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Melt percolation, concentration and dyking in the Hawaiian mantle plume and overriding lithosphere: links to the evolution of lava composition along the volcanic chain

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1 **Melt percolation, concentration and dyking in the Hawaiian mantle plume**
2 **and overriding lithosphere: links to the evolution of lava composition along**
3 **the volcanic chain**

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11 **ABSTRACT**

12 Oceanic island basalts and related magmatic rocks from Hawaii are derived from a
13 compositionally heterogeneous mantle plume. Here we describe how this heterogeneity results
14 from the transport of filaments of a specific composition in the plume, representing a relatively
15 small volume of rocks (~15%) interbedded inside a dry peridotite mantle. Four types of
16 filaments are considered: sub-primitive mantle, ultra-low-velocity zone, fertilized-harzburgite
17 and eclogite type filaments. We present a model that describes the flow within a plume and the
18 stress field in the overriding viscoelastic lithosphere and that can determine, from depth to the
19 surface, the melting rate, composition and trajectory of melts produced within each type of
20 filament. Our model shows that (i) the filaments melt at a depth greater than 5 GPa, where the
21 temperature gap between the solidus and liquidus is narrow (~ 40°C to 80°C), and (ii) the
22 volume of filaments is small relative to the total volume of mantle, which therefore allows the
23 latent heat required for the partial melting to be provided via conduction inside the hot plume.

24 The primitive melts produced inside the filaments, occasionally mixed with the melt derived
25 from an eclogite filament, represent a volume comparable with that expected in a plume
26 composed only of dry peridotite that partially melts to a degree of ~10% at the interface between
27 the spinel and garnet fields (60–70 km depth). In particular, in the centre of the plume, sub-
28 primitive mantle filaments produce up to 30% tholeiite-picrite melts, whereas in fertilized-
29 harzburgite filaments, the mantle melts completely to produce a melt having a meimechite-like
30 composition. A key finding is that the fractional crystallization of these melts likely forms the
31 so-called “primary mantle-derived alkaline magmas” along with dunites and olivine-rich
32 cumulates. Our plume model shows that the mantle flow divides into two parts. The first
33 corresponds to hot flowlines that originate at a depth of ~200 km and at a distance of less than
34 25 km from the plume axis. Along these flowlines, when the mantle reaches a pressure of 5
35 GPa, the partially molten horizon in filaments is sufficiently thick for the interstitial melt to be
36 squeezed out via dykes. This melt eventually ponds as sills in a subrectangular zone that is
37 located inside the overlying lithosphere, between 70 and 50 km depth and centred over a
38 distance of less than 40 km on either side of the axis. This zone is designated as the shield
39 magmatic reservoir. The volatile-rich melt inside the sills infiltrates the surrounding mantle
40 lithosphere and partially melts it. After ~0.1 Ma, the melt resumes its vertical ascent via dykes
41 and eventually ponds and differentiates within subcrustal magma chambers located below
42 active shield volcanoes. This sequence of processes matches the expected volume, petrology
43 and geochemistry recorded for shield volcanoes. The second part of the melt flow does not pond
44 within the shield magmatic reservoir. Rather, the mantle cold flowlines, originating at ~200 km
45 depth and at 25 to 35 km from the plume axis, discharge their interstitial melt through dykes
46 that were initially generated deeper, at ~5 GPa. The melt reaches the Moho at 100 to 150 km
47 from the plume axis, where it forms magmatic bodies within which the melt differentiates. This
48 melt likely represents that observed in pre- and postshield volcanoes. Finally, at ~70 km from

49 the plume axis and at a depth greater than 200 km, the flowlines are subvertical. They then
50 deflect at ~180 km depth and rotate toward the horizontal and eventually transit at 10° to 20°
51 to the horizontal across a ~200 km distance from the axis and reach ~140 km depth. The
52 fertilized-harzburgite and sub-primitive mantle/ultra-low-velocity zone filaments that flow
53 along these elbows partially melt by a few to several percent. The resulting interstitial melt has
54 a kimberlite-like composition. Thereafter, the excess pressure at the top of the filament at ~200
55 km from the axis overcomes the threshold for dyking and thus allows the escape of the
56 interstitial melt via dykes ponding in subcrustal magma chambers or emerging directly at the
57 surface. These melts have a composition similar to that associated with rejuvenated volcanism.
58 We use the nature and the composition of whole erupted magmas and the seismic structure
59 along the Hawaiian chain to validate this model.

60 **KEYWORDS:** meimechite, kimberlite-like melt, oceanic island alkali basalts, Hawaiian
61 plume, heterogeneous mantle, melt percolation

62 INTRODUCTION

63 The long-standing debate regarding the origin of oceanic island basalts (OIBs) remains largely
64 unresolved. In particular, melting processes (pressure, temperature and degree of melting) fail
65 to explain the full spectrum of petrological and geochemical characteristics of Hawaiian
66 magmas. This failure stems from the mantle being composed of streaks of material, which we
67 designate here as “*filaments*”. These filaments are transported by mantle convection and differ
68 in their lithologies and volatile contents (e.g. Farnetani & Hofmann, 2010; Hofmann *et al.*,
69 2011). Accordingly, these materials partially melt at different depths and temperatures, and the
70 resulting interstitial melts percolate within the mantle and eventually escape through fast-
71 propagating dykes before ponding at various depths (Spera, 1987; Grégoire *et al.*, 2006;
72 Soltanmohammadi *et al.*, 2018). Moreover, the interaction of the ponded melts with the

73 surrounding rocks gives rise to new streaks that produce the specific petrological and
74 geochemical signatures of the resulting magmas. Recent progress in the modelling of mantle
75 convection, together with plate tectonics, has advanced the understanding as to how the most
76 important lithological and compositional streaks form within the mantle and then partially melt
77 at depth (e.g. Farnetani & Hofmann, 2010; Hofmann *et al.*, 2011; Ballmer *et al.*, 2013). In
78 contrast, the modelling of melt percolation, migration and concentration within the various
79 streaks during convective transport remains limited. Here we link the nature and composition
80 of erupted magmas along the Hawaiian chain and present more precise melt migration models.
81 Our models lead to reliable quantitative explanations for the final composition of the various
82 Hawaiian magmas. In this paper, we consider the following types of filament and mantle
83 reservoirs:

84 “*Eclogitic*” and “*fertilized-harzburgite filaments*”: A commonly invoked paradigm of both
85 geochemistry and geodynamics is that subduction processes have a major effect on the
86 evolution of mantle composition, in particular for magmas originating from intraplate
87 volcanism (e.g. Sobolev *et al.*, 2005). This paradigm relates to the melting processes and solid-
88 phase changes during subduction that transform subducting crust—mostly oceanic, but also
89 some continental crust—and the associated sediments into eclogites *sensu lato* at >100 km
90 depth (e.g. Gill, 1981; Hacker *et al.*, 2003). These ~5 km thick eclogite layers ultimately detach
91 from the previously attached mantle portion of the lithosphere (e.g. Rüpke *et al.*, 2004). A ~30
92 km thick layer of serpentinites produced during hydrothermalism at the ridge crest along major
93 faults and during plate bending prior to subduction are transformed at depth into harzburgite
94 and are refertilized, in particular by volatile and alkaline elements (Elthon, 1992; Ross *et al.*,
95 2005; Faccenda, 2014; DeFelice *et al.*, 2019). We refer to these subduction-derived recycled
96 serpentinites as fertilized harzburgite, possibly associated with recycled non-serpentinized
97 oceanic peridotites or mantle wedge-related peridotites and serpentinites. Both the eclogites

98 and the fertilized harzburgites are particularly stable through the lower mantle, but become
99 unstable when convection returns them to the upper mantle within or above the mantle transition
100 zone (MTZ) (e.g. Hirschmann, 2006; Motoki & Ballmer, 2015; Soltanmohammadi *et al.*, 2018).

101 “*Sub-primitive mantle filament*”: The preservation of primitive mantle geochemical
102 signatures, e.g. high He^3/He^4 values and $\mu^{182}\text{W} \sim 0$ (Mundl-Petermeier *et al.*, 2020), of some
103 plume-derived lavas (e.g. OIBs) provide evidence for the existence (survival) of a primitive
104 mantle reservoir. This mantle reservoir is believed to be in the lower mantle. It eventually
105 penetrates the upper mantle via deep-rooted mantle plumes (Courtillot *et al.*, 2003;
106 McNamara, 2019). In this text, we designate this reservoir as the sub-primitive mantle
107 reservoir. In particular, the sub-primitive mantle reservoir used in this study has a lherzolite-
108 like composition but is considered relatively rich in volatile, alkaline and incompatible
109 elements. Moreover, the geochemical signature indicates this reservoir was never processed
110 by melting or degassing after the formation of the Earth’s core (e.g. Kurtz *et al.*, 1983;
111 Collerson *et al.*, 2010; Mundl-Petermeier *et al.*, 2020). The presence of a sub-primitive mantle
112 reservoir indicates that the convective processes that tend to homogenize the entire Earth
113 mantle did not affect this reservoir. This idea is supported by seismic tomography that shows
114 the mantle convective plumes in the lower mantle being very broad; thus, the viscosity in the
115 lower mantle is several orders of magnitude greater than that within the upper mantle. The
116 convective velocities are therefore reduced in the lower mantle (French & Romanowicz,
117 2015; Liu *et al.*, 2018). Consequently, the overturn time of convection for an entire mantle
118 cell represents ~ 1 Ga, and a large volume of the lower mantle stagnates mainly inside the
119 several 1000 km wide interior of these convective cells, thereby preventing the mechanical
120 homogenization of the sub-primitive mantle with dry peridotite. Eventually, the shear at the
121 borders of the plume entrains some sub-pristine mantle inside the whole-mantle convection
122 into the main plume flow. This process forms a sub-primitive filament having a thickness

123 comparable to the vertical dynamic boundary layer surrounding the mantle plume, i.e. a few
124 to several tens of kilometres (Farnetani *et al.*, 2002).

125 “*ULVZ filament*”: Seismic wave reflection along the core-mantle interface shows the presence
126 of patchy ultra-low-velocity zones (ULVZs) (Li *et al.*, 2017). These zones are thin (~5–40 km
127 thick and several hundred kilometres wide) and are generally found near or within large low-
128 seismic-velocity provinces (LLSVPs). These provinces represent the ~500 km thick hot
129 boundary layer of whole-mantle convection, a layer that contains the roots of various plumes
130 set below the two antipolar African and Central Pacific superswells (French & Romanowicz,
131 2015, McNamara *et al.*, 2019). Because mantle flow along the core converges toward the axis
132 of the plume, the ULVZs pile up near the root of the plume as the hottest region in the lower
133 mantle. Accordingly, the ULVZs are linked to partial melting. As similar structures—
134 although having lower S-velocities—have also been detected well outside LLSVPs,
135 particularly beneath subduction zones, it is thought that the ULVZs are not only partially
136 melted but also compositionally heterogeneous, i.e. consist of eclogite and/or volatile-rich
137 mantle (Collerson *et al.*, 2010; Andraut *et al.*, 2014; Nomura *et al.*, 2014). In addition, some
138 volatiles and metals are directly provided to the mantle by the core. This is the case for CO₂
139 (Collerson *et al.*, 2010; Dobretsov & Shatskiy, 2012), He (showing an elevated ³He/⁴He
140 ratio), and also evidenced by highly negative $\mu^{182}\text{W}$ (Mundl-Petermeier *et al.*, 2020), Fe (Mao
141 *et al.*, 2017) and ¹⁸⁶Os (Brandon *et al.*, 1999). In short, all necessary conditions for developing
142 ULVZ filaments are achieved at the root of the Hawaiian plume. These filaments are likely
143 entrained upward by the plume and eventually solidify well above the core-mantle interface
144 (Collerson *et al.*, 2010). This process would form solid volatile-rich filaments having a
145 lherzolite-like composition with an elevated ³He/⁴He ratios and a $\mu^{182}\text{W} < 0$. These filaments
146 are thereafter propelled upward by the plume flow (Farnetani *et al.*, 2018). As we do not have
147 more quantitative information related to the true composition of the ULVZ filament, we

148 assume that it contains the same volatile content and $^3\text{He}/^4\text{He}$ ratio as a sub-primitive mantle
149 filament.

150 “*The ambient mantle*”: A well-accepted paradigm is that the upper mantle is mostly depleted,
151 thought to represent the mechanically homogenized MORB mantle reservoir (Harpp & Weis,
152 2020). According to these authors, the Hawaiian plume pushes it apart to collide with the
153 lithosphere. Nonetheless, the model does not indicate what occurs after more than 100 Ma of
154 quasi-steady life of the plume when a huge volume of the MORB dry mantle reservoir,
155 representing the volume of a ~ 2000 km section (assuming that the ascending velocity of the
156 plume at the upper mantle/lower mantle interface is 10 cm/yr), has been driven back into the
157 lower mantle. To address this problem, we can first turn to the case of the Asian continent,
158 which is surrounded by oceanic plates including the Pacific. An oceanic plate does not
159 penetrate inside the lower mantle and eventually floats along the MTZ below the China block
160 up to 2000 km distant from the Japan arc. There, it is highly likely that the upper convective
161 mantle remained isolated from the lower convective mantle during an orogeny, i.e. during
162 ~ 200 Ma (Agrusta, *et al.*, 2017). Then, because the upper-mantle convection is much faster
163 than that of the lower mantle (~ 10 cm/yr), the overturn time for convection takes only ~ 20
164 Ma. After ~ 200 Ma it has therefore overturned 10 times, sufficient to mechanically
165 homogenize both eclogites and fertilized harzburgites to eventually attain a centimetre-scale
166 size (van Keken *et al.*, 2003). The marble-cake model, used to explain the formation of
167 several centimetre-thick pyroxenite bands, supports this paradigm (Allegre & Turcotte, 1986).
168 This model leads to the idea of a specific composition of a fully homogenized mantle low in
169 volatiles and other incompatible elements, sometimes designated as the DMM (depleted
170 MORB mantle). In this text, we name it the “dry peridotite” reservoir. Although, at present,
171 the stratification of convection on both sides of the mantle appears relatively exceptional, it
172 does likely still occur beneath both the west coast of the United States (Rigo *et al.* 2015) and

173 the central Atlantic (Courtillot *et al.*, 2003). It is suspected, however, that before 3 Ga,
174 layering was systematic (Breuer and Spohn, 1995). The key problem is to determine what
175 happened after 3 Ga when the lower mantle flushed inside the upper mantle (i.e. Breuer &
176 Spohn, 1995). Given the data considered in this study, it is likely that borders of the entire
177 mantle convective cell, e.g. the Hawaiian case, are constituted of dry peridotite, i.e. the axis of
178 the plume, the cell top below the lithosphere, the vertical return flow in the mantle and along
179 the cell bottom above the core. Finally, the cylindrical interior of the convective cell must
180 contain an essential volume of primitive mantle. In this context, the dry peridotitic border of
181 the entire mantle convection cell constitutes the envelope in which the eclogite, fertilized-
182 harzburgite, sub-primitive mantle and ULVZ filaments circulate. In particular, these filaments
183 all converge toward the axis of the Hawaiian plume where they rise inside the plume towards
184 the sublithospheric mantle.

185 The aim of our paper is to detail what happens to these filaments when they partially melt.
186 The present model agrees well with the latest geochemical conception of the Hawaiian plume.
187 For instance, the main large-scale heterogeneity of the shield magmatism, representing by far
188 the largest volume of Hawaiian magmas, is demonstrated by the compositional differences of
189 lava sources for volcanoes located along a line linking Loihi and Kauai (Loa trend line) and
190 those located along a line joining Kilauea to East Moluka'i (Kea trend line) (e.g. Dana, 1849;
191 Appleman and Dana, 1987; Frey *et al.*, 2016). The radiogenic isotopic ratios of most
192 Hawaiian volcanoes indicate that the Loa- and Kea-type mantles originate from two
193 geochemically distinct deep mantle reservoirs (Weis *et al.*, 2011; Harrison *et al.*, 2017; Weis
194 *et al.*, 2020). The Loa-type reservoir is proposed to originate from the central Pacific LLSVP,
195 particularly rich in ULVZ filaments, whereas the Kea-type reservoir originates from the
196 "Pacific mantle," away from the LLSVP (see Fig. 5 of Weis *et al.*, 2011; Fig. 4 of French and
197 Romanowicz, 2015; Fig. 4 of Harrison *et al.*, 2017; and Fig. 5 of Weis *et al.*, 2020). Note that

198 for these authors, the Pacific mantle likely represents the relatively cold, ~500 km thick,
199 bottom flow that transports eclogite, fertilized-harzburgite and sub-primitive mantle filaments
200 toward the axis of the Hawaiian plume set along the border of the central Pacific LLSVP.
201 They consider that the geochemical characteristics of each volcano are a direct fingerprint of a
202 specific filament, whereas our approach holds that the Hawaiian lavas are derived from
203 various filaments and from the melts' percolation, ponding/mixing and interaction with the
204 surrounding mantle and/or overriding lithosphere.

205 Because of the thick lithosphere overlying the Hawaiian plume, partial melting of the dry
206 peridotite inside the plume is limited if it occurs at all (as will be discussed below). In the
207 Hawaiian context, this indicates that the primitive melts of Hawaii do not originate from a
208 normal dry mantle. Therefore, it implies that even in the hot centre of the plume, the mantle
209 must contain a notable quantity of volatiles to partially melt (Wallace, 1998; Putirka, 1999; Li
210 *et al.*, 2000). Further, the composition of some metasomatized sources indicates that the sources
211 originate either from fertilized harzburgite (Dawson *et al.*, 1980; Grégoire *et al.*, 2002) or from
212 the sub-primitive mantle (Collerson *et al.*, 2010). Estimates of the volatile content of the
213 fertilized-harzburgite and sub-primitive-mantle components of the Hawaiian plume range from
214 450 ppm to 750 ppm for H₂O and up to 1300 ppm for CO₂ (Dixon *et al.*, 1997; Wallace, 1998;
215 Marty, 2012; Tucker *et al.*, 2019). Therefore, inside the lower mantle, the interstitial H₂O
216 concentrates in dense hydrous magnesium silicate phases (DHMS), and CO₂ is reduced to
217 carbide or diamond. When the mantle migrates up within the centre of the plume and crosses
218 the interface between the lower and upper mantle, its temperature is very high, likely ranging
219 between 1800°C and 2100°C (Liu *et al.*, 2018). The DHMS, i.e. phases D and B identified as
220 holding most of the water in the deep mantle, therefore melt (Ringwood and Major, 1967; Liu,
221 1987; Gasparik, 1993; Ohtani *et al.*, 2001). In addition, during the ascent of the plume within
222 the horizon lying between the ringwoodite-perovskite and the garnet-perovskite phase

223 transitions, the production of majorite likely leads to the oxidation of C (Stagno & Frost, 2010;
224 Kiseeva *et al.*, 2018). Because the melting temperature of the carbonates at these depths is less
225 than 1800°C to 2100°C, we deduce that a H₂O- and CO₂-rich melt is produced inside the garnet-
226 perovskite transition zone. The plume model of Ballmer *et al.* (2013) suggests that the mantle
227 plume flow contains 15% and ~20% of eclogite and hydrated lherzolite filaments, respectively,
228 to explain the produced crust volume and the distribution of the volcanoes along the Hawaiian
229 chain. Furthermore, this model assumes that the eclogite transported by the Hawaiian plume
230 melts at a pressure of ~5 GPa (~150 km depth) and then infiltrates the surrounding peridotites
231 to form pyroxenites, which are suspected to be a major source of OIB melts (e.g. Cordery *et al.*,
232 1997; Sobolev *et al.*, 2005; Ballmer *et al.*, 2013). Considering the various interactions between
233 melts and mantle reservoirs at a relatively crude scale, this model does not account for the large
234 diversity in terms of composition or the geochemical signature of lavas observed among the
235 various Hawaiian volcanoes (e.g. Frey & Rhodes, 1993; Bianco *et al.*, 2008).

236 Following the detailed studies of Frey & Clague (1983) and Clague (1987), all studies that
237 focused on the geological, petrographical, mineralogical, geochemical and isotopic features of
238 Hawaiian magmatic rocks agree that the Hawaiian volcanoes are affected by four magmatic
239 stages (see Clague & Sherrod (2014) for a synthesis): (i) a preshield stage, characterized mainly
240 by non-feldspathoidal silica-undersaturated to saturated alkaline magmas as well as tholeiite-
241 transitional magmas; (ii) a shield stage having eruptions of volumetrically large, dominant
242 silica-saturated tholeiitic-picritic lavas, which represent 98% of the plume-driven melts; (iii)
243 occasionally, a postshield stage, consisting of non-feldspathoidal silica-undersaturated alkaline
244 lavas and highly differentiated alkaline lavas; and (iv) a rejuvenated stage, comprising
245 feldspathoidal silica-undersaturated alkaline to highly alkaline lavas. The four stages reflect
246 variations in magma supply over time from the underlying mantle—as well as variations in the
247 degree of partial melting of this mantle—and therefore the stages reflect variable amounts and

248 rates of heat supply to the lithosphere when the latter overrides the hot spot (e.g. Clague, 1987;
249 Ribe & Christensen, 1999; Clague & Sherrod, 2014). Hofmann *et al.* (2011) developed the
250 concept that the along-track evolution of the composition of these different types of volcanoes
251 results from (i) the low-viscosity metasomatic melts caused by the dehydration and
252 decarbonation of the mantle and (ii) the more viscous tholeiitic melts produced by
253 decompression and partial melting in the elongated tail of the plume driven by plate drift. Here
254 we will illustrate that the production, percolation and local concentration of melts differ within
255 each type of filament. Then, assuming a plume overlain by a motionless plate, we calculate the
256 final trajectory of these concentrated melts via dykes up to their ponding in subsurface magma
257 chambers. Finally, we show that the eventual escape to the surface of these melts greatly
258 improve our understanding of the causes of the along-track evolution of the composition of
259 Hawaiian volcanism. In the present study, we first discuss the link between a 2D numerical
260 convective model of the Hawaiian plume and the trajectories of dykes that eventually cross the
261 overriding lithosphere. Then, we study the relationship between the temperature field and the
262 melting rate in sub-primitive mantle/ULVZ, fertilized-harzburgite and eclogite filaments. We
263 compare this relationship with that of dry peridotite, which is assumed to represent the bulk
264 composition of the mantle flowing inside the plume. Subsequently, we decipher how the
265 composition of primitive melts produced in sub-primitive mantle, fertilized-harzburgite and
266 eclogitic filaments evolves during their percolation and eventual concentration below the
267 lithosphere. We describe how various melts mix during their respective ascents through the
268 lithosphere and explore how these melts then evolve during their partial crystallization and
269 eventual eruption from the various volcanoes along the Hawaiian chain. Further, we evaluate
270 the volume of recycled sub-primitive mantle filaments required to explain the crustal volume
271 produced along the Hawaiian chain. Finally, we combine seismic tomographic data and
272 geochemical data to evaluate the melt compositions observed in preshield, shield, postshield

273 and rejuvenated-stage volcanoes of the Hawaiian chain. Specifically, our model explains the
274 development of rejuvenated volcanism around Oahu, i.e. 300 km from Kilauea. Furthermore,
275 we demonstrate that the elongate type of plume stem predicted by Ballmer *et al.* (2013) explains
276 the survival of Niihau and Kaua'i rejuvenated volcanism for up to several 100 kyrs and to a
277 distance of more than 500 km from Kilauea.

278 **STRESS FIELD ACROSS THE LITHOSPHERE AND THE UNDERLYING** 279 **ASTHENOSPHERE IN HAWAII**

280 Magma reaching the surface from below a thick lithosphere, such in the case of the Hawaiian
281 volcanoes, requires dykes and fractures to cross the lithosphere. The trajectories of dykes and
282 fractures are thus closely related to the stress field of the vertical push of the lithosphere,
283 induced by the plume at the top of the asthenosphere. Fig. 1 represents a 2D numerical model
284 of the Hawaiian plume that uses a temperature-dependent, power-law, dry elastoplastic
285 rheology for the mantle and where any effects due to partial melting are assumed to be
286 negligible. More precisely, we fix the shear modulus value at 50 GPa and use the common
287 dislocation-creep law of dry dunite determined by Chopra and Peterson (1984) to compute
288 viscosity. The model describes the development of a plume rising from a 100 km wide hot
289 (~1650°C) region at the base of the upper mantle (Gerbault *et al.*, 2017). The top surface
290 represents the seafloor, which is considered to be motionless. This figure presents a snapshot
291 of the thermal, stress and effective viscosity fields a few million years after the collision of the
292 plume with a ~100 Ma old lithosphere. Figs 1b and 1d display what we designate in the
293 following sections as the bottom of the “lithosphere.” We use the term lithosphere as designated
294 by its rheological definition, i.e. the top zone where the mantle is weakly or not at all
295 deformable. It represents the elastic zone, where heat is transported exclusively by conduction
296 (~80 km thick for Hawaii), plus the highly viscous layer, where the plume flow is sluggish

297 enough to have the distribution of temperature controlled essentially by conduction. In our
298 model, the bottom of the lithosphere corresponds to a horizon where the viscosity of the mantle
299 η is $\sim 10^{20}$ Pa·s, and the asthenosphere is a zone where the viscosity is slightly lower than 10^{20}
300 Pa·s. The bottom of the lithosphere lies above ~ 120 km (~ 4 GPa) and is not flat. At the plume
301 axis, the horizon attains a depth of 120 km, where the temperature is $\sim 1200^\circ\text{C}$; along its sides,
302 however, it is about 20 km thinner and has a temperature $< 1100^\circ\text{C}$. Such variability at the base
303 of the lithosphere is specific to the non-linearity of rheology. This non-linearity forms a quasi-
304 motionless region between the lithosphere and the zone where the flow of the 50 km radius
305 plume rotates to the horizontal, and it causes a 1–2 order of magnitude increase in viscosity
306 across a triangular region at ~ 100 km depth over a distance less than ± 50 km from the plume
307 axis (Fig. 1b). Below this region, the mantle flow within the plume is purely vertical within less
308 than ± 25 km from its axis, resulting in a vertically oriented maximum compressive stress σ_1 to
309 the base of the lithosphere (Fig. 1a and d). In a zone within the lithosphere, at a horizontal
310 distance of less than ~ 40 km from the plume axis, the direction of σ_1 becomes horizontal above
311 ~ 80 km depth. From 70 to 50 km depth, σ_1 is completely horizontal, and it returns to the vertical
312 above 50 km (Fig. 1a and d). These superimposed horizons of horizontal compression (σ_1
313 horizontal) and extension (σ_1 vertical) result from the bending of the elastic part of the
314 lithosphere induced by the upward push of the plume. During its rotation to the horizontal, the
315 flow divides into two regimes (Fig. 1d). The flow originates from a depth of > 200 km, where it
316 is vertically oriented at ~ 25 km from the plume axis. Thereafter, it rotates to the horizontal and
317 attains a depth of ~ 150 km at ~ 50 km from the plume axis. Finally, the flow rises along a very
318 gentle slope to reach 120 km depth at 150 km from the plume axis. In the following text, we
319 designate this flowline as the “hot border flowline” of the plume because it initially follows the
320 1500°C isotherm. Above that flowline, the intensity of the flow diminishes, as it is slowed down
321 by the motionless lithosphere. There, stress is oriented at $> 45^\circ$ to the horizontal to become

322 oriented towards the direction of flow. This implies that, above the hot border flowline, dyking
323 may drive melt upward. The flowline that moulds the border of the plume at a distance of 50
324 km from its axis is the cold border of the plume (Fig. 1d). The flow flattens to the horizontal at
325 ~70 km from the plume axis and 180 km below the surface. Around that flowline, vertical
326 velocity shows limited variability, and the horizontal velocities decrease moving away from the
327 axis, implying that the orientation of σ_1 at this location is quasi-horizontal and that the condition
328 is fulfilled for the development of sills rather than dykes. As we will see later, all of these
329 conditions are important for evaluating melt trajectories within the plume.

330 Our model is comparable to the 3D Hawaiian plume models of Ballmer *et al.* (2011, 2013), a
331 major difference occurring in the investigated mantle rheological laws. In Ballmer *et al.* (2011,
332 2013), the rheology of the mantle is dependent on temperature, pressure, composition, H₂O
333 content and melting rate but is independent of stress and does not integrate an elastic
334 component. Therefore, this model describes the flow and temperature only for depths >90 km,
335 i.e. well below the elastic lithosphere, which is ~30 km thick. Finally, the 3D box models of
336 Ballmer *et al.* (2011, 2013) are overlain by a top interface moving at the velocity of the Pacific
337 plate (8 cm/yr) and are partially open from below and from the sides to permit the free vertical
338 rise of the plume from the lower mantle. Given the need to access the stress field through the
339 entire lithosphere, we concluded that only the upper-mantle plume model of Gerbault *et al.*
340 (2017) can be used to decipher how the melts that accumulate below the lithosphere can make
341 their way through the lithosphere and erupt at the surface. Nonetheless, many important
342 characteristics of our model are also found in those of Ballmer *et al.* (2011, 2013); in the latter
343 models, for example, the bottom of the lithosphere, defined as the $\sim 10^{20}$ Pas high- versus low-
344 viscosity interface, is also ~120 km deep, and the temperature and effective viscosity fields of
345 the plumes are very similar to those of Fig. 1. However, a detailed comparison of the flow
346 derived from our model with that of these authors shows notable differences. First, in the

347 Ballmer *et al.* (2011, 2013) models, the viscosity inside the top diverging portion of the plume
348 displays a strong lateral variation of amplitude. In the 2011 model, for instance, the minimum
349 viscosity at the core of the plume is 10^{18} Pa·s and then grades up to 3×10^{19} Pa·s at the border
350 of the plume stem. Further, outside the plume, the viscosity steadily rises with depth—from 240
351 km to 420 km—where it reaches a fairly uniform amplitude of 10^{20} Pa·s, i.e., the value
352 recovered in our model throughout the entire asthenosphere (Fig 1). Such values likely result
353 from the lack of a stress-dependent rheology in their models; a lower viscosity due to
354 temperature is not compensated by an increased viscosity owing to a drop in stress within low-
355 viscosity areas (Christensen, 1984). Further, the specific lateral boundary conditions adopted
356 by these authors drive the downstream tail of the plume parallel to but faster than plate motion;
357 thus, the width of the plume becomes similar in dimension to the height of the convective zone
358 (~1000 km). This effect could explain volcanic activity along the Hawaiian track and, in
359 particular, the rejuvenated volcanism up to 500 km from Kilauea (Hofmann *et al.*, 2011;
360 Ballmer *et al.*, 2011). Seismic tomography below the Hawaiian swell confirms the presence
361 below the lithosphere of a ~1000 km long low anomaly, oriented parallel to the plate drift, that
362 originates at the east of the island of Hawaii (Wolfe *et al.*, 2009; 2015). Such an anomaly likely
363 includes the downflow long tails recovered in the Hofmann *et al.* (2011) and Ballmer *et al.*
364 (2011) models. The low-velocity seismic anomaly is, however, clearly much deeper than the
365 one displayed in these models because it reaches a wide horizon at the base of the upper mantle.
366 It indicates that a several hundred kilometre-long hot sheet running parallel to the drift direction
367 crosses the main upper mantle downflow of the island of Hawaii (Agius *et al.*, 2017). Seismic
368 tomography models show that such structures exist in multiple zones beneath the Pacific plate
369 and likely represent ascending limbs of 1000 km wide rolls having axes of rotation parallel to
370 the drift direction (French *et al.*, 2013; French & Romanowicz, 2014). These observations are
371 consistent with 3D convective models confined to the upper mantle showing that the plate drift

372 (i) shifts, in the upflow direction, the axis of the plumes. These plumes eventually collide with
373 the downwelling cold limbs of polygonal convective cells; and (ii) eventually lines up the hot,
374 buoyant plumes inside hot sheets crossing the entire upper mantle (Rabinowicz *et al.*, 1990).
375 Ballmer *et al.* (2013) introduced a denser eclogitic component to their plume model that melts
376 above 410 km. This melting leads to the pooling of the plume inside a long horizon above this
377 depth and eventually to a double layering of the convective flow above and below 410 km
378 depth. It ends with the generation of ~300 km wide rolls having their axes parallel to the plate
379 drift; these rolls are eventually superimposed by the already complex 3D-plume flow (Ballmer
380 *et al.*, 2013; Dubuffet *et al.*, 2000). This result demonstrates (i) why a narrower and much faster
381 vertical plume anchored above 410 km depth requires the relay of the deeper plume and
382 eventually collides at 120 km depth with the bottom of the overriding lithosphere; and why (ii)
383 the downstream tail is so long and locally flows much faster than the top plate. To explain the
384 volume of melt of the shield volcanoes, the vertical velocity of the axis of the narrowed plume
385 must reach ~1 m/yr compared with the ~8 cm/yr of the Pacific plate drift. Therefore, plate drift
386 eventually plays a negligible dynamic effect at the very top of the narrowed plume and explains
387 a posteriori the strong analogy between the top plume morphology of the Ballmer *et al.* (2013)
388 model (size, temperature, viscosity fields) and our Fig. 1 model topped by a motionless
389 lithosphere.

390 **ADIABATIC TEMPERATURE FIELD INSIDE THE PLUME AND MELTING**

391 **RATES IN THE ASTHENOSPHERE ABOVE 5 GPA**

392 In this study, we assume that (1) the H₂O and CO₂ concentrations in sub-primitive/ULVZ and
393 fertilized-harzburgite filaments are 500 ppm and 1000 ppm, respectively, in agreement with the
394 estimate of the mantle source contents of Hawaiian lavas studied by Dixon *et al.* (1997); (2) the
395 eclogite filaments are formed by material that was fully dehydrated during subduction (e.g.

396 Godard, 2001); (3) the temperature at the border of the plume follows the mantle adiabat, as
397 determined by Katsura *et al.* (2010), which has a potential temperature at atmospheric pressure
398 of 1360°C and a 0.45°C/km gradient above the MTZ (Fig. 2). Finally, the most controversial
399 point concerns the temperature profile along the plume axis. Here, we adopt the same adiabatic
400 gradient as that of the plume border (0.45°C/km), whereas the potential mantle temperature
401 inside the plume axis is assumed to exceed the adiabatic temperature by several hundred
402 degrees. For the consequences of the modification of those assumed values within reasonable
403 ranges, we refer to supplementary material A1.

404 By using a thermometric method based on olivine-liquid equilibria ($K_D(\text{Fe-Mg})_{\text{ol-liq}}$), Putirka
405 (2005) inferred a potential axial temperature for the Hawaiian mantle plume of 1688°C ± 50°C.
406 More recent studies have shown that this approach is characterized by large uncertainties and,
407 in the case of the Putirka (2005) method, could overestimate the potential mantle temperature
408 (Herzberg *et al.*, 2007; Herzberg, 2011; Matthews *et al.*, 2016). Fig. 2 displays the dry solidus
409 of lherzolite, as determined by Herzberg *et al.* (2010), which shows that (i) at a depth
410 corresponding to 3 GPa, the dry solidus is ~1480°C, i.e. ~200°C lower than the lowest estimate
411 of the axial temperature; (2) at 5 GPa the dry lherzolite solidus is ~1670°C, below the minimum
412 axial temperature estimated by Putirka (2005) (1705°C or a potential temperature of 1640°C);
413 and finally (3) at 8 GPa depth, the solidus crosses the maximum axial temperature at ~1850°C.
414 As well, at 3 GPa, 5 GPa and 8 GPa, the excess temperatures with respect to the lherzolic
415 solidus required to reach the cpx-out threshold are 125°C, ~80°C and ~40°C, respectively.
416 These estimates are based on Eqn. 9 and the parameter values given by Katz *et al.* (2003) and
417 assume that cpx-out occurs at a degree of melting ~35% in agreement with the data of Jaques
418 & Green (1980). Using these last numbers, we can deduce that at 5 GPa, the cpx-out of dry
419 lherzolite is 1670°C + 80°C = 1750°C, a value in between the extreme values (1705°C and
420 1805°C) of the axial temperature of Putirka (2005) (Fig. 2). Furthermore, at greater depth, that

421 threshold notably exceeds the maximum possible temperature along the plume axis; for
422 example, at 8 GPa it is $\sim 1850^{\circ}\text{C} + 40^{\circ}\text{C} = 1890^{\circ}\text{C}$, whereas the maximum possible axial
423 temperature is 1850°C (Fig 2). It is important to note that at 3 GPa the situation is the opposite:
424 the range of possible axial temperature ($1680\text{--}1780^{\circ}\text{C}$) notably exceeds the estimate for cpx-
425 out (here $\sim 1480^{\circ}\text{C} + 125^{\circ}\text{C} = \sim 1600^{\circ}\text{C}$). Considering that the Hawaiian plume is overlain by a
426 120 km thick (4 GPa) plate, the temperature in the plume at 3 GPa is $<1200^{\circ}\text{C}$, implying that
427 no dry melting is possible at this depth. Between 4 GPa and 5 GPa depth, the situation is more
428 difficult to evaluate, because both in our model and that of Ballmer *et al.* (2013) the convective
429 boundary layer, separating the centre of the plume from the lithosphere, is located at these
430 depths. According to these models, however, the temperature in the boundary layer is lower
431 than 1600°C , which corresponds to the solidus of the mantle only at 4 GPa. Such an estimate
432 opens the possibility that dry melting may occur at these depths; regardless, this dry melting
433 must be quite limited. Finally, all these temperatures indicate that between 5 and 8 GPa, a
434 marked melting of dry lherzolite is only possible if one assumes an axial temperature exceeding
435 by several 10°C the lower range of that estimated by Putirka (2005), which, according to Fig.
436 2, corresponds to a potential temperature of 1640°C . These estimates show that the melting rate
437 in the plume axis below the 4 GPa thick lithosphere is very low but that, if the potential
438 temperature exceeds 1640°C by several tens of degrees, significant volumes of tholeiitic or
439 picritic melts must be produced at about 5 GPa or above. This scenario contradicts the
440 petrological and geochemical analyses of the shield-stage melts of Hawaii, which show that
441 they result essentially from a partial-melting process occurring along the spinel/garnet interface,
442 i.e. at 70 km depth, and at a potential temperature of 1500°C (McKenzie & Bible, 1988;
443 McKenzie & O'Nions, 1991). Accordingly, we assume here that the potential temperature at
444 the axis of the plume follows the potential temperature of 1640°C , for which dry lherzolite

445 mantle melting is expected to be negligible or even impossible immediately below the
446 lithosphere.

447 The presence of CO₂ decreases the solidus of lherzolite by up to 600°C (Dasgupta &
448 Hirschmann, 2006). Further, the solidus of mantle peridotites is nearly independent of the bulk
449 CO₂ concentrations of the mantle, provided CO₂ exceeds 5 ppm (Dasgupta & Hirschmann,
450 2007). It is inferred that the CO₂-rich lherzolite solidus profile lies below the adiabatic
451 temperature through the entire sublithospheric upper mantle (Fig. 2). However, the composition
452 of interstitial melt depends strongly on temperature. When the temperature is much lower than
453 that of the dry mantle solidus, the composition of the interstitial melt is carbonatitic, i.e. the
454 CO₂ concentration of the melt in equilibrium with the rock represents ~40% of the melt mass.
455 For temperatures relatively close to that of the dry solidus, however, kimberlitic-like melts with
456 ~10% CO₂ are produced (Dalton & Presnall, 1998). More precisely, at 5 GPa and 10 GPa, i.e.
457 ~150 and ~300 km depth, respectively, the kimberlite-like interstitial melt forms at ~1700°C
458 and above 1800°C, respectively (Brey *et al.*, 2009; Shatskiy *et al.*, 2013). Between 5 GPa and
459 10 GPa, a temperature of ~1450°C is sufficient to form interstitial melts having a composition
460 that is “intermediate between that of kimberlite and carbonatite” but containing a CO₂
461 concentration of ~40% (Ghosh *et al.*, 2014). As the solubility of CO₂ in mantle rocks is very
462 low (~5 ppm), the quantity of CO₂ dissolved in the melt likely represents that of the entire solid
463 and interstitial melt volume. As stated above, assuming that the total concentration of CO₂ in a
464 sub-primitive/ULVZ mantle or fertilized harzburgite is ~1000 ppm, we can infer that the ratio
465 of the mass of the interstitial melt to that of the mush, which we note as ϕ_r , ranges between
466 ~0.25% and 1% when the interstitial melt is intermediate between a carbonatite-like and a
467 kimberlite-like melt.

468 The influence of H₂O on the mantle solidus is more complex. Experiments testing the mantle
469 conditions of the MTZ show that if the rock contains 2 wt.% water, i.e. exceeding the water
470 saturation at that depth, the wet solidus is ~200°C lower than that of the dry solidus, up to a
471 pressure of 14 GPa (~420 km depth; Fig. 2). Given that the maximum concentration of H₂O
472 assumed in our filaments is ~500 ppm, and thus much lower than the water solubility of the
473 mantle in the MTZ (>2000 ppm, Bercovici & Karato, 2005), we can infer that hydrous melting
474 is not expected inside the MTZ. Further, between 3 GPa and 8 GPa, a 150°C temperature drop
475 of the solidus is expected for a lherzolite containing 500 ppm of H₂O (see Fig. 1 from Katz *et*
476 *al.*, 2003). Assuming that the excess of temperature above the solidus required to reach cpx-out
477 remains the same for a dry and a H₂O-rich lherzolite (see above), we can also deduce that the
478 temperature needed to reach the cpx-out threshold at 3 GPa, 5 GPa and 8 GPa for wet lherzolite
479 is ~1500°C – 150°C + 125°C = ~1470°C, 1670°C – 150°C + 80°C = ~1600°C and ~1850°C –
480 150°C + 40°C = 1740°C, respectively (Fig. 2). To summarize, we therefore hypothesize that at
481 ~5 GPa, the adiabatic temperature is ~1440°C, the wet solidus of lherzolite is ~1520°C and the
482 cpx-out temperature is ~1600°C. Such a reasoning maintains that when the temperature is
483 >1600°C, picritic melts are produced at 5 GPa owing to the drop of the solidus because of the
484 presence of H₂O. However, relatively close to the axis where the temperature is slightly greater
485 than that of the wet solidus (1520°C) but lower than 1600°C, tholeiites could be produced
486 instead of picrites. Eventually, these tholeiites incorporate all CO₂ and H₂O from the sub-
487 primitive/ULVZ mantle filaments. When the temperature is slightly lower than 1520°C, the
488 mantle no longer melts, and the interstitial carbonate-rich melt keeps its initial mass fraction, ϕ_r
489 = 0.25%. It is striking to note that because of the H₂O concentration of the sub-primitive/ULVZ
490 mantle, the temperature at which the interstitial melt is kimberlite-like is greater than 1600°C
491 and 1700°C at 5 GPa and 10 GPa, respectively (Brey *et al.*, 2009). We therefore propose that
492 through a 100°C interval centred on the plume axis, the interstitial melt is kimberlite-like and

493 the mass fraction is $\phi_r = 1\%$. Away from this interval, however, the interstitial melt has a greater
494 CO_2 concentration and, thus, ϕ_r decreases down to the border of the plume where $\phi_r = \sim 0.25\%$.
495 These results are important because they demonstrate that at ~ 5 GPa, a tube of flow, having a
496 significant radius compared with that of the plume (~ 25 km according to Fig. 1d), can produce
497 tholeiite-picrite melts.

498 The above-mentioned estimates depend highly on pressure (and therefore depth); at 4 GPa for
499 instance, the dry solidus occurs at 1600°C , thereby suggesting that even a dry lherzolithic mantle
500 could partially melt along the plume axis. Conversely, the dry solidus reaches 1720°C at 6 GPa,
501 and from the above reasoning, we can hypothesize that even dry partial melting at that depth
502 can occur. We must also mention that the Herzberg *et al.* (2010) dry solidus profile produces
503 higher temperatures than other estimates; at 5 GPa; for instance, the dry solidus occurs at
504 1670°C , 1650°C and 1630°C for Herzberg *et al.* (2010), Hirschmann (2000) and Katz *et al.*
505 (2003), respectively. This drop of the estimated dry solidus implies that the window for wet
506 partial melting simply drifts down to 6 GPa when the Hirschmann (2000) or the Katz *et al.*
507 (2003) solidus is assumed.

508 An important question does arise from the above reasoning: Does fertilized harzburgite have
509 the same melting behaviour as lherzolite? Unfortunately, we cannot use the studies of Katz *et al.*
510 (2003) and Ballmer *et al.* (2013) to answer this question, as they did not explicitly consider
511 the case of harzburgitic mantle. Maaloe (2004) performed laboratory experiments at pressures
512 < 3.5 GPa and demonstrated that the dry solidus of mantle harzburgite clearly lies 150°C above
513 that of dry lherzolite, i.e. at $\sim 1670^\circ\text{C} + 150^\circ\text{C} = 1820^\circ\text{C}$ and $\sim 1850^\circ\text{C} + 150^\circ\text{C} = 2000^\circ\text{C}$ at 5
514 GPa and 8 GPa, respectively (Fig. 2). Moreover, it has been observed that immediately above
515 the solidus, melts having a tholeiitic composition can be produced (Maaloe, 2004; Herzberg &
516 Asimow, 2015). If we assume, such as for lherzolite, that a 150°C drop of the hydrated solidus

517 occurs when the H₂O concentration is 500 ppm, i.e. it coincides approximately with the dry
518 lherzolite solidus, we can deduce that the hydrated solidus for fertilized harzburgite may be
519 ~1670°C and ~1850°C at 5 GPa and 8 GPa, respectively, i.e. about 80°C below and just equal
520 to the maximum temperature of the plume at 5 GPa and 8 GPa, respectively (Fig. 2). It remains
521 that inside the Hawaiian mantle plume stem, the partial melting of fertilized-harzburgite
522 filaments depends essentially on the CO₂ concentration. Unfortunately, there have been no
523 laboratory experiments in which materials resembling hydrated and carbonated fertilized
524 harzburgites have been partially melted. We only know that between 5 GPa and 10 GPa, the
525 melting temperature of sodium carbonate (Na₂CO₃) increases from ~1400°C to 1520°C,
526 whereas that of calcium carbonate (CaCO₃) fluctuates around 1720°C (Li *et al.*, 2017). As the
527 fertilization of harzburgite implies an enrichment in alkali and other trace elements such as the
528 light rare earth elements (LREEs), we deduce that the partial melting of such filaments must
529 follow approximately the melting temperature of Na₂CO₃; thus between 5 GPa to 10 GPa,
530 carbon-rich interstitial melts are produced, regardless of their position within the plume. As
531 well, because the temperature along the plume axis is close to that of the harzburgite hydrated
532 solidus, kimberlite-like melts may be produced, whereas to the sides of the axis, the melts
533 should be carbonatite-like.

534 Finally, we must consider the melting temperature of the recycled eclogite. Given the
535 dehydration of eclogite during subduction, the recycled eclogite is likely dry (Dixon *et al.*,
536 2002). As well, at 5 GPa the dry solidus versus liquidus interval of eclogite and residual eclogite
537 is at 1400°C and 1550°C, respectively (Rosenthal *et al.*, 2014; Fig. 2). These temperatures
538 imply that for a large fraction of the plume section, the dry eclogite melts completely along the
539 plume axis and partially melts along its border (Fig. 2).

540 **FATE OF THOLEIITE-PICRITE MELTS OF SUB-PRIMITIVE/ULVZ MANTLE**
541 **FILAMENTS FLOWING INSIDE THE HOT CENTRE OF THE PLUME**

542 There is now a good understanding of the physics of melt-rock separation during partial
543 melting. The explicit incorporation of these processes into mantle plume models is, nonetheless,
544 impossible because it requires grid sizes that are two or three orders of magnitude smaller than
545 those required for modelling mantle convection. Studies of the transition zone between the
546 mantle peridotites and the gabbroic crust in the Oman ophiolites have revealed multiple
547 consequences of what can occur when ~20% of interstitial tholeiitic melt is produced below an
548 oceanic ridge (Rabinowicz *et al.*, 1987; Ceuleneer & Rabinowicz, 1992; Rabinowicz &
549 Ceuleneer, 2005). This initially consolidated mantle mush, overlain by a gabbro horizon,
550 rapidly separates into two phases: one phase composed of a melt having few crystals (near zero
551 to 10%) and a viscosity similar to that of pure melt; the other phase is a consolidated rock having
552 an interstitial melt concentration buffered at 5% (Rabinowicz *et al.*, 2001; Rabinowicz &
553 Toplis, 2009). The transient, interstitial upward percolation of tholeiitic melt dissolves the opx
554 of the surrounding mantle to form restites having a dunite composition (Ceuleneer &
555 Rabinowicz, 1992). This is evidenced by the presence, in the same dunitic layer, of relics of
556 troctolitic horizons—up to 10 m thick—illustrating the transient concentration of the tholeiitic
557 melt inside compaction waves; the melt eventually intrudes as dykes within the overlying
558 gabbro horizon (Rabinowicz & Ceuleneer, 2005).

559 Is it possible to have a situation similar to that observed in ophiolites during the partial melting
560 via decompression of sub-primitive/ULVZ mantle filaments, between 4 and 5 GPa and at
561 >1550°C, leading to tholeiite-picrite melts representing up to 35% of the mass of the mush?
562 Numerical modelling-based studies have led to the concept that, during compaction, interstitial
563 melt within the mantle may be collected within very narrow and raised vertical tubes (Connolly

564 & Podladchikov, 1998; Connolly & Podladchikov, 2007; Räss *et al.*, 2019). One of these
565 models is presented in Fig. 3a. The model requires two ingredients: (1) a mantle mush where
566 the initial melt concentration drops above a given depth and (2) a mush viscosity that drops by
567 a few to several orders of magnitude when the melt pressure p_f of the mush is greater than that
568 of the solid p_s . The former condition represents an obstruction to the upward percolation of melt
569 because of the drop in permeability owing to a decrease in interstitial melt, whereas the latter
570 corresponds to a decompaction viscosity drop due to the solid grains that form the mush being
571 weakly stuck together and therefore implying that the mush is close to its deconsolidation
572 threshold. Therefore, it is easier to separate the grains using the excess pressure p_e ($p_e = p_f - p_s$)
573 than to deform the grains plastically to permit the upward percolation of melt (Connolly &
574 Podladchikov, 2007).

575 Connolly & Podladchikov (1998) investigated this hydrodynamic situation more than 20 years
576 ago; however, there are no convincing natural cases to which this description of melt
577 percolation in the mantle can be applied. This lack of application relates to upward obstruction
578 being generally ascribed to a horizon where the mush permeability is low, whereas a marked
579 drop in viscosity because of decompaction is generally associated with a high permeability
580 horizon. We argue, however, that this situation can occur when tholeiites-picrites are produced
581 within a mantle plume at 5 GPa. First, the upward obstruction to melt percolation cannot be
582 attributed to a drop of upward melt concentration. A single order of magnitude drop in viscosity
583 of the mush occurs when the interstitial melt concentration within the mantle exceeds ~5%
584 (Hirth & Kohlstedt, 1995). According to our model, this situation must occur inside filaments
585 located close to the plume axis at a pressure somewhere above 6 GPa (see above). Such a
586 viscosity drop increases the upward ductile velocity inside the filament and, consequently,
587 locally reduces the thickness of the filament and generates compressive horizontal stresses σ_1
588 (see Fig. 3b). In return, the compression leads to a horizontal coalescence of the grains that

589 compose the mush and, thus, a marked reduction of the vertical permeability of the mush
590 without implying a macroscopic reduction of the bulk melt concentration of the mush
591 (Rabinowicz & Toplis, 2009). This reasoning explains the local development of an obstruction
592 inside the plume at ~ 6 GPa depth (Fig. 3b). Then, because of the accumulation of melt below
593 the zone of reduced filament thickness—arising because of the later obstruction and increased
594 melting rate owing to decompression—the interstitial fluid progressively wraps a greater
595 surface of the grain boundaries within the mush. This process leads to a progressive reduction
596 of the cohesion of the mush, which can eventually be reduced to zero when the local melt
597 concentration surpasses the deconsolidation threshold in the mush, at a degree of melting of
598 $\sim 30\%$, i.e., at a melt fraction of lherzolites where the interstitial melt has a picrite composition
599 (Rabinowicz & Toplis, 2009). When the excess pressure p_e exceeds the stress required to
600 separate the grains, the mush is then fully deconsolidated. The implication is that its viscosity
601 becomes that of a suspension, i.e., dramatically lower than that of the bottom of the compaction
602 wave, where p_e is negative and the mush continues to behave as a consolidated medium. As a
603 result, inside the core of the wet filaments, the selected conditions required to generate vertical
604 melt tubes are fulfilled, providing an efficacious way for the melt to propagate up to the base
605 of the lithosphere (Fig. 3a). An important question concerns the estimation of the tube thickness
606 δ . According to Connolly & Podladchikov (2007), δ is $\sim L\sqrt{R}$, where L represents the
607 compaction length defined by Eqn. 2 (supplementary material Table S1), and R represents the
608 amplitude drop of the solid viscosity during decompaction, i.e., when the local melt
609 concentration reaches $\sim 30\%$ over that at the initiation of the obstruction (i.e. $\sim 5\%$). According
610 to Rabinowicz and Ceuleneur (2005), L is on the order of kilometres. Further, the R factor may
611 be extremely small, in the order of 10^{-6} , indicating that δ can be on the scale of metres.

612 To hold back this melt along the bottom of the lithosphere, the effective pressure p_e at the top
613 of the tubes must be lower than the fracturing limit of the lithosphere, which is ~ 50 – 100 MPa

614 (Grégoire *et al.*, 2006). Because of the low viscosity of the melt, the excess pressure p_e drop
615 due to the melt flow in tubes is negligible. Therefore, p_e at the top of the tubes approximates
616 that of the buoyancy of the melt integrated along the total height H of the tubes (here ~ 30 km,
617 i.e. the distance separating the base of the lithosphere from the zone of production of picrites at
618 5 GPa):

$$619 \quad p_e = \delta\rho gH = \sim 150 \text{ MPa}, \quad (\text{Equ. 1})$$

620 where g is the gravity constant, and $\delta\rho$ is the solid-melt density contrast ($\sim 500 \text{ kg/m}^3$). This last
621 value is consistent with estimates of densities of carbonatitic to hydrous silicate melts, which
622 range from 2600 kg/m^3 to 3000 kg/m^3 at 5 GPa (Dobson *et al.*, 1996; Spera, 1984 and Matsukage
623 *et al.*, 2005), and a mantle density at this depth of $\sim 3420 \text{ kg/m}^3$ (Kaban *et al.*, 2016).
624 Consequently, with these parameters, upon the height of the tubes reaching a value H_{dyke} less
625 than ~ 20 km, the vertical dykes arise in the mantle, and the melt drains through these dykes.
626 The dyke thickness, depending on both the elastic parameters and the amplitude of excess
627 pressure, is on the scale of metres (i.e. Grégoire *et al.*, 2006). After penetrating the lithosphere,
628 the dykes follow the trajectory of σ_1 , which is oriented vertically above the plume at the bottom
629 of the lithosphere (Figs. 1 and 3c). When σ_1 is not perfectly vertical, the excess pressure p_e at
630 the top of the dyke tends to drop. Accordingly, dyke thickness decreases. To maintain the total
631 melt volume transported by the dykes, however, the heights of the dykes increase and eventually
632 preserve an excess pressure at their tops, buffered to the critical threshold for fracturing. Such
633 a process allows the upward free movement of dykes. Being metres in thickness, dykes not only
634 transport melt but also solid crystals and xenoliths. Finally, when the apices of the dykes rise
635 above 70 km, where σ_1 rotates to the horizontal (Figs. 1 and 3c), they transform into sills of
636 several tens of kilometres in length and probably a few metres in thickness (Maccaferri *et al.*,

637 2011). The superposition of these sills, likely set between 70 and 50 km depth, produces a
638 laterally extensive magma chamber, which we label the “shield magma reservoir”.

639 Before melt intrusion into the shield magma reservoir, a dry mantle horizon at this depth is
640 characterized by temperatures between 600°C and 750°C, and an effective viscosity of $\sim 10^{23}$
641 Pa·s (Fig. 1). Moreover, the ponded tholeiite and picrite melts are highly reactive because they
642 contain 0.3% CO₂ and 0.15% H₂O since they comprise up to 30% partial melt of a mantle
643 containing 1000 ppm and 500 ppm of CO₂ and H₂O, respectively. Thus, three mechanisms
644 contribute simultaneously to heat and eventually reduce the effective viscosity of the solid
645 fraction of the shield magmatic reservoir: (1) heating of the mantle at the contact of the sills;
646 (2) partial crystallization of the ponded melts during cooling; and (3) diffusion of the volatiles
647 from the melt into the surrounding mantle. Eventually, these processes reduce, by a few orders
648 of magnitude, the effective mantle viscosity of the shield magmatic reservoir. As the reservoir
649 is relatively thick (~ 20 km), its timescale of thermal relaxation are long (>10 Ma). If we assume
650 that the effective viscosity of the lithosphere, where the shield magmatic reservoir becomes
651 established, is 10^{20} Pa·s (instead of 10^{23} Pa·s prior to the accumulation of melt from the dykes),
652 then the compression induced by the bending of the plate in the shield magmatic reservoir,
653 having an amplitude of 30 MPa (Fig. 1) and a time-lapse of 0.1 Ma, produces a total shear ϵ of
654 $\sim 10^{20}$ Pa·s / $3 \cdot 10^7$ Pa \times $3 \cdot 10^{12}$ s = ~ 1 . If we recall that $\epsilon = \sim 0.5$ yields a total deformation γ (the
655 ratio of the long axis of the deformation with the short axis) of ~ 4 , we therefore expect $\epsilon = 1$ to
656 produce a $\gamma = \sim 16$. Such deformation rates likely induce a strong folding of the sills that form
657 the reservoir (Fig. 3d). This folding provokes a marked concentration of melt at the axis of the
658 folds. At the top of these folds, vertical fractures are generated and eventually drain the ponded
659 melt of the shield magmatic reservoir (Barraud *et al.*, 2004; Weinberg *et al.*, 2015). Finally,
660 when the tops of the fractures reach a depth of ~ 50 km, where σ_1 is again vertical, vertical dykes
661 are generated, transporting melts to near the surface (Fig. 3d). Therefore, their trajectories

662 depend strongly on the elastic stress field induced by the volcanic structure or by the crust itself.
663 This stress field can both curve the trajectories of dykes, which eventually pond as sills inside
664 a subvolcanic magma chamber or cause an eruption at the surface (Pinel & Jaupart, 2003;
665 Maccaferri *et al.*, 2011; Pinel *et al.*, 2017). The evolution of the composition of transported
666 lavas depends greatly on what occurs during this last stage of the melt flow. In supplementary
667 material A, we show that the temperature inside the ‘shield reservoir’ can account for the
668 compositions of the melts erupting at the surface. Finally, the situation is similar for tholeiite-
669 picrite melts produced in the hot diverging part of the plume, i.e. above the hot flowline (Figs
670 1d and 3c) but at a sufficient distance from the plume axis, i.e. greater than ~40 km, to allow
671 dykes carrying the melts to avoid the shield magmatic reservoir. These melts are not affected
672 by melt mixing or interactions with the mantle lithosphere occurring in and around the shield
673 magmatic reservoir. Melt-mixing and fractional crystallization processes may eventually occur
674 when these dykes reach a subcrustal horizon at a distance of ~100 to ~180 km from the plume
675 axis.

676 **FATE OF THE CARBON-RICH INTERSTITIAL MELTS OF FERTILIZED-**
677 **HARZBURGITE FILAMENTS FLOWING INSIDE THE HOT CENTRE OF THE**
678 **PLUME**

679 The above-mentioned migration processes are specific to sub-primitive/ULVZ mantle filaments
680 flowing close to the plume axis. In this section, we concentrate on the possible mobilization,
681 inside the plume at 5 GPa of CO₂-H₂O-rich melts produced within fertilized-harzburgite
682 filaments. Here, the temperatures of the solidus and the cpx-out of the 500 ppm-rich fertilized
683 harzburgite are 1670° and 1760°C, respectively, whereas the axial temperature is 1710°C (see
684 above). It is inferred that the melting rate, even along the axis of the plume itself, remains small
685 at 5 GPa depth in the fertilized harzburgite (Fig. 2). Accordingly, the rheological softening

686 expected when the interstitial melt concentration surpasses 5% does not occur. Thus, one of the
 687 conditions required to generate narrow vertical tubes is not fulfilled. Because of the severe
 688 obstruction of the upward percolation of melt located at the bottom of the lithosphere, i.e. below
 689 4 GPa, however, compaction waves are generated. This process agglomerates melts within
 690 separate pockets having thicknesses of $\sim 2 \times L$ (Rabinowicz & Ceuleneer, 2005; and Fig. 4a).
 691 Here, L designates the compaction length (Eqn. 2):

$$692 \quad L = \sqrt{\frac{k(\phi_0)\eta}{\phi_0\mu}}, \quad (\text{Equ. 2})$$

693 where η and μ represent the viscosity of the solid fraction of the mush and the melt, respectively.
 694 $k(\phi_0)$ is the permeability of the mush evaluated at ϕ_0 , the initial volume of the interstitial melt
 695 ($\phi_0 = 1.2\phi_r$, with ϕ_r ranging between 0.25 and 1% when the interstitial melt is intermediate
 696 between carbonatitic-like and kimberlite-like melts; see above). A good approximation of $k(\phi_0)$
 697 is given by Wark & Watson (1998):

$$698 \quad k(\phi_0) \approx \frac{a^2\phi_0^3}{200}, \quad (\text{Equ. 3})$$

699 where a is the solid-grain size (here ~ 3 mm). From Eqn. 3, we can derive $k(1.2\%) = 10^{-13} \text{ m}^2$
 700 and $k(0.3\%) = 1.6 \cdot 10^{-15} \text{ m}^2$. As well, the viscosity of a hydrated kimberlite melt μ at 1700°C
 701 and a depth of 5.5 GPa is $\sim 0.5 \text{ Pa}\cdot\text{s}$ (Persikov *et al.*, 2017), whereas the viscosity at a similar
 702 depth and temperature for a carbonatite melt is slightly less than $10^{-2} \text{ Pa}\cdot\text{s}$ (Kono *et al.*, 2014).
 703 Considering that η , the effective viscosity of a hydrated lherzolite mantle, is about $\sim 10^{19} \text{ Pa}\cdot\text{s}$,
 704 i.e. a value consistent with the $10^{20} \text{ Pa}\cdot\text{s}$ value observed below the lithosphere in our dry mantle
 705 plume model (Fig 1) combined with an order of magnitude extra drop linked to the water
 706 content of the filament (Karato & Wu, 1993), then $L = \sim 15 \text{ km}$ for both values of the mass of

707 melt to bulk rock ratio ϕ_r , and the viscosity of the kimberlite-like or carbonatite-like interstitial
 708 melt μ .

709 Fig. 4a presents a 1D snapshot of the distribution of melts after a period of $15 \times \frac{L}{v_{int}} = 80$ ka,
 710 where v_{int} is the interstitial melt velocity, as derived from Darcy's law (Eqn. 4):

$$711 \quad v_{int} = \frac{\delta\rho g k(\phi)}{\mu\phi} = \sim 2.5 \text{ m/yr.} \quad (\text{Equ. 4})$$

712 Over 80 kyr, the interstitial melt percolates upward by ~ 200 km. The melt accumulates below
 713 the base of the lithosphere, i.e. at a depth of ~ 120 km (pressure of ~ 4 GPa; see above). This
 714 accumulation initiates a first uppermost compaction wave, $2 \times L \approx 30$ km thick, below 120 km,
 715 followed by subsequent waves situated 60 km below each other (Fig. 4a). The upward
 716 percolating melt accumulates initially on the highest wave. Between two successive waves, the
 717 melt concentration decreases progressively, and thus the Darcy velocity here becomes quasi-
 718 zero. Finally, after 80 kyr, the amplitude of the uppermost wave approaches $\sim 10 \times \phi_r$. It is
 719 important to note that all values adopted in our model are consistent with those used by
 720 Hofmann *et al.* (2011). The main difference between the two models is that Hofmann *et al.* only
 721 studied the Darcy percolation of the interstitial melt; they did not include the simultaneous
 722 formation of compaction waves.

723 Assuming that the vertical distribution of the melt in waves is sinusoidal, we can estimate that
 724 the mean concentration of melt inside the top wave is $\sim 2 \times \left(\frac{10}{\pi}\right)\phi_r = \sim 7 \times \phi_r$. It is inferred that
 725 the mean concentration of H₂O in the mush is $\sim 7 \times 500$ ppm = 0.35 wt.%, i.e. H₂O represents
 726 $\sim 0.35\%$ of the mass of the mush. Similarly, the CO₂ concentration is ~ 0.7 wt.%. Considering
 727 the above estimate of the dry solidus of harzburgite at 1820°C and the liquidus of olivine, which
 728 is the most refractory mineral of peridotites, at 1950°C (Herzberg & Asimow, 2015), we

729 propose that the gap between the dry solidus and the liquidus of the harzburgitic mantle at 5
730 GPa defines a very narrow range from 1820°C to 1950°C. As proposed by Katz *et al.* (2003)
731 for lherzolites, we assume that the solidus versus liquidus interval of a mantle harzburgite
732 containing 0.3 wt.% of H₂O is reduced by 500°C compared with that of a dry harzburgite. We
733 therefore deduce that at 5 GPa, the solidus versus liquidus interval of a harzburgite containing
734 ~0.3 wt.% (3000 ppm) H₂O ranges between 1320°C and 1450°C. The latter value is very close
735 to the adiabatic temperature of the mantle at 5 GPa (1440°C, Fig. 2). Therefore, the presence of
736 the $\sim 7 \times \phi_r$ of kimberlitic or carbonatitic interstitial melt inside the top compaction wave leads
737 to the complete melting of the solid fraction of the mantle within the wave. The composition of
738 the produced melt is a mix of 7% kimberlite-like or 1.8% carbonatite-like melt and the
739 remaining melt being molten fertilized harzburgite. These melts have similar major element
740 compositions as those of meimechites from the Siberian large igneous province (Table S2A).
741 Our proposition also agrees with Panina & Motorina (2013), who studied melt inclusions
742 occurring in olivine phenocrystals from meimechite lavas and associated cumulative dunites.
743 They determined that these melt inclusions represented a meimechite melt produced by a high
744 degree of partial melting of a mantle source. Accordingly, for the following discussion, we use
745 the term “meimechite-like” melts. As a consequence of the complete melting of the harzburgite
746 along the plume axis in the uppermost compaction wave, a 30 km thick layer of pure melt
747 accumulates below the lithosphere, here at ~120 km depth (Fig. 4b). It also imposes an excess
748 pressure p_e of 0.15 GPa at the base of the lithosphere, which ends with the generation of
749 meimechite-like dykes having the same trajectory as the tholeiite-picrite dykes generated above
750 the sub-primitive/ULVZ mantle filaments. In particular, when the dykes eventually pond in the
751 shield magmatic reservoir, the folding of the sills must cause the meimechite-like and tholeiite-
752 picrite melts to mix. Finally, in supplementary material A, we discuss melt segregation when

753 meimechite-like dykes avoiding the shield reservoir pond in sub-crustal magma chambers
754 below pre-shield and post-shield volcanoes.

755 **TRAJECTORY OF MELT PRODUCED IN FERTILIZED-HARBURGITE AND SUB-**
756 **PRIMITIVE MANTLE/ULVZ FILAMENTS ALONG THE PLUME BORDER**

757 Determining the upward percolation of volatile-rich melts close to the border of the plume is
758 more complex because this vertical movement also depends on trajectories of the ductile flow
759 within the plume, which are not purely vertical. A first consequence is that filaments tend to
760 become more horizontal. This is clearly the case for filaments that follow the cold border
761 flowline and become more horizontal at a distance of ~ 70 km from the plume axis and at a
762 depth >180 km (~ 6 GPa; Figs. 1d and 4b). Hence, the upper interface of each filament
763 corresponds to a zone composed of dry peridotite, which does not partially melt and thus has a
764 zero permeability, i.e. it fulfils the condition for concentrating the carbonatitic interstitial melt
765 in compaction waves in the underlying vertically oriented fertilized-harzburgite or sub-
766 primitive mantle filaments that form because the local temperature is close to the mantle
767 adiabatic temperature, i.e. $\sim 1410^\circ\text{C}$, a value lower than the 500 ppm H_2O -hydrated solidus of
768 lherzolite at 6 GPa ($\sim 1600^\circ\text{C}$) and thus also of harzburgite ($\sim 1750^\circ\text{C}$). It is inferred that at the
769 entrance of the elbow formed during the fast rotation of the flow line to the horizontal at the
770 plume border, at a depth corresponding to 6 GPa, there is an accumulation of interstitial
771 carbonatitic melt that had previously dissolved the 500 ppm of H_2O initially present in the bulk
772 of the filament. This melt is reactive with both the filament and the surrounding dry peridotite
773 and thus contributes to metasomatizing them. Accordingly, mafic silicate melt can form and
774 mix with the carbonatitic interstitial melt to produce kimberlite-like melts. In the present
775 situation, the interstitial melt cannot escape upward via dykes because of the horizontally
776 oriented σ_1 , in contrast with flowlines inside the plume core where σ_1 is vertically oriented. (see

777 Fig.1). Beyond the entrance of the horizontalized section of the elbow, the compaction leads to
778 accumulation of volatile-rich melt along the top of the filament. At a distance of 75 km from
779 the plume axis, the cold flowline has a 10° – 20° slope leading to a horizontal upper limit at about
780 140 km depth, i.e. ~ 10 km less than the depth corresponding to 5 GPa and at a distance of 175
781 to 250 km from the plume axis (Fig. 4b). Melt percolation experiments inside a sloped partially
782 molten layer show that (1) compaction pockets migrate upward and accumulate at the apex of
783 the layer and (2) the pressure excess p_e there reflects the buoyancy of the melt integrated from
784 the bottom to the top of the inclined layer (here ~ 40 km, between 180 and 140 km depth). This
785 value represents a pressure excess p_e of 0.2 GPa, a value exceeding the threshold limit for
786 fracturing by a factor of ~ 2 (Rabinowicz & Ceuleneer, 2005; Grégoire *et al.*, 2006). Finally,
787 because of the ascent of the cold flowline, the cold convective boundary of the plume is
788 somewhat pinched at a distance of ~ 150 to 200 km from the plume axis, suggesting that
789 trajectories of σ_1 along that boundary are vertically oriented (Figs. 1d and 4b). At the top of the
790 sloped filaments, therefore, there is the possibility for the kimberlitic-like interstitial melts to
791 discharge via dykes, which eventually reach the subcrustal horizon.

792 **PARTIAL MELTING OF ECLOGITE FILAMENTS INSIDE THE PLUME**

793 A complex question remains: What is the topology of the distribution of eclogite material inside
794 the plume? Considering that eclogite filaments after one convective cycle are relatively thin
795 (~ 500 m; see Introduction) and that eclogite has a stiff rheology relative to that of the
796 surrounding recycled hydrated mantle, a deformation γ of ~ 10 for the plume during its ascent
797 through the upper mantle (Farnetani *et al.*, 2002) likely leads to boudinage of the recycled
798 eclogitic filaments (Ramberg, 1955). Consequently, the resulting bodies of eclogite must be
799 less than a few kilometres in height. When they melt, the excess pressure p_e at their apex is
800 therefore likely too low (< 10 MPa, Eqn. 1) to fracture the mantle located above. It is very likely

801 that close to the plume axis, the eclogitic melt mixes with the tholeiite-picrite melts, which
802 accumulate within tubes that develop inside sub-primitive/ULVZ mantle filaments. Further,
803 melt derived from eclogite filaments may also mix inside the 30 km high meimechite-like melt
804 horizon, which develops below the lithosphere, or with the kimberlite-like interstitial melt along
805 the cold border flowline or the plume. Occasionally, the eclogitic filament can connect over a
806 vertical distance of several tens of kilometres. Accordingly, dykes transporting the derived melt
807 could end either in the shield magmatic reservoir, where it mixes with tholeiite-picrite and
808 meimechite-like melts, or, if the eclogitic filament follows the cold flowline of the plume, it
809 mixes with carbonatite or kimberlite melts. The presence of CO₂ and H₂O inside the resulting
810 hybrid melts allows them to be eruptable and eventually reach the surface. This process agrees
811 with Sobolev *et al.* (2009) who demonstrated in a melt-inclusion study that meimechite melts
812 from the Siberian platform are volatile-rich at great depths (200–270 km; CO₂: 6 wt.%; H₂O: 2
813 wt.%). When the residual partially molten eclogite filaments pond at ~150 km depth and at
814 ~200 km from the plume axis, they likely partially melt, and the surrounding kimberlitic-like
815 dykes likely host both eclogitic melts and eclogitic xenoliths. We note that meimechite melts
816 have not been found as melt inclusions in Hawaiian tholeiitic picrites; the presence of
817 meimechite melts is a prediction of our model that remains to be verified by further petrological
818 studies of Hawaiian lavas and associated olivine-rich cumulates and dunites.

819 **ABUNDANCE OF RECYCLED FILAMENTS REQUIRED TO EXPLAIN THE** 820 **VOLUME OF MELT PRODUCED IN THE HAWAIIAN CHAIN**

821 According to the model from Fig. 1, we assume that the mantle plume below Hawaii has a
822 radius of 50 km with an axial upward velocity of 1 m/yr. These estimates are consistent with a
823 1 cm/yr velocity of a 500 km radius plume in the lower mantle, which eventually transforms
824 into a plume having a 1 m/yr velocity and a 50 km radius once it reaches the upper mantle. The

825 solid flux in the plume is therefore $2 \pi VR^2/3 = \sim 5 \text{ km}^3/\text{yr}$ (to obtain this last equation, we assume
826 that the velocity is parabolic within the tube of radius R). The actual volume of crust produced
827 by the Hawaiian plume activity, as estimated using gravimetric data, is $\sim 0.2 \text{ km}^3/\text{yr}$ (van Arc &
828 Lin, 2004). Considering that tholeiite-picrite melts result from a degree of partial melting of up
829 to 30%, we find that $0.6 \text{ km}^3/\text{yr}$ of mantle is partially molten to account for the present partial
830 melt productivity of the Hawaiian plume. Such melt is produced inside a flow tube, which at 5
831 GPa depth is delimited by the 1550°C isotherm up to the plume axis where the temperature is
832 1700°C . When the 50 km radius plume cools, it imposes, along its border, the adiabatic
833 temperature of 1440°C . Therefore, it can be shown that during the time needed for the mantle
834 located within the plume to move up from the bottom of the MTZ to near the top of the upper
835 mantle ($\sim 600 \text{ km}$), a temperature of 1550°C is attained at a distance less than 4 km from the
836 border of the plume, i.e. 45 km from the plume axis at depth (see Fig. 1). Therefore, $\sim 80\%$ of
837 the flux of the plume could provide the necessary conditions for the production of tholeiite-
838 picrite melts. If the plume was composed entirely of sub-primitive/ULVZ mantle filaments,
839 then $4 \text{ km}^3/\text{yr}$ of mantle should have produced the tholeiite-picrite melt. As only $0.6 \text{ km}^3/\text{yr}$ of
840 mantle is needed to explain Hawaiian volcanism, we can deduce that the sub-primitive/ULVZ
841 mantle filaments represent a fraction ($0.6/4 = 15\%$) of the “effective” flux of the plume. This
842 estimate agrees with that proposed by Ballmer *et al.* (2013). Consequently, we conclude that
843 the Hawaiian plume is relatively poor in sub-primitive/ULVZ mantle. This result may explain
844 a posteriori why the seismic tomographic models of the Hawaiian plume display a bottom-to-
845 top maximum variation of amplitude of only 1% for V_p seismic velocities (Wolfe *et al.*, 2011).
846 This is also consistent with Ballmer *et al.* (2013) who explained the entire seismic signal of the
847 mantle plume using only thermal anomalies. For us, this result is striking because a 1%
848 interstitial melt may potentially lead to a drastic drop in seismic velocity (up to 8%; i.e.
849 Hammond & Humphreys, 2000). We suggest that because sub-primitive/ULVZ mantle

850 filaments are sparse, the large-scale tomographic models, which have a resolution power for
851 observing anomalies of a >100 km wavelength, simply do not image these filaments. The low
852 volume of sub-primitive/ULVZ mantle filaments needed to explain the productivity of the
853 Hawaiian plume is helpful because it indicates that the temperature field of our model is
854 compatible with the thermodynamics of the melting process. More precisely, we have shown
855 above that at 5 GPa, the temperature gap between the hydrated lherzolite solidus and the cpx-
856 out curve ranges from $\sim 1520^{\circ}\text{C}$ to $\sim 1600^{\circ}\text{C}$. This indicates that the 20% to 30% melting of the
857 mantle material consumes an amount of energy corresponding to a temperature of less than 0.3
858 $\times 250^{\circ}\text{C} = 75^{\circ}\text{C}$ (Lange *et al.*, 1994; Sleep & Warren, 2014). As the temperature at the plume
859 axis is close to $\sim 1700^{\circ}\text{C}$, we see that if the energy required to melt the rock is fully furnished
860 by temperature, picrites are produced only very close to the plume axis. However, because the
861 sub-primitive/ULVZ mantle filaments are sparse—representing $\sim 15\%$ of the plume flux—
862 approximately 85% of the energy required to partially melt the filaments may be provided by
863 conduction through the solid framework of the plume (e.g. Ballmer *et al.*, 2013). We can infer
864 that $\sim 12^{\circ}\text{C}$ of the bulk temperature drop, due to energy extraction by melting, is expected to
865 produce picrites, a value that is negligible considering the uncertainty of all values considered
866 in this study. Finally, because of the incorporation of H_2O and CO_2 into the produced tholeiite-
867 picrite melts, the dry restite of the sub-primitive mantle filament has an elevated solidus,
868 preventing remelt up to ~ 90 km depth even along the plume axis (Fig. 2).

869 According to our model, if we assume that the entire mantle flux of a plume is composed of
870 sub-primitive mantle, the melt productivity of the plume can exceed $1.3 \text{ km}^3/\text{yr}$, a value
871 comparable to that estimated for the Deccan traps, the North Atlantic volcanic province and the
872 Kerguelen plateau (e.g. Coffin & Eldholm, 1994). Such a hypothesis is consistent with the so-
873 called primitive mantle sources proposed for various LIPS, including the Kerguelen-Heard
874 plateaus, the Siberian and Deccan traps, the Karoo province and the Baffin Island and West

875 Greenland flood basalts (e.g. Neal *et al.*, 2002; Jackson & Carlson, 2011). Recently, Garcia *et*
876 *al.* (2020) provided evidence that Puhahonu, a 12.5–14.1 Ma volcano in the north-western-
877 western portion of the Hawaiian Ridge, is the largest shield volcano on Earth. To explain its
878 development, they considered various testable mechanisms for increasing magma production,
879 including a more fertile source and a hotter mantle. By elimination, Garcia *et al.* (2020)
880 concluded that only fluctuations in plume temperature can explain this anomaly. According to
881 our model, the mantle that led to the construction of Puhahonu was simply more fertile in sub-
882 primitive/ULVZ mantle filaments than the filaments currently transported by the Hawaiian
883 plume. When the volume of filaments increases, a higher quantity of sensible heat is required
884 to partially melt the filaments. Such an effect likely leads to a temperature decrease in the plume
885 and reduced melting rate of filaments, favouring the production of tholeiitic melt—the
886 dominant type of mantle-derived melts in the case of LIPs (e.g. Coffin & Eldhom, 1994; Ernst,
887 2014).

888 **QUANTITATIVE MODEL COMPARISON WITH GEOPHYSICAL AND** 889 **GEOCHEMICAL DATA FROM THE HAWAIIAN CHAIN**

890 **Geophysical data**

891 Fig. 5a, b and c display the 140 km, 80 km and 60 km depth slices, respectively, of the Rayleigh
892 wave tomographic model of Laske *et al.* (2011) around the Hawaiian region. The slices show
893 that the 100 km wide island of Hawaii is contoured somewhat by a crown-shaped, low-velocity
894 zone. The latter may result from the melt—produced inside the hot centre of the plume below
895 the lithosphere—being squeezed out from the mantle and eventually ponding up to 100 to 150
896 km from the present axis of the Hawaiian plume. This process is corroborated by the Tilman *et*
897 *al.* (2001) P-wave velocity structure at 65 km depth below the main island (Fig. 5d), which
898 illustrates that Kilauea, Mauna Loa, Mauna Kea, Lahalai and Kohala are located well inside the
899 low seismic anomaly crown. Strikingly, these volcanoes lie over P-wave local maxima; this

900 indicates that these major active or recently dormant Hawaiian shield volcanoes mostly drain
901 accumulated melt from the shield magma reservoir and, consequently, are probably domains
902 where the mush is concentrated anomalously in olivine-rich cumulates.

903 A 3D seismic velocity structure of the south-eastern part of the main island, above 20 km, is
904 displayed in Fig. 6 (Park *et al.*, 2007). Below ~5 km, V_P velocities range between ~6 and 8 km/s
905 (Fig. 6c and d). These velocities have been related to ultramafic cumulates that crystallized
906 inside subcrustal magma reservoirs from which the volcanism derives (Park *et al.*, 2007).
907 However, the shape of the structures suggests rather the development of 20 km wavelength
908 diapirs composed of a high-velocity material intruding a low-velocity layer. In particular, the
909 tops of the diapirs culminate below the volcanoes of Mauna Loa, Kilauea and Loihi. Given that
910 the high-velocity material corresponds to ultramafic olivine-rich cumulates, we can deduce that,
911 because of their density, the tholeiitic-picritic-meimechite melts transported within the swarms
912 of dykes described in the previous sections pond below the crust (Maccaferri *et al.*, 2011).
913 Thereafter, melts partially crystallize and form 20 km wide compaction pockets composed of a
914 mush, having a basaltic interstitial melt and a solid fraction composed essentially of olivine
915 crystals (Rabinowicz & Ceuleneer, 2005). These pockets eventually rise by diapirism to collide
916 with and deform the Moho interface. The resulting interstitial melts, squeezed out during these
917 collisions, explain the observed abnormally high velocity of diapirs, which are capped by
918 relatively low-velocity gabbroic material. Another striking observation is that the set of high-
919 velocity diapirs is concentrated mainly along two parallel vertically oriented arcs of cylinders
920 centred around the axis of the Hawaiian plume (Fig. 6a). The first arc includes Loihi and
921 eventually intercepts the coast of the main island at the location where the strongest seismic
922 activity is recorded (Fig. 6a). The second arc, situated ~20 km seaward (Fig. 6a), likely
923 represents subcrustal melt transported by existing dykes at ~100 km from the present plume
924 axis. These dykes avoid the shield magma reservoir (Fig. 6). Currently, the pooling in a cold

925 environment of a high-density mantle mush immediately beneath the crust shows strong
926 similarities with the crust/mantle transition zone beneath the ridge over the Iceland plume
927 (Rubin, 1990, Rabinowicz *et al.*, 2012). These observations therefore provide information
928 related to the ultimate ponding of dykes that drain mantle primitive melt extraction from any
929 plume.

930 Unfortunately, Fig. 6 does not indicate from where the latter dykes originate. The depth
931 distribution of the loci of the earthquakes below the rectangular horizons QQ' and RR' of the
932 main island is drawn in Fig. 7a, and we can deduce four deep zones of earthquakes occurring
933 in the south-eastern section of the island of Hawaii. One zone is very apparent and vertical
934 down to 30 km below Kilauea (Fig. 7b). Two other zones of the same vertical profile correspond
935 to ~50 km deep sources, one located beneath Mauna Loa and the other situated along the east
936 coast of the island of Hawaii (Fig. 7a), where the first seismic low-velocity cylinder of Fig. 6a
937 crosses the coast. A final zone corresponds to seismic loci rooted at 30 to 50 km below Loihi
938 (Fig. 7c). It is noteworthy that all these loci are connected to swarms of earthquakes rooting at
939 ~50 km beneath the eastern limit of the main island (Klein, 1982; Wright & Klein, 2006).
940 Previously, we stated that the ponding of sills within the shield magma reservoir leads to the
941 relaxation of the elastic stresses inside this domain and that the melt stored in the axis of the
942 horizontally folded sills eventually escapes above 50 km via vertically oriented dykes (Fig. 3).
943 Fig. 7b and c suggest that these dykes follow a path through the lithosphere created when the
944 shield volcanoes were initiated during their preshield period (Wright & Klein, 2006). The
945 persistence of the original preshield path of the melt through the entire lithosphere for such a
946 long period (>1 Ma) is surprising. We speculate that, because of the persistence of the
947 orthogonality of the σ_3 -direction in the lithosphere of the Hawaiian track, it is likely that the
948 fault plane created during the preshield stage remains largely open to drive the rejuvenated
949 volcanism to the same shield volcano (Solomon & Sleep, 1975). For instance, the oldest ages

950 of Kilauea vary between 0.6 and 1 Ma, whereas the rejuvenated stage for West Maui and East
951 Molokai began ~0.6 My ago (Garcia *et al.*, 2016; Fig. 8). These three volcanoes lie along the
952 Kea line with the two latter volcanoes located ~200 to 250 km downflow from Kilauea (Bianco
953 *et al.*, 2005). The rejuvenated stage for West Maui and East Molokai may be related to melt
954 transport via deep open faults following the same cold flowline (Fig. 4b) at a ~200 km distance
955 from the plume axis.

956 Fig. 8 displays the Watts & ten Brink (1989) estimate of the evolution of the thickness of the
957 volcanic load and the depth of the Moho along the Hawaiian volcanic chain (Loihi volcano used
958 as the reference point). Between Hawaii and Molokai, located at a distance of ~200 km from
959 the plume axis, both parameters are negatively correlated. The decrease of the thickness of the
960 load with volcanic age is consistent with the erosion of volcanoes, whereas the increase in the
961 depth of the Moho is indicative of an underplating process related to ultramafic cumulates that
962 crystallized within subcrustal magma reservoirs, similar to that occurring along the east coast
963 of Hawaii (Fig. 6b). However, the negative correlation between both profiles ends between
964 Molokai and Oahu (Fig. 8). There, the Moho depth decreases, whereas the thickness of the
965 volcanic load continues to decrease. The decrease of volcanic load coupled with the decrease
966 of the Moho depth is indicative of a reduction of the effective topographical weight applied to
967 the top of the 25 km thick elastic Hawaiian plate—Watts & ten Brink (1989) explained this
968 layer as representing the horizon where the elastic bending permanently sustains the loading
969 induced by the various volcanic charges. The resulting elastic stress is compressive above ~12.5
970 km and extensional below. If we consider the horizontal tension induced at the surface of the
971 plate by the plume, its intensity also decreases with distance from the plume axis and partly
972 balances the elastic compressive stress generated by the volcanic loading above ~12.5 km, i.e.
973 at crustal levels. The pressure of the volcanic load generally exceeds the vertical dynamic
974 stresses induced by the plume; below the main island, the pressure is ~100 Mpa, whereas that

975 of the plume is ~ 30 Mpa, i.e. the magnitude of the vertical push of the ~ 1 km high and 1200 km
976 wide swell surrounding the main island. The elevation of the coral reef deposits of the central
977 Hawaiian islands, i.e. occurring around Oahu, Molokai and Lanai, is an indication that the
978 volcanic load at ~ 200 km from the main island becomes negative, i.e. the elastic deflection due
979 to the loading is reversed because of the uplift of the volcanic structure (Grigg & Jones, 1997).
980 Above 12.5 km, the elastic stress due to the loading is, therefore, extensional rather than
981 compressive and thus reinforces the extension induced by the plume (Watts & ten Brink, 1989;
982 Bianco *et al.*, 2005). We can infer that here, the melt that accumulates below the crust is free to
983 erupt at the surface and, therefore, this stress reversal is the likely cause of the deep origin of
984 the rejuvenated volcanism, which is generally marked by the eruption of kimberlite-like melt
985 dykes commonly transporting upper-mantle xenoliths to the surface (~ 110 – 150 km; Keshav *et*
986 *al.*, 2007, and also see above). These details are specific to rejuvenated volcanism, implying
987 that it constitutes the only case where mantle melts may directly erupt at the surface without
988 differentiation in a magma chamber located either immediately below the crust
989 (preshield/postshield and most shield magmas) and in the shield magma reservoir (lavas of the
990 shield reservoir erupting directly at the surface). Bianco *et al.* (2005) suggest that the
991 rejuvenated volcanism at North-Arch, Niihau and Kauai, which are also downflow volcanic
992 structures but located quite far (up to 500 km) from the present axis of the Hawaiian plume, can
993 be interpreted similarly. This supposes that the upward push of the Hawaiian plume is
994 asymmetric and reaches ~ 500 km at a downflow distance from Kilauea. As stated in the
995 Introduction, such an effect is not taken into consideration in our model, as it is in Balmer *et al.*
996 (2011) and Hofmann *et al.* (2011). Accordingly, we suggest that the downflow plume proposed
997 by these authors may lead to trajectories of kimberlite-like melt dykes initiated beneath the cold
998 flowlines at a downflow distance of up to 500 km from the axis of the Hawaiian plume and
999 therefore eventually explains the rejuvenated volcanism of North-Arch, Niihau and Kauai.

1000 **Geochemical features of Hawaiian magmatic rocks**

1001 *Characterization of the primary melt interaction with the overlying mantle and lithosphere*

1002 *Shield volcanism.* The shield stage corresponds to the by far most voluminous magmatic activity
1003 of the Hawaiian plume. According to our model, the shield-stage parental melts—including the
1004 high-SiO₂ tholeiites, volumetrically the most abundant during that stage (Rhodes, 2016), are
1005 produced between 5 and 4 GPa (150–120 km depth) and then pond in the shield reservoir
1006 located at about 70 km depth (about 2.5 GPa), i.e. within the spinel peridotite stability field of
1007 the Hawaiian lithospheric mantle. We therefore propose that the reaction proposed by Rhodes
1008 (2016) between melts produced within the garnet peridotite stability field and the harzburgite
1009 residues occurs in the shield magmatic reservoir within the lithosphere. The lithospheric
1010 harzburgite residues proposed by Rhodes (2016) correspond, in that case, to the dry peridotite
1011 component of our model, a component consisting mostly of lherzolites and harzburgites. We
1012 finally propose that the shield magmatic reservoir, located within the mostly anhydrous spinel-
1013 bearing lithospheric mantle and where the shield-building stage parental melts pond,
1014 corresponds to a zone of more or less intense reactions (e.g. assimilation, local melting) between
1015 those melts and the surrounding lithospheric mantle. These reactions lead to variations in major
1016 elements (e.g. SiO₂, CaO, TiO₂), isotopic ratios (e.g. Sr, Nd, Pb) and trace element ratios (e.g.
1017 Zr/Nb, Zr/Y, La/Nb) observed in shield lavas occurring at the surface of the Hawaiian
1018 archipelago. Indeed, we propose that the mantle lithosphere beneath Hawaii, which is 120 km
1019 thick at the axis of the plume, is heterogeneous, consisting of a “puzzle” composed in particular
1020 of 1) anhydrous peridotite components, including the mantle source of MORB, and 2) Loa and
1021 Kea components accreted in the lithosphere during its 90 My subsidence (Fig. 9). Harrison &
1022 Weis (2018) demonstrated that the enriched Loa component was largely absent from ~11–47
1023 Ma, whereas the continuous record of shield magmatism was not modified. Therefore, it is not

1024 the nature of the dry mantle of the shield reservoir that determines the volume of the shield
1025 magmatism but rather the volume of melts derived from the filaments generated at 5 GPa. The
1026 geochemical signature is the fingerprint of the mantle composition within the shield reservoir.
1027 Regardless, it is likely that the distribution of filaments inside the plume is heterogeneous. This
1028 is suggested by Ballmer *et al.* (2013) who proposed that the northern side of the Hawaiian chain
1029 is characterized by a higher volume of melt, which can be linked to the higher volume of
1030 filaments in the northern side of the Hawaiian plume (e.g. Yu *et al.*, 2018).

1031 *Postshield volcanism.* The postshield magmas are typically small volume alkaline melts
1032 erupting immediately after (and sometimes overlapping with) the shield-stage magmas of some
1033 Hawaiian volcanoes (e.g. Clague, 1987a; Haase *et al.*, 2019). Farnetani and Hofmann (2010)
1034 suggested that they represent deeper melts than shield magmas, the latter forming at depths
1035 >4 GPa in the garnet peridotite field. This suggestion agrees with their lower SiO₂ content and
1036 higher (Ce/Yb)_N than those of most shield melts (Haase *et al.*, 2019). Rhodes (2016), on the
1037 other hand, proposed that the former originated by a garnet-free melting of the mantle at a
1038 relatively shallow depth (< 3 GPa). This assertion was based on the characteristics of postshield
1039 lavas from Mauna Kea, Kohala and Hualalai, especially their higher Al₂O₃ content relative to
1040 shield-stage lavas. Our model could reconcile these two divergent interpretations if we consider
1041 that dykes arriving from the melting zone (5 to 4 GPa depth) and transporting postshield
1042 magmas may or may not interact with the shield reservoir. If they do not interact, they will
1043 maintain the imprint of their genesis in the garnet peridotite stability field, whereas if they do
1044 interact, they lose those imprints owing to the reaction with shield reservoir products.

1045 The postshield activity of two volcanic sites, Mahukona and Haleakala, are not compatible with
1046 the volcanic classification of Clague (1987) and Clague & Dalrymple (1987). Both sites clearly
1047 show a delay to the start of their activity compared to their postshield stage. At Mahukona, the

1048 postshield stage lasts till 325 ± 25 ka, whereas the shield stage ends at ~ 400 ka (Clague &
1049 Calvert, 2009). It implies that the postshield stage may be related to melts provided to the
1050 subsurface by the Hawaiian plume from outside the shield magmatic reservoir ~ 325 ky ago. In
1051 particular, at that time, the axis of the plume was located 24 km downflow of its present position
1052 at a distance to the Mahukona volcano, clearly within the 100 to 150 km interval where melt
1053 trajectories initiated at 150 km depth drive out to pond within the shield magma reservoir (Fig.
1054 6a). This idea explains that the volcanism around 0.3 Ma at Mahokona can be compared with
1055 that of the preshield stage of Loihi (Garcia *et al.*, 2012).

1056 “*Downflow-postshield volcanism.*” The situation is more complex for Haleakala volcanism.
1057 The lavas of the submarine Hana Ridge, on which the Haleakala volcano rests, are tholeiitic
1058 basalts and picrites, similar to lavas from the Kilauea shield-building stage (Ren *et al.*, 2004).
1059 The postshield stage began with Kula volcanism, composed essentially of hawaiites and dated
1060 at 1 Ma to 0.15 Ma, followed by the still-active Hana volcanism. The latter is characterized by
1061 alkaline lavas such as basanites. The closest volcano is the extinct Kohala shield volcano
1062 located 100 km from Haleakala and which began its activity ~ 1 Ma (Fig. 7a). It can be inferred
1063 that when Kula volcanism was initiated, the plume lay in the immediate vicinity of Kohala, i.e.
1064 100 km from Haleakala. It is thus reasonable to assume that dykes transporting the erupted
1065 melts avoided ponding in the active shield magma reservoir underneath Kohala during this
1066 period. These dykes probably emerged near the subsurface below Haleakala. Further, it is
1067 tempting to assume that the recent Hana volcanism is related to the present plume, now located
1068 at ~ 160 km from Haleakala. Phillips *et al.* (2016) studied the isotopic composition of the
1069 nepheline-normative basanites from the Hana suite. These basanites are enriched in
1070 incompatible trace elements and have a more depleted isotopic signature than the similarly aged
1071 Hawaiian shield lavas from Kilauea and Mauna Loa. These differences relate to the time needed
1072 for the parental mantle melt of the Hana lavas to reach the subsurface, this time being $\sim 20\times$

1073 greater than that necessary for the Kilauea and Mauna Loa shield lavas. Our model explains
1074 this through the dykes generated at ~150 km from the plume axis and that likely root at 120 km
1075 along the hot flowline trajectory (Fig. 4b). These melts were produced at 150 km and 50 km
1076 from the plume axis and were thereafter transported via ductile plume flow before being
1077 eventually transported to the subsurface through dykes generated at 150 km from the plume
1078 axis (Figs 1d, 3c and 4b). It is clear that the time required for the melt to move 50 to 150 km
1079 away from the plume axis requires ~1 Ma, an order of magnitude greater than the 0.1 My needed
1080 for the melt produced at 150 km depth along the plume axis to erupt at the surface (Fig. 3).
1081 Accordingly, the trace element and depleted isotopic signatures of the Hana lavas are explained
1082 by the specific trajectory of their parental melt within the mantle. We thus propose that the
1083 postshield volcanism of Mahukona and Haleakala is caused by about 20% of the tholeiite-
1084 picrite and meimechite-like primitive melts produced within the hot centre of the plume not
1085 ponding within the shield magma reservoir centred along this hot centre of the plume. To
1086 distinguish this new type of Hawaiian volcanism represented by this pair of volcanoes from the
1087 normal postshield volcanism at the border of the shield magma reservoir, we designate this type
1088 as downflow-postshield volcanism.

1089 *The evolution of He^3/He^4 along the Hawaiian chain*

1090 The observation of $^3He/^4He$ ratios at more than eight times the Ra atmospheric record and a
1091 $\mu^{182W} \sim 0$ provide evidence of a major role for sub-primitive mantle filaments or particularly
1092 old, i.e. a few Ga, fertilized-harzburgite source to generate most lavas along the Hawaiian chain.
1093 Regarding the latter, some incompatible trace element ratios, such as Nb/U, appear to agree
1094 with the occurrence of such a type of fertilized-harzburgite source (Hofmann, 1986). This is
1095 illustrated in particular by the east-to-west decrease of $^3He/^4He$ in lavas along the chain. For
1096 instance, Loihi preshield lavas have the highest helium ratio values, sometimes more than $30 \times$

1097 Ra in tholeiitic samples, whereas transitional, alkaline to basanitic samples have an elevated
1098 but relatively lower helium ratio ($>20 \times Ra$) (Staudigel *et al.*, 1984). These observations support
1099 the idea that melts from dykes avoid the shield magma reservoir and only pond below the crust.
1100 Thus, some tholeiites derived from the melting of sub-primitive mantle or old fertilized
1101 filaments may not eventually mix with other melts when they erupt at the surface through the
1102 Loihi fracture. During the shield stage, the helium ratio decreases from $\sim 20 \times Ra$ to $\sim 8 \times Ra$
1103 from the time of construction of the volcano to the end of the shield stage (Kurz *et al.*, 1987).
1104 This is consistent with the idea that during their ponding period as sills in the shield magma
1105 reservoir, the mixing of tholeiite-picrite melts (high $^3He/^4He$) with meimechite-like melts
1106 derived from the partial melting of a low $^3He/^4He$ and recent (<1 Ga) filament or interaction
1107 with the dry peridotite mantle lithosphere ($8 \times Ra$) promoted a drop from a ratio of $30 \times Ra$ (the
1108 maxima at Loihi) to $20 \times Ra$ (value obtained after magma differentiation in chambers below
1109 Loihi). On the other hand, we propose that values as low as $8 \times Ra$ agree well with a preferential
1110 accumulation of meimechite-like melts associated with a recent fertilized harzburgite in the
1111 subcrustal magma reservoir below the main island at the end of shield magma reservoir activity.
1112 These low values match those of alkaline and transitional basaltic, ankaramite, hawaiiite,
1113 mugearite, benmoreite and trachyte melts, which characterize the postshield stage (Clague,
1114 1987). Although this type of volcanism represents a very small fraction of the total volume of
1115 the Hawaiian volcanism, its occurrence is a strong argument that melts provided by young
1116 fertilized-harzburgite filaments effectively contribute to the Hawaiian volcanism. However,
1117 downflow-postshield volcanism at Mahukona and Haleakala represent a striking exception to
1118 the monotonic decrease of the $^3He/^4He$ ratio with distance from Kilauea, as they display lavas
1119 and xenoliths having a $^3He/^4He$ ratio of $21 Ra$ and $32 Ra$, respectively, i.e. similar to those
1120 observed for the Loihi preshield lavas (Garcia *et al.*, 1990; Rison & Craig, 1983). This last

1121 observation is consistent with dykes that provide the subcrust melt at downflow-postshield
1122 volcanoes not interacting with the shield magma reservoir.

1123 Finally, the He^3/He^4 of rejuvenated lavas is commonly low (ranging between $7 \times \text{Ra}$ to $9 \times \text{Ra}$)
1124 except at South Arch, where the values range between $16 \times \text{Ra}$ and $22 \times \text{Ra}$, i.e. a ratio similar
1125 to that of the shield stage. This observation requires an explanation within the framework of
1126 our model. We propose linking the $7 \times \text{Ra}$ to $9 \times \text{Ra}$ range to a possible infiltration of a
1127 carbonatitic (or kimberlitic) melt that concentrates inside young fertilized-harzburgite filaments
1128 (≤ 1 Ga; $\text{He}^3/\text{He}^4 \sim 0$) that flow along the weakly tilted section of the cold flowline (Fig 4b).

1129 From laboratory experiments, we see that the infiltration process of carbonatites behaves as a
1130 diffusive process, having an equivalent diffusivity D of $4 \cdot 10^{-9}$ m²/s and $5 \cdot 10^{-8}$ m²/s when
1131 reacting with olivine or pyroxene minerals, respectively, within the overriding mantle (Shatskiy
1132 *et al.*, 2013). It is deduced that a diffusion front develops on the dry peridotitic horizon (He^3/He^4
1133 = $7-9 \times \text{Ra}$), which caps the filaments at the height of $\sim \sqrt{Dt}$, where t represents the time elapsed
1134 between the generation of the carbonatitic impregnation at 75 km from the plume axis to its
1135 discharge via dykes at ~ 200 km from that axis. According to our model, when $t \sim 1.2$ Ma and
1136 $\sqrt{Dt} = 500$ to 1500 m when the carbonatite melt interacts with a dry peridotite, the He^3/He^4 of
1137 reacting melt increases close to $7-9 \times \text{Ra}$. At the same time, because the carbonatitic melt—
1138 initially confined in a filament—infiltrates its overlying dry mantle cap, the mush percolates
1139 upward through this infiltrated front (Keller & Katz, 2016). According to Shatskiy *et al.* (2013),
1140 such a process likely leads to an increase in the height of the infiltration front by approximately
1141 an order of magnitude, i.e. 5 to 15 km. Such a distance is sufficient to buffer the He^3/He^4 of the
1142 carbonatitic melt to the value of the dry mantle cap. We note that deep xenoliths brought to the
1143 surface at Oahu likely show the imprint of such an infiltration process related to
1144 carbonatitic/kimberlitic melts (see, for example, Fig. 1 in Keshav & Sen (2003) and Plate 17a
1145 in Sen (1987)). Further, the rejuvenated magmas from Molokai, Oahu, Kauai and Niihau, which

1146 have He^3/He^4 values from 7 to $9 \times \text{Ra}$, i.e. in the range of the values of tholeiitic melts from
1147 mid-oceanic ridges centred on $8 \times \text{Ra}$ (Poreda *et al.* 1986 and Hofmann *et al.*, 2011), can be
1148 explained by a reacting carbonatite (or kimberlite). This hypothesis is consistent with that of
1149 Bizimis *et al.* (2005), who interpreted Salt Lake (Oahu) mantle xenoliths as originating from
1150 old, depleted, recycled lithosphere, and DeFelice *et al.* (2019), who generalized these results to
1151 rejuvenated volcanism occurring along the Hawaiian chain. Finally, the South Arch rejuvenated
1152 lavas, which occur outside and somewhat upflow of the Hawaiian track are the only rejuvenated
1153 lavas that show a high $^3\text{He}/^4\text{He}$, ranging between $16 \times \text{Ra}$ and $22 \times \text{Ra}$. We propose that these
1154 rejuvenated lavas originate from a sub-primitive/ULVZ mantle filament, which is expected to
1155 have a high $^3\text{He}/^4\text{He}$.

1156 CONCLUSIONS

- 1157 1- In our plume model, four types of volatiles-rich filaments are considered; sub-primitive
1158 mantle, ultralow-velocity zone, fertilized-harzburgite and eclogite type filaments. The
1159 partial melting process exclusively occurs within those filaments.
- 1160 2- The obstruction of the upward percolation of the partial melting-derived melts at the base
1161 of the lithosphere causes a series of propagating dykes. From the stress field, we
1162 distinguish two types of dykes. The first is generated at ~ 120 km depth and at a distance
1163 greater than 40 km from the plume axis and crosses the entire mantle section of the
1164 lithosphere before ponding below the crust at least at 100 km from the plume axis. The
1165 second type is generated at the same depth at a distance less than 40 km from the plume
1166 axis and ponds within the lithosphere at a depth ranging between 50 and 70 km, this final
1167 ponding horizon is called the shield magma reservoir.
- 1168 3- The dykes generated at the top of hot flowlines carry melts resulting from almost
1169 complete melting of fertilized-harzburgite filaments. The composition of those melts is

1170 similar to that of meimechites; which in our model is compatible with the prediction of
1171 high-MgO and volatile-rich melts rarely being erupted at the surface. On the other hand,
1172 20% partial melting of the sub-primitive/ULVZ filaments result in primary tholeiite-
1173 picrite melts.

1174 4- The dykes generated at the top of cold flowlines are linked to the rejuvenated volcanism.
1175 The composition of melt is kimberlite/carbonatite-like, resulting from few percent of
1176 melting regardless of the type of filaments.

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1183 **DATA AVAILABILITY**

1184 No new data were generated in support of this research.

1185 **SUPPLEMENTARY MATERIALS**

1186 Supplementary materials are available at Journal of Petrology online.

1187

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1569 **FIGURE CAPTIONS**

1570 **Fig. 1** Mantle plume convective model incorporating a viscoelastic lithosphere (Gerbault *et al.*,
1571 2017); (a) Isotherms, direction of maximum compression σ_1 (arrows with length proportional
1572 to stress amplitude) and colour code of the isocontours of “dynamic pressure,” i.e., the
1573 difference between the mantle and the lithostatic pressure. Values in blue correspond to zones
1574 that are in dynamic compression, and values in red correspond to zones in depression; (b)
1575 isotherms and isocontours (in colour) of the effective viscosity of the mantle. Note that the base
1576 of the lithosphere follows approximately the border of the intense blue zone, i.e. the $\sim 10^{20}$
1577 Pa·s horizon; (c) isotherms, vectors representing the convecting velocity (arrows, with length
1578 proportional to velocity amplitude) and isocontours of the temperature (in colour); (d) plume
1579 model characteristics necessary for understanding the melt-mantle separation processes. The
1580 1500°C isotherm delimits the high temperature centre of the plume, and some velocity vectors

1581 of the plume flow are drawn with directions of σ_1 throughout the plume and the overriding
1582 lithosphere. The elastically compressed zone, due to the flexion of the lithosphere (shown as
1583 red rectangular box), is designated in the text as the “shield magma reservoir.” The solid red
1584 line presents the trajectory of the dyke bringing melt to the surface of preshield volcanoes. The
1585 melt is initiated at 5 GPa and at <40 km from the plume axis, then ponds in the shield magma
1586 reservoir and eventually emerges at the surface, 100 km from the plume axis. Dashed blue lines
1587 illustrate the hot and cold flowlines that follow the trajectory of the mantle at the border of the
1588 hot centre and cold boundary of the plume, respectively.

1589 **Fig. 2** Pressure (depth)-temperature diagram illustrating the different curves used in the present
1590 study. The solidus of the dry lherzolite (from Herzberg *et al.*, 2000) is also estimated to be that
1591 of the fertilized harzburgite having 500 ppm H₂O. The solidus of dry fertilized harzburgite is
1592 estimated as the adjustment of the dry solidus by +150°C, whereas the dry lherzolite having
1593 500 ppm H₂O is estimated as the same dry solidus reduced by 150°C. Lines in the figure include
1594 the solidi of i) vapour-saturated peridotite (from Green, 2015); ii) pyrolite having 2 wt.% H₂O
1595 (from Litasov & Ohtani, 2002); and iii) carbonate-bearing fertile peridotite (from Dasgupta *et*
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1597 adiabatic temperature at the boundary and along the plume axis (from Katsura *et al.*, 2010). The
1598 axial temperature profile is assumed to exceed the adiabatic profile by 280°C (Putirka, 2005).
1599 Estimates of the pyroxene-out of lherzolite having 500 ppm H₂O and the dry lherzolite at 3
1600 GPa, 5 GPa and 8 GPa are deduced from Katz *et al.* (2003).

1601 **Fig. 3 a)** Model by Räss *et al.* (2019) showing the development of elongated melt tubes when
1602 the interstitial melt concentration drops by a factor of 3 at 5 GPa and the viscosity of the mush
1603 drops by a factor of 500 when the excess pressure p_e is positive. The scale of the model is
1604 adapted to describe the percolation process of tholeiite–picrite melts formed around the plume

1605 axis. b) Probable shape of the flowlines in the plume when the melt concentration exceeds ~5%
1606 in volume (from Rabinowicz & Toplis, 2009). This leads to a horizontal orientation of the
1607 direction of maximum compression σ_1 . If the mush results from the superposition of spherical
1608 grains of equal size, the σ_1 stress leads to a horizontal coalescence of these grains, and the melt
1609 moves inside planes that separate successive horizons of coalesced grains. Such a process leads
1610 to a marked drop in the vertical permeability of the mush, which leads to a marked obstruction
1611 of the vertical melt percolation. In addition to considering that the mush deconsolidates when
1612 the interstitial melt concentration is high (~30%) and when $p_e > 0$, Fig. 3a describes tholeiite–
1613 picrite melt concentration in tubes within the plume above the 5% partially molten horizon
1614 rooting at a depth of ~150 km (~5 GPa). Fig 3c displays (black lines) the trajectories of dykes
1615 that drain the melt concentrated at the top of the melt tubes. This melt develops in evenly
1616 distributed sub-primitive filaments inside the plume centre. They follow the orientation of σ_1
1617 (black segments) and eventually pond below the crust within magma chambers (red ellipses).
1618 The different trajectories provide melts to preshield, shield, postshield and downflow-
1619 postshield volcanoes as described in the text. Fig 3d explains why—because of the elastic twist
1620 of the lithosphere induced by the underlying plume—sills lodged in the shield magma reservoir
1621 fold and generate dykes above 50 km to provide melt to shield volcanoes.

1622 **Fig. 4** Evolution (1D) of melt concentration related to the obstruction of upward melt
1623 percolation, which occurs at the bottom of the lithosphere (from Rabinowicz & Ceuleneer,
1624 2005). The initial melt concentration profile of the computation is drawn in a thin plain line,
1625 above 420 km depth, where it is equal to ϕ_r , the initial melt/mush mass ratio. The bold plain
1626 line represents the evolution of melt concentration after 80 ka. Note that below the lithosphere,
1627 the uppermost compaction wave is ~30 km thick and reaches a maximum melt concentration
1628 of $\sim 10\phi_r$. b) The trajectories of the dykes that develop at the top of the compaction waves and
1629 in which carbonatite/kimberlite-like melts are concentrated. This process stems from the

1630 dehydration and decarbonation of the fertilized-harzburgite filaments above the mantle
1631 transition zone (see text). The intensity of interstitial melt concentration is represented by the
1632 colour gradient from intense red to light pink. Along the plume axis, a train of compaction
1633 waves develops because of the permeability drop at the base of the lithosphere. Away from the
1634 plume axis, in contrast, a train of compaction waves results from the horizontalization of
1635 flowlines due to the rotation of the flow below the lithosphere (see text). Inside the hot centre
1636 of the plume, the mantle in the upper wave melts completely to form meimechite-like melts that
1637 apply a pressure at the plume top to trigger dykes that follow the direction of σ_1 to transport the
1638 melt up to the base of the crust. The trajectories are very similar to those of sub-primitive
1639 filaments (Fig. 3) and thus feed the same volcanoes. Similarly, along the elbow formed by the
1640 cold flowline, carbonatitic melt may accumulate; however, because of the low temperature,
1641 fertilized-harzburgite filaments only experience moderate partial melting. The produced melt is
1642 channelled along the weakly tilted (to horizontal) arm of the elbow. The resulting kimberlite-
1643 like melt concentrates at its apex, where excess pressure is sufficient to promote melt transport
1644 via dykes up to the subcrust or even to the surface at a distance >200 km from the plume axis.
1645 This process explains rejuvenated volcanism (see text).

1646 **Fig. 5** V_p seismic anomalies (in %), derived from the Rayleigh wave tomographic model of
1647 Laske *et al.* (2012), at a) 60 km, b) 80 km and c) 140 km depth. d) V_p seismic anomalies (in %)
1648 at 65 km depth, as derived from the P-volume wave tomography model of Tillman *et al.* (2001).
1649 Note that the seismic anomalies are positive and represent local maxima above both the main
1650 island and the main shield volcanoes.

1651 **Fig. 6** Three-dimensional onshore–offshore V_p seismic tomography of the eastern margin of
1652 the main island (from Park *et al.*, 2007), with velocity fields at a) 10 km and b) 12 km depth. A
1653 velocity <6 km/s corresponds to crustal material, whereas a velocity >6 km/s corresponds to

1654 mafic or mantle materials. Also provided are the vertical velocity fields along the profiles c)
1655 AA' (in Fig. 6a) and d) BB' (in Fig. 6b). The positions of Mauna Loa, Loihi, Kilauea, Keikikea
1656 and Hohonu are indicated.

1657 **Fig. 7** a) Representation of the surface distribution of the loci of various earthquakes around
1658 the main island (from Tillman *et al.*, 2001). The V_p -anomaly map at 10 km depth identifies the
1659 location of Loihi and Keikikea volcanoes and associated seismic activity (the full map is
1660 displayed in Fig. 6a). Also provided is the distribution of the loci of earthquakes along a vertical
1661 plane linking b) Mauna Loa to Kilauea (Tillman *et al.*, 2001) and c) Loihi to Kilauea (Klein,
1662 1982).

1663 **Fig. 8** Model of Watts & ten Brink (1989) of the evolution of the Moho depth along the
1664 Hawaiian chain, including topography, positions of the top of the oceanic crust, base of the
1665 oceanic crust and depth of the Moho discontinuity along the axis of the Hawaiian chain. The
1666 numbers above the islands are the radiogenic ages of surface lavas (Clague & Dalrymple, 1987).
1667 The velocity–depth structure based on ESP 5 is also presented (ten Brink & Brocher, 1988).

1668 **Fig. 9** Diagrams illustrating the flow of various mantle materials of the Hawaiian plume along
1669 the Mauna Kea and Mauna Loa lines. The basic flow (cream color) is assumed to consist of dry
1670 peridotite; the light blue stripes represent devolatilized LLSVP mantle, i.e. Loa-type mantle, and
1671 the red stripes represent devolatilized Pacific, i.e. Kea-type mantle. Sub-primitive/ULVZ (deep
1672 blue), eclogite (yellow) and fertilized-harzburgite (grey) filaments are transported by the mantle
1673 flow and eventually partially melt beneath the lithosphere. The melts provided by filaments are
1674 collected in dykes rooting between 120 and 150 km depth and are eventually transported
1675 upward to pond in the shield reservoir and subcrustal magma chambers (preshield, postshield
1676 and downflow-postshield dykes) or directly erupt at the surface (rejuvenated dykes). It is
1677 important to note that the type of mantle composing the lithosphere depends on its nature when

1678 it was integrated to the lithosphere during its subsidence, i.e., far from the present position of
1679 the plume. Because the type of mantle in the shield magma reservoir determines the Kea (light
1680 blue) or Loa (dark pink) nature of shield magma, it is deduced that, at present, the nature of the
1681 mantle within the shield reservoir and the sublithospheric plume flow is identical. This is not
1682 the case downflow of the plume, however, where the nature of mantle sources of shield melts
1683 fluctuates along both lines.

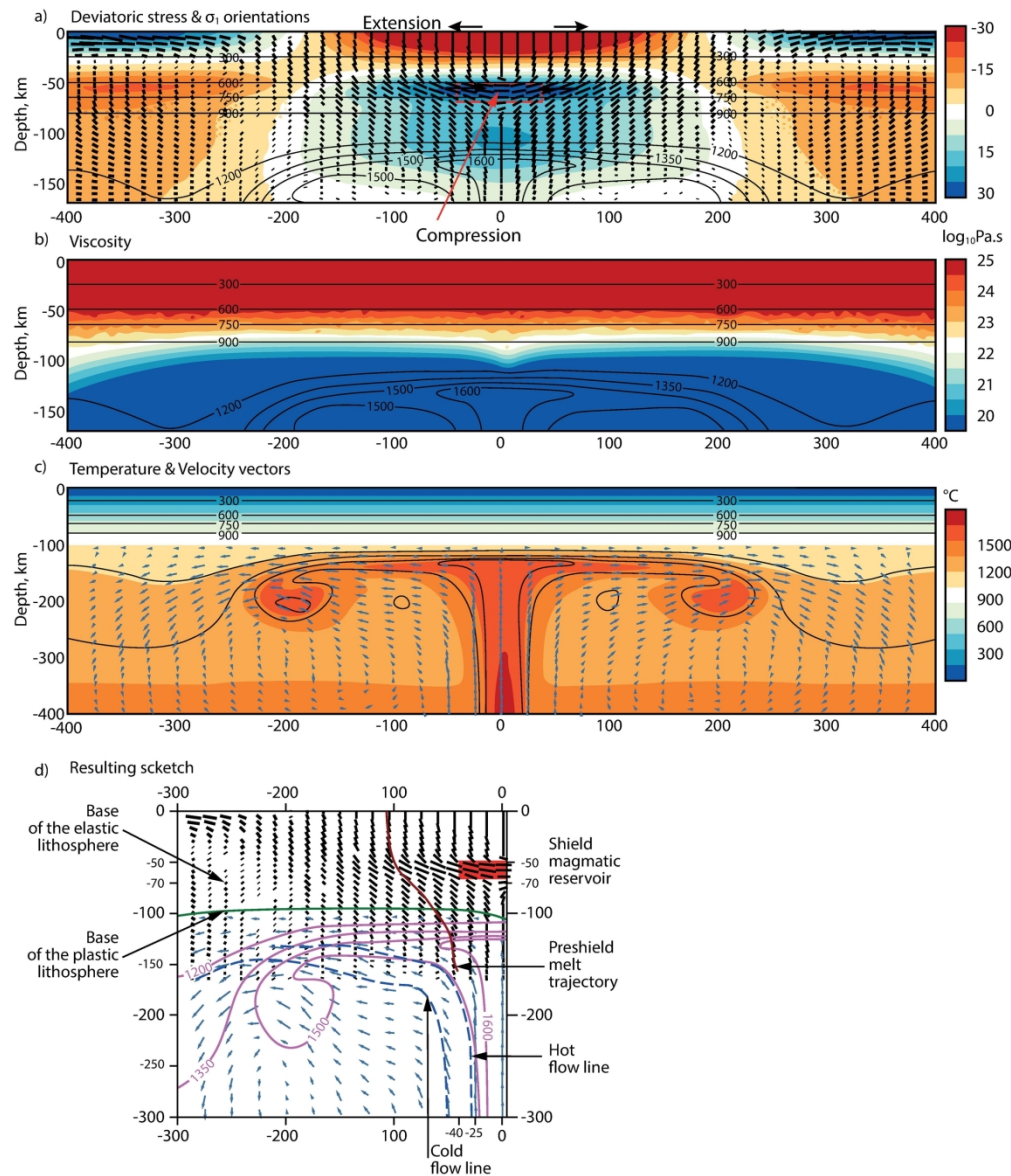


Fig. 1 Mantle plume convective model incorporating a viscoelastic lithosphere (Gerbault et al., 2017); (a) Isotherms, direction of maximum compression σ_1 (arrows with length proportional to stress amplitude) and colour code of the isocontours of "dynamic pressure," i.e., the difference between the mantle and the lithostatic pressure. Values in blue correspond to zones that are in dynamic compression, and values in red correspond to zones in depression; (b) isotherms and isocontours (in colour) of the effective viscosity of the mantle. Note that the base of the lithosphere follows approximately the border of the intense blue zone, i.e. the $\sim 1020 \text{ Pa}\cdot\text{s}$ horizon; (c) isotherms, vectors representing the convecting velocity (arrows, with length proportional to velocity amplitude) and isocontours of the temperature (in colour); (d) plume model characteristics necessary for understanding the melt-mantle separation processes. The 1500°C isotherm delimits the high temperature centre of the plume, and some velocity vectors of the plume flow are drawn with directions of σ_1 throughout the plume and the overriding lithosphere. The elastically compressed zone, due to the flexion of the lithosphere (shown as red rectangular box), is designated in the text as the "shield magma reservoir." The solid red line presents the trajectory of the dyke bringing melt to the surface of preshield volcanoes. The melt is initiated at 5 GPa and at $<40 \text{ km}$ from the plume axis, then ponds in the shield magma reservoir and eventually emerges at the surface, 100 km from the plume axis. Dashed blue

lines illustrate the hot and cold flowlines that follow the trajectory of the mantle at the border of the hot centre and cold boundary of the plume, respectively.

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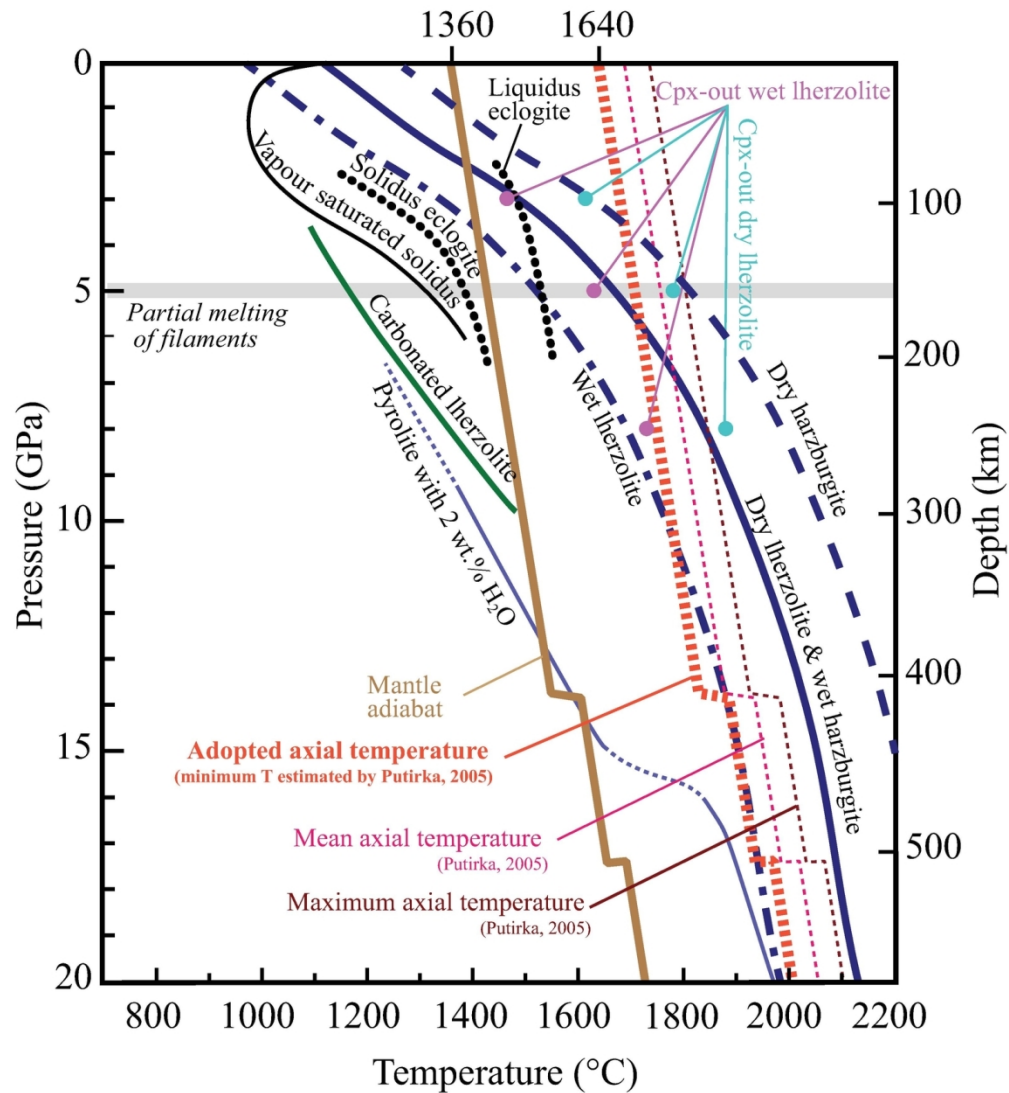


Fig. 2 Pressure (depth)-temperature diagram illustrating the different curves used in the present study. The solidus of the dry lherzolite (from Herzberg et al., 2000) is also estimated to be that of the fertilized harzburgite having 500 ppm H₂O. The solidus of dry fertilized harzburgite is estimated as the adjustment of the dry solidus by +150°C, whereas the dry lherzolite having 500 ppm H₂O is estimated as the same dry solidus reduced by 150°C. Lines in the figure include the solidi of i) vapour-saturated peridotite (from Green, 2015); ii) pyrolite having 2 wt.% H₂O (from Litasov & Ohtani, 2002); and iii) carbonate-bearing fertile peridotite (from Dasgupta et al., 2006). iv) The solidus and liquidus of eclogite (from Rosenthal et al., 2014) and v) the adiabatic temperature at the boundary and along the plume axis (from Katsura et al., 2010). The axial temperature profile is assumed to exceed the adiabatic profile by 280°C (Putirka, 2005). Estimates of the pyroxene-out of lherzolite having 500 ppm H₂O and the dry lherzolite at 3 GPa, 5 GPa and 8 GPa are deduced from Katz et al. (2003).

134x146mm (300 x 300 DPI)

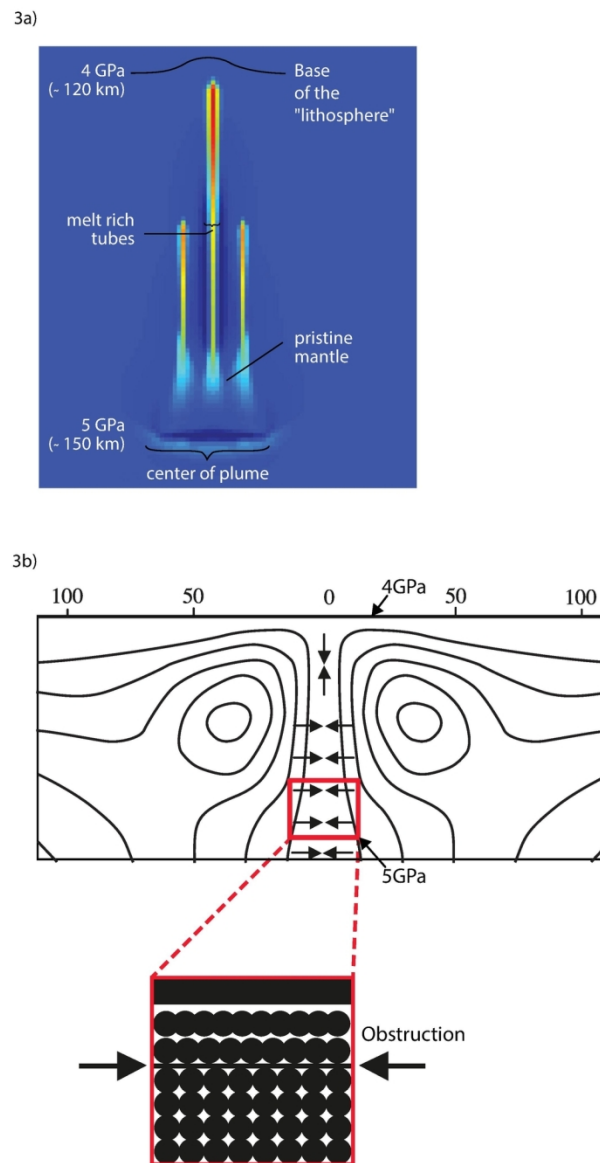


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centre. They follow the orientation of σ_1 (black segments) and eventually pond below the crust within magma chambers (red ellipses). The different trajectories provide melts to preshield, shield, postshield and downflow-postshield volcanoes as described in the text. Fig 3d explains why—because of the elastic twist of the lithosphere induced by the underlying plume—sills lodged in the shield magma reservoir fold and generate dykes above 50 km to provide melt to shield volcanoes.

113x182mm (300 x 300 DPI)

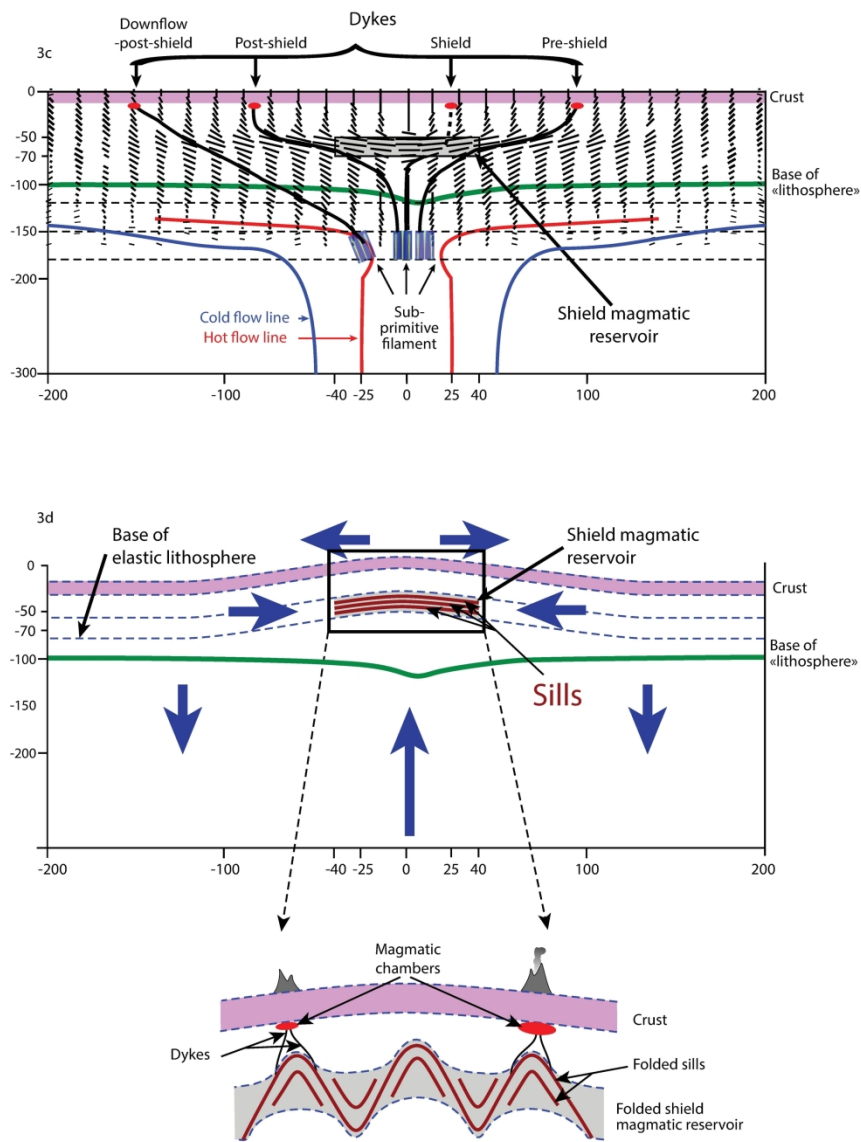


Fig. 3c and d

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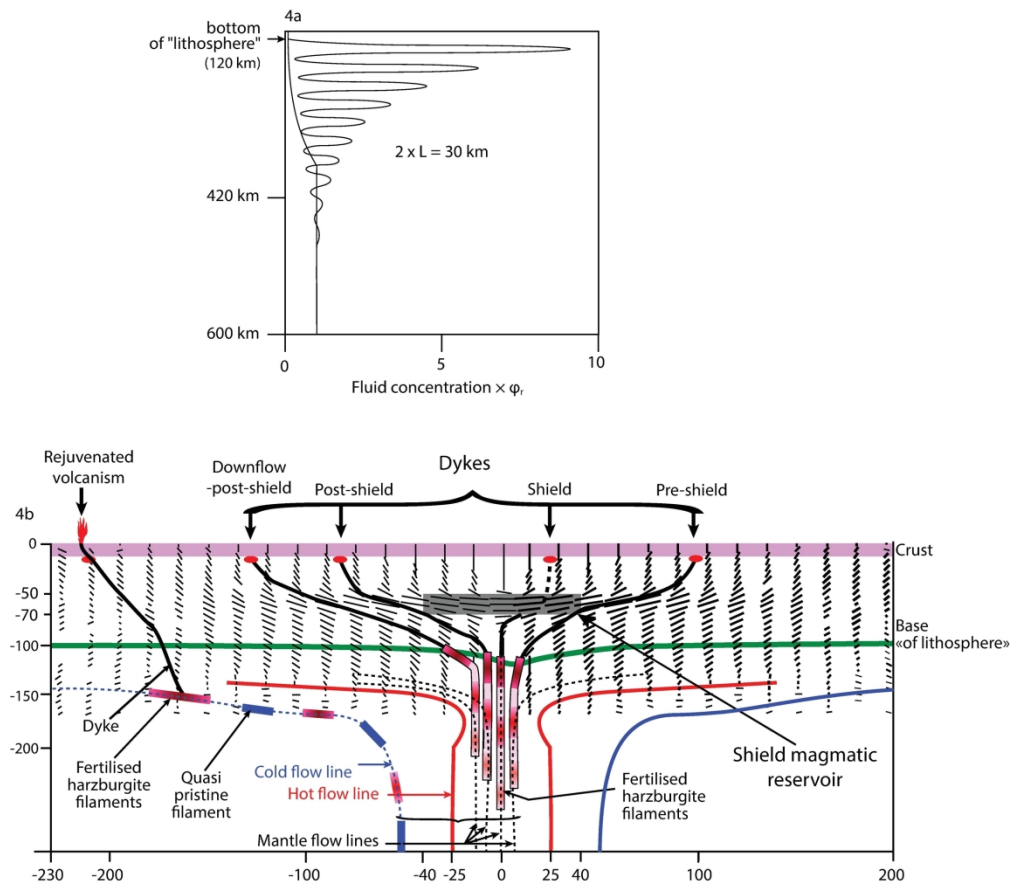


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204x238mm (300 x 300 DPI)

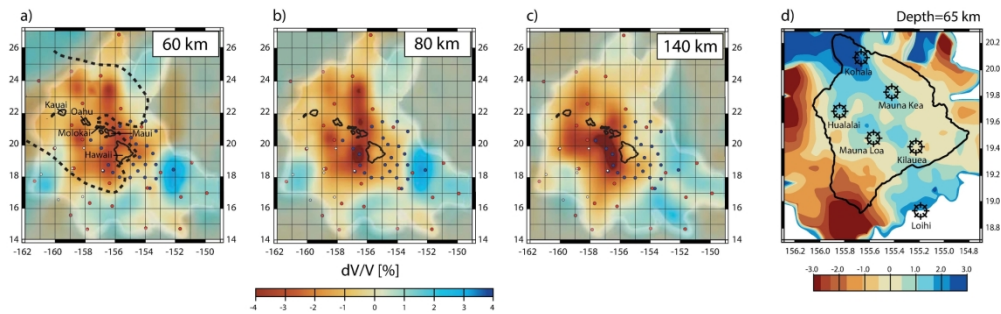


Fig. 5 V_p seismic anomalies (in %), derived from the Rayleigh wave tomographic model of Laske et al. (2012), at a) 60 km, b) 80 km and c) 140 km depth. d) V_p seismic anomalies (in %) at 65 km depth, as derived from the P-volume wave tomography model of Tillman et al. (2001). Note that the seismic anomalies are positive and represent local maxima above both the main island and the main shield volcanoes.

216x65mm (300 x 300 DPI)

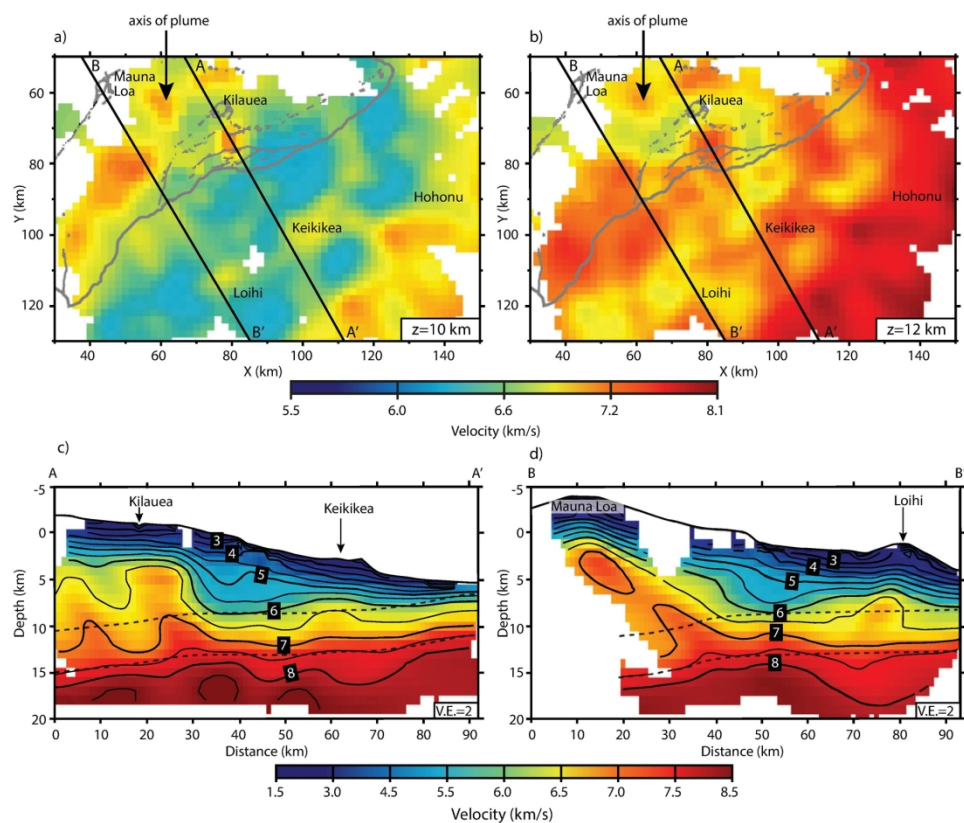


Fig. 6 Three-dimensional onshore-offshore V_p seismic tomography of the eastern margin of the main island (from Park et al., 2007), with velocity fields at a) 10 km and b) 12 km depth. A velocity <6 km/s corresponds to crustal material, whereas a velocity >6 km/s corresponds to mafic or mantle materials. Also provided are the vertical velocity fields along the profiles c) AA' (in Fig. 6a) and d) BB' (in Fig. 6b). The positions of Mauna Loa, Loihi, Kilauea, Keikikea and Hohonu are indicated.

170x139mm (300 x 300 DPI)

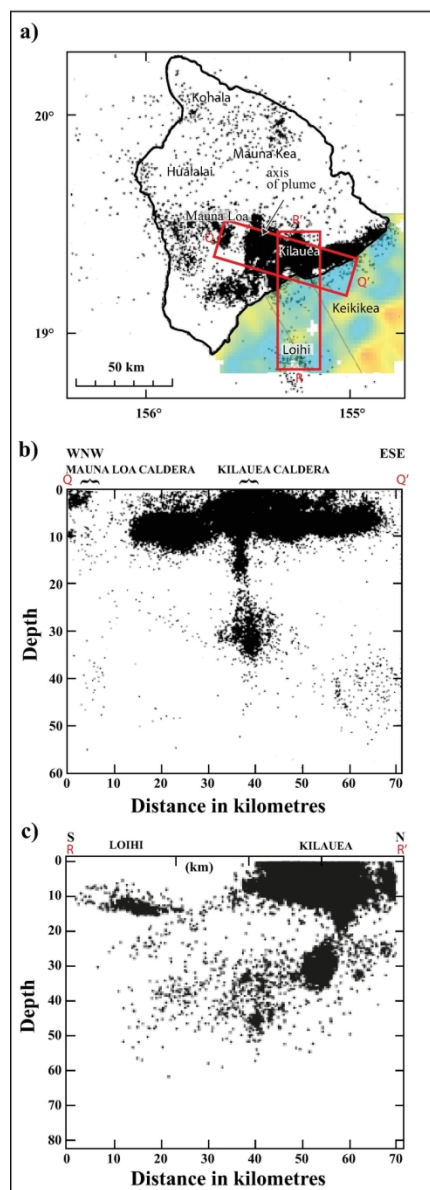


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129x184mm (300 x 300 DPI)

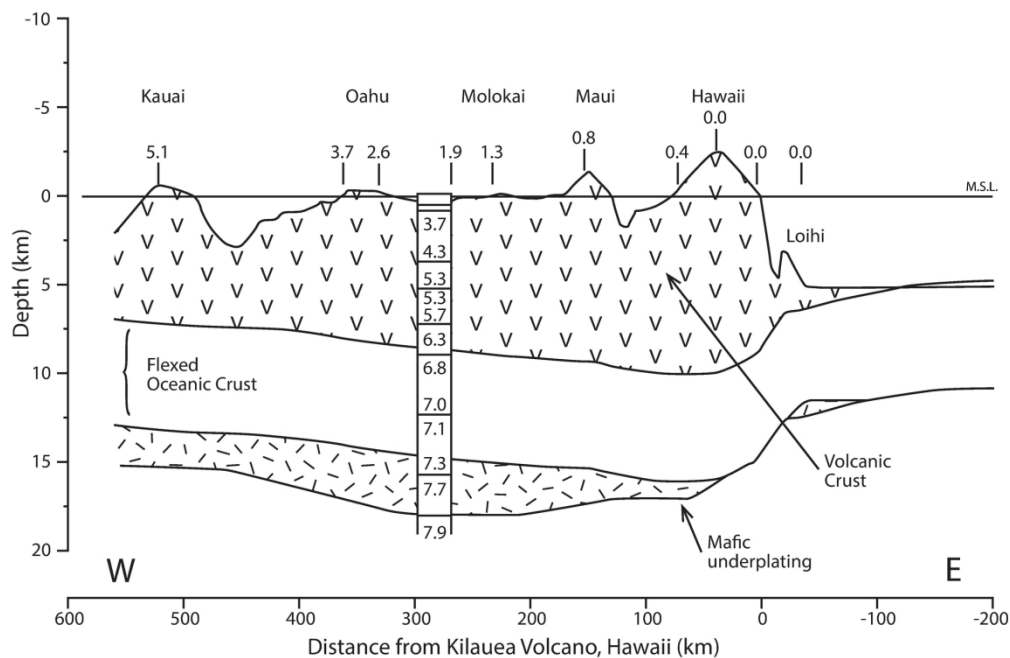


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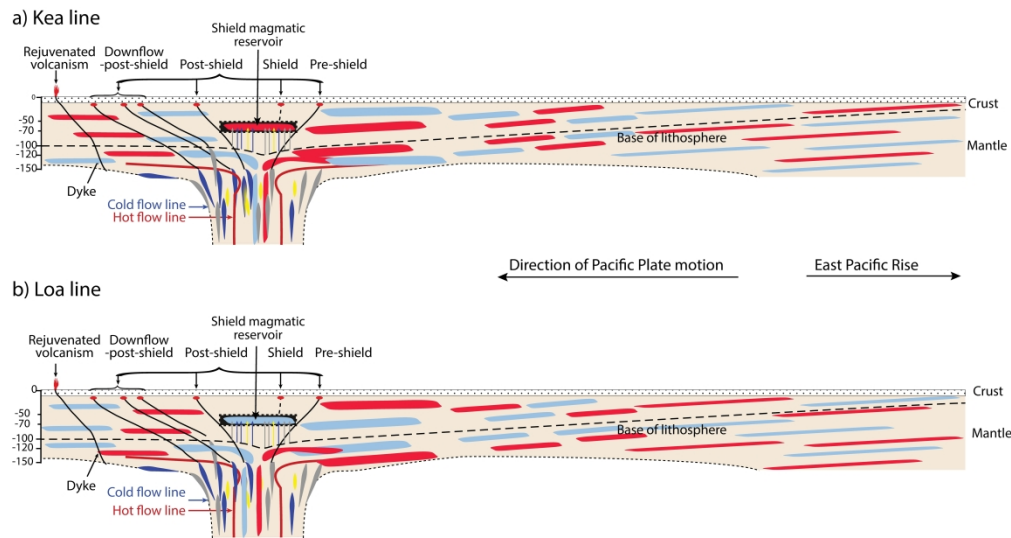


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628x334mm (300 x 300 DPI)