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# Plate tectonics and mantle controls on plume dynamics

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## Abstract

Mantle plumes provide valuable information about whole Earth convection: they originate at the core-mantle boundary, cross Earth's mantle and interact with the lithosphere. For instance, it has been proposed that the mobility/stability of plumes depends on plume intrinsic properties, on how slabs interact with the basal boundary layer, on mantle flow, or on their proximity to mid-ocean ridges.

Here, we use 3D-spherical models of mantle convection generating self-consistent plate-like behaviour to investigate the mechanisms linking tectonics and mantle convection to plume dynamics. Our models produce fully-dynamic mantle plumes that rise vertically with deflection  $< 10^\circ$  and present excess temperatures, rising speeds, buoyancy and heat fluxes comparable to

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observations. In the absence of plate tectonics, plumes are stable and their lifetime exceeds hundreds of million years. With plate tectonics, plumes are more mobile, and we identify four physical mechanisms controlling their stability. 1/ Fixed plumes are located at saddle points of basal mantle flow. 2/ Plumes moving at speeds between 0.5-1 cm yr<sup>-1</sup> are slowly entrained by passive mantle flow. 3/ Fast plume motions between 2-5 cm yr<sup>-1</sup> lasting several tens of million years are caused by slab push. 4/ Plumes occasionally drift at speeds > 5 cm yr<sup>-1</sup> over < 10 Myr through plume merging. We do not observe systematic anchoring of plumes to mid-oceanic ridges. Independent of the presence of a dense basal layer, plate-like regimes decrease the lifetime of plumes compared to stagnant-lid models. Plume age, temperature excess or buoyancy flux are not diagnostic of plume lateral speed. The fraction of plumes moving by less than 0.5 cm yr<sup>-1</sup> is > 25%, which suggests that fixed hotspot reference frames can be defined from carefully selected hotspot tracks.

*Keywords:* Mantle plumes, plume drift, numerical modelling, mantle convection, plate-like tectonics

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## 1. Introduction

Since Morgan (1972) linked deep mantle plumes to tectonics motions, the combination of seismology, petrology and geophysics have led to the characterisation of hotspots and deep mantle plumes. Recent full-waveform tomography suggests that Earth's major plume conduits are vertical and broad, > 600 km in diameter in the lower mantle, preferentially tilted in the upper mantle and likely anchored at the base of the mantle (French and Romanow-

8 icz, 2015). The composition of olivine phenocrysts indicates upper mantle  
9 plume excess temperatures between +150 K and +300 K (*e.g.* Putirka, 2005).  
10 The vertical deflection of oceanic lithosphere by mantle plumes (*e.g.* Sleep,  
11 1990; Crosby and McKenzie, 2009) and the propagation velocity of plume-  
12 related V-shaped ridges (Poore et al., 2009; Parnell-Turner et al., 2014) con-  
13 strain plume buoyancy fluxes to between  $0.3 \times 10^3 \text{ kg s}^{-1}$  (Bowie, Sleep, 1990)  
14 and  $> 70 \times 10^3 \text{ kg s}^{-1}$  (Iceland, Parnell-Turner et al., 2014). Combining infor-  
15 mation on plume radius, excess temperature and buoyancy flux gives plume  
16 rising speeds between  $23 \text{ cm yr}^{-1}$  and  $54 \text{ cm yr}^{-1}$  (Poore et al., 2009; Turcotte  
17 and Schubert, 2014) and plume heat flow anomalies between  $10\text{-}20 \text{ mW m}^2$   
18 (Sleep, 1990).

19 Plumes can provide valuable information about the physics of mantle  
20 convection since they potentially interact with the whole mantle, including  
21 both the basal and top boundary layer. Several studies have focused on  
22 characterising the temporal stability of mantle plumes, because fixed plumes  
23 can serve as an absolute reference for global tectonic reconstructions (*e.g.*  
24 Wilson, 1963). However, paleomagnetic, geochronological and petrological  
25 studies suggest contrasting plume stability/mobility. Early geochronologi-  
26 cal and paleomagnetic observations (*e.g.* Morgan, 1981), and studies of the  
27 uncertainties of plate-reconstruction circuits (*e.g.* Duncan, 1981) suggested  
28 negligible Indo-Atlantic plume motions during the last 100 Myr. In con-  
29 trast, more recent analyses of geochronological and paleomagnetic datasets  
30 suggested either a true polar wander episode (*e.g.* Koivisto et al., 2014), a  
31 change in Pacific plate motion (*e.g.* Torsvik et al., 2017), a southward mo-  
32 tion of the Hawaiian plume reaching  $4 \text{ cm yr}^{-1}$  between 81 and 47 Ma (*e.g.*

33 Tarduno et al., 2003), or a combination of plume and plate motion (*e.g.* Fin-  
34 layson et al., 2018; Konrad et al., 2018) to explain the bent Hawaii-Emperor  
35 hotspot track. Petrological data also suggest that the Azores plume has  
36 drifted northwards by 1-2 cm yr<sup>-1</sup> along the Mid-Atlantic ridge over the last  
37 85 Myr (Arnould et al., 2019).

38 Numerical and laboratory experiments give independent constraints on  
39 mantle plume behaviour. The viscosity contrast between plume conduits and  
40 the ambient mantle (*e.g.* Jellinek and Manga, 2004)) may influence their dy-  
41 namics and stability. Mantle convection has also been shown to contribute  
42 to plume motion. A highly viscous lower mantle (Richards, 1991) or the  
43 anchoring of plume conduits along the edges of dense basal thermochemi-  
44 cal heterogeneities (*e.g.* Davaille et al., 2002) representing Large Low Shear  
45 Velocity Provinces (LLSVPs), is expected to stabilise plumes, while lateral  
46 mantle flow, sometimes called *mantle wind* (Duncan and Richards, 1991;  
47 Richards and Griffiths, 1988), would favour highly tilted plumes (Steinberger  
48 and O’Connell, 1998; O’Neill et al., 2005). Finally, plate tectonics have been  
49 proposed to promote plume stability via ridge capture (*e.g.* Tarduno et al.,  
50 2009). Subduction can also indirectly induce plume motions through the  
51 effect of supercontinent cycles on the planform of global convection (Zhong  
52 et al., 2007) or the lateral push of plume conduits by lower mantle slabs  
53 (Hassan et al., 2016).

54 Collectively, the above-mentioned studies highlight the need to investigate  
55 the coupled behaviour of mantle plumes, plate tectonics, large-scale mantle  
56 flow and basal thermochemical structures. Here we use time-dependent 3D-  
57 spherical numerical models of whole-mantle convection at Earth-like con-

58 vective vigour and self-generating plate tectonics to jointly investigate the  
59 mechanisms linking plate tectonics, mantle convection and plume dynam-  
60 ics. In these models, drifting plumes arise self-consistently and dynamically  
61 interact with surface tectonics, large-scale and small-scale convection. Fo-  
62 cusing on model plume conduits, we show that their number, lifetime, shape,  
63 temperature excess, rising speed, buoyancy and heat fluxes are comparable  
64 to observations. This serves as a basis to investigate the role of plate tec-  
65 tonics, plate layout, convective vigour of the lowermost mantle and basal  
66 thermochemical structures on plume spatio-temporal dynamics.

## 67 **2. Methods**

### 68 *2.1. Numerical models of mantle convection with plate-like tectonics*

69 We solve the equations of mass, momentum, energy and advection of  
70 compositional fields under the Boussinesq approximation to produce a series  
71 of 3D-spherical models of mantle convection with plate-like tectonics using  
72 the code StagYY (Tackley, 2008).

73 The reference Rayleigh number  $Ra_0$  used to model Earth-like mantle con-  
74 vective vigour is

$$Ra_0 = \frac{\alpha_0 \rho_0 g \Delta T D^3}{\kappa_0 \eta_0} = 10^7 \quad (1)$$

75 with  $\alpha_0$  the surface thermal expansivity,  $\rho_0$  the reference density,  $g$  the gravi-  
76 tational acceleration,  $\Delta T$  the temperature difference across the whole mantle  
77 domain,  $D$  the mantle domain thickness,  $\eta_0$  the reference viscosity and  $\kappa_0$   
78 the reference thermal diffusivity. All parameters are listed in Table 1.

79 Viscosity varies both radially and laterally, and depends on both temper-

80 ature and pressure as:

$$\eta(T, z) = \eta_0(z) \exp\left(A + \frac{E_a + \Pi(z)V_a}{RT}\right) \quad (2)$$

81 with  $A$  chosen *a priori* so that  $\eta = \eta_0$  when  $T = 0.64$  (dimensional equivalent  
 82 1,600 K) at the lithosphere-asthenosphere boundary,  $E_a$  the activation en-  
 83 ergy,  $V_a$  the activation volume,  $\Pi(z)$  the static pressure,  $R$  the gas constant,  $z$   
 84 the depth and  $T$  the temperature. The activation energy and volume used in  
 85 the models produce seven orders of magnitude of viscosity variation through-  
 86 out the domain (Fig. S1). We impose a viscosity increase by a factor of 30 at  
 87 660 km depth, consistent with geoid and postglacial rebound studies (Naki-  
 88 boglu and Lambeck, 1980; Ricard et al., 1993). Although it remains debated  
 89 (*e.g.* Rudolph et al., 2015), the chosen viscosity structure is comparable to  
 90 most numerical studies (*e.g.* Steinberger and Calderwood, 2006). In order  
 91 to achieve self-generation of a plate-like behaviour, we use a pseudo-plastic  
 92 rheology in which viscosity depends on the yield stress (Tackley, 1998):

$$\eta_Y = \frac{\sigma_Y}{2\dot{\epsilon}_{II}} \quad (3)$$

93 with  $\eta_Y$  the yielded viscosity,  $\sigma_Y = \sigma_{Y_0} + z \times d\sigma_Y$  the yield stress where  
 94  $\sigma_{Y_0}$  is the yield stress at the surface and  $d\sigma_Y$  is its depth-dependence.  $\dot{\epsilon}_{II} =$   
 95  $\sqrt{0.5\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}}$  is the second invariant of the strain rate.

96 We consider five models to investigate the effect of mantle and subduc-  
 97 tion dynamics on mantle plumes characteristics and drift (Fig. 1, Table  
 98 2). In Model 1 (no yielding), we test the effect of a stagnant lid on plume  
 99 stability. This model does not consider continents, contrary to Models 2-  
 100 5. The surface yield stress of oceanic lithosphere is equal to 48 MPa and  
 101 to 27 MPa in Models 2 and 3, respectively. These yield stress values are

102 within the range of those determined from earthquake stress release (All-  
103 mann and Schearer, 2009) and from rock mechanics experiments (*e.g.* Brace  
104 and Kohlstedt, 1980), although caution is required when comparing values  
105 inferred from mechanisms occurring at different spatial and temporal scales.  
106 They result in a mobile plate-like regime with slow (Model 3) or fast (Model  
107 2) plate motions. In Models 1-3 thermal expansivity decreases with depth  
108 by a factor of three, to account for less vigorous convection in the lowermost  
109 mantle (Chopelas and Boehler, 1992). In Model 4 the surface yield stress is  
110 the same as in Model 2 but thermal expansivity is constant with depth, to  
111 test the effect of a more vigorous lowermost convection on plume drift. In  
112 Model 5, thermal expansivity varies with depth, the surface oceanic litho-  
113 sphere yield stress is equal to 61 MPa and an uppermost weak-crust layer of  
114 14 km with a yield stress equal to 13 MPa favours the formation of asymmet-  
115 ric subduction (Cramer et al., 2012). Model 5 is the last model of Coltice  
116 et al. (2019) for the period 555 to 905 Myr. It includes two basal chemical  
117 heterogeneities that are about 3% denser and 10 times more viscous than  
118 the ambient mantle, initially 500 km thick and laterally homogeneous with  
119 a configuration comparable to present-day LLSVPs (see Fig. S3 of Coltice  
120 et al. (2019)). In Models 2-5, the yield-stress slightly increases with depth  
121 (Table 2). In Models 2, 3 and 4, the average radial resolution is 31 km while  
122 it is 23 km in Model 1 and 5; it is less than 15 km and 10 km near the surface  
123 and the core-mantle boundary respectively. The horizontal resolution at the  
124 surface varies between 26 km (Models 1 and 5) and 35 km (Model 2 - 4).

125 We track the evolution of compositional fields using the tracer-ratio method  
126 (Tackley and King, 2003). Model continents are low-density lithospheric

127 rafts, thicker and stiffer than the oceanic lithosphere to prevent their en-  
128 trainment by the convective flow (Table 1). Despite these characteristics,  
129 their margins are recycled through time (Supplementary Movies 1-3), con-  
130 sistent with estimates of continental recycling at subduction zones (Coltice  
131 et al., 2019). Their initial shapes correspond to continents reconstructed  
132 either 80 Ma ago (separated, Models 2-4 , Fig. S2c) or 200 Ma ago (ag-  
133 gregated, Model 5 Coltice et al. (2019)). In Models 1-4, after a preliminary  
134 equilibration phase of approximately 200 Myr during which tracers are not  
135 advected, we allow continents to dynamically interact within the convective  
136 environment for at least 200 Myr. We then analyse mantle plume behaviour  
137 over a subsequent period of approximately 350 Myr.

138 The use of pseudo-plasticity favours Earth-like surface velocities and tec-  
139 tonics (*e.g.* Coltice et al., 2017). As in Arnould et al. (2018), we verify that  
140 Models 2-5 favour plate-like behaviour with Earth-like plateness (proxy for  
141 the degree of surface deformation localisation, Tackley, 2000), surface mobil-  
142 ity, plate velocities, heat flow and topography (Table 2, Fig. 1, Fig. S3(a),  
143 Movie S1 and S2).

## 144 *2.2. Automated detection of mantle plumes*

145 We developed a plume detection scheme to track plume positions and  
146 characteristics through time in map view. We located mantle plume con-  
147 duits in the upper mantle (at 250, 350 and 670 km depth) and in the lower  
148 mantle (at 1,000 km depth) from their anomalously large radial heat ad-  
149 vection, which is proportional to  $v_z T$  (with  $v_z$  the radial velocity and  $T$  the  
150 temperature, Fig. S3b). Plumes were defined by upper and lower mantle heat  
151 advection equal to  $190 \text{ K m yr}^{-1}$  (Fig. S4) and  $48 \text{ K m yr}^{-1}$  respectively.

152 The greater upper mantle threshold value reflects greater rising speeds. The  
153 use of the temperature (*e.g.* Labrosse, 2002) or radial velocity field alone  
154 (*e.g.* Hassan et al., 2015) is not selective enough to detect mantle plumes in  
155 our models. We chose a unique threshold at each depth to directly compare  
156 different models (Fig. S4). Results were verified visually for artifacts that  
157 could be introduced by the use of threshold values.

158 We used the connectivity method described in Labrosse (2002) in order  
159 to identify distinct plume conduits. We assigned identities (IDs) to plume  
160 conduits, and we inferred their location from their geometrical centroid. We  
161 tracked plumes in 1 Myr intervals by connecting plumes located within a  
162 radius of 500 km with a method similar to that of Hassan et al. (2015),  
163 although with a larger radius to follow plumes moving at several tens of  
164  $\text{cm yr}^{-1}$ . We used this plume tracking method to recover plume positions  
165 and characteristics in the upper and the lower mantle over time.

166 We used *pyGPlates* (Muller et al., 2018) to extract the time-dependent  
167 motion of each mantle plume conduit. We calculated absolute model plume  
168 velocities averaged over 5 Myr to eliminate artificial plume wobbling caused  
169 by rapid readjustments of plume conduit shape. We calculated the relative  
170 motions of all plumes with respect to one another at 5 Myr intervals to  
171 compare them to observations.

172 In StagYY, net rotation is zero at the surface. Therefore, in our models,  
173 “net rotation” is the net rotation of the mantle with respect to a fixed surface.

### 174 **3. Phenomenology of model mantle plumes**

175 We first describe and quantify the spatio-temporal characteristics of model  
176 mantle plumes. The goal is to compare them to those deduced from obser-  
177 vations to assess the applicability of our models to Earth.

#### 178 *3.1. Number*

179 On average, we detected between 15 (Models 1 and 5) and 35 upper man-  
180 tle plumes (Model 4) (Fig. 2a). The number of plumes remains relatively sta-  
181 ble over the considered timescale. However, longer model evolutions suggest  
182 that supercontinent cycles influence the number of active plumes (Coltice  
183 et al., 2019). The number of mantle plumes increases with plate-like be-  
184 haviour (Models 2 and 3 versus Model 1) and with increasing core mantle  
185 boundary (CMB) heat flow. Indeed, the 15 % larger CMB heat flow due to  
186 more vigorous lower mantle convection in Model 4 (Fig. 3f) leads to approx-  
187 imately 30 % more plumes than in Model 2 (Fig. 2a) while the presence of a  
188 basal thermochemical layer in Model 5 results in a CMB heat flow half that  
189 of Model 2 (Fig. 3f), and in a smaller number of plumes than in Model 2  
190 (Fig. 2a).

191 The number of detected plumes is in broad agreement with other numer-  
192 ical models of whole-mantle convection although model parameters (includ-  
193 ing physical assumptions and surface boundary conditions) differ (*e.g.* Davies  
194 and Davies, 2009; Hassan et al., 2015; Li and Zhong, 2019). On Earth, the  
195 number of hotspots of lower mantle origin remains debated. Courtillot et al.  
196 (2003) used strict criteria to propose that only seven hotspots have a deep  
197 origin. Doubrovine et al. (2012) used only five deep mantle plumes to con-

198 strain absolute plate motions. Wang et al. (2018) proposed 24 deep-rooted  
199 mantle plumes from criteria less strict than Courtillot et al. (2003).

### 200 *3.2. Lifetime*

201 In Models 2-4, 30% of detected plumes exist for at least 200 Myr (Fig.  
202 2b). In Model 1, all but one mantle plumes exist throughout the model  
203 evolution, suggesting that the absence of plate-like tectonics leads to long-  
204 lived mantle plumes. In contrast, the lifetime of almost all plumes in Model  
205 5 is shorter than 150 Myr (Fig. 2b), potentially because the deformation of  
206 basal thermochemical structures by slabs perturbs the buoyancy field that  
207 triggers and sustains mantle plumes (Heyn et al., 2018).

208 These results are compatible with the longest lived oceanic hotspot tracks  
209 for St-Helena (about 120 Myr), Tristan (about 120 Myr) and New England  
210 (about 130 Ma) plumes (Duncan, 1984; Williams et al., 2015). Shorter-  
211 lived hotspot tracks (e.g. Hawaii and Louisville active at least from 87 Ma)  
212 terminate at subduction zones, but may have been active longer, the po-  
213 tential oldest parts of their tracks having been subducted (*e.g.* Portnyya-  
214 gin et al., 2008). Finally, geochemical analyses suggest that South Pacific  
215 Cook–Austral plumes have been active for at least 120 Ma (*e.g.* Konter et al.,  
216 2008).

### 217 *3.3. Shape and radius*

218 In all models, large plume heads initiate in the lowermost mantle and  
219 reach the uppermost mantle (Movies S1 and S2). Most plumes consist of one  
220 individual conduit (Fig. 1). Some distinct upper mantle plume conduits are  
221 merged in the lower mantle (Fig. S8 and Movie S3). This is consistent with

222 seismic tomography models that suggest a connection of Ascension and St-  
223 Helena plumes below the transition zone (Montelli et al., 2006), of the Azores,  
224 the Canaries and Cape-Verde below 1,400 km depth and of Kerguelen and  
225 Crozet below 2,300 km depth (Davaille and Vatteville, 2005).

226 We quantified mantle plume tilt angle to investigate the possible effect of  
227 lateral mantle flow on their motions (Richards and Griffiths, 1988). Model  
228 plume conduits are generally vertical: the average tilt angle between 1,000 km  
229 depth and 350 km depth is approximately  $5.5^\circ$  in Models 1-4 (Fig. 4a) and  
230  $16.5^\circ$  in Model 5. In all models, only a limited proportion of plume conduits is  
231 deflected by more than  $10^\circ$ ; tilt angles rarely reach the  $60^\circ$  maximum used in  
232 plume-particle backward advection models (*e.g.* Steinberger and O’Connell,  
233 2000). Mantle plume conduits are more likely deflected in the upper mantle  
234 than in the lower mantle, consistently with seismic tomography (French and  
235 Romanowicz, 2015): in Model 2, the upper-mantle average tilt angle is equal  
236 to  $15^\circ$ , compared to  $28^\circ$  in Model 5 (Fig. 4b), due to upper mantle shearing  
237 by surface plates, which can also cause deformation of the thermal trail of  
238 plumes (Fig. 4b).

239 The average upper mantle plume radius varies from 130 km for stagnant-  
240 lid convection (Model 1) to 75 km for vigorous lower mantle convection  
241 (Model 4). Buoyancy fluxes deduced from hotspot topographic swells (Sleep,  
242 1990) indicate that the average radius of upper mantle plume conduits is of  
243 the order of 100 km. This radius is small compared to seismic tomography  
244 resolution, which explains the difficulties in imaging plume conduits.

245 The radius of plume conduits is systematically greater in the lower mantle  
246 than in the upper mantle (377 km in Model 1 at 1,000 km depth to 131 km

247 in Model 4, Fig. 1 and S6a). The difference in plume conduit radius between  
248 the lower and upper mantle is due to the 30 fold viscosity jump imposed at  
249 670 km depth (*e.g.* Leng and Gurnis, 2012). The presence of thermochem-  
250 ical heterogeneities within plumes (Davaille and Vatteville, 2005) and the  
251 visco-plastic rheology of the lower mantle (Davaille et al., 2018) might partly  
252 explain why lower mantle plumes are broader in tomography models (French  
253 and Romanowicz, 2015).

254 In Model 5, all plumes but one are anchored to basal thermochemical  
255 heterogeneities (about 120 different hotspots were detected over 350 Myr).  
256 Plumes originate from topographic crests (Fig. 1), either along the edges or  
257 on the top of the basal thermochemical heterogeneities, as in previous models  
258 that consider dense and hot basal material (Garnero and McNamara, 2008;  
259 Hassan et al., 2015; Li and Zhong, 2017, 2019).

### 260 3.4. *Excess temperature*

261 Model plume average excess temperature is defined as the difference be-  
262 tween the average temperature over their cross-sectional area and the aver-  
263 age mantle temperature at 350 km depth. The average excess temperature  
264 of model plume conduits over their cross-sectional area is + 140 K in Model  
265 1 and + 200 – 225 K in Models 2-5 (Fig. 3b), which is consistent with petro-  
266 logical studies (*e.g.* Putirka, 2005). The lower excess temperature in Model  
267 1 results from the stagnant-lid regime leading to a larger ambient mantle  
268 temperature in this model. The average excess temperature decreases with  
269 depth along plume conduits (Fig. S6b): it is for example 450 K in average  
270 for Model 2 at 1000 km depth.

271 *3.5. Buoyant rising speed*

272 In Models 1-4 (thermal plumes only), plume material rises in the upper  
273 mantle at approximately 30 cm yr<sup>-1</sup> on average (Fig. 3c).

274 The average rising speed is consistent with the modified Stokes velocity  
275 (Richards and Griffiths, 1988):

$$v_p = \frac{k_p \Delta \rho g r_p^2}{\eta} \approx 24 \text{ cm yr}^{-1} \quad (4)$$

276 with  $\eta \approx 1 \cdot 10^{20}$  Pa s the viscosity of the upper mantle in our models (Fig.  
277 S1),  $k_p \approx 0.54$  a geometrical constant determined experimentally,  $\Delta \rho =$   
278  $\alpha \rho_m \Delta T \approx 22 \text{ kg m}^{-3}$  the density contrast between thermal plumes and the  
279 surrounding upper mantle (where  $\Delta T \approx 200 \text{ K}$ ,  $\alpha$  is the thermal expansivity  
280 and  $\rho_m$  is the reference density),  $g$  the gravitational acceleration and  $r_p \approx$   
281 80 km the plume radius.

282 In Model 5 (thermochemical plumes), buoyant rising upper mantle speeds  
283 are only approximately 17 cm yr<sup>-1</sup> because entrained denser basal material  
284 decreases the positive thermal buoyancy of plume material.

285 *3.6. Buoyancy flux*

286 We calculated the buoyancy flux of mantle plumes as in Sleep (1990) (Fig.  
287 3d):

$$B_p = \rho_m \alpha \Delta T A_p v_p \quad (5)$$

289 with  $\rho_m$  the reference mantle density,  $\alpha$  the reference thermal expansivity,  
290  $\Delta T$  the mantle plume temperature excess,  $A_p$  the cross-sectional area of man-  
291 tle plume conduits and  $v_p$  the buoyant rising speed. In Models 2-4, plume  
292 buoyancy flux ranges between 0.01 and  $20 \times 10^3 \text{ kg s}^{-1}$  and is  $\sim 5 \times 10^3 \text{ kg s}^{-1}$

293 on average. In Model 1, it is  $8 \times 10^3 \text{ kg s}^{-1}$  on average because of the larger  
294 cross-sectional area of mantle plumes, and in Model 5 it is  $2.5 \times 10^3 \text{ kg s}^{-1}$   
295 because of the lower plume rising speed. This range of model plume buoy-  
296 ancy fluxes is comparable to estimates for present-day hotspots (Crosby and  
297 McKenzie, 2009).

### 298 3.7. Heat flow

299 Plume heat flow was derived from the buoyancy flux as in Sleep (1990):

$$300 \quad F_p = B_p \frac{c_p}{\alpha} \quad (6)$$

301 where  $c_p$  is the mantle heat capacity in  $\text{J kg}^{-1}\text{K}^{-1}$ . Individual plumes ad-  
302 vect about 0.08-0.23 TW of heat on average in Models 1-5 (Fig. 3e). In  
303 Models 1-5, the core heat flow varies between 3 TW (Models 5) and 6.3 TW  
304 (Model 4, Fig. 3f), and never exceeds 20% of the surface heat flow (Fig.  
305 S5), which is at the lowest end of observational constraints (5-15 TW, Lay  
306 et al., 2008). The amount of heat carried by mantle plumes at 350 km depth  
307 represents  $\sim 3$  TW in Models 1-4 and  $\sim 1.5$  TW in Model 5, about 2 TW  
308 less than the corresponding heat flow (Fig. 3f). Davies (1988) and Sleep  
309 (1990) estimated the contribution of plumes to the total surface heat flow as  
310 2.5 TW from hotspot swells. At 350 km depth, model mantle plumes carry  
311 a similar amount of heat. The ratio of upper-mantle plume heat flow to core  
312 heat flow (about 50-60% in Models 1-5) is consistent with numerical models  
313 of incompressible and isoviscous (Labrosse, 2002) or temperature-dependent  
314 viscosity convection with different internal heating rates (Mittelstaedt and  
315 Tackley, 2006).

316 The amount of heat carried by each plume is about 1.5 times larger at  
317 1000 km depth than at 350 km depth (Fig. S6). Mittelstaedt and Tackley  
318 (2006) showed that plumes contribute to heat up sinking slabs, therefore  
319 losing some heat on their way up (Fig. S6). Using extended-Boussinesq (Leng  
320 and Zhong, 2008; Zhong, 2006) and compressible models (Bunge, 2005), it  
321 was also suggested that the proportion of core heat flux advected by plumes  
322 decreases towards the surface following a steep plume adiabatic gradient.

### 323 *3.8. Pulses of activity*

324 The activity of a given mantle plume progressively decreases through  
325 time, as evident in the decrease of plume maximum temperature excess from  
326 + 500 K to + 200 K for a hotspot in Model 2 (Fig. 2d). This temperature  
327 decrease is consistent with geological estimates of the temperature evolu-  
328 tion of the Galapagos and Iceland plumes (Herzberg and Gazel, 2009). The  
329 long-term decrease of the activity of a plume is punctuated by shorter-term  
330 fluctuations in flux by about  $\pm 1 \times 10^3 \text{ kg s}^{-1}$  on 5-10 Myr timescales. Changes  
331 in plume buoyancy flux tend to occur after plume merging events or after  
332 interaction with a ridge (Fig. 2d).

### 333 *3.9. Absolute motions of plumes*

334 In Model 1 (stagnant lid), the average absolute drift of plumes in the  
335 upper mantle is  $0.6 \text{ cm yr}^{-1}$ . In contrast, the average absolute lateral  
336 speed of upper mantle plumes is approximately  $1.8 \text{ cm yr}^{-1}$  in Models 2  
337 and 3,  $2.2 \text{ cm yr}^{-1}$  in Model 4 (more vigorous lower mantle convection), and  
338  $2.4 \text{ cm yr}^{-1}$  in Model 5 (deep thermochemical piles), which is characterised  
339 by more slowly rising plumes (Table 2). Average absolute lateral speeds of

340 plume conduits in the lower mantle are 15 to 40% lower than in the up-  
341 per mantle, depending on model parameters; they vary between  $0.4 \text{ cm yr}^{-1}$   
342 (Model 1) and  $2 \text{ cm yr}^{-1}$  (Model 5). This reflects that some plumes are tilted  
343 by vigorous upper mantle convection.

### 344 *3.10. Relative motion between plumes*

345 We quantified the evolution of the relative motion of pairs of plumes  
346 (Table 2 and Fig. 5) to study the processes controlling plume lateral motion.  
347 This allowed us to statistically compare the lateral motions of model plumes  
348 with observations without considering the global reference frame of tectonic  
349 reconstructions. The average relative speed between mantle plumes in the  
350 upper mantle is  $0.6 \text{ cm yr}^{-1}$  in Model 1,  $1.6 \text{ cm yr}^{-1}$  in Models 2 and 3, and  
351  $2.2 \text{ cm yr}^{-1}$  in Models 4 and 5. Relative velocities between mantle plumes are  
352 systematically lower than the average absolute motion of plumes in models  
353 with a plate-like tectonic behaviour, which suggests a coherent drift of at  
354 least some mantle plumes (see section 4.2). However, in these models, the  
355 net motion of the mantle relative to the surface has a small effect on plume  
356 lateral motion because it is only about  $0.03\text{-}0.04 \text{ }^\circ \text{ Myr}^{-1}$  (Table 2), at the  
357 low end of net rotation estimates (Tetley et al., 2019).

358 More than 76% of model plumes move by less than  $2 \text{ cm yr}^{-1}$  depending  
359 on model parameters (Fig. 5a). This is comparable to results from Li and  
360 Zhong (2019), after the removal of the net motion of the mantle from their  
361 modeled plume motions; in their models, large mantle net rotation results  
362 from both imposed surface boundary conditions containing net-rotation up  
363 to  $0.55 \text{ }^\circ \text{ Myr}^{-1}$ , and stronger viscous coupling between the lithosphere and  
364 asthenosphere.

365 Doubrovine et al. (2012) produced a global moving hotspot reference  
366 frame from the best fit of advected tracers within a reconstructed mantle  
367 flow during the last 130 Ma with the tracks of 44 well-known hotspots. They  
368 found that limited plume motion (drift of less than  $3 \text{ cm yr}^{-1}$  for any plume,  
369 Fig. 5a) and large plume deflection could explain observed hotspot tracks.  
370 This scenario requires plumes to drift slower than in Models 2-5. How-  
371 ever, this reference frame results in large lithospheric net rotation (Table 2),  
372 and it neglects potential large uncertainties on both hotspot track ages and  
373 paleo-locations, as well as uncertainties in backward-advection mantle flow  
374 models. To overcome the limitations associated with the use of uncertain pa-  
375 leomagnetic estimates and incomplete geological record, Tetley et al. (2019)  
376 generated an absolute plate motion model assuming fixed hotspots and opti-  
377 mising trench motion, lithospheric net rotation and hotspot track geometry.  
378 From the difference between nine modeled and observed hotspot tracks, they  
379 estimated relative mantle plume motions as  $2.07 \pm 0.8 \text{ cm/yr}$ , which is 85%  
380 larger than Doubrovine et al. (2012). Our results are compatible with that  
381 of Tetley et al. (2019) (Fig. 5a and Table 2).

### 382 *3.11. Topographic swell*

383 In all models, the buoyancy force of mantle plumes leads to the formation  
384 of a surface topographic swell 1000-2000 km across and 1-2 km high (Fig. 1,  
385 S3 and S5). The swell extent and amplitude depend on 1/ plume buoyancy  
386 flux that decreases over time (Fig. 2 and Movies S1 and S2), 2/ the nature of  
387 the impacted lithosphere (thickness, type, plate boundary proximity) and 3/  
388 the relative motion between the plume and the lithosphere, which can shear  
389 the conduit and result in an asymmetric shape of the plume trail (Fig. S2,

390 Movies S1 and S2; see also Arnould et al., 2019).

### 391 3.12. *Plume-ridge interactions*

392 In Models 2-4, 30 to 50% of mantle plumes interact with ridges at one  
393 point. This proportion increases to 45-80% for Model 5 in which the weaker  
394 lithosphere is more likely to break above mantle plumes. The increasing  
395 number of plumes interacting with ridges during the last 150 Myr in Model 5  
396 represents a rise of total ridge length during a significant rifting event, with  
397 the newly formed ridges connecting over preexisting mantle plumes (Coltice  
398 et al., 2019). This preferential relation between ridges and mantle plumes  
399 was first noted by Wilson (1965).

400 Some model plumes interacting with ridges are relatively stable ( $< 2 \text{ cm yr}^{-1}$ )  
401 and stay in the vicinity of a long-lived ridge for at least 50 Myr (Fig. 6a).  
402 This case represents the proposed long-term interactions between the Mid-  
403 Atlantic ridge and Tristan, St-Helena and the Azores (*e.g.* Gente et al., 2003).  
404 Plumes moving along or near a ridge can favour ridge-jumps (Fig. 6b), in  
405 a configuration comparable to the interaction of the Nazca ridge with the  
406 Galapagos hotspot (Wilson and Hey, 1995). Such jumps can also occur  
407 when a plume reaches the surface in the vicinity of an existing ridge, result-  
408 ing in the relocation of the ridge axis above the newly-formed hotspot. This  
409 phenomenon has been invoked for the formation of the Elan bank during  
410 the emplacement of the Kerguelen plateau and in the Tasman Sea (Gaina  
411 et al., 2003), although paleomagnetic measurements in these regions are de-  
412 bated (Sandwell and Smith, 1997). This mechanism possibly explains why  
413 slowly-migrating ridges often interact with plumes (Whittaker et al., 2015).  
414 Finally, some plumes can cross ridge axes and contribute to the propagation

415 of a new spreading axis (Fig. 6c). This case shares similarities with the  
416 Réunion hotspot crossing the Central Indian Ridge about 30 Myr ago and  
417 its putative link to the formation of the Rodriguez ridge (Morgan, 1978).

#### 418 **4. Sources of plume drift**

419 We used the above analysis of the characteristics of model plumes to  
420 classify the lateral plume motions into categories. We used Model 1 as a  
421 reference to define four categories reflecting four different processes.

##### 422 *4.1. "Fixed" plumes*

423 In Model 1, 60% of plumes move by less than  $0.5 \text{ cm yr}^{-1}$  relatively to each  
424 other (Fig. 5a). These plumes are rooted at saddle points of basal flow and  
425 therefore remain stable throughout their lifetime. This stability is favoured  
426 by the absence of plate tectonics at the surface, since the number of fixed  
427 plumes decreases to between 25% (Model 5) and 50% (Model 3) when plate-  
428 like behaviour occurs (Fig. 5a). The most stable plumes are still located at  
429 saddle points of lowermost mantle flow (such as Plume IDs 1, 4, 7, 16 and 21  
430 on Fig. S9), which is consistent with the numerical models of Zhong et al.  
431 (2000). In Li and Zhong (2019), the proportion of fixed plumes is 10-20%,  
432 possibly due to the deep mantle rotation rotation induced by imposed plate  
433 velocities (Rudolph and Zhong, 2014).

##### 434 *4.2. Basal mantle flow entrainment*

435 In Model 1, about 30% of model plumes move at relative speeds comprised  
436 between  $0.5$  and  $1 \text{ cm yr}^{-1}$  (Fig. 5a). Such lateral speeds are consistent with  
437 average lower mantle lateral motions (Table 2), suggesting entrainment of

438 these plumes by lateral flow in the lowermost mantle. In Models 2-5, the  
439 number of slowly-moving plumes varies from 25% of the total number of  
440 plumes (Model 4) to 40% (Model 2). The lateral motion of such plumes (e.g.  
441 IDs 18 and 18) coincides in direction and magnitude with lowermost mantle  
442 flow in their vicinity (Fig. 5a and Table 2).

#### 443 *4.3. Slab-induced drift*

444 This category of plumes drifting at speeds between 1-5 cm yr<sup>-1</sup> over sev-  
445 eral tens of million years only occurs with plate-like behaviour (Models 2-5)  
446 and is indirectly related to subduction dynamics. 28% (Model 2) to 40%  
447 (Model 5) of model plumes belong to this category (Fig. 5). In these models,  
448 slabs penetrating in the lower mantle as far as 5,000 km away from man-  
449 tle plumes modify the mantle pressure field by generating regional positive  
450 pressure gradients in the basal thermal boundary layer (Fig. 7a) resulting  
451 in plume motions at about 2 cm yr<sup>-1</sup>. Some close conduits can therefore  
452 be pushed in a similar directions by these regional reorganisations of mantle  
453 flow, such as Plume ID 6, 10 and 15 in Model 2 (Fig. S9). Faster lateral  
454 motions result from complex interactions between plume conduits (which are  
455 subsequently highly deformed) and slabs in the upper mantle and the upper-  
456 most lower mantle (Movies S1, S2 and S3). In Model 5, slabs also contribute  
457 to deforming the edges of thermochemical heterogeneities and therefore in-  
458 directly push mantle plumes. This process is consistent with the scenario  
459 proposed by Hassan et al. (2016) to explain the fast (> 4 cm yr<sup>-1</sup>) drift of  
460 the Hawaii plume between 81 and 47 Ma.

#### 461 4.4. *Rapid drift due to plume merging events*

462 Independent of the presence of surface plate-like motions, all models fea-  
463 ture plumes with lateral speeds exceeding  $5 \text{ cm yr}^{-1}$  and reaching  $10 \text{ cm yr}^{-1}$   
464 (Model 1) to  $22 \text{ cm yr}^{-1}$  (Models 2 and 4). Less than 5% of all mantle plumes  
465 drift at such speeds which occur when two mantle plumes, whose roots are  
466 within a few hundreds of kilometres of one another, merge (Fig. 7b). Plume  
467 merging events have a recurrence time varying between 35 Myr (Models 3  
468 and 5) and 200 Myr (Models 1 and 2). Merging between two mantle plumes  
469 always starts at the base of the mantle due to a basal pressure difference be-  
470 tween two close conduits, and propagates to the upper mantle, where the last  
471 merging steps occur over less than 10 Myr (Fig. 5b and Fig. 7b). Merging  
472 between two mantle plumes has been described in mantle convection models  
473 (Davies and Davies, 2009), but fast lateral motions of merging mantle plumes  
474 have not yet been documented on Earth, which could be explained by the  
475 rare occurrence and short duration of such model events.

## 476 5. Discussion

### 477 5.1. *Limitations of models and analysis*

478 We tested a limited number of parameters potentially affecting plume  
479 drift. For example, we only considered one parameterisation (initial thick-  
480 ness, density and viscosity contrasts) to model basal thermochemical het-  
481 erogeneities, although the nature of LLSVPs remains debated (*e.g.* Garnero  
482 et al., 2016). Moreover, although the resolution at 350 km depth in Mod-  
483 els 2-4 (33 km) is lower than the one of Models 1 and 5 (24 km, similar  
484 to the lateral resolution of Li and Zhong (2019)), the range of the detected

485 cross-sectional areas for all models are consistent (Fig. 3a). Furthermore,  
486 we achieved radial and lateral viscosity contrasts as large as  $10^7$  Pa s at the  
487 expense of model computational time: each model required about 145,000  
488 CPU hours over three months of calculation on a parallel supercomputer to  
489 model mantle flow over 350 Myr.

490 Our plume detection algorithm is based on the hypothesis that all mantle  
491 plumes are characterised by significant heat advection, and therefore by both  
492 substantial temperatures and large rising speeds. This method makes it  
493 possible to characterise model plume dynamics, however results depend on  
494 1/ the ability to detect a potentially large variety of plumes in models with  
495 distinct parameters, and 2/ on the threshold that is set for plume detection.  
496 This threshold was defined by comparing the number of plumes detected by  
497 the algorithm to the number of mantle plumes detected visually.

## 498 *5.2. What controls plume dynamics?*

499 Here, we focus on the causes of lateral motions of already developed  
500 plumes and do not investigate the controls on the position of their source,  
501 which is likely to also affect their dynamics (*e.g.* Heyn et al., 2018; Li and  
502 Zhong, 2017).

### 503 *5.2.1. Indirect control by plate tectonics*

504 Despite these limitations, we show that planetary surface dynamics exerts  
505 a first-order control on plume drift: if the surface is in stagnant lid, stable  
506 or slowly moving plumes predominate. The impossibility for mantle plumes  
507 to drift rapidly in the absence of plate tectonics and lithospheric thickness  
508 heterogeneities was noted by Zhong (2009) who studied the origin of Martian

509 volcanism. Plate-like behaviour promotes faster plume motion (Fig. 5) due  
510 to the interaction of plume conduits with slabs in the upper or the lower  
511 mantle. Nevertheless, we see no significant effect of the number of plates  
512 (comparison between Model 2 and 3) on the statistical distribution of lateral  
513 plume motions.

#### 514 *5.2.2. Limited stabilisation of mantle plumes by mid-ocean ridges*

515 It has been proposed that mantle plumes can be pinned to stable ridges  
516 (Tarduno et al., 2009). In our models, three plume behaviours arise depend-  
517 ing on the type of spreading centre they interact with. A limited number of  
518 plumes pond below and remain fixed to long and stable ridges (e.g. Hotspot  
519 7 in Fig. S9, from 90 Ma), consistently with Whittaker et al. (2015). In con-  
520 trast, plumes may move along a stable ridge due to slab-induced lower mantle  
521 flow during several tens of million years (Hotspots 10 and 15 of Fig. S9). The  
522 possible slow motion of the Azores plume could fall in that category (Arnould  
523 et al., 2019). Finally, mantle plumes interacting with fast-migrating ridges,  
524 usually neighbouring small plates affected by fast reorganisations, tend to  
525 move laterally faster and more erratically (Fig. 6d). We find no statistical  
526 difference in the lateral motion distribution of plumes as a function of their  
527 interacting with ridges (Fig. 8a).

#### 528 *5.2.3. The role of the mantle environment on plume dynamics*

529 Lower mantle convection vigour seems to play a limited role in plume  
530 drift (comparison between Model 2 and 4) while affecting the number of  
531 active plumes. This result is consistent with that of Li and Zhong (2019),  
532 who did not detect any change in the statistical distribution of plume lateral

533 motions when changing the Rayleigh number in mantle convection models  
534 with imposed plate histories.

535 We do not observe significant and systematic plume conduit deflection by  
536 mantle wind (Fig. 4). The strong temperature-dependence of mantle viscos-  
537 ity causes plume material to rise 10-100 times faster than typical ambient flow  
538 in the upper mantle (Fig. 3c). Therefore, plume material rises too quickly to  
539 be substantially deflected (Fig. 4b). This conclusion differs from the results  
540 of numerical models of mantle convection with imposed plate history (*e.g.*  
541 Steinberger and O’Connell, 1998; O’Neill et al., 2005), which advocate for  
542 a strong effect of mantle wind on plume conduit tilting and plume motions  
543 to fit hotspot tracks. However, these models do not consider fully-dynamic  
544 mantle plumes. Instead, passive tracers are advected within a seismically and  
545 tectonically reconstructed mantle flow at a constant rising speed ( $2.2 \text{ cm yr}^{-1}$   
546 in Steinberger and O’Connell (1998) to  $10 \text{ cm yr}^{-1}$  in Steinberger and Antret-  
547 ter (2006)) and with an assigned maximal tilt angle of  $60^\circ$ , constrained by  
548 experimental values on chemical plumes (Whitehead, 1982). We note that  
549 our model plumes generate Earth-like buoyancy fluxes, that our findings are  
550 consistent with other numerical studies of mantle convection with a free-slip  
551 surface (Zhong et al., 2000; Davies and Davies, 2009) and that models with  
552 imposed plate history also generate mainly vertical fully-dynamic plumes (Li  
553 and Zhong, 2019) as opposed to significantly deflected plumes.

554 Model mantle plumes rooted at the edges of basal thermochemical het-  
555 erogeneities tend to be more mobile (Fig. 5a) and are more likely to be  
556 deflected (Fig. 4) than purely thermal plumes (Davaille et al., 2002; Li and  
557 Zhong, 2019) because the entrainment of dense material by plume conduits

558 slows their rise by a factor of two in our experiments.

#### 559 *5.2.4. Plume characteristics are not diagnostic of plume motions*

560 We assess the relationships between model plume lateral velocities and  
561 age (defined as the time since a new plume was first detected), tilt angle,  
562 temperature excess, rising speed and buoyancy flux, to explore whether it  
563 may be possible to estimate plume drifting rates from such observations  
564 for Earth (Fig. 8 and S10). We find that plume age, temperature excess,  
565 rising speed and buoyancy flux do not correlate with absolute plume velocity.  
566 Moreover, plume tilt angles do not correlate with plume drifting rates (Fig.  
567 8b). Instead, fast-moving plumes are characterized by small tilt angles. This  
568 suggests that intrinsic plume characteristics cannot be used as a diagnostic  
569 for plume drift.

#### 570 *5.3. Implications for absolute plate reconstructions*

571 The study of the statistical behaviour of mantle plumes is critical to in-  
572 terpret the paleomagnetic and geochronological record and to build absolute  
573 plate reference frames based on moving hotspots for the last  $\sim 80$  Myr. In  
574 our study, 25% (Model 5) to 50% (Model 3) of plumes move at  $< 0.5 \text{ cm yr}^{-1}$   
575 during several tens of millions of years in models with plate tectonics. This  
576 significant proportion of slow-moving plumes is comparable to the propor-  
577 tion of stable hotspots of Doubrovine et al. (2012) and Tetley et al. (2019),  
578 in which 52% and 30% of Earth plumes move by less than  $0.5 \text{ cm yr}^{-1}$ ,  
579 respectively (Fig. 5a), although we note that observations of plume abso-  
580 lute motions are uncertain, notably for motion along the direction of hotspot  
581 tracks (Li and Zhong, 2019). We can nevertheless propose that such hotspots

582 can be used as a robust non-moving reference to reconstruct plate motions.

583 Moreover, our results show that plumes with different drifting rates can  
584 coexist within the same global convective system, depending on their loca-  
585 tion and potential interactions with slabs, plate tectonics and regional con-  
586 vective flow (Fig. S9). Indeed, model plume velocities can exceed  $2 \text{ cm yr}^{-1}$ ,  
587 consistent with geochronological and paleomagnetic observations for Hawaii  
588 (Tarduno et al., 2003; Konrad et al., 2018) and Kerguelen (Antretter et al.,  
589 2002). Our study therefore reconciles contradictory observations of plume  
590 drift and suggests that defining a global reference frame based on hotspot  
591 tracks to reconstruct past absolute plate motions requires the careful selec-  
592 tion of slow-moving plumes based on paleomagnetic and geochronological  
593 data.

## 594 **6. Conclusion**

595 We presented mantle plumes arising in models of mantle convection self-  
596 generating plate-like tectonics that display excess temperatures, rising speed  
597 and buoyancy fluxes comparable with Earth’s major hotspots. In the absence  
598 of plate tectonics, mantle plumes are long-lived, vertical and fixed. In models  
599 with plate tectonics, plumes rise vertically with deflection  $< 10^\circ$  on average,  
600 usually limited to the more vigorously convecting upper mantle, consistently  
601 with recent tomographic models (French and Romanowicz, 2015). Plume  
602 lifetime is shorter with tectonics than in stagnant lid mode, with half of  
603 plumes existing for less than 50 Myr, although some plumes may exist for  
604 hundreds of million years. The coexistence of several long-lived plumes with  
605 short-lived ones is consistent with observations from LIPs (Large Igneous

606 Provinces) and hotspot tracks.

607 With plate tectonics, plumes are in general mobile. We identified four  
608 distinct groups of plumes:

- 609 1. 25-50% of plumes are fixed to saddle points of the basal mantle flow.  
610 Their velocity is  $<0.5 \text{ cm yr}^{-1}$ . They can be used as reference to  
611 reconstruct absolute plate motions.
- 612 2. 25 to 40% of plumes are entrained by passive mantle flow; they move  
613 at speeds between  $0.5\text{-}1 \text{ cm yr}^{-1}$  without interaction with dynamic  
614 instabilities such as slabs or other plumes.
- 615 3. 30-40 % of plumes are pushed by slabs sinking in the lowermost man-  
616 tle that generate high pressures in the basal boundary layer, pushing  
617 plumes towards lower pressure regions. Such plume motion is between  
618  $1 \text{ and } 5 \text{ cm yr}^{-1}$ , can last several tens of million years, and may explain  
619 the large speeds of the Hawaiian plume (*e.g.* Konrad et al., 2018).
- 620 4. Rare (once every 200 Myr in some models) merging of close plumes due  
621 to a pressure difference in the basal boundary layer. Merging starts in  
622 the thermal boundary layer at a slow pace. Plumes drift at speeds  $>$   
623  $5 \text{ cm yr}^{-1}$  over less than 10 Myr once merging propagates to the upper  
624 mantle.

625 The limited deflection of plume conduits results from their strong tem-  
626 perature dependence of viscosity. The entrainment of basal thermochemical  
627 material by mantle plumes decreases their buoyant rising speed, leading to  
628 enhanced plume deflection, lateral plume motion and shorter lifetime.

629 Our models suggest that ridges are passive enough to interact with plumes  
630 but not to systematically influence the flow where plumes originate. Some

631 plumes leave ridge segments, other move along with a fast-drifting ridge, and  
632 some move along a stable ridge axis.

633 The intrinsic properties of plumes observed from the surface (age, excess  
634 temperature, buoyancy flux, rising speed, tilt angle) are not diagnostic of  
635 plume motion since they do not correlate with lateral plume velocity.

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Table 1: Non-dimensional and dimensional model parameters

<b>Parameter</b>	<b>Non-dim. value</b>	<b>Dim. value</b>
Surface temperature ( $T_{top}$ )	0.12	255 K
Basal temperature ( $T_{bot}$ )	1.12	2240 – 2645 K
Mantle thickness (D)	1	2890 km
Reference thermal expansivity ( $\alpha_0$ )	1	$3 \times 10^{-5} \text{ K}^{-1}$
Reference density ( $\rho_0$ )	1	$4400 \text{ kg m}^{-3}$
Reference diffusivity ( $\kappa_0$ )	1	$1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$
Reference heat capacity ( $c_{p0}$ )	1	$715 \text{ J kg}^{-1} \text{ K}^{-1}$
Reference viscosity ( $\eta_0$ )	1	$1 \times 10^{22} \text{ Pa s}$
Internal heating rate ( $H$ )	40 – 50	$7.54 – 9.43 \times 10^{-12} \text{ W kg}^{-1}$
Activation energy ( $E_a$ )	8	$142 \text{ kJ mol}^{-1}$
Activation volume ( $V_a$ )	3	$13.8 \text{ cm}^3 \text{ mol}^{-1}$
Maximum viscosity cut-off	$10^4$	$10^{26} \text{ Pa s}$
Viscosity factor below 660 km depth	30	
Yield stress gradient for all materials ( $d\sigma_Y$ )	$2.34 \times 10^6$	$1088 \text{ Pa m}^{-1}$
Oceanic lithosphere surface yield stress ( $\sigma_{Y_{0,oc}}$ )	$2 – 200 \times 10^4$	$27 – 270 \text{ MPa}$
Weak-crust surface yield stress (Model 5, $\sigma_{Y_{0,wc}}$ )	$1 \times 10^4$	$13 \text{ MPa}$
Weak-crust thickness (Model 5)	0.005	$14 \text{ km}$
Continental interior surface yield stress ( $\sigma_{Y_{0,cont}}$ )	$7 \times 10^5$	$932 \text{ MPa}$
Continental interior viscosity factor	100	
Continental interior buoyancy number ( $B_{cont}$ )	–0.32	$–150 \text{ kg m}^{-3}$
Continental interior thickness	0.0692	$200 \text{ km}$
Continental belt surface yield stress ( $\sigma_{Y_{belt}}$ )	$3 \times 10^5$	$400 \text{ MPa}$
Continental belt viscosity factor	50	
Continental belt buoyancy number ( $B_{belt}$ )	–0.4	$–188 \text{ kg m}^{-3}$
Continental belt thickness	0.0432	$125 \text{ km}$
Basal thermochemical layer buoyancy number ( $B_{llsvp}$ )	0.25	$117 \text{ kg m}^{-3} (2.7 \%)$
Basal thermochemical layer initial thickness	0.17	$500 \text{ km}$
Basal thermochemical layer viscosity factor	10	

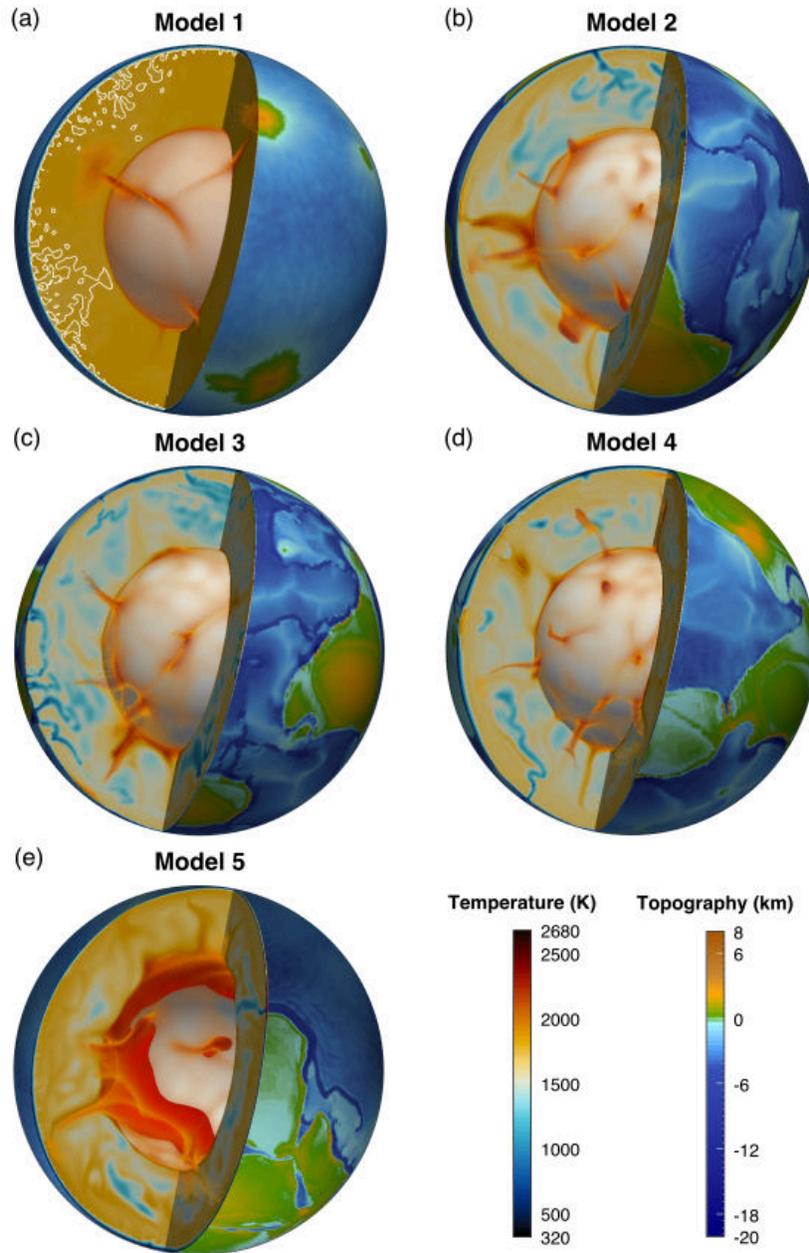


Figure 1: 3D snapshots of (a) Model 1, (b) Model 2, (c) Model 3, (d) Model 4 and (e) Model 5. Temperature is shown in the interior of the shells and topography at their surface. The white isotherm on (a) highlights small-scale convection. The red isosurface on (e) delineates basal thermochemical heterogeneities.

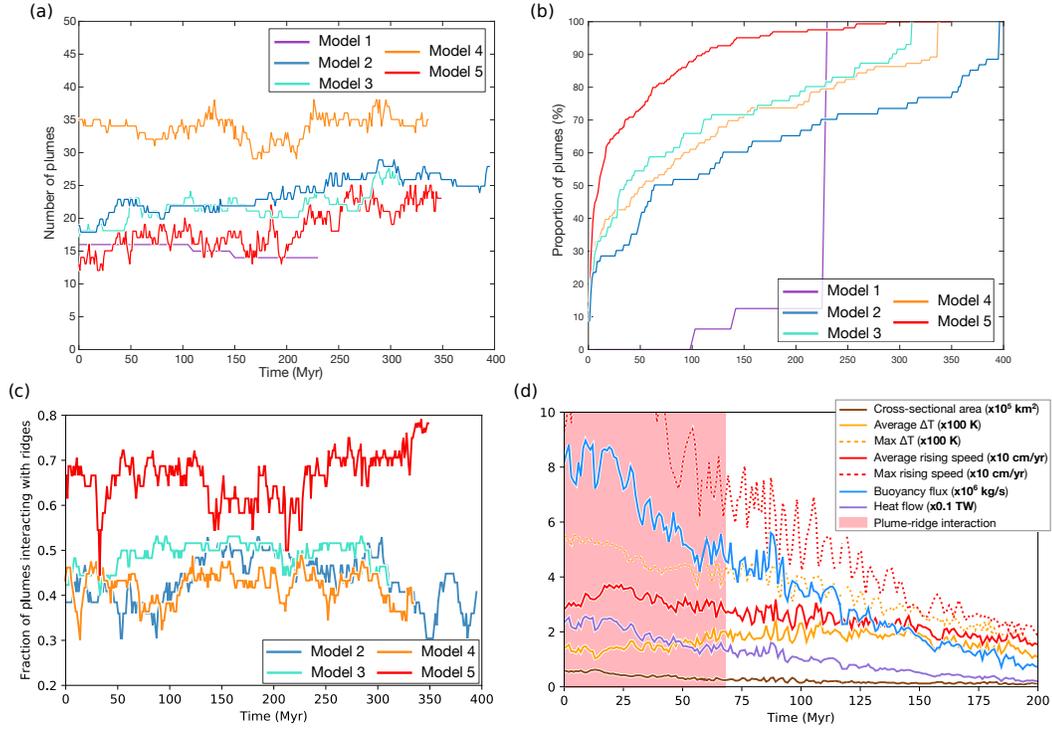


Figure 2: Evolution, at 350 km depth, of (a) the number of plumes, (b) the cumulative lifetime of plumes and (c) the fraction of plumes interacting with ridges. (d) Temporal evolution of plume characteristics for Plume ID 6 in Model 2 (Fig. S3(a) and (b)): cross-sectional area (brown), average (orange) and maximum (orange, dashed) temperature excess, average (red) and maximum (red, dashed) buoyant rising speed, buoyancy flux (blue) and heat flux (purple). The units of each curve is listed in the figure key. The duration of plume-ridge interaction is highlighted by the pink area.

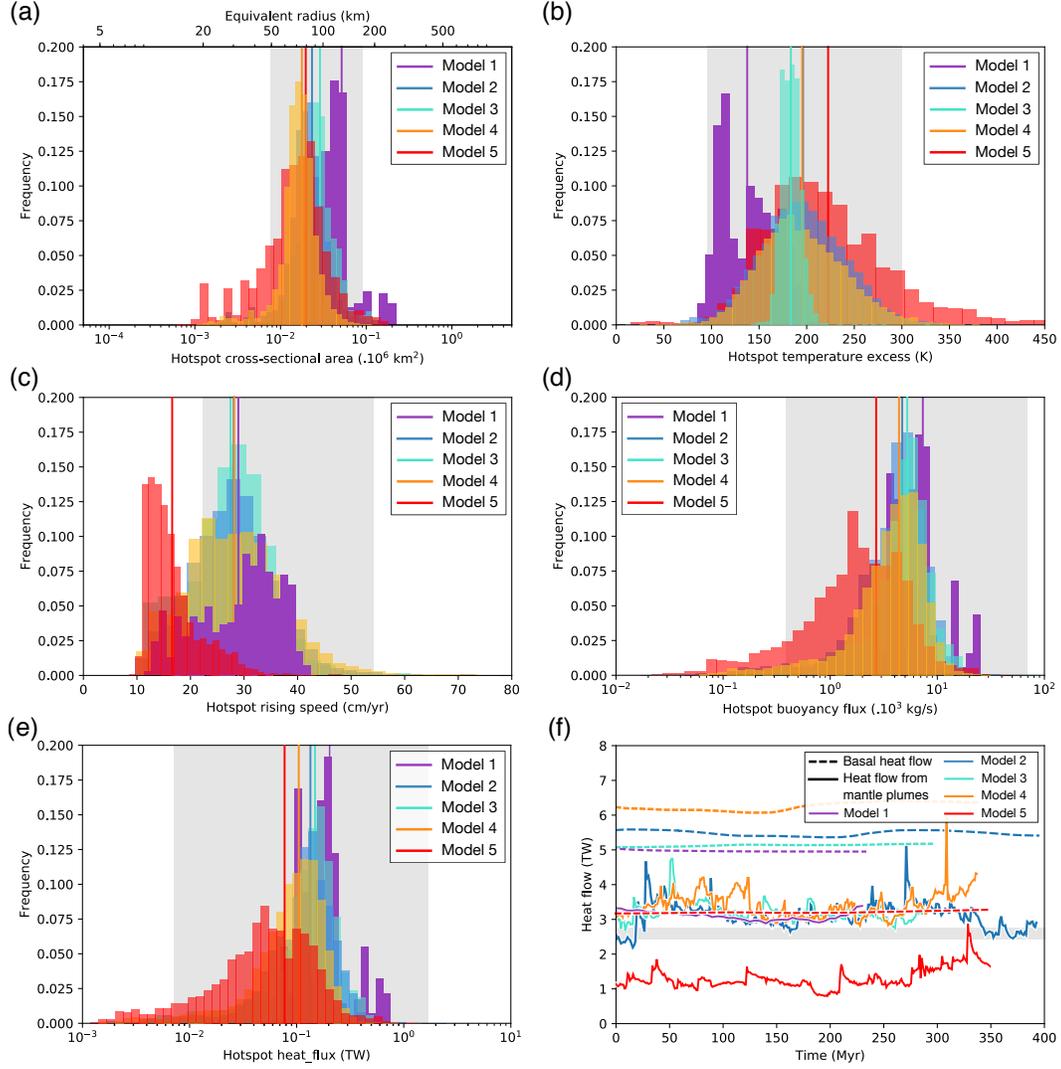


Figure 3: Distribution of (a) the radius, (b) the excess temperature, (c) the buoyant rising speed, (d) the buoyancy flux and (e) the heat flow of mantle plumes for all models at 350 km depth. Temporal evolution of (f) the core and plume heat flow in all models at 350 km depth. Spikes in plume heat flow correspond to the occasional birth of highly-vigorous plumes. The grey areas highlight the range of observational values for each plume characteristic (see text for details).

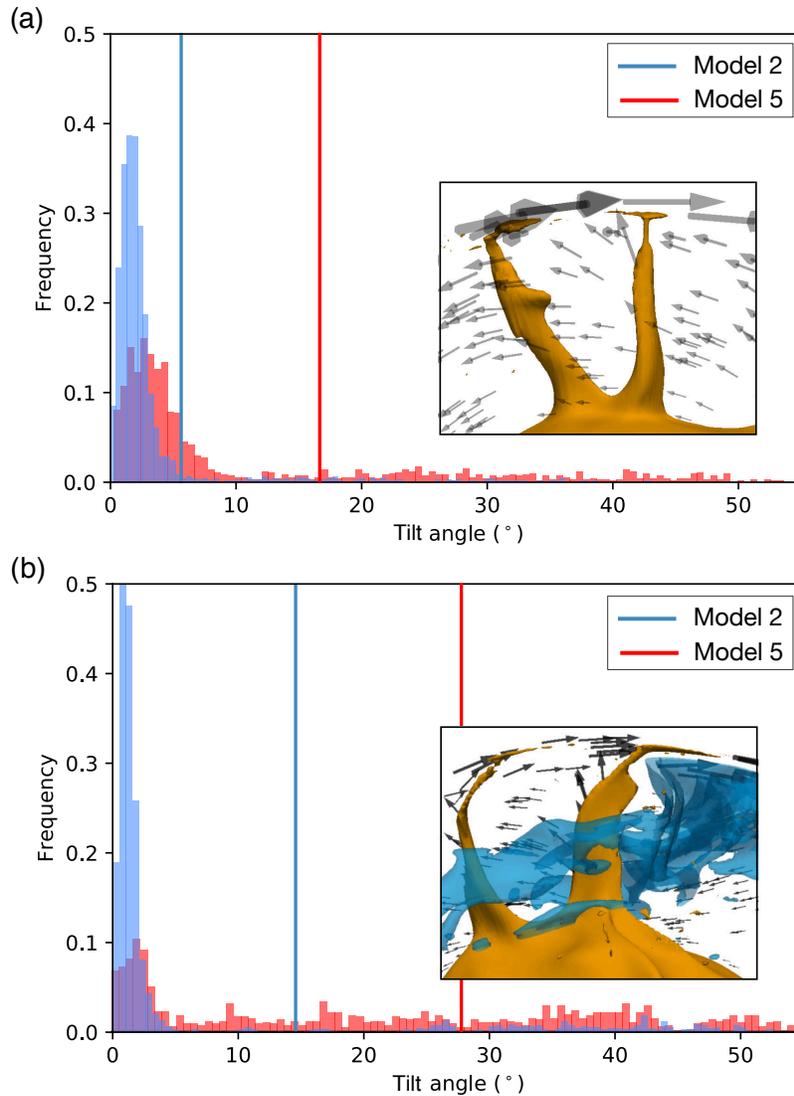


Figure 4: (a) Upper-lower mantle (350-1000 km depth) and (b) upper mantle (150-670 km depth) plume tilt angle distribution for Models 2 and 5. The average tilt angle is shown as a vertical line. Insets show (a) near vertical and (b) tilted conduits (orange isotherm) in Model 5. Transparent black arrows show the direction of mantle flow. In (b) mantle flow deflects plume conduits in the vicinity of subducting lithosphere (transparent blue isotherm). Note that plume deflection mainly occurs in the uppermost mantle.

Table 2: Key model parameters for all models (w.c. = weak crust, depth dep. = depth-dependent thermal expansivity, Continent ages = ages of their initial configuration). Average plateness and surface mobility of all models. Modeled average and standard deviation of surface ( $\mathbf{v}_{\mathbf{rms}}$ ), mantle horizontal ( $\mathbf{v}_{\mathbf{m}}$ ), mantle net rotation (NR) and plume absolute ( $\mathbf{v}_{\mathbf{ha}}$ ) and relative ( $\mathbf{v}_{\mathbf{hr}}$ ) velocities in the upper (UM, between 150 and 410 km depth) and the lower (LM, between 670 km depth and the CMB) mantle. Mobility and plateness are dimensionless.

<b>Parameter</b>	<b>Model 1</b>	<b>Model 2</b>	<b>Model 3</b>	<b>Model 4</b>	<b>Model 5</b>	<b>Dobrovine et al. (2012)</b>	<b>Tetley et al. (2019)</b>
Yield stress (MPa)	no yielding	48	27	48	61 (w.c.)		
Thermal expansivity	depth dep.	depth dep.	depth dep.	constant	depth dep.		
LLSVPs	no	no	no	no	yes		
Continents	no	80 Ma	80 Ma	80 Ma	200 Ma		
Plateness		0.90	0.77	0.84	0.87		
Mobility		1.29	1.69	1.62	1.60		
Surface $\mathbf{v}_{\mathbf{rms}}$ (cm/yr)	0.002	$2.90 \pm 0.85$	$3.32 \pm 0.22$	$3.79 \pm 0.60$	$5.04 \pm 1.83$	2 – 6	1.5 – 4.8
Mean UM $\mathbf{v}_{\mathbf{m}}$ (cm/yr)	$2.34 \pm 0.82$	$3.19 \pm 0.57$	$2.77 \pm 0.32$	$3.60 \pm 0.57$	$4.85 \pm 1.52$		
Mean LM $\mathbf{v}_{\mathbf{m}}$ (cm/yr)	$0.85 \pm 0.25$	$0.91 \pm 0.26$	$0.74 \pm 0.15$	$1.01 \pm 0.23$	$1.23 \pm 0.35$		
Average NR ( $^{\circ}$ /Myr)	$0.001 \pm 0.0003$	$0.039 \pm 0.014$	$0.031 \pm 0.013$	$0.039 \pm 0.016$	$0.029 \pm 0.009$	0.13 – 0.53	0.05 – 0.11
Mean UM $\mathbf{v}_{\mathbf{ha}}$ (cm/yr)	$0.57 \pm 0.45$	$1.80 \pm 1.13$	$1.74 \pm 0.78$	$2.22 \pm 0.82$	$2.40 \pm 0.77$	$1.12 \pm 0.31$	$2.07 \pm 0.80$
Mean UM $\mathbf{v}_{\mathbf{hr}}$ (cm/yr)	$0.63 \pm 0.64$	$1.55 \pm 1.58$	$1.61 \pm 1.18$	$2.20 \pm 1.28$	$2.22 \pm 1.15$	$0.87 \pm 0.35$	$1.46 \pm 1.07$
Mean LM $\mathbf{v}_{\mathbf{ha}}$ (cm/yr)	$0.39 \pm 0.51$	$1.32 \pm 0.56$	$1.05 \pm 0.32$	$1.79 \pm 0.90$	$2.04 \pm 1.25$		
Mean LM $\mathbf{v}_{\mathbf{hr}}$ (cm/yr)	$0.44 \pm 0.68$	$1.02 \pm 0.83$	$0.77 \pm 0.47$	$1.66 \pm 1.37$	$2.01 \pm 1.90$		

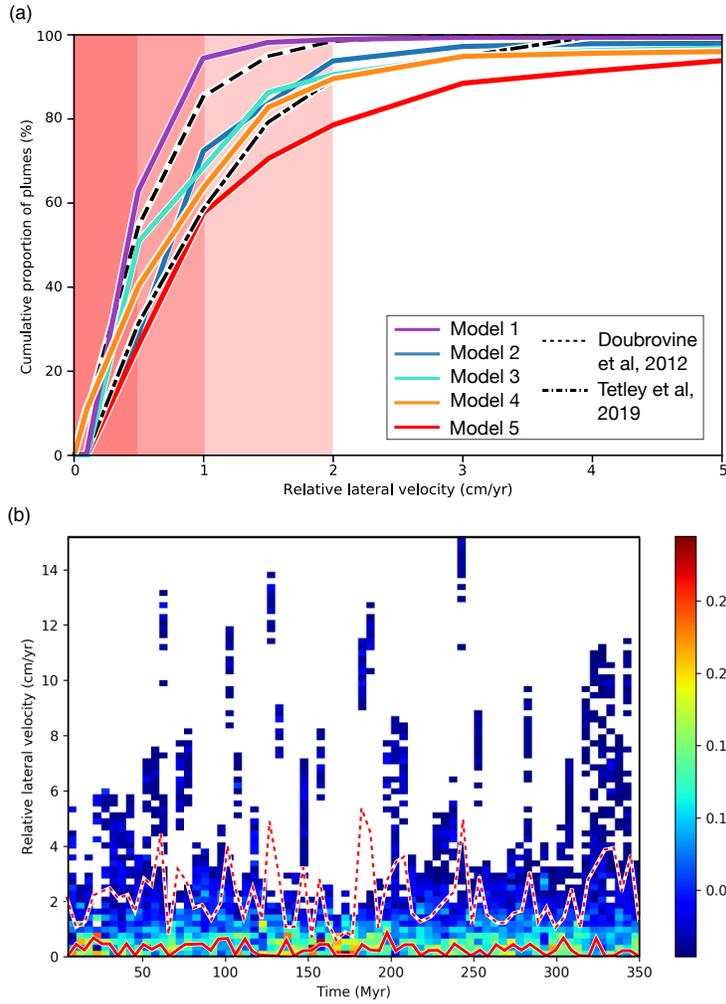


Figure 5: (a) Time-averaged cumulative distribution of relative plume velocities in all models and as estimated for Earth by Doubrovine et al. (2012) and OptAPM1-M16 of Tetley et al. (2019). The transparent red background rectangles highlight the cumulative proportion of plumes moving at less than 0.5, 1 and 2  $\text{cm yr}^{-1}$ . (b) Density plot of the temporal evolution of upper mantle RMS pair-wise lateral relative velocity distribution between mantle plumes in Model 5. Relative velocities are calculated in 5 Myr intervals. The thick red line is the temporal evolution of the mode of relative mantle plume velocities. The dashed red line is the temporal evolution of the average relative velocities between mantle plumes.

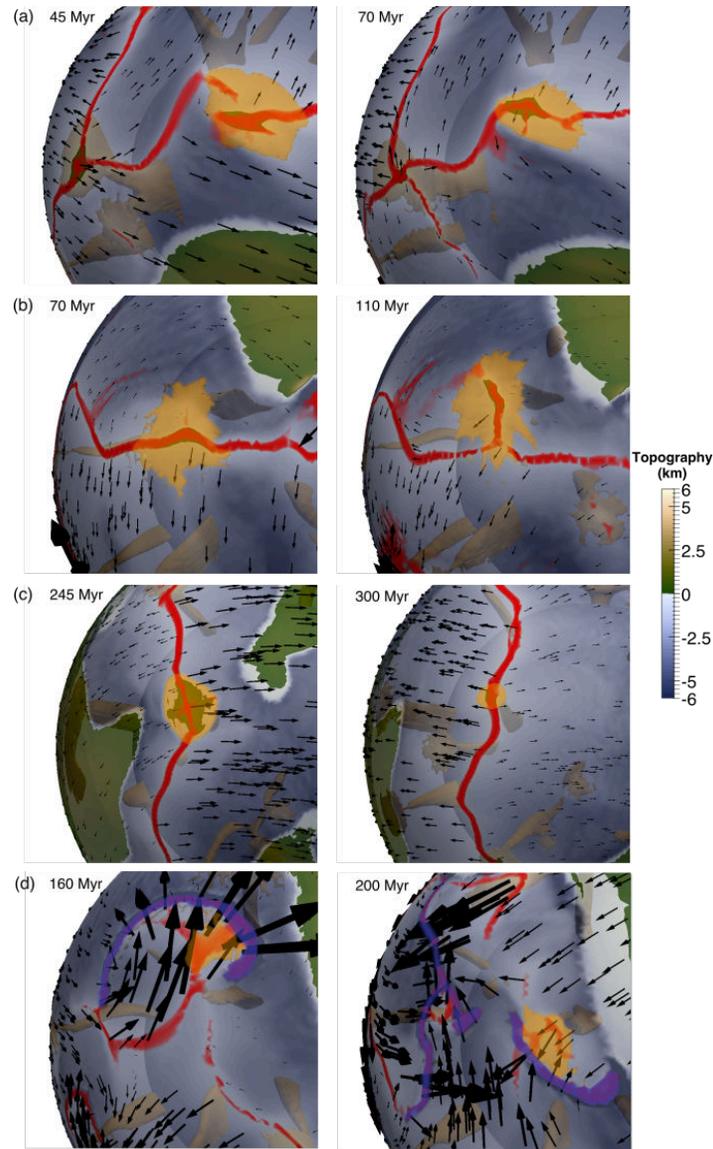


Figure 6: Four cases of ridge-plume interactions in Model 2: (a) plume ponding below a stable ridge (b) ridge-jump in the vicinity of a plume, (c) ridge-propagation induced by a plume, and (d) backarc-plume interaction. Surface topography is outlined as a transparent field. Arrows represent surface velocities. Plume isotherms are seen by transparency in orange. The thermal trail of considered plumes are highlighted in brighter orange. Ridges are in red and subduction zones are in transparent blue. The camera is fixed for each temporal evolution.

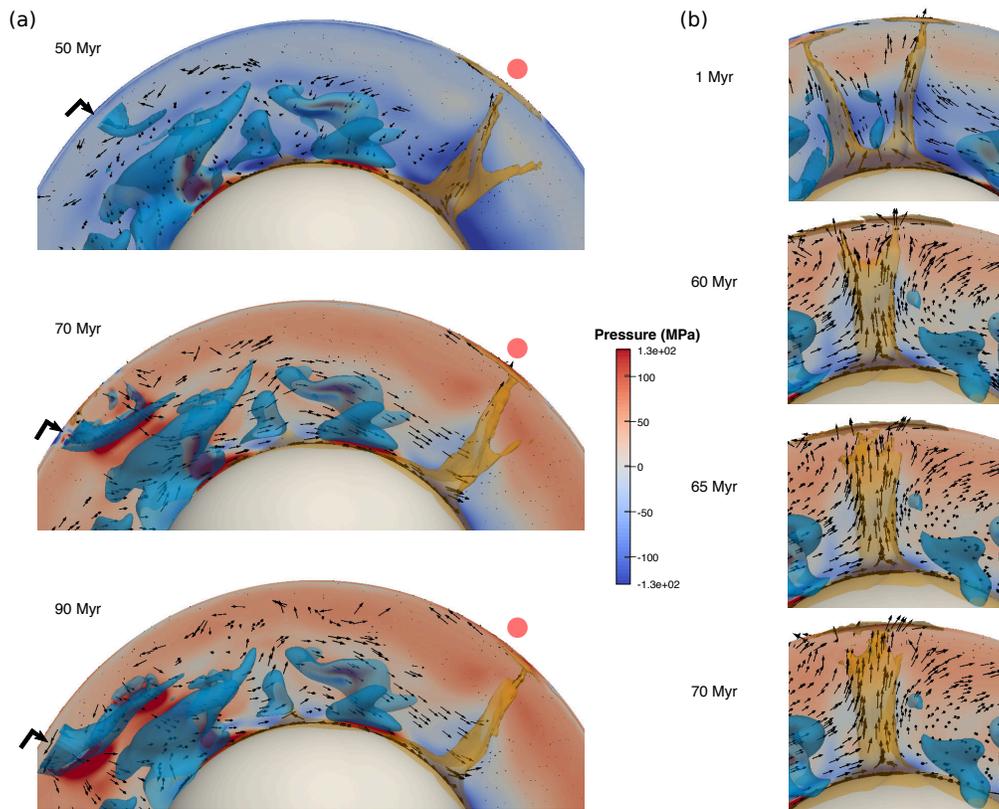


Figure 7: Examples of pressure-gradient-induced plume drifts in Model 2. Mantle plumes are highlighted by orange transparent isotherms and subducting lithosphere is shown as transparent blue isotherms. The velocity field is shown as black arrows. In (a), the position of the plume is indicated at each timestep with a red circle. Subduction (the thick black arrow shows the location of the trench) initiates on the left-hand-side, which results in far-field compression and horizontal lower mantle flow directed towards the plume. (b) Merging of two mantle plumes starting from the base of the mantle and propagating upwards. The camera is fixed.

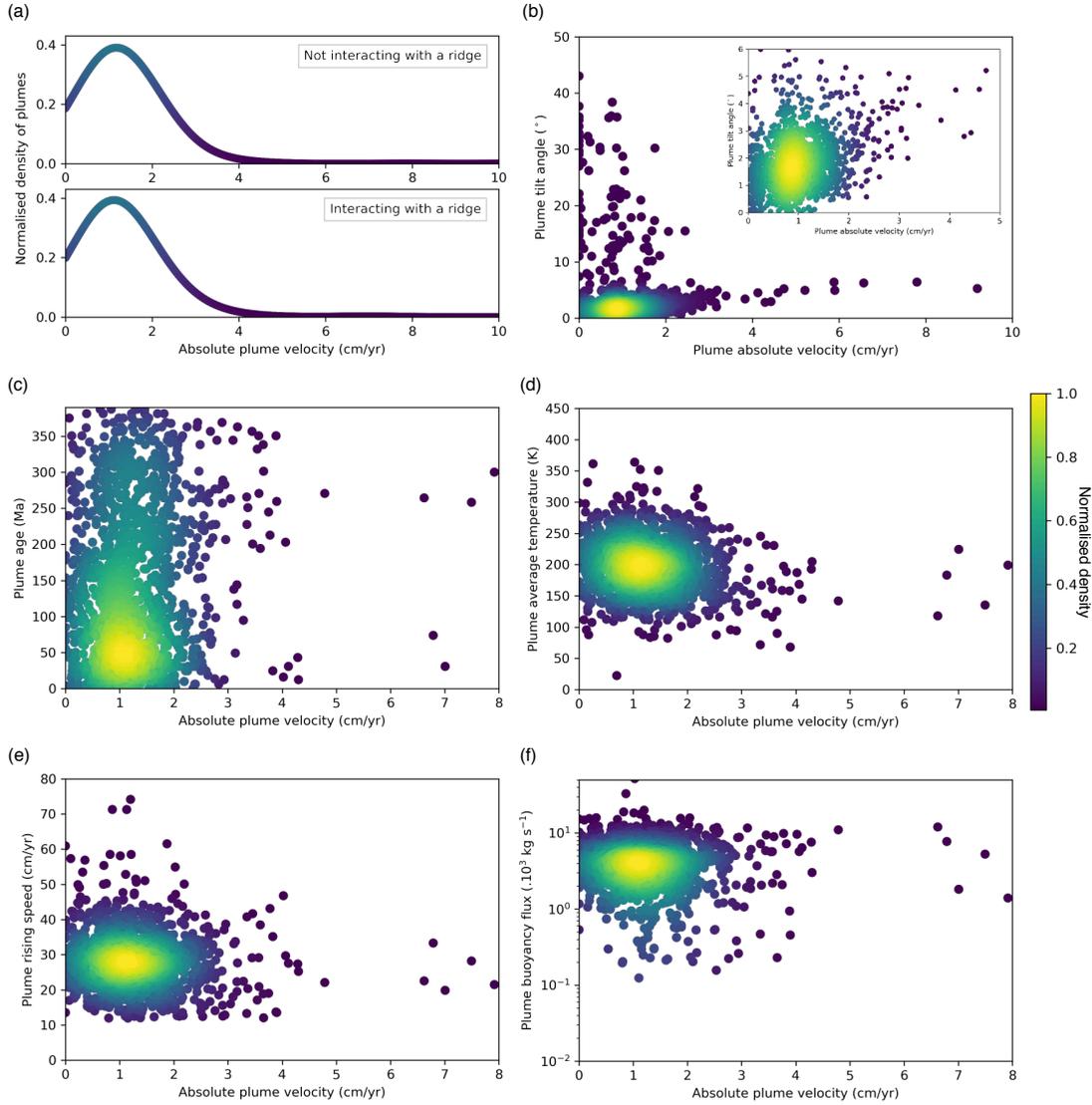


Figure 8: Distribution of (a) plumes interacting (bottom panel) or not (top panel) with ridges, (b) plume tilt angle between 150 and 670 km depth, (c) plume age, (d) plume temperature excess, (e) plume rising speed and (f) plume buoyancy flux versus plume absolute velocity for Model 2. The color scale shows the normalised density of mantle plumes on each plot. The inset on (b) shows a closeup view on the largest density of plumes of that plot.