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Contributions of Scientific Ocean Drilling to Understanding the Emplacement of Submarine LARGE IGNEOUS PROVINCES and Their Effects on the Environment

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Thin section production in progress, Integrated Ocean Drilling Program Expedition 324, Shatsky Rise. *Photo credit: John Beck, IODP/TAMU*

ABSTRACT. The Ontong Java Plateau (OJP), Shatsky Rise (SR), and Kerguelen Plateau/Broken Ridge (KP/BR) represent three large igneous provinces (LIPs) located in oceanic settings. The basement lavas have been investigated through scientific ocean drilling and, in the case of the OJP, fieldwork on the emergent obducted portions of the plateau in the Solomon Islands. Such studies show that these three LIPs have very different characteristics. For example, the KP/BR still has an active hotspot, whereas the OJP and the SR do not. The OJP is remarkable in its compositional monotony across the plateau (the Kwaimbaita geochemical type), with minor compositional variation found at the margins (the Kroenke, Singgalo, and Wairahito types). Shatsky Rise shows more compositional variation and, like the OJP, has a dominant lava type (termed the "normal" type) in the early stages (Tamu Massif), but subsequent eruptions at the Ori and Shirshov massifs comprise isotopically and trace element enriched lavas, likely reflecting a change in mantle source over time. The KP/BR has highly variable basement lava compositions, ranging from lavas slightly enriched above that of normal mid-ocean ridge basalt in the northern portion (close to the South East Indian Ridge) to more enriched varieties to the south and on Broken Ridge, with a continental crust signature present in lavas from the southern and central KP/BR. The OJP and the KP/BR appear to have formed through punctuated magmatic events, whereas the SR was formed by one relatively long, drawn out event. The formation of oceanic LIPs has in many (but not all) cases been synchronous with oceanic anoxic events. This paper focuses on three oceanic plateaus to emphasize the debate surrounding the environmental impact such LIPs may have had, and also highlights the contributions of scientific ocean drilling to our knowledge of oceanic LIP formation and evolution. This new knowledge allows planning for future oceanic LIP drilling.

INTRODUCTION

Volcanic plateaus on the ocean crust represent the oceanic equivalents of continental flood basalts. Some of these formed in nascent ocean basins (e.g., Kerguelen Plateau/Broken Ridge in the southern Indian Ocean), and the earliest outpourings contain the geochemical signature of continental crust contamination (e.g., Storey et al., 1992; Neal et al., 2002; Kinman et al., 2009). Others formed within an oceanic setting and are free from the contamination of continental crust, which allows an examination of the deeper mantle source regions for these voluminous eruptions (e.g., Shatsky Rise, Ontong Java Plateau, Hikurangi Plateau, and Manihiki Plateau in the Pacific Ocean; Heydolph et al., 2014; Fitton et al., 2004; Hoernle et al., 2010; Timm et al., 2011). This paper outlines scientific ocean drilling contributions to understanding the origins, evolution, and environmental impacts of three oceanic LIPs—the Ontong Java Plateau and the Shatsky Rise in the Pacific Ocean, and the Kerguelen Plateau/Broken Ridge in the Indian Ocean.

Ontong Java Plateau

The Ontong Java Plateau (OJP) is situated in the Southwest Pacific and covers an area of 1.86×10^6 km^2 (Coffin and Eldholm, 1994; Table 1), about the size of Alaska, Greenland, or Western Europe (Fitton and Goddard, 2004). However, OJP-like basalts also have been recovered from the Nauru, East Mariana, and Pigafetta basins that surround the OJP (e.g., Saunders, 1986; Castillo et al., 1992, 1994), more than doubling the area of these basalts to ~4×10^6 km^2. Based on a variety of geophysical data, the crustal thickness of the OJP is estimated to be between 30 km and 43 km, with an average around 36 km (e.g., Furumoto et al., 1970, 1976; Murauchi et al., 1973; Hussong et al., 1979; Miura et al., 1996; Richardson and Okal, 1996). Gladczenko et al. (1997) estimated the volume of the OJP to be between 44.4×10^6 km^3 and 56.7×10^6 km^3; the lower estimate is assuming the plateau formed on preexisting older oceanic crust and the higher assumes it formed on young crust at a spreading center. The OJP consists of two parts: the main, or High Plateau in the west and north and the Eastern Salient to the east, the latter having been split during the opening of the Stewart Basin (e.g., Neal et al., 1997; Figure 1a). The OJP lies generally between 2 km and 3 km water depth, although the central High Plateau region rises to ~1,700 m below sea level. The OJP is isostatically compensated (e.g., Sandwell and McKenzie, 1989), with much of the plateau surface being relatively smooth, but the top of the plateau is punctuated by several large seamounts, including Ontong Java atoll,

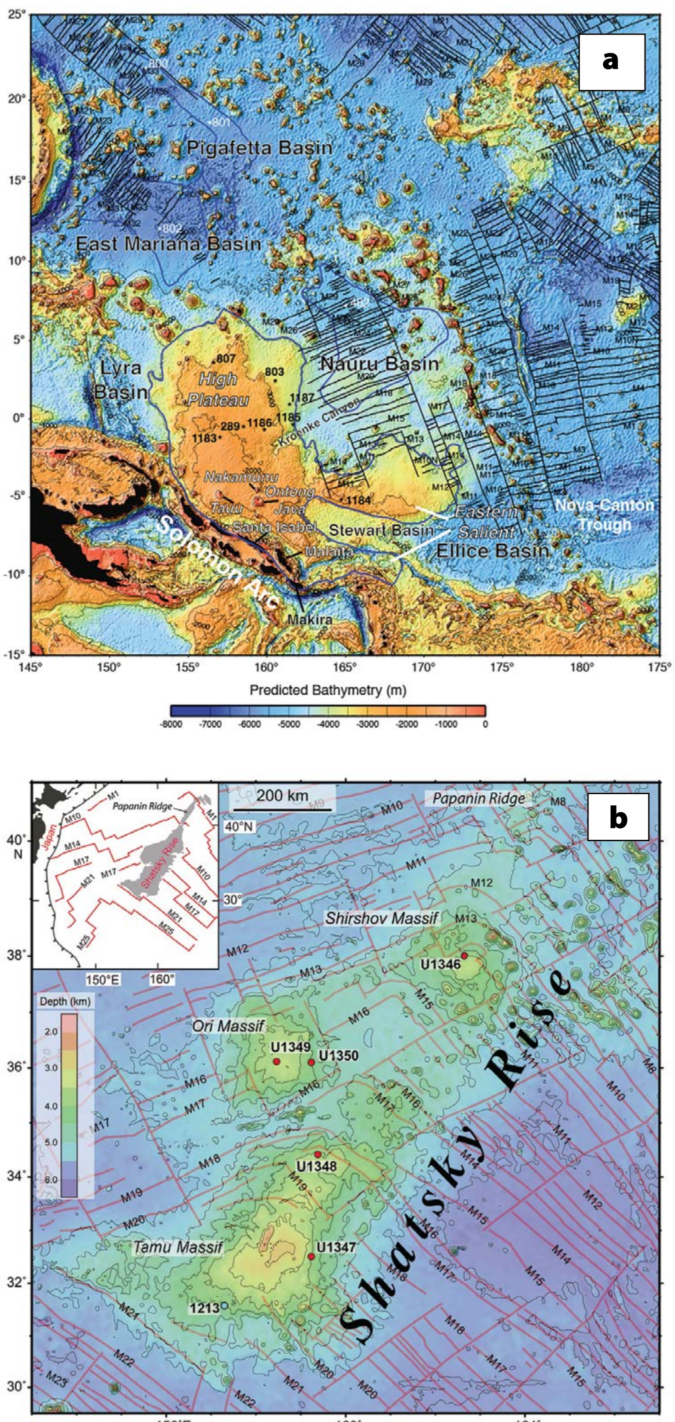
TABLE 1. Sizes of oceanic large igneous provinces.

Name	Area (10^6 km^2)	Volume (10^6 km^3)	References
Ontong Java Plateau	1.86	44–57	1,2,3
Manihiki Plateau	0.77	9–14	1,4
Hikurangi Plateau	0.4–0.8	6.4–18.8	5
Shatsky Rise	0.48	6.9	6
Kerguelen Plateau/Broken Ridge	2.3	15–24	1

1 = Coffin and Eldholm (1994) 2 = Neal et al. (1997)
3 = Gladczenko et al. (1997) 4 = Eldholm and Coffin (2000)
5 = Hoernle et al. (2010) 6 = J. Zhang et al. (2016)

Tauu atoll to the west, and Nukumanu atoll to the north (Figure 1a). The plateau has been considered the product of plume volcanism, formed through the pressure release melting of a large packet of mantle material rising from the deep interior as a bulbous plume head with a long tail that extends back to the source region (e.g., Richards et al., 1989; Campbell and Griffiths, 1990). In this scenario, the impact of the plume head on the rigid lithosphere should induce uplift (e.g., Griffiths et al., 1989; Hill, 1991), but the lack of subaerial volcanism at the thickest part of the OJP High Plateau, coupled with a lack of any identifiable plume tail evidence, has called the origin of the OJP via a surfacing plume head into question (e.g., Korenaga, 2005).

The OJP collided with the Solomon Arc along its southern and southwestern borders (Figure 1a), which promoted a change in subduction direction from the southwest to the northeast around 27–23 million years ago (Ma; Coleman and Kroenke, 1981; Cooper and Taylor, 1985; Petterson et al., 1997, 1999). Consequently, subaerial outcrops of OJP basalt are present on the islands of Malaita, Santa Isabel, Makira,



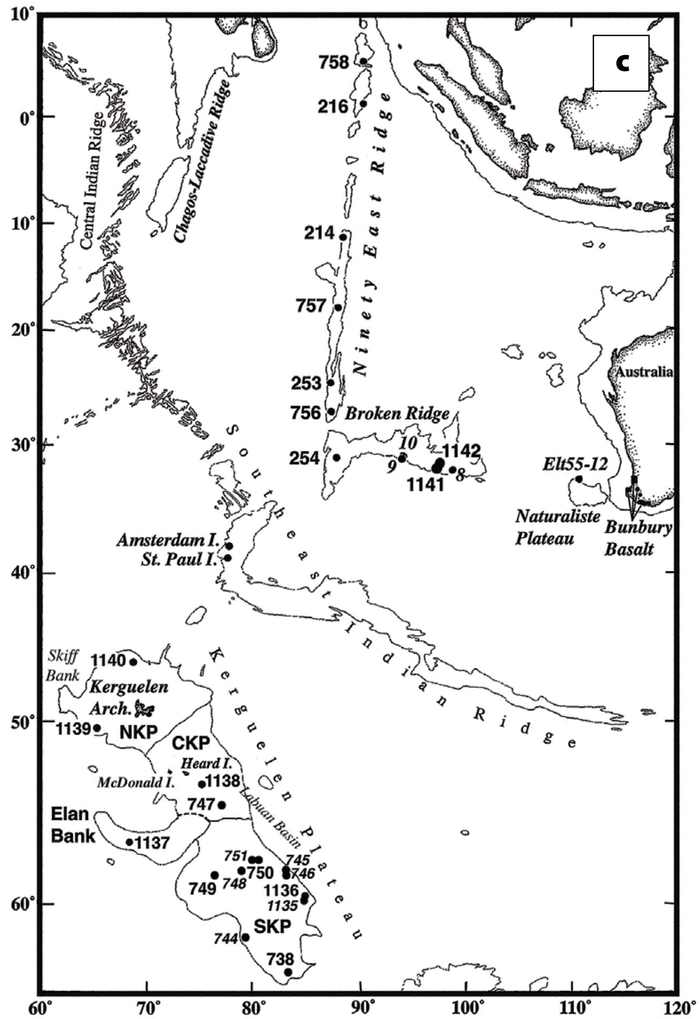


FIGURE 1. (a) Ontong Java Plateau with location names and previous scientific ocean drilling sites. (b) Shatsky Rise with previous scientific ocean drilling sites identified (modified from Sano et al., 2012). (c) Locations of the Kerguelen Plateau/Broken Ridge, the Ninetyeast Ridge, and the Southeast Indian Ridge with scientific ocean drilling and dredge sites identified (modified from Neal et al., 2002).

and Ramos, which have been examined through a number of field seasons (e.g., Petterson et al., 1997, 1999, 2009; Petterson, 2004; Tejada et al., 1996, 2002). Recently, it has been argued that the OJP may be much bigger than previously thought, as the Manihiki (~0.8×10^6 km^2; Coffin and Eldholm, 1994. Table 1) and Hikurangi (~0.5×10^6 km^2; Hoernle et al., 2010; Table 1) plateaus have been hypothesized to have formed with the OJP, but were subsequently rifted apart (Taylor, 2006; Chandler et al., 2012, 2015; Hochmuth et al., 2015).

Shatsky Rise

Shatsky Rise (SR), located ~1,500 km east of Japan, stretches over ~1,700 km and covers an area of ~0.5×10^6 km^2 (about the size of California), with an estimated igneous volume of 6.9×10^6 km^3 (J. Zhang et al., 2016; Table 1). It is proposed to have erupted at a triple junction of divergent plate boundaries due to the convergence of magnetic lineation groups at Shatsky Rise (e.g., Sager et al., 1988, 1999; Nakanishi et al., 1989). It is comprised of three principal volcanic massifs, Tamu, Ori, and Shirshov (Figure 1c; Sager et al., 1999). Tamu was built on oceanic crust of Late Jurassic-Early Cretaceous age, while the Ori and Shirshov massifs were built on progressively younger crust. The Tamu Massif, with a volume of 2.5×10^6 km^3, is thought to have erupted at flood basalt rates of ~1.7 km^3 yr^{-1} (Sager and Han, 1993), which is three orders of magnitude greater than the eruption rate of Kilauea (~0.002 km^3 yr^{-1}). This massif represents a shield volcano that could be the largest on Earth, rivaling the size of the largest such edifice in the solar system, Olympus Mons on Mars (Sager et al., 2013), with a maximum thickness of ~30 km (Korenaga and Sager, 2012). The size, morphology, and trend of decreasing edifice volume over time appears to fit the plume head hypothesis, with a transition from voluminous plume head eruptions at Tamu Massif to smaller-scale eruptions from the narrower plume tail farther along the plateau (Nakanishi et al., 1999; Sager, 2005). However, Shatsky Rise is unique among the large western Pacific oceanic plateaus because it formed during a time of magnetic reversals (Late Jurassic and Early Cretaceous, ~145–125 Ma) so that spreading ridge magnetic anomalies were recorded within the plateau, promising to allow the connection with mid-ocean ridges to be examined (Sager, 2005). This is corroborated by geochemical evidence pointing to a system with strong mid-ocean-ridge basalt characteristics.

Kerguelen Plateau/Broken Ridge

The conjugate Kerguelen Plateau/Broken Ridge (KP/BR) in the southern Indian Ocean (Figure 1b) together cover a vast area (~2.3×10^6 km^2—a little over three times the size of California; Coffin and Eldholm, 1994; Table 1), stand 2 km to 4 km above the surrounding ocean floor, and have thick mafic crusts of 15 km to 25 km (Charvis et al., 1995; Operto and Charvis, 1995, 1996; Borissova et al., 2003) and an estimated volume of 15–24 × 10^6 km^3 (Coffin and Eldhom, 1994; Table 1). In addition, it has been estimated that in total ~2.5×10^7 km of mafic crust has been produced from the Kerguelen hotspot source(s) since ~130 Ma (Coffin et al., 2002). The Cretaceous Kerguelen Plateau/Broken Ridge large igneous province (LIP) is interpreted to represent voluminous volcanism associated with arrival of the Kerguelen plume head below young Indian Ocean lithosphere (e.g., Coffin et al., 2002; Whittaker et al., 2015). Subsequently, rapid northward movement of the Indian Plate over the plume tail formed a 5,000 km long hotspot track from ~82 to 38 Ma, the Ninetyeast Ridge (e.g., Seton et al., 2012). The KP itself is divided into distinct domains: the southern (SKP), central (CKP), and northern Kerguelen Plateau (NKP); Elan Bank; and the Labuan Basin (Figure 1b). Multichannel seismic reflection data show that numerous dipping intra-basement reflections interpreted as subaerial flood basalts form the uppermost igneous crust of the Kerguelen Plateau/Broken Ridge (Coffin et al., 1990; Schaming and Rotstein, 1990). Magma output has varied significantly through time, beginning with low volumes contemporaneous with or postdating continental breakup in Early Cretaceous time, extending through at least one and possibly two peaks in Early and Late Cretaceous time into a preexisting and growing ocean basin, and finally tapering to relatively steady state output in Late Cretaceous and Cenozoic times.

At ~40 million years ago, the newly formed Southeast Indian Ridge (SEIR) intersected the Kerguelen plume's position. As the SEIR migrated northeast relative to the plume, hotspot magmatism became confined to the Antarctic Plate. From ~40 Ma to the present, the Kerguelen Archipelago, Heard and McDonald Islands, and a northwest-southeast trending chain of submarine volcanoes between these islands were constructed on the northern and central sectors of the Kerguelen Plateau/Broken Ridge. Taken together, a ~130 million-year long record of volcanism is attributed to the Kerguelen plume.

COMPOSITIONAL AND AGE DATA

Ontong Java Plateau

The OJP basement has been studied through Deep Sea Drilling Project (DSDP) Leg 30 (Site 289; Stoeser, 1975) and Ocean Drilling Program (ODP) Leg 130 (Sites 803 and 807; Kroenke et al., 1991; Mahoney et al., 1993a,b) and Leg 192 (Sites 1183–1187; Fitton et al., 2004). In addition, fieldwork has been conducted on the obducted portions of the OJP where they outcrop in the Solomon Islands on Santa Isabel, Malaita, Ulawa, Ramos, and San Cristobal (Makira) (Tejada et al., 1996, 2002; Birkhold-VanDyke et al., 1996; Neal et al., 1997; Petterson et al., 1997, 1999, 2009; Birkhold-VanDyke, 2000; Petterson, 2004). Four very similar basalt formations have been recognized: Kroenke, Kwaimbaita, Wairahito, and Singgalo, with the Kwaimbaita basalts being the most voluminous. The basalt compositions are similar (Figure 2a), but there are key differences in the incompatible trace element (ITE) abundances, ratios, and observed isotopic ratios, which give insights into magma chamber processes underlying OJP formation, as well as subtle differences in mantle source regions for these different basalt formations. The Singgalo and Wairahito basalts sit on top of the Kwaimbaita Formation, as do the Kroenke basalts (as demonstrated at Site 1185; Mahoney et al., 2001). The Kroenke Formation basalts are the most primitive (and ITE depleted; Figure 2a,d) recovered from the OJP, being magnesian tholeiites containing 11.8–13.2 wt.% MgO and relatively low abundances of ITEs (Fitton and Goddard, 2004). The Kwaimbaita and Singgalo Formations are named after type locality rivers on the island of Malaita (Tejada et al., 2002). Singgalo basalts correspond to the Unit A basalts and Kwaimbaita to Units C–G basalts (Unit B is a 1–2 m limestone interbed) from ODP Leg 130, Site 807, in the north of the High Plateau (Figure 1a; Mahoney et al., 1993a,b), and the Singgalo composition is also present at ODP Leg 192, Site 1183, as a vitric tuff in the sediment above the Kwaimbaita basalts (Tejada et al., 2004). The Wairahito Formation basalts are named after the type locality, Wairahito River, on the island of Makira (San Cristobal). The Kwaimbaita basalts are the most abundant type, found across the High Plateau and the Eastern Salient (Mahoney et al., 2001). They are evolved tholeiites containing generally 6–8 wt.% MgO, although the Kwaimbaita Formation basalts from the island of Ulawa are more primitive (9.8–11.2 wt.% MgO), but all Kwaimbaita type have higher abundances of ITEs than those from the Kroenke Formation (Figure 2a). The Wairahito Formation contains basalts that are even more evolved with higher ITE abundances (notably Nb) than the Kwaimbaita basalts (Figure 2a) and 4.5–7 wt.% MgO (Birkhold-VanDyke, 2000; Shafer et al., 2004; Petterson et al., 2009). Basalts from the Kroenke, Kwaimbaita, and Wairahito formations are isotopically indistinguishable from each other, indicating they were derived from similar mantle sources (e.g., Shafer et al., 2004; Tejada et al., 2004), and at least the Kroenke and Kwaimbaita basalts have been related by crystal fractionation of olivine (Fitton and Goddard, 2004). The Singgalo Formation

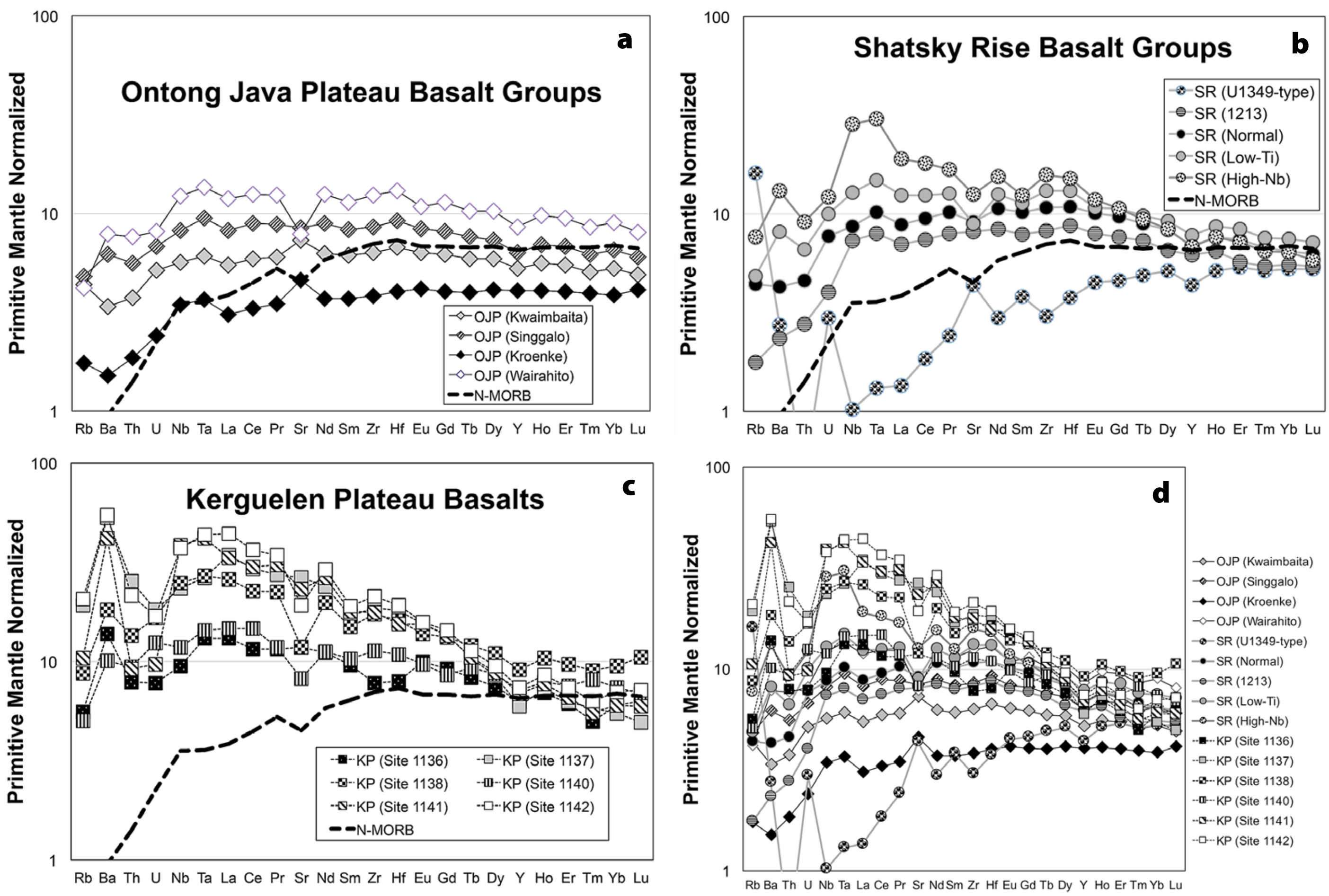


FIGURE 2. Primitive mantle normalized (Sun and McDonough, 1989) trace element plots of average basalt compositions from: (a) the Ontong Java Plateau (OJP; Birkhold-VanDyke et al., 1996; Birkhold-VanDyke, 2000; Petterson et al., 2009; Tejada et al., 1996, 2002, 2004; Fitton and Goddard, 2004; Shafer et al., 2004); (b) Shatsky Rise (SR; Mahoney et al., 2005; Sano et al., 2012); (c) Kerguelen Plateau/Broken Ridge (KP/BR) ODP Leg 183 data (Neal et al., 2002; Weis and Frey, 2002); and (d) all data on the same plot. Normal mid-ocean ridge basalt (N-MORB) composition taken from Sun and McDonough (1989).

is stratigraphically higher than the Kwaimbaita Formation and has similar MgO contents (6.3–7.8 wt.%), but higher abundances of the ITEs (Figure 2a). This is seen across the plateau from the islands of Malaita, Makira, and Santa Isabel in the south of the OJP to ODP Leg 130 Site 807 in the north, and since the Singgalo basalts are isotopically distinct, they must have been derived from a different source region than the Kroenke, Kwaimbaita, and Wairahito basalts (Mahoney et al., 1993a,b; Tejada et al., 1996, 2002; Birkhold-VanDyke, 2000).

The bulk of the OJP formed around 122 Ma with vast outpourings of submarine lava flows that are remarkable for the overall homogeneity of their basalt compositions (e.g., Tejada et al., 1996, 2002; Chambers et al., 2002, 2004; Fitton and Goddard, 2004). Kwaimbaita Formation basalts are found across the plateau, including from the Solomon Islands in the south (Santa Isabel, Malaita, Ulawa, Makira) and at all cored sites (except Site 1187 that recovered only Kroenke basalts). Distinct plagioclase-rich cumulate xenoliths are also found in some Kwaimbaita flows across the plateau (Kinman and Neal, 2006). Although the Singgalo Formation basalts are stratigraphically above those of the Kwaimbaita formation, the $^{40}Ar/^{39}Ar$ ages of the former are indistinguishable within the relatively large (±1–2 million year) analytical age uncertainties (e.g., Mahoney et al., 1993a,b; Tejada et al., 2002). The Kroenke Formation basalts, potentially the parent to the Kwaimbaita basalts (Fitton and Goddard, 2004), are found only at Sites 1185 and 1187 from ODP Leg 192, and a few basaltic clasts from Site 1184 have the Kroenke basalt signature (Shafer et al., 2004). The Wairahito Formation basalts are known only from Makira in the Solomon Islands and again as clasts in the volcaniclastic sequence from Site 1184 (Birkhold-VanDyke, 2000; Shafer et al., 2004). The unaltered basaltic glass in the lower portion of the Site 1184 core is of Kwaimbaita composition (White et al., 2004). Thus, from what we know so far about the OJP, the variation in basaltic types occurs around the margins of the plateau. There were also periodic eruptions subsequent to the main ~122 Ma outpouring of magma, again concentrated around the edge of the plateau, that produced Kwaimbaita-type lavas. This could occur through periodic remelting of the Kwaimbaita source at progressively lower degrees of partial melting (Birkhold-VanDyke et al., 1996; Birkhold-VanDyke, 2000). Compiling the Ar-Ar ages for OJP basalts yields the following eruption periods (all uncertainties are 1 sigma of the mean):

- 121.1 ± 3.8 Ma, n = 26 (ODP Sites 289, 807, 1183, 1184, 1185, 1886, 1187; islands of Malaita, Ramos, Santa Isabel). Data from Mahoney et al. (1993a,b), Chambers et al. (2002, 2004), Tejada et al. (1996, 2002).
- 90.5 ± 3.3 Ma, n = 13 (ODP Site 803; islands of Makira, Santa Isabel). Data from Mahoney et al. (1993a,b), Tejada et al. (1996), Birkhold-VanDyke et al. (1996), Birkhold-VanDyke (2000).
- 61.1 ± 4.6 Ma, n = 8 (islands of Makira and Santa Isabel). Data from Tejada et al. (1996), Birkhold-VanDyke et al. (1996), Birkhold-VanDyke (2000).
- 36.6 ± 1.2 million years ago, n = 4 (island of Makira). Data from Birkhold-VanDyke et al. (1996), Birkhold-VanDyke (2000).

Shatsky Rise

This oceanic plateau was drilled sporadically by DSDP and ODP, principally for paleoceanographic data from its carbonate sediment cap (e.g., Bralower et al., 2006). The basement of Shatsky Rise was drilled during ODP Leg 198 (Site 1213 on Tamu; Shipboard Scientific Party, 2002; Mahoney et al., 2005) and Integrated Ocean Drilling Program (IODP) Expedition 324 (Sites U1346 on Shirshov; U1347, U1348 (volcaniclastics only) on Tamu; U1349, U1350 on Ori; Expedition 324 Scientists, 2010; Sano et al., 2012). Cores recovered both pillow lavas and massive flows, with a trend from thick, massive flows at Tamu Massif to mainly pillow lavas at Shirshov Massif. The core interval recovered at Site U1347, with intervals of massive flows separated by pillow lavas, appears much like that from Leg 192 Sites 1185 and 1186 on the OJP (Shipboard Scientific Party, 2001), implying similar volcanic emplacement. The shift in volcanic style with time is consistent with the expected waning of volcanism from high effusion (thick, massive flows) to lesser effusive outpourings (pillow lavas) with the transition from plume head to tail (Sager et al., 2011, 2016).

The picture that emerged was that Shatsky Rise, although a large LIP, has strong links to and geochemical similarities with mid-ocean ridges (Mahoney et al., 2005; Sager, 2005). Five magma types have been described from the plateau: normal, low-Ti, high-Nb, and U1349 types (Figure 2b; Sano et al., 2012). Additionally, the basalts from Site 1213 (Leg 198) are also distinct, being depleted relative to the normal-type (Figure 2b). The normal type basalt composition is the most abundant in volume, appears on all three massifs, and is similar to normal mid-ocean ridge basalt (N-MORB) composition, but with a slight relative enrichment of the highly ITEs. The low-Ti type is distinguished from the normal type basalt by slightly lower Ti contents at a given MgO, and slight enrichment of the more incompatible ITEs (Figure 2b). The compositions of high-Nb basalts are characterized by distinctively higher contents of the ITEs relative to the other types. U1349 type basalts are composed of more primitive and depleted compositions compared with other SR basalts (Figure 2b). Modeling demonstrates that compositions of the normal-, low-Ti-, and high-Nb-type basalts evolved through fractional crystallization of olivine, plagioclase, and augite in shallow magma chambers (Sano et al., 2012; Heydolph et al., 2014), akin to mid-ocean ridge volcanism.

Ages (Ar-Ar plateau) for Shatsky Rise were derived from two samples of basalts from Tamu Massif Site 1213 (Leg 198) by Mahoney et al. (1995) of 143.7 ± 3.0 Ma and 144.8 ± 1.2 Ma. A longer section of cored basalt was recovered by Expedition 324 at Tamu Massif and yielded basalt ages of 143–144 Ma in the lower portion, but a significantly younger age of 133.9 ± 2.3 Ma was obtained in the upper section (Geldmacher et al., 2014). Compiling of $^{40}Ar/^{39}Ar$ (Mahoney et al., 2005; Koppers, 2010; Geldmacher et al., 2014; Heaton and Koppers, 2014; Tejada et al., 2016) and magnetic anomaly (Sager et al., 1999; Nakanishi et al., 1999) ages shows an age progression to the northeast from Tamu Massif (~144–129 Ma) and the Ori Massif (142–134 Ma) to Shirshov Massif (137–136 Ma), with the Papanin Ridge yielding ages of 128–121 Ma.

Kerguelen Plateau/Broken Ridge

Basement material has been recovered through drilling on Broken Ridge and each part of the Kerguelen Plateau during ODP Legs 119 (Site 738), 120 (Sites 747, 749, 750), and 183 (Sites 1136-1142). Only Leg 183 recovered basalt from Broken Ridge (Sites 1141–1142), with scientific dredging on Broken Ridge affording additional samples from this part of the KP/BR (e.g., Mahoney et al., 1995). Compositions are dominantly tholeiitic, but are highly variable across the Kerguelen LIP (Figure 2c), with alkali basalts sitting atop Broken Ridge, and trachytes, dacites, and rhyolites found at Sites 1137 (as clasts in conglomerate horizons) and 1139 on Skiff Bank (Frey et al, 2000). The latter is interpreted to be part of a later shield volcano constructed on top of the basaltic plateau (Kieffer et al., 2002). The geochemical data show that a continental component must be present in some of the KP/BR basement lavas (e.g., Frey et al., 2002; Ingle et al., 2002a; Neal et al., 2002), perhaps derived from the Eastern Ghats of eastern India for Site 1137 (Nicolaysen et al., 2001; Ingle et al., 2002a,b). Continental remnants must therefore have occurred at shallow depths within the Indian Ocean lithosphere and contaminated the rising mantle-derived melts. The basalts recovered from CKP Site 1138 and NKP Site 1140, however, do not contain the continental crustal signature (Neal et al., 2002).

Duncan (2002) reported basement ages for the Leg 183 basalts, and Whitechurch et al. (1992) for Leg 120 basalts. Their basalt $^{40}Ar/^{39}Ar$ plateau ages from SKP Sites 749, 750, and 1136 are 109 ± 0.7 Ma, 118.2 ± 5 Ma, and 118.99 ± 2.11 Ma, respectively; from Elan Bank Site 1137, the age is 107.53 ± 1.04 Ma; from CKP Sites 1138 and 1139, the ages are 100.51 ± 1 Ma and 68.57 ± 0.61 Ma, respectively; from NKP Site 1140 is 34.34 ± 1.22 Ma, and from Broken Ridge Sites 1141 and 1142, the ages are 95.17 ± 0.77 Ma and 94.87 ± 0.91 Ma, respectively. On the basis of these data, the Kerguelen Plateau/Broken Ridge has been built in several stages over the last ~130 million years.

Paleolatitudes of Kerguelen Plateau/Broken Ridge and Ninetyeast Ridge basalts suggest 3°–10° southward motion of the hotspot relative to the rotation axis, a finding that can be modeled by large-scale mantle flow influencing the location of the plume conduit (Antretter et al., 2002). At ~40 Ma, the newly formed SEIR intersected the plume's position. As the SEIR migrated northeast relative to the plume, hot spot magmatism became confined to the Antarctic plate. From ~40 Ma to the present, the Kerguelen Archipelago, Heard and McDonald Islands, and a northwest-southeast trending chain of submarine volcanoes between these islands were constructed on the northern and central sectors of the Kerguelen Plateau/Broken Ridge.

Combined with the above results and age determinations for basalt from the Ninetyeast Ridge (Duncan, 1978, 1991), age determinations from basalt and lamprophyre attributed to the Kerguelen hotspot in India, Western Australia, and Antarctica (Coffin et al., 2002; Kent et al., 2002) make the ~130 million-year-long record of Kerguelen hotspot activity the best documented of any hotspot trace on Earth. Magma output has varied significantly through time, beginning with low volumes contemporaneous with or postdating continental breakup in Early Cretaceous time, extending through at least one and possibly two peaks in Early and Late Cretaceous time into a preexisting and growing ocean basin, and finally tapering to relatively steady state output in Late Cretaceous and Cenozoic time. The 25 million-year-long duration of peak hotspot output at geographically and tectonically diverse settings is challenging to reconcile with current plume models. Coffin et al. (2002) proposed two alternatives to the standard Hawaii model for hotspots, one involving multiple mantle plume sources and the other a single, but dismembered, plume source. Alternatively, Lin and van Keken (2005) proposed a model of secondary instabilities resulting from the interaction between thermal and compositional buoyancy forces in a thermochemical mantle plume.

LIP PETROGENESIS

The three oceanic LIPs described here represent three different examples of flood magmatism. The OJP and SR contain basalts that have ~MORB-like compositions, but both have a "high-Nb" magma type ("Wairahoito-type" on the OJP; "high-Nb-type" on the SR) that is distinct from MORB (Figure 2a,b). No basalt from the OJP or SR contains any evidence of a continental signature in their basalt compositions. The KP/BR does contain a continental crustal signature in some of the basalts so far recovered, predominantly in the southern and central portions. The basalt ages for the different LIPs also suggest differences: the OJP appears to have erupted regularly every ~30 million years after the ~122 Ma eruption (which was the largest), whereas SR basalt ages indicate that the main edifices were created near the time of surrounding lithosphere formation (Geldmacher et al., 2014; Heaton and Koppers, 2014; Tejada et al., 2016), implying that volcanism occurred at or near the spreading ridges (consistent with the MORB-like character of the lavas). The KP was initiated at the breakup of India, Antarctica, and Australia, with trapped continental selvages at least in the SKP and CKP.

It also appears that magmatic flux waxed and waned in that the KP/BR was built in stages (Duncan, 2002). Explaining the origin of these three LIPs through a unified model has proven difficult.

Ontong Java Plateau

Much of the OJP was erupted in deep water, which was used to argue against a plume head origin for the plateau (e.g., Korenaga, 2005). The central portion of the OJP was taken to be on the High Plateau (Figure 1a) where the crust is thickest, but lavas recovered here at Site 1183 still showed deepwater eruptions, as did other sites on the high plateau (Fitton and Goddard, 2004; Roberge et al., 2004, 2005). Evidence for subaerial eruptions, as expected from a surfacing thermal plume (Campbell, 2007), surprisingly came from the only Eastern Salient drilling, at Site 1184 (Figure 1a), where the volcaniclastic sediment contain glass shards indicating shallow eruption (Roberge et al., 2005), and several horizons of carbonized wood were recovered in the volcaniclastic sequence (Shipboard Scientific Party, 2001; Thordarson, 2004). Neal et al. (1997) predicted there was between 1 km and 4 km of uplift if the OJP formed from a surface plume head. As noted above, this was not over the thickest part of the OJP, as Site 1183 gave an eruption depth of over a kilometer (Roberge et al., 2005), but actually in the Eastern Salient at Site 1184. Using the basalt compositions, estimates of partial melting for the OJP range from ~23%–30% of garnet peridotite (Fitton and Goddard, 2004) or melting over a pressure range (i.e., polybaric melting) that started in garnet peridotite and ended shallower in spinel peridotite (Neal et al., 1997). These melting conditions are consistent with a rising plume head to explain the OJP, but there is no evidence of a long-lived hotspot track associated with the OJP, as plate reconstructions show that the nearby Louisville seamount chain is not a viable candidate (e.g., Yan and Kroenke, 1993).

Taylor (2006) suggested that, based upon seafloor fabric data, three oceanic plateaus in the western Pacific (Ontong Java-Manihiki-Hikurangi; Figure 3) were all formed by a singular huge magmatic event and subsequently rifted apart. For example, comparison of seafloor fabric data between the Hikurangi and the Manihiki plateaus shows they have conjugate margins separated by a former spreading center, the Osbourn Trough (Billen and Stock, 2000; Figure 3) that opened up and separated them (Taylor, 2006). Ages of Integrated Ocean Drilling Program Expedition 329 basalts collected just north of the Osbourn Trough indicate that rifting of the Hikurangi and Manihiki plateaus was "superfast" (~190 mm yr^{-1}), and that the ocean floor basalts produced by this now extinct spreading center were akin to basalts from the Ontong Java and Manihiki plateaus (G.-L. Zhang and Li, 2016). Subsequent kinematic plate reconstructions also show the plausibility of this hypothesis (Chandler et al., 2012, 2015; Hochmuth et al., 2015), which has been termed the Ontong Java Nui or Greater Ontong Java event. These three plateaus sit on ocean crust of similar ages, show similarities in their basalt compositions (Figure 4) and seismic velocity structures, and formed at roughly the same time (Ingle et al., 2007; Hoernle et al., 2010; Timm et al., 2011; Chandler et al., 2012, 2015; Hochmuth et al., 2015; Golowin et al., 2018). Interestingly, reconstruction of these oceanic plateaus to when they were conjoined shows the central portion located around the Eastern Salient of the OJP, the only part that once was subaerial. While this is consistent with the plume model, the lack of a plume tail for these plateaus is not. Also,

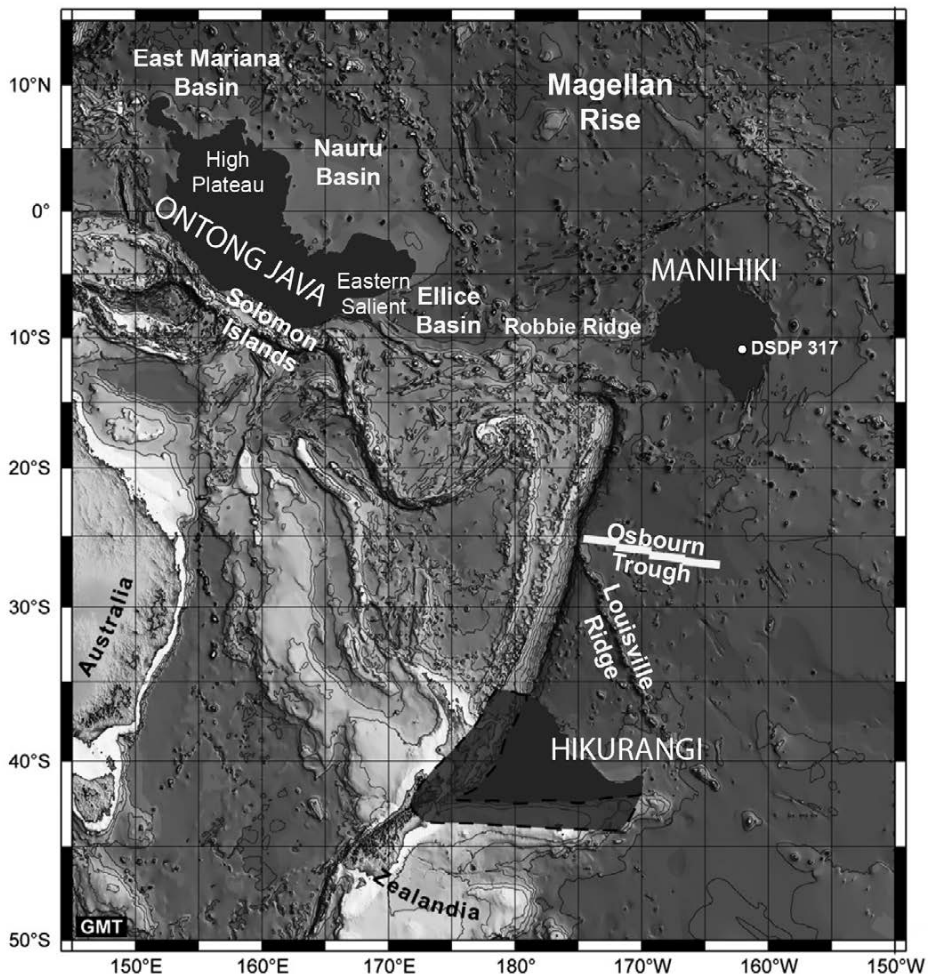


FIGURE 3. Map of the Western Pacific with Ontong Java, Manihiki, and Hikurangi Plateaus and location names. *Map modified from Hoernle et al. (2010)*

if the three plateaus formed at the same time, it would represent the largest magmatic event recorded (~59–90 × 10^6 km^3 of magma production; Table 1).

Shatsky Rise

Geochemical data from most cored lavas give major element ratios that are near normal MORB but trace elements that imply deeper melting than at normal mid-ocean ridges (Sano et al., 2012). Both major element geochemistry and immobile trace element geochemistry indicate 15%–23% partial melting, greater than normal mid-ocean ridge values (Sano et al., 2012; Husen et al., 2013) but less than the 30% estimated for the OJP (Neal et al., 1997; Fitton and Goddard, 2004). The degree of melting implies slightly (~50°C) higher mantle temperatures than normal (Sano et al., 2012; Sager et al., 2016).

IODP Expedition 324 was envisioned as a test between competing hypotheses for the formation of oceanic plateaus: a thermal mantle plume head (Richards et al., 1989; Coffin and Eldholm, 1994) versus shallow, plate-controlled volcanism (Foulger, 2007). Many of the results from the expedition can be framed by the plume head model (e.g., Heydolph et al., 2014). Physical characteristics, including crustal thickness, large magmatic emplacement, apparent rapid emplacement, and possible formation at the edge of the Pacific LLSVP (Large Low Shear-wave Velocity Province; Burke et al., 2008) are all consistent with the plume head hypothesis. Deeper melting than normal MORB, greater percentage of partial melt, and greater than normal temperature are also consistent with this model. Other characteristics can be similarly interpreted. Whereas Tamu Massif lavas are homogeneous in composition with nearly normal MORB chemistry, those from the other massifs are more heterogeneous, including some that are enriched both isotopically and in incompatible trace elements, interpreted as evidence that the source contained recycled oceanic crust, possibly brought up from the lower mantle (Heydolph et al., 2014). The shift to more heterogeneous geochemistry with the lesser volcanic flux of smaller Shatsky Rise edifices was suggested to be indicative of a plume head to plume tail transition because modeling indicates that lower mantle chemical heterogeneities can be preserved in plumes with predominantly vertical motion and limited stirring and at lower degrees of melting (Farnetani et al., 2002; Heydolph et al., 2014). Shatsky Rise samples are also anomalous in $^3He/^4He$ ratios, which were found to be lower than MORB values (Hanyu et al., 2015). Similarly, vanadium isotopes were also found to be different than those found in MORB (Prytulak et al., 2013).

Despite the seeming preponderance of evidence pointing toward a plume source, the connection with ridge volcanism was also strengthened, and some of the plume indications are unequivocal. For example, while the predominant lava type is close to MORB in major element and isotopic chemistry (Sano et al., 2012), some characteristics interpreted as favoring a plume origin, including volume, flux, volume and flux variations over time, and high degrees of partial melting, could also occur from shallow melting of a fertile source (e.g., King and Anderson, 1995; Foulger, 2007). However, geochemical modeling indicates a shallow source could not have generated the erupted basalt compositions (Sano et al., 2012; Husen et al., 2013).

Other indicators are equivocal. Although basal sediments cored on Expedition 324 were deposited in shallow water (Sager et al., 2011), in accord with evidence from volatiles that eruptions were in water <1 km deep (Shimizu et al., 2013), there is little core or geophysical evidence of significant subaerial exposure (Sager et al., 2013, 2016). This does not support the prediction of significant uplift by a thermal plume (Campbell, 2007). Furthermore, the inferred anomalous temperature is much less than expected (~100°–200°C) for a strong thermal plume. Other inferences are model-dependent. Compositional heterogeneity, although indicative of source heterogeneity, does not necessarily imply a deep plume. Likewise, recycled crust can be found at shallow depths (Foulger, 2007), and is not necessarily material recycled to the deep mantle. Moreover, the Jurassic position of Shatsky Rise is probably

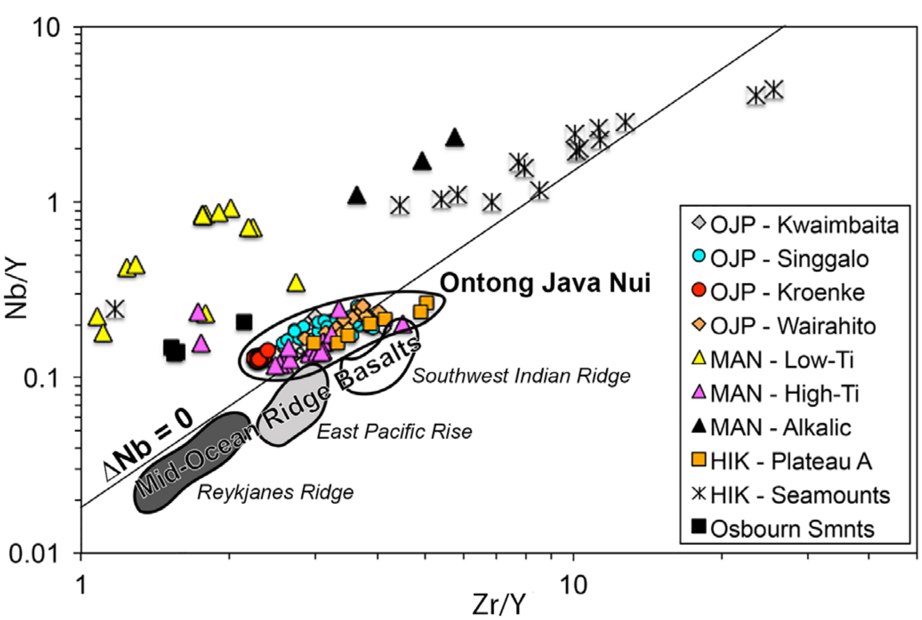


FIGURE 4. Zr/Y vs. Nb/Y plot for Ontong Java, Manihiki, and Hikurangi basalts after Fitton et al. (1997). Data sources as in Figure 2 plus Ingle et al. (2007), Hoernle et al. (2010), Timm et al. (2011), Golowin et al. (2018). OJP = Ontong Java Plateau. MAN = Manihiki. HIK = Hikurangi.

uncertain by ~1,000 km because of poor constraint on the total northward drift of the Pacific Plate during the Cretaceous. As a result, its reconstructed position relative to the LLSVP is also not well known.

Recent research into improving the mapping of magnetic anomalies over Shatsky Rise provides a stronger link to mid-ocean ridge volcanism, indicating that all of the massifs record linear anomalies and thus were formed by spreading (Huang et al., 2018). These results suggest that whatever the source of the Shatsky Rise volcanism, it occurred through a spreading center (which was the path of least resistance), and this may be why many oceanic plateaus apparently formed near spreading ridges. The reason that Expedition 324 research was unable to separate plume from plate mechanisms may be that they are inextricably intertwined (Sager et al., 2016).

Kerguelen Plateau/Broken Ridge

The uppermost basement lavas forming the LIP range widely in Sr, Nd, and Pb isotopic ratios, and each scientific ocean drilling site has distinctive isotopic characteristics (Frey et al., 2003). This points toward differences in source materials and their proportions in the Kerguelen mantle source(s). Site 1140 basalt erupted within 50 km of the SEIR axis at 34 Ma, and the geochemical characteristics of Site 1140 lavas can be explained by mixing, in varying proportions, components derived from the Kerguelen plume and the source of SEIR MORB (Weis and Frey, 2002). In contrast, lavas from Site 738 on the SKP (Mahoney et al., 1995), and from Site 1137 on Elan Bank have radiogenic isotopic ratios that reflect a small and variable but significant role for continental crust in their petrogenesis (Weis et al., 2001; Ingle et al., 2002b). The isotopic evidence indicating a role for continental crust correlates with a relative depletion in abundance of Nb. As to the origin of the continental components contributing to the LIP lavas, intercalated within the basaltic flows at Site 1137 is ~26 m of fluvial conglomerate with clasts of garnet-biotite gneiss containing zircon and monazite (Frey et al., 2000) of Proterozoic age (Nicolaysen et al., 2001). This constitutes the first and only unequivocal evidence of the presence of continental crust within the Kerguelen Plateau/Broken Ridge and in any other oceanic plateau drilled so far. The geochemical data indicate that the signature of continental crust is widely distributed in Cretaceous basalt forming the uppermost basement of the LIP. The most compelling examples are at Site 738 (SKP), Site 1137 (Elan Bank), and Site 747 (CKP) (Mahoney et al., 1995; Ingle et al., 2002b; and Frey et al., 2002, respectively).

A plume source, but more complex than a single plume head and tail model, for the basalt forming the Kerguelen Plateau/Broken Ridge remains a viable hypothesis. Like the active plumes of Hawaii, Iceland, and the Galápagos, lavas forming the Kerguelen Plateau/Broken Ridge are isotopically heterogeneous. Challenging questions posed here are: to what extent is this heterogeneity intrinsic to the plume, and to what extent does the heterogeneity reflect mixing between quite different components, such as oceanic and continental lithosphere. For lavas from some of the Kerguelen Plateau/Broken Ridge sites, the isotopic heterogeneity undoubtedly reflects mixing of plume-related components with components derived from depleted asthenosphere (Site 1140) or continental lithosphere (Sites 738, 747, and 1137; Frey et al., 2003).

Evidence from ODP Legs 119, 120, and 183 clearly demonstrates that large parts of the SKP and CKP that are now submarine were originally subaerial during at least the final stages of plateau construction (Coffin et al., 2000; Mohr et al., 2002). Subsidence estimates for ODP drill sites indicate that the various parts of the Kerguelen Plateau/Broken Ridge subsided at a rate comparable to that for normal Indian Ocean lithosphere (Coffin, 1992; Wallace, 2002). Hence, the original maximum elevations would have been 1 km to 2 km above sea level, and much of the SKP's ~500,000 km^2 area would at one time have been above sea level (Coffin, 1992). The SKP and CKP supported a dense conifer forest with various fern taxa and early angiosperms in late Albian to earliest Cenomanian time (Francis and Coffin, 1992; Mohr et al., 2002). By latest Cenomanian time, the CKP had subsided to a depth that allowed open marine sediments to accumulate.

On Broken Ridge, the vesicularity and oxidative alteration of basement basalts at Sites 1141 and 1142, which formed close to the CKP (Figure 1), are also consistent with a subaerial environment (Keszthelyi, 2002). At SKP Site 1136, inflated pāhoehoe lavas lack features of submarine volcanism (e.g., pillows and quenched glassy margins), suggesting subaerial eruption. The igneous basement complex of Elan Bank (Site 1137) includes basaltic lava flows that erupted subaerially, as indicated by oxidation zones and the presence of inflated pāhoehoe flows. Some interbedded volcaniclastic rocks were deposited in a fluvial environment, consistent with subaerial eruption of the basalt. The NKP (Site 1139) was also subaerial during its final stages of formation, as indicated by a succession of variably oxidized volcanic and volcaniclastic rock. After volcanism ceased, paleoenvironments changed from intertidal (beach deposits) to very high-energy, nearshore (grainstone and sandstone), to low-energy offshore (packstone), to bathyal pelagic (ooze) (Coffin et al., 2000).

ENVIRONMENTAL IMPACT

It has been hypothesized that eruption of flood basalts affected the surface environment (e.g., Self et al., 2005, 2008) and potentially prompted mass extinctions (e.g., Keller, 2005). The extent of environmental impact related to LIP volcanism depends on whether eruptions were subaerial or submarine, and whether the magma passed through coal, petroleum, and/or evaporite deposits on the way to the surface, supplementing the magma's volatile content (Neal et al., 2008; Svensen et al., 2009). The other important factor in assessing the environmental impact of LIP formation is determining the flux of volcanism. It is unfortunate that many of the erupted lavas are low-K tholeiites, as the uncertainty of $^{40}Ar/^{39}Ar$ ages is usually between one and two million years. This uncertainty makes it impossible to assess the flux of volcanism during LIP formation and to determine how many eruptive episodes there were, though these topics remain top priorities in understanding LIP formations and their impacts on the environment (Neal et al., 2008).

Evidence indicates a link between LIP formation and environmental crises. For example, the largest mass extinction at the Permo-Triassic boundary was synchronous with the eruption of the Siberian and Emeishan Traps (Wignall, 2005; Wignall et al., 2009). Oceanic LIP formation would have had a more subtle environmental impact, given the general submarine nature of the eruptions (e.g., Ernst and Youbi, 2017). In order for submarine LIP formation to generate a global impact on the ocean, the ocean needs to be well mixed, but LIP volcanism alone could not have released enough CO_2 to have a global environmental impact (e.g., Kerr, 2005; Kerr and Mahoney, 2007; Naafs et al., 2016). However, it is likely that a complex positive feedback mechanism triggered by the volcanically derived CO_2 led to increased CO_2 and elevated temperatures associated with submarine oceanic plateau volcanism (e.g., Kerr, 2005). The initial emissions from oceanic plateau volcanism were probably a mixture of CO_2, SO_2, and halogens (Self et al., 2005), which would have made the ocean at least more anoxic and acidic locally. This increased acidity would have led to the dissolution of shallow-water carbonates, thus releasing more CO_2 to the ocean and the atmosphere (Kerr, 2005). In addition, the main phase of Ontong Java Plateau formation (~120 Ma) appears to correlate with a massive release of methane, which may have come from warming of methane hydrate that had been stored beneath the ocean floor (Jahren, 2002; Naafs et al., 2016). As CO_2 solubility decreases, the warmer the ocean water becomes. Kerr (1998) proposed that such a scenario would relatively rapidly result in the establishment of a runaway greenhouse effect. Also contributing to oceanic anoxia is the fact that a warmer ocean dissolves less O_2 (de Boer, 1986).

The reaction of O_2 with trace metals and sulfides in hydrothermal fluids as well as enhanced phytoplankton growth also decrease the amount of oxygen in seawater (Sinton and Duncan, 1997). Injection of a large hydrothermal plume could easily rise through the water column and spread laterally over a significant proportion of the ocean's surface. The metal-rich waters of such massive hydrothermal plumes may well have stimulated increased levels of organic productivity (e.g., increased iron can stimulate phytoplankton productivity) in nutrient-poor surface waters (Coale et al., 1996; Sinton and Duncan, 1997). Such productivity could have led to further oxygen reduction in ocean waters as organic material decayed and sank through the water column.

While there is evidence for the subaerial eruption of parts of the KP/BR and OJP, most of these plateaus (and SR) formed through submarine eruptions. If the Ontong Java, Manihiki, and Hikurangi plateaus formed from a "superplume" event (see Larson, 1991), the environmental impact may have been immense. A global oceanic anoxic event (OAE-1a or the "Selli" event at 124–122 Ma; Coccioni et al., 1987; Méhay et al., 2009) was coincident with the initial (and largest) eruption of the OJP at ~122 Ma. Tejada et al. (2009) show a causative link between OJP emplacement and OAE-1a using osmium isotopes, where two large influxes of non-radiogenic osmium were observed within a period of ~2 million years starting in the lower Aptian and ending just above OAE-1a. These Os influxes are consistent with huge outpourings of mantle-derived submarine basalts though major eruptions at the formation of the Ontong Java Nui. While the formation of the OJP (and potentially the formation of the Manihiki and Hikurangi Plateaus) had global implications for the surface environment, this event also changed the nature of the upper mantle in the western Pacific. The oceanic crust that pre-dates this event is different in composition from that which post-dates it (Janney and Castillo, 1997).

Shatsky Rise erupted prior to the OJP over a period of ~24 million years, from 145 Ma to 121 Ma, with the largest eruptions occurring earlier and volcanism waning thereafter (e.g., Sager et al., 2016; Tejada et al., 2016). Two OAEs over this period have been reported: the Weissert Event (late Valanginian at ~140.5 Ma) and the Faraoni Event (late Hauterivian at ~131 Ma) (e.g., Erba et al., 2004; Jenkyns, 2010). OAEs were also forming during the construction of the KP/BR (OAE-1b,c,d occurring at approximately 111.5 Ma, 103–104 Ma, and 100 Ma, respectively; e.g., Erba, 2004). This suggests that all three LIPs studied here had major effects on ocean chemistry, although the links between the SR and the KP/BR with OAEs are not as compelling as the link between the OJP and OAE-1a (e.g., Tejada et al., 2009).

FUTURE OCEANIC LIP DRILLING

In 2007, a multidisciplinary group of 80 scientists met July 22–25 at the University of Ulster in Coleraine, Northern Ireland, to discuss strategies for advancing our understanding of LIP formation, evolution, and environmental impacts (Neal et al., 2008). This workshop produced a series of questions that could be addressed, at least in part, by scientific ocean drilling:

- To what extent do melting anomalies reflect excess fertility in the mantle rather than excess mantle temperature?
- Do thermo-chemical plume models account for the observations of uplift and basalt chemistry around oceanic plateaus?
- What is the internal architecture of an oceanic LIP?
- Do LIPs initiate continental breakup, or does continental breakup initiate LIP development?

- What was the mode(s) of LIP emplacement? Fissure eruptions (e.g., OJP)? Large volcanic centers (e.g., SR)?
- Was there more than one pulse of voluminous volcanism associated with LIP formation or were there a series of smaller magmatic pulses?
- How long did the pulse(s) of LIP volcanism last?
- What were the environmental impacts of such voluminous volcanism?
- Are there differences between subaerial and submarine LIP emplacement?
- What is the overall architecture of an oceanic plateau sequence?
- What is the nature of seaward-dipping reflectors that are highly significant components of several LIPs (e.g., North Atlantic, Kerguelen)?
- What is the relationship between felsic LIP magmas and the more common mafic (basaltic) varieties?
- Are environmental perturbations, including mass extinctions, directly caused by LIP emplacement?

While it can be argued that in the ~12 years since the last LIP workshop, progress has been made in addressing at least some of these questions, they still remain valid for developing future scientific ocean drilling strategies. Neal et al. (2008) highlighted a number of ways scientific ocean drilling could advance our understanding of LIPs:

- Obtaining deep sections within multiple LIPs to examine magmatic (and therefore mantle source) variability through time
- Defining the nature of melting anomalies (i.e., compositional vs. thermal) that produce LIPs
- Defining precise durations of oceanic LIP volcanic events
- Defining modes of eruption-constant effusion over several million years or several large pulse events over the same time interval
- Establishing relationships among oceanic LIPs, OAEs, and other major environmental changes (e.g., ocean acidification and fertilization)

Obtaining a continuous record of syn-LIP sediments may be one of the most important endeavors for the future. With such materials, we will be able to address the last three points above. Potential sites include the sediments on top of the Magellan Rise in the Southwest Pacific, which could be used to examine the duration of volcanic events, modes of eruption, and environmental impacts of the Ontong Java, Hikurangi, and Manihiki oceanic plateaus. Mercury may be a promising new tracer for fingerprinting major volcanic events recorded in sediment (Font et al., 2016; Jones et al., 2017).

An example of what has changed since 2007 is provided in work on the OJP, beneath which there was reported to be an anomalous seismically slow region that extends to 300 km beneath the plateau (Richardson et al., 2000). This had been interpreted to represent the depleted mantle root that remained after a superplume (see Larson, 1991) formed the OJP, and that is now (still) moving with the plateau (Klosko et al., 2001). Subsequent seismic data have questioned this interpretation. For example, Covellone et al. (2015) used a combination of Rayleigh wave data extracted from ambient noise and earthquakes (and an iterative finite-frequency tomography method) to show that shear wave speeds were actually faster beneath the OJP, rather than slower as found in earlier work (Richardson et al., 2000; Klosko et al., 2001). This was interpreted as being consistent with eclogite entrainment by the surfacing plume head, which also retarded surface uplift. The fast seismic signature in the mantle beneath the OJP down to ~300 km was also supported by Japanese ocean bottom seismometer data (Isse et al., 2018; Obayashi et al., 2018). However, the geochemistry of erupted OJP basalts is not consistent with eclogite in the mantle sources (e.g., Tejada et al., 2004). Further research is required to resolve this issue.

> "It is evident that large igneous provinces can have global environmental impacts, and understanding their origin and evolution is important for understanding how our planet has evolved, potentially from the core-mantle boundary to the surface environment."

It is important to realize that the last scientific ocean drilling conducted on the OJP was 19 years ago (Leg 192 to the OJP in 2000), 10 years ago for the SR (Expedition 324 in 2009), and 20 years ago for the KP/BR (Leg 183 in 1998/1999). Since that time and based upon the data already obtained through drill cores, new questions can be asked that can be only or best addressed by scientific ocean drilling. Examples include:

- What is (are) the possible interaction(s) of LIP emplacement and the plate tectonic cycle (continental breakup, the subduction process, enhancement of mid-ocean ridge spreading)?
- What, if any, is the temporal relationship between LIP events and changes in the reversal frequency of Earth's magnetic field (e.g., the formation of the OJP is approximately synchronous with the beginning of the Cretaceous Normal Superchron; e.g., Granot et al., 2012, and references therein)?

- What role do LIPs have in initiation of continent formation and continental growth (e.g., Wrangellia and OJP; Samson et al., 1990; Wignall, 2001; Kerr, 2003; Miura et al., 2004)?
- Do all LIPs have the chemical inventory of LIPs to produce ore deposits, such as those associated with the Siberian Traps (e.g., Naldrett et al., 1996; Malitch et al., 2014)?
- Can the magnitude of the environmental effects induced by LIP eruption be quantified?

The first 50 years of scientific ocean drilling have informed us about the formation, evolution, and environmental impacts of oceanic LIPs. Based upon previous results, the next 50 years will see new expeditions to investigate oceanic LIPs that will address more sophisticated questions. Such investigations would also inform us of similar volcanic constructs on the Moon, Mars, Mercury, and Venus (e.g., Head and Coffin, 1997; Hansen, 2007; Head et al., 2011; Neal et al., 2017).

SUMMARY

Scientific ocean drilling has given us major insights into the origin and evolution of the three LIPs highlighted here. However, the data thus far collected show each has unique characteristics that are difficult to reconcile with a unifying petrogenetic model. For example, the KP/BR still has an active hotspot, whereas the OJP and SR do not. The OJP is remarkable in that its compositional homogeneity is transitional between normal (N-)MORB and enriched (E-)MORB across the plateau, whereas SR in the early stages (Tamu Massif) is similar to N-MORB, but subsequent eruptions at the Ori and Shirshov massifs include both geochemically more enriched and depleted compositions. The KP/BR has highly variable basement lava compositions, with a continental crust signature present in lavas from the southern and central KP, and there is strong evidence of subaerial eruptions across the plateau. The OJP and KP/BR appear to have formed through punctuated volcanic events, whereas SR was formed by one relatively long event.

Neal et al. (2008) summarized future LIP drilling targets that are still valid today. One issue, however, that could be resolved by future LIP drilling (and not included in the compilation by Neal et al., 2008) is that the OJP may be much larger than originally thought, as the Manihiki and Hikurangi plateaus are hypothesized to have formed from the same magmatic event that formed the OJP. Only one site has been drilled on Manihiki (DSDP Leg 33, Site 317) and none have been drilled on Hikurangi. Most of the data we have for each of these plateaus are from dredged samples. This huge magmatic event also affected the Pacific upper mantle as MORB prior to the OJP was of a different composition than MORB that post-dates its formation.

What is common among the OJP, the SR, and the KP/BR is that coincident with or shortly after their emplacements, a crisis in the world ocean resulted in an unprecedented die-off that is highlighted by global black shale horizons representing oceanic anoxic events. It is evident that LIPs can have global environmental impacts, and understanding their origin and evolution is important for understanding how our planet has evolved, potentially from the core-mantle boundary to the surface environment.

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