Geodynamic Investigation of a Cretaceous Superplume in the Pacific Ocean

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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 ~~giant~~ plume head ~~, rising in southern Pacific,~~ and the lithosphere would have been divided into ~~segments~~ smaller plateaus by spreading ridges, and end up at the present locations of Ontong-Java ~~Plateau~~, 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

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igneous plateaus.

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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; **?**). 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; **?**), the North Atlantic Tertiary basalts have been associated with the Iceland hotspot (**??**), the Rajmahal flood basalts have been associated with the Kerguelen hotspot (Mahoney et al., 1983; **?**; **?**), and the Paraná (Southeast Brazil) and Etendeka (southwest Africa) flood basalts have been associated with the Walvis hotspot (Gibson et al., 2006; **?**). In each case, there is a ~~strong~~ geographical, geochronological and geochemical connection between the present location of active volcanism and the associated LIP (Duncan and Richards, 1991) with an age-progressive volcanic chain connecting the location of active volcanism and LIP.

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

The formation of the Pacific LIPs coincided with continental breakup, development of new ocean basin and mid-ocean ridge spreading system, and may indeed have caused the subsequent ocean opening (Bryan and Ernst, 2008). The similarity in both age and geochemistry of the OJP, HIK, and MAN plateaus suggests that these formed as a single superplateau from a unique mantle source (Hoernle et al., 2010; Timm et al., 2011). Subsequently, this superplateau, referred to as the Great Ontong-Java Plateau, broke up into the present day OJP, HIK, and MAN plateaus as new divergent plate boundaries formed (Taylor, 2006; Hoernle et al., 2010; Timm et al., 2011; Chandler et al., 2012).

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

In addition, OIB-type volcanism with mid-Cretaceous ages (about 120 Ma) have been reported in Costa Rica (Hauff et al., 2000; **?**; **?**). The OIB type volcanism in Costa Rica has been associated with Caribbean plateau; however the age is inconsistent with the age of the Caribbean Plateau. Because the basalts in Costa Rica have a similar age with the Great Ontong-Java plateau, these Mid-Cretaceous segments in Costa Rica may belong to the Great Ontong-Java plateau. The reconstructed single Ontong Java-Manihiki-Hikurangi LIP would have had an original size of nearly 80 Mkm^3 (Bryan and Ernst, 2008), and there could be as much as 30% melting of the initial peridotite source (Fitton and Godard, 2004).

At present day, seismic tomography reveals large wave speed anomalies beneath Pacific Ocean, known as Large Low Shear Velocity Provinces (LLSVP) (Masters et al., 2000; Houser et al., 2008; Lekic et al., 2012), providing a possible deep source for the Cretaceous volcanism on the Pacific plate. The Pacific LLSVP is as wide as 3000 km, and extends from the CMB up to about 1000 km above the CMB (Masters et al., 2000; Romanowicz and Gung, 2002; Ishii and Tromp, 2004; Houser et al., 2008; Ritsema et al., 2011; **?**). The seismic anomaly in the lowermost mantle has also been interpreted as a cluster of plumes (Schubert et al., 2004; **?**). Montelli et al. (2004, 2006) demonstrated that Pacific superplume can be regarded as a cluster of five individual plumes in the lowermost 1000 km of the mantle. The debate over the thermal or thermal-chemical origin of the LLSVPs is still going on (Cobden et al., 2012; Davies et al., 2012, 2015). Karato and Karki (2001) and Brodholt et al. (2007) pointed out that the ratio between S- and P-wave perturbation in the lowermost mantle derived from tomography models exceeds the threshold for pure thermal anomalies. The anti-correlation between bulk sound speed and shear wave speed perturbations in tomography has been explained as compositional anomalies (Deschamps and Trampert, 2003; Thorne and Garnero, 2004; Tan and Gurnis, 2005; **?**; **?**).

The thermo-chemical origin of superplumes has also been investigated numerically (McNamara and Zhong, 2004; **?**; Tan and Gurnis, 2005; **?**; **?**; **?**; Bower et al., 2013) and experimentally (Davaille, 1999; Davaille et al., 2005). Davies et al. (2012) argued that even the inclusion of chemical heterogeneities cannot produce a large-scale anti-correlation between bulk sound speed and shear wave speed anomalies in the lowermost mantle, and the seismically observed anti-correlation more likely to be caused by phase transitions.

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

2. Method

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

$$\nabla \cdot \vec{u} = 0 \tag{1}$$

$$-\nabla P + \nabla \cdot \eta'(\nabla \vec{u} + \nabla \vec{u}^T) + RaT\vec{e_r} = 0 \tag{2}$$

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

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

The Bousinesq convection problem is governed by the non-dimensional parameter called the Rayleigh Number,

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

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

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

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

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

$$E_\eta = \frac{E^*}{R\Delta T}, \qquad T_\eta = \frac{T_{surf}}{\Delta T} \tag{6}$$

where E^*, R, T_{surf} are the activation energy (Karato and Wu, 1993), universal gas constant, and surface temperature (Table 1). We use a four-layered viscosity structure in which ~~the pre-exponential~~ the $\eta(r)$ term in the Arrhenius law take on the following values $\eta_{lith} : \eta_{upper} : \eta_{tz} : \eta_{lower} = 1 : 0.03 : 1 : 30$, where the lithosphere is defined to be from 0-90 km, the upper mantle is from 90-400 km, the transition zone is from 400-660 km and the lower mantle is from 660 km to the CMB. This depth-dependent viscosity profile is consistent with the long-wavelength geoid (King, 1995; **?**) and global plate motions (**??**).

2.1. Boundary Conditions

We use a free-slip boundary condition at the core-mantle boundary with a temperature of $T = 2773$ K and a kinematic boundary condition at the surface with a temperature of $T = 273$ K. For the surface velocity boundary conditions, we use a time-dependent plate velocity model exported from GPlates (Gurnis et al., 2012) to constrain the surface movement of the calculation to be consistent with geological observations. ~~The plate velocity and Rayleigh number are carefully balanced so that we do not impart energy into the convecting system from the imposed plate velocity boundary conditions (Davies, 1988).~~ Even though imposing plate velocities imparts energy into the system, the advantages of using plate reconstructions for this problem are significant and outweigh the small error in the energy balance over the relatively short time period of these calculations.

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

2.2. Initial Condition and Tracers

We use an error function with a temperature of 1747 K in the interior of the mantle to which we add a single Gaussian perturbation in the lowermost mantle. This perturbation generates a massive 'super plume' that may lead to the formation of the Pacific plateaus as envisioned by Sheridan (1987), Larson (1991b), Larson (1991a) and Condie et al. (2002). The initial lateral radius of the temperature anomaly is 1500 km, and the initial thickness of the temperature anomaly is 300 km. The geographic location of the initial anomaly is chosen so that the resulting plume head will reach the lithosphere at the location of the Pacific-Manihiki-Hikurangi triple junction in the Seton reconstruction (52°S, 128°W) (See Fig. 1). We vary the temperature perturbation from non-dimensional temperatures of 0.1 to 0.3, representing maximum dimensional temperature perturbations of 250° to 750° at the center of anomaly. We then consider a case where there is no initial thermal anomaly to test the hypothesis put forward by Anderson (1994) that the Pacific plateaus are the result of plate reorganization and do not require excess temperature from a plume head.

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

3. Results

In order to address the question of whether plate motion history can explain the formation of the large Cretaceous plateaus in the Pacific without invoking ~~an excess~~ a thermal anomaly, we consider 3D spherical geodynamic simulations using the plate reconstruction models from Gurnis et al. (2012) and Seton et al. (2012). We consider cases with and without a starting thermal anomaly at the core-mantle boundary and we use tracers to track the movement of material initially at the core-mantle boundary. We compare the temperature patterns and rates of melt production along with dymamic topography through time to test these numerical experiements. We consider successful calculations to be ones that have no more than 30% partial melt of an initial peridotite source (Fitton and Godard, 2004) and ones where topographic uplift did not produce subaerial eruptions (**?**). Furthermore, we rule out models that produce substantial partial melt in other regions of the globe where LIPs are not observed both at the time of the LIP formation (120 Ma) and continuing to present day.

3.1. Role of Plate Tectonics

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

To understand the interaction between plumes and plate tectonics, we plot 3D-isothermal structures in dark red covered by translucent plates at surface (Fig. 2). ~~The translucent colors at surface represent plate velocities.~~ Fig. 2(a) shows that the 'head' of the plume as it reaches the lithosphere at about 120 Ma. The yellow tracers demarcate material that comes from initial thermal anomaly in the lowermost mantle. We note that when the resulting plume head interacts with the lithosphere, it is not directly above the original temperature anomaly. The geographic location of the starting anomaly is actually centered on (52°S, 128°W) and as the hot superplume rises, it moves to southeast, swept along by the large-scale plate-driven flow consistent with (?).

~~The resulting plume head material breaks into smaller segments~~ We theorize that the interaction between the plume head and the lithosphere will create a superplateau at the Pacific-Manihiki-Hikurangi triple junction that will break up ~~(plateaus)~~ into smaller plateaus which move away from each other as a result of ~~due to~~ the breakup of the Phoenix plate, which occurs between 120 Ma and 86 Ma (Fig. 1). The plume head reaches the surface at the Pacific-Manihiki-Hikurangi triple junction and the spreading ridges break up and push the resulting ~~pieces of the plume material~~ plateaus away from each other. When the Pacific-Manihiki, Pacific-Hikurangi ridge complex stops spreading at 86 Ma, the smaller plates (i.e., Manihiki, Hikurangi) retain the same velocity as the Pacific plate. At 90 Ma (Fig. 2 (b)) the original thermal anomaly begins to die away and material from the original upwelling, indicated by tracers, has cooled to nearly the same temperature as the surrounding mantle. In our model there is no secondary pulse of widespread volcanism at 90 Ma beneath these three plateaus, as suggested by (?). There are several other upwellings (plumes) that are beginning to rise from the CMB. One of these is located near the present day location of the St. Felix and Juan Fernandez hotspots; another is beneath the present day location of the Samoan hotspot; a third is northeast of the Marquesas; and the last is midway between the present day location of the Tasminid and Kerguelen hotspots in the present day Indian ocean (see maps in supplemental online material).

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There are few additional constraints that we can apply to test our assumption that a superplume arises beneath the Pacific-Manihiki-Hikurangi triple junction forming a superplateau that subsequently breaks into the Ontong-Java, Manihiki, and Hikurangi plateaus. First, as the calculations continue to present day they should not produce large-scale melting in places where there is no evidence for melt. Second, the mantle should have regions of present-day melt under the same conditions we used at 120 Myrs. At 60 Ma (Fig. 2 (c)) the plumes that were in the lower mantle at 90 Ma have now reached the surface with conduits (plume tails) that are still clearly visible extending at least from 900 km to the core mantle boundary. By the time these anomalies reach the upper mantle it is difficult to distinguish them from the anomalies related to plate scale flow. We also observe new instabilities forming: one 500-1,000 km north of Hawaii; another 500-1,000 km south of Cape Verde; a third in the equatorial western Pacific; and a fouth 1,000-

2,000 km east of the present day location of the Reunion hotspot (see maps in supplemental online material).

At present day (Fig. 2 (d)), the plumes that we described forming at 90 and 60 Ma are still present throughout the lower mantle. In addition there are new instabilities northeast of Nova Scotia and beneath the surface expression of the Balleny hotspot although neither of these reaches the surface. Several of the plumes in this model end up near the present day location of hotspots– notably, Hawaii, Pitcairn, Louisville, Cape Verde and St. Helena. However, an equal or greater number of plumes in this calculation have little or no correlation with present day location of hotspots. This is not to suggest that these specific hotspots are the result of a deep mantle plume, however, starting from an isothermal mantle and a single starting anomaly centered on ($52°$ *S,* $128°$ *W), the present day mantle produces small pockets of melt. It is possible that the return flow from plate motions pushes anomalies at the core mantle boundary to these locations, however this requires further investigation.*

The locations of tracers mark the ~~region~~ area where the superplume ~~material~~ first interacts with the lithosphere and where it would be most likely for a LIP to form. The locations of tracers at present day, shown in Fig. 2 (d), are consistent with the current positions of the Ontong-Java, Manihiki, and Hikurangi plateaus and, we also note the presences of a few tracers outside these plateaus. For example, some tracers are seen near the west edge of Caribbean plateau, and others distribute along the subduction zone between the Pacific and Indo-Australian plates.

In order to address the effect of the plate reconstruction, we consider a second calculation where we use the Gurnis et al. (2012) plate reconstruction model. Because the Gurnis et al. (2012) plate reconstruction does not account for the breakup of the Phoenix plate, we move the location of initial temperature anomaly to (20.0°S, 81.86°W) so that the superplume head arrives at the Pacific-Farallon-Phoenix triple junction in their model. Unlike the previous calculations which used the Seton et al. (2012) plate reconstruction model, the tracers in this new calculation divide into two clusters (Fig. 3); one cluster gathers at the approximate location of the Manihiki plateau while the other gathers at the approximate location of the Caribbean plateau. We have considered other locations for the initial temperature anomaly with the Gurnis et al. (2012) plate reconstruction, however we have not been able to produce clusters of tracers at the Ontong-Java, Manihiki, and Hikurangi 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 ~~model~~ 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 viscosity so that it is consistent with geophysical observations (King, 1995). While we use a single activation energy for the entire mantle, the activation energies for creep mechanisms for silicate minerals do not vary widely. The upper mantle viscosity in the high Rayleigh number calculations is slightly lower than viscosity in the low Rayleigh number calculations because the temperature in the high Rayleigh number calculations is slightly greater than the low Rayleigh number calculations. Even though all calculations have the same starting temperature, in the low Rayleigh number calculations it takes longer for the plume to rise and there is more time for the upper mantle to cool. The deviation between the temperature profiles is small and has almost no impact on the resulting solutions. Similarly the lower mantle viscosity in the high Rayleigh number calculations is slightly greater than that for the low Rayleigh number calculations because the temperature in the high Rayleigh number calculations is slightly lower than the low Rayleigh number calculations. The most notable difference between the two groups of calculations is that the calculations with the higher Rayleigh number have larger horizontal-averaged horizontal and radial velocities in the upper mantle and larger radial velocities throughout the lower mantle. This affects the rise time of the superplume and the degree to which the rising superplume (and subsequent secondary plumes) are affected by the large-scale plate flow.

Because we adjust the starting time of the calculations so that the plume head reaches the Pacific-Manihiki-Hikurangi triple junction at 120 Ma, and the plate history is the same for all eight calculations, we find that all of the calculations reproduce the present day locations of the Ontong-Java, Manihiki, and Hikurangi plateaus as was the case for the calculation shown in Fig. 2. To assess whether the degree and locations of ~~location of regions of~~ partial melt in these calculations are consistent with observations we ~~present present-day temperature (Fig. 5) and~~ melt fraction (Fig. 6) maps from the calculations at 90 km depth. We calculate the melt fraction resulting from temperature and pressure in the geodynamic simulations using pMELTS (Ghiorso et al., 2002) and we estimate the temperature dependence of the liquid fraction formed by partial melting of KLB-1 peridotite (Hirschmann, 2000) at 3 GPa, which corresponds to a depth of 90 km. Because we do not explicitly include the effect of melting in our calculation, we account for the latent heat of melting by reducing the temperature above the melting point from our calculations by a factor

$$(T_{effective} - T_m) = (T_{code} - T_m)\left(1 + \frac{L}{c_p * 0.005}\right) \tag{7}$$

where $T_{effective}$ is the effective temperature above the solidus, T_m is the temperature of the solidus, T_{code} is the actual temperature from the code (which does not include latent heating), L is the latent heat of melting (600 kJ/kg)(**?**), and c_p is the specific heat at constant pressure (1200 J/kg-K). The value 0.005 is the fraction of melting per degree(Hirschmann, 2000). Anomalously high temperatures at this depth produce pockets of high melt fraction and we compare these with regions of present day volcanism. ~~We do not include the latent heat of melting or the impact of melt on viscosity in these calculations. Because we are looking at global flow patterns and small amounts and regions of melting should have no impact on our results.~~

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The patterns of high and low temperature are quite similar in both the four low Rayleigh number calculations (Fig. 5a-d) and the four high Rayleigh number calculations (Fig. 5e-h). For comparison the present day locations of hotspots are shown as black circles on the maps using the locations in Sleep (1990). Subduction zones are the most prominent low temperature features in all eight calculations and the system of mid-ocean ridges is the most prominent high-temperature features in the low Rayleigh number calculations. While mid-ocean ridges are present in the high Rayleigh number calculations, there is a smaller temperature contrast beneath the ocean basins and the ridges do not stand out as prominently in these plots as they do in the low Rayleigh number calculations. In the low Rayleigh number calculations, the temperature difference between eastern Pacific and western Pacific at 90 km depth is approximately 400 °C, ~~which~~ 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). Therefore we do not discuss these cases further.

In the high Rayleigh number calculations, the difference between the temperatures in the eastern and western Pacific at 90 km depth is on the order of 100-200 K, which is consistent with the predicted wave speeds from surface wave tomography models (Zhou et al., 2006; French et al., 2013) (Fig. 5e-h).

In all of the high Rayleigh number calculations there are high temperature anomalies near Iceland; in the eastern Atlantic between the Azores and Canary islands; in the south Atlantic near the Walvis ridge; at the Reunion hotspot; and approximately 1,000 km southeast of Hawaii in the central Pacific. We also observe anomalously hot linear features in the central Pacific extending northwest from the East Pacific rise which coincide with the anomalous slow fingers seen in the surface wave tomography of French et al. (2013), and there is a small region of anomalously high-temperature beneath central Asia near the Baikal rift in all four high Rayleigh number calculations. There is no anomalously high temperature beneath the Afar region. The similarity of the temperature patterns for all four of the high Rayleigh number calculations demonstrates that the presence (HR1-3) or absence (HR0) of the superplume at 120 Ma has a minimal impact on the present day upper mantle thermal structure. However there are some differences between the calculations, in calculations HR2 and HR3, we observe a high-temperature anomaly in the southeastern Pacific, beneath the present day Nazca plate, far from the present day location of the superswell. We note that these high-temperature anomalies are consistent with slow seismic anomalies beneath the Nazca plate in tomographic models (Zhou et al., 2006; Ritsema et al., 2011; French et al., 2013). This feature is not present in calculations HR0 or HR1. Finally, imposing plate motions over the last 200 Myr reproduces the loca-

tion of subducted slabs, as has been previously shown (Lithgow-Bertelloni and Richards, 1995; Ricard et al., 1996). However, the present day distribution of hotspots cannot be explained solely by the past 200 Myrs of plate motion.

If the Ontong-Java, Manihiki, and Hikurangi plateaus can be tracked back to a single Great Ontong-Java plateau, this requires the presence of a large amount of melt in the southern Pacific Ocean at about 120 Ma. Therefore we present temperature (Fig. 7) and melt (Fig. 8) maps at 90 km depth from the eight calculations described above at 120 Ma, when the Great Ontong-Java plateau formed. In the lower Rayleigh number calculations, the larger temperature anomalies in the initial condition produce hotter plume heads that spread along the Pacific-Phoenix plate boundary at 120 Ma. In calculation LR0, which does not have an initial temperature anomaly at the CMB, there is no unusually hot anomaly in Pacific Ocean basin and the temperature pattern reflects the pattern of sea floor spreading. In contrast, the model HR0, in which the mantle is more active, has hot anomalies in both the northeast and southeast Pacific. The hot anomaly in the northeast Pacific would eventually be subducted beneath, or obducted onto, the North American continent. The calculation HR0 begins at 199 Ma, the start of the Seton et al. (2012) plate reconstruction, while the other calculations begin somewhat later–at the time so that the superplume reaches the Pacific-Manihiki-Hikurangi triple junction at 120 Ma. Either the superplume disrupts the plate scale flow in the lower mantle, that would otherwise form the plumes in HR0 beneath the northeast and southeast Pacific, or the missing plate history impacts their formation. With the exception of the calculations without an initial thermal instability (LR0 and HR0), the calculations produce up to 30% partial melt in the Southern Pacific Ocean at 120 Ma (Fig. 8). The temperature and melt fraction maps indicate that the difference in initial anomaly does not change the global temperature or melt pattern, but it does affect melting in the south Pacific Ocean at about 120 Ma.

An additional constraint on the Pacific plateau formation is that the lavas erupted submarine, not subaerially. In order to address this constraint we consider the dynamic topography at 120 Ma from the eight calculations (LR0-3 and HR0-3) (Fig. 9). For the low Rayleigh number cases (LR0-3) the resulting dynamic topography is too large, not only in the region of the Pacific-Manihiki-Hikurangi triple junction but for all the ridges in the Pacific. For the high Rayleigh number calculations the resulting dynamic topography is more reasonable. These calculations do not take into account the reduction in viscosity due to the melt pockets or local flexure/isostatic effect; however the range of dynamic topography is encouraging.

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

3.3. Location of Initial Instability

It is reasonable to ask why a superplume should happen to form beneath the Pacific-Manihiki-Hikurangi triple junction. In an attempt to understand the stability of the results shown above, we move the initial thermal instability at the CMB 500 km northwest, southeast, northeast and southwest (Fig. 10). The other parameters in these four calculations are identical to calculation LR3 and we use the Seton et al. (2012) plate reconstruction model and the parameters in Table 1. The isothermal structures at present day and location of tracers in model P1, P2, P3 and P4 are very similar to each other, indicating that the resulting temperature and melt pattern is not strongly sensitive to the position of the initial anomaly forming the superplume within a 10 degree area at the core-mantle boundary.

3.4. Bottom Boundary Temperature

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

Fig. 11 shows that the isothermal structures of models with different bottom boundary temperature are similar except for the case with the 2050 K CMB temperature where the total number of plumes is less than the other cases. In this case, the top of the outer core is only 300 K hotter than the average mantle temperature and there is only a small thermal boundary layer. Multiple thermal upwellings and ridge-like structures are seen in all calculations. The bottom boundary temperature does not significantly affect the activity of mantle convection. While studies have suggested that regions above core may be thermochemically distinct from the rest of the lower mantle (Deschamps and Trampert, 2003; Thorne and Garnero, 2004; McNamara and Zhong, 2004; Tan and Gurnis, 2005; Bower et al., 2013), we have not include this complexity. Torsvik et al. (2006) show that the location of LIPs and hotspots can be traced back to the edges of the proposed thermochemical structures and this additional complexity may be required to achieve a better correlation between the plumes in our models and hotspot locations.

4. Conclusions

We consider a series of 3D spherical calculations with imposed plate reconstructions in order to investigate the formation of the Ontong-Java, Hikurangi, and Manihiki plateaus. The calculations here consider two different plate reconstructions and consider cases with and without an imposed instability to form a superplume. The two plate reconstructions ~~models~~ differ significantly between 120 Ma and 86 Ma where the Seton et al. (2012) model explicitly includes the breakup of the Great Ontong-Java plateau into the Ontong-Java, Manihiki and Hikurangi plateaus, following Taylor (2006) and Chandler et al. (2012). In contrast, the Gurnis et al. (2012) model does not explicitly account for the plateau break-up. We find that the Seton et al. (2012) reconstruction is necessary to produce a pattern of mantle temperatures necessary to explain the Ontong-Java, Hikurangi, and Manihiki plateaus from a single superplateau.

The geochemistry of these plateaus suggests they formed from the same mantle source (Hoernle et al., 2010; Timm et al., 2011). Further, we find that plate reorganizations alone, at least the reconstructions by Seton et al. (2012) and Gurnis et al. (2012) are not sufficient in and of themselves to produce a temperature anomaly in the south Pacific that could explain the formation of the superplateau. The calculations here are exploratory in nature and do not consider the feedback between melt and viscosity nor do they consider the possibility of thermo-chemical piles above the core mantle boundary. These are logical steps for follow on calculations.

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

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References

Anderson, D. L., 1994. Superplumes or supercontinents. Geology 22 (1), 39–42.

Bower, D. J., Gurnis, M., Seton, M., 2013. Lower mantle structure from paleogeographically constrained dynamic Earth models. Geochemistry Geophysics Geosystems 14 (1), 44–63.

Brodholt, J. P., Helffrich, G., Trampert, J., 2007. Chemical versus thermal heterogeneity in the lower mantle: The most likely role of anelasticity. Earth and Planetary Science Letters 262 (3), 429–437.

Bryan, S. E., Ernst, R. E., 2008. Revised definition of Large Igneous Provinces (LIPs). Earth-Science Reviews 86 (1-4), 175–202.

Campbell, I. H., Griffiths, R. W., 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth and Planetary Science Letters 99 (1-2), 79–93.

Chandler, M. T., Wessel, P., Taylor, B., Seton, M., Kim, S. S., Hyeong, K., 2012. Reconstructing Ontong Java Nui: Implications for Pacific absolute plate motion, hotspot drift and true polar wander. Earth and Planetary Science Letters 331, 140–151.

Clouard, V., Bonneville, A., 2001. How many Pacific hotspots are fed by deep-mantle plumes? Geology 29 (8), 695–698.

Cobden, L., Mosca, I., Trampert, J., Ritsema, J., 2012. On the likelihood of post-perovskite near the core–mantle boundary: A statistical interpretation of seismic observations. Physics of the Earth and Planetary Interiors 210, 21–35.

Coffin, M. F., Eldholm, O., 1994. Large Igneous Provinces - Crustal Structure, Dimensions, and External Consequences. Reviews of Geophysics 32 (1), 1–36.

Coffin, M. F., Gahagan, L. M., 1995. Ontong-Java and Kerguelen Plateaus - Cretaceous Iceland. Journal of the Geological Society 152, 1047–1052.

Condie, K. C., Abbott, D., Des Marais, D. J., 2002. Preface. Journal of Geodynamics 34 (2), 159–162.

Courtillot, V., Besse, J., Vandamme, D., Montigny, R., Jaeger, J.-J., Cappetta, H., 1986. Deccan flood basalts at the Cretaceous/Tertiary boundary? Earth and Planetary Science Letters 80 (3), 361–374.

Davaille, A., 1999. Simultaneous generation of hotspots and superswells by convection in a heterogenous planetary mantle. Nature 402 (6763), 756–760.

Davaille, A., Stutzmann, E., Silveira, G., Besse, J., Courtillot, V., 2005. Convective patterns under the Indo-Atlantic << box >>. Earth and Planetary Science Letters 239 (3-4), 233–252.

Davies, D. R., Goes, S., Davies, J. H., Schuberth, B. S. A., Bunge, H. P., Ritsema, J., 2012. Reconciling dynamic and seismic models of Earth's lower mantle: The dominant role of thermal heterogeneity. Earth and Planetary Science Letters 353, 253–269.

Davies, D. R., Goes, S., Lau, H. C. P., 2015. Thermally dominated deep mantle LLSVPs: a review. In: The Earth's Heterogeneous Mantle. Springer, pp. 441–477.

Davies, G. F., 1988. Role of the lithosphere in mantle convection. Journal of Geophysical Research-Solid Earth and Planets 93 (B9), 10451–10466.

Deschamps, F., Trampert, J., 2003. Mantle tomography and its relation to temperature and composition. Physics of the Earth and Planetary Interiors 140, 277–291.

Duncan, R. A., Richards, M. A., 1991. Hotspots, mantle plumes, flood basalts, and true polar wander. Reviews of Geophysics 29 (1), 31–50.

Fitton, J. G., Godard, M., 2004. Origin and evolution of magmas on the Ontong Java Plateau. Geological Society, London, Special Publications 229 (1), 151–178.

French, S., Lekic, V., Romanowicz, B., 2013. Waveform tomography reveals channeled flow at the base of the oceanic asthenosphere. Science 342 (6155), 227–230.

Ghiorso, M. S., Hirschmann, M. M., Reiners, P. W., Kress, V. C., 2002. The pMELTS: A revision of MELTS for improved calculation of phase relations and major element partitioning related to partial melting of the mantle to 3 GPa. Geochemistry Geophysics Geosystems 3.

Gibson, S. A., Thompson, R. N., Day, J. A., 2006. Timescales and mechanisms of plumelithosphere interactions: 40Ar/39Ar geochronology and geochemistry of alkaline igneous rocks from the ParanEtendeka large igneous province . Earth and Planetary Science Letters 251 (12), 1 – 17.

Gurnis, M., Turner, M., Zahirovic, S., DiCaprio, L., Spasojevic, S., Muller, R. D., Boyden, J., Seton, M., Manea, V. C., Bower, D. J., 2012. Plate tectonic reconstructions with continuously closing plates. Computers & Geosciences 38 (1), 35–42.

Hauff, F., Hoernle, K., van den Bogaard, P., Alvarado, G., Garbe-Schnberg, D., 2000. Age and geochemistry of basaltic complexes in western Costa Rica: Contributions to the geotectonic evolution of Central America. Geochemistry, Geophysics, Geosystems 1 (5).

Hirschmann, M. M., 2000. Mantle solidus: Experimental constraints and the effects of peridotite composition. Geochemistry Geophysics Geosystems 1.

Hirth, G., Kohlstedt, D., 2003. Rheology of the upper mantle and the mantle wedge: A view from the experimentalists. Inside the subduction Factory, 83–105.

Hoernle, K., Hauff, F., van den Bogaard, P., 2004. 70 m.y. history (139-69 Ma) for the Caribbean large igneous province. Geology 32 (8), 697–700.

Hoernle, K., Hauff, F., van den Bogaard, P., Werner, R., Mortimer, N., Geldmacher, J., Garbe-Schonberg, D., Davy, B., 2010. Age and geochemistry of volcanic rocks from the Hikurangi and Manihiki oceanic Plateaus. Geochimica Et Cosmochimica Acta 74 (24), 7196–7219.

Houser, C., Masters, G., Shearer, P., Laske, G., 2008. Shear and compressional velocity models of the mantle from cluster analysis of long-period waveforms. Geophysical Journal International 174 (1), 195–212.

Ingle, S., Coffin, M. F., 2004. Impact origin for the greater Ontong Java Plateau? Earth and Planetary Science Letters 218 (1-2), 123–134.

Ishii, M., Tromp, J., 2004. Constraining large-scale mantle heterogeneity using mantle and inner-core sensitive normal modes. Physics of the Earth and Planetary Interiors 146 (1-2), 113–124.

Jones, A. P., Price, G. D., Price, N. J., DeCarli, P. S., Clegg, R. A., 2002. Impact induced melting and the development of large igneous provinces. Earth and Planetary Science Letters 202 (3-4), 551–561.

Karato, S., Karki, B. B., 2001. Origin of lateral variation of seismic wave velocities and density in the deep mantle. Journal of Geophysical Research-Solid Earth 106 (B10), 21771–21783.

Karato, S., Wu, P., 1993. Rheology of the upper mantle - a synthesis. Science 260 (5109), 771–778.

King, S. D., 1995. The viscosity structure of the mantle. Reviews of Geophysics 33 (S1), 11–17.

King, S. D., 2007. Hotspots and edge-driven convection. Geology 35 (3), 223–226.

King, S. D., Anderson, D. L., 1995. An alternative mechanism of flood basalt formation. Earth and Planetary Science Letters 136 (3-4), 269–279.

King, S. D., Anderson, D. L., 1998. Edge-driven convection. Earth and Planetary Science Letters 160 (34), 289 – 296.

King, S. D., Ritsema, J., 2000. African hot spot volcanism: Small-scale convection in the upper mantle beneath cratons. Science 290 (5494), 1137–1140.

Korenaga, J., 2005. Why did not the Ontong Java Plateau form subaerially? Earth and Planetary Science Letters 234 (3-4), 385–399.

Larson, R. L., 1991a. Geological consequences of superplumes. Geology 19 (10), 963–966.

Larson, R. L., 1991b. Latest pulse of Earth - Evidence for a mid-Cretaceous superplume. Geology 19 (6), 547–550.

Lekic, V., Cottaar, S., Dziewonski, A., Romanowicz, B., 2012. Cluster analysis of global lower mantle tomography: A new class of structure and implications for chemical heterogeneity. Earth and Planetary Science Letters 357, 68–77.

Lin, S.-C., van Keken, P. E., 2005. Multiple volcanic episodes of flood basalts caused by thermochemical mantle plumes. Nature 436 (7048), 250–252.

Lithgow-Bertelloni, C., Richards, M. A., 1995. Cenozoic plate driving forces. Geophysical Research Letters 22 (11), 1317–1320.

Mahoney, J., Macdougall, J., Lugmair, G., Gopalan, K., 1983. Kerguelen hotspot source for Rajmahal traps and Ninetyeast Ridge. Nature 303, 385–389.

Mahoney, J. J., Spencer, K. J., 1991. Isotopic evidence for the origin of the Manihiki and Ontong Java oceanic plateaus. Earth and Planetary Science Letters 104 (2-4), 196–210.

Masters, G., Laske, G., Bolton, H., Dziewonski, A., 2000. The Relative Behavior of Shear Velocity, Bulk Sound Speed, and Compressional Velocity in the Mantle: Implications for Chemical and Thermal Structure. American Geophysical Union, pp. 63–87.

McNamara, A. K., Zhong, S. J., 2004. Thermochemical structures within a spherical mantle: Superplumes or piles? Journal of Geophysical Research-Solid Earth 109 (B7).

Montelli, R., Nolet, G., Dahlen, F. A., Masters, G., 2006. A catalogue of deep mantle plumes: New results from finite-frequency tomography. Geochemistry Geophysics Geosystems 7.

Montelli, R., Nolet, G., Dahlen, F. A., Masters, G., Engdahl, E. R., Hung, S. H., 2004. Finite-frequency tomography reveals a variety of plumes in the mantle. Science 303 (5656), 338–343.

Müller, R. D., Royer, J. Y., Lawver, L. A., 1993. Revised plate motions relative to the hotspots from combined Atlantic and Indian-Ocean hotspot tracks. Geology 21 (3), 275–278.

Nakanishi, M., Tamaki, K., Kobayashi, K., 1992. Magnetic Anomaly Lineations from Late Jurassic to Early Cretaceous in the West-Central Pacific-Ocean. Geophysical Journal International 109 (3), 701–719.

Neal, C., Mahoney, J., Kroenke, L., Duncan, R., Petterson, M., 1997. The Ontong Java Plateau. Vol. 100. American Geophysical Union, pp. 183–216.

O'Neill, C., Müller, D., Steinberger, B., 2005. On the uncertainties in hot spot reconstructions and the significance of moving hot spot reference frames. Geochemistry Geophysics Geosystems 6.

Ricard, Y., Nataf, H.-C., Montagner, J.-P., 1996. The three-dimensional seismological model a priori constrained: Confrontation with seismic data. Journal of Geophysical Research: Solid Earth 101 (B4), 8457–8472.

Richards, M. A., Duncan, R. A., Courtillot, V. E., 1989. Flood basalts and hot-spot tracks - plume heads and tails. Science 246 (4926), 103–107.

Richards, M. A., Jones, D. L., Duncan, R. A., Depaolo, D. J., 1991. A mantle plume initiation model for the Wrangellia flood-basalt and other oceanic plateaus. Science 254 (5029), 263–267.

Ritsema, J., Deuss, A., van Heijst, H. J., Woodhouse, J. H., 2011. S40rts: a degree-40 shear-velocity model for the mantle from new rayleigh wave dispersion, teleseismic traveltime and normal-mode splitting function measurements. Geophysical Journal International 184 (3), 1223–1236.

Romanowicz, B., Gung, Y., 2002. Superplumes from the core-mantle boundary to the lithosphere: Implications for heat flux. Science 296 (5567), 513–516.

Schubert, G., Masters, G., Olson, P., Tackley, P., 2004. Superplumes or plume clusters? Physics of the Earth and Planetary Interiors 146 (1-2), 147–162.

Seton, M., Muller, R. D., Zahirovic, S., Gaina, C., Torsvik, T. H., Shephard, G., Talsma, A., Gurnis, M., Turner, M., Maus, S., Chandler, M., 2012. Global continental and ocean basin reconstructions since 200 ma. Earth-Science Reviews 113 (3-4), 212–270.

Sheridan, R. E., 1987. Pulsation tectonics as the control of long-term stratigraphic cycles. Paleoceanography 2 (2), 97–118.

Sleep, N. H., 1990. Hotspots and mantle plumes - some phenomenology. Journal of Geophysical Research-Solid Earth and Planets 95 (B5), 6715–6736.

Staudigel, H., King, S. D., 1992. Ultrafast subduction: the key to slab recycling efficiency and mantle differentiation? Earth and Planetary Science Letters 109 (3), 517 – 530.

Steinberger, B., Torsvik, T. H., 2008. Absolute plate motions and true polar wander in the absence of hotspot tracks. Nature 452 (7187), 620–U6.

Tan, E., Choi, E., Thoutireddy, P., Gurnis, M., Aivazis, M., 2006. Geoframework: Coupling multiple models of mantle convection within a computational framework. Geochemistry Geophysics Geosystems 7.

Tan, E., Gurnis, M., 2005. Metastable superplumes and mantle compressibility. Geophysical Research Letters 32 (20).

Tarduno, J. A., McWillams, M., Sleep, N., 1990. Fast instantaneous oceanic plate velocities recorded by the Cretaceous Laytonville limestone: Paleomagnetic analysis and kinematic implications. Journal of Geophysical Research: Solid Earth 95 (B10), 15503–15527.

Taylor, B., 2006. The single largest oceanic plateau: Ontong Java-Manihiki-Hikurangi. Earth and Planetary Science Letters 241 (3-4), 372–380.

Thorne, M. S., Garnero, E. J., 2004. Inferences on ultralow-velocity zone structure from a global analysis of $SPdKS$ waves. Journal of Geophysical Research-Solid Earth 109 (B08301).

Timm, C., Hoernle, K., Werner, R., Hauff, F., van den Bogaard, P., Michael, P., Coffin, M. F., Koppers, A., 2011. Age and geochemistry of the oceanic Manihiki Plateau, SW Pacific: New evidence for a plume origin. Earth and Planetary Science Letters 304 (1-2), 135–146.

Torsvik, T. H., Smethurst, M. A., Burke, K., Steinberger, B., 2006. Large igneous provinces generated from the margins of the large low-velocity provinces in the deep mantle. Geophysical Journal International 167 (3), 1447–1460.

Vanderkluysen, L., Mahoney, J. J., Koppers, A. A. P., Beier, C., Regelous, M., Gee, J. S., Lonsdale, P. F., 2014. Louisville Seamount Chain: Petrogenetic processes and geochemical evolution of the mantle source. Geochemistry Geophysics Geosystems 15 (6), 2380–2400.

Zhong, S. J., Zuber, M. T., Moresi, L., Gurnis, M., 2000. Role of temperature-dependent viscosity and surface plates in spherical shell models of mantle convection. Journal of Geophysical Research-Solid Earth 105 (B5), 11063–11082.

Zhou, Y., Nolet, G., Dahlen, F. A., Laske, G., 2006. Global upper-mantle structure from finite-frequency surface-wave tomography. Journal of Geophysical Research-Solid Earth 111 (B4).

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. Plate velocities at the surface are shown in color. In (d) Grey line is coastline at present day. The present-day locations of the Ontong-Java, Manihiki, and Hikurangi plateaus are shown as green polygens.

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. Plate velocities at the surface are shown in color. The present-day locations of the Ontong-Java, Manihiki, and Hikurangi plateaus are shown as green polygens.

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

Figure 5: Present day temperature 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 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: Temperature 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 8: 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 9: Dynamic topography for calculation with lower Rayleigh Number in Table 2 at 120 Ma.

Figure 10: 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 11: 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 $\Delta T = 2500^{\circ}$.

Tables

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 lowermost 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