Pin-pricking the elephant: evidence on the origin of the Ontong Java Plateau from Pb–Sr–Hf–Nd isotopic characteristics of ODP Leg 192 basalts

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Abstract: Age-corrected Pb, Sr and Nd isotope ratios for early Aptian basalt from four widely separated sites on the Ontong Java Plateau that were sampled during Ocean Drilling Program Leg 192 cluster within the small range reported for three earlier drill sites, for outcrops in the Solomon Islands, and for the Nauru and East Mariana basins. Hf isotope ratios also display only a small spread of values. A vitric tuff with $\varepsilon_{Nd}(t) = +4.5$ that lies immediately above basement at Site 1183 represents the only probable example from Leg 192 of the Singgalo magma type, flows of which comprise the upper 46–750 m of sections in the Solomon Islands and at Leg 130 Site 807 on the northern flank of the plateau. All of the Leg 192 lavas, including the high-MgO (8–10 wt%) Kroenke-type basalts found at Sites 1185 and 1187, have $\varepsilon_{Nd}(t)$ between +5.8 and +6.5. They are isotopically indistinguishable from the abundant Kwaimbaita basalt type in the Solomon Islands, and at previous plateau, Nauru Basin and East Mariana Basin drill sites. The little-fractionated Kroenke-type flows thus indicate that the uniform isotopic signature of the more evolved Kwaimbaita-type basalt (with 5–8 wt% MgO) is not simply a result of homogenization of isotopically variable magmas in extensive magma chambers, but instead must reflect the signature of an inherently rather homogeneous (relative to the scale of melting) mantle source. In the context of a plume-head model, the Kwaimbaita-type magmas previously have been inferred to represent mantle derived largely from the plume source region. Our isotopic modelling suggests that such mantle could correspond to originally primitive mantle that experienced a rather minor fractionation event (e.g. a small amount of partial melting) approximately 3 Ga or earlier, and subsequently evolved in nearly closed-system fashion until being tapped by plateau magmatism in the early Aptian. These results are consistent with current models of a compositionally distinct lower mantle and a plume-head origin for the plateau. However, several other key aspects of the plateau are not easily explained by the plume-head model. The plateau also poses significant challenges for asteroid impact, Icelandic-type and plate separation (perisphere) models. At present, no simple model appears to account satisfactorily for all of the observed first-order features of the Ontong Java Plateau.

Several massive volcanic plateaus appeared at equatorial to mid-southern latitudes in the Pacific Basin between the latest Jurassic and the middle Cretaceous. Of these, the Ontong Java Plateau (OJP; Fig. 1) in the western Pacific is the world's largest (the 'elephant' in our title), with

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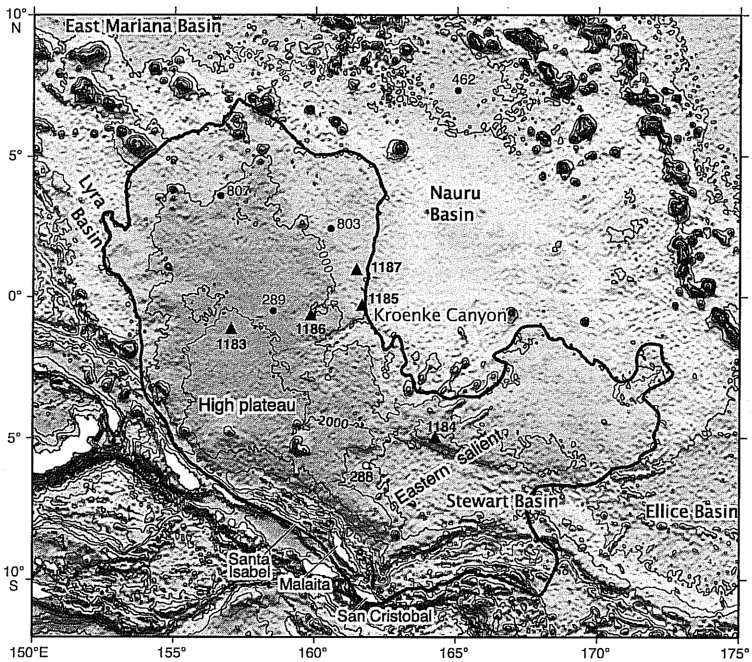

Fig. 1. Map of the Ontong Java Plateau (outlined) showing the locations of sites drilled during Leg 192 (triangles). Dots are previous drill sites that reached basement. The white dot represents Site 288, which did not reach basement but bottomed in Aptian limestone. The bathymetric contour interval is 1000 m (predicted bathymetry from Smith & Sandwell 1997).

a Greenland-size area of approximately 2 × 10^6 km^2 and an average crustal thickness of about 32 km (e.g. Gladczenko *et al.* 1997; Richardson *et al.* 2000; J.G. Fitton & M.F. Coffin pers. comm. 2003; Miura *et al.* 2004). Despite their great size, the origin of the Pacific plateaus is only poorly understood, having been attributed variously to: (1) cataclysmic melting in the inflated heads of newly risen mantle plumes (e.g. Richards *et al.* 1989) or even a single 'super' plume (Larson 1991); (2) formation above near-ridge plume tails over much longer periods of time (e.g. Mahoney & Spencer 1991; Ito & Clift 1998); (3) plate separation above extensive, near-solidus, but non-plume regions of the shallow asthenosphere (e.g. Anderson *et al.* 1992; Smith & Lewis 1999; Hamilton 2003); and (4) asteroid impact (Rogers 1982). The variety of models that have been applied in part reflects a lack of detailed knowledge of Pacific plateau composition and age, which in turn is a result of the very sparse sampling of crustal basement. Although it is by far the largest, the OJP is also presently the best sampled of any Pacific plateau.

Along its southern margin the plateau has collided with the Solomon island arc, where fragments of OJP crustal basement have been uplifted and exposed in several places, particularly in the islands of Santa Isabel, Malaita and San Cristobal (also known as Makira) (see Petterson *et al.* 1999). Away from the collision zone, however, basement on the plateau is buried under a thick marine sedimentary section, itself submerged approximately 1.7–4 km below sea level. Thus, drilling is the only effective means of sampling volcanic basement, in general. Until recently, it had been reached at only three drill sites; penetration of 149 m into the volcanic section was achieved at one site (Site 807), but the other two holes penetrated only 9 (Site 289) and 26 m (Site 803) into basement (Fig. 2) (Andrews *et al.* 1975; Kroenke *et al.* 1991). Sampling of basement was augmented considerably in September and October of 2000, when Ocean Drilling Program (ODP) Leg 192 cored sections at four sites on the OJP's main or high plateau (Sites 1183, 1185, 1186 and 1187) to subbasement depths ranging from 65 to 217 m (Mahoney *et al.* 2001). A fifth site, Site 1184, cored 338 m of a basaltic volcaniclastic sequence on the eastern lobe or salient of the plateau.

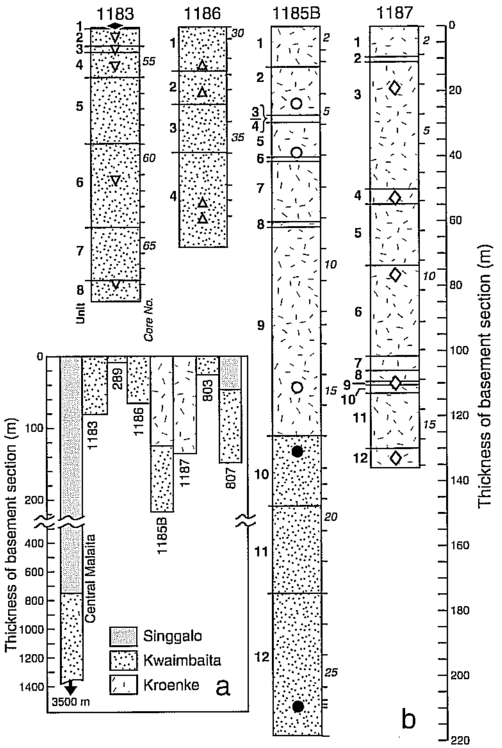

Fig. 2. (**a**) Basement thickness and magma types in OJP drill sites and central Malaita. (**b**) Basement sections of Leg 192 sites on the high plateau, showing drill-core number, unit boundaries and magma type. Symbols indicate sample locations.

Prior to Leg 192, study of samples from the Solomon Islands and the three previous drill sites had established that basement at all of these locations is composed of massive and pillowed submarine lava flows. The rock is low-K tholeiitic basalt with only a small range of major-element, trace-element and Nd–Pb–Sr isotopic composition, a surprising result in view of the immensity of the plateau (e.g. Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002; Babbs 1997; Neal *et al.* 1997). Dating by $^{40}Ar–^{39}Ar$ revealed that a major plateau-forming event occurred in the early Aptian, with ages clustering around 122 Ma; however, ages near 90 Ma were obtained for Site 803 and parts of Santa Isabel and San Cristobal (Mahoney *et al.* 1993; Birkhold-VanDyke *et al.* 1996; Parkinson *et al.* 1996; Tejada *et al.* 1996, 2002). All of the basalts were found to be distinct from both N-MORB (normal mid-ocean ridge basalt) and OIB (ocean island basalt). They have low, MORB-like concentrations of many incompatible elements, but OIB-like isotopic characteristics rather similar to those of the Hawaiian shield volcanoes of Kilauea and Mauna Loa; moreover, unlike either N-MORB or OIB, their primitive-mantle-normalized incompatible-element patterns and chondrite-normalized rare-earth patterns are relatively flat. Despite the limited compositional variability, two isotopically distinct, stratigraphically separate groups of basalt, termed the Kwaimbaita and Singgalo types by Tejada *et al.* (2002), were identified at Site 807, in Santa Isabel and in Malaita. The stratigraphically lower Kwaimbaita type is characterized by higher age-corrected $\varepsilon_{Nd}(t)$ (+5.4–+6.4), higher $(^{206}Pb/^{204}Pb)_t$ (18.21–18.42) and lower $(^{87}Sr/^{87}Sr)_t$ (0.7034–0.7039) than the overlying Singgalo type (with +3.7–+5.3, 17.71–17.99 and 0.7039–0.7044, respectively). Kwaimbaita-type basalts also tend to have slightly lower ratios of highly incompatible elements to moderately incompatible elements. The thickest basement section (3.5 km) is found in central Malaita (Fig. 2), where the two groups are defined as formations; the lower group, the Kwaimbaita Formation, is >2.7 km thick and the upper group, the Singgalo Formation, reaches a thickness of 750 m (Tejada *et al.* 2002). At Site 807, 1600 km to the north, the thickness of Singgalo-type flows is only 46 m. At Site 289, located between Site 807 and the Solomons, the single flow sampled at the top of basement is isotopically Kwaimbaita type. Isotopic data for glasses from the 640 m-thick basalt sequence drilled at Site 462A in the Nauru Basin to the NE of the OJP proper (Mahoney 1987; Castillo *et al.* 1994) show that they, too, are of the Kwaimbaita type. North of the OJP, 51 m of Kwaimbaita-type flows were drilled at Site 802 in the East Mariana Basin (Castillo *et al.* 1994). Both Singgalo and Kwaimbaita magma types appear to be the products of high amounts of partial melting; pre-Leg 192 estimates yielded values in the 20–30% range (assuming peridotite source rock), with the Kwaimbaita representing the upper end of this range (Mahoney *et al.* 1993; Neal *et al.* 1997; Tejada 1998).

Shipboard analysis of TiO_2, Zr and several other elements during Leg 192 suggested that basalt recovered from Sites 1183, 1186 and the lower 92 m of the 217 m-thick lava section at Site 1185 was of the Kwaimbaita type (Fig. 2), whereas biostratigraphic evidence indicated an early Aptian basement age (Shipboard Scientific Party 2001). The volcaniclastic deposits at Site 1184 also appeared chemically Kwaimbaita-like. No obvious Singgalo-type compositions were found at any of the sites. In contrast, the 136 m-thick basement section drilled at Site 1187 and the upper 125 m of flows at Site 1185, 146 km south of Site 1187, were discovered to be a low-TiO_2, high-MgO type of basalt not seen previously on the OJP. We term this magma type, the least differentiated of any found thus far, the Kroenke type, after the location of Site 1185 adjacent to Kroenke Canyon, a large submarine canyon just south of the site. Shipboard measurements showed it to be characterized by approximately 0.75 wt% TiO_2, 8–10 wt% MgO, *c.* 200 ppm Ni and *c.* 500 ppm Cr; in contrast, Kwaimbaita-type basalts average around 1 wt% TiO_2, 7 wt% MgO, and have <120 ppm Ni and <250 ppm Cr.

In this chapter we present Sr, Pb, Hf and Nd isotopic data for samples from Sites 1183, 1185, 1186 and 1187, and, in combination with previous data, discuss their implications for the source and origin of the OJP. Complementary to our work, Fitton & Godard (2004) have carried out a detailed major- and trace-element study of basalt from these sites, and mineral and glass compositions have been determined by Sano & Yamashita (2004). Elemental and isotopic data for the volcaniclastic rocks of Site 1184 are reported by Shafer *et al.* (2004) and White *et al.* (2004).

Analytical methods

During the cruise, we obtained small slabs from the least altered portions of crystalline basalt and, in some cases, glassy flow margins recovered from a representative number of basement units from each site (see Fig. 2) (Mahoney *et al.* 2001). Both massive and pillowed lava units were sampled, as was one of two vitric tuff beds lying immediately above basement at Site 1183. Preparation and analysis onshore followed our standard procedures for glass and for moderately altered bulk-rock samples (e.g. see descriptions and references of Castillo *et al.* 1994; Mahoney *et al.* 1998; Ingle *et al.* 2004). Basalts from Sites 1183 and 1186 were processed for Pb, Nd and Sr isotope analyses at the Scripps Institution of Oceanography, whereas those from Sites 1185 and 1187, and the tuff from Site 1183, were processed and analysed at the University of Hawaii. In both laboratories, parent and daughter element abundances were obtained on the same dissolution of sample analysed for isotope ratios. Concentration measurements at Scripps employed single-collector, high-resolution, inductively coupled plasma-mass spectrometry (ICP-MS), whereas at Hawaii parent and daughter concentrations were determined by isotope dilution using a multi-collector thermal-ionization mass spectrometer. Analysis of Hf isotopes was carried out at the Université Libre de Bruxelles using multi-collector ICP-MS (Nu Plasma). In addition to the Leg 192 samples, Hf isotopes also were measured for four samples of Singgalo Formation basalt and one sample of Kwaimbaita Formation basalt from Malaita previously analysed by Tejada *et al.* (2002) for Nd–Pb–Sr isotopes, and for major and trace elements. Concentrations of Lu and Hf for the Leg 192 samples were determined by high-resolution ICP-MS at the Université de Montpellier II on separate splits of sample from those used for isotope analysis (see Fitton & Goddard 2004). Our results are presented in Tables 1 and 2, along with estimated analytical uncertainties, isotopic fractionation corrections and standard values.

Results

Pb, Nd and Sr isotopes

An important, and unexpected, result of our study is that the high-MgO Kroenke-type basalt flows at Site 1187 and comprising the upper 125 m of basement units at Site 1185 are isotopically indistinguishable from previous data for the Kwaimbaita magma type (Figs 3 and 4). Recent ^{187}Re–^{187}Os dating indicates that basement at these sites and at Sites 1183 and 1186 is early Aptian, 121.7 ± 1.4 Ma (Parkinson *et al.* 2001), although ^{40}Ar–^{39}Ar and biostratigraphic data suggest a slight progression of ages within the Aptian (L. Chambers pers. comm. 2003). For age-correcting our isotopic data we have assumed an age (t) of 120 Ma. The Kroenke-type lavas have age-corrected $\varepsilon_{Nd}(t)$ = +6.1–+6.4, $(^{87}Sr/^{87}Sr)_t$ = 0.70374–0.70381, $(^{206}Pb/^{204}Pb)_t$ = 18.317–18.360, $(^{207}Pb/^{204}Pb)_t$ = 15.514–15.522 and $(^{208}Pb/^{204}Pb)_t$ = 38.169–38.220 (where t = 120 Ma). This range is within, or only slightly greater than, the propagated analytical errors for age-corrected Nd and Sr isotope ratios and double-spike $^{207}Pb/^{204}Pb$ (see the footnotes to Tables 1 and 2), which is remarkable considering the distance of 146 km between these two sites. Although relative variability in $(^{206}Pb/^{204}Pb)_t$ and $(^{208}Pb/^{204}Pb)_t$ is slightly greater, among the glasses it is only 18.320–18.340 and 38.180–38.194, respectively.

The age-corrected Pb, Nd and Sr isotope ratios for basalts from Sites 1183, 1186 and the lower 92 m of Site 1185 also fall within, or very close to, the small range measured for Kwaimbaita-type basalts from Malaita, Santa Isabel, Sites 289 and 807, and glasses from the Nauru and East Mariana basins. The Kwaimbaita-type nature of

Table 1. *Sr, Nd and Hf isotope data*

Sample			$(^{87}Sr/^{86}Sr)_t$	$(^{143}Nd/^{144}Nd)_t$	$\varepsilon_{Nd}(t)$	$(^{176}Hf/^{177}Hf)_t$	$\varepsilon_{Hf}(t)$	Rb	Sr	Sm	Nd	Lu	Hf
Site 1183A													
54–3 (28–34) tuff	L	Sg	0.70254	0.512714	+4.5			55.39	28.95	1.106	3.599		
54–4 (2–6) Unit 2B	L	Kw	0.70329	0.512812	+6.4	0.28298	+10.1	8.46	145.0	2.10	5.17	0.33	1.72
55–1 (77–78) Unit 3B	L	Kw	0.70347	0.512820	+6.5	0.28301	+11.0	0.91	104.1	0.84	1.83	0.37	1.83
55–3 (120–122) Unit 4B	L	Kw	0.70337	0.512786	+5.9	0.28301	+11.2	3.72	95.60	1.19	2.27	0.33	1.75
61–2 (6–7) Unit 6	L	Kw	0.70345	0.512789	+5.9	0.28300	+10.7	0.39	90.89	0.93	2.02	0.35	1.71
	R		0.70348										
	R		0.70347										
67–3 (93–95) Unit 8	L	Kw	0.70349	0.512802	+6.2	0.28301	+11.0	0.49	85.88	1.15	2.69	0.33	1.67
Site 1185B													
5–5 (142–143) Unit 2		Kr	0.70381	0.512799	+6.1	0.28301	+11.0	1.290	77.37	1.518	4.393	0.28	1.07
6–4 (95–96) Unit 5		Kr	0.70380	0.512803	+6.2	0.28307	+13.3	1.035	63.56	1.273	3.678	0.24	1.05
15–2 (113–114) Unit 9	L	Kr	0.70375	0.512813	+6.4	0.28304	+12.2	0.5998	82.15	1.485	4.333	0.28	1.10
17–3 (33–34) Unit 10		Kw	0.70352	0.512788	+5.9	0.28304	+12.1	0.8742	104.9	2.290	6.855	0.38	1.82
28–1 (56–57) Unit 12		Kw	0.70354	0.512785	+5.8	0.28303	+11.8	0.7935	104.1	2.163	6.424	0.33	1.62
Site 1186A													
32–2 (100–101) Unit 1	L	Kw	0.70343	0.512798	+6.1	0.28301	+11.2	3.70	124.0	1.82	3.84	0.36	1.79
34–1 (84–86) Unit 2B	L	Kw	0.70346	0.512801	+6.1	0.28301	+11.0	0.26	87.81	0.91	1.99	0.33	1.52
38–1 (38–40) Unit 4	L	Kw	0.70342	0.512820	+6.5	0.28301	+11.0	1.14	88.87	0.90	2.28	0.34	1.58
GRS-39–1 Unit 4	L	Kw	0.70345	0.512792	+6.0	0.28301	+11.1	0.19	71.69	0.95	2.06	0.34	1.58
Site 1187A													
3–4 (93–95) Unit 3	G	Kr	0.70376	0.512803	+6.2	0.28302	+11.4	1.198	83.47	1.556	4.571	0.29	1.17
7–7 (97–98) Unit 4	L	Kr	0.70377	0.512799	+6.1	0.28302	+11.6	0.4247	79.82	1.498	4.395	0.28	1.17
10–1 (124–126) Unit 6	G	Kr	0.70375	0.512807	+6.3	0.28300	+10.8	1.281	88.88	1.663	4.892	0.29	1.18
14–2 (4–6) Unit 9	G	Kr	0.70381	0.512799	+6.1	0.28301	+11.2	1.347	84.73	1.572	4.630	0.29	1.08
16–2 (87–89) Unit 12	G	Kr	0.70374	0.512802	+6.2			2.445	86.21	1.538	4.511		
Central Malaita													
SG-1		Sg	0.70405	0.512694	+4.1	0.28298	+10.0	1.277	107.8	1.689	4.607	0.43	2.50
SGB-17	L	Sg	0.70414	0.512690	+4.0	0.28299	+10.3	0.9675	93.10	1.540	3.425	0.38	2.47
SGB-18	L	Kw	0.70379	0.512783	+5.8	0.28305	+12.4	0.6878	112.5	1.439	3.253	0.32	2.00
ML-421	L	Sg	0.70416	0.512691	+4.1	0.28298	+10.1	0.0990	109.6	1.326	2.885	0.43	2.94
KF-4	L	Sg	0.70416	0.512687	+4.0	0.28298	+10.2	0.7995	81.30	1.558	3.673	0.47	3.02

Notes: All elemental concentrations are in ppm. Sg, Singgalo type; Kw, Kwaimbaita type; Kr, Kroenke type. G, glass; R, replicate analysis; L, strongly acid-leached powder for Sr and Nd isotope analysis, except for the two samples from Sites 1185B and 1187A, for which only Sr isotopes and Rb and Sr concentrations were measured on strongly leached powder. Isotopic fractionation corrections are $^{148}NdO/^{144}NdO = 0.242436$ ($^{148}Nd/^{144}Nd = 0.241572$), $^{86}Sr/^{88}Sr = 0.1194$ and $^{179}Hf/^{177}Hf = 0.7325$. Sm, Nd, Lu and Hf concentrations, and Nd isotope data for Central Malaitan samples are from Tejada *et al.* (2002). Nd and Sr isotope data are reported relative to $^{143}Nd/^{144}Nd = 0.511850$ for La Jolla Nd and $^{87}Sr/^{86}Sr = 0.71025$ for NBS 987 Sr. At the University of Hawaii, the total range measured for La Jolla Nd is ± 0.000008 (0.2 ε units); for NBS 987 it is ± 0.000020 over a 2 year-period; at Scripps, it is ± 0.000014 (0.3 e units) and 0.000018, respectively. At the Free University of Brussels, $^{176}Hf/^{177}Hf = 0.282160 \pm 0.000020$ (0.7 ε units) for the JMC-475 standard. Within-run errors on the isotopic data above are less than or equal to the external uncertainties on these standards. Estimated relative uncertainties on Sm and Nd concentrations are <0.2% and *c.*1%, respectively, at Hawaii and Scripps; for Sr the respective uncertainties are 0.5% and *c.*2%, and for Rb, 1% and *c.*2%. Total procedural blanks are negligible: for Hf, <20 pg; for Nd and Sr, <10 pg and <35 pg. $\varepsilon_{Nd} = 0$ today corresponds to $^{143}Nd/^{144}Nd = 0.51264$; for $^{147}Sm/^{144}Nd = 0.1967$, $\varepsilon_{Nd}(t) = 0$ corresponds to $^{143}Nd/^{144}Nd = 0.512486$ at 120 Ma. $\varepsilon_{Hf} = 0$ today corresponds to $^{176}Hf/^{177}Hf = 0.28277$; for $^{176}Lu/^{177}Hf = 0.0332$, $\varepsilon_{Hf}(t) = 0$ corresponds to 0.282696 at 120 Ma.

Table 2. *Pb isotope data*

Sample		$(^{206}Pb/^{204}Pb)_t$	$(^{207}Pb/^{204}Pb)_t$	$(^{208}Pb/^{204}Pb)_t$	Th	U	Pb	$(^{206}Pb/^{204}Pb)_0$	$(^{207}Pb/^{204}Pb)_0$	$(^{208}Pb/^{204}Pb)_0$
Site 1183A										
54–3 (28–34) tuff	Sg	18.400	15.556	38.436	0.2877	0.5119	1.1750	18.915	15.581	38.531
54–4 (2–6) Unit 2B	Kw	18.517	15.528	38.256	0.1701	0.0825	0.3122	18.829	15.543	38.467
55–1 (77–78) Unit 3B	Kw	18.495	15.514	38.286	0.1971	0.0778	0.6595	18.634	15.521	38.401
55–3 (120–122) Unit 4B	Kw	18.429	15.509	38.161	0.1788	0.0718	0.2562	18.759	15.525	38.431
61–2 (6–7) Unit 6	Kw	18.369	15.511	38.138	0.2175	0.0690	0.2463	18.700	15.528	38.479
67–3 (93–95) Unit 8	Kw	18.510	15.525	38.248	0.4968	0.1135	0.9453	18.651	15.532	38.451
Site 1185B										
5–5 (142–143) Unit 2	Kr	18.320	15.518	38.185	0.1526	0.0471	0.4057	18.456	15.525	38.329
6–4 (95–96) Unit 5	Kr	18.317	15.514	38.169	0.1416	0.0433	0.1012	18.823	15.539	38.712
15–2 (113–114) Unit 9	Kr	18.360	15.515	38.220	0.1496	0.0450	0.2470	18.574	15.537	38.453
17–3 (33–34) Unit 10	Kw	18.398	15.516	38.201	0.2411	0.0711	0.4350	18.590	15.525	38.414
28–1 (56–57) Unit 12	Kw	18.390	15.515	38.219	0.2292	0.0690	0.1674	18.879	15.539	38.751
Site 1186A										
32–2 (100–101) Unit 1	Kw	18.279	15.512	38.190	0.1936	0.1716	0.2278	19.173	15.556	38.520
34–1 (84–86) Unit 2B	Kw	18.465	15.509	38.240	0.1958	0.0685	0.4061	18.663	15.519	38.426
38–1 (38–40) Unit 4	Kw	18.427	15.521	38.244	0.2110	0.0686	0.6348	18.554	15.528	38.372
GRS-39–1 Unit 4	Kw	18.376	15.518	38.180	0.1658	0.0526	0.1496	18.792	15.538	38.609
Site 1187A										
3–4 (93–95) Unit 3 G	Kr	18.336	15.521	38.194	0.1564	0.0485	0.2402	18.573	15.533	38.445
7–7 (97–98) Unit 4	Kr	18.357	15.514	38.201	0.1376	0.0445	0.2494	18.567	15.524	38.413
10–1 (124–126) Unit 6 G	Kr	18.320	15.522	38.184	0.1978	0.0606	0.2595	18.595	15.540	38.478
14–2 (4–6) Unit 9 G	Kr	18.333	15.517	38.180	0.1634	0.0495	0.2427	18.573	15.529	38.439
16–2 (87–89) Unit 12 G	Kr	18.340	15.517	38.192	0.1551	0.0478	0.2402	18.574	15.528	38.441

Notes: G, glass; Sg, Singgalo type; Kw, Kwaimbaita type; Kr, Kroenke type. All elemental abundances are in ppm. For Scripps data, measured Pb isotopic ratios are corrected for fractionation using the NBS 981 standard values of Todt *et al.* (1996); the long-term errors measured for this standard are ±0.008 for $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$, and ± 0.024 for $^{208}Pb/^{204}Pb$. For Hawaii data, a double-spike method (Galer 1999) was employed; the total range on approximately 5 ng loads of NBS 981 Pb in the last 3 years is 230 ppm for each ratio, and mean ratios measured are 16.937, 15.492 and 36.710. For both Scripps and Hawaii data, the within-run errors on measurements above are less than or equal to the external uncertainties on the standard. Estimated uncertainties on concentrations are <2% on Th, ≈1% on U and ≈0.5% on Pb for the Hawaii isotope-dilution data, and <2% for these elements for the Scripps data. Total procedural blanks are negligible: <3 pg for Th, <5 pg for U and <30 pg and <60 pg for Pb at Hawaii and Scripps, respectively.

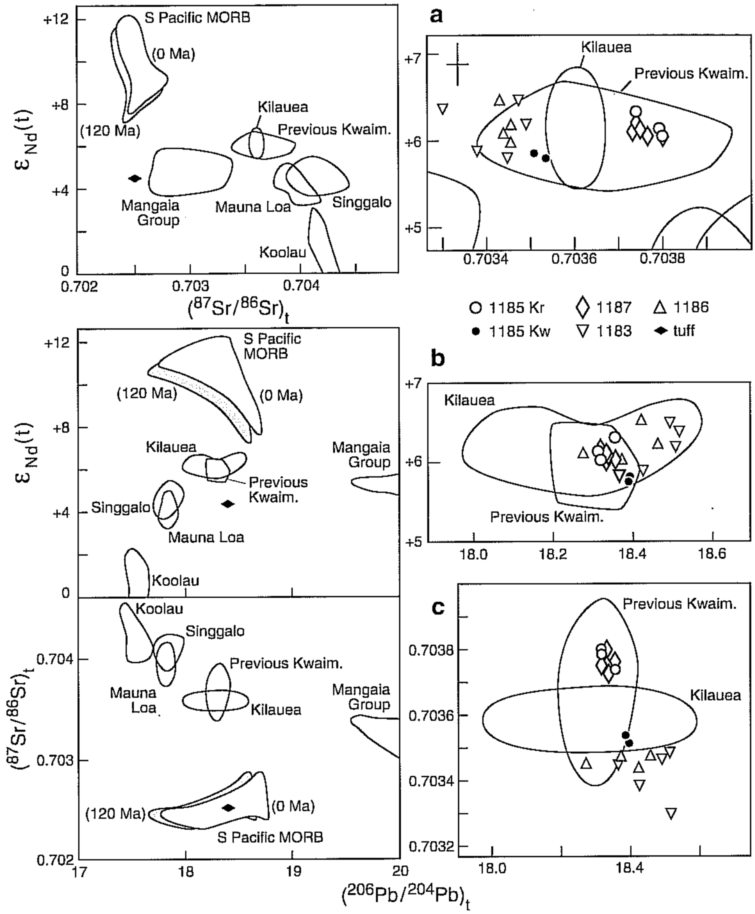

Fig. 3. Age-corrected (**a**) Sr v. Nd, (**b**), $^{206}Pb/^{204}Pb$ v. Nd and (**c**) $^{206}Pb/^{204}Pb$ v. Sr isotopic data for the Leg 192 lavas and tuff. Panels on the right are expanded portions of those on the left. Kw, Kwaimbaita type; Kr, Kroenke type. Fields are shown for previous Kwaimbaita- (Kwaim.) and Singgalo-type basalt from the pre-Leg 192 drill sites, Malaita and Santa Isabel, and glass from the Nauru and East Mariana basins (data sources: Mahoney 1987; Castillo *et al.* 1991, 1994; Mahoney & Spencer 1991; Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002). Pb isotope data for two previously analysed samples with suspected analytical problems were not used. See Tejada *et al.* (2002) for data sources used for the fields of South (S) Pacific MORB, Kilauea, Mauna Loa (subaerial portion), Koolau (subaerial portion) and the Mangaia Group islands. The shaded 120 Ma field is for estimated South Pacific MORB source mantle (see Tejada *et al.* 2002). Error bars in (**a**) are for data in this paper (see Fig. 4 for Pb error bars). Note that symbols are the same as in Figure 2.

these Leg 192 lavas is confirmed by the incompatible-element data of Fitton & Godard (2004). For the Site 1183 basalts, the isotopic range is $\varepsilon_{Nd}(t)$ = +5.9–+6.5, $(^{87}Sr/^{87}Sr)_t$ = 0.70329–0.70349, $(^{206}Pb/^{204}Pb)_t$ = 18.369–18.517, $(^{207}Pb/^{204}Pb)_t$ = 15.509–15.528 and $(^{208}Pb/^{204}Pb)_t$ = 38.138–38.286. Samples from Site 1186 have $\varepsilon_{Nd}(t)$ = +6.0–+6.5, $(^{87}Sr/^{87}Sr)_t$ = 0.70342–0.70346, $(^{206}Pb/^{204}Pb)_t$ = 18.279–18.465, $(^{207}Pb/^{204}Pb)_t$ = 15.509–15.521 and $(^{208}Pb/^{204}Pb)_t$ = 38.180–38.244. Values for the lower basalt units at Site 1185 are +5.8–+5.9; 0.70352–0.70354; 18.390–18.398; 15.515–15.516 and 38.201–38.219, respectively.

Although the range of age-corrected isotopic values for all the Leg 192 and other Kwaimbaita-type basalt flows is small, subtle variations are apparent between and within sites, particularly in Sr isotopes (Fig. 3a). The $(^{87}Sr/^{86}Sr)_t$ values of lavas from Sites 1183, 1186 and the lower portion of Site 1185, together with the Kwaimbaita-type units at Site 807 and Site 289, are less than 0.7035. In contrast, Kwaimbaita-type lavas in

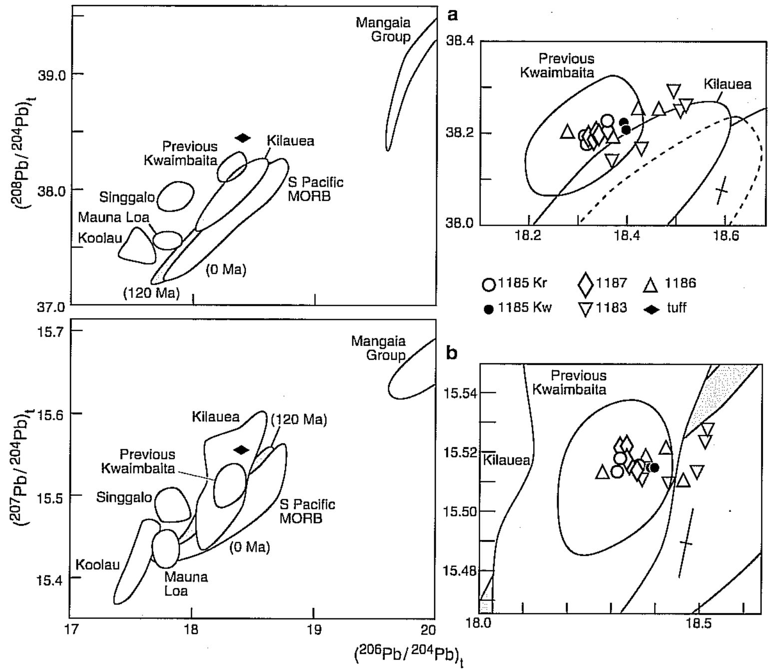

Fig. 4. (**a**) $(^{206}Pb/^{204}Pb)_t$ v. $(^{208}Pb/^{204}Pb)_t$ and (**b**) $(^{207}Pb/^{204}Pb)_t$ for the Leg 192 samples. Kw, Kwaimbaita type; Kr, Kroenke type. Panels on the right are expanded portions of those on the left. Fields for previous Kwaimbaita- and Singgalo-type basalt are for the pre-Leg 192 drill sites, Malaita and Santa Isabel, and glass from the Nauru and East Mariana basins. Note that for some of the previous data and for the Site 1183 and 1186 results, some of the range in age-corrected Pb isotope values probably reflects the determination of parent–daughter ratios by methods other than isotope dilution. In addition, for some previous samples and several in our data set, variable alteration of U/Pb ratios, in particular, has probably caused over- or under-corrections in initial Pb isotope ratios. See Figure 3 and Tejada *et al.* (2002) for data sources. Error bars are for measured values of Site 1183 and 1186 samples.

Malaita and Santa Isabel, the Site 803 basalts, and the Kroenke-type flows at Sites 1185 and 1187 have $(^{87}Sr/^{86}Sr)_t$ between 0.7036 and 0.7039. The Nauru Basin and East Mariana Basin data set straddles both ranges. Some of the higher values for these basins and the plateau proper no doubt reflect the effects of sea-water alteration not removed by acid-leaching, but values as high as 0.7038 are found in fresh glasses, indicating a small amount of real, pre-eruptive variability in Sr isotopes. The lowest value, 0.70329, is for sample 1183A-54-4 (2–6). This sample has a much higher Rb concentration (8.47 ppm) and $^{87}Rb/^{86}Sr$ value (0.169) than unaltered or slightly altered Kwaimbaita-type basalts (generally <2 ppm and <0.06, respectively), suggesting that alteration- or leaching-related disturbance of the Rb–Sr system may have caused its age-adjusted Sr isotope ratio to be over-corrected slightly. Three Site 1183 samples and one Site 1186 sample have slightly higher age-corrected $(^{206}Pb/^{204}Pb)_t$ (by up to 0.1) than seen previously, although their $(^{207}Pb/^{204}Pb)_t$ and $(^{208}Pb/^{204}Pb)_t$ values are within the previous range (Fig. 4). The reason for these small differences is not clear. They could represent pre-eruptive variability; however, disturbance of U/Pb ratios and, for these particular samples (and some of the pre-Leg 192 samples), measurement of U and Pb concentrations by methods other than isotope dilution probably account for a significant portion of the range in age-corrected $(^{206}Pb/^{204}Pb)_t$ values. Mis-corrections also may result when different splits of sample are used for concentration and Pb isotope measurements, as was the case for several pre-Leg 192 samples.

The only evidence of Singgalo-type magmatism at any of the Leg 192 sites is provided by the vitric tuff of Site 1183. The Singgalo magma type is characterized by higher $(^{87}Sr/^{86}Sr)_t$ and lower $\varepsilon_{Nd}(t)$ and $(^{206}Pb/^{204}Pb)_t$ than the Kwaimbaita type (Figs 3 and 4), and by slight relative enrichment in highly incompatible elements. The tuff's $\varepsilon_{Nd}(t)$ = +4.5, which is well within the Singgalo range. Although it is possible that the tuff could represent a volcanic source unrelated to the OJP, alteration-resistant incompatible elements also indicate that it belongs to the Singgalo magma type (Fitton & Godard 2004). However, its age-corrected $(^{87}Sr/^{87}Sr)_t$ is only 0.70254, much lower than values for either Singgalo or Kwaimbaita types. Likewise, its $(^{206}Pb/^{204}Pb)_t$ ratio (18.400) is well above the Singgalo range, and its $(^{207}Pb/^{204}Pb)_t$ (15.556) and $(^{208}Pb/^{204}Pb)_t$ (38.436) values are higher than for Singgalo- or Kwaimbaita-type basalts. The tuff is much more altered than any of the basalts, and these isotopic differences are probably a combination of the high level of alteration coupled with disturbance of the Rb–Sr system, and possibly the U–Pb and Th–Pb systems, by acid leaching during sample preparation. For example, relatively small changes in the high $^{87}Rb/^{86}Sr$ ratio (5.535) caused by leaching would lead to large under- or over- (as appears likely in this case) corrections in the calculated initial Sr isotope ratio of this sample.

Hf isotopes

Hafnium isotope ratios for all the Leg 192 basalts and the Kwaimbaita Formation sample from Malaita (SGB-18) fall within a limited range between $\varepsilon_{Hf}(t)$ = +10.1 and +13.3; however, values for all but two Leg 192 samples are within a much narrower range, from +10.7 to +12.2 (Table 1). As with Nd, Pb and Sr isotopes, the Kroenke-type lavas are indistinguishable from the Kwaimbaita type in their Hf isotope characteristics. The Singgalo Formation samples from Malaita possess slightly lower values of $\varepsilon_{Hf}(t)$, from +10.0 to +10.3.

The combined Hf–Nd isotope results place the OJP lavas within the field of OIB in Figure 5. However, relative to their $\varepsilon_{Nd}(t)$ values, the $\varepsilon_{Hf}(t)$ values of the Singgalo Formation samples are slightly higher (by approximately 1–2 ε units) than the corresponding values on a regression line fitted through the global Nd–Hf isotope array for oceanic basalts (Vervoort *et al.* 1999). Thus, the combined OJP data form an array with a slightly shallower slope than the mean slope for OIB globally, and in this respect, as in Figure 3, again appear broadly similar to the Kilauea and Mauna Loa shield volcanoes of Hawaii.

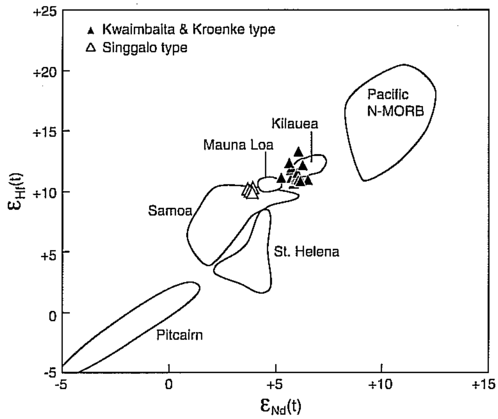

Fig. 5. $\varepsilon_{Hf}(t)$ v. $\varepsilon_{Nd}(t)$ for Ontong Java basalts. Fields for some Pacific and Atlantic oceanic island volcanoes and for high-precision data for Pacific N-MORB are shown for comparison. Data sources are Nowell *et al.* (1998), Salters & White (1998) and Chauvel & Blichert-Toft (2001) for the MORB field; Stracke *et al.* (1999), Blichert-Toft *et al.* (1999) and references therein for the Kilauea and Mauna Loa fields; Patchett & Tatsumoto (1980) and White & Hofmann (1982) for Samoa; Salters & White (1998) for St Helena; and Eisele *et al.* (2002) for Pitcairn. Note that, as in Figure 3, the Kwaimbaita- and Kroenke-type rocks are similar in their Hf–Nd isotope compositions to values reported for Kilauea, Hawaii, whereas the Singgalo-type compositions are closer to those of Mauna Loa, Hawaii.

Discussion

Was the OJP derived from a huge (almost) primitive mantle reservoir?

The small isotopic range defined by lavas from the OJP and surrounding basins is remarkable, given the great distances between sampling locations. In previous studies it was not clear to what extent the range for each magma type represents homogenization of more variable primary magma compositions by efficient mixing in extensive open-system magma chambers, or reflects an enormous, isotopically near-uniform mantle source (Tejada *et al.* 1996, 2002; Neal *et al.* 1997). The Singgalo magma type, although widespread and of considerable thickness in some OJP basement sections, now appears to have been volumetrically very minor relative to the Kwaimbaita type, at least during formation of the upper levels of basement crust. In considering the origin of the OJP, we therefore focus here on the Kwaimbaita mantle source.

To what extent was the Kwaimbaita source homogeneous? The Kroenke-type basalt, having lost only olivine by crystal fractionation (Fitton & Godard 2004; Sano & Yamashita 2004), is much closer to a primary magma composition than the Kwaimbaita type, which has lost substantial amounts of olivine, augite and plagioclase (e.g. Neal *et al.* 1997). Thus, the Kroenke-type isotopic signature cannot be a product of homogenization via open-system magmatic differentiation to any significant extent. That it is identical to the Kwaimbaita isotopic signature indicates that both groups of basalt were derived from the same mantle source and, although magma mixing may have played some role in damping isotopic variability, this source must indeed have been isotopically quite homogeneous relative to the scale of melting. The small amount of non-alteration-related site-to-site variability observed probably reflects small-scale heterogeneity present in the source region that, although largely averaged out at the high extents of partial melting involved, was not distributed perfectly uniformly.

Fitton & Godard (2004) show that Kwaimbaita-type chemical compositions can be derived by fractionation of olivine, plagioclase and augite from Kroenke-type magmas. For the Kroenke magma type, these authors estimate the percentage of partial melting to be approximately 30%. This value, which is at the high end of previous estimates for the Kwaimbaita magma type, is based on Zr content in primary Kroenke-type magma, the composition of which is calculated by incremental addition of equilibrium olivine to a Kroenke-type glass composition. In good agreement, a nearly identical value is obtained by Herzberg (2004) using a completely different method that relies on phase equilibria and major elements.

What might be the explanation for the particular combination of isotopic and incompatible-element characteristics in this mantle source? Specifically, the flat chondrite-normalized rare-earth patterns and, for all but the most incompatible elements, flat primitive-mantle-normalized element patterns of the Kwaimbaita- and Kroenke-type basalts (see Fitton & Godard 2004) point to a mantle source not too different from estimated primitive mantle in most of its inter-element ratios. However, the observed isotopic values (e.g. $\varepsilon_{Nd}(t)$ *c.* +6) are clearly far removed from those estimated for primitive mantle ($\varepsilon_{Nd} = 0$). Qualitatively similar characteristics are seen in some basalts from some other oceanic plateaus, but in those plateaus the range of isotopic and chemical variation is substantially greater than found for the OJP; the general explanation given is in terms of variable mixing of OIB mantle end-member types, with or without involvement of MORB-type mantle (e.g. Kerr *et al.* 2002). Contrary to early predictions (Mahoney 1987), no evidence for involvement of MORB-type mantle has yet been found in the OJP, including the Leg 192 basalts. Although mixing involving, for example, an EM-1-like (low $^{206}Pb/^{204}Pb$) end-member and anciently recycled oceanic lithosphere can explain the observed OJP isotopic ratios, it is not particularly supported by the lack of any discernible mixing arrays in the data, or by trace- or major-element modelling (Tejada *et al.* 2002).

An alternative possibility is that the Kwaimbaita source represents originally primitive mantle (i.e. originally of bulk silicate earth composition) that underwent minor fractionation long ago, after which it evolved isotopically in an essentially closed-system manner until the early Aptian. Indeed, simple two-stage evolution models assuming a fractionation event in the 3–4 Ga range can reproduce the Sr, Pb, Hf and Nd isotopic characteristics of the Kwaimbaita- and Kroenke-type basalts. An example is summarized in Table 3 and illustrated in Figure 6. In this case, the first stage of isotopic evolution occurs in a reservoir of primitive-mantle chemical and isotopic composition until about 3 Ga, when a fractionation event occurs that modifies the reservoir's chemical composition slightly. Subsequent closed-system isotopic evolution transpires until 120 Ma in this slightly modified mantle reservoir, which possesses parent–daughter ratios similar to those measured in the OJP basalts (i.e. we assume that because the OJP basalts represent high-degree partial melts, their parent–daughter ratios are not too different from those of their source). The *c.* 3 Ga fractionation event in this model is assumed to occur by removal of a small (1%) partial melt under upper-mantle conditions, leaving residual mantle (the future Kwaimbaita/Kroenke source) with the same normative phase proportions as those estimated for the Kwaimbaita-type source by Neal *et al.* (1997). In general agreement, Fitton & Godard (2004) show that the incompatible-element patterns of Kroenke- and Kwaimbaita-type basalts can be explained quite well by past removal of a similarly small amount of melt from primitive mantle. Such a modified primitive-mantle source also would be predicted to be depleted in volatiles and to be slightly enriched in highly compatible elements. Few data on volatiles in OJP basalt are available as yet, but those that exist show that water contents are indeed low (MORB-like or lower; Michael 1999; Roberge *et al.* 2004). Likewise, data for platinum-group elements suggest a source slightly enriched in highly siderophilic elements relative to most estimates of primitive-mantle composition (Chazey & Neal 2004).

Arguably, the most likely part of the planet in which a large volume of ancient, chemically near-primitive material might survive would be the lower mantle. (Note that the model

Table 3. *Parameters and results for two-stage model evolution of Kwaimbaita-type mantle*

	Start, stage 1	Start, stage 2	OJP eruption
Time (Ma)	4450	3050	120
Nd isotopes			
$^{147}Sm/^{144}Nd$	0.1967	0.2126	0.2126
OJP source $^{143}Nd/^{144}Nd$	0.506829	0.508675	0.512791
$\varepsilon_{Nd}(t)$ for OJP source	0.0	0.0	+6.0
Hf isotopes			
$^{176}Lu/^{176}Hf$	0.0332	0.0388	0.0388
OJP source $^{176}Hf/^{177}Hf$	0.279795	0.280759	0.283022
Sr isotopes			
$^{87}Rb/^{86}Sr$	0.0850	0.0538	0.0538
OJP source $^{87}Sr/^{86}Sr$	0.69956	0.70134	0.70363
Pb isotopes			
$^{238}U/^{204}Pb$	9	9.5	9.5
$^{232}Th/^{204}Pb$	36	36.575	36.575
OJP source $^{206}Pb/^{204}Pb$	9.311	12.792	18.361
OJP source $^{207}Pb/^{204}Pb$	10.301	14.210	15.522
OJP source $^{208}Pb/^{204}Pb$	29.476	32.478	38.218

Notes: Stage 1 source is estimated primitive mantle, and evolution to the beginning of Stage 2 is closed-system. Changes in parent–daughter ratios at 3.05 Ga (the start of Stage 2) are assumed to result from partial melting. Stage 2 evolution is also closed-system. Partial melting event at 3.05 Ga involves 1% batch melting, leaving behind a residue with normative phase proportions inferred for the OJP source (0.6 ol: 0.2 opx: 0.1 cpx: 0.1 gt; Neal *et al.* 1997) (ol, olivine; opx, orthopyroxene; cpx, clinopyroxene; gt, garnet). Phase proportions assumed entering the melt are 0.15 ol: 0.3 opx: 0.25 cpx: 0.3 gt (note that a sulphide-bearing residue would change the above Pb isotope results somewhat). Solid–liquid distribution coefficients used are from Kennedy *et al.* (1993) and the compilation of Green (1994). Starting isotope compositions are determined from meteorite (for Nd, Jacobsen & Wasserburg 1980; for Hf, Blichert-Toft & Albarède 1997; for Pb, Chen & Wasserburg 1983) and estimated original bulk earth (for Sr; McCulloch 1994) values.

calculations in Table 3 do not specify how such material would come to reside in the lower mantle, and the assumption of *c.* 3 Ga melting at upper-mantle depths is made simply because very little is known of solid–liquid distribution coefficients for lower-mantle minerals and conditions.) Recent seismological results, indeed, suggest the bottom *c.* 1000 km of today's mantle is chemically distinct, perhaps relatively primitive, separated from the rest of the mantle by a thermochemical interface with large undulations, and that most of the time little mixing occurs across the interface (e.g. Kellogg *et al.* 1999). Hypotheses involving an ultimately lower-mantle origin for the plateau are, of course, inevitably coupled to plume-head or mantle-overturn models.

Results of recent models of plume formation in the lower mantle are broadly consistent with the isotopic homogeneity exhibited by the voluminous Kwaimbaita (and Kroenke) magma type in that they suggest large plume heads should be well mixed, should entrain little non-plume mantle during their ascent and thus should be significantly more homogeneous than plume tails (e.g. Van Keken 1997; Farnetani *et al.* 2002). More generally, the OIB-like isotopic signature and the apparently rapid formation of the bulk of the OJP by high-degree partial melting are consistent with predictions of plume-head (plume-impact) models (e.g. Richards *et al.* 1989; Campbell & Griffiths 1990; Saunders *et al.* 1992); thus, this sort of model has been the type explored most commonly in previous attempts to understand the origin of the plateau (Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002; Babbs 1997; Gladczenko *et al.* 1997; Neal *et al.* 1997).

However, these studies also noted significant discrepancies between observation and model. Tejada *et al.* (2002) recently emphasized several of the most important discrepancies, which include the following: (1) after an apparent eruptive hiatus of *c.* 30 Ma, the puzzling *c.* 90 Ma volcanic episode, which in several widespread locations produced tholeiitic basalts with

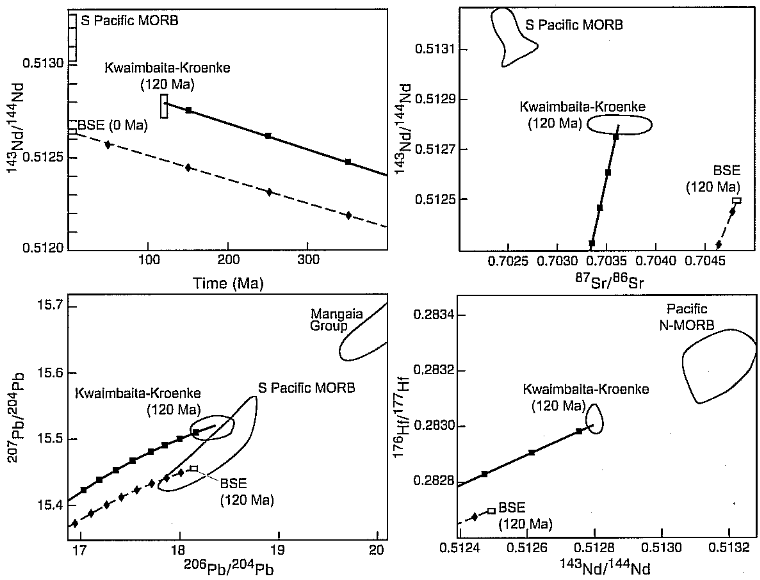

Fig. 6. Upper left panel: evolution of $^{143}Nd/^{144}Nd$ during the last several hundred million years in the two-stage model of Table 3. Other panels illustrate model evolution in Nd v. Sr isotope, $(^{206}Pb/^{204}Pb)_t$ v. $(^{207}Pb/^{204}Pb)_t$ and Nd v. Hf isotope space. Squares and diamonds on lines indicate successive 100 Ma increments since a fractionation event at 3.05 Ga. BSE is model bulk silicate Earth or primitive mantle.

isotopic and incompatible-element compositions closely similar to those of the *c.* 122 Ma Kwaimbaita-type magmas; (2) the lack of a post-plateau seamount chain corresponding to the plume-tail stage of hot-spot development theorized to follow the plume-head stage; (3) the lack of any presently active hot spot that can be linked unambiguously to the plateau; (4) post-eruptive subsidence of the plateau appears to have been much less than expected for oceanic lithosphere; and (5) all OJP lavas sampled thus far were erupted well below sea level, yet the standard form of the plume-head model (e.g. Richards *et al.* 1989; Campbell & Griffiths 1990) predicts that much of the surface should originally have been shallow or subaerial.

Some of these discrepancies may provisionally be accommodated by assuming a number of case-specific modifications to the standard plume-head model. (1) Following Leg 192, it now appears that the *c.* 90 Ma event was volumetrically minor (Shipboard Scientific Party 2001; L. Chambers pers. comm. 2003). It could represent plume-tail-related volcanism, assuming that the plateau did not drift much between 122 and 90 Ma (Neal *et al.* 1997), but additional assumptions are required to account for the apparent lack of any eruptive activity between 122 and 90 Ma. (2) The absence of a post-plateau chain of seamounts perhaps may be explained by appealing to a 'mid-mantle' plume of the type modelled by Cserepes & Yuen (2000). Alternatively, it can be explained by assuming a hypothetical jump or migration of a spreading centre near the OJP after the 90 Ma event, placing a different plate, now largely subducted, above the hot spot (Neal *et al.* 1997). Unfortunately, the precise relationship of the OJP to spreading centres in its vicinity at approximately 90 Ma is still largely conjectural (Gladczenko *et al.* 1997; Kroenke *et al.* 2004). (3) That the OJP cannot be linked clearly to a present-day hot spot could mean that the plume that produced the OJP no longer exists (e.g. Neal *et al.* 1997); alternatively, the problem could arise from large cumulative uncertainties in the amount of 120–0 Ma plume motion and true polar wander (Antretter *et al.* 2004). (4) Large amounts of intrusion and underplating during the 122–90 Ma period, and perhaps subsequently, have been postulated to explain the anomalous subsidence record of the plateau (Ito & Clift 1998; Ito & Taira 2000), but persistent lithospheric stress conditions that would prevent any significant accompanying volcanism during this *c.* 30 Ma period also must be assumed. (5) The lack of a large area initially at shallow depths may partly be explained by appealing to an eclogite-rich (Tejada *et al.* 2002) or, to a lesser extent, a volatile-rich plume head. Both would allow large-scale melting at lower plume temperatures than required for a dry or purely peridotitic plume head, and thus might lead to significantly less lithospheric uplift. Both seem unlikely, however, as OJP glasses point to a volatile-poor source (Michael 1999; Roberge *et al.* 2004) and chemical characteristics of OJP lavas are not matched well by assuming an eclogite-rich head, which additionally would require truly enormous amounts of eclogite to be concentrated in one deep-mantle location and then entrained within the plume head (Tejada *et al.* 2002). The amount of eclogite that plumes can carry is debated, but is likely to be rather small, in general (e.g. Gibson 2002). In any case, the high density of eclogite implies that eclogite-rich plume heads would have to be hotter, not cooler, than purely peridotitic ones in order to be sufficiently buoyant to rise through the mantle. For the OJP, Fitton & Godard (2004) point out that to produce the high-MgO Kroenke magma type from an eclogite-rich source would require approximately 100% fusion of the eclogite component and thus very high potential temperatures.

Plate separation and ridge-centred hot-spot hypotheses for the origin of the OJP

Proposed alternatives to the plume-head model also are problematic in the case of the OJP. The plate separation hypothesis (e.g. Anderson *et al.* 1992; Smith & Lewis 1999) posits an extensive layer of shallow, volatile-rich, near-solidus, OIB-like (but not plume-derived) asthenosphere ('perisphere') to have been residing beneath a region of the Pacific lithosphere that was rifted suddenly by a ridge jump around 122 Ma, causing cataclysmic melting. However, the pre-122 Ma seafloor within several hundred kilometres of the plateau is not much older than the plateau itself, having been formed only *c.* 2–35 Ma earlier (e.g. Taylor 1978; Nakanishi *et al.* 1992); thus, during this period, a spreading system was not too distant from the (future) location of the OJP. It is difficult to understand why such perisphere, assuming it existed, was not drained earlier by the nearby ridge. Also, the hypothesis predicts that normal MORB-type mantle lying just beneath the perisphere should have been tapped progressively more as OJP volcanism proceeded; yet, as noted above, no evidence of MORB-type mantle has been found thus far in OJP basalts. Indeed, the topmost part of the lava pile in several widespread locations is composed of the Singgalo-type basalts, which are even less MORB-like than the Kwaimbaita and Kroenke types. Also, as noted above, present indications are that the OJP's source mantle was poor in volatiles.

Similarly, the 'Icelandic' hypothesis that the OJP formed gradually, over several tens of millions of years, as the product of a large, approximately steady-state, ridge-centred plume (e.g. Mahoney & Spencer 1991; Ito & Clift 1998) is not supported by the isotopic and elemental homogeneity of OJP basalts over great distances, by the extensive Kwaimbaita-type basalts filling the Nauru and West Mariana basins, or by evidence suggesting that much of the plateau itself may have formed in an off-axis position (Coffin & Gahagan 1995). Also, in contrast to earlier plate-motion models, recent modelling suggests that the OJP drifted WNW as much as 2000 km in the 122–90 Ma period (Kroenke *et al.* 2004), implying that it could not have been situated above one hot spot the whole time. Furthermore, it now appears that the great bulk of the OJP formed rapidly in the Aptian (e.g. Tejada *et al.* 1996, 2002; Parkinson *et al.* 2001; L. Chambers pers. comm. 2003). A super-fast spreading rate would shorten the time needed for a ridge-centred hot spot to form the plateau, but the Aptian spreading rate in the vicinity of the OJP is unknown. However, for the Pacific–Phoenix ridge segments east of the plateau, a super-fast (*c.* 150 km Ma^{-1}) Barremian–late Hauterivian (*c.* 122–129 Ma) rate is indeed indicated by magnetic lineations M0–M7 (Larson 1997). What we can say at present is that the lack of any evidence for a long period of significant constructional volcanism appears to rule out the hypothesis as originally presented.

Meteorite impact instead of plume impact?

Although the isotopic and elemental characteristics of OJP lavas can be accommodated relatively well by plume-head models, we are impressed by the number of characteristics that are not explained satisfactorily by such models without the *ad hoc* postulation of special conditions. In the light of presently available evidence, we revisit a proposed alternative mechanism for the formation of oceanic plateaus by meteorite impact (Rogers 1982).

Noting that impact sites should be more numerous in the ocean than on land, Rogers (1982) proposed that the major Pacific plateaus represent massive outpourings of basalt formed by the cataclysmic excavation of asthenosphere by large, but rare, oceanic impacts. This hypothesis is attractive in that it can explain, without the special pleading required in the plume-head model, the absence of a post-plateau seamount chain and any obvious present-day hot spot that can be linked with the formation of the OJP (the same applies to several of the other Pacific plateaus). Also, the apparent lack of large areas at shallow-water depths during the construction of the OJP is not necessarily a problem, because inherently hotter-than-normal mantle is not required for widespread magmatism. Nor are huge amounts of eclogite- or volatile-rich mantle necessary. Post-volcanic subsidence likewise might be less than otherwise expected (although we question whether the pervasive serpentinization of shallow mantle suggested by Rogers (1982) is a viable mechanism for a feature the size of the OJP). Moreover, the limited range of elemental and isotopic variation among both the Singgalo- and Kwaimbaita-type basalts might be attributable in part to melt homogenization in extensive magma pools created by the impact. For additional discussion of the potential advantages of the impact hypothesis in explaining the OJP, we refer the reader to Ingle & Coffin (2004).

However, just as simple versions of the plume-head and other models for the OJP have significant shortcomings, the same appears to be true of the asteroid (or comet) impact hypothesis. Large-volume, high-degree partial melting of the upper few hundred kilometres of subseafloor mantle resulting from a large impact would ordinarily be expected to produce basalt with essentially N-MORB-type isotopic and incompatible-element characteristics. Although sampled in relatively few places, pre-OJP Pacific MORB are isotopically and chemically indistinguishable from modern Pacific MORB (Janney & Castillo 1997; Mahoney *et al.* 1998, J. Mahoney unpublished data). In contrast, the OJP is characterized by enormous volumes of basalt with OIB-like isotopic signatures and rather flat incompatible-element patterns, and there is as yet no evidence for any involvement of MORB-type mantle in the plateau (see above, and Tejada *et al.* 2002). Note also that neither chondritic, achondritic nor iron meteorites have a suitable combination of Sr–Pb–Nd isotopic characteristics (e.g. Kerridge & Matthews 1988 and references therein) to explain the isotopic signature of the OJP basalts via contamination of MORB-type mantle with meteoritic material. (Moreover, the isotopic and incompatible element contribution of an impacting body would be quite small in most plateau magmas, as most of the body's mass would be expelled in the impact ejecta, the volume of the object would be miniscule compared to that of the OJP, and meteoritic concentrations of most of these elements are low (e.g. Wolf *et al.* 1980; Schuraytze *et al.* 1996).)

Of course, just as with plume-head models, it is possible to appeal to special circumstances. For example, the impact site might have been above a geochemically anomalous region of asthenosphere containing a substantial amount of OIB-type mantle. Suitable isotopic compositions are notably rare among modern South Pacific hot spots and non-hot-spot volcanic areas, but it is conceivable that an impact fortuitously occurred near an area dominated by such material (cf. Ingle & Coffin 2004). Alternatively, as some authors have speculated, a large impact may actually trigger a deep-mantle plume beneath the impact site (e.g. Alt *et al.* 1988; Glikson 1999). In such a 'hybrid' scenario, it is not clear whether significant initial uplift of a plateau's surface would result or not.

Another potential difficulty for the impact hypothesis includes the reportedly systematic patterns in the gravity field and bathymetry of the OJP that Winterer & Nakanishi (1995), Neal *et al.* (1997) and Kroenke *et al.* (2004) have suggested represent a seafloor-spreading fabric. This interpretation remains to be evaluated rigorously but, if correct, is very difficult to reconcile with the expected widespread destruction and disruption of pre-existing oceanic lithosphere by a large impact, or with the short-lived outpouring of magma following an impact, which would be too rapid to allow formation of significant amounts of new seafloor.

To our knowledge, key features diagnostic of other large impact events, such as microspherules and siderophile element anomalies, have not been found in terrestrial or marine sediments around the Barremian–Aptian boundary. In contrast to the impact-linked Cretaceous–Tertiary boundary, no mass extinction occurred at the time of OJP formation. The statistical likelihood of an impact large enough to form the OJP can be estimated from cratering rates (e.g. Glikson 1999), but significant extrapolation is required. It is not clear that a sufficiently large object has been available in the Earth's vicinity in the last few hundred million years. The largest known Phanerozoic impact crater, the 65 Ma, approximately 200–250 km-wide crater at Chicxulub, Mexico, is thought to have been created by an object about 10 km in diameter (Grieve & Therriault 2000 and references therein). Although a large impact in relatively young, thin oceanic lithosphere would have different consequences than one on a continent (Rogers 1982; Glikson 1999; Jones *et al.* 2002), any object capable of creating the OJP must have been several times larger. However, no near-earth objects of such size are observed today (Binzel *et al.* 2002). Moreover, Venus, with a surface age estimated at approximately 600 Ma (e.g. Nimmo & McKenzie 1998), lacks any impact craters larger than Chicxulub (Schaber *et al.* 1992), whereas all lunar craters larger than about 100 km in diameter appear to be older than approximately 800 Ma (e.g. Eberhardt *et al.* 1973; Grier *et al.* 2001). Nevertheless, despite these potentially serious problems with the impact hypothesis, we regard it as deserving of further study in view of the difficulties encountered with any simple form of the plume-head or other proposed models for the OJP.

Conclusions

Age-corrected Nd, Pb and Sr isotope ratios of early Aptian basalt flows cored in four widely separated sites on the OJP during Leg 192, including the primitive Kroenke type, display only a small range of variation (e.g. $\varepsilon_{Nd}(t)$ = +5.8–+6.5, $(^{206}Pb/^{204}Pb)_t$ = 18.28–18.52, $(^{87}Sr/^{87}Sr)_t$ = 0.7033–0.7038). Moreover, all of the values fall within, or very near, the small field defined by the Kwaimbaita basalt type of the eastern Solomon Islands, previous OJP basement drill sites, and the adjacent Nauru and East Mariana basins. Age-corrected Hf isotope ratios display a correspondingly small range of variation ($\varepsilon_{Hf}(t)$ = +10.7–+12.2 for all but two samples), and the combined data indicate the Kwaimbaita/Kroenke-type mantle source was both immense and quite homogeneous relative to the scale of melting. Among the Leg 192 sites evidence for the Singgalo magma type, which forms lava piles lying stratigraphically above sections of Kwaimbaita-type basalt in the Solomon Islands and at Leg 130 Site 807, is confined to a thin interval of vitric tuff atop basement at Site 1183. It now seems apparent that the Singgalo magma type was a relatively minor component in the upper portions of crustal basement over much of the high plateau.

The isotopic characteristics of the Kwaimbaita/Kroenke mantle source can be modelled by simple two-stage evolution involving originally primitive mantle that underwent minor fractionation in the 3–4 Ga period, followed by closed-system radiogenic ingrowth until being tapped by plateau magmatism in the early Aptian. In the context of a plume-head model, such a source is compatible with recent geophysical evidence for a compositionally distinct lower-mantle layer, assuming this layer consists of chemically slightly modified primitive mantle. Although such a model can account for the isotopic and chemical characteristics of the OJP basalts and – with a large enough plume head – the sheer size of the plateau and the rapid formation of most of it in the early Aptian, a number of other first-order features require the assumption of *ad hoc* adjustments to the plume-head model. At least some of these appear unlikely or untenable on the basis of presently available evidence. However, alternative hypotheses, including an origin by asteroid impact, formation by plate separation, and gradual formation above a ridge-centred plume tail, also appear inadequate or require significant *ad hoc* modifications.

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