GEOLOGY | THE GEOLOGICAL SOCIETY OF AMERICA®

https://doi.org/10.1130/G46754.1

Manuscript received 12 July 2019
Revised manuscript received 10 October 2019
Manuscript accepted 21 October 2019

Published online 22 November 2019

Coupled supercontinent–mantle plume events evidenced by oceanic plume record

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ABSTRACT

The most dominant features in the present-day lower mantle are the two antipodal African and Pacific large low-shear-velocity provinces (LLSVPs). How and when these two structures formed, and whether they are fixed and long lived through Earth history or dynamic and linked to the supercontinent cycles, remain first-order geodynamic questions. Hotspots and large igneous provinces (LIPs) are mostly generated above LLSVPs, and it is widely accepted that the African LLSVP existed by at least ca. 200 Ma beneath the supercontinent Pangea. Whereas the continental LIP record has been used to decipher the spatial and temporal variations of plume activity under the continents, plume records of the oceanic realm before ca. 170 Ma are mostly missing due to oceanic subduction. Here, we present the first compilation of an Oceanic Large Igneous Provinces database (O-LIPdb), which represents the preserved oceanic LIP and oceanic island basalt occurrences preserved in ophiolites. Using this database, we are able to reconstruct and compare the record of mantle plume activity in both the continental and oceanic realms for the past 2 b.y., spanning three supercontinent cycles. Time-series analysis reveals hints of similar cyclicity of the plume activity in the continent and oceanic realms, both exhibiting a periodicity of ~500 m.y. that is comparable to the supercontinent cycle, albeit with a slight phase delay. Our results argue for dynamic LLSVPs where the supercontinent cycle and global subduction geometry control the formation and locations of the plumes.

INTRODUCTION

Mantle plumes are responsible for the formation of hotspots, flood basalts, and large igneous provinces (LIPs) (e.g., Ernst, 2014). Seismic tomography reveals the presence of two antipodal large low-shear-velocity provinces (LLSVPs) below Africa and the central Pacific, respectively (Dziewonski, 1984; Becker and Boschi, 2002). It has also been demonstrated that the majority of hotspots and LIPs over the past 200 m.y. have been generated above the two LLSVPs but away from subduction zones (Hager et al., 1985; Torsvik et al., 2016). These LLSVPs are not only hotter than the ambient mantle, but are also believed to be geochemically distinct from it (Boyet and Carlson, 2005; White, 2015). Geophysical studies argue that the LLSVPs cover ~25% of the core-mantle boundary surface and might extend up to 1000 km above it (Ni et al., 2002; He and Wen, 2012; Davies et al., 2015). Volume estimates indicate that LLSVPs could represent between 1% (Burke et al., 2008) and 10% of the entire mantle (Cottaar and Lekic, 2016).

Opposing views exist regarding how and when the two LLSVPs formed, and whether they are fixed and long-lived (>2 Ga) through Earth history (Burke et al., 2008; Dziewonski et al., 2010; Torsvik et al., 2016; Niu, 2018) or dynamic in their formation, evolution (including demise), and geographic locations, linked to the assembly and breakup of supercontinents (Anderson, 1994; Zhong et al., 2007; Li and Zhong, 2009; Zhang et al., 2010; Hassan et al., 2015; Zhong and Liu, 2016; Flament et al., 2017; Bono et al., 2019). Answers to such questions have direct implications on how mantle dynamics interacts with plate tectonics. Paleomagnetic data and mantle plume records from the last two supercontinent cycles have been used to suggest that mantle plumes produced by the antipodal LLSVPs are related to whole-mantle convection due to circum-supercontinent subduction (Li and Zhong, 2009). As such, the formation of mantle plumes both in time and space is dynamically linked to the formation of supercontinents and therefore not necessarily always at the same position. Li et al. (2008, 2004) and Li and Zhong (2009) further speculated that when the antipodal LLSVPs, divided and controlled by the subduction girdle, are off the equator, true polar wander (Gold, 1955; Goldreich and Toomre, 1969) would bring them to the equator.

To further test such contrasting models about the longevity and stability of the LLSVPs in both the continental and oceanic realms, we need to examine the occurrence of mantle plume products on Earth's surface through time. The dynamic LLSVP model predicts a cyclic occurrence of plume activity linked to the 500–700 m.y. supercontinent cycle (Li and Zhong, 2009). The fixed and long-lived model, on the other hand, would predict a stochastic plume intensity (i.e., randomly distributed) through Earth history because in such a case, mantle plumes are unrelated to the supercontinent cycle or plate motion in general. However, the mantle plume history prior to ca. 200 Ma is predominantly preserved in continental records

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CITATION: Doucet, L.S., et al., 2020, Coupled supercontinent–mantle plume events evidenced by oceanic plume record: Geology, v. 48, p. 159–163, https://doi.org/10.1130/G46754.1

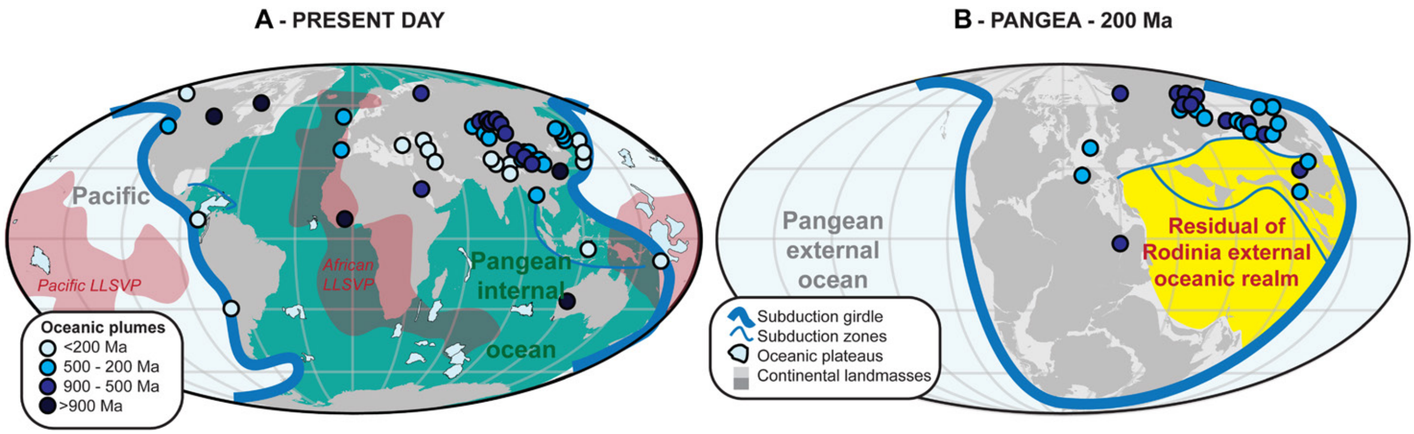

Figure 1. Maps of oceanic mantle plume occurrences in the geological record; locations of ophiolites have been plotted in present-day configuration (A) and in Pangea configuration at 200 Ma (B) (Matthews et al., 2016), highlighting the plume record of the Rodinian external superocean that was consumed during Pangea assembly (Li et al., 2019). Ages of the plume magmatic record in the oceanic realm are color-coded for the following intervals: <200 Ma, 500–200 Ma, 900–500 Ma, and >900 Ma. Also shown are African and Pacific large low-shear-velocity provinces (LLSVPs) (red shading in A), subduction zones (blue lines, both A and B), remanence of the Rodinian external superocean (the Mirovoi ocean; yellow shading in B), and Pangean internal oceans (green shading in A) and external ocean (light blue field, both A and B). Note that in B, only occurrences with magmatic ages corresponding to Rodinian oceans (i.e., 900–200 Ma) are shown.

dominated by continental large igneous provinces (C-LIPs) (Bryan and Ernst, 2008; Ernst et al., 2013), whereas that in the oceanic realm, recording the activity of the sub-oceanic mantle plumes (e.g., plumes above the present-day Pacific LLSVP), is largely missing due to the destruction of oceanic crust and associated plume record in subduction zones.

Nonetheless, work over the past decade has revealed an increasing number of oceanic plume records in ophiolitic belts of past suture zones, where fragments of oceanic crust were accreted to continental margins during subduction (Dilek and Ernst, 2008; Furnes et al., 2014, 2015). Such records include preserved fragments of oceanic LIPs (O-LIPs) (e.g., those reported in the Wrangellia terrane [northwestern North America] by Greene et al. [2010], in the Capricorne orogen [Australia] by Pirajno [2004], and in the Tibetan Plateau by Zhang et al. [2014]), and of oceanic island basalts (OIBs) (e.g., those reported in the central Asian orogenic belts by Safonova [2017] and Yang et al. [2017]). These fragments of largely lost oceanic plume records provide the opportunity to retrieve evidence of mantle plume activity in the oceanic realm for the last 2 b.y. Here, we present the first compilation of an Oceanic Large Igneous Provinces database (O-LIPdb) (see Table DR2 in the GSA Data Repository[1]) that reports both the O-LIP and OIB occurrences preserved in ophiolite belts, including their geochemical, tectonic, and age information. This O-LIPdb, together with an updated C-LIP database (Table DR3, updated from http://www.largeigneousprovinces.org), allows us to analyze and compare mantle plume activities in both the continental and oceanic realms during the last three supercontinent cycles and test the opposing global geodynamic models.

THE O-LIP RECORD OF THE PAST TWO BILLION YEARS

To establish the O-LIPdb, we present a review of the reported geological and geochemical evidence of plume-related materials in ophiolitic belts (Kerr et al., 2000; Bryan and Ernst, 2008; Dilek and Furnes, 2011) (Table DR1), as well as their magmatic ages. The resulting O-LIPdb includes 58 oceanic mantle plume occurrences accreted in ophiolitic belts and 24 O-LIPs preserved on the present-day seafloor (Fig. 1A; Table DR2).

The O-LIPdb encompasses the three last supercontinent cycles over the past 2200 m.y. (Fig. 2) (Evans et al., 2016). We found a low number of O-LIP records for the time period of 2200–1900Ma before the formation of the supercontinent Nuna at 1.6 Ga (Pisarevsky et al., 2014; Pourteau et al., 2018), followed by an ~900 m.y. gap between 1900Ma and 1030 Ma that encompasses the assembly and breakup of Nuna (Fig. 2) (Evans and Mitchell, 2011; Zhang et al., 2012). The recorded gap may have resulted from (1) preservation bias (Stern, 2005), (2) suppressed oceanic plume activity during the "Boring Billion" (Roberts, 2013) or a Proterozoic introversion assembly of Rodinia with a relatively lower level of tectonism (Li et al., 2019), or (3) a lack of studies on rocks of that age. We compiled a more comprehensive record for the past 1 b.y. (Fig. 2) and therefore focus our analysis and interpretation on that time interval.

The frequency of O-LIP activity during the past 1 b.y. is not uniformly distributed, as the stable and long-lived LLSVPs model would predict, assuming that the plates drifted randomly over the dynamically unrelated LLSVPs in the lower mantle. Rather, the O-LIP record exhibits an episodic pattern with a 400–600 m.y. periodicity, not dissimilar to that of the C-LIP record (Prokoph et al., 2004, 2013; Li and Zhong, 2009; Ernst et al., 2013) (Fig. 3; Table DR3). We identified five major plume episodes in the oceanic realm: ca. 1000 Ma, ca. 800 Ma, ca. 600–500 Ma, ca. 350–250 Ma, and ca. 200–40 Ma. These peaks are broadly coeval with plume activity in the continental realm for the same period of time (Fig. 3). More intriguingly, there appears to be a phase shift of plume activity between the continent and oceanic realms during the breakup of Rodinia after 800 Ma (Fig. 3). The Rodinia sub-supercontinent mantle plumes started at least by ca. 825 Ma, if not by ca. 860 Ma, 40–75 m.y. after Rodinia's final assembly at ca. 900 Ma (Li et al., 2008), and resulted in two main C-LIP pulses at ca. 825–800 Ma and 780–750 Ma, and continued during a protracted breakup of Rodinia after 720 Ma (Li and Zhong, 2009). For the concurrent superocean realm (the Mirovoi ocean; McMenamin and McMenamin, 1990), the record (Fig. 3) shows a first post–1 Ga plume activity at 825–800 Ma, overlapping with the major 825–720 Ma continental plume peak. The next episode of oceanic plume activity started at 650 Ma, reaching a local maximum at ca. 600–450 Ma, prior to Pangea assembly at ca. 320 Ma. This latter and more intense oceanic plume peak occurred ~100 m.y. later than that of the continental plume peak. For Pangea breakup, the continental plume breakout started at ca. 300–260 Ma (Li and Zhong, 2009), again ~20–60 m.y. after the final assembly of Pangea. The oceanic plume

[1]GSA Data Repository item 2020046, supplementary information regarding the O-LIP and C-LIP database and time-series analyses, Figures DR1–DR4, and Tables DR1–DR3, is available online at http://www.geosociety.org/datarepository/2020/, or on request from editing@geosociety.org.

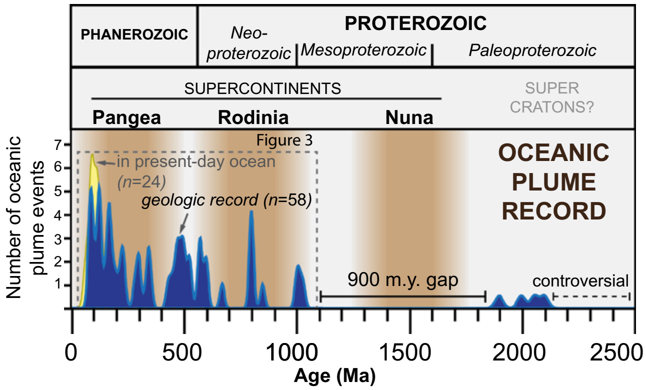

Figure 2. Age distribution of oceanic mantle plume occurrences for the past 2500 m.y. plotted against known supercontinents. Note that Gondwana is not considered a true supercontinent here because it consists of only half of all continents at the time, and is therefore regarded as an intermediate stage during the assembly of the true supercontinent Pangea (Evans et al., 2016). The ophiolitic nature of Archean greenstone belts remains controversial.

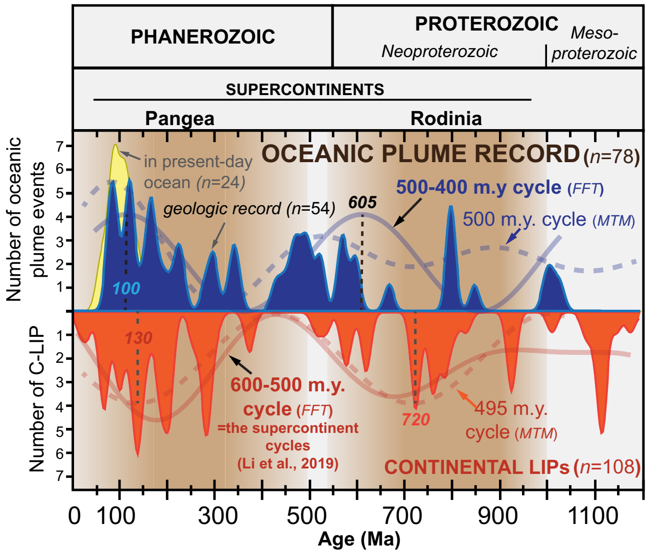

Figure 3. Distribution of both oceanic (blue) and continental (red) mantle plume occurrences for the past 1200 m.y., plotted against the life cycles of supercontinents Rodinia and Pangea, and time-series analyses (fast-Fourier transform [FFT] and multi-taper method [MTM]) of both continental and oceanic large igneous province (C-LIP and O-LIP, respectively) records (see text, and Data Repository figures [see footnote 1] for details).

activity started simultaneously with the Pangean continental plume activity, but the peak of the Pacific oceanic plume peak at ca. 100 Ma occurred ~30 m.y. later than that of the African (Pangean) continental plume peak.

To examine the statistical significance of the potential episodic nature of plume activities in both the continental and oceanic records for the past 1 b.y., we conducted time-series analysis on the two sets of plume records. We used both a basic fast-Fourier transform (FFT) and a more advanced multi-taper method (MTM) (Ghil et al., 2002; Muller and MacDonald, 2000). The results are presented in Figures DR1–DR4. For the basic FFT, the spectral power used is the complex conjugate of the Fourier coefficients, normalized to unit mean power (Muller and MacDonald, 2000). Based on these results, we ran bandpass filters with Gaussian windows to encapsulate the significant peaks identified (Fig. DR1). Both the C-LIP and O-LIP records, following the FFT, can be deconvolved into several periodic components with cycle period ranges of ~2600–1600 m.y., ~1000–900 m.y., ~600–500 m.y., ~350–300 m.y., and ~170–150 m.y. for the C-LIP, and ~1450–710 m.y., ~500–400 m.y., ~250–200 m.y., and ~185–165 m.y. for the O-LIP (Fig. DR1). Due to the limited time interval of the oceanic plume record (i.e., 1200 m.y.), the significance of the 1450–710 m.y. period, centered at 1000 m.y., is difficult to assess but could be related to the longer superocean cycle (Li et al., 2019), where the birth and destruction of a superocean span three supercontinents and a new superocean forms every second supercontinent. The 600–500 and 500–400 m.y. periods that encompass the last two supercontinent cycles are clearly present in both the C-LIP and O-LIP records. Accordingly, the MTM shows a clear statistically robust 500 m.y. period (Fig. DR3) in both the C-LIP and O-LIP records, although this could not statistically be called a cycle for the short O-LIP record featuring only two recurrence intervals. Also, the difference in the spectral analysis between O-LIP and C-LIP could result from the data coverage of each record because FFT and MTM are sensitive to such coverage. Nonetheless, taking such limitations into account, we observe that oceanic and continental plume activities are episodic and almost coeval, as described above, and the time-series analysis highlights a slight time lag between the oceanic and continental plumes for Rodinia (~115 m.y.) and Pangea (~30 m.y.) (Fig. 3). The ~300 m.y. cycles also reproduced what we observed as bursts of oceanic plume events, one occurring in between supercontinents and the other during supercontinent breakup (Fig. DR2). The shorter-wavelength ~170 m.y. cycles had already been identified on the Phanerozoic C-LIP record (Prokoph et al., 2013) and could represent second-order mantle plume pulses due to whole-mantle convective cycles related to subduction (Prokoph et al., 2013).

FIXED VERSUS DYNAMIC LLSVPS—EVIDENCE FROM THE OCEANIC MANTLE PLUME RECORD

The global plume record can be used to test the two mutually exclusive models regarding the origin and evolution of the observed LLSVPs in the lower mantle: (1) the *fixed* model (Burke et al., 2008; Dziewonski et al., 2010; Torsvik et al., 2016; Niu, 2018), where the LLSVPs that generate most of the mantle plumes are long-lived and anchored to the core-mantle boundary; and (2) the *dynamic* model (Zhong et al., 2007; Li and Zhong, 2009; Zhang et al., 2010), where the formation and demise of LLSVPs are dynamically linked to the supercontinent cycle. According to the fixed model, the LLSVPs are independent of plate tectonics. Accordingly, the LLSVPs could have resulted from the primordial differentiation of the Earth and remained largely unchanged in their positions and shapes since their initial formation in deep time (see review by White, 2015). This model predicts that the global mantle plume activity expressed by hotspots, flood basalts, and oceanic plateaus should be stochastic through time, and neither the C-LIP nor the O-LIP records should display systematic correlation with the supercontinent cycle. In contrast, the dynamic model predicts a coupling in time and space between plume activity and the supercontinent cycle (Li and Zhong, 2009). According to this model, the circum-supercontinent subduction girdle produces the antipodal LLSVPs (and thus mantle plumes) soon (i.e., tens of millions of years) after a supercontinent is assembled (Zhong et al., 2007). LLSVPs not only can move rapidly through true polar wander (Li et al., 2004), but can also disappear if the mechanism supporting their presence (the subduction girdle) has vanished (Zhong et al., 2007).

Our newly compiled oceanic mantle plume record, together with the C-LIP record, exhibit episodic variations similar in recurrence rate to the supercontinent cycle itself (Fig. 4), and the O-LIPdb is therefore in agreement with the dynamic model for LLSVPs. In fact, the slight phase delays between supercontinent

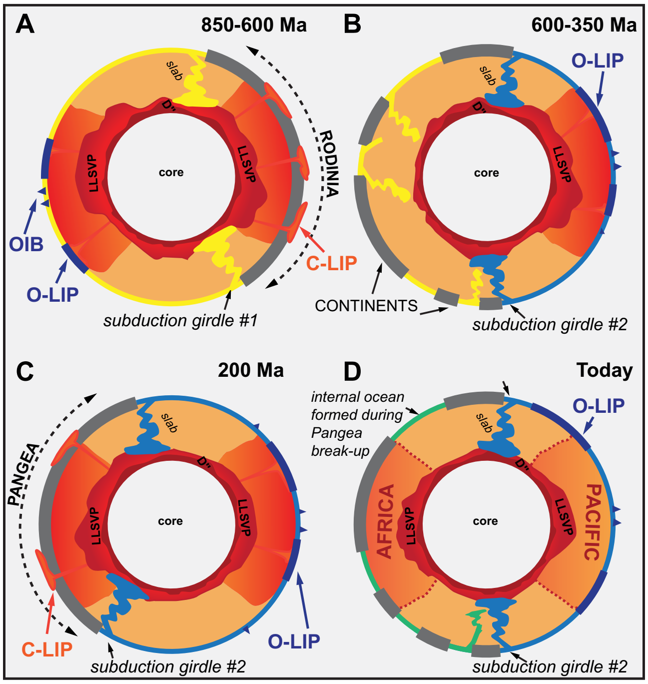

Figure 4. Cartoon of dynamic and coupled supercontinent-plume cycles model (after Li et al., 2019). (A) 850–600 Ma, formation of antipodal large low-shear-velocity provinces (LLSVPs) (and mantle plumes above them) by the circum-Rodinia subduction girdle (yellow) after Rodinia's final assembly at 900 Ma (Li et al., 2004, 2008). "D" marks the lowermost portion of the mantle, at the core-mantle boundary. (B) 600–350 Ma, the protracted and complete breakup of Rodinia, where Rodinia LLSVP (by now under the paleo-Pacific ocean) was still active while Gondwana and Pangea started to assemble above the vanishing antipodal oceanic LLSVP. (C) 200 Ma, formation of antipodal LLSVPs by the circum-Pangea subduction girdle (blue) after Pangea's final assembly at 320 Ma. (D) Present-day configuration of residual antipodal, equatorial African and Pacific LLSVPs after Pangea breakup, but before the assembly of the future supercontinent. O-LIP—oceanic large igneous province; C-LIP—continental large igneous province; OIB—oceanic island basalt.

cycle and the C-LIP record, and also between the C-LIP peaks and O-LIP peaks, are exactly what the dynamic model predicts. According to the dynamic model (see Li and Zhong [2009, p. 150] for a more detailed description), the topography of the LLSVP (and thus the intensity of mantle plume activity) beneath the supercontinent should be more prominent than that of the sub-oceanic LLSVP during the early stage of the supercontinent breakup because it is driven by a small-circle subduction girdle surrounding the supercontinent. As circum-supercontinent subduction begins to retreat with supercontinent breakup, the subduction girdle moves toward the opposing superocean realm, thus gradually driving the formation of a more prominent sub-superocean LLSVP (therefore enhanced O-LIP generation). Our work therefore lends support to the dynamic model over the static and long-lived LLSVP model.

ACKNOWLEDGMENTS

Doucet thanks all members of the Earth Dynamics Research Group (EDRG, Curtin University, Perth, Australia) for discussion and comments. We also thank Dietmar Müller, Shijie Zhong, and an anonymous reviewer for their insightful comments that helped us to improve the quality of the manuscript, and Judith Totman Parrish for editorial handling. This work was supported by the Australian Research Council Laureate Fellowship grant to Li (FL150100133). Ernst was supported in this research by the Russian Science Foundation (grant No. 18-17-00240). This is International Geoscience Programme (IGCP) contribution 648.

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