reported as pale blue areas (see [Supplementary Material](#)).

244

245 4. Discussion

246 4.1. Predominance of interconnected, volatile-rich melts in the LVZ

247 Our experiments directly evidence incipient melting where minute amounts of both CO₂ and H₂O
248 generate minute amounts of (CO₂+H₂O)-rich melts in mantle rocks down to lithosphere temperatures
249 (Figs. 2-4). They corroborate studies that have long predicted this phenomenon from volatile-doped
250 experiments at much higher melt fractions (Wallace and Green, 1988; Hirshmann, 2010; Dasgupta et
251 al., 2013; Massuyeau et al., 2015; Gaillard et al., 2019) (Figs. 1, 4). Note that, although our experiments
252 were performed at 3 GPa, volatile-assisted melting was also observed at much lower and much higher
253 pressures covering the range of LVZ pressures (see e.g. Massuyeau et al., 2015 and references therein).
254 Hence, incipient melting occurs in the largest part of the oceanic mantle, generating minute amounts of
255 melt highly enriched in both CO₂ and H₂O and depleted in SiO₂ (e.g. Exp. 1 to 3 from 1050 to 1327°C).
256 At the vicinity of hot settings like ridges or hotspots, melt fraction and SiO₂ content increase while
257 volatile content dilutes as temperature enhances silicate melting (e.g. Exp. 4 at 1360°C).

258 As volatile content is very low and highly heterogeneously distributed in the upper mantle (Le Voyer et
259 al., 2017), the question of melt interconnection in highly volatile-depleted mantle regions with extremely
260 low melt fraction arises. Interconnection is found invariably complete down to the ~0.1 vol.% volatile-
261 rich melt produced in our experiments, with no evidence of deterioration (Fig. 5). This is confirmed by
262 the ~30° contact angle of partially molten mantle rock in the (CO₂+H₂O)-assisted melting domain which
263 is significantly smaller than the 60° threshold below which interconnection occurs at any melt fraction
264 (von Bargen and Waff, 1986) (Fig. 5).

265 The persistence of interconnection at extremely low melt fractions was also observed in experiments in
266 simplified petrological systems and at higher temperatures. The smallest silicate melt fraction in
267 monomineralic olivine aggregate ever investigated, i.e. ~0.01 vol.%, was found to be interconnected at
268 1200°C and 0.2 GPa (Faul et al., 2004). Minarik and Watson (1995) investigated bulk mass transport as
269 a function of Na-carbonate melt fractions in monomineralic olivine aggregates at 1300°C and 1 GPa,
270 and reported transport enhancements down to ~0.007 vol.% melt. The absence of detection of mass

271 transport enhancement at lower melt fractions does not imply that melt interconnection stops, but that
272 the enhancement could be too weak to be distinguished from mass transport in melt-free sample. It
273 should be noted that [Minarik and Watson \(1995\)](#) inferred an interconnection cut-off at ~0.07 vol.% from
274 their longest experiments (~100 h). However these conclusions were very likely misled by a loss of melt
275 to the surrounding graphite medium in the longest experiments given the extremely fast migration of
276 Na-carbonate in olivine aggregates ([Hammouda and Laporte, 2000](#)). [Laumonier et al. \(2017\)](#)
277 investigated the electrical conductivity of mixtures of monomineralic olivine aggregate with silicate
278 melt and reported conductivity enhancement already at the smallest added melt fraction, i.e. 0.15 vol.%.
279 The 0.5 vol.% threshold pointed by [Laumonier et al. \(2017\)](#) thus does not represent a connection
280 threshold, but might rather be related to an evolution of melt geometry such as the increase in grain
281 boundary wetness observed between our samples with ~0.1-0.15 vol.% melt (Exp. 1-3) and with 1.5
282 vol.% melt (Exp. 4) ([Fig. 5, Table 2](#)).

283 Melt thus form interconnected networks down to extremely low melt fractions, without evidence for any
284 interconnection threshold, as predicted for systems with contact angles below 60° ([von Bargaen and](#)
285 [Waff, 1986](#)). Melt interconnection is also predicted to be independent on grain size ([von Bargaen and](#)
286 [Waff, 1986](#)), being micrometric as in experiments or millimetric as in the mantle. Note that, aside from
287 melt interconnection, melt mobility should be enhanced at large mantle grain size as it implies melt
288 channels with larger cross-sectional areas, and thus smaller surface-to-volume ratio, favouring buoyancy
289 over capillarity ([von Bargaen and Waff, 1986](#); [Holtzman, 2016](#)). It is also worth noticing that melt
290 interconnection has been reported over the range of LVZ pressures, in experiments from e.g. 0.2 GPa
291 ([Faul et al., 2004](#)) to 7 GPa ([Yoshino et al., 2009](#)). Therefore, interconnected, volatile-rich melts should
292 prevail in the mantle, in both the asthenosphere and the lithosphere, down to the coldest temperatures of
293 the (CO₂+H₂O)-assisted melting domain, and in highly volatile-depleted regions.

294 *4.2 Correspondence of geophysical anomalies and volatile-assisted melting in the LVZ*

295 The nature of the low velocity zone (LVZ) in oceanic mantle can be reassessed in light of our findings.
296 Sharp drops in shear wave velocity (Vs) are detected using SS precursors in the shallow mantle beneath
297 Pacific, with Vs lowering ranging from ~3 to 22% ([Rychert and Shearer, 2011](#); [Schmerr, 2012](#);

298 [Tharimena et al., 2017](#)). The top of these seismic anomalies locates at 30-80 km depth and only slightly
299 deepens as a function of lithosphere age ([Fig. 6](#)). The onset of increase in Vs radial anisotropy in Pacific
300 locates at similar depths ([Burgos et al., 2014](#)), and also only slightly deepens with age ([Fig. 6](#)).
301 Moreover, though scarcer than seismic data, very high electrical conductivities > 0.1 S/m were detected
302 at similar depths, such as locally beneath the edge of the Cocos plate at 45-70 km depth ([Naif et al.,](#)
303 [2013](#)) ([Fig. 6](#)). The top of these anomalies was referred to as the Gutenberg discontinuity (G) (e.g.
304 [Kawakatsu et al., 2009](#); [Schmerr, 2012](#)). Comparison with model of the thermal structure of oceanic
305 mantle (G13R1350 model with 1350°C potential temperature from [Grose and Afonso, 2013](#)) evidences
306 that the G-discontinuity locates at depths where temperature can be as low as $\sim 950^\circ\text{C}$, within the
307 lithosphere ([Fig. 6](#)). Actually, the depth of the G-discontinuity, together with its slight deepening with
308 age, is in remarkable agreement with the $(\text{CO}_2+\text{H}_2\text{O})$ -assisted melting curve ([Wallace and Green, 1988](#);
309 [Sifré et al., 2014](#)) ([Fig.6](#)). Since both CO_2 and H_2O are sampled in the mantle ([Le Voyer et al., 2017](#))
310 and since we show here they induce fully interconnected volatile-rich melts, this spatial correspondence
311 evidences that the G-discontinuity delimits the top of a volatile-assisted melting zone.

312 It is worth recalling that the hydration of olivine is unable to produce such strong anomalies in both
313 seismic velocities ([Cline et al., 2018](#)) and electrical conductivities ([Gardés et al., 2014, 2015](#)).
314 Temperature excess can enhance partial melting but it appears to be of second order compared to volatile
315 content. Temperature anomalies would not generate sharp discontinuities such as G-discontinuity
316 ([Kawakatsu et al., 2009](#)) and would not yield a spatial correlation of G-discontinuity with the
317 $(\text{CO}_2+\text{H}_2\text{O})$ -solidus ([Fig. 6](#)). Moreover, in a mantle with average volatile content of 140 wt. ppm CO_2
318 and 240 wt.ppm H_2O ([Le Voyer et al., 2017](#)), reproducing the ~ 0.2 S/m electrical conductivities at 45-
319 70 km depths in the ~ 25 Ma lithosphere reported by [Naif et al. \(2013\)](#) would require a partially molten
320 mantle with temperatures above 1550°C ([Sifré et al., 2014](#)), while temperatures are expected to be 1100-
321 1300°C in this setting ([Grose et Afonso, 2013](#); [Fig. 6](#)). On the other hand, those high electrical
322 conductivities can be reached at temperatures around 1200°C in a partially molten mantle enriched by
323 3-4 times the average volatile content ([Sifré et al., 2014](#)).

324 Although sharp Vs drops mainly distributes at G-discontinuity, they are also found down to ~180 km
325 depth beneath Pacific (Fig. 6). Experimental petrology predicts that the decrease of oxygen fugacity as
326 a function of depth in the upper mantle should limit the range of stability of oxidised forms of carbon,
327 and thus CO₂-bearing melts. Estimations for this redox freezing boundary (RFB) range between 150 and
328 250 km depth (Stagno and Frost, 2010; Rohrbach and Schmidt, 2011), in line with the ~180 km depth
329 below which no more sharp Vs drops are observed. The distribution of sharp Vs drops thus coincides
330 with a melting zone, which extends from G-discontinuity in the lithosphere, i.e. at the location of the
331 (CO₂+H₂O)-solidus, down to the RFB in the asthenosphere beneath the whole Pacific basin (Fig. 6). It
332 is worth noticing that such a spatial extent can only result from (CO₂+H₂O)-assisted melting. Volatile-
333 free-melting and H₂O-only-assisted melting are not possible at high depth or beneath mature to old
334 lithosphere (Figs. 1, 6), and they are irrelevant since both CO₂ and H₂O are detected in the mantle (Le
335 Voyer et al., 2017).

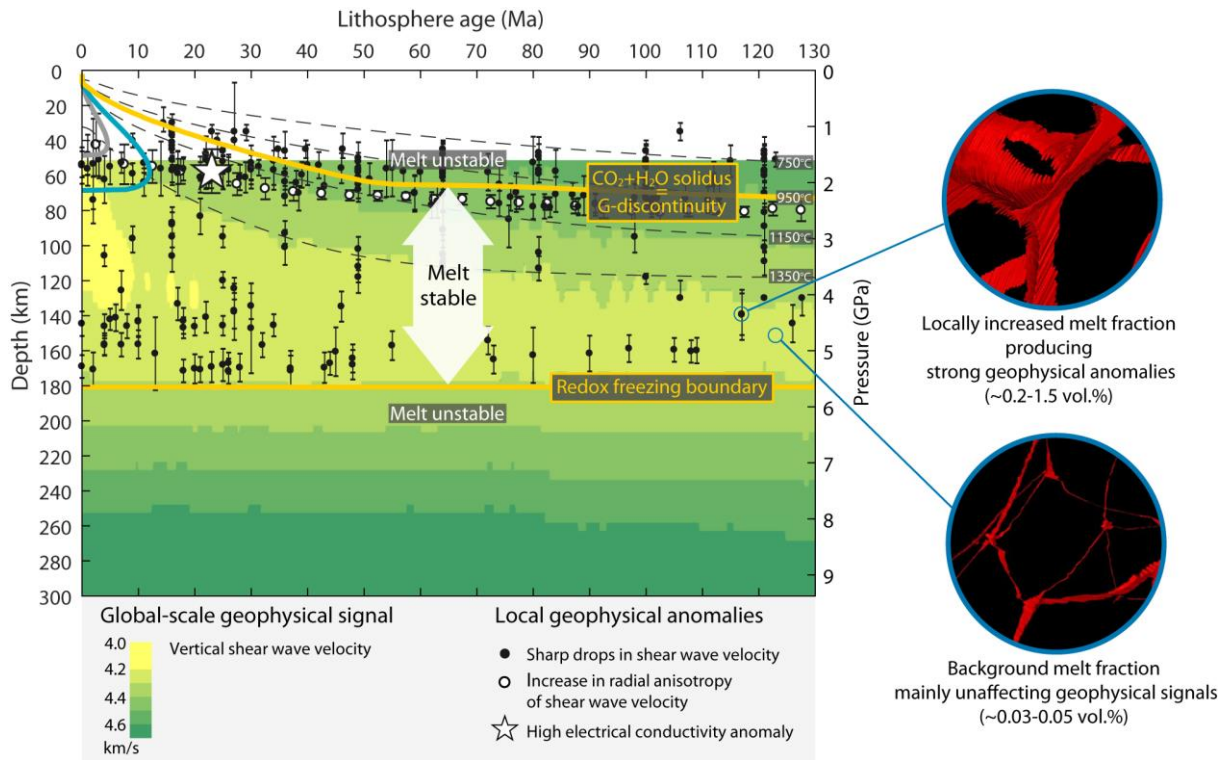
336 *4.3 Low and heterogeneous distribution of volatile-rich melt imaged by geophysical surveys in the LVZ*

337 The debate on the origin of the LVZ long opposed solid state mechanisms (e.g. Hirth and Kohlstedt,
338 1996; Prestley and McKenzie, 2013; Takei, 2017; Karato and Park, 2019) and partial melting (e.g.
339 Hirshmann, 2010; Schmerr, 2012; Dasgupta et al., 2013; Naif et al., 2013; Sifré et al., 2014; Chantel et
340 al., 2016; Holtzman, 2016; Gaillard et al., 2019). Nevertheless, the magnitude of geophysical signals
341 from the LVZ must result from the combination of both phenomena, with respective contributions
342 depending on melt fraction. Apart from the strong, local geophysical anomalies discussed above, most
343 of the locations investigated by Schmerr (2012) do not yield measurable sharp drops in shear wave
344 velocity using SS precursors. Moreover, negligible amount of melt is required to account for the
345 electrical conductivity beneath the mature lithosphere at NoMelt area in Pacific (Sarafian et al., 2015).
346 Away from hot settings such as ridges or hotspot, the smooth evolution of global-scale Vs to its
347 minimum at ~150 km depth (Fig. 6) is compatible with the response of solid mantle rocks to the changes
348 of temperature and pressure in the LVZ (Prestley and McKenzie, 2013; Takei, 2017). This indicates
349 that, on average, melt content is very low in the mantle.

350 This is consistent with the very low volatile contents reported by geochemical surveys, with average
351 values estimated at ~140 wt.ppm CO₂ and ~240 wt.ppm H₂O (Le Voyer et al., 2017). These values are
352 ~3 times less than in present experiments where incipient melting of mantle rock with ~500 wt.ppm CO₂
353 and ~630 wt.ppm H₂O bulk content induced ~0.1-0.15 vol.% melt. This yields an average melt fraction
354 of ~0.03-0.05 vol.% in the upper mantle. In absence of measurement of Vs as a function of small
355 fractions of (CO₂+H₂O)-rich melts, one can tentatively estimate a threshold for local, sharp Vs drops
356 detection from Vs measurements on olivine aggregates mixed with basaltic melt by Chantel et al. (2016).
357 Note that the similar ~30° contact angles for the (CO₂+H₂O)-rich melts measured here indicate a
358 comparable interconnection geometry with basaltic melts (von Barga and Waff, 1986). The detection
359 of sharp Vs drops over less than ~5 km using SS precursors requires Vs lowering by at least ~2%
360 according to Schmerr (2012). This requires a melt fraction of at least ~0.1 vol.% according to Chantel
361 et al. (2016). The ~0.03-0.05 vol.% average melt fraction in the mantle is thus below threshold for sharp
362 Vs drops detection using SS precursors, and consistent with mainly unaffected global-scale geophysical
363 signals (Preistley and McKenzie, 2013; Takei, 2017) (Fig. 6). Nevertheless, geochemical surveys
364 evidence that mantle volatiles distribution is highly heterogeneous, and volatile contents reported above
365 are averages from highly dispersed values (Le Voyer et al., 2017). The local detections of sharp Vs
366 lowering of ~3 to 22% using SS precursors (Rychert and Shearer, 2011; Schmerr, 2012; Tharimena et
367 al., 2017) are thus in line with a mantle with local melt enrichments, corresponding to regions with ~0.2
368 to 1.5 vol.% melt according to Chantel et al. (2016) (see their preferred curve at anelasticity attenuation
369 factor $\alpha = 0.26$ in their Fig. 3). These values fall in the range of melt fractions produced in our
370 experiments (Table 2), indicating local volatile enrichments in the mantle of more than ~3 times the
371 average mantle content.

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374

375 **Fig. 6.** Correspondence of volatile-assisted melting domain and geophysical anomalies domain in the LVZ of Pacific upper
 376 mantle. Apart from ridge, global-scale shear wave velocity is mainly unaffected by incipient, interconnected melt because of a
 377 too low background melt fraction, estimated at $\sim 0.03\text{-}0.05$ vol.%. Sporadic, localized geophysical anomalies arise from areas
 378 with increased melt fractions, estimated to range from ~ 0.2 to ~ 1.5 vol.% for the detection of sharp Vs drops using SS
 379 precursors. These anomalies lay in the stability domain of $(\text{CO}_2+\text{H}_2\text{O})$ -assisted melting. This domain extends beneath the whole
 380 ocean basin, and is delimited at its top by the $(\text{CO}_2+\text{H}_2\text{O})$ -solidus (Wallace and Green, 1988), which coincides with G-
 381 discontinuity, and at its bottom by the redox freezing boundary at ~ 180 km depth, in the range of petrological estimations (150-
 382 250 km; Stagno and Frost, 2010; Rohrbach and Schmidt, 2011). Volatile-free-melting (grey curve) and H_2O -only-assisted
 383 melting (blue curve) domains (Sarafian et al., 2017) are restricted to the vicinity of ridge. Note that the G-discontinuity locates
 384 at depths where temperature can be as low as $\sim 950^\circ\text{C}$, within the lithosphere. Vertical shear wave velocity contours are
 385 PM_v2_2012 tomographic model from Prestley and McKenzie (2013). Sharp drops in shear wave velocity obtained from SS
 386 precursors are from Rychert and Shearer (2011), Schmerr (2012) and Tharimena et al. (2017). Increase in radial anisotropy of
 387 shear wave velocity is from Burgos et al. (2014). High electrical conductivity anomaly is from Naif et al. (2013). Mantle
 388 thermal structure is G13R1350 model with 1350°C potential temperature from Grose and Afonso (2013). The illustrations of
 389 grain scale melt interconnection at increased and background melt fractions are based on the 3D imaging of our samples (not
 390 to scale).

391

392 Geophysical surveys thus appears to image *in situ* the very low and highly heterogeneous distribution
393 of melt in the mantle corresponding to the very low and highly heterogeneous distribution of volatiles
394 reported by surficial geochemical surveys. While partial melts are probably ubiquitous and must
395 interconnect in the LVZ, the global-scale geophysical signature of the LVZ mainly derives from solid
396 state processes because the background melt fraction corresponding to mantle average volatile content
397 is very low, being ~0.03-0.05 vol.% melt for 140 wt.ppm CO₂ and 240 wt.ppm H₂O (Fig. 6). Enhanced
398 signals arise from sporadic, localized areas where melt fraction is increased (Fig. 6), e.g. above ~0.2
399 vol.% melt for detecting sharp Vs drops using SS precursors which corresponds to mantle volatile
400 enrichments above ~500 wt.ppm of both CO₂ and H₂O. Geophysical surveys also appear to reveal the
401 dynamics of melt redistribution and localization in the LVZ. This is most striking at G-discontinuity
402 where the concentration of geophysical anomalies suggests a zone of melt accumulation, though
403 heterogeneous and likely corresponding to an average melt fraction well below 1 vol.%. One might
404 speculate that volatile-rich melts generated from the asthenosphere migrate upwards to the lithosphere
405 by percolation via interconnected networks and/or mantle convection (Rabinowicz et al., 2002). They
406 are slowed down at G-discontinuity, i.e. at (CO₂+H₂O)-solidus, above which they must solidify, unless
407 local setting allows for fast, out of equilibrium transfer to shallower depths (Hirano et al., 2006; Machida
408 et al., 2017).

409 In-depth deciphering of the dynamics of melt and volatiles in the LVZ calls for investigations on the
410 seismic velocity, permeability and rheology of partially molten mantle rocks covering the diversity of
411 mantle melt compositions, fractions and temperatures. This is also critical for deciphering Earth's global
412 dynamics. If small fractions of volatile-rich melts significantly lower the viscosity of mantle rocks (Hirth
413 and Kohlstedt 1995a, 1995b, 2003; Takei and Holtzman, 2009a, 2009b; Holtzman, 2016), the LVZ
414 could be a weakening zone between the rigid lithospheric plates and the convective mantle, playing key
415 role in plate tectonics.

416

417

418 **5. Conclusion**

419 (i) We experimentally evidence the long predicted incipient melting regime where minute amounts of
420 both CO₂ and H₂O generate minute amounts of (CO₂+H₂O)-rich melts in mantle rocks down to the
421 coldest temperatures of the (CO₂+H₂O)-assisted melting domain, i.e. in both the asthenosphere and the
422 lithosphere.

423 (ii) We experimentally evidence that those minute amounts of volatile-rich melts do interconnect in
424 mantle rocks, enabling thus the modification of geophysical signals from the mantle. No interconnection
425 threshold can be evidenced or predicted. Interconnection of volatile-rich melts must prevail even at
426 extremely low melt fractions in highly volatile-depleted regions of the mantle.

427 (iii) The distribution of geophysical anomalies in the oceanic mantle corresponds to a melting zone
428 which extends from the (CO₂+H₂O)-solidus at ~30-80 km depth in the lithosphere, corresponding to
429 Gutenberg discontinuity, down to the redox freezing boundary in the asthenosphere at ~180 km depth.
430 G-discontinuity locates at depths where temperature can be as low as ~950°C, within the lithosphere.

431 (iv) Geophysical surveys appears to image *in situ* the very low and highly heterogeneous distribution of
432 melt in the mantle generated by the very low and highly heterogeneous distribution of volatiles probed
433 by surficial geochemical surveys. The global-scale geophysical signature of the LVZ appears mainly
434 unaffected because the average background melt fraction is very low, estimated at ~0.03-0.05 vol.%
435 melt. Nevertheless, enhanced geophysical signals arise from sporadic, localized areas where melt
436 fraction is above a threshold, estimated at ~0.2 vol.% melt for sharp Vs drop detection using SS
437 precursors.

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562

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