

# Experimental Investigation on the Transport of Sulfide Driven by Melt-rock Reaction in Partially Molten Peridotite

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## Abstract

Extraction of sulfide liquid from partially molten mantle is vital to elucidate the cycling of metal and sulfur elements between different geochemical circles but has not been investigated systematically. Using the reaction couple method of laboratory experiments and theoretical calculations, this study documents systematical variations in lithologies and compositions of silicate minerals and melts, which are approximately consistent with the results of thermodynamically-constrained model. During melt-peridotite reaction, dissolution of olivine and precipitation of new orthopyroxene produce an orthopyroxene-rich layer between melt source and peridotite. With increasing reaction degree, more melt is infiltrated into and reacts with upper peridotite, which potentially enhances the concomitant upward transport of dense sulfide droplets. Theoretical analyses suggest an energetical focused melt flow with a high velocity ( $\sim 170.9 \mu\text{m/h}$ ) around sulfide droplet through pore throat. In this energetic melt flow, we, for the first time, observed the mechanical coalescence of sulfide droplets, and produced drag force was likely driving upward entrainment of fine  $\mu\text{m}$ -scale sulfide. For coarse sulfide droplets whose sizes are larger than the pore throat in partially molten peridotite, their entrainment through narrow constrictions in crystal framework seems to be physically possible only when high-degree melt-peridotite reaction drives high porosity of peridotite and some channelized melt flows with extremely high velocity. Hence, melt-rock reaction could drive and enhance upward entrainment of  $\mu\text{m}$ - to  $\text{mm}$ -scale sulfide in the partially molten mantle, potentially contributing to the fertilization of the sub-continental lithospheric mantle and the endowment of metal-bearing sulfide for the formation of magmatic sulfide deposits.

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41 **Plain Language Summary**

42 Sulfides are a pivotal potential reservoir for sulfur and economically important metals. Their  
43 transport in the Earth's mantle plays a vital role in understanding many crucial geological and  
44 environmental processes, especially the formation of mineral deposits, and the environmental  
45 damage and health hazards related to volcanic eruptions. This work proposes a new driving force  
46 for upward transport of dense sulfide drops in the upper mantle that experiences partial melting.  
47 Reaction between melt and rock potentially leads to focused melt flow in new-forming channels  
48 with three orders of magnitude higher velocity than that of melt flowing among crystal  
49 framework of peridotite. This energetic melt flow drives upward transport of tiny  $\mu\text{m}$ -scale

50 sulfide droplets in peridotite and may also facilitate amalgamation of droplets contacting each  
51 other. Coarse sulfide droplets could be possibly entrained upward through narrow pore throats,  
52 especially when high-degree melt-peridotite reaction drives fast-flowing melt in the mantle with  
53 high porosity.

## 54 **1 Introduction**

55 Sulfide is ubiquitous in mantle rocks (Alard et al., 2011) and an important repository for  
56 sulfur and geochemically and economically important chalcophile metals, which plays an  
57 important role in the partitioning behaviors of PGE, Cu, and Ni (Mungall & Brenan, 2014; Patten  
58 et al., 2013). Understanding the factors that control the fate of sulfide phases in the partially  
59 molten mantle is of fundamental importance in exploring the recycling of sulfur and chalcophile  
60 elements among different geochemical reservoirs (Ding & Dasgupta, 2017; Farquhar et al., 2002;  
61 Yao et al., 2018; Chen et al., 2022) and identifying the re-fertilization of the depleted  
62 lithospheric mantle, potentially providing the metal endowment for the formation of Cu-rich  
63 porphyry and/or Ni-rich magmatic ore system (Holwell et al., 2022; Lee & Tang, 2020; Mungall  
64 et al., 2015; Zhao et al., 2022). However, it remains highly contentious for driving forces of the  
65 transport of metal-bearing sulfide liquid, which severely blocks our understanding of the details  
66 of fertilization processes occurred in the source of sulfide-related magmatic and hydrothermal  
67 deposits.

68 Conventionally, the removal of sulfide liquid from a partially molten peridotite requires  
69 sulfide liquid to be progressively dissolved by the departing silicate melt (Holzheid & Grove,  
70 2002; Mungall & Brenan, 2014; Yao et al., 2018). Nevertheless, the mechanical entrainment of  
71 sulfide liquid could also potentially be an efficient process driving the redistribution and local  
72 enrichment of sulfur and chalcophile elements in the mantle source (Iacono-Marziano et al.,  
73 2022; Heinrich & Connolly, 2022; Z. Wang et al., 2020; Yao & Mungall, 2020). Although a  
74 small amount ( $<$  percolation thresholds) of sulfide liquid under hydrostatic conditions cannot be  
75 entrained through the porous flow of silicate melt due to the high surface tension (Bagdassarov et  
76 al., 2009; Holzheid et al., 2000; Yoshino et al., 2003, 2004), it has been argued that the  
77 extraction of sulfide liquid driven by stress in a partially molten peridotite could be further  
78 facilitated by the directional porous flow of silicate melt, strongly affecting the fractionation and  
79 abundances of chalcophile metals in the mantle melting products, *i.e.*, basalts (Ballhaus et al.,

80 2006; Bockrath et al., 2004; Z. Wang et al., 2020). Except for the deviatoric stress, a preliminary  
81 experimental study (Wang & Jin, 2020) proposed that during melt-rock reaction, the entrainment  
82 of sulfide liquid may be enhanced by the reaction infiltration instability (RII) in the partially  
83 molten peridotite when a reacting melt percolates through a dissolvable, porous, melt-mineral  
84 mixture. The mechanical transport and enrichment of sulfide during this melt-rock reaction have  
85 been extensively demonstrated in mantle peridotites, which may clarify some geochemical  
86 paradoxes of the chalcophile/siderophile elements (Ciazela et al., 2018; Lorand & Luguet, 2016).  
87 The RII theory proposed by Chadam et al. (1986) also applies to magmatic systems by  
88 introducing viscous compaction of the matrix and driving a solubility gradient along the flow  
89 direction in place of the propagating reaction front (Aharonov, 1995), which induces positive  
90 feedback between the increasing permeability due to melt-rock reaction and the associated  
91 increase of melt flux in reacting regions. The reaction between peridotite and silicate melt is well  
92 known as a pervasive process even forming melt-rich channels with an extremely speedy melt  
93 flow due to the positive feedback in partially molten regions of the mantle, which has been  
94 broadly demonstrated by geological investigations (*e.g.*, Kelemen et al., 1995; Sundberg et al.,  
95 2010), high-temperature, high-pressure (HTHP) laboratory works (*e.g.*, Daines & Kohlstedt,  
96 1994; Pec et al., 2015) and theoretical analyses (Aharonov, 1995; Chadam et al., 1986;  
97 Spiegelman et al., 2001). During a reaction of melt with a depleted mantle, melt re-fertilization  
98 processes potentially occur, thereby impregnating the mantle with sulfides, which have been  
99 widely found in mantle xenoliths, orogenic and abyssal peridotites (*e.g.*, Ciazela et al., 2018;  
100 Luguet et al., 2003; Niu, 2004; Wang et al., 2009).

101 Although it has been tentatively demonstrated that the rapid enough ascending magma  
102 flow in those melt-rich channels could hold upwards entrainment of large olivine phenocrysts or  
103 even dense sulfide droplets during the melt-rock reaction (Pec et al., 2017; Wang & Jin, 2020),  
104 the extraction of sulfide liquid driven by the RII has received very little attention, and the precise  
105 physical and chemical constraints during these processes appear to be worth studying  
106 thoroughly, which is vital to the fertilization of lithospheric mantle, as well as the formations of  
107 magmatic and/or hydrothermal sulfide deposits. Here we conducted systematically two-layer  
108 reaction experimental studies in which a partially molten peridotite is placed on a sulfide-bearing  
109 silicate melt source to explore the entrainment of sulfide liquids driven by the RII and the  
110 physical and chemical changes of liquid and solid phases during these processes.

111 **2 Materials and Methods**

## 112 2.1 Starting materials

113 As a melt source, the starting materials were a mixture of the powdered calc-alkaline  
 114 tholeiitic basalt (70 wt.%) from the East Pacific Rise (102.7044 °W, 2.64961 °S), olivine crystals  
 115 (20 wt.%) from fresh spinel lherzolite xenoliths at Damaping (Hannuoba region), North China,  
 116 and sulfide aggregates (10 wt.%) from Jinchuan Ni-Cu sulfide deposit, NW China. Additionally,  
 117 0 or 2 wt.% oceanic sediments (mainly carbonate) were mixed into the melt source to explore the  
 118 effect of volatile (CO<sub>2</sub>) on the transport of sulfides. The Ni-Cu sulfides were composed of  
 119 pyrrhotite (59 wt.%), pentlandite (36 wt.%), and chalcopyrite (5 wt.%) separates, similar to the  
 120 composition of base-metal sulfide aggregates in massif peridotites (Lorand et al., 2010), and  
 121 were grounded in alcohol for about 6h to < 10 μm measured by using the scanning electron  
 122 microscope (SEM).

123 To prepare a partially molten rock, olivine and clinopyroxene crystals from Damaping  
 124 spinel lherzolite xenoliths were mixed in a 50:50 ratio by weight. Subsequently, ~ 0 or ~ 5 wt.%  
 125 calc-alkaline tholeiitic basalts were added to change the initial permeability of the partially  
 126 molten rock. All silicate mineral grains were ground to 10-20 μm grain size in an agate mortar.

127 The partially molten rock was placed on the melt source, and the length ratio of these two  
 128 parts was slightly larger than ~1:1 to avoid chemical equilibration between the melt source and  
 129 the partially molten rock over experimental time scales. Compositions of these starting materials  
 130 were described by Z. Wang et al. (2020), and experimental conditions are listed in Table 1.

131 **Table 1.** Summary of experimental conditions and observed lithologies.

NO.	Starting material (upper peridotite/lower melt source)	Experimental conditions (P/GPa, T/°C, t/h)	Capsule	Lithologies (upper peridotite/reaction interface /melt source)
PC520	Ol:Cpx(1:1)/Ol(20%)+basalt(70%)+sulfide(10%)	1.5,1250,12	Pt-graphite	Ol+Cpx+SM+SL/ORL (Opx+SM+SL)/ Opx+SM+SL
PC527	Ol:Cpx(1:1)/Ol(20%)+basalt(70%)+sulfide(10%)	1.5,1300,12		Ol+Cpx+SM+SL/olivine-melt layer (Ol+SM+SL)/ORL (Opx+SM+SL)/ Opx+SM+SL
PC528	Ol:Cpx(1:1)/Ol(20%)+basalt(70%)+sulfide(10%)	1.5,1250,48		Ol+Cpx+SM+SL/ORL (Opx+SM+SL)/ Opx+SM+SL
PC537	Ol:Cpx(1:1)/Ol(20%)+basalt(70%)+sulfide(10%)	1.5,800,12		Ol+Cpx/Ol+basalt+sulfide

PC545	Ol:Cpx(1:1)/Ol(20%)+basalt(70%) +sulfide(10%)	1.5,1250,72		Ol+Cpx+SM+SL/olivine-melt layer (Ol+SM+SL)/ORL (Opx+SM+SL)/Opx+SM+SL
PC548	Ol:Cpx(1:1)+basalt(5wt%)/Ol(20%) +basalt(70%)+sulfide(10%)	1.5,1250,48		Ol+Cpx+SM+SL/ORL (Opx+SM+SL)/Opx+SM+SL
PC559	Ol:Cpx(1:1)+basalt(5wt%)/Ol(20%) +basalt(70%)+sulfide(10%)	0.5,1200,6		Ol+Cpx+SM+SL/Ol+SM+SL
PC560	Ol:Cpx(1:1)+basalt(5wt%)/Ol(20%) +basalt(68%)+carbonate(2%) +sulfide(10%)	1.5,1250,24		Ol+Cpx+SM+SL/olivine-melt layer (Ol+SM+SL)/ORL (Opx+SM+SL)/Opx+SM+SL

132 Note: Ol-olivine; Cpx-clinopyroxene; SM-silicate melt; SL-sulfide liquid; ORL-orthopyroxene-rich reaction  
133 layer

## 134 2.2 Experimental methods

135 The two-layer reaction experiments were conducted at the State Key Laboratory of  
136 Geological Processes and Mineral Resources (GPMR) of China University of Geosciences using  
137 a 150 Ton non-end-loaded type piston-cylinder press. Starting materials were loaded into a 3.75-  
138 mm-diameter and 5-6-mm-high cylindrical platinum (Pt) capsule with a graphite inner sleeve  
139 (Figure 1-inset in the lower left). In this assembly, the sample was in contact only with the  
140 graphite inner sleeve and therefore, the oxygen fugacity of this experimental system was  
141 maintained at about CCO-0.8 (the graphite-CO<sub>2</sub> buffer), which corresponds to a log  $fO_2 < FMQ$ -  
142 2 log unit at ~1360 °C and 1.5 GPa (e.g., Médard et al., 2008). A low friction assembly  
143 consisting of NaCl and Pyrex sleeves, a graphite heater, sintered MgO spacers, and an Al<sub>2</sub>O<sub>3</sub>  
144 plug, was used for the piston-cylinder experiments. Pressure calibration was conducted against  
145 the quartz/coesite phase transition, and was accurate to ± 0.1 GPa. The temperature was  
146 monitored by a W/Re type C thermocouple located at the bottom of the capsule. These  
147 experiments (Table 1) were conducted under pressures of 0.5-1.5 GPa and temperatures of 800-  
148 1300 °C for 12-72 h before being quenched to room temperature. All capsules were heated  
149 before being sealed at 120 °C in a vacuum oven for more than 12 h to remove absorbed water  
150 vapor in the specimens, and no apparent sulfur loss was observed during this drying.

## 151 2.3 Analytical techniques

152 Polished sections were prepared from the recovered experimental specimens cut parallel  
153 to the specimen axis. The microstructure of experimental run products was observed using a  
154 Quanta 450 field-emission scanning electron microscope (FE-SEM) at the GPMR with an

155 accelerating voltage of 20 kV, a spot size of 6.0  $\mu\text{m}$ , and a working distance of  $\sim 12$  mm. The  
156 backscattered electron (BSE) images were used to measure the morphological characteristics of  
157 experimental products, such as the thickness of reactive boundary layers, the dissolution  
158 distance, and the size of sulfide droplets.

159 Quantitative compositional analyses for olivine and clinopyroxene were performed using  
160 a JEOL JXA-8230 electron probe microanalyzer with four wavelength-dispersive spectrometers  
161 (WDS) at the Center for Global Tectonics, School of Earth Sciences, China University of  
162 Geosciences (Wuhan). Fifteen kV accelerating voltage, 20 nA probe current, and a 1  $\mu\text{m}$  beam  
163 diameter had been used. Sulfides were analyzed with 20 kV acceleration voltage and 20 nA  
164 beam current, whereas quenched silicate melts were analyzed with 15 kV/10 nA. A defocused  
165 beam of 20-30  $\mu\text{m}$  diameter was used for all the standardizations and quenched melts and  
166 sulfides.

### 167 **3 Results**

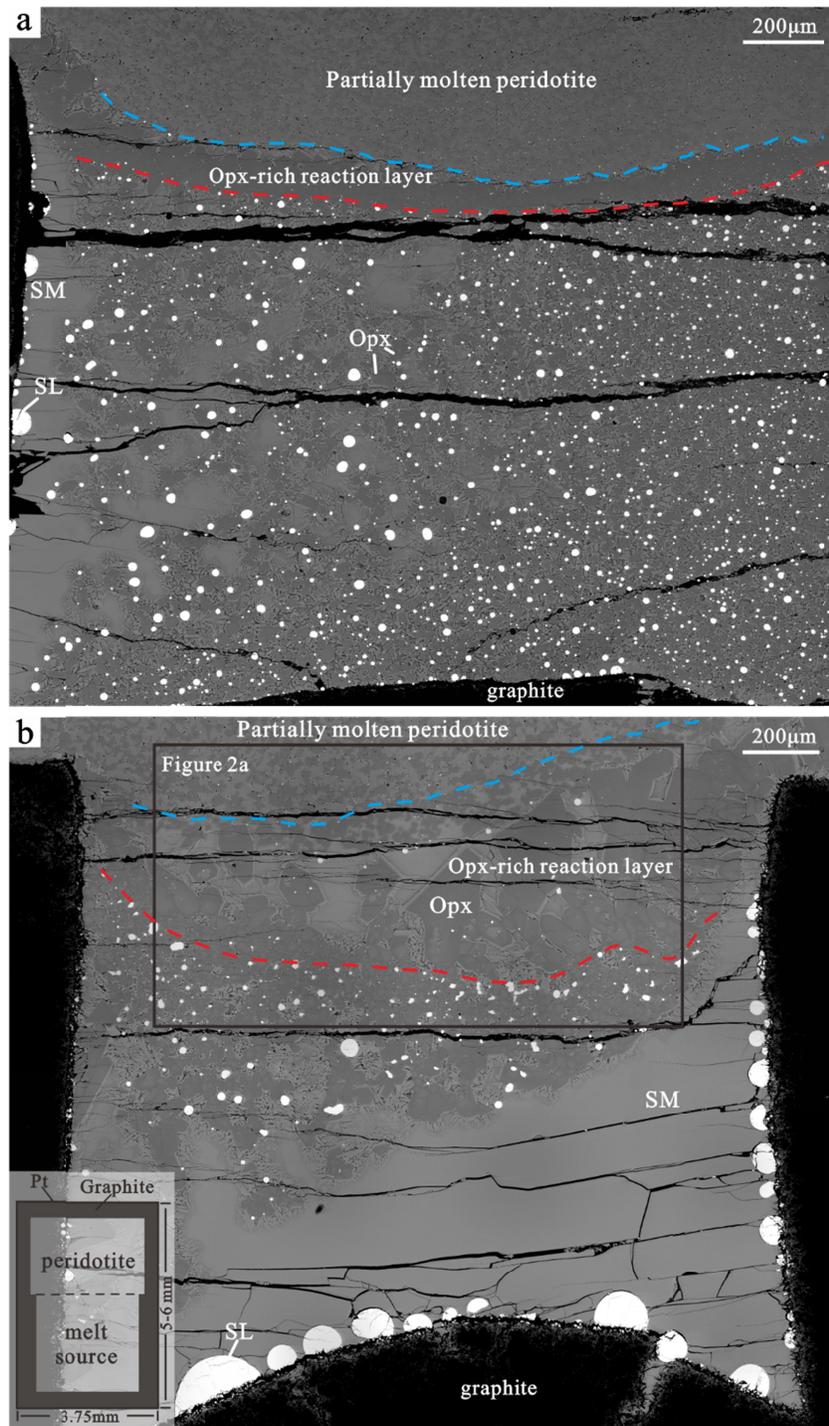
168 Several two-layer reaction experiments (Table 1) were conducted at 0.5-1.5 GPa, 800-  
169 1300  $^{\circ}\text{C}$ , and 12-72h to investigate the transport of sulfide liquid during reaction infiltration of  
170 silicate melt. Generally, several distinct lithological regions were developed and, from bottom to  
171 top, respectively separated by a diffuse-dominated, mineralogical interface in these two-layer  
172 reaction experiments (Figure 1). All two-dimensional images were analyzed using the freeware  
173 ImageJ developed by the National Institute of Health (NIH) (<http://imagej.nih.gov/ij/>).

#### 174 3.1 Phase assemblages and textures

175 Under the low-temperature, high-pressure conditions (800  $^{\circ}\text{C}$  and 1.5 GPa), both  
176 minerals and sulfides were generally polygonal and not molten, and sulfide droplets in the lower  
177 part of the sample were uniformly distributed in pores among silicate grains. At 1.5 GPa, when  
178 the experimental temperatures exceed  $\sim 1250$   $^{\circ}\text{C}$ , the basaltic material was completely molten,  
179 and the diameter of sulfide droplets ranges from several  $\mu\text{m}$  to  $> 100$   $\mu\text{m}$  (Figure 1).  
180 Gravitational segregation of sulfide was hardly ever observed in all samples, but obviously, some  
181 huge sulfide droplets were attached to the graphite capsule wall, especially at high-temperature  
182 conditions (Figure 1). These coarse droplets were removed from the collected data to estimate  
183 the average droplet size. Tiny exsolution lamellae with higher backscattered electron (BSE)

184 intensity were occasionally observed in some quenched large sulfide droplets. Sometimes silicate  
185 melts were pooled at the edge of sample (Figure 1).

186 In all high-pressure experiments (1.5 GPa), orthopyroxene (Opx)-rich reaction layer  
187 (ORL) was formed between the lower melt source and the upper partially molten peridotite due  
188 to the reaction between them, and it chiefly consists of Opx grains, silicate melt, and sulfide  
189 droplets (Figure 1). With increasing the annealing time and/or run temperature, the morphology  
190 of ORLs became more irregular (Figure 2), and the ORL at 1300°C was even ruptured due to the  
191 presence of several enormous Opx grains ( $> 100 \mu\text{m}$ ) with few or no sulfide inclusions (Figure  
192 2b). We also observed the bulge of ORL toward the melt source (Figure 1), which can be  
193 attributed to the volumetric shrinkage of the melt source after being molten and the upward  
194 percolation of silicate melt. Thus, the original interface between the melt source and peridotite  
195 before the experimental reaction cannot be marked by the offset of the graphite inner sleeve. At  
196 1250 °C, the thickness of ORL increases from  $\sim 118.4 \pm 11.1 \mu\text{m}$  ( $1\sigma$ ) at 12h to  $\sim 416.3 \pm 118.6$   
197  $\mu\text{m}$  at 72h, which is likely a linear function of the square root of the run duration (slope =  $0.78 \pm$   
198  $0.05$ ) (Figure 3a-blue line). At 1300 °C, the thickness increases considerably to  $500.3 \pm 127.9 \mu\text{m}$   
199 at 12h (Figure 3a). Additionally, based on the linear fit, the addition of 5 wt.% basalts into the  
200 partially molten peridotite (run PC548) and/or 2 wt.% oceanic sedimentary into the melt source  
201 (run PC560) seems to have no significant effect on the thickness of ORL (Figure 3a).

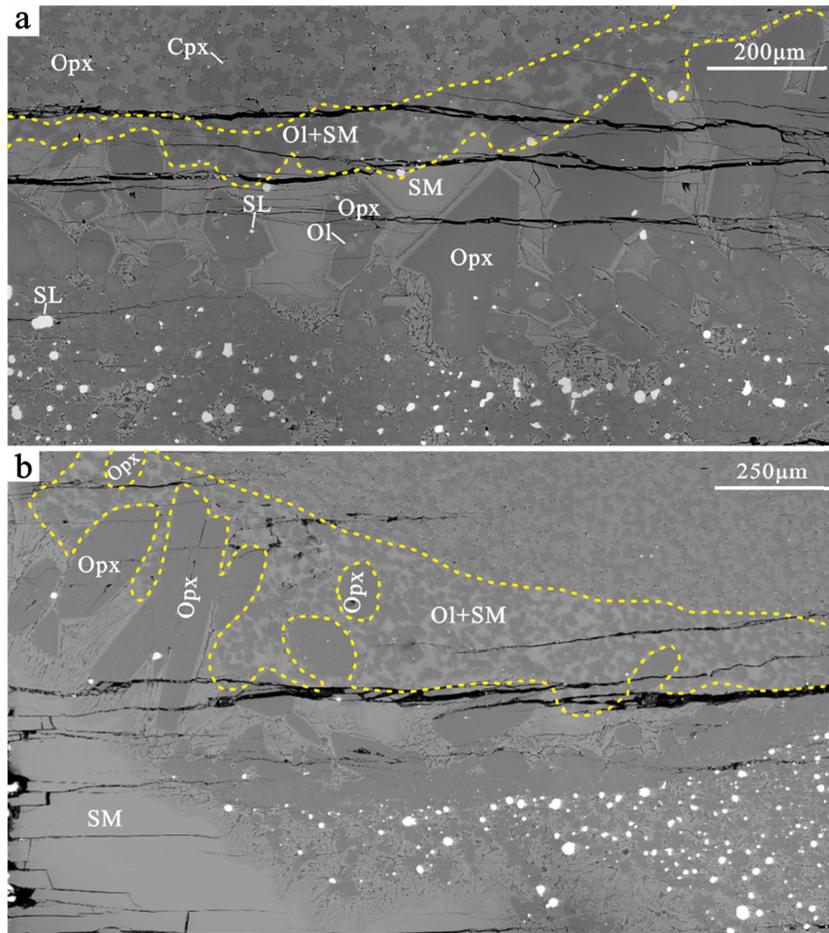


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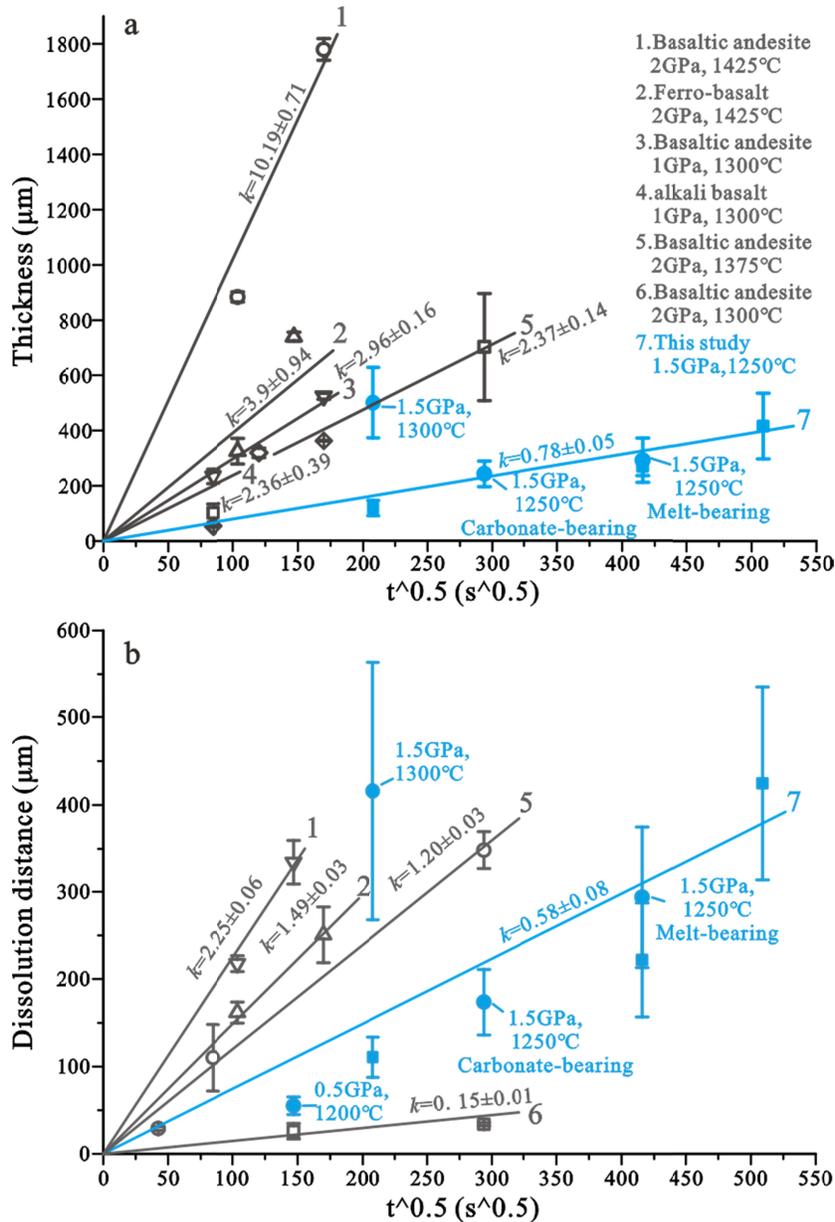
203 **Figure 1.** Back-scattered electron images (BSE) of experimental products from experiment  
 204 PC520 (a) (annealing time ~ 12h) and PC545 (b) (annealing time ~ 72h) under the conditions of  
 205 1.5 GPa and 1250 °C. Red and blue dotted lines respectively denote the original and current  
 206 interfaces between the melt source and peridotite before and after the melt-rock reaction,

207 marking the Opx-rich reaction layer due to the melt-rock reaction. The region between the melt  
208 source and partially molten peridotite of experiment PC545 is exaggerated to observe the  
209 microstructure of the melt-rock reaction interface in Figure 2a. The specimen setup of these  
210 experiments is shown as the inset in the left lower part. Mineral abbreviations: Opx-  
211 orthopyroxene; SM- silicate melt; SL-sulfide liquid.

212 On the other hand, peridotites in the upper part of sample were in disequilibrium with  
213 silicate melt at these run conditions, thereby should be dissolved essentially into the melt. The  
214 amount of dissolution can be indirectly quantified by the dissolution distance (Figure 3b), which  
215 is the distance between the current interface (Figure 1-blue dotted lines) and the original  
216 interface (Figure 1-red dotted lines) that is represented by a sharp decrease in the size of sulfide  
217 droplets. All these measurements were conducted around the central part of each experimental  
218 charge at least 5 times. Similarly, dissolution distance increases linearly with the square root of  
219 time (Figure 3b-blue line) at 1250 °C, and the increase of temperature to 1300 °C significantly  
220 enlarges the dissolution distance (Figure 3b). The addition of oceanic sedimentary (PC560)  
221 and/or basalt (PC548) has no substantial effect on the dissolution distance (Figure 3b).



223 **Figure 2.** Microstructures of the interface of melt-rock reaction in experiment PC545 (1250 °C)  
 224 (a) and PC527 (1300 °C) (b). Yellow dotted lines show the olivine-melt layer consisting of  
 225 olivine, sulfide, and silicate melt above the Opx-rich reaction layer. Mineral abbreviations: Ol-  
 226 olivine. Note that some huge Opx grains are present in the Opx-rich reaction layer of experiment  
 227 PC527 with high reaction temperature, leading to the destruction of the layer.

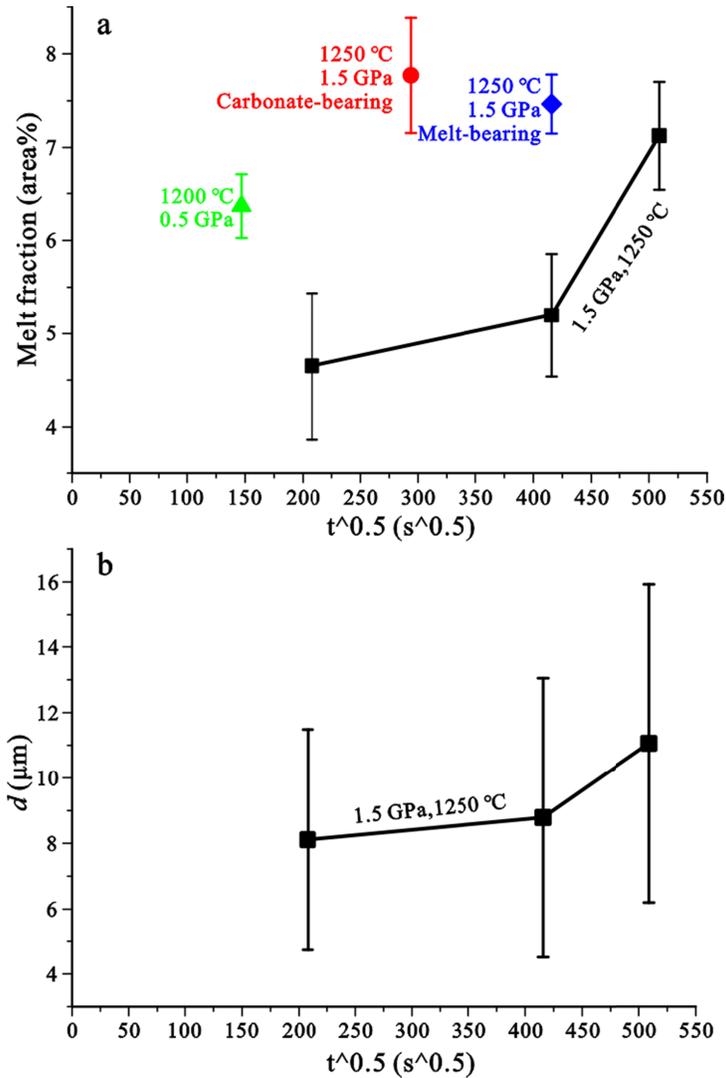


228

229 **Figure 3.** Plots of thicknesses of Opx-rich reaction layers (a) and dissolution distance (b) as a  
 230 function of the square root of run time for the melt-rock reaction experiments. The lines are the

231 best fit for the data. The data denoted as 3 and 4 are from Morgan & Liang (2005), and other data  
232 are from C. Wang et al. (2020).

233         Moreover, based on analyses of SEM images, with increasing run duration or  
234 temperature, more silicate melt penetrated the Opx reaction layer into the partially molten  
235 peridotite ( $\sim 4.65 \pm 0.78$  area% at 12h;  $\sim 7.13 \pm 0.58$  area% at 72h), resulting in the formations  
236 of melt junctions and channels among silicate minerals, and even some large melt pools just  
237 above the ORL (Figure 2 and Figure 4a). At 1.5 GPa, with increasing annealing time from 12h to  
238 72h, the grain size of silicate minerals increases from  $8.1 \pm 3.4$   $\mu\text{m}$  to  $11.0 \pm 4.9$   $\mu\text{m}$  in the  
239 partially molten peridotite (Figure 4b; Supporting information; Table S1). Under conditions of  
240 high temperature (1300 °C) and long-run duration (72h), an olivine-melt layer, consisting of  
241 sulfide, olivine, and silicate melt, is present above the ORL (Figure 2), which may be attributed  
242 to the dissolution of clinopyroxene and reprecipitation of olivine during the melt-peridotite  
243 reaction. Additionally, crystal faces between olivine grains were open and full of silicate melt,  
244 forming abundant melt channels among the crystal framework (Figure 2).

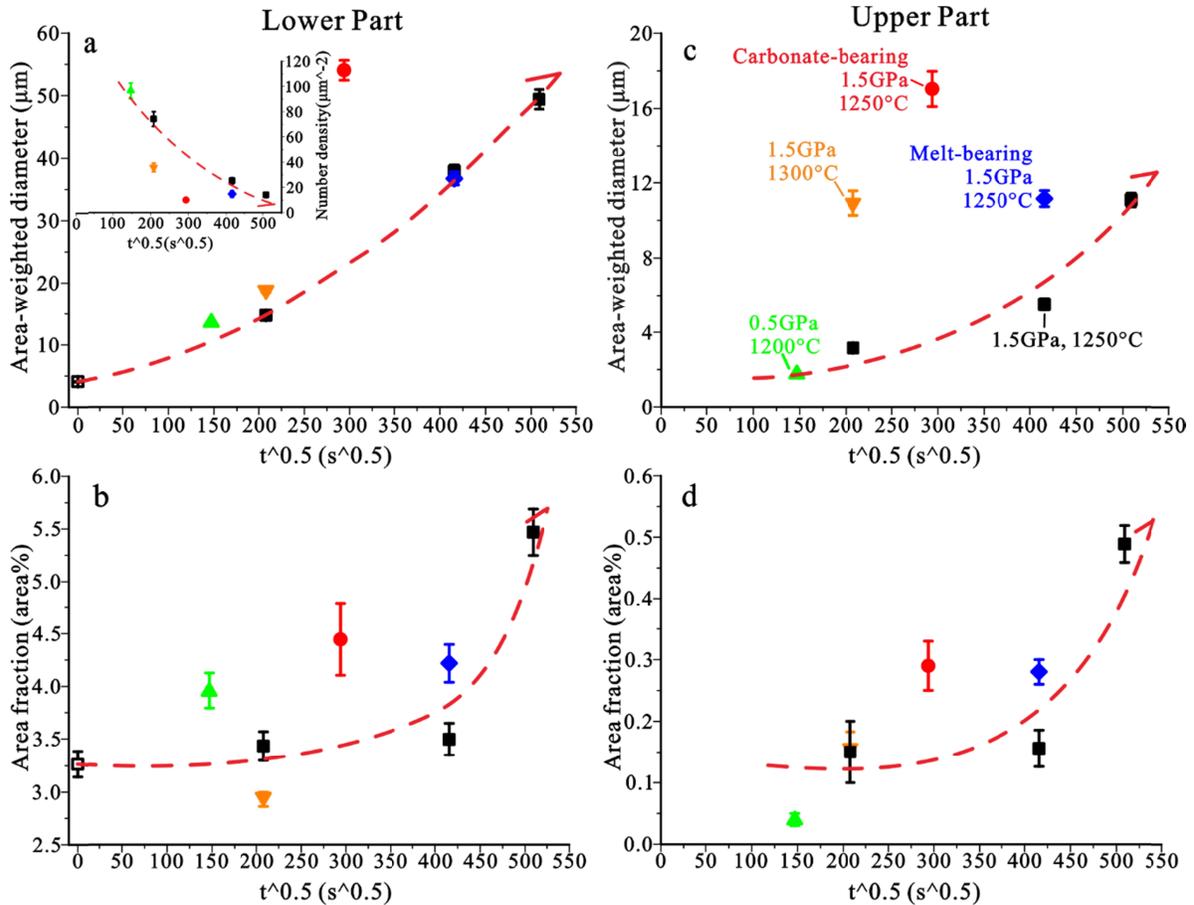


245

246 **Figure 4.** Plots of melt fraction (a) and grain size of silicate minerals (b) as a function of the  
 247 square root of run time in the partially molten peridotite from these experiments of this study.  
 248 Curves are drawn only to guide the eye.

249 In the melt source, the mean area-weighted diameter of sulfide droplets was  $\sim 4.1 \pm 0.1$   
 250  $\mu\text{m}$  in a low-temperature hot-press experiment (PC537, 800 °C), which potentially denotes the  
 251 initial size of sulfide droplets in the starting material. At a higher temperature (1250 °C), the  
 252 sulfide size increases from  $\sim 14.7 \pm 0.2 \mu\text{m}$  to  $\sim 49.4 \pm 1.6 \mu\text{m}$  with increasing the annealing time  
 253 from 12h to 72h (Figure 5a; Table S2), in which the larger standard deviations from the long  
 254 annealing time (72h) experiments may be partly because of the presence of some enormous ( $> 50$   
 255  $\mu\text{m}$ ) and tiny ( $< 1 \mu\text{m}$ ) sulfide droplets. Under the same annealing time (48h), there is no

256 conspicuous increase in sulfide size when 5 wt.% basalt was added to the peridotite (PC548),  
257 while a slight increase was observed if the experimental temperature increased to 1300 °C  
258 (PC527) (Figure 5a). The addition of carbonate into the melt source (PC560) resulted in a  
259 significant increase in sulfide size to  $\sim 54.0 \pm 1.6 \mu\text{m}$  at 24h (Figure 5a). Additionally, we also  
260 observed that the number density of sulfide droplets uniformly decreases with increasing  
261 annealing time in the lower melt source (Figure 5a-inset). The area fraction of sulfide liquid in  
262 the low-temperature experiment (800 °C) was  $\sim 3.26 \pm 0.12 \text{ area}\%$  of the melt source. At  
263 1250°C, the area fraction was essentially constant ( $3.44 \pm 0.14 \text{ area}\%$  at 12h;  $3.50 \pm 0.15 \text{ area}\%$   
264 at 48h) when the annealing time was less than 48h, but a visible increase ( $5.46 \pm 0.22 \text{ area}\%$ ) can  
265 be observed for the experiments with annealing time of 72h (Figure 5b). A similar trend was also  
266 shown in the upper part of samples (Figure 5d), which may be due to more silicate melts from  
267 the melt source infiltrating into the upper part of samples with increasing annealing time to 72h  
268 (Figure 4). As previously observed (Yoshino & Watson, 2005), the diffuse addition of Fe and Ni  
269 components from silicate phases into sulfide liquids may be one of the reasons for the increase of  
270 the sulfide area fraction in the melt source. In contrast, with increasing temperature to 1300 °C, a  
271 lower area fraction ( $2.93 \pm 0.07 \text{ area}\%$ ) of sulfide was observed even in the experiment with a  
272 short annealing time (12h) (Figure 5b), potentially due to partial dissolution of sulfide droplet  
273 driven by the higher sulfur content at sulfide saturation (SCSS) at a higher temperature (*e.g.*, Liu  
274 et al., 2007; Mavrogenes & O'Neill, 1999). Similarly, the additions of basalt and carbonate  
275 enhance the permeation of more silicate melt into the upper part of samples, thereby increasing  
276 the area fraction of sulfide in the melt source (Figure 5b).



277

278 **Figure 5.** Plots of grain size ( $\mu\text{m}$ ) of sulfide droplets weighted by the area (a and c) and area  
 279 fraction of sulfide droplets (b and d) as a function of the square root of run time in the melt  
 280 source (lower part) and the partially molten peridotite (upper part) from these experiments of this  
 281 study. Inset in (a) shows the relationship between the number density of sulfide droplets in the  
 282 melt source and the square root of run time. Curves are drawn only to guide the eye.

283 Based on the SEM images of products from these high-temperature experiments (1250  
 284 °C), it becomes evident that sulfide droplets in the melt source were entrained into the partially  
 285 molten peridotite by the porous flow of silicate melt during the melt-peridotite reaction (Figure  
 286 1b and Figure 2). With permeating more silicate melt upwards, more and larger sulfide droplets  
 287 were observed in the partially molten peridotite, and meanwhile their sizes (area-weighted  
 288 diameter  $\sim 3.1 \pm 0.1 \mu\text{m}$  at 12h to  $\sim 11.1 \pm 0.4 \mu\text{m}$  at 72h) and area fractions ( $\sim 0.15 \pm 0.05$   
 289 area% at 12h to  $\sim 0.49 \pm 0.03$  area% at 72h) increase with the increasing annealing time at 1250  
 290 °C (Figure 5c and d; Table S2). Apparently, the addition of carbonate and 5 wt.% basalt into the  
 291 melt source and partially molten peridotite, respectively, drives a higher area fraction and larger

292 size of sulfide droplet in the upper part of samples (Figure 5c and d). Hence, these observations  
293 propose that the ORL may not efficiently prevents the upward transport of silicate melt and  
294 sulfide droplets from the melt source to the upper molten peridotite.

295 Notably, under the condition of lower pressure (0.5 GPa), some vapor bubbles have been  
296 found in the partially molten peridotite due to the exsolution of volatile (mainly H<sub>2</sub>O) in the  
297 starting basaltic material, and they mostly absorb on sulfide droplets to form compound drops,  
298 which has been proposed to potentially enhance the upward transport of sulfide (Mungall et al.,  
299 2015; Yao & Mungall, 2020).

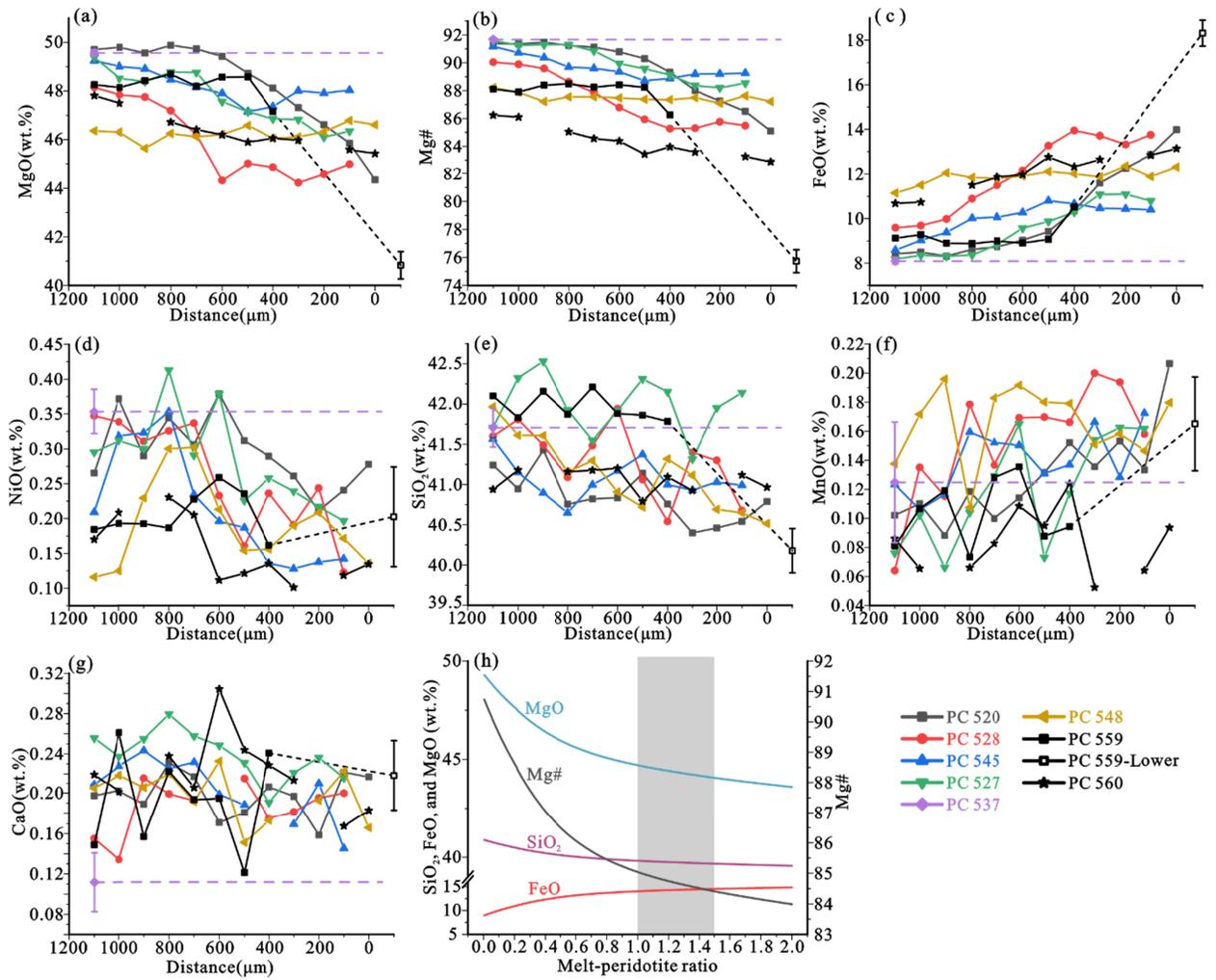
### 300 3.2 Phase compositions

#### 301 3.2.1 Mineral compositions

302 Figure 6 and Figure 7 respectively showed the compositional variations of olivine and  
303 clinopyroxene as a function of distance away from the final melt-peridotite interface. Under  
304 conditions of 1.5 GPa and 1250-1300 °C, from the far-field region to the interface, olivine grains  
305 became gradually lower in the Mg# (defined as molar Mg/(Mg+Fe)\*100) and concentrations of  
306 SiO<sub>2</sub>, MgO and NiO, and meanwhile had increased in FeO and MnO contents (Figure 6).  
307 Compared with the low-temperature hot-press experiment (PC537) (Figure 6 purple dotted  
308 lines), olivine grains from high-temperature experiments contained lower MgO and NiO, and  
309 higher FeO and CaO concentrations. By contrast, olivine compositions in the experiment PC548  
310 (5 wt.% basalts) and PC559 (0.5 GPa) were roughly constant across the partially molten  
311 peridotite, whereas olivine grains crystallized in the melt source at lower-pressure (0.5 GPa,  
312 PC559) experiment had lower Mg#, MgO, SiO<sub>2</sub> and higher FeO than those in the peridotite  
313 region (Figure 6).

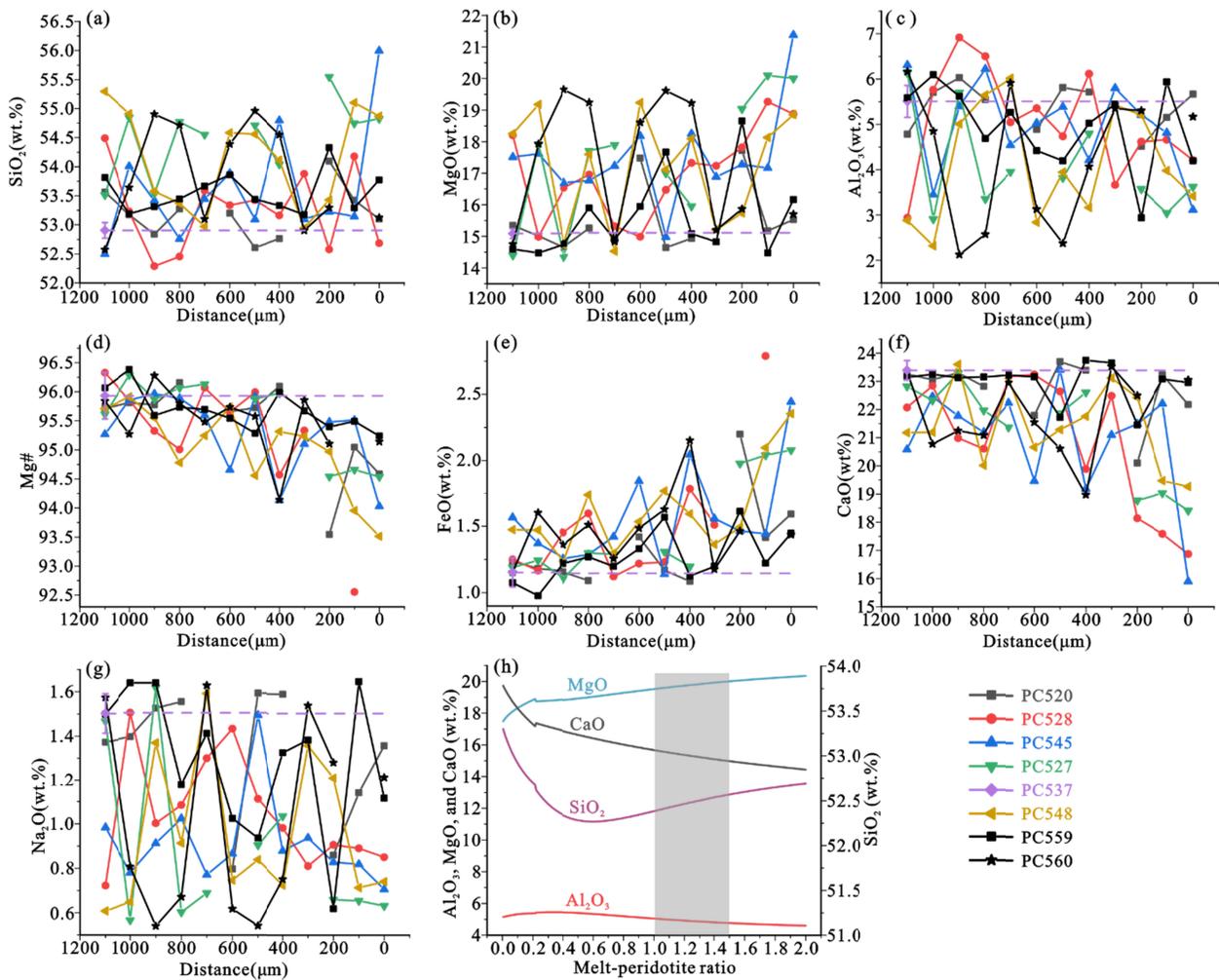
314 In contrast to olivine, more scatter compositions were observed in the clinopyroxene  
315 (Cpx) grains from the partially molten peridotite (Figure 7). Only near the melt-rock interface,  
316 the trends of decreasing Mg# and CaO and increasing FeO were present in Cpx from all high-  
317 pressure experiments, whereas the Cpx compositions in the low-pressure and carbonate-bearing  
318 experiment (PC559 and PC560) were essentially constant across the peridotite region (Figure 7).  
319 On the other hand, the Opx grains in the reaction layer and melt source were the products of

320 silicate melt-olivine reaction. No distinct compositional difference was observed in these Opx  
 321 grains.



322

323 **Figure 6.** Plots (a-g) of measured oxide abundance (in wt.%) and Mg# in olivine as a function of  
 324 distance (in  $\mu\text{m}$ ) away from the melt-rock interface. The composition variations of MgO, Mg#,  
 325 FeO, and  $\text{SiO}_2$  are also simulated thermodynamically as a function of melt-peridotite ratio in (h).  
 326 When the melt-rock ratio is between 1 and 1.5, these measured compositions are roughly  
 327 consistent with those results simulated thermodynamically (the gray shadow region in h). At low  
 328 pressure (PC559, 0.5 GPa), the compositions of olivine recrystallized in the melt source during  
 329 the melt-rock reaction are present as PC559-Lower (black hollow square). The composition of  
 330 olivine grains in the low-temperature hot-press experiment (PC537) was almost constant across  
 331 the partially molten peridotite region and thus can be used as the reference line (purple dotted  
 332 lines) for the composition variations of olivine in those high-temperature reaction experiments.



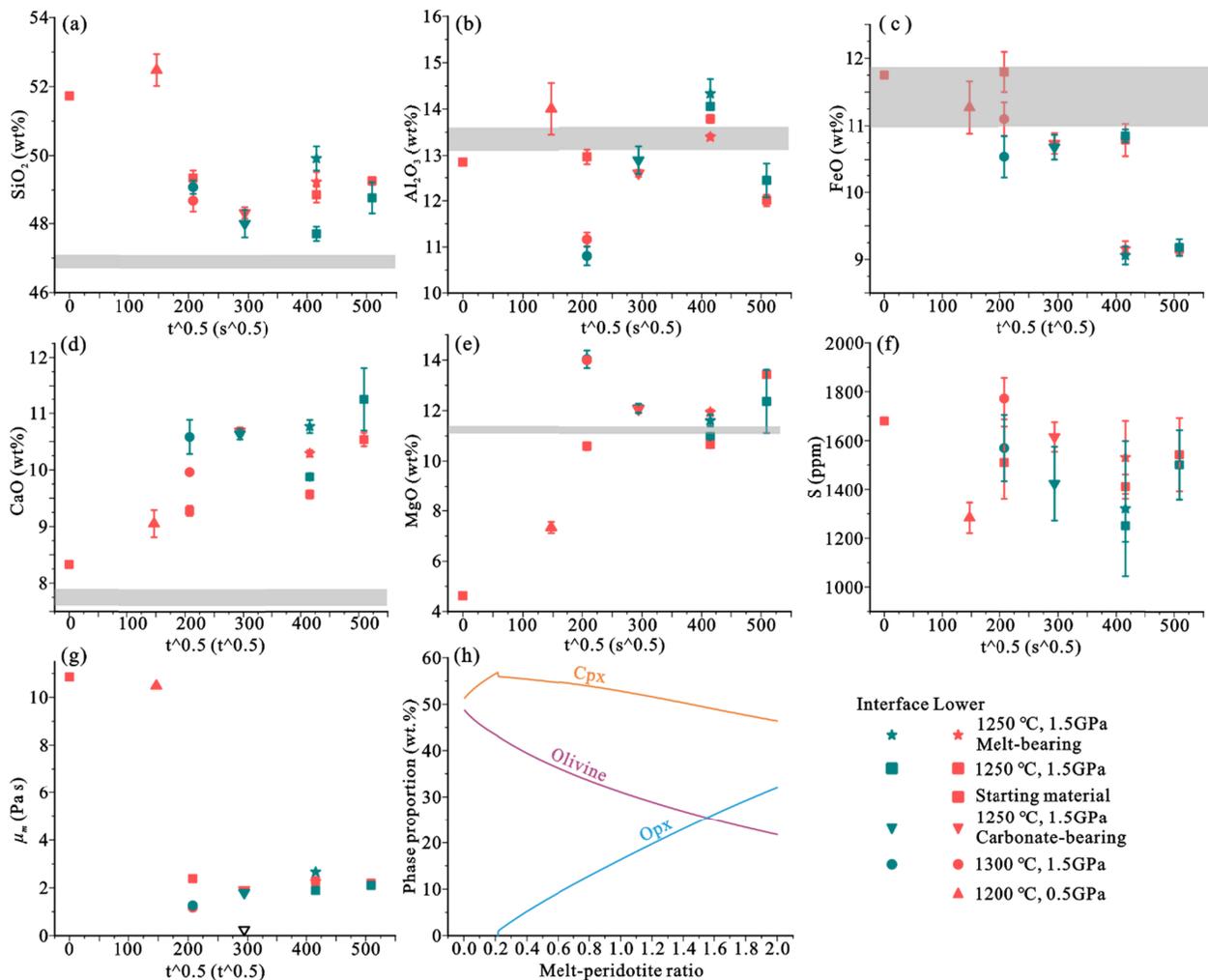
333

334 **Figure 7.** Plots (a-g) of measured oxide abundance (in wt.%) and Mg# in clinopyroxene as a  
 335 function of distance (in  $\mu\text{m}$ ) away from the melt-rock interface. The composition variations of

336 MgO, Al<sub>2</sub>O<sub>3</sub>, CaO, and SiO<sub>2</sub> are also simulated thermodynamically as a function of the melt-  
337 peridotite ratio in (h). The legends are the same as in Figure 6.

### 338 3.2.2 Silicate melt and sulfide liquid compositions

339 The evolution of the reacted melt composition in these layered experiments mostly  
340 depends on the extent of melt-rock reaction and the major phase formed by the reaction. In the  
341 melt source, with increasing run time and/or temperature, the compositions of silicate melt  
342 became higher in CaO and MgO concentrations, and lower in SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, and FeO (Figure 8),  
343 implying the effect of high-degree melt-peridotite reaction on the melt composition. No distinct  
344 variation of the melt sulfur contents was observed under the conditions of 1250 °C and 1.5 GPa  
345 (Figure 8f), potentially indicative of sulfur-saturated silicate melt in these experiments. The  
346 decrease of pressure and temperature respectively to 0.5 GPa and 1200 °C caused the obvious  
347 increases in SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> concentrations and a slight decrease of S concentration in silicate  
348 melt. In addition, no significant variations of melt compositions were observed between the melt  
349 source and melt-peridotite interface within the error of measurement (Figure 8), suggesting the  
350 convective flow is strong enough to drive the chemical equilibrium of silicate melt. The  
351 composition of primary silicate melt infiltrating into the peridotite cannot be analyzed due to the  
352 small scale. In addition,  $\sim 1.57 \pm 0.15$  wt.% H<sub>2</sub>O was detected in the reacted melts from  
353 experiment PC520 and PC528 using the Fourier Transform Infrared Microscopy (Supporting  
354 Information, Mercier et al., 2010), which is slightly higher than that of basalt ( $\sim 1.19$  wt.%)  
355 added in the starting materials.



356

357 **Figure 8.** Plots of oxide abundance (in wt.%) (a-e), sulfur content (f), and viscosity ( $\mu_m$ ) (g) of  
 358 silicate melt versus run time in melt-rock interface and lower melt source of these reaction  
 359 experiments. The ranges of composition variations of SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, CaO, and MgO  
 360 simulated thermodynamically are denoted as gray shadow regions in (a)-(e) when the melt-  
 361 peridotite ratio varies from 1 to 1.5. Produced phase proportions of different silicate minerals  
 362 during the melt-rock reaction are also simulated thermodynamically as a function of melt-  
 363 peridotite ratio in (h). A lower melt viscosity ( $\sim 0.25$  Pa s) of carbonate-bearing experiment  
 364 (PC560) is roughly estimated using a simple model from Di Genova et al. (2014) and denoted as  
 365 a hollow inverted triangle in (g). Cpx-clinopyroxene, Opx-orthopyroxene.

366 Except for several large sulfide droplets containing the Ni-rich quenched phases, the  
 367 composition of sulfide was homogeneous in the lower melt source of all experiments. In the 12h  
 368 experiment, the S content, Ni/S and Fe/S ratios of sulfide liquid were  $\sim 35.16 \pm 0.72$  wt.%,  $\sim$

369  $0.39 \pm 0.03$ , and  $\sim 1.43 \pm 0.04$ , respectively, and these values ( $S \sim 35.74 \pm 0.74$  wt.%,  $Ni/S \sim$   
 370  $0.37 \pm 0.09$ ,  $Fe/S \sim 1.42 \pm 0.06$ ) remained constant within the error of measurement when the  
 371 annealing time increases to 72h. However, the Cu/S ratio of sulfide liquid slightly decreased  
 372 from  $0.017 \pm 0.004$  at the 12h to  $0.009 \pm 0.002$  at the 72h at 1250 °C, while an obvious increase  
 373 in the Cu/S ratio ( $\sim 0.046 \pm 0.010$ ) was observed at 1300 °C.

## 374 **4 Discussion**

### 375 **4.1 Grain-scale processes in melt-peridotite reaction**

376 Firstly, based on the thermodynamic model via the pMELTS (Ghiorso et al., 2002), if the  
 377 upper peridotite reaches equilibrium at 1.5 GPa and 1250-1300 °C, the corresponding melt  
 378 fraction is less than 0.04-0.08 wt.%, implying that the self-partial melting of peridotite at the  
 379 experimental conditions here is negligible. Therefore, the observed variations of melt area  
 380 fraction, phase proportions, and minerals' sizes and compositions in the experimental products  
 381 are likely to be due primarily to the melt-peridotite reaction. On the other hand, the upper  
 382 peridotites are not in equilibrium with the starting lower melts at the experimental conditions,  
 383 and hence olivine and clinopyroxene in the peridotite part will partially or completely dissolve in  
 384 the reacting and upward percolating melt (Liang, 1999; C. Wang et al., 2020). In  
 385 multicomponent partially molten systems, the melt-peridotite reaction is widely suggested to  
 386 occur via multi-scale mass transfer processes that inevitably involve both the grain-scale  
 387 dissolution-precipitation-reprecipitation and diffusion, in conjunction with large-scale advective  
 388 transport (Cascio et al., 2008; Morgan & Liang, 2005).

389 In our high-temperature, high-pressure experiments (1.5 GPa and 1250-1300 °C) here,  
 390 due to the disequilibrium between melt and peridotite, the consequent grain-scale processes can  
 391 mostly occur via the dissolution, precipitation, and reprecipitation of mineral grains, which is  
 392 evident from the formation of ORL. In the lower melt source, the basaltic melt in starting  
 393 materials is olivine-undersaturated, and thus the melt-rock reaction between basaltic melt and  
 394 olivine produces new Opx grains (Figure 1), which is consistent with the following reaction:



396 where subscripts 0 and 1 designate melt and mineral grains from starting materials and the  
 397 reaction products, respectively. This reaction should occur via the dissolution of olivine, and

398 precipitation of new orthopyroxene, largely contributing to the formation of ORL in the interface  
399 between the peridotite and melt source (Figure 1a). It is well known that dissolution and  
400 precipitation occur simultaneously during the development of the ORL under high pressure ( $> 1$   
401 GPa) (Cascio et al., 2008; Morgan & Liang, 2003). With increasing the degree of melt-peridotite  
402 reaction (a longer run duration and/or higher temperature), the olivine-melt layer above the ORL  
403 is formed by the further dissolution of olivine and few clinopyroxene grains, with concomitant  
404 reprecipitation of fresh olivine grains during melt reaction flow (Figure 2) (Kelemen et al.,  
405 1995).

406 On the other hand, the grain size of silicate minerals is a vital role in affecting the  
407 physical properties of upper partially molten peridotite, such as permeability (Faul, 2001). Here,  
408 the growth of silicate mineral sizes in the upper partially molten peridotite with increasing  
409 annealing time (Figure 4b) is mainly driven by a reduction of grain boundary energy (Faul &  
410 Scott, 2006) and demonstrates that textural coarsening by which small size grains are consumed  
411 as supply material for the growth of coarser grains (Higgins, 1998, 2011; Yao et al., 2017), may  
412 overweigh the decreasing-size effect driven by the reprecipitation of new, fine-grained crystals.  
413 The transference of material from dissolved olivine to other grains can occur via diffusion  
414 through crystals themselves, grain boundaries, and the interstitial melt (Yao et al., 2017), while  
415 the last one is much faster and becomes the main pathway of these grain-scales processes.  
416 Finally, the large-scale advective transport of the reacting melt is also evident from the variations  
417 of melt area fraction in the partially molten peridotite and the formations of melt junctions and  
418 channels (Figure 1; Figure 4a).

419 Based on the parabolic law of diffusive dissolution (Liang, 1999; Zhang et al., 1989), the  
420 slope (*i.e.*, the diffusive dissolution constant) of the fitted linear regression line for our  
421 experiments is used to estimate the growth rate of ORL ( $k_{ORL}$ ), which is about  $\sim 0.78 \pm 0.05$   
422  $\mu\text{m/s}^{0.5}$  here at 1.5 GPa and 1250 °C and far lower than those of some previous experiments  
423 conducted under higher temperatures and/or pressures (Figure 3a) (Morgan & Liang, 2005; C.  
424 Wang et al., 2020). However, the thickness of ORL in experiment PC527 at 1.5 GPa and 1300  
425 °C has a faster-increasing tendency ( $k_{ORL} = 2.37 \pm 0.14 \mu\text{m/s}^{0.5}$ ), consistent with the experiments  
426 under 2 GPa and 1375 °C (C. Wang et al., 2020). This tendency of ORL growth is likely  
427 unchanged in the experiments with the carbonate-bearing melt source (PC560) and silicate melt-  
428 bearing peridotite (PC548) (Figure 3a), implying an insignificant effect of the melt composition

429 and the porosity on the growth of ORL in these experiments. These results indicate that the  
430 increases in temperature and pressure will potentially enhance the growth rate of the ORL  
431 thickness during the melt-rock reaction, which can be mostly attributed to the higher diffusivities  
432 of elements in silicate melt at a higher temperature (Mungall, 2002; Zhang et al., 2010).

433 Similarly, we observed the linear increase of dissolution distance with the square root of  
434 time (Figure 3b), suggesting that a substantial amount of peridotite was dissolved in the  
435 percolating melt. At 1.5 GPa and 1250 °C, the dissolution rate of peridotite evaluated by the  
436 slope of the linear regression line ( $k_{diss} = 0.58 \pm 0.08 \mu\text{m/s}^{0.5}$ ) is lower than those in previous  
437 experiments with higher temperatures and pressures (Figure 3b, C. Wang et al., 2013, 2020),  
438 indicating that relatively high temperature and pressure could also increase the dissolution rate.  
439 Moreover, the higher dissolution rate of peridotite at experiment PC527 (1300 °C) implies that  
440 the increase in temperature significantly enhances the melt-rock reaction (Mitchell & Grove,  
441 2016). The high content of water ( $\sim 1.57 \pm 0.15 \text{ wt.}\%$ ) in silicate melt could also enhance the  
442 formation of several enormous Opx grains in the ORL (Figure 2) (Wang et al., 2016), partly  
443 because the addition of water can strongly depress the peridotite solidus. This may be also the  
444 reason for the large standard deviations of the dissolution rate in experiment PC527 and PC545  
445 with high temperature (1300 °C) and long annealing time (72h). Overall, the growth rate of the  
446 ORL is generally higher than the peridotite dissolution rate, which is partly due to the growth of  
447 ORL not only involves the precipitation of some new mineral grains in the process of melt-  
448 enhanced dissolution of peridotite, but also is produced by the volume-increasing reaction during  
449 the replacement of olivine by orthopyroxene (Milke et al., 2009).

#### 450 4.2 Compositional variations of melt and minerals

451 During the melt-peridotite reaction, besides the systematic changes in mineralogy and  
452 texture of peridotite, we also observed the compositional variations of melt and minerals in the  
453 reaction couple (Figure 6-8), which have been widely used to outline the grain-scale processes  
454 governing this melt-peridotite reaction (Mallik & Dasgupta, 2012; C. Wang et al., 2020). On the  
455 other hand, a thermodynamically-constrained mixing model has been recently adopted to  
456 examine the variations of major element compositions during the melt-peridotite interaction  
457 (Lambart et al., 2012; Pin et al., 2022; Shaw et al., 2018), and this forward model may offer a  
458 key to testing and understanding the compositional evolution of our experimental products.

459 In this regard, the melt-peridotite reaction is modeled as a simplified, thermodynamic  
460 process in which the peridotite is continually impregnated by a finite amount of basaltic melt  
461 from the lower melt source, which is the same as the assumption in previous works (Lambart et  
462 al., 2012; Pin et al., 2022; Shaw et al., 2018). Once each increment of melt impregnation, the  
463 infiltrated melt will eventually be equilibrated with the surrounding peridotite, and the  
464 thermodynamic properties of the whole system should be adopted as a reference for the next  
465 increment. In each increment, the proportions and compositions of melt and solid phases after the  
466 chemical re-equilibrium can be modeled by minimizing the Gibbs energy of the whole system  
467 from constraints on bulk composition, temperature, pressure, enthalpy, and oxygen fugacity (Pin  
468 et al., 2022; Yao et al., 2018). This process is simulated by constantly adding up to 200 g of the  
469 lower basaltic melt by increments of 0.5 g, to 100 g of the upper peridotite, using the pMELTS  
470 of alphaMELTS (Ghiorso et al., 2002; Ghiorso & Sack, 1995; Smith & Asimow, 2005) in  
471 isenthalpic mode at 1.5 GPa, 1250-1300 °C and  $\Delta Q_{FM} = 2.0$ . Here, the incremental addition of  
472 basaltic melt corresponds to the increase of melt: peridotite ratio from 0.005 to 2.

473 Our simulation shows that the mass fraction of olivine decreases from ~ 49 wt.% to ~  
474 21.8 wt.% with the increasing melt: peridotite ratio from 0.005 to 2 (Figure 8h). The silicate melt  
475 becomes saturated with orthopyroxene when the melt: peridotite ratio reaches a value of ~ 0.22.  
476 After this point, the mass fraction of new orthopyroxene precipitated from silicate melt quickly  
477 increases to ~ 32 wt.% at the melt: peridotite ratio of 2 (Figure 8h). In contrast to the initial  
478 peridotite, the mass fraction of clinopyroxene has a quick increase from ~ 51 wt.% to ~ 57 wt.%  
479 before the precipitation of orthopyroxene, but then slowly decreases to ~ 46 wt.% at melt:  
480 peridotite ratio = 2 (Figure 8h). Hence, the increase in orthopyroxene proportion is mostly due to  
481 the dissolution of olivine, accompanied by a limited contribution from the consumption of no  
482 more than ~ 5 wt.% clinopyroxene, which approximately coincides with the hypothetical  
483 Reaction 1 mentioned above.

484 Concurrently, the melt penetrates upwards and further reacts with the peridotite to result  
485 in systematic variations of the compositions of olivine and Cpx grains towards the melt-rock  
486 interface (Figure 6 and Figure 7). The modeled compositional evolutions of olivine show  
487 decreasing trends of MgO and SiO<sub>2</sub> contents respectively from ~ 49.3 wt.% to ~ 44.1 wt.% and  
488 40.9 wt.% to 39.7 wt.%, an increase of FeO content from ~ 9.0 to ~ 14.5 wt.%, and the  
489 associated decrease of Mg# from ~ 90.7 to ~ 84.4, when the melt: peridotite ratio increases from

490 ~ 0.005 to ~ 1.5 (Figure 6h). In the same range of melt: peridotite ratio, the clinopyroxene has an  
491 obvious increase in its MgO content (from ~ 17.5 to ~ 20.0 wt.%) and approximately constant  
492 SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> contents (Figure 7h). The trends of these compositional variations of minerals  
493 are nearly in agreement with those of our experimental measurements, except for the low-  
494 pressure experiment (PC 559, Figure 6h and Figure 7h). Therefore, the measured compositional  
495 variations of olivine and clinopyroxene as a function of distance away from the final melt-  
496 peridotite interface may be referred to the melt: peridotite ratio at any position of the upper  
497 peridotite. The measured compositions of olivine and Cpx grains just above the final melt-  
498 peridotite interface are approximately consistent with those results simulated thermodynamically  
499 (Figure 6h and 7h gray shadow regions) in the range of melt-peridotite ratio of 1.0 to 1.5, which  
500 suggests that they have obtained locally chemical re-equilibrium with the reacting melt that is  
501 about ~ 1.0-1.5 times the initial mass of peridotite in the same region. As the distance away from  
502 the final melt-peridotite interface grows, the increasing MgO and Mg# of olivine reflect the  
503 decrease of melt: peridotite ratio, and consequently imply the gradual weakening of melt-  
504 peridotite reaction. Hence, when the upward-flowing melt encounters the partially molten  
505 peridotite that initially has a low permeability, strong convection and backflow of silicate melt  
506 and the associated high-degree melt-peridotite reaction mostly occur at the bottom of peridotite  
507 and thus lead to an increase of porosity of the reaction region and a larger buoyancy-driven  
508 upward motivation of silicate melt, which produces positive feedback between the reaction and  
509 melt flow contributing to the upward infiltration of more silicate melt into the partially molten  
510 peridotite.

511         Along with the increasing melt: peridotite ratio from ~ 1.0 to ~ 1.5, the modeled silicate  
512 melt that is reacted with the peridotite shows a decrease of SiO<sub>2</sub> content from ~ 47.1 wt.% to ~  
513 46.7 wt.%, which is similar to the measured ranges in our experiments with long annealing time  
514 but has a slightly lower SiO<sub>2</sub> content at high melt-peridotite ratio (Figure 8a). The Al<sub>2</sub>O<sub>3</sub> content  
515 of silicate melt also varies from ~ 13.6 wt.% to ~ 13.1 wt.%, while the MgO content of melt  
516 shows a narrow variation range of ~11.1 to ~11.4 wt.%, which both roughly match those results  
517 measured in this study (Figure 8b and e). The possible iron loss to Pt capsule in these  
518 experiments due to the incomplete isolation of separate upper and lower graphite capsules with  
519 different diameters for silicate melt from Pt capsule (J. Wang et al., 2020) is likely to be the main  
520 reason for the lower FeO content (Figure 8c) measured in those experiments with long annealing

521 time (> 48h). The model suggests that silicate melt contains ~ 7.6-7.9 wt.% CaO at the melt-  
522 peridotite of 1.0-1.5, but this range is largely underestimated in contrast to our measurements  
523 (Figure 8d), partly because the slight differences in the Gibbs free energy among various  
524 compositional models of pyroxenes (Yao et al., 2021) may drive large errors in the CaO content  
525 of clinopyroxene in the AlphaMELTS. Although there still have some weaknesses in modeling  
526 the compositional evolution of silicate melt, the use of AlphaMELTS is encouraging here, and  
527 helps to examine how mineral model and chemical composition evolve as silicate melts are  
528 added into the peridotite part.

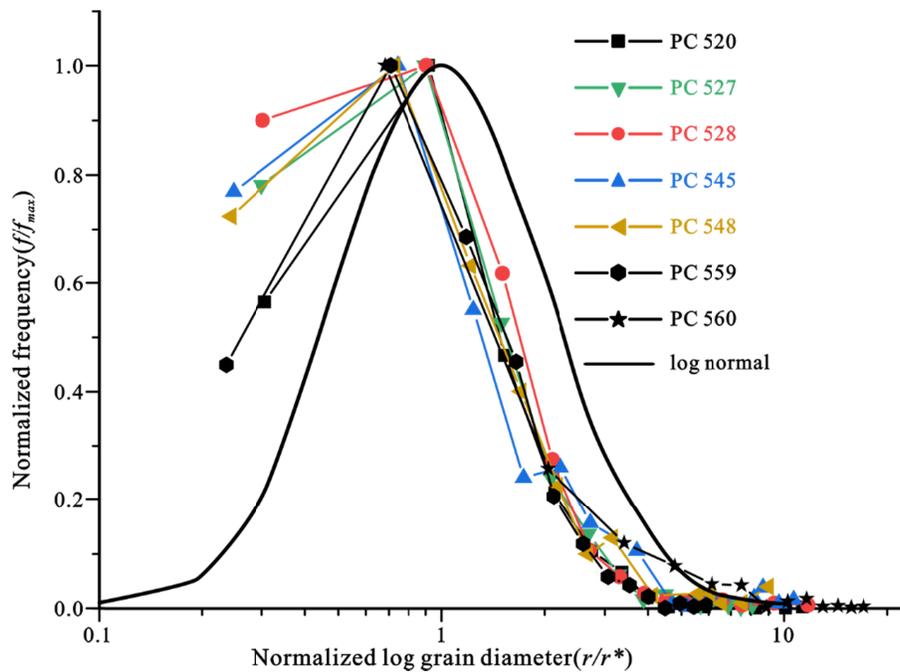
### 529 **4.3 Coarsening processes of sulfide droplets**

530 In a partially molten system, the nucleation/growth rate and associated size distribution of  
531 sulfide droplets are vital constraints on their dynamics (Yoshino & Watson, 2005). Along with  
532 the increasing annealing time, the size of sulfide droplets shows a larger increase (from  $\sim 3.2 \pm$   
533  $0.1 \mu\text{m}$  to  $\sim 11.1 \pm 0.4 \mu\text{m}$ ) (Figure 5c) compared with that (from  $\sim 8.1 \pm 3.4 \mu\text{m}$  to  $\sim 11.0 \pm 4.9$   
534  $\mu\text{m}$ ) (Figure 4b) of the surrounding silicate minerals. The growth of silicate minerals has been  
535 suggested to be mostly controlled by the melt-enhanced diffusion of material, but sulfur that is  
536 the one of major elements in the sulfide has a much lower diffusivity than those of other  
537 elements (Freda et al., 2005; Yao & Mungall, 2021). Given that S diffusion rather than Fe  
538 diffusion limits the growth rate of sulfide droplets (Zhang, 2015), if the growth of droplets  
539 completely occurred via the diffusion of S, it should have a smaller growth rate than that of  
540 silicate minerals, which is opposite to our measurements. Therefore, besides the diffusion-driven  
541 growth, there must be other mechanisms here contributing to the fast growth of sulfide droplets.

542 On the other hand, the actual 3D size distribution of sulfide droplets in the melt source  
543 shows a uniform, nearly log-normal size distribution independent of annealing times with a  
544 skewed peak to a slightly smaller size than the average one (Figure 9), which can be attributed to  
545 the possible contribution of the Lifshitz-Slyozov-Wagner (LSW) coarsening and the occurrence  
546 of normal grain growth (Honour et al., 2019; Yoshino & Watson, 2005).

547 Based on the classical LSW theory (Lifshitz & Slyozov, 1961; Wagner, 1961), the  
548 kinetics of particle growth can be approximately described as  $r \sim t^{1/m}$ , where  $r$  is the grain size,  
549 with the exponent  $m = 3$  for a diffusion-controlled process and  $m = 2$  under the control of an  
550 interface reaction. At 1250 °C and 1.5 GPa, coarsening rates of sulfide droplets are slightly

551 slower than that predicted ( $m = 2$ ) by the interface-reaction-controlled process due to a transient  
552 regime ripening before static-state ripening (Figure 5a), implying the likely combination of  
553 multiple coarsening processes here (Lautze et al., 2011; Sun, 2007). The higher run temperature  
554 (1300 °C) likely contributes to a higher degree of interface-controlled reaction (Yoshino &  
555 Watson, 2005), which can potentially facilitate the coarsening processes of sulfide liquid in the  
556 melt source (Figure 5a). A dramatic decrease in the number density of sulfide droplets (Figure  
557 5a-inset) and the presence of sulfide droplets in contact with each other (Figure 2b and Figure  
558 S1) in the melt source both imply that the contribution of mechanical coalescence of sulfide  
559 droplets (agglomeration and successive coalescence) to their coarsening processes may be more  
560 important at droplet contacts (Yoshino & Watson, 2005). Moreover, silicate melt structure  
561 strongly affects the tendency of sulfide droplets to coalesce, and the coalescence of sulfide  
562 droplets could be enhanced at lower viscous melts (Holzheid, 2010), potentially addressing a  
563 significant increase in the coarsening rate (Figure 5a) in the carbonate-bearing experiment  
564 (PC560) with a lower melt viscosity (Figure 8g) (Di Genova et al., 2014). Although the previous  
565 analog experiment (de Bremond d’Ars et al., 2001) and some static high-temperature and high-  
566 pressure experiments (Bockrath et al., 2004; Holzheid, 2010) did not observe significant  
567 coalescence of sulfide droplets, the mechanical coalescence of sulfide droplets, for the first time,  
568 has been demonstrated to be feasible in our experiments with strong melt convection and melt-  
569 peridotite reaction. This mechanical coalescence process may potentially occur via the impaction  
570 of sulfide droplets with each other, and drainage and rupture of the infinitesimally liquid film  
571 separating them during the reacting melt flow, as suggested by theoretical analyses (Yao et al.,  
572 2019).



573

574 **Figure 9.** Normalized 3D size distributions of sulfide droplets in the melt source. The size of  
 575 sulfide droplets ( $r$ ) and frequency ( $f$ ) are normalized to average size ( $r^*$ ) and maximum  
 576 frequency ( $f_{max}$ ), respectively.

#### 577 4.4 Dynamics of melt flow and entrainment of fine sulfide droplets

578 These dissolution-precipitation processes not only change the composition and  
 579 proportion of solid phases, but also increase the porosity of peridotite in the upper part of  
 580 samples, thereby enhancing the upward mobilities of silicate melt and sulfide droplets away from  
 581 the lower melt source (Figure 4). Moreover, the porosity of ORL is enough high for the  
 582 entrainment of sulfide droplets into peridotite (Figure 1 and Figure 2). In addition, with  
 583 increasing annealing time to 72h, more basaltic melt infiltrating into the peridotite region from  
 584 the melt source (Figure 4) could potentially enhance the reaction between basalt melt and  
 585 peridotite. Melt-rock reaction may produce a high flow capacity of reacting melt through a  
 586 dissolvable peridotite (Chadam et al., 1986; Jackson et al., 2018), resulting in the formation of  
 587 finger- or tree-like melt-rich channels in which a fast-ascending magma flow can even drive the  
 588 antigravitational migration of some coarse olivine grains (Pec et al., 2017). Hence, the  
 589 development of positive feedback between the permeability due to the melt-peridotite reaction  
 590 and the associated increase of melt flux in reacting regions produces a higher flow velocity of

591 melt and wider melt channels among crystal framework (Aharonov, 1995; Pec et al., 2017),  
 592 thereby potentially facilitating the upward entrainment of sulfide droplets (Wang & Jin, 2020).

593 To quantitatively outline the upward physical migration of sulfide droplets along with the  
 594 reactive melt flows in the dissolvable peridotite, some theoretical parameters of physical  
 595 properties of the partially molten rock should be considered first (e.g., von Bagen & Waff,  
 596 1986; Mckenzie, 1984). For an ideal porous crystal framework, the permeability ( $k$ ) can be cast  
 597 by using a simple power-law relationship between grain size ( $d$ ) and porosity ( $\Phi$ ) of the crystal  
 598 matrix (von Bagen & Waff, 1986):

$$599 \quad k = \frac{\Phi^n d^2}{C}, \quad (2)$$

600 where  $C$  and  $n$  are constant parameters depending on the topology of melt phase and the  
 601 geometry of an individual melt channel. Based on the previous works (Miller et al., 2014; Z.  
 602 Wang et al., 2020), here some suitable ranges of  $C$  (36-94) and  $n$  (2.4-2.8) are adopted for our  
 603 experiments that contain the multiphase-mineral assemblage.

604 On the basis of the numerical model from McKenzie (1989), the extraction velocity of  
 605 melt relative to the stationary crystal framework can be estimated by:

$$606 \quad v = \frac{k \Delta \rho_m g}{\Phi \mu_m}, \quad (3)$$

607 where  $\Delta \rho_m$  - the density contrast between silicate melt and solid mineral,  $g$  - the gravitational  
 608 acceleration, and  $\mu_m$  - the melt viscosity, which decreases with increasing temperature and  
 609 volatile content (mainly CO<sub>2</sub> and H<sub>2</sub>O) (Dingwell, 1996; Di Genova et al., 2014). Due to the  
 610 models from Giordano et al. (2008), the viscosity ( $\mu_m$ ) of the starting basalt can be calculated as  
 611  $\sim 10.8$  Pa·s at 1250 °C, whereas the  $\mu_m$  of reacting melt after melt-rock reaction decreases to  $\sim$   
 612  $1.2$ - $2.7$  Pa·s under conditions of 1.5 GPa and 1250-1300 °C (Figure 8g). The reactive melt has a  
 613 higher viscosity of  $\sim 10.5$  Pa·s when the pressure and temperature respectively decrease to 0.5  
 614 GPa and 1200 °C (Figure 8g). Based on equation (2) and (3), the porous flow velocity of melt  
 615 through grain-scale percolation is  $\sim 0.76$   $\mu\text{m}/\text{h}$  in the partially molten peridotite of experiment  
 616 PC545 ( $\Phi \sim 7.13$  %;  $\mu_m \sim 2.2$  Pa s;  $d \sim 11.0$   $\mu\text{m}$ ;  $C \sim 36$ ,  $n \sim 2.6$ , Miller et al., 2014;  $\Delta \rho_m \sim 600$   
 617  $\text{kg}/\text{m}^3$ ). Apparently, the transport distance ( $\sim 54.7$   $\mu\text{m}$ ) of this slow-extracting ( $\sim 0.76$   $\mu\text{m}/\text{h}$ ) melt  
 618 within the longest experimental time (72h) is much shorter than the region of melt-peridotite

619 reaction observed in experiments (Figure 1a), which implies that other mechanisms of melt  
 620 extraction occur concurrently to drive a higher flow velocity. In addition to the melt junctions  
 621 among minerals, we also observed thick melt channels (Figure 2) between some opening mineral  
 622 grain boundaries in the ORL and olivine + melt layer where a new crystal framework may have  
 623 not been built, indicating the possible contribution of melt channel flow, which have been  
 624 demonstrated by some experimental studies (*e.g.*, Pec et al., 2017; Zhu et al., 2011) and the field  
 625 investigations on dunite channels in MORB (*e.g.*, Kelemen et al., 1995). The melt velocity in  
 626 these channels is suggested to be much higher relative to other regions (Pec et al., 2017; Wang &  
 627 Jin, 2020).

628 The physical mobility of sulfide liquids along with the sulfur-saturated silicate melt  
 629 within the crystal framework of partially molten peridotite largely depends on the size of sulfide  
 630 droplets, the permeability of crystal framework, the connectivity of melt network, and the  
 631 processing degree of melt-rock reaction (Z. Wang et al., 2020; Yao & Mungall, 2020). Chung &  
 632 Mungall (2009) has simplified this complex process and proposed that the migration of sulfide is  
 633 mostly controlled by the relative sizes of sulfide droplets and surrounding crystals within  
 634 partially molten peridotite. When a sulfide droplet is smaller than the size of the most constricted  
 635 part of melt channel or pore throat in the crystal framework, it can be easily entrained by  
 636 ascending melt through peridotite without any obstruction if the terminal settling velocity of  
 637 sulfide droplets is less than the upward velocity of melt flow. For the spherical sulfide droplet,  
 638 the terminal settling velocity will achieve when its gravity-driven buoyancy force ( $F_B$ ) can be  
 639 balanced by the vertical drag force ( $F_D$ ).

640 The melt flows passing through sulfide droplets that are nearly stranded in the melt  
 641 channels or port throat, as a first approximation, can be regarded as the flow of a viscous  
 642 incompressible fluid around a circular cylindrical post (*i.e.*, the sulfide droplet) confined between  
 643 two parallel flat plates (*i.e.*, the melt channel) (Figure 10a) (Lee & Fung, 1969). The drag force  
 644 ( $F_D$ ) acting on this sulfide droplet can be expressed by a dimensionless coefficient ( $f_D$ ):

$$645 \quad F_D = f_D \times 4\pi\mu_m UR, \quad (4)$$

646 where  $R$  is half the distance between the two plates, *i.e.*, the radius of the melt channel, and  $U$  is  
 647 the stokes flow velocity of melt around sulfide droplet. Based on the modeling results from Lee  
 648 & Fung (1969), this dimensionless coefficient,  $f_D$ , can be estimated via the ratio between  $R$  and  $r_s$

649 that is the radius of an initial undeformed sulfide droplet before entering the melt channel (Figure  
 650 10b). For simplification, here we assume that spherical sulfide droplets have a radius ( $r_s$ )  
 651 equaling to or being slightly higher than  $R$ , and thus the corresponding values of  $f_D$  should  
 652 exceed  $\sim 4.8$  (Figure 10b) (Lee & Fung, 1969).

653 On the other hand, sulfide droplets always have a trend of settling to the base of the  
 654 capsule due to gravity, and the corresponding buoyancy force ( $F_B$ ) can be expressed as:

$$655 \quad F_B = \frac{4}{3}\pi r_s^3 g \Delta\rho_s, \quad (5)$$

656 where  $\Delta\rho_s = 1700 \text{ kg/m}^3$  is the density difference between the sulfide and silicate melt (Kress et  
 657 al., 2008; Z. Wang et al., 2020). Hence, constrained by fluid dynamical arguments, the velocity  
 658 of melt flow that just successfully entrains sulfide droplets whose sizes are close to that of melt  
 659 channel/pore throat can be calculated via:

$$660 \quad U = \frac{F_D}{4\pi\mu_m R \cdot f_D} = \frac{F_B}{4\pi\mu_m R \cdot f_D} = \frac{r_s^3 g \Delta\rho_s}{3\mu_m R \cdot f_D}, \quad (6)$$

661 Because the measured maximum radius of sulfide droplets in the upper peridotite part is  
 662  $\sim 9.5 \text{ }\mu\text{m}$  under the conditions of 72h and 1250 °C, the maximum value of  $U$  can be estimated as  
 663 about  $\sim 170.9 \text{ }\mu\text{m/h}$  ( $\sim 1.5 \text{ m/year}$ ), when the  $R/r_s$  equals to 1. This value, at least to some extent,  
 664 represents the localized velocity of channelized melt flow in the regions of melt-rock reaction  
 665 and is nearly three orders of magnitude higher than the velocity of melt porous flow ( $\sim 0.76$   
 666  $\mu\text{m/h}$ ) driven by the compaction of dense crystal framework. Given the velocity of melt flow  
 667 should be changeable due to the complex morphology of interconnected melt channels and  
 668 variable fluid dynamic environments in porous peridotite (McKenzie, 1989; Miller et al., 2014;  
 669 Z. Wang et al., 2020; Zhu et al., 2011), our calculation ( $\sim 1.5 \text{ m/year}$ ) here can be used as an  
 670 order of magnitude estimate for the velocity of melt flow around sulfide droplet through narrow  
 671 pore throat in these experiments.

672 First of all, we must use some reasonable geometrical parameters of the natural partially  
 673 molten rocks, such as the grain size, and the distribution of relative sizes of pores and pore  
 674 throats (Chung & Mungall, 2009; Yao & Mungall, 2020), before extrapolating our models to the  
 675 partially molten mantle. Commonly, the porosity ( $\Phi$ ) is approximately estimated to be  $\sim 1\text{-}15 \%$   
 676 in the partially molten mantle (e.g., Mei et al., 2002; Yoshino et al., 2010; Zhu et al., 2011), and  
 677 the high-degree melt-rock reaction regions would be expected to obtain a higher porosity (Pec et

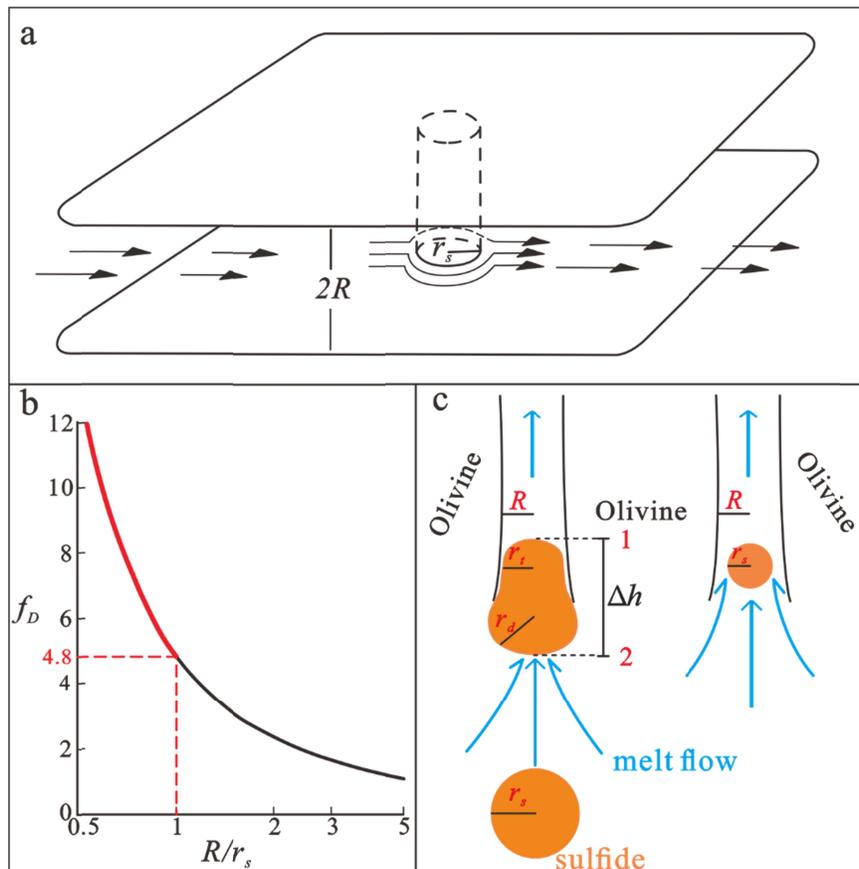
678 al., 2017). Thus, the  $\Phi$  in melt-rock reaction regions is properly set as 3-20 % in this study. The  
 679 observed grain size of upper mantle peridotite is about mm-scale (Ave Lallemand et al., 1980;  
 680 Karato, 1984), and thus the crystal radius ( $R_c$ ) could be reasonably set as  $\sim 1.5$  mm, which is  
 681 about two orders of magnitude higher than the grain size ( $\sim 10$   $\mu\text{m}$ ) in our experiments. Hence,  
 682 the permeability of upper mantle peridotite is much higher than the estimated value for our  
 683 experiments, and consequently, the extraction velocity of melt relative to the stationary crystal  
 684 framework could even increase from about  $\sim 4.88 \times 10^{-6} - 5.73 \times 10^{-5}$  m/s ( $\sim 153.89 - 1807.20$   
 685 m/year) with the growth of  $\Phi$  from 3 to 20 %, if other parameters in equation (3) remain constant.

686 On the other hand, for the partially molten mantle, a diversity of properties and  
 687 mechanisms contribute to the heterogeneous physics of melt extraction (Katz et al., 2022; Z.  
 688 Wang et al., 2020; Zhu et al., 2011), and the melt transport velocities can be estimated to be a  
 689 wide range of 1 to 1000 m/year by analyses of uranium-series (U-series) disequilibria in lavas  
 690 and Icelandic deglaciation (Katz et al., 2022; Rees Jones & Rudge, 2020; Rubin et al., 2005).  
 691 Commonly, channelized melt flow potentially characterizes a high melt extraction velocity in  
 692 melt channels with high porosity, which could even reach  $\sim 10$  km/year (McKenzie, 2000). In  
 693 contrast, recent laboratory experiments further estimate some lower velocities ( $\sim 2.4-29.7$   
 694 m/year, Z. Wang et al., 2020;  $\sim 2-150$  m/year, Connolly et al., 2009;  $\sim 0.5$  m/year, Zhu et al.,  
 695 2011). Obviously, these above values ( $\sim 0.5-10000$  m/year) are orders of magnitude almost the  
 696 same as or significantly larger than those ( $\sim 153.89 - 1807.20$  m/year) of our experiments, and  
 697 the high estimated value could be attributed to the heterogeneous melt flow, the larger grain size  
 698 and associated higher permeability in the natural peridotite than those in our experiments.

699 Combined with the numerical model from Sweeney & Martin (2003) estimating the  
 700 relationship between the  $R/R_c$  ratio and  $\Phi$ , we depict the radius ( $r_s$ ) of initial undeformed sulfide  
 701 droplet that can pass through the pore throat as a function of  $\Phi$  of the partially molten mantle in  
 702 Figure 11. When the radius ( $r_s$ ) of a sulfide droplet is smaller than and equal to the minimum  
 703 constriction ( $R$ ) of the melt channel, the relationship between  $r_s$  and  $\Phi$  is displayed in the regions  
 704 of No. 1 and 2 of Figure 11. If the flow velocity is less than  $\sim 1.16 \times 10^{-5}$  m/s (about 366 m/year),  
 705 the vertical drag force ( $F_D$ ) from this slow-flowing melt starts to be smaller than the gravity-  
 706 driven buoyancy force ( $F_B$ ) of sulfide droplet that has the same or similar radius with that of pore  
 707 throat at high porosity. In this situation, the maximum size ( $\sim 148.5$   $\mu\text{m}$ ) of upward migrating  
 708 sulfide droplet is limited by the low flow velocity and will keep constant at high porosity, but is

709 still controlled by the radius of pore throat at low porosity. Assuming that the velocity of the melt  
 710 flow increases from about  $5.1 \times 10^{-6}$  m/s ( $\sim 160$  m/year) to  $2.1 \times 10^{-5}$  m/s ( $\sim 660$  m/year), the  
 711 maximum radius of sulfide droplets that could be carried along with silicate melt increases from  
 712 about  $\sim 98.2$   $\mu\text{m}$  to  $\sim 200.0$   $\mu\text{m}$  (Figure 11-No. 1 region), and with increasing  $\Phi$  to 20%, the  
 713 value of  $r_s$  significantly increases up to 246.7  $\mu\text{m}$ , which is equal to the radius of pore throat at  $\Phi$   
 714 = 20%, with a lower limit of velocity  $\sim 3.2 \times 10^{-5}$  m/s ( $\sim 1010$  m/year) (Figure 11-No. 2 region).  
 715 These indicate that the size of sulfide droplets is dominated by the grain size and porosity of  
 716 peridotite, that is the size of pore throat, and the velocity of melt porous flow under the condition  
 717 of  $r_s \leq R$ .

718 Therefore, during the melt-peridotite reaction, the accompanying melt flow in the  
 719 dissolved peridotite is energetic, and its flow velocity seems to be high enough to drive the  
 720 upward transport of fine  $\mu\text{m}$ -scale sulfide droplets (Figure 10c-right side; Figure 11-the No. 1  
 721 and 2 regions).



723 **Figure 10.** Schematic map illustrating the flow of a viscous incompressible fluid around a  
 724 circular cylindrical post confined between two parallel flat plates (a), the dimensionless  
 725 coefficients  $f_D$  representing the drag force acting on sulfide droplet as a function of the  $R/r_s$  ratio  
 726 (b) (modified from Lee & Fung, 1969), and the extrusion and entrainment of large (left side) and  
 727 small (right side) sulfide droplets driven by focused melt flow (c). In (a), the two parallel flat  
 728 plates and a circular cylindrical post are considered as the walls of the pore throat/melt channel  
 729 and sulfide droplet, respectively.

#### 730 4.5 Potentially upward transport of coarse sulfide droplet

731 When a coarse sulfide that is sufficiently larger than the minimum constriction of the  
 732 melt channel is gradually squeezed into the small “throat” part along with the ascending reactive  
 733 melt flow, the sulfide droplet will invade this throat and be deformed into a pear shape with a  
 734 smaller radius of curvature (Figure 10c-left side). Obviously, the upward migration of this coarse  
 735 sulfide droplet in the partially molten peridotite becomes complicated, and the associated  
 736 dynamic process can simplistically involve a competition between pressure gradient, viscous,  
 737 buoyancy, capillary, and drag forces (Yao & Mungall, 2020). This squeezing process introduces  
 738 the additional capillary pressure  $P_c$ , which is a measure of the pressure discontinuity existing at  
 739 the interface of two immiscible phases (Chung & Mungall, 2009). The additional pressure  
 740 imposed on this deformed sulfide droplet within the squeezing process equals the difference in  
 741 capillary pressure between the top and bottom interfaces of the sulfide droplet and can be  
 742 calculated by:

$$743 \quad \Delta P_c = 2\gamma_{ms} \times \left( \frac{1}{r_t} - \frac{1}{r_d} \right), \quad (7)$$

744 where  $\gamma_{ms} = 0.21$  N/m is the melt-sulfide surface tension (Mungall et al., 2015),  $r_t$  is the radius of  
 745 the upper spheric cap, and  $r_d$  is the radius of the lower part of the pear-shaped droplet (Figure  
 746 10c). The droplet cannot be forced through the pore throat unless this excess pressure  $\Delta P_c$  is  
 747 balanced by an equal or greater pressure ( $\Delta P$ ) exerted by the ascending melt flow on the trailing  
 748 edge of sulfide droplet, which tends to push the droplet to rise through the melt channel.

749 Because the silicate melt flows at low Mach number, the simple form of Bernoulli’s  
 750 principle is valid for the quasi-incompressible flows in melt-peridotite reaction, and hence the

751 pressure of flowing melt around a sulfide droplet can be highly related to the peripheral flow  
 752 velocity via the Bernoulli's equation (Bauman & Schwaneberg, 1994):

$$753 \quad P_1 + \frac{\rho_m U_1^2}{2} + \rho_m g h_1 = P_2 + \frac{\rho_m U_2^2}{2} + \rho_m g h_2, \quad (8)$$

754 where the subscripts 1 and 2 correspond to the top and bottom outside points of deformed sulfide  
 755 droplets along the axis of symmetry, respectively;  $P$  is the pressure caused by the flow of silicate  
 756 melt;  $\rho_m$  is the density of melt;  $h$  is the height of 1 and 2 relative to a reference point beneath  
 757 sulfide droplet. Hence, the flowing-melt-driven external pressure imposed on the deformed  
 758 sulfide droplet equals the difference ( $\Delta P$ ) between  $P_1$  and  $P_2$ . In an extreme case, the coarse,  
 759 deformed sulfide droplet completely blocks the ascending melt flow, and the flow velocity at  
 760 bottom of droplets ( $U_2$ ) is close enough to zero, leading to the maximum value of external  
 761 pressure difference ( $\Delta P$ ) that can be calculated as:

$$762 \quad \Delta P = P_2 - P_1 = \frac{\rho_m U_1^2}{2} + \rho_m g (h_1 - h_2) = \frac{\rho_m U_1^2}{2} + \rho_m g \Delta h, \quad (9)$$

763 where  $U_1$  is the flow velocity of melt at the top of sulfide droplet, and  $\Delta h$  is the vertical distance  
 764 between the top and bottom points of the droplet. Assuming that the volume of a sulfide droplet  
 765 is unchanged during the squeezing process, the height of this invading sulfide droplet ( $\Delta h$ ) can be  
 766 described by Chung & Mungall (2009):

$$767 \quad \Delta h = r_t + r_d + \sqrt{r_d^2 - R^2} - \sqrt{r_t^2 - R^2}, \quad (10)$$

768 For the deformed sulfide droplet, it bears the upward pressure driven by flow melt (*i.e.*,  
 769  $\Delta P$ ), which should balance or even exceed the total of downward capillary pressure ( $\Delta P_C$ ) and  
 770 gravity-driven pressure ( $P_G$ ) for the droplet to rise through the constriction:

$$771 \quad \Delta P \geq P_G + \Delta P_C = \rho_s g \Delta h + 2\gamma_{ms} \times \left(\frac{1}{r_t} - \frac{1}{r_d}\right), \quad (11)$$

772 where  $\rho_s$  is the density of sulfide liquid. Thus, combining the above equations (7-10), under the  
 773 critical condition for migration of sulfide droplet into the pore throat to be the case ( $r_t = R$ )  
 774 (Chung & Mungall, 2009), the equation 11 substituted with the appropriate values for  $\rho_m = 2600$   
 775  $\text{kg/m}^3$  (Robertson et al., 2016) is simplified into:

$$776 \quad 1300 \times U_1^2 \geq 16660 \times \left(R + r_d + \sqrt{r_d^2 - R^2}\right) + 0.42 \times \left(\frac{1}{R} - \frac{1}{r_d}\right), \quad (12)$$

777 And finally, the initial radius ( $r_s$ ) of this sulfide droplet before its invasion into the pore  
778 throat can be calculated as Chung & Mungall (2009):

$$779 \quad r_s = (0.5 \times (r_d^3 + R^3 + (r_d^2 + \frac{1}{2}R^2) \times \sqrt{r_d^2 - R^2}))^{1/3}, \quad (13)$$

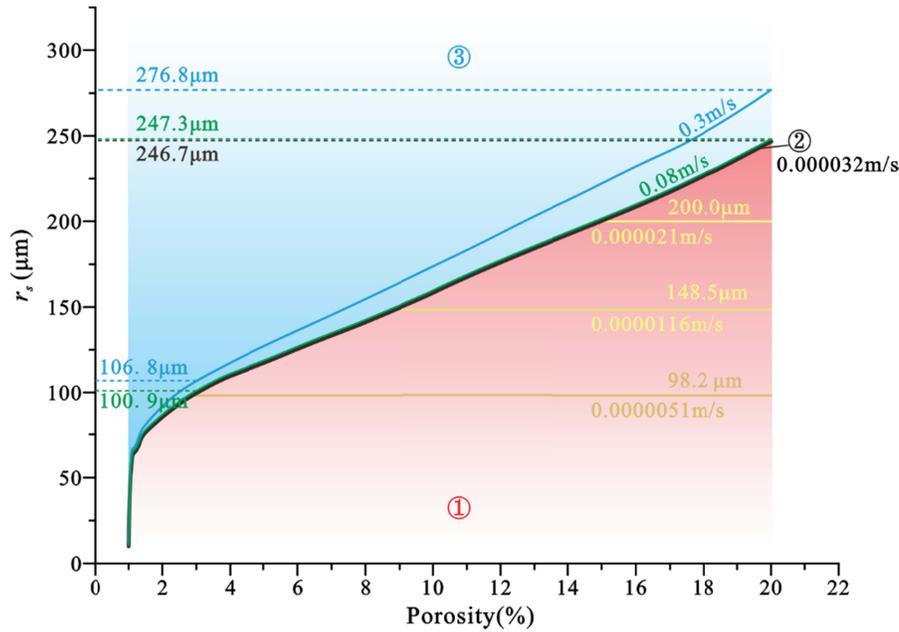
780 When silicate melt flows around a sulfide droplet through a narrow pore throat in the  
781 partially molten peridotite, the associated velocity field is suggested to be not homogeneous. In  
782 addition, it is well known that exactly deciphering the flow velocity field of silicate melt  
783 infiltrating through multi-mineral phase system plays a critically important role in understanding  
784 geodynamic properties of the partially molten upper mantle, but it is really challenging and  
785 beyond the scope of this study. The increases in grain size and permeability may also strongly  
786 enhance the melt flow velocity in the new channel of crystal framework. Here, the flow velocity  
787 in the channel has been confirmed to be about three orders of magnitude higher than that of  
788 porous flow through grain-scale percolation, which may be extrapolated to the mm-scale upper  
789 mantle peridotite. Therefore, the velocity of melt flow around a sulfide droplet in the channel in  
790 upper mantle peridotite may be within the magnitude of 0.001-0.1 m/s, and thus the local  
791 velocity ( $U_l$ ) of focused melt flow around the top of sulfide droplet through the narrow pore  
792 throat could be analogously assumed to be  $\sim 0.3$  m/s and  $\sim 0.08$  m/s here (Figure 11-No.3  
793 region) as the possible values in the localized high-degree melt-rock reaction regions of the  
794 partially molten mantle with high porosity. This range of flow velocity is also consistent with the  
795 estimates from previous studies discussed in Section 4.4.

796 Equation 12 and 13 are numerically solved for the above value of  $U_l$  by programming a  
797 loop using Maple<sup>®</sup>. Based on the above numerical calculation, when the velocity of focused melt  
798 flow around a sulfide droplet is within the range of  $\sim 0.08$  m/s and  $3.2 \times 10^{-5}$  m/s (corresponding  
799 to the range of the velocity of melt porous flow from  $\sim 2500$  m/year to  $\sim 1$  m/year), larger sulfide  
800 droplets than the radius ( $R$ ) of melt channel/pore throat are difficult to overcome the capillary  
801 pressure and will be stranded in the inter-grain pores (Figure 11-region No. 3). However, these  
802 flow velocities are still strong enough to drive the upward transport of sulfide droplets whose  
803 radii are smaller than or similar to the radius ( $R$ ) of pore throat, and as shown in the region of No.  
804 3 of Figure 11, the estimated maximum value of  $r_s$  increases from  $\sim 100.9$  to  $247.3$   $\mu\text{m}$  at the  
805 velocity of 0.08 m/s, which is roughly similar to  $R$  (from  $\sim 99.7$  to  $246.7$   $\mu\text{m}$  in the region No. 2)  
806 with the growth of  $\Phi$  from 3 to 20 %. In this range of flow velocity, the maximum size of

807 upward migrating sulfide droplets is limited by the radius of pore throat and consequently  
808 depends on the grain size and porosity of the upper mantle.

809 In contrast, for a powerful melt flow with extremely high velocity (*e.g.*,  $\sim 0.3$  m/s in  
810 Figure 11-region No. 3), coarse-grained sulfide droplet may overcome the capillary pressure  
811 driven by the pore throat, and its maximum radius for upward migration increases from  $\sim 106.8$   
812  $\mu\text{m}$  to  $\sim 276.8$   $\mu\text{m}$  with increasing the porosity from 3 % to 20 % (Figure 11-region No.3), where  
813 the sulfide droplets can pass through pore throat narrower than themselves. Although the  
814 estimates of the maximum sulfide size that are capable of upward migrating through the porous  
815 molten peridotite are divided into three conditions, all of them are mostly dominated by the melt  
816 flow velocity, porosity, and grain size of peridotite, which would be strongly enhanced by the  
817 high-degree melt-rock reaction.

818 On the other hand, based on these previous studies on the abyssal and orogenic  
819 peridotites, and peridotite xenoliths (Lorand & Luguet, 2016; Lorand et al., 2010; Luguet et al.,  
820 2003), highly variable grain size ( $< 20$   $\mu\text{m}$  to  $500$   $\mu\text{m}$ ) of sulfides is widely observed in the upper  
821 mantle, while most sulfides are likely smaller than about  $100$   $\mu\text{m}$  in diameter. Hence, it is  
822 plausible that almost all sulfide droplets are stranded among the crystal framework of the upper  
823 peridotite mantle with low porosity, while the high-degree partial melting and/or melt-peridotite  
824 reaction will lead to high porosity, large flow velocity, and even the new-forming melt channel,  
825 which proposes that efficient entrainment of most sulfide droplets into ascending magma flow  
826 among porous peridotite is physically possible, especially among the channelized extraction of  
827 silicate melt in melt-peridotite reaction. Conceivably, the mechanical entrainment of sulfide  
828 droplets during the melt-rock reaction would be more efficient than the transport of sulfide liquid  
829 by dissolving in departing silicate melt. The sub-continental lithospheric mantle that undergoes  
830 large-scale partial melting and melt-peridotite reaction under tectonically active craton margins  
831 may be favorable conjunction of these factors and potentially conducive to the entrainment of  
832 large sulfide droplets, thereby contributing to the fertilization of sub-continental lithospheric  
833 mantle and the primary enrichment of metal-bearing sulfides for magmatic sulfide deposits (*e.g.*,  
834 Griffin et al., 2013). This may be one of the reasons to account for the issue of why the spatial  
835 and temporal distribution of magmatic Ni-Cu-(PGE) sulfide deposits are genetically related to  
836 these regions (Maier and Groves, 2011).



837

838 **Figure 11.** Plots of the radius ( $r_s$ ) of initial undeformed sulfide droplets that could extrude the  
 839 pore throat as a function of the porosity ( $\Phi$ ) of the partially molten mantle. The relationship  
 840 between  $r_s$  and  $\Phi$  is divided into three parts according to the relative size of sulfide droplets'  
 841 radius ( $r_s$ ) and the pore throat's minimum constriction ( $R$ ). At  $r_s < R$ , the relationship is shown in  
 842 region No. 1, whereas regions No. 2 and No.3 are respectively under the conditions of  $r_s = R$  and  
 843  $r_s > R$ .

844 **5 Conclusions**

845 Reaction-infiltration of silicate melt and concomitant transport of sulfide droplets in the  
 846 partially molten peridotite are examined experimentally and thermodynamical-quantitatively in  
 847 this study, which provides important insights into the fertilization of sub-continental lithospheric  
 848 mantle and the efficient recycling of sulfur and metal elements from the partially molten mantle.  
 849 The reaction between peridotite and basalt leads to the preferential dissolution of olivine and the  
 850 precipitation of orthopyroxene at high temperature (1250-1300 °C) and pressure (1.5 GPa),  
 851 forming an orthopyroxene-rich reaction layer (ORL) with a high growth rate ( $\sim 0.78 \pm 0.05$   
 852  $\mu\text{m/s}^{0.5}$  at 1250 °C) in the melt-rock interface. With increasing the degree of melt-peridotite  
 853 reaction, more silicate melt infiltrates through the ORL into the upper partially molten peridotite,  
 854 forming an olivine-melt layer above the ORL, and coarse sulfide droplets could also be entrained  
 855 along with the infiltration of silicate melt. Systematic variations observed in the compositions of

856 minerals and reactive melt are approximately consistent with the thermodynamically-constrained  
857 mixing model. Meanwhile, the sizes of sulfide droplets have fast growth in the melt source,  
858 which is partly attributed to the multiple coarsening processes (including the mechanical  
859 coalescence) of sulfide droplets in the dynamic reactive melt flow.

860         Suspension and entrainment of sulfide droplets in the partially molten peridotite indicate  
861 a fast-flowing velocity of reactive melt ( $\sim 170.9 \mu\text{m/h}$ ) in our experiments, which demonstrates  
862 the occurrence of focused melt flows with high velocity in the reactive-forming melt channels,  
863 besides the melt porous flow driven by the density difference between silicate melt and minerals  
864 forming the crystal framework. It is convincible that the melt flow velocity within the melt-rock  
865 reaction is potentially high enough to drive the upward transport of fine  $\mu\text{m}$ -scale sulfide droplets  
866 with smaller diameters than that of the pore throat in the partially molten peridotite. In this  
867 condition, the maximum size of sulfide for upward entrainment is limited by the radius of pore  
868 throat in the crystal framework of peridotite and increases with the growth of porosity during  
869 partial melting and melt-peridotite reaction. Only in channelized melt flow with extremely high  
870 velocity, coarse-grained sulfide droplets could upward migrate through the pore throats narrower  
871 than themselves. Hence, the high-degree melt-rock reaction regions in the upper mantle, such as  
872 the sub-continental lithospheric mantle of craton margins, are likely to have wide pore throats,  
873 high porosity, and even channelized melt flow with high velocity, which can potentially drive  
874 upward entrainment and/or extrusion of  $\mu\text{m}$ - to  $\text{mm}$ -scale sulfide droplets in the partially molten  
875 mantle, and thereby fertilize the lithospheric mantle and lead to the endowment of Cu- and Ni-  
876 bearing sulfide for the formation of associated deposits.

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886 **Data Availability Statement**

887 All data are shown in figures, supplemental materials, and tables can also be found in the  
888 Supporting Information.

889 **References**

- 890 Aharonov, E. (1995). Channeling instability of upwelling melt in the mantle. *Journal of*  
891 *Geophysical Research: Solid Earth*, 100(B10), 20433–20450.  
892 <https://doi.org/10.1029/95JB01307>
- 893 Alard, O., Lorand, J. P., Reisberg, L., Bodinier, J. L., Dautria, J. M., & O'Reilly, S. Y. (2011).  
894 Volatile-rich metasomatism in Montferrier xenoliths (Southern France): Implications for the  
895 abundances of chalcophile and highly siderophile elements in the subcontinental mantle.  
896 *Journal of Petrology*, 52(10), 2009–2045. <https://doi.org/10.1093/petrology/egr038>
- 897 Ave Lallemand, H. G., Mercier, J.-C. C., Carter, N. L., & Ross, J. V. (1980). Rheology of the  
898 upper mantle: Inferences from peridotite xenoliths. *Tectonophysics*, 70(1–2), 85–113.  
899 [https://doi.org/10.1016/0040-1951\(80\)90022-0](https://doi.org/10.1016/0040-1951(80)90022-0)
- 900 Bagdassarov, N., Solferino, G., Golabek, G. J., & Schmidt, M. W. (2009). Centrifuge assisted  
901 percolation of Fe–S melts in partially molten peridotite: Time constraints for planetary core  
902 formation. *Earth and Planetary Science Letters*, 288(1–2), 84–95.  
903 <https://doi.org/10.1016/j.epsl.2009.09.010>
- 904 Ballhaus, C., Bockrath, C., Wohlgemuth-Ueberwasser, C., Laurenz, V., & Berndt, J. (2006).  
905 Fractionation of the noble metals by physical processes. *Contributions to Mineralogy and*  
906 *Petrology*, 152(6), 667–684. <https://doi.org/10.1007/s00410-006-0126-z>
- 907 Bauman, R. P., & Schwaneberg, R. (1994). Interpretation of Bernoulli's equation. *The Physics*  
908 *Teacher*, 32(8), 478–488. <https://doi.org/10.1119/1.2344087>
- 909 Bockrath, C., Ballhaus, C., & Holzheid, A. (2004). Fractionation of the platinum-group elements  
910 during mantle melting. *Science*, 305(5692), 1951–1953.  
911 <https://doi.org/10.1126/science.1100160>
- 912 Cascio, M. Lo, Liang, Y., Shimizu, N., & Hess, P. C. (2008). An experimental study of the grain-  
913 scale processes of peridotite melting: Implications for major and trace element distribution  
914 during equilibrium and disequilibrium melting. *Contributions to Mineralogy and Petrology*,  
915 156(1), 87–102. <https://doi.org/10.1007/s00410-007-0275-8>

- 916 Chadam, J., Hoff, D., Merino, E., Ortoleva, P., & Sen, A. (1986). Reactive infiltration  
917 instabilities. *IMA Journal of Applied Mathematics (Institute of Mathematics and Its  
918 Applications)*, 36(3), 207–221. <https://doi.org/10.1093/imamat/36.3.207>
- 919 Chen, C., Yao, Z.S., Wang, C.Y. (2022). Partitioning behaviors of cobalt and manganese along  
920 diverse melting paths of peridotitic and MORB-like pyroxenitic mantle. *Journal of Petrology*,  
921 63, 1-22. <https://doi.org/10.1093/petrology/egac021>
- 922 Chung, H. Y., & Mungall, J. E. (2009). Physical constraints on the migration of immiscible  
923 fluids through partially molten silicates, with special reference to magmatic sulfide ores. *Earth  
924 and Planetary Science Letters*, 286(1–2), 14–22. <https://doi.org/10.1016/j.epsl.2009.05.041>
- 925 Ciazela, J., Koepke, J., Dick, H. J. B., Botcharnikov, R., Muszynski, A., Lazarov, M., et al.  
926 (2018). Sulfide enrichment at an oceanic crust-mantle transition zone: Kane Megamullion  
927 (23°N, MAR). *Geochimica et Cosmochimica Acta*, 230, 155–189.  
928 <https://doi.org/10.1016/j.gca.2018.03.027>
- 929 Connolly, J. A. D., Schmidt, M. W., Solferino, G., & Bagdassarov, N. (2009). Permeability of  
930 asthenospheric mantle and melt extraction rates at mid-ocean ridges. *Nature*, 462(7270), 209–  
931 212. <https://doi.org/10.1038/nature08517>
- 932 Daines, M. J., & Kohlstedt, D. L. (1994). The transition from porous to channelized flow due to  
933 melt/rock reaction during melt migration. *Geophysical Research Letters*, 21(2), 145–148.  
934 <https://doi.org/10.1029/93GL03052>
- 935 de Bremond d’Ars, J., Arndt, N. T., & Hallot, E. (2001). Analog experimental insights into the  
936 formation of magmatic sulfide deposits. *Earth and Planetary Science Letters*, 186(3–4), 371–  
937 381. [https://doi.org/10.1016/S0012-821X\(01\)00254-0](https://doi.org/10.1016/S0012-821X(01)00254-0)
- 938 Di Genova, D., Romano, C., Alletti, M., Misiti, V., & Scarlato, P. (2014). The effect of CO<sub>2</sub> and  
939 H<sub>2</sub>O on Etna and Fondo Riccio (Phlegrean Fields) liquid viscosity, glass transition  
940 temperature and heat capacity. *Chemical Geology*, 377, 72–86.  
941 <https://doi.org/10.1016/j.chemgeo.2014.04.001>
- 942 Ding, S., & Dasgupta, R. (2017). The fate of sulfide during decompression melting of peridotite  
943 – implications for sulfur inventory of the MORB-source depleted upper mantle. *Earth and  
944 Planetary Science Letters*, 459, 183–195. <https://doi.org/10.1016/j.epsl.2016.11.020>
- 945 Dingwell, D. B. (1996). Volcanic Dilemma--Flow or Blow? *Science*, 273(5278), 1054–1055.  
946 <https://doi.org/10.1126/science.273.5278.1054>

- 947 Farquhar, J., Wing, B. A., McKeegan, K. D., Harris, J. W., Cartigny, P., & Thiemens, M. H.  
948 (2002). Mass-independent sulfur of inclusions in diamond and sulfur recycling on early Earth.  
949 *Science*, 298(5602), 2369–2372. <https://doi.org/10.1126/science.1078617>
- 950 Faul, U. H. (2001). Melt retention and segregation beneath mid-ocean ridges. *Nature*, 410(6831),  
951 920–923. <https://doi.org/10.1038/35073556>
- 952 Faul, U. H., & Scott, D. (2006). Grain growth in partially molten olivine aggregates.  
953 *Contributions to Mineralogy and Petrology*, 151(1), 101–111. [https://doi.org/10.1007/s00410-](https://doi.org/10.1007/s00410-005-0048-1)  
954 005-0048-1
- 955 Freda, C., Baker, D. R., & Scarlato, P. (2005). Sulfur diffusion in basaltic melts. *Geochimica et*  
956 *Cosmochimica Acta*, 69(21), 5061–5069. <https://doi.org/10.1016/j.gca.2005.02.002>
- 957 Ghiorso, M. S., & Sack, R. O. (1995). Chemical mass transfer in magmatic processes IV. A  
958 revised and internally consistent thermodynamic model for the interpolation and extrapolation  
959 of liquid-solid equilibria in magmatic systems at elevated temperatures and pressures.  
960 *Contributions to Mineralogy and Petrology*, 119(2–3), 197–212.  
961 <https://doi.org/10.1007/BF00307281>
- 962 Ghiorso, M. S., Hirschmann, M. M., Reiners, P. W., & Kress, V. C. (2002). The pMELTS: A  
963 revision of MELTS for improved calculation of phase relations and major element partitioning  
964 related to partial melting of the mantle to 3 GPa. *Geochemistry, Geophysics, Geosystems*, 3(5),  
965 1–35. <https://doi.org/10.1029/2001gc000217>
- 966 Giordano, D., Russell, J. K., & Dingwell, D. B. (2008). Viscosity of magmatic liquids: A model.  
967 *Earth and Planetary Science Letters*, 271(1–4), 123–134.  
968 <https://doi.org/10.1016/j.epsl.2008.03.038>
- 969 Griffin, W. L., Begg, G. C., & O'Reilly, S. Y. (2013). Continental-root control on the genesis of  
970 magmatic ore deposits. *Nature Geoscience*, 6(11), 905–910. <https://doi.org/10.1038/ngeo1954>
- 971 Heinrich, C. A., & Connolly, J. A. D. (2022). Physical transport of magmatic sulfides promotes  
972 copper enrichment in hydrothermal ore fluids. *Geology*, 50(10), 1101–1105.  
973 <https://doi.org/10.1130/G50138.1>
- 974 Higgins, M. D. (1998). Origin of Anorthosite by Textural Coarsening: Quantitative  
975 Measurements of a Natural Sequence of Textural Development. *Journal of Petrology*, 39(7),  
976 1307–1323. <https://doi.org/10.1093/petrology/39.7.1307>

- 977 Higgins, M. D. (2011). Textural coarsening in igneous rocks. *International Geology Review*,  
978 53(3–4), 354–376. <https://doi.org/10.1080/00206814.2010.496177>
- 979 Holwell, D. A., Fiorentini, M. L., Knott, T. R., McDonald, I., Blanks, D. E., Campbell McCuaig,  
980 T., & Gorczyk, W. (2022). Mobilisation of deep crustal sulfide melts as a first order control on  
981 upper lithospheric metallogeny. *Nature Communications*, 13(1).  
982 <https://doi.org/10.1038/s41467-022-28275-y>
- 983 Holzheid, A. (2010). Separation of sulfide melt droplets in sulfur saturated silicate liquids.  
984 *Chemical Geology*, 274(3–4), 127–135. <https://doi.org/10.1016/j.chemgeo.2010.03.005>
- 985 Holzheid, A., & Grove, T. L. (2002). Sulfur saturation limits in silicate melts and their  
986 implications for core formation scenarios for terrestrial planets. *American Mineralogist*, 87(2–  
987 3), 227–237. <https://doi.org/10.2138/am-2002-2-304>
- 988 Holzheid, A., Schmitz, M. D., & Grove, T. L. (2000). Textural equilibria of iron sulfide liquids  
989 in partly molten silicate aggregates and their relevance to core formation scenarios. *Journal of*  
990 *Geophysical Research: Solid Earth*, 105(B6), 13555–13567.  
991 <https://doi.org/10.1029/2000JB900046>
- 992 Honour, V. C., Holness, M. B., Partridge, J. L., & Charlier, B. (2019). Microstructural evolution  
993 of silicate immiscible liquids in ferrobasalts. *Contributions to Mineralogy and Petrology*,  
994 174(9). <https://doi.org/10.1007/s00410-019-1610-6>
- 995 Iacono-Marziano, G., Le Vaillant, M., Godel, B. M., Barnes, S. J., & Arbaret, L. (2022). The  
996 critical role of magma degassing in sulphide melt mobility and metal enrichment. *Nature*  
997 *Communications*, 13(1). <https://doi.org/10.1038/s41467-022-30107-y>
- 998 Jackson, M. D., Blundy, J., & Sparks, R. S. J. (2018). Chemical differentiation, cold storage and  
999 remobilization of magma in the Earth's crust. *Nature*, 564(7736), 405–409.  
1000 <https://doi.org/10.1038/s41586-018-0746-2>
- 1001 Karato, S.-I. (1984). Grain-size distribution and rheology of the upper mantle. *Tectonophysics*,  
1002 104(1–2), 155–176. [https://doi.org/10.1016/0040-1951\(84\)90108-2](https://doi.org/10.1016/0040-1951(84)90108-2)
- 1003 Katz, R. F., Jones, D. W. R., Rudge, J. F., & Keller, T. (2022). Physics of Melt Extraction from  
1004 the Mantle: Speed and Style. *Annual Review of Earth and Planetary Sciences*, 50, 507–540.  
1005 <https://doi.org/10.1146/annurev-earth-032320-083704>

- 1006 Kelemen, P. B., Shimizu, N., & Salters, V. J. M. (1995). Extraction of mid-ocean-ridge basalt  
1007 from the upwelling mantle by focused flow of melt in dunite channels. *Nature*, *375*(6534),  
1008 747–753. <https://doi.org/10.1038/375747a0>
- 1009 Kress, V., Greene, L. E., Ortiz, M. D., & Mioduszewski, L. (2008). Thermochemistry of sulfide  
1010 liquids IV: Density measurements and the thermodynamics of O-S-Fe-Ni-Cu liquids at low to  
1011 moderate pressures. *Contributions to Mineralogy and Petrology*, *156*(6), 785–797.  
1012 <https://doi.org/10.1007/s00410-008-0315-z>
- 1013 Lambart, S., Laporte, D., Provost, A., & Schiano, P. (2012). Fate of pyroxenite-derived melts in  
1014 the peridotitic mantle: Thermodynamic and experimental constraints. *Journal of Petrology*,  
1015 *53*(3), 451–476. <https://doi.org/10.1093/petrology/egr068>
- 1016 Lautze, N. C., Sisson, T. W., Mangan, M. T., & Grove, T. L. (2011). Segregating gas from melt:  
1017 An experimental study of the Ostwald ripening of vapor bubbles in magmas. *Contributions to*  
1018 *Mineralogy and Petrology*, *161*(2), 331–347. <https://doi.org/10.1007/s00410-010-0535-x>
- 1019 Lee, C. T. A., & Tang, M. (2020). How to make porphyry copper deposits. *Earth and Planetary*  
1020 *Science Letters*, *529*, 115868. <https://doi.org/10.1016/j.epsl.2019.115868>
- 1021 Lee, J. S., & Fung, Y. C. (1969). Stokes flow around a circular cylindrical post confined between  
1022 two parallel plates. *Journal of Fluid Mechanics*, *37*(4), 657–670.  
1023 <https://doi.org/10.1017/S0022112069000796>
- 1024 Liang, Y. (1999). Diffusive dissolution in ternary systems: Analysis with applications to quartz  
1025 and quartzite dissolution in molten silicates. *Geochimica et Cosmochimica Acta*, *63*(23–24),  
1026 3983–3995. [https://doi.org/10.1016/s0016-7037\(99\)00203-3](https://doi.org/10.1016/s0016-7037(99)00203-3)
- 1027 Lifshitz, I. M., & Slyozov, V. V. (1961). The kinetics of precipitation from supersaturated solid  
1028 solutions. *Journal of Physics and Chemistry of Solids*, *19*(1–2), 35–50.  
1029 [https://doi.org/10.1016/0022-3697\(61\)90054-3](https://doi.org/10.1016/0022-3697(61)90054-3)
- 1030 Liu, Y., Samaha, N.-T., & Baker, D. R. (2007). Sulfur concentration at sulfide saturation (SCSS)  
1031 in magmatic silicate melts. *Geochimica et Cosmochimica Acta*, *71*(7), 1783–1799.  
1032 <https://doi.org/10.1016/j.gca.2007.01.004>
- 1033 Lorand, J. P., & Luguët, A. (2016). Chalcophile and siderophile elements in Mantle Rocks:  
1034 Trace elements controlled by trace minerals. *Reviews in Mineralogy and Geochemistry*, *81*(1),  
1035 441–488. <https://doi.org/10.2138/rmg.2016.81.08>

- 1036 Lorand, J. P., Alard, O., & Luguet, A. (2010). Platinum-group element micronuggets and  
1037 refertilization process in Lherz orogenic peridotite (northeastern Pyrenees, France). *Earth and*  
1038 *Planetary Science Letters*, 289(1–2), 298–310. <https://doi.org/10.1016/j.epsl.2009.11.017>
- 1039 Luguet, A., Lorand, J. P., & Seyler, M. (2003). Sulfide petrology and highly siderophile element  
1040 geochemistry of abyssal peridotites: A coupled study of samples from the Kane Fracture Zone  
1041 (45°W 23°20N, MARK area, Atlantic Ocean). *Geochimica et Cosmochimica Acta*, 67(8),  
1042 1553–1570. [https://doi.org/10.1016/S0016-7037\(02\)01133-X](https://doi.org/10.1016/S0016-7037(02)01133-X)
- 1043 Mallik, A., & Dasgupta, R. (2012). Reaction between MORB-eclogite derived melts and fertile  
1044 peridotite and generation of ocean island basalts. *Earth and Planetary Science Letters*, 329–  
1045 330, 97–108. <https://doi.org/10.1016/j.epsl.2012.02.007>
- 1046 Mavrogenes, J. A., & O'Neill, H. S. C. (1999). The relative effects of pressure, temperature and  
1047 oxygen fugacity on the solubility of sulfide in mafic magmas. *Geochimica et Cosmochimica*  
1048 *Acta*, 63(7–8), 1173–1180. [https://doi.org/10.1016/S0016-7037\(98\)00289-0](https://doi.org/10.1016/S0016-7037(98)00289-0)
- 1049 Mckenzie, D. (1984). The generation and compaction of partially molten rock. *Journal of*  
1050 *Petrology*, 25(3), 713–765. <https://doi.org/10.1093/petrology/25.3.713>
- 1051 McKenzie, D. (1989). Some remarks on the movement of small melt fractions in the mantle.  
1052 *Earth and Planetary Science Letters*, 95(1–2), 53–72. [https://doi.org/10.1016/0012-](https://doi.org/10.1016/0012-821X(89)90167-2)  
1053 [821X\(89\)90167-2](https://doi.org/10.1016/0012-821X(89)90167-2)
- 1054 McKenzie, D. (2000). Constraints on melt generation and transport from U-series activity ratios.  
1055 *Chemical Geology*, 162(2), 81–94. [https://doi.org/10.1016/S0009-2541\(99\)00126-6](https://doi.org/10.1016/S0009-2541(99)00126-6)
- 1056 Médard, E., McCammon, C. A., Barr, J. A., & Grove, T. L. (2008). Oxygen fugacity,  
1057 temperature reproducibility, and H<sub>2</sub>O contents of nominally anhydrous piston-cylinder  
1058 experiments using graphite capsules. *American Mineralogist*, 93(11–12), 1838–1844.  
1059 <https://doi.org/10.2138/am.2008.2842>
- 1060 Mei, S., Bai, W., Hiraga, T., & Kohlstedt, D. L. (2002). Influence of melt on the creep behavior  
1061 of olivine-basalt aggregates under hydrous conditions. *Earth and Planetary Science Letters*,  
1062 201(3–4), 491–507. [https://doi.org/10.1016/S0012-821X\(02\)00745-8](https://doi.org/10.1016/S0012-821X(02)00745-8)
- 1063 Mercier, M., Muro, A. di, Métrich, N., Giordano, D., Belhadj, O., & Mandeville, C. W. (2010).  
1064 Spectroscopic analysis (FTIR, Raman) of water in mafic and intermediate glasses and glass  
1065 inclusions. *Geochimica et Cosmochimica Acta*, 74(19), 5641–5656.  
1066 <https://doi.org/10.1016/j.gca.2010.06.020>

- 1067 Milke, R., Abart, R., Kunze, K., Koch-Müller, M., Schmid, D., & Ulmer, P. (2009). Matrix  
1068 rheology effects on reaction rim growth I: Evidence from orthopyroxene rim growth  
1069 experiments. *Journal of Metamorphic Geology*, 27(1), 71–82. <https://doi.org/10.1111/j.1525-1314.2008.00804.x>
- 1071 Miller, K. J., Zhu, W. lu, Montési, L. G. J., & Gaetani, G. A. (2014). Experimental quantification  
1072 of permeability of partially molten mantle rock. *Earth and Planetary Science Letters*, 388,  
1073 273–282. <https://doi.org/10.1016/j.epsl.2013.12.003>
- 1074 Mitchell, A. L., & Grove, T. L. (2016). Experiments on melt–rock reaction in the shallow mantle  
1075 wedge. *Contributions to Mineralogy and Petrology*, 171(12). <https://doi.org/10.1007/s00410-016-1312-2>
- 1077 Morgan, Z., & Liang, Y. (2003). An experimental and numerical study of the kinetics of  
1078 harzburgite reactive dissolution with applications to dunite dike formation. *Earth and*  
1079 *Planetary Science Letters*, 214(1–2), 59–74. [https://doi.org/10.1016/S0012-821X\(03\)00375-3](https://doi.org/10.1016/S0012-821X(03)00375-3)
- 1080 Morgan, Z., & Liang, Y. (2005). An experimental study of the kinetics of lherzolite reactive  
1081 dissolution with applications to melt channel formation. *Contributions to Mineralogy and*  
1082 *Petrology*, 150(4), 369–385. <https://doi.org/10.1007/s00410-005-0033-8>
- 1083 Mungall, J. E., Brenan, J. M., Godel, B., Barnes, S. J., & Gaillard, F. (2015). Transport of metals  
1084 and sulphur in magmas by flotation of sulphide melt on vapour bubbles. *Nature Geoscience*,  
1085 8(3), 216–219. <https://doi.org/10.1038/ngeo2373>
- 1086 Mungall, J. E. (2002). Empirical models relating viscosity and tracer diffusion in magmatic  
1087 silicate melts. *Geochimica et Cosmochimica Acta*, 66(1), 125–143.  
1088 [https://doi.org/10.1016/S0016-7037\(01\)00736-0](https://doi.org/10.1016/S0016-7037(01)00736-0)
- 1089 Mungall, J. E., & Brenan, J. M. (2014). Partitioning of platinum-group elements and Au between  
1090 sulfide liquid and basalt and the origins of mantle-crust fractionation of the chalcophile  
1091 elements. *Geochimica et Cosmochimica Acta*, 125, 265–289.  
1092 <https://doi.org/10.1016/j.gca.2013.10.002>
- 1093 Niu, Y. (2004). Bulk-rock major and trace element compositions of abyssal peridotites:  
1094 Implications for mantle melting, melt extraction and post-melting processes beneath Mid-  
1095 Ocean ridges. *Journal of Petrology*, 45(12), 2423–2458.  
1096 <https://doi.org/10.1093/petrology/egh068>

- 1097 Patten, C., Barnes, S. J., Mathez, E. A., & Jenner, F. E. (2013). Partition coefficients of  
1098 chalcophile elements between sulfide and silicate melts and the early crystallization history of  
1099 sulfide liquid: LA-ICP-MS analysis of MORB sulfide droplets. *Chemical Geology*, 358, 170–  
1100 188. <https://doi.org/10.1016/j.chemgeo.2013.08.040>
- 1101 Pec, M., Holtzman, B. K., Zimmerman, M. E., & Kohlstedt, D. L. (2015). Reaction infiltration  
1102 instabilities in experiments on partially molten mantle rocks. *Geology*, 43(7), 575–578.  
1103 <https://doi.org/10.1130/G36611.1>
- 1104 Pec, M., Holtzman, B. K., Zimmerman, M. E., & Kohlstedt, D. L. (2017). Reaction infiltration  
1105 instabilities in mantle rocks: An experimental investigation. *Journal of Petrology*, 58(5), 979–  
1106 1003. <https://doi.org/10.1093/petrology/egx043>
- 1107 Pin, J., France, L., Lambart, S., & Reisberg, L. (2022). Thermodynamic modeling of melt  
1108 addition to peridotite: Implications for the refertilization of the non-cratonic continental  
1109 mantle lithosphere. *Chemical Geology*, 609, 121050.  
1110 <https://doi.org/10.1016/j.chemgeo.2022.121050>
- 1111 Rees Jones, D. W., & Rudge, J. F. (2020). Fast magma ascent, revised estimates from the  
1112 deglaciation of Iceland. *Earth and Planetary Science Letters*, 542, 116324.  
1113 <https://doi.org/10.1016/j.epsl.2020.116324>
- 1114 Robertson, J. C., Barnes, S. J., & Le Vaillant, M. (2016). Dynamics of magmatic sulphide  
1115 droplets during transport in silicate melts and implications for magmatic sulphide ore  
1116 formation. *Journal of Petrology*, 56(12), 2445–2472. <https://doi.org/10.1093/petrology/egv078>
- 1117 Rubin, K. H., van der Zander, I., Smith, M. C., & Bergmanis, E. C. (2005). Minimum speed limit  
1118 for ocean ridge magmatism from  $^{210}\text{Pb}$ – $^{226}\text{Ra}$ – $^{230}\text{Th}$  disequilibria. *Nature*, 437(7058), 534–  
1119 538. <https://doi.org/10.1038/nature03993>
- 1120 Shaw, C. S. J., Lebert, B. S., & Woodland, A. B. (2018). Thermodynamic modelling of mantle-  
1121 melt interaction evidenced by veined wehrlite xenoliths from the Rockeskyllerkopf Volcanic  
1122 Complex, West Eifel volcanic field, Germany. *Journal of Petrology*, 59(1), 59–86.  
1123 <https://doi.org/10.1093/petrology/egy018>
- 1124 Smith, P. M., & Asimow, P. D. (2005). Adibat-1ph: A new public front-end to the MELTS,  
1125 pMELTS, and pHMELTS models. *Geochemistry, Geophysics, Geosystems*, 6(2), 1–8.  
1126 <https://doi.org/10.1029/2004GC000816>

- 1127 Spiegelman, M., Kelemen, P. B., & Aharonov, E. (2001). Causes and consequences of flow  
1128 organization during melt transport: The reaction infiltration instability in compatible media.  
1129 *Journal of Geophysical Research: Solid Earth*, *106*(B2), 2061–2077.  
1130 <https://doi.org/10.1029/2000JB900240>
- 1131 Sun, W. (2007). Kinetics for coarsening co-controlled by diffusion and a reversible interface  
1132 reaction. *Acta Materialia*, *55*(1), 313–320. <https://doi.org/10.1016/j.actamat.2006.07.045>
- 1133 Sundberg, M., Hirth, G., & Kelemen, P. B. (2010). Trapped melt in the Josephine peridotite:  
1134 Implications for permeability and melt extraction in the upper mantle. *Journal of Petrology*,  
1135 *51*(1–2), 185–200. <https://doi.org/10.1093/petrology/egp089>
- 1136 Sweeney, S. M., & Martin, C. L. (2003). Pore size distributions calculated from 3-D images of  
1137 DEM-simulated powder compacts. *Acta Materialia*, *51*(12), 3635–3649.  
1138 [https://doi.org/10.1016/S1359-6454\(03\)00183-6](https://doi.org/10.1016/S1359-6454(03)00183-6)
- 1139 von Bagen, N., & Waff, H. S. (1986). Permeabilities, interfacial areas and curvatures of partially  
1140 molten systems: Results of numerical computations of equilibrium microstructures. *Journal of*  
1141 *Geophysical Research*, *91*(B9), 9261–9276. <https://doi.org/10.1029/JB091iB09p09261>
- 1142 Wagner, C. (1961). Theorie der Alterung von Niederschlägen durch Umlösen (Ostwald-  
1143 Reifung). *Zeitschrift Für Elektrochemie, Berichte Der Bunsengesellschaft Für Physikalische*  
1144 *Chemie*, *65*(7–8), 581–591. <https://doi.org/10.1002/bbpc.19610650704>
- 1145 Wang, C., Liang, Y., Xu, W., & Dygert, N. (2013). Effect of melt composition on basalt and  
1146 peridotite interaction: Laboratory dissolution experiments with applications to mineral  
1147 compositional variations in mantle xenoliths from the North China Craton. *Contributions to*  
1148 *Mineralogy and Petrology*, *166*(5), 1469–1488. <https://doi.org/10.1007/s00410-013-0938-6>
- 1149 Wang, C., Liang, Y., Dygert, N., & Xu, W. (2016). Formation of orthopyroxenite by reaction  
1150 between peridotite and hydrous basaltic melt: an experimental study. *Contributions to*  
1151 *Mineralogy and Petrology*, *171*(8–9). <https://doi.org/10.1007/s00410-016-1287-z>
- 1152 Wang, C., Lo Cascio, M., Liang, Y., & Xu, W. (2020). An experimental study of peridotite  
1153 dissolution in eclogite-derived melts: Implications for styles of melt-rock interaction in  
1154 lithospheric mantle beneath the North China Craton. *Geochimica et Cosmochimica Acta*, *278*,  
1155 157–176. <https://doi.org/10.1016/j.gca.2019.09.022>

- 1156 Wang, J., Xiong, X., Zhang, L., & Takahashi, E. (2020). Element loss to platinum capsules in  
1157 high-temperature-pressure experiments. *American Mineralogist*, *105*(10), 1593–1597.  
1158 <https://doi.org/10.2138/am-2020-7580>
- 1159 Wang, K.-L., O'Reilly, S. Y., Griffin, W. L., Pearson, N. J., & Zhang, M. (2009). Sulfides in  
1160 mantle peridotites from Penghu Islands, Taiwan: Melt percolation, PGE fractionation, and the  
1161 lithospheric evolution of the South China block. *Geochimica et Cosmochimica Acta*, *73*(15),  
1162 4531–4557. <https://doi.org/10.1016/j.gca.2009.04.030>
- 1163 Wang, Z. J., & Jin, Z. (2020). Reaction Infiltration Instabilities in Partially Molten Peridotite and  
1164 Implications for Driving the Transport of Sulfide Liquid. *Journal of Earth Science*, *31*(3),  
1165 447–455. <https://doi.org/10.1007/s12583-020-1301-2>
- 1166 Wang, Z. J., Jin, Z., Mungall, J. E., & Xiao, X. (2020). Transport of coexisting Ni-Cu sulfide  
1167 liquid and silicate melt in partially molten peridotite. *Earth and Planetary Science Letters*,  
1168 *536*, 116162. <https://doi.org/10.1016/j.epsl.2020.116162>
- 1169 Yao, Z. S., & Mungall, J. E. (2021). Kinetic controls on the sulfide mineralization of komatiite-  
1170 associated Ni-Cu-(PGE) deposits. *Geochimica et Cosmochimica Acta*, *305*, 185–211.  
1171 <https://doi.org/10.1016/j.gca.2021.05.009>
- 1172 Yao, Z. S., Qin, K. Z., & Xue, S. C. (2017). Kinetic processes for plastic deformation of olivine  
1173 in the Poyi ultramafic intrusion, NW China: Insights from the textural analysis of a ~ 1700 m  
1174 fully cored succession. *Lithos*, *284–285*, 462–476. <https://doi.org/10.1016/j.lithos.2017.05.002>
- 1175 Yao, Z. S., & Mungall, J. E. (2020). Flotation mechanism of sulphide melt on vapour bubbles in  
1176 partially molten magmatic systems. *Earth and Planetary Science Letters*, *542*, 116298.  
1177 <https://doi.org/10.1016/j.epsl.2020.116298>
- 1178 Yao, Z. S., Qin, K. Z., & Mungall, J. E. (2018). Tectonic controls on Ni and Cu contents of  
1179 primary mantle-derived magmas for the formation of magmatic sulfide deposits. *American*  
1180 *Mineralogist*, *103*(10), 1545–1567. <https://doi.org/10.2138/am-2018-6392>
- 1181 Yao, Z. S., Mungall, J. E., & Qin, K. Z. (2019). A preliminary model for the migration of sulfide  
1182 droplets in a magmatic conduit and the significance of volatiles. *Journal of Petrology*, *60*(12),  
1183 2281–2316. <https://doi.org/10.1093/petrology/egaa005>
- 1184 Yao, Z. S., Mungall, J. E., & Jenkins, M. C. (2021). The Rustenburg Layered Suite formed as a  
1185 stack of mush with transient magma chambers. *Nature Communications*, *12*(1), 1–14.  
1186 <https://doi.org/10.1038/s41467-020-20778-w>

- 1187 Yoshino, T., & Watson, E. B. (2005). Growth kinetics of FeS melt in partially molten peridotite:  
1188 An analog for core-forming processes. *Earth and Planetary Science Letters*, 235(1–2), 453–  
1189 468. <https://doi.org/10.1016/j.epsl.2005.04.021>
- 1190 Yoshino, T., Walter, M. J., & Katsura, T. (2003). Core formation in planetesimals triggered by  
1191 permeable flow. *Nature*, 422(6928), 154–157. <https://doi.org/10.1038/nature01459>
- 1192 Yoshino, T., Walter, M. J., & Katsura, T. (2004). Connectivity of molten Fe alloy in peridotite  
1193 based on in situ electrical conductivity measurements: Implications for core formation in  
1194 terrestrial planets. *Earth and Planetary Science Letters*, 222(2), 625–643.  
1195 <https://doi.org/10.1016/j.epsl.2004.03.010>
- 1196 Yoshino, T., Laumonier, M., McIsaac, E., & Katsura, T. (2010). Electrical conductivity of  
1197 basaltic and carbonatite melt-bearing peridotites at high pressures: Implications for melt  
1198 distribution and melt fraction in the upper mantle. *Earth and Planetary Science Letters*,  
1199 295(3–4), 593–602. <https://doi.org/10.1016/j.epsl.2010.04.050>
- 1200 Zhang, Y. (2015). Toward a quantitative model for the formation of gravitational magmatic  
1201 sulfide deposits. *Chemical Geology*, 391, 56–73.  
1202 <https://doi.org/10.1016/j.chemgeo.2014.10.025>
- 1203 Zhang, Y., Walker, D., & Leshner, C. E. (1989). Diffusive crystal dissolution. *Contributions to*  
1204 *Mineralogy and Petrology*, 102(4), 492–513. <https://doi.org/10.1007/BF00371090>
- 1205 Zhang, Y., Ni, H., & Chen, Y. (2010). Diffusion data in silicate melts. *Reviews in Mineralogy*  
1206 *and Geochemistry*, 72(1), 311–408. <https://doi.org/10.2138/rmg.2010.72.8>
- 1207 Zhao, Y., Liu, S. A., Xue, C., & Li, M. L. (2022). Copper isotope evidence for a Cu-rich mantle  
1208 source of the world-class Jinchuan magmatic Ni-Cu deposit. *American Mineralogist*, 107(4),  
1209 673–683. <https://doi.org/10.2138/am-2021-7911>
- 1210 Zhu, W., Gaetani, G. A., Füsseis, F., Montési, L. G. J., & de Carlo, F. (2011). Microtomography  
1211 of partially molten rocks: Three-dimensional melt distribution in mantle peridotite. *Science*,  
1212 332(6025), 88–91. <https://doi.org/10.1126/science.1202221>