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 $0.5\text{-}1\text{ cm yr}^{-1}$ are slowly entrained by passive mantle flow. 3/ Fast plume motions between $2\text{-}5\text{ cm yr}^{-1}$ lasting several tens of million years are caused by slab push. 4/ Plumes occasionally drift at speeds $> 5\text{ cm yr}^{-1}$ over $< 10\text{ Myr}$ 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^{-1} 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

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, $> 600\text{ km}$ 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 cm yr^{-1} and 54 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 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⁻¹ 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 α_0 the surface thermal expansivity, ρ_0 the reference density, g the gravi-
 76 tational acceleration, ΔT the temperature difference across the whole mantle
 77 domain, D the mantle domain thickness, η_0 the reference viscosity and κ_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-

ature and pressure as:

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

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

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

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

We consider five models to investigate the effect of mantle and subduction dynamics on mantle plumes characteristics and drift (Fig. 1, Table 2). In Model 1 (no yielding), we test the effect of a stagnant lid on plume stability. This model does not consider continents, contrary to Models 2-5. The surface yield stress of oceanic lithosphere is equal to 48 MPa and 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 (Crameri 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

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

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

2.2. Automated detection of mantle plumes

We developed a plume detection scheme to track plume positions and characteristics through time in map view. We located mantle plume conduits in the upper mantle (at 250, 350 and 670 km depth) and in the lower mantle (at 1,000 km depth) from their anomalously large radial heat advection, which is proportional to $v_z T$ (with v_z the radial velocity and T the temperature, Fig. S3b). Plumes were defined by upper and lower mantle heat advection equal to 190 K m yr^{-1} (Fig. S4) and 48 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 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.* Portnaya-
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° in Models 1-4 (Fig. 4a) and
230 16.5° in Model 5. In all models, only a limited proportion of plume conduits is
231 deflected by more than 10° ; tilt angles rarely reach the 60° 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° , compared to 28° 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⁻¹ 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 $\Delta T \approx 200 \text{ K}$, α is the thermal expansivity
280 and ρ_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⁻¹ 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 ρ_m the reference mantle density, α the reference thermal expansivity,
290 ΔT 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 $\sim 5 \times 10^3 \text{ kg s}^{-1}$

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

3.7. Heat flow

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

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

where c_p is the mantle heat capacity in $\text{J kg}^{-1}\text{K}^{-1}$. Individual plumes advect about 0.08-0.23 TW of heat on average in Models 1-5 (Fig. 3e). In Models 1-5, the core heat flow varies between 3 TW (Models 5) and 6.3 TW (Model 4, Fig. 3f), and never exceeds 20% of the surface heat flow (Fig. S5), which is at the lowest end of observational constraints (5-15 TW, Lay et al., 2008). The amount of heat carried by mantle plumes at 350 km depth represents ~ 3 TW in Models 1-4 and ~ 1.5 TW in Model 5, about 2 TW less than the corresponding heat flow (Fig. 3f). Davies (1988) and Sleep (1990) estimated the contribution of plumes to the total surface heat flow as 2.5 TW from hotspot swells. At 350 km depth, model mantle plumes carry a similar amount of heat. The ratio of upper-mantle plume heat flow to core heat flow (about 50-60% in Models 1-5) is consistent with numerical models of incompressible and isoviscous (Labrosse, 2002) or temperature-dependent viscosity convection with different internal heating rates (Mittelstaedt and 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 cm yr^{-1} . In contrast, the average absolute lateral
 336 speed of upper mantle plumes is approximately 1.8 cm yr^{-1} in Models 2
 337 and 3, 2.2 cm yr^{-1} in Model 4 (more vigorous lower mantle convection), and
 338 2.4 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 cm yr^{-1}
 342 (Model 1) and 2 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 cm yr^{-1} in Model 1, 1.6 cm yr^{-1} in Models 2 and 3, and
 351 2.2 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 $0.03\text{-}0.04^\circ \text{ Myr}^{-1}$ (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 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 $0.55^\circ \text{ Myr}^{-1}$, 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 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 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 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 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⁻¹ 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⁻¹. 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⁻¹) drift of
 460 the Hawaii plume between 81 and 47 Ma.

4.4. *Rapid drift due to plume merging events*

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

5. Discussion

5.1. *Limitations of models and analysis*

We tested a limited number of parameters potentially affecting plume drift. For example, we only considered one parameterisation (initial thickness, density and viscosity contrasts) to model basal thermochemical heterogeneities, although the nature of LLSVPs remains debated (*e.g.* Garnero et al., 2016). Moreover, although the resolution at 350 km depth in Models 2-4 (33 km) is lower than the one of Models 1 and 5 (24 km, similar 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 cm yr^{-1}
546 in Steinberger and O’Connell (1998) to 10 cm yr^{-1} in Steinberger and Antret-
547 ter (2006)) and with an assigned maximal tilt angle of 60° , 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 ~ 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 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 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 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

Parameter	Non-dim. value	Dim. value
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 (α_0)	1	$3 \times 10^{-5} \text{ K}^{-1}$
Reference density (ρ_0)	1	4400 kg m $^{-3}$
Reference diffusivity (κ_0)	1	$1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$
Reference heat capacity (c_{p0})	1	715 J kg $^{-1} \text{ K}^{-1}$
Reference viscosity (η_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 kJ mol $^{-1}$
Activation volume (V_a)	3	13.8 cm $^3 \text{ mol}^{-1}$
Maximum viscosity cut-off	10^4	10^{26} Pa s
Viscosity factor below 660 km depth	30	
Yield stress gradient for all materials ($d\sigma_Y$)	2.34×10^6	1088 Pa m $^{-1}$
Oceanic lithosphere surface yield stress ($\sigma_{Y_{0,oc}}$)	$2 - 200 \times 10^4$	27 – 270 MPa
Weak-crust surface yield stress (Model 5, $\sigma_{Y_{0,wc}}$)	1×10^4	13 MPa
Weak-crust thickness (Model 5)	0.005	14 km
Continental interior surface yield stress ($\sigma_{Y_{0,cont}}$)	7×10^5	932 MPa
Continental interior viscosity factor	100	
Continental interior buoyancy number (B_{cont})	–0.32	–150 kg m $^{-3}$
Continental interior thickness	0.0692	200 km
Continental belt surface yield stress ($\sigma_{Y_{belt}}$)	3×10^5	400 MPa
Continental belt viscosity factor	50	
Continental belt buoyancy number (B_{belt})	–0.4	–188 kg m $^{-3}$
Continental belt thickness	0.0432	125 km
Basal thermochemical layer buoyancy number (B_{llsvp})	0.25	117 kg m $^{-3}$ (2.7 %)
Basal thermochemical layer initial thickness	0.17	500 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 ($\mathbf{v}_{\text{rms}}$), mantle horizontal ($\mathbf{v}_{\text{m}}$), mantle net rotation (NR) and plume absolute ($\mathbf{v}_{\text{ha}}$) and relative ($\mathbf{v}_{\text{hr}}$) 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.

Parameter	Model 1	Model 2	Model 3	Model 4	Model 5	Dobrovine et al. (2012)	Tetley et al. (2019)
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 $\mathbf{v}_{\text{rms}}$ (cm/yr)	0.002	2.90 ± 0.85	3.32 ± 0.22	3.79 ± 0.60	5.04 ± 1.83	2 – 6	1.5 – 4.8
Mean UM $\mathbf{v}_{\text{m}}$ (cm/yr)	2.34 ± 0.82	3.19 ± 0.57	2.77 ± 0.32	3.60 ± 0.57	4.85 ± 1.52		
Mean LM $\mathbf{v}_{\text{m}}$ (cm/yr)	0.85 ± 0.25	0.91 ± 0.26	0.74 ± 0.15	1.01 ± 0.23	1.23 ± 0.35		
Average NR ($^{\circ}$ /Myr)	0.001 ± 0.0003	0.039 ± 0.014	0.031 ± 0.013	0.039 ± 0.016	0.029 ± 0.009	0.13 – 0.53	0.05 – 0.11
Mean UM $\mathbf{v}_{\text{ha}}$ (cm/yr)	0.57 ± 0.45	1.80 ± 1.13	1.74 ± 0.78	2.22 ± 0.82	2.40 ± 0.77	1.12 ± 0.31	2.07 ± 0.80
Mean UM $\mathbf{v}_{\text{hr}}$ (cm/yr)	0.63 ± 0.64	1.55 ± 1.58	1.61 ± 1.18	2.20 ± 1.28	2.22 ± 1.15	0.87 ± 0.35	1.46 ± 1.07
Mean LM $\mathbf{v}_{\text{ha}}$ (cm/yr)	0.39 ± 0.51	1.32 ± 0.56	1.05 ± 0.32	1.79 ± 0.90	2.04 ± 1.25		
Mean LM $\mathbf{v}_{\text{hr}}$ (cm/yr)	0.44 ± 0.68	1.02 ± 0.83	0.77 ± 0.47	1.66 ± 1.37	2.01 ± 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 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.