Geodynamic Investigation of a Cretaceous Superplume in the Pacific Ocean

Jing Xue

*Institute of Theoretical and Applied Geophysics, Peking University, Beijing, China
Virginia Polytechnic Institute and State University, Blacksburg, Virginia, USA*

Scott D. King

Virginia Polytechnic Institute and State University, Blacksburg, Virginia, USA

Abstract

The similarity in both age and geochemistry of the Ontong-Java, Hikurangi, and Manihiki plateaus suggests that they formed as a single superplateau from a unique mantle source. We investigate the necessity of a thermal superplume to form the Great Ontong-Java plateau at about 120 Ma using 3D spherical models of convection with imposed plate reconstruction models. The numerical simulations show that the giant plateau which formed as a result of melting due to the interaction of a plume head and the lithosphere would have been divided into smaller plateaus by spreading ridges, and end up at the present locations of Ontong-Java, Manihiki, and Hikurangi plateaus as well as a fragment in the western Caribbean. By comparing temperature and melt fraction between models with and without an initial thermal superplume, we propose that a Cretaceous superplume in Pacific at 120 Ma is required to form large igneous plateaus.

Email addresses: `jingx87@vt.edu` (Jing Xue), `sking07@vt.edu` (Scott D. King)

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1. Introduction

Large Igneous Provinces (LIPs) are regions on Earth where large accumulations of mafic rock were erupted, or emplaced at depth, over a geologically short time interval of a few million years (Coffin and Eldholm, 1994). LIPs are widely distributed, occurring both on continental plates, where they are often referred to as Continental Flood Basalts (CFB), and on oceanic plates where they are often referred to as oceanic plateaus (Coffin and Eldholm, 1994; Coffin and Gahagan, 1995). LIPs have been interpreted as initial outburst resulting from the interaction of a plume ‘head’ with the lithosphere followed by a continuing plume conduit, or ‘tail’, which produces an oceanic seamount chain (Richards et al., 1989; Campbell and Griffiths, 1990; Ballmer et al., 2015). For example, in the Atlantic and Indian oceans, the Deccan traps have been associated with the Réunion hotspot (Courtillot et al., 1986; Duncan and Richards, 1991; Roy, 2003), the North Atlantic Tertiary basalts have been associated with the Iceland hotspot (Saunders et al., 1997; Smallwood and White, 2002), the Rajmahal flood basalts have been associated with the Kerguelen hotspot (Mahoney et al., 1983; Coffin et al., 2002; Frey et al., 2002), and the Paraná (Southeast Brazil) and Etendeka (southwest Africa) flood basalts have been associated with the Walvis hotspot (Gibson et al., 2006; Adam et al., 2007). In each case, there is a geographical, geochronological and geochemical connection between the present location of active volcanism and the associated LIP (Duncan and Richards, 1991) with

23 an age-progressive volcanic chain connecting the location of active volcanism
24 and LIP.

25 In contrast, the connection between basaltic plateaus in the Pacific ocean
26 basin and long-lived hotspots is less clear (Mahoney et al., 1993; Neal et al.,
27 1997; Clouard and Bonneville, 2001; Ballmer et al., 2015). The Hawaii-
28 Emperor, Cobb, and Kodiak-Bowie volcanic chains terminate at subduc-
29 tion zones, leaving no trace of an oceanic plateau. The Louisville hotspot
30 has been suggested to be the source of the Ontong-Java (OJP), Hikurangi
31 (HIK) and Manihiki (MAN) plateaus (Mahoney and Spencer, 1991; Richards
32 et al., 1991). However, there is no clear geochemical evidence connecting
33 the Louisville hotspot and Ontong-Java plateau (Mahoney et al., 1993; Neal
34 et al., 1997; Vanderkluysen et al., 2014). Thus, the Louisville hotspot may
35 not have been involved in the formation of these plateaus or the hotspot
36 may have changed compositionally over time (Mahoney and Spencer, 1991;
37 Vanderkluysen et al., 2014). A meteorite impact (Ingle and Coffin, 2004;
38 Jones et al., 2002), edge-driven convection at craton boundaries (King and
39 Anderson, 1995, 1998; King and Ritsema, 2000; King, 2007), and melting
40 of fertile eclogitic material (Korenaga, 2005; Anderson, 2005; Foulger et al.,
41 2005), have also been proposed to explain the formation of LIPs.

42 The formation of the Pacific LIPs coincided with continental breakup,
43 development of new ocean basin and mid-ocean ridge spreading system, and
44 may indeed have caused the subsequent ocean opening (Bryan and Ernst,
45 2008). The similarity in both age and geochemistry of the OJP, HIK, and
46 MAN plateaus suggest that these formed as a single superplateau from a
47 unique mantle source (Hoernle et al., 2010; Timm et al., 2011). Subsequently,

48 this superplateau, referred to as the Great Ontong-Java plateau, broke up
49 into the present day OJP, HIK, and MAN plateaus as new divergent plate
50 boundaries formed (Taylor, 2006; Hoernle et al., 2010; Timm et al., 2011;
51 Chandler et al., 2012).

52 At the time of formation of the Great Ontong-Java superplateau (120
53 Ma), the supercontinent Gondwana also broke up and the ridge bounding
54 the Pacific plate reorganized (Nakanishi et al., 1992). The global rate of
55 oceanic crustal production increased by more than 50% between 120 Ma
56 and 100 Ma, which then decreased to present day rates over the period of
57 100 Ma-80 Ma (Tarduno et al., 1990; Larson, 1991b; Staudigel and King,
58 1992). This was dominated by crustal production in the Pacific ocean. An-
59 derson (1994) argued that the mid-Cretaceous pulse of Pacific oceanic crust
60 formation was due to plate reorganization and that plate-tectonic forces lo-
61 cated at lithospheric discontinuities controlled the timing and location of the
62 Pacific oceanic plateaus while Larson (1991b) attributed the plateau and in-
63 creased crustal production in the Pacific basin to a superplume rising from
64 the CMB. Korenaga (2005) pointed out that the plume head and bolide im-
65 pact models fail to explain the origin of the Great OJP event because he
66 argues that they form a subaerial, as opposed to submarine, plateau. Ko-
67 renaga (2005) and Lin and van Keken (2005) proposed a thermo-chemical
68 superplume/dome beneath the present South Pacific superswell in the lower
69 mantle.

70 In addition, OIB-type volcanism with mid-Cretaceous ages (about 120
71 Ma) have been reported in Costa Rica (Hauff et al., 2000; Hoernle et al.,
72 2004; Madrigal et al., 2015). The OIB type volcanism in Costa Rica has

73 been associated with Caribbean plateau; however the age is inconsistent with
74 the age of the Caribbean plateau. Because the basalts in Costa Rica have
75 a similar age with the Great Ontong-Java plateau, these Mid-Cretaceous
76 segments in Costa Rica may belong to the Great Ontong-Java plateau. The
77 reconstructed single Ontong Java-Manihiki-Hikurangi LIP would have had an
78 original size of nearly 80 Mkm³ (Bryan and Ernst, 2008), and there could be
79 as much as 30% melting of the initial peridotite source (Fitton and Godard,
80 2004).

81 At present day, seismic tomography reveals large wave speed anoma-
82 lies beneath Pacific ocean, known as Large Low Shear Velocity Provinces
83 (LLSVP) (Masters et al., 2000; Houser et al., 2008; Lekic et al., 2012), pro-
84 viding a possible deep source for the Cretaceous volcanism on the Pacific
85 plate. The Pacific LLSVP is as wide as 3000 km, and extends from the CMB
86 up to about 1000 km above the CMB (Masters et al., 2000; Romanowicz and
87 Gung, 2002; Ishii and Tromp, 2004; Houser et al., 2008; Ritsema et al., 2011;
88 Mosca et al., 2012). The seismic anomaly in the lowermost mantle has also
89 been interpreted as a cluster of plumes (Schubert et al., 2004; Bull et al.,
90 2009). Montelli et al. (2004, 2006) demonstrated that Pacific superplume
91 can be regarded as a cluster of five individual plumes in the lowermost 1000
92 km of the mantle. The debate over the thermal or thermal-chemical origin of
93 the LLSVPs is still going on (Cobden et al., 2012; Davies et al., 2012, 2015).
94 Karato and Karki (2001) and Brodholt et al. (2007) pointed out that the
95 ratio between *S*- and *P*-wave perturbation in the lowermost mantle derived
96 from tomography models exceeds the threshold for pure thermal anomalies.
97 The anti-correlation between bulk sound speed and shear wave speed pertur-

98 bations in tomography has been explained as compositional anomalies (De-
99 schamps and Trampert, 2003; Thorne and Garnero, 2004; Tan and Gurnis,
100 2005; Lassak et al., 2007; Bull et al., 2009).

101 The thermo-chemical origin of superplumes has also been investigated nu-
102 merically (McNamara and Zhong, 2004, 2005; Tan and Gurnis, 2005; Lassak
103 et al., 2007; Tan and Gurnis, 2007; Bull et al., 2009; Bower et al., 2013)
104 and experimentally (Davaille, 1999; Davaille et al., 2005). Davies et al.
105 (2012) argued that even the inclusion of chemical heterogeneities cannot pro-
106 duce a large-scale anti-correlation between bulk sound speed and shear wave
107 speed anomalies in the lowermost mantle, and the seismically observed anti-
108 correlation more likely to be caused by phase transitions.

109 In this study, we investigate the necessity of a thermal superplume to
110 form the Great Ontong-Java plateau at about 120 Ma. We apply a new
111 global tectonic reconstruction starting from 200 Ma (Seton et al., 2012) with
112 continuously closing plate polygons (Gurnis et al., 2012). The numerical
113 simulations show that the giant LIP generated by a superplume in southern
114 Pacific can be divided into segments by spreading ridges, and would end up
115 at the present locations of Ontong-Java, Manihiki, Hikurangi plateaus with
116 an additional piece at the western Caribbean. By comparing temperature
117 and melt fraction between models with and without an initial thermal su-
118 perplume, we propose that a Cretaceous superplume interacting with the
119 lithosphere in Pacific at 120 Ma is required to form large igneous plateaus.
120 The Ontong-Java, Manihiki and Hikurangi plateaus can be tracked back to
121 the same thermal source from the CMB.

122 2. Method

123 We solve the equations of thermal convection appropriate for Earth's
 124 mantle using the finite element code CitcomS-3.2.0 (Zhong et al., 2000; Tan
 125 et al., 2006) with $65 \times 65 \times 65$ elements in each of 12 spherical caps. The
 126 non-dimensional equations of conservation of mass, momentum and energy
 127 for incompressible thermal convection under the Boussinesq Approximation
 128 are:

$$129 \quad \nabla \cdot \vec{u} = 0 \quad (1)$$

$$130 \quad -\nabla P + \nabla \cdot \eta'(\nabla \vec{u} + \nabla \vec{u}^T) + RaT\vec{e}_r = 0 \quad (2)$$

$$\frac{\partial T}{\partial t} + \vec{u} \cdot \vec{\nabla} T = \nabla^2 T \quad (3)$$

131 where $\vec{u}$ is velocity, P is dynamic pressure, T is temperature, η' is viscosity,
 132 t is time, and $\vec{e}_r$ is the unit vector in the radial direction.

133 The Boussinesq convection problem is governed by the non-dimensional
 134 parameter called the Rayleigh Number,

$$Ra = \frac{\alpha \rho g \Delta T R_o^3}{\eta_o \kappa}, \quad (4)$$

135 where α is the thermal expansion coefficient, ρ is density, g is gravitational ac-
 136 celeration, ΔT is temperature difference between surface and the core-mantle
 137 boundary (CMB), R_o is radius of the Earth, η_o is the reference viscosity, and
 138 κ is thermal diffusivity. The parameters used in the model are given in
 139 Table 1.

140 The temperature-dependent viscosity η' is calculated using an Arrhenius
 141 form of the viscosity law (Hirth and Kohlstedt, 2003):

$$\eta' = \eta_0 \eta(r) \exp \left(\frac{E_\eta}{T^* + T_\eta} - \frac{E_\eta}{1 + T_\eta} \right) \quad (5)$$

142 where $T^* = \min(\max(T, 0), 1)$, $\eta(r)$ is a non-dimensional factor allowing
 143 for depth-dependent changes in viscosity, and η_0 is the reference viscosity in
 144 Table 1. The non-dimensional parameters E_η and T_η are normalized using

$$E_\eta = \frac{E^*}{R\Delta T}, \quad T_\eta = \frac{T_{surf}}{\Delta T} \quad (6)$$

145 where E^* , R , T_{surf} are the activation energy (Karato and Wu, 1993), univer-
 146 sal gas constant, and surface temperature (Table 1). We use a four-layered
 147 viscosity structure in which the $\eta(r)$ term in the Arrhenius law take on the
 148 following values $\eta_{lith} : \eta_{upper} : \eta_{tz} : \eta_{lower} = 1 : 0.03 : 1 : 30$, where the litho-
 149 sphere is defined to be from 0-90 km, the upper mantle is from 90-400 km,
 150 the transition zone is from 400-660 km and the lower mantle is from 660 km
 151 to the CMB. This depth-dependent viscosity profile is consistent with the
 152 long-wavelength geoid (King, 1995; Liu and Zhong, 2015) and global plate
 153 motions (Becker, 2006; Ghosh et al., 2010).

154 2.1. Boundary Conditions

155 We use a free-slip boundary condition at the core-mantle boundary with
 156 a temperature of $T = 2773$ K and a kinematic boundary condition at the
 157 surface with a temperature of $T = 273$ K. For the surface velocity bound-
 158 ary conditions, we use a time-dependent plate velocity model exported from
 159 GPlates (Gurnis et al., 2012) to constrain the surface movement of the calcu-
 160 lation to be consistent with geological observations. Even though imposing
 161 plate velocities imparts energy into the system, the advantages of using plate
 162 reconstructions for this problem are significant and outweigh the small error
 163 in the energy balance over the relatively short time period of these calcula-
 164 tions.

165 We compare two different plate reconstructions and the choice of these
 166 plate reconstructions deserves further discussion. Gurnis et al. (2012) created
 167 a continuously closed plate reconstruction model from 140 Ma to present. In
 168 the Gurnis et al. (2012) model, a moving hotspot reference frame (O'Neill
 169 et al., 2005) is used for 0-100 Ma and a fixed hotspot reference frame is
 170 used for 100-140 Ma (Müller et al., 1993). Seton et al. (2012) constructed
 171 a global plate reconstruction based on a true-polar wander corrected paleo-
 172 magnetic reference frame prior to 100 Ma (Steinberger and Torsvik, 2008),
 173 which extends the plate reconstruction back to 200 Ma, the time of Pangea
 174 break-up. The plate reconstruction models of Gurnis et al. (2012) and Seton
 175 et al. (2012) are nearly identical from 80 Ma to present (Fig. 1). The two
 176 models differ significantly between 120 Ma to 86 Ma where the Seton et al.
 177 (2012) model explicitly includes the breakup of the Phoenix plate (Fig. 1).
 178 In the Seton et al. (2012) model, the plateau breaks up into the Ontong-Java,
 179 Manihiki and Hikurangi plateaus between 120 Ma and 86 Ma in Pacific ocean
 180 (Taylor, 2006; Chandler et al., 2012). Ridges in the Phoenix plate cross-
 181 connect with Pacific spreading boundary, forming multiple ridge-ridge-ridge
 182 triple junctions at 120 Ma. In contrast, the Gurnis et al. (2012) model does
 183 not explicitly account for the plateau break-up. We will compare these plate
 184 reconstructions in greater detail in the next section.

185 *2.2. Initial Condition and Tracers*

186 We use an error function with a temperature of 1747 K in the interior of
 187 the mantle to which we add a single Gaussian perturbation in the lowermost
 188 mantle. This perturbation generates a massive ‘super plume’ that may lead
 189 to the formation of the Pacific plateaus as envisioned by Sheridan (1987),

190 Larson (1991b), Larson (1991a) and Condie et al. (2002). The initial lateral
 191 radius of the temperature anomaly is 1500 km, and the initial thickness of
 192 the temperature anomaly is 300 km. The geographic location of the initial
 193 anomaly is chosen so that the resulting plume head will reach the lithosphere
 194 at the location of the Pacific-Manihiki-Hikurangi triple junction in the Se-
 195 ton reconstruction (52°S , 128°W) (See Fig. 1). We vary the temperature
 196 perturbation from non-dimensional temperatures of 0.1 to 0.3, representing
 197 maximum dimensional temperature perturbations of 250° to 750° at the cen-
 198 ter of anomaly. We then consider a case where there is no initial thermal
 199 anomaly to test the hypothesis put forward by Anderson (1994) that the Pa-
 200 cific plateaus are the result of plate reorganization and do not require excess
 201 temperature from a plume head.

202 We introduce passive tracers in the initial temperature anomaly to track
 203 the motion of the ‘plume head’ material. The properties of tracers are the
 204 same as surrounding materials so that inclusion of these tracers does not
 205 affect the buoyancy. At the start of the model, the tracers are uniformly
 206 distributed throughout the initial temperature anomaly at the CMB. Once
 207 the Large Igneous Province (LIP) is created by the initial upwelling super-
 208 plume, we only keep those tracers that remain near the surface because we
 209 are interested in the migration of the LIPs as the plate motions evolve.

210 **3. Results**

211 In order to address the question of whether plate motion history can ex-
 212 plain the formation of the large Cretaceous plateaus in the Pacific without
 213 invoking a thermal anomaly, we consider 3D spherical geodynamic simula-

214 tions using the plate reconstruction models from Gurnis et al. (2012) and
 215 Seton et al. (2012). We consider cases with and without a starting ther-
 216 mal anomaly at the core-mantle boundary and we use tracers to track the
 217 movement of material initially at the core-mantle boundary. We compare
 218 the temperature patterns and rates of melt production along with dynamic
 219 topography through time to test these numerical experiments. We consider
 220 successful calculations to be ones that have no more than 30% partial melt
 221 of an initial peridotite source (Fitton and Godard, 2004) and ones where
 222 topographic uplift did not produce subaerial eruptions (Neal et al., 1997).
 223 Furthermore, we rule out models that produce substantial partial melt in
 224 other regions of the globe where LIPs are not observed both at the time of
 225 the LIP formation (120 Ma) and continuing to present day.

226 *3.1. Role of Plate Tectonics*

227 We first apply the Seton et al. (2012) plate reconstruction model as a
 228 boundary condition along with the initial condition with a 500 K temper-
 229 ature anomaly just above the CMB as described in Section 2. The initial
 230 temperature anomaly above the CMB is dynamically unstable and produces
 231 a buoyant upwelling. We adjust the starting time of the calculation and the
 232 initial placement of the anomaly above the core-mantle boundary so that the
 233 rising hot mantle reaches the surface at 120 Ma and the location of the initial
 234 anomaly coincides with the location of the Pacific-Manihiki-Hikurangi triple
 235 junction in the Seton et al. (2012) reconstruction (46°S, 123°W).

236 To understand the interaction between plumes and plate tectonics, we
 237 plot 3D-isothermal structures in dark red covered by translucent plates at
 238 surface (Fig. 2). Fig. 2(a) shows that the ‘head’ of the plume as it reaches

239 the lithosphere at about 120 Ma. The yellow tracers demarcate material
240 that comes from initial thermal anomaly in the lowermost mantle. We note
241 that when the resulting plume head interacts with the lithosphere, it is not
242 directly above the original temperature anomaly. The geographic location
243 of the starting anomaly is actually centered on (52°S, 128°W) and as the
244 hot superplume rises, it moves to southeast, swept along by the large-scale
245 plate-driven flow consistent with (Steinberger and O’Connell, 2000).

246 We theorize that the interaction between the plume head and the litho-
247 sphere will create a superplateau at the Pacific-Manihiki-Hikurangi triple
248 junction that will break up into smaller plateaus which move away from
249 each other as a result of the breakup of the Phoenix plate, which occurs
250 between 120 Ma and 86 Ma (Fig. 1). The plume head reaches the surface
251 at the Pacific-Manihiki-Hikurangi triple junction and the spreading ridges
252 break up and push the resulting plateaus away from each other. When the
253 Pacific-Manihiki, Pacific-Hikurangi ridge complex stops spreading at 86 Ma,
254 the smaller plates (i.e., Manihiki, Hikurangi) retain the same velocity as the
255 Pacific plate. At 90 Ma (Fig. 2 (b)) the original thermal anomaly begins
256 to die away and material from the original upwelling, indicated by tracers,
257 has cooled to nearly the same temperature as the surrounding mantle. In
258 our model there is no secondary pulse of widespread volcanism at 90 Ma
259 beneath these three plateaus, as suggested by (Hoernle et al., 2010). There
260 are several other upwellings (plumes) that are beginning to rise from the
261 CMB. One of these is located near the present day location of the St. Felix
262 and Juan Fernandez hotspots; another is beneath the present day location
263 of the Samoan hotspot; a third is northeast of the Marquesas; and the last

264 is midway between the present day location of the Tasminid and Kerguelen
265 hotspots in the present day Indian ocean (see maps in supplemental online
266 material).

267 The locations of tracers mark the area where the superplume first inter-
268 acts with the lithosphere and where it would be most likely for a LIP to
269 form. The locations of tracers at present day, shown in Fig. 2 (d), are con-
270 sistent with the current positions of the Ontong-Java, Manihiki, and Hiku-
271 rangi plateaus and, we also note the presences of a few tracers outside these
272 plateaus. For example, some tracers are seen near the west edge of Caribbean
273 plateau, and others distribute along the subduction zone between the Pacific
274 and Indo-Australian plates.

275 In order to address the effect of the plate reconstruction, we consider a
276 second calculation where we use the Gurnis et al. (2012) plate reconstruction
277 model. Because the Gurnis et al. (2012) plate reconstruction does not account
278 for the breakup of the Phoenix plate, we move the location of initial tem-
279 perature anomaly to (20.0°S, 81.86°W) so that the superplume head arrives
280 at the Pacific-Farallon-Phoenix triple junction in their model. Unlike the
281 previous calculations which used the Seton et al. (2012) plate reconstruction
282 model, the tracers in this new calculation divide into two clusters (Fig. 3);
283 one cluster gathers at the approximate location of the Manihiki plateau while
284 the other gathers at the approximate location of the Caribbean plateau. We
285 have considered other locations for the initial temperature anomaly with the
286 Gurnis et al. (2012) plate reconstruction, however we have not been able
287 to produce clusters of tracers at the Ontong-Java, Manihiki, and Hikurangi
288 plateaus with a single superplume as we were able to do with the Seton

et al. (2012) plate reconstruction. Based on these results, it appears that the break-up of Phoenix plate is a necessary condition to create the Ontong-Java, Manihiki, and Hikurangi plateaus from a single superplume event. Therefore for the rest of the calculations in this paper we will use the Seton et al. (2012) plate reconstruction.

3.2. *Temperature and Melt Fraction*

While the motion of tracers in the previous calculations show that Pacific LIPs can be tracked back to a deep source in the lowermost mantle, we investigate role of the initial temperature anomaly in the deep mantle by considering a series of calculations where we vary the Rayleigh number, start time, and initial temperature anomaly (Table 2). The start times for each calculation are chosen such that the initial plume head reaches the surface at 120 Ma, the formation time of the Ontong-Java, Manihiki, and Hikurangi plateaus. For calculations without a superplume (LR0 and HR0), we start the calculation at 199 Ma, which is the start of the Seton et al. (2012) plate reconstruction. These calculations allow us to test whether plate tectonic reconstructions alone are sufficient to form the Pacific LIPs as suggested by Anderson (1994).

Fig. 4 shows the present day horizontal-averaged temperature, viscosity, horizontal velocity and radial velocity for models in Table 2. The viscosity structure, which is a combination of the imposed depth-dependent viscosity and the temperature-dependent Arrhenius law, is similar for all the calculations. Because we use the Boussinesq approximation and do not add an adiabatic gradient to the temperature field, we also do not include the pressure-dependent activation volume and instead choose the depth-dependence of

314 viscosity so that it is consistent with geophysical observations (King, 1995).
 315 While we use a single activation energy for the entire mantle, the activation
 316 energies for creep mechanisms for silicate minerals do not vary widely. The
 317 upper mantle viscosity in the high Rayleigh number calculations is slightly
 318 lower than viscosity in the low Rayleigh number calculations because the
 319 temperature in the high Rayleigh number calculations is slightly greater than
 320 the low Rayleigh number calculations. Even though all calculations have the
 321 same starting temperature, in the low Rayleigh number calculations it takes
 322 longer for the plume to rise and there is more time for the upper mantle
 323 to cool. The deviation between the temperature profiles is small and has
 324 almost no impact on the resulting solutions. Similarly the lower mantle vis-
 325 cosity in the high Rayleigh number calculations is slightly greater than that
 326 for the low Rayleigh number calculations because the temperature in the
 327 high Rayleigh number calculations is slightly lower than the low Rayleigh
 328 number calculations. The most notable difference between the two groups of
 329 calculations is that the calculations with the higher Rayleigh number have
 330 larger horizontal-averaged horizontal and radial velocities in the upper man-
 331 tle and larger radial velocities throughout the lower mantle. This affects the
 332 rise time of the superplume and the degree to which the rising superplume
 333 (and subsequent secondary plumes) are affected by the large-scale plate flow.

334 Because we adjust the starting time of the calculations so that the plume
 335 head reaches the Pacific-Manihiki-Hikurangi triple junction at 120 Ma, and
 336 the plate history is the same for all eight calculations, we find that all of the
 337 calculations reproduce the present day locations of the Ontong-Java, Mani-
 338 hiki, and Hikurangi plateaus similar to the calculation shown in Fig. 2. To

339 assess whether the degree and locations of partial melt in these calculations
340 are consistent with observations we present melt fraction map (Fig. 5) from
341 the calculations at 90 km depth. We calculate the melt fraction resulting
342 from temperature and pressure in the geodynamic simulations using pMELTS
343 (Ghiorso et al., 2002) and we estimate the temperature dependence of the
344 liquid fraction formed by partial melting of KLB-1 peridotite (Hirschmann,
345 2000) at 3 GPa, which corresponds to a depth of 90 km. Anomalously high
346 temperatures at this depth produce pockets of high melt fraction (see tem-
347 perature maps in supplemental online material).

348 If the Ontong-Java, Manihiki, and Hikurangi plateaus can be tracked back
349 to a single Great Ontong-Java plateau, this requires the presence of a large
350 amount of melt in the southern Pacific ocean at about 120 Ma. Therefore
351 we present melt (Fig. 5) maps at 90 km depth from the eight calculations
352 described above at 120 Ma, when the Great Ontong-Java plateau formed.

353 The calculation LR0 and HR0 begin at 199 Ma, the start of the Seton
354 et al. (2012) plate reconstruction, while the other calculations begin some-
355 what later—at the time so that the superplume reaches the Pacific-Manihiki-
356 Hikurangi triple junction after 120 Ma. There is no melt present in calcula-
357 tion LR0. Either the superplume disrupts the plate scale flow in the lower
358 mantle, that would otherwise form the plumes in HR0 beneath the northeast
359 and southeast Pacific, or the missing plate history impacts their formations.
360 With the exception of the calculations without an initial thermal instabil-
361 ity (LR0 and HR0), the calculations produce up to 30% partial melt in the
362 southern Pacific ocean at 120 Ma (Fig. 5). The melt fraction maps indicate
363 that the difference in initial anomaly does not change the global melt pat-

tern, but it does affect melting in the south Pacific ocean at about 120 Ma
(see temperature maps in supplemental online material).

Continuing the calculations to present day, the present-day melt fraction maps for the four high Rayleigh number calculations (Fig. 6e-h) illustrate that large pockets of melt only occur in a few isolated locations and that those locations are consistent in all four calculations. There is a greater degree of melting in the low Rayleigh number calculations when compared with the high Rayleigh number calculations; however, the temperature difference between eastern Pacific and western Pacific at 90 km depth in the low Rayleigh number calculations is approximately 400 °C, a value that is too large to be consistent with the predicted wave speeds from surface wave tomography models (Zhou et al., 2006; French et al., 2013) (see temperature maps in online supplemental material). We specifically shade areas with large areas of melt (i.e., greater than a few percent) in the plots in order to focus on the regions where large melt volumes, consistent with LIP formation, are present. The largest locations of melt are near the equator along the western coast of Africa, 1,000 km east-southeast of Hawaii in the central Pacific, and in the southeastern Pacific, south of the Juan Fernandez hotspot, and beneath the southwestern Indian ocean between the Kerguelen and Bouvet hotspots. Even though we note relatively high temperatures beneath regions such as Iceland, those temperatures are not sufficient to generate significant regions of partial melt. With a small change in the melt calculation or geotherm, we could produce melt at these locations as well. However, then we would produce an even greater degree of melt in locations where there is no evidence supporting large pockets of melt. There is a greater difference

389 between the calculation with no superplume (HR0-(Fig. 6e)) and those that
 390 include a superplume (HR1-3-(Fig. 6f-h)) than between the different models
 391 that have a superplume (HR1-3-(Fig. 6f-h)). In particular the melt anomaly
 392 along the west African coast is located at 30 degrees south in HR0 as opposed
 393 to the equator in HR1-3 and there is a small melt anomaly between the Cape
 394 Verde and Canary hotspots in HR0 that doesn't show up in HR1-3. The pat-
 395 tern of melt bears no obvious resemblance to the present day distribution of
 396 intraplate volcanism. This does not support the hypothesis that intraplate
 397 volcanism can be explained by plate processes alone.

398 An additional constraint on the Pacific plateau formation is that the
 399 lavas erupted submarine, not subaerially. In order to address this constraint
 400 we consider the dynamic topography at 120 Ma from the eight calculations
 401 (LR0-3 and HR0-3) (Fig. 7). For the low Rayleigh number cases (LR0-3)
 402 the resulting dynamic topography is too large, not only in the region of the
 403 Pacific-Manihiki-Hikurangi triple junction but for all the ridges in the Pacific.
 404 For the high Rayleigh number calculations the resulting dynamic topography
 405 is more reasonable. These calculations do not take into account the reduction
 406 in viscosity due to the melt pockets or local flexure/isostatic effect; however
 407 the range of dynamic topography is encouraging.

408 *3.3. Location of Initial Instability*

409 It is reasonable to ask why a superplume should happen to form beneath
 410 the Pacific-Manihiki-Hikurangi triple junction. In an attempt to understand
 411 the stability of the results shown above, we move the initial thermal insta-
 412 bility at the CMB 500 km northwest, southeast, northeast and southwest
 413 (Fig. 8). The other parameters in these four calculations are identical to cal-

414 culation LR3 and we use the Seton et al. (2012) plate reconstruction model
 415 and the parameters in Table 1. The isothermal structures at present day and
 416 location of tracers in model P1, P2, P3 and P4 are very similar to each other,
 417 indicating that the resulting temperature and melt pattern is not strongly
 418 sensitive to the position of the initial anomaly forming the superplume within
 419 a 10 degree area at the core-mantle boundary.

420 3.4. Bottom Boundary Temperature

421 In order to investigate the effect of bottom boundary temperature to the
 422 activity of mantle convection, we compare calculations with different bottom
 423 boundary temperatures. We decrease the core-mantle boundary temperature
 424 from 2773 K to 2023 K and follow numerical modeling processes described
 425 above using the Seton et al. (2012) plate reconstruction model and the pa-
 426 rameters in Table 1.

427 Fig. 9 shows that the isothermal structures of models with different bot-
 428 tom boundary temperature are similar except for the case with the 2050 K
 429 CMB temperature where the total number of plumes is less than the other
 430 cases. In this case, the top of the outer core is only 300 K hotter than
 431 the average mantle temperature and there is only a small thermal boundary
 432 layer. Multiple thermal upwellings and ridge-like structures are seen in all
 433 calculations. The bottom boundary temperature does not significantly affect
 434 the activity of mantle convection. While studies have suggested that regions
 435 above core may be thermochemically distinct from the rest of the lower man-
 436 tle (Deschamps and Trampert, 2003; Thorne and Garnero, 2004; McNamara
 437 and Zhong, 2004; Tan and Gurnis, 2005; Bower et al., 2013), we have not
 438 include this complexity. Torsvik et al. (2006) show that the location of LIPs

439 and hotspots can be traced back to the edges of the proposed thermochem-
440 ical structures and this additional complexity may be required to achieve a
441 better correlation between the plumes in our models and hotspot locations.

442 **4. Conclusions**

443 We consider a series of 3D spherical calculations with imposed plate re-
444 constructions in order to investigate the formation of the Ontong-Java, Hiku-
445 rangi, and Manihiki plateaus. The calculations here consider two different
446 plate reconstructions and consider cases with and without an imposed insta-
447 bility to form a superplume. The two plate reconstructions differ significantly
448 between 120 Ma and 86 Ma where the Seton et al. (2012) model explicitly in-
449 cludes the breakup of the Great Ontong-Java plateau into the Ontong-Java,
450 Manihiki and Hikurangi plateaus, following Taylor (2006) and Chandler et al.
451 (2012). In contrast, the Gurnis et al. (2012) model does not explicitly account
452 for the plateau break-up. We find that the Seton et al. (2012) reconstruc-
453 tion is necessary to produce a pattern of mantle temperatures necessary to
454 explain the Ontong-Java, Hikurangi, and Manihiki plateaus from a single
455 superplateau.

456 The geochemistry of these plateaus suggests they formed from the same
457 mantle source (Hoernle et al., 2010; Timm et al., 2011). Further, we find that
458 plate reorganizations alone, at least the reconstructions by Seton et al. (2012)
459 and Gurnis et al. (2012) are not sufficient in and of themselves to produce a
460 temperature anomaly in the south Pacific that could explain the formation of
461 the superplateau. The calculations here are exploratory in nature and do not
462 consider the feedback between melt and viscosity nor do they consider the

463 possibility of thermo-chemical piles above the core mantle boundary. These
464 are logical steps for follow on calculations.

465 After the interaction of the superplume with the surface, our calculations
466 produce numerous other plumes. While not the goal of our study, it is inter-
467 esting to note that some of these additional plumes can be related to present
468 day location of hotspots; however many if not more of these plumes bear no
469 relationship to the distribution of intraplate volcanism. This demonstrates
470 that influences beyond the pattern of plate motions over the past 200 Myr are
471 responsible for the pattern of intraplate volcanism. Whether these are related
472 to thermo-chemical structure above the core mantle boundary, lithospheric
473 structure, our limited knowledge of the thermal structure of the mantle (as
474 our initial conditions), the role of water on melting or other factors we can
475 not speculate.

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Figure 1: Comparison of plate reconstruction models from 140 Ma to present day generated by Gurnis et al. (2012) and Seton et al. (2012), respectively. Red lines represent ridges and transforms while blue lines represent subduction zones. Plates in figures are: PAC-Pacific, FAR-Farallon, AUS-Australia, ANT-Antarctica, IZA-Izanagi, PHO-Phoenix, EGO-East Gondwana, HIK-Hikurangi, MAN-Manihiki, CHA-Chazca, CAT-Catequil, CAR-Caribbean, WANT-West Antarctica, EANT-East Antarctica, KUL-Kula, VAN-Vancouver, COC-Cocos, JUA-Juan De Fuca, NAZ-Nazca.

Figure 2: Isothermal structures at calculations times of:(a) 120 Ma,(b) 90 Ma, (c) 60 Ma, and (d) present day. The Initial temperature anomaly above CMB is 500 °C and the Seton et al. (2012) plate reconstruction model is used as the surface boundary condition. Green lines are plate boundaries. The present-day locations of the Ontong-Java, Manihiki, and Hikurangi plateaus are shown as yellow polygons.

Figure 3: Isothermal structures at (a) 120 Ma and (b) present day. The Initial temperature anomaly above CMB is 500 °C, and the Gurnis et al. (2012) plate reconstruction model is used as the surface boundary condition. Green lines are plate boundaries. The present-day locations of the Ontong-Java, Manihiki, and Hikurangi plateaus are shown as yellow polygons.

Figure 4: Horizontally averaged (a) temperature, (b) viscosity, (c) horizontal velocity and (d) radial velocity profiles for the calculations in Table 2.

Figure 5: Melt fraction maps at 90 km depth for calculations in Table 2 at 120 Ma. The black circles denote the present-day locations of hotspots taken from Sleep (1990).

Figure 6: Present-day melt fraction maps at 90 km depth for calculations in Table 2. The black circles denote the present-day locations of hotspots taken from Sleep (1990).

Figure 7: Dynamic topography for calculation with lower Rayleigh Number in Table 2 at 120 Ma.

Figure 8: Isothermal structures for calculations with different locations of initial thermal anomaly. (e) The location of initial anomalies in model P1, P2, P3 and P4. The position 0 represent (52S, 128W).

Figure 9: Isothermal structures of calculations with different bottom boundary temperature: (a) $T_{cmb} = 1.0$, (b) $T_{cmb} = 0.9$, (c) $T_{cmb} = 0.8$, (d) $T_{cmb} = 0.7$. The temperature dimensional number is $\Delta T = 2500^\circ$.

Table 1: Model parameters

Parameter	Symbol	Value	Unit
Radius of the Earth	R_0	6.371×10^6	m
Density of the mantle	ρ	4.0×10^3	kg/m^3
Thermal expansion coefficient	α	2.0×10^{-5}	K^{-1}
Gravitational acceleration	g	10	m/s^2
Temperature difference across the mantle	ΔT	2500	K
Thermal diffusivity	κ	10^{-6}	m^2/s
Reference viscosity	η_0	10^{21}	$Pa \cdot s$
Specific heat capacity	C_p	1200	$J/kg \cdot K$
Surface temperature	T_{surf}	300	K
Activation energy	E	300	kJ/mol
Universal gas constant	R	8.3144	$J/(K \cdot mol)$

Table 2: Model-specific parameters, δT_0 is the initial temperature anomaly in the lower-most mantle.

Model	Ra	$\eta_0(Pas)$	$\delta T_0(^{\circ}C)$	Start Age (Ma)
LR0	5×10^8	1×10^{21}	0	199
LR1	5×10^8	1×10^{21}	250	199
LR2	5×10^8	1×10^{21}	500	170
LR3	5×10^8	1×10^{21}	750	160
HR0	2.5×10^9	2×10^{20}	0	199
HR1	2.5×10^9	2×10^{20}	250	199
HR2	2.5×10^9	2×10^{20}	500	170
HR3	2.5×10^9	2×10^{20}	750	160