

Large igneous province magma petrogenesis from source to surface: platinum-group element evidence from Ontong Java Plateau basalts recovered during ODP Legs 130 and 192

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Abstract: A total of 16 Ontong Java Plateau (OJP) basalt samples from Ocean Drilling Program Legs 192 and 130 were analysed for major, trace and platinum-group elements (PGEs; Ir, Ru, Rh, Pt and Pd). Major- and trace-element compositions determined by our study confirm Leg 192 shipboard analyses that indicated a new group of more primitive or 'Kroenke-type' basalts, with higher MgO, Ni and Cr, and lower incompatible-element abundances than the more common Kwaimbaita-type basalts. The PGE abundances quantified here extend the range of the continuum of compositions found in previously analysed OJP basalts and are similar to those present in some komatiites. The PGEs, therefore, cannot be used to differentiate definitively between OJP basalt groups. The two samples analysed from Leg 130 (one from Site 803 and one from Site 807) are akin to the Kwaimbaita-type basalts.

Low-temperature alteration has not affected Pd abundances in the Leg 192 basalts as it has in the Solomon Island and the Leg 130 samples. Elemental abundances and ratios along with petrography reveal that the OJP basalts have not experienced sulphide saturation. Positive correlations of Ir and Ru with Cr and Ni attest to the lithophile behaviour of the PGEs and lend more credence to studies suggesting compatibility of these elements in oxide and silicate phases, such as Cr-spinel and olivine. Estimates of sulphur abundance in the mantle, degree of partial melting and pressure of melt initiation were used in conjunction with the model of Mavrogenes & O'Neill to calculate a minimum initial excess temperature of +185–+235°C (1465–1515°C at 3.5–4.0 GPa) above ambient mantle for the OJP source. This is in broad agreement with a fossil geotherm preserved in megacrysts and peridotite xenoliths found in pipe-like intrusives of alnöite that outcrop on the island of Malaita, Solomon Islands. Using the PGEs as a guide, the OJP basalts were modelled using a three-source component melt mix: a 10% garnet peridotite melt of primitive mantle composition, which then passed through the garnet–spinel transition and melted a further 20%, a 30% partial melt of fertile upper mantle and 0–1% of outer core material. The core component was included only in the plume source, and the ratio of plume source to upper mantle source was 19: 1. It is evident from this study that the PGE contents of at least some of the OJP basalts are too high to be generated by primitive mantle sources alone. A PGE-enriched component is required and we suggest that this is outer core material. While a sulphide-rich mantle component could also increase the PGE abundances (assuming that the sulphide is exhausted during partial melting), the sulphur-undersaturated nature of these basalts argues against this. Variations in the amount of outer core in the source (from 0 to 1 wt%) and degree of fractional crystallization can account for the entire range in PGE abundances of OJP basalts.

Large igneous provinces (LIPs) are massive crustal emplacements of predominantly mafic igneous rock (extrusive and intrusive), which are generated at rates far quicker than can be accounted for by plate margin volcanism (e.g. Coffin & Eldholm 1993, 1994). They include such manifestations as continental flood basalts, volcanic passive margins, oceanic plateaus, submarine ridges and seamount groups. The origin of LIPs has generally been attributable to surfacing mantle plumes, which originate via heat transfer across a mechanical boundary layer,

causing thermal instability and rising of superheated mantle material forming a plume head (e.g. Campbell & Griffiths 1990), the melting of which produces a flood basalt province. This is followed by a less voluminous and longer-lived 'plume tail' stage (e.g. Richards *et al.* 1989). Alternative models have been proposed, such as the rift-induced melting of fixed, anomalously hot upper-mantle 'hot cells' (e.g. Anderson *et al.* 1992) and magmatism induced by meteorite impact (Rogers 1982; Jones *et al.* 2002).

In terms of eruptive mechanism and duration,

there are two types of oceanic LIPs: those that occur on or near mid-ocean ridge axes (e.g. Iceland) and those that occur mid-plate (e.g. Hawaii). Both the Hawaiian and Icelandic hot spots have been almost continuously active for millions of years (e.g. Clague and Dalrymple 1987, 1989; Chalmers *et al.* 1995). The Kerguelen Plateau hot spot has displayed both periods of pulsed and continuous volcanism; the plateau formed both off-ridge and near the SE Indian Ridge during periods of its evolution (Royer *et al.* 1991; Frey & Weis 1995; Pringle & Duncan 2000; Duncan 2002). The Ontong Java Plateau, however, appears to have erupted in at least two and perhaps four discrete pulses at approximately 122, 90, 62 and 34 Ma (Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002; Birkhold 2000), although the initial pulse at *c.* 122 Ma seems to be by far the most voluminous. This causes difficulties in using a plume model for OJP petrogenesis (see Tejada *et al.* 2004 for further details), although Bercovici & Mahoney (1994) developed a model that could accommodate two magmatic events approximately 30 Ma apart within the framework of a rising plume head. Birkhold (2000) suggested further magmatic events that tap the OJP source could occur through plate reorganization producing pressure-release partial melting in the fossil plume head material that sits beneath the OJP (cf. Richardson *et al.* 2000; Klosko *et al.* 2001). These post-90 Ma events would be progressively smaller in magnitude and in degree of partial melting as the source is progressively depleted.

The depth of origin and chemical composition of plumes are intrinsically linked. The predominance of hot upper mantle in the Icelandic plume has been used to suggest a partial origin at the 670 km discontinuity (Fitton *et al.* 1997). Osmium, O, Sr, Nd, Pb and Hf isotopes have been used to suggest that ancient recycled crustal material has been incorporated into the Hawaiian plume (Hauri *et al.* 1996; Lassiter *et al.* 1996; Lassiter & Hauri 1998; Blichert-Toft *et al.* 1999). Others have suggested that outer core material is responsible for suprachondritic $^{187}\text{Os}/^{188}\text{Os}$ (Widom & Shirey 1996), and coupled suprachondritic $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$, in both Hawaii and other LIPs (Walker *et al.* 1995, 1997; Brandon *et al.* 1999). High $^3\text{He}/^4\text{He}$ values are also interpreted to suggest that the Hawaiian plume originated in the undegassed lower mantle (e.g. Kurz *et al.* 1996; Brandon *et al.* 1999). Geophysical and experimental evidence suggests that core-mantle interaction is not only feasible, but occurs in the D" layer (Shannon & Agee 1998; Vidale & Hedlin 1998; Vinnik *et al.* 1998). Ely & Neal (2003) have modelled the platinum-group

elements (PGEs) of OJP basalts cropping out on the Solomon Islands to infer a core-mantle boundary origin for the OJP, assuming an initial origin as a rising plume head. If the OJP were a product of either hot-cell activity or meteorite impact, only upper-mantle material would be involved and the erupted lavas would be expected to be MORB-like in character; Sr, Nd and Pb isotopes demonstrate that this is not the case (Mahoney & Spencer 1991, Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002; Neal *et al.* 1997). While there are still problems with a simple plume origin for the OJP (see Tejada *et al.* 2004), it appears that these are less severe than with the alternatives. Therefore, for this chapter we assume the origin of the OJP via a surfacing plume head.

Owing to their siderophile nature, the PGEs would be partitioned strongly into the metallic core during planetary differentiation. Incorporation of less than 1% of such material into the plume source would dominate the PGE budget. However, it has been suggested that the lower mantle may equilibrate with the core with respect to Os isotopes, but that admixtures resulting in high overall PGE abundances are unlikely (Puchtel & Humayun 2000). In this paper, we attempt to test the idea that PGE abundances can be used to indicate a core-mantle boundary origin for plumes, and expand upon the work of Ely & Neal (2003) using OJP basalts recovered from Ocean Drilling Program (ODP) Legs 130 and 192.

Geological background

Located in the SW Pacific (Fig. 1), the OJP is the world's most expansive oceanic large igneous province at approximately $2.0 \times 10^6 \text{ km}^2$, with a total volume of *c.* $5 \times 10^7 \text{ km}^3$ (e.g. M. Coffin pers. comm. 2003). The OJP is almost entirely submarine, although subaerial outcrops occur on the islands of Malaita, Makira, Ulawa and Santa Isabel in the Solomon Islands. This exposure is a result of uplift after collision between the OJP and the Australian plate (Coleman & Kroenke 1981; Petterson *et al.* 1997, 1999).

Remarkably, the chemical variation of the OJP basalts is extremely limited given the size of this LIP. Prior to ODP Leg 192, only two significant types of basalt were known for the OJP. Tejada *et al.* (2002) used the terms Singgalo Formation and Kwaimbaita Formation (from type localities exposed on Malaita) for the two isotopically distinct basalt types. Singgalo-type basalts are characterized by $(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7040\text{--}0.7042$, $(\epsilon_{\text{Nd}})_i = +3.8\text{--}+5.4$ and $^{206}\text{Pb}/^{204}\text{Pb} = 18.245\text{--}18.521$, and include the Unit A basalt at Site 807, whereas Kwaimbaita-type basalt

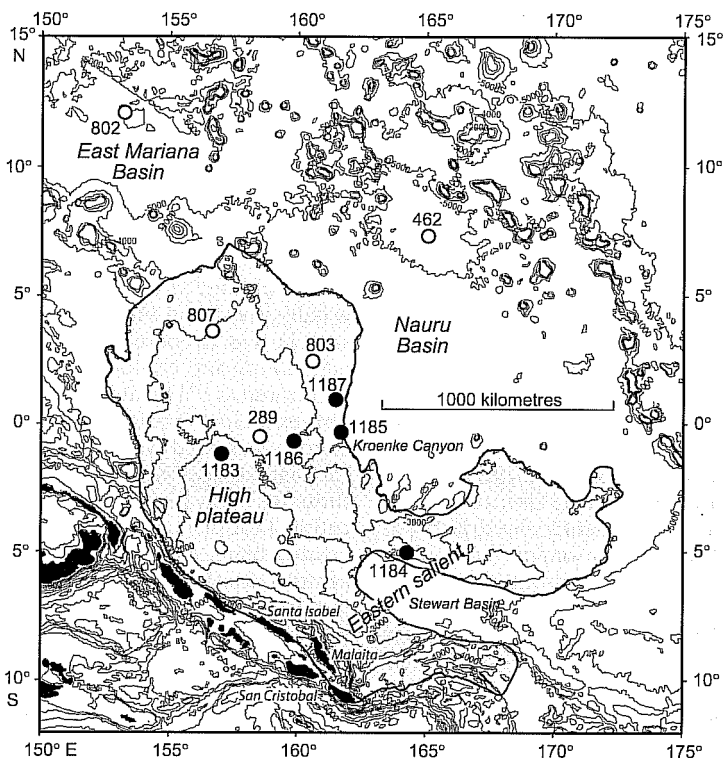


Fig. 1. Map showing the locations of all Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) drill sites on the Ontong Java Plateau (adapted from Mahoney *et al.* 2001). Leg 192 drill sites are marked by black circles; open circles represent pre-Leg 192 drill sites.

incompatible trace-element abundances are slightly depleted relative to those of the Singgalo-type basalts, and have $(^{87}\text{Sr}/^{86}\text{Sr})_t = 0.7034\text{--}0.7039$, $(\epsilon_{\text{Nd}})_t = +5.4\text{--}+6.5$ and $^{206}\text{Pb}/^{204}\text{Pb} = 18.626\text{--}18.708$, and include basalts of Units C–G at Site 807 and the c. 90 Ma basalts at Site 803 (Mahoney 1987; Mahoney & Spencer 1991; Mahoney *et al.* 1993; Tejada *et al.* 1996, 2002; Neal *et al.* 1997). The term $(\epsilon_{\text{Nd}})_t$ represents $^{143}\text{Nd}/^{144}\text{Nd}$ normalized to the chondritic uniform reservoir, both being corrected for the age of the sample. Abundances of MgO are slightly lower in Singgalo-type basalts, typically showing a range of 6–7.3 wt%, while the range in the Kwaimbaita-type basalts is usually from 7 to 8 wt%. Notably, Singgalo-type basalts have been found only on the northern and southern margins of the plateau, at ODP Site 807 and as subaerial outcrops in the Solomon Islands, respectively. Kwaimbaita-type basalts appear to be present across the plateau, being found in cores from Deep Sea Drilling Program Site 289, and ODP Sites 803, 807, 1183, 1185 and 1186, as well as subaerial outcrops on the Solomon Islands (Fig. 1). Birkhold (2000) described basalts on Makira

from the Wairahito Formation, which contain higher abundances of the incompatible trace elements than other OJP basalts, but were isotopically similar to Kwaimbaita basalts. These are interpreted as a more evolved lava type ($\text{MgO} = 4.5\text{--}6.6$ wt%) derived from the Kwaimbaita source and, to date, basalts of this type have only been described from Makira. Several basalt flows recovered during ODP Leg 192 at Sites 1185 and 1187 have higher MgO abundances (8–11 wt%), and lower contents of incompatible trace elements than those of the Kwaimbaita-type basalts ($\text{Nb} = \text{c. } 2.1\text{--}2.4$ v. $\text{c. } 2.9\text{--}3.5$ ppm, respectively; Fitton & Godard 2004). However, these primitive basalts are isotopically similar to the more evolved Kwaimbaita-type basalts (Tejada *et al.* 2004).

Analytical techniques

Sixteen basalt samples were analysed for PGEs, and major and trace elements. Sample collection began onboard the *JOIDES Resolution* during Leg 192, where precautions against contamination were taken from the moment the cores

were brought onboard. Staff and Leg 192 scientists removed all jewellery and, where possible, contact with metal was avoided. Samples were selected from each volcanic unit as determined on board the ship, making sure to select only the least altered samples. Samples were selected adjacent to those used in Fitton & Godard's (2004) study. Each sample was then cut from the working half of the core with a diamond-impregnated saw and sealed in a plastic bag for transport. Once at Notre Dame, samples were first cut into approximately 0.5 cm-thick wafers with a diamond-impregnated saw. Care was taken to avoid veins, especially those with black or brown haloes, in which secondary sulphide (pyrite) or Fe-oxyhydroxides, respectively, were observed in thin section. All surfaces were subsequently ground away with a diamond grinding disk to remove any contamination imparted by the drill bit or the saw during cutting. The samples were then crushed using alumina jaw crushers, washed in dilute (0.6 M) HCl in an ultrasonic bath for 20 min to remove secondary minerals and then rinsed several times with 18 M Ω H₂O, until all flocculants were removed.

Heterogeneous distribution of trace primary sulphides in basaltic samples can produce the 'nugget' effect, so care was taken to obtain a sample that truly represents the whole-rock composition. Reducing this effect is accomplished by choosing fine-grained samples and/or powdering a large amount of sample (10–20 g) in an alumina ball mill, followed by remixing each powdered sample (if it was left unagitated for more than a week) before analysing a given aliquot. This study used the PGE standard UMT-1, which was re-mixed on a roller for 24 h before each split was taken. Two-hundred-and-fifty milligrams of each sample powder was then dissolved in sequential stages of 16 M HNO₃ + 29 M HF, aqua regia (made from 16 M HNO₃ and 12 M HCl), concentrated HCl and finally 0.6 M HCl (see Ely *et al.* 1999; Ely & Neal 2002 for full details of the PGE analytical techniques and data reduction methods employed). All acids used were double distilled in-house from reagent-grade stocks. For details regarding the analytical procedures used in determining the major and trace elements of the samples discussed in this paper, see Jenner *et al.* (1990), Neal (2001) and Shafer *et al.* (2004).

Where possible, multiple isotopes of each of the PGEs were analysed and the interference-free isotope of highest abundance used to quantify PGE contents (Table 1). For example, while ¹⁹³Ir is the most abundant Ir isotope (62.7%), it is interfered with by ¹⁷⁷Hf¹⁶O if Hf is present in the

whole rock at >1 ppm, as Hf is not completely removed by the cation-exchange procedure (Ely *et al.* 1999). The PGE abundances are based on results obtained for ¹⁰²Ru, ¹⁰³Rh, ¹⁰⁵Pd and ¹⁹¹Ir. The Pt isotope used depends on the abundance of Hf in the sample (and the magnitude of the HfO interference), but is either ¹⁹⁵Pt or ¹⁹⁸Pt. The low abundance of ¹⁹⁸Pt generates larger errors than those with ¹⁹⁵Pt, but the oxide interferences are typically present on ¹⁹⁵Pt (¹⁷⁹Hf¹⁶O) and absent on ¹⁹⁸Pt (¹⁸²W¹⁶O), as very little W is usually present in the sample, especially as the sample was prepared using alumina crushing and powdering equipment. Mercury was also monitored as ¹⁹⁸Hg interferes with ¹⁹⁸Pt, but was always below limits of detection. The ¹⁹⁵Pt isotope was used when Pt abundances derived for that isotope were within statistical error of those for ¹⁹⁸Pt. Limits of detection and quantification (Table 1) were calculated from a procedural blank analysed immediately before the sample. In this way, memory effects can be monitored (although with the wash-out procedure described in Ely *et al.* 1999, they were non-existent) and, if needed, applied to the calculation of PGE abundances in the sample. Oxide formation was monitored pre-run using a 1 ppb tuning solution containing Ce. The instrument was tuned such that CeO was no greater than 1–2% of the Ce counts before a run proceeded; typically CeO was around 0.5% of the Ce counts.

Results

Trace-element and selected major-element data for the samples analysed are presented in Table 1. Major- and trace-element data presented are from this study (all data for Leg 130 samples; V, Cr and Zn for all samples), M. Godard (all remaining Leg 192 trace-element data; pers. comm.) and Fitton & Godard (all Leg 192 major-element data; 2004); MgO for some samples are presented from within the same unit, within a few centimetres of the sample selected for PGE analysis. The primitive or 'Kroenke-type' basalts from Sites 1185 and 1187 are characterized by MgO and TiO₂ abundances of 8.0–10.9 wt% and 0.69–0.76 wt%, while those of Kwaimbaita-type basalts range from 7.3 to 8.0 wt% and 1.04–1.10 wt%, respectively. The Kroenke-type basalts are named after the location of Site 1185 adjacent to Kroenke Canyon, a large submarine canyon just south of this site. Kroenke-type basalts contain Ni and Cr abundances (177–250 and 371–494 ppm, respectively) that indicate that these lavas have experienced a lower degree of fractional crystallization

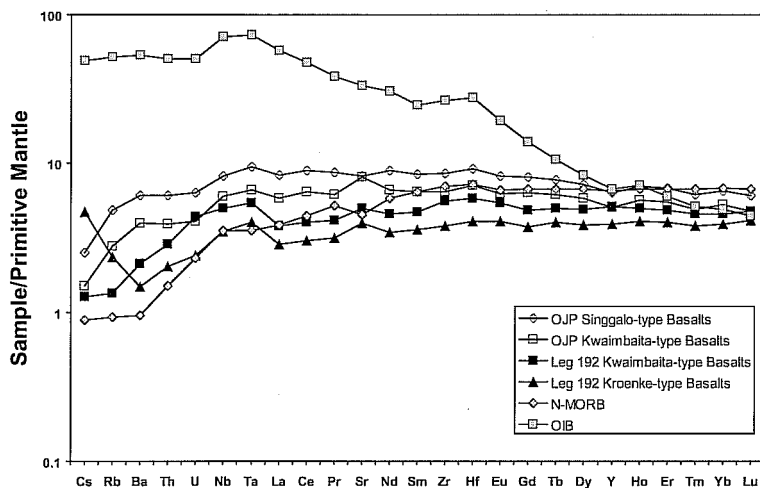


Fig. 2. Incompatible trace-element abundances normalized to primitive mantle for the Ontong Java Plateau basalts. Data are from Mahoney *et al.* (1993), Tejada *et al.* (1996, 2002), Neal *et al.* (1997) and Fitton & Godard 2004. Normalization values are from Sun & McDonough (1989).

than the Kwaimbaita-type basalts (122–135 and 142–294 ppm, respectively). Incompatible trace-element abundances for both the Kroenke-type and Kwaimbaita-type basalts are also consistent with these observations (Fig. 2). In addition to these Leg 192 samples, two samples from ODP Leg 130 (one each from Sites 803 and 807) were also analysed, both of which have major and trace elements in the range of Kwaimbaita-type basalts. The Site 803 basalt, however, is apparently approximately 32 Ma younger (*c.* 90 Ma) than the other Kwaimbaita-type basalts analysed (*c.* 122 Ma; e.g. Mahoney *et al.* 1993).

Platinum-group element abundance data are presented in Table 1. All values are reported with $\pm 2\sigma$ errors (after Ely & Neal 2002). Values falling below the limit of detection (LOD: background + 3σ) are reported as 'BDL' (below detection limit). Values presented in italics fall between the LOD and the limit of quantification (LOQ: background + 10σ). The LOD and LOQ are calculated using a blank analysed immediately before the sample or reference material. Average values for UMT-1 are also reported. As the PGEs were quantified using standard addition, reference materials were analysed at least once for every two unknown samples. For this study, UMT-1 data are reported as averages of 13 separate analyses and the errors here represent $\pm 2\sigma$ of the mean value.

In general, the Leg 192 Kroenke- and Kwaimbaita-type basalts contain overlapping PGE abundances (Fig. 3a–d), which are roughly equivalent to those found in komatiites (e.g.

Brügmann *et al.* 1987; Rehkämper *et al.* 1999; Puchtel & Humayun 2001). While Pt and Pd values for these basalt types are essentially identical, a slight distinction can be made between Kwaimbaita- and Kroenke-type basalts in terms of Ir, Ru and Rh in Hole 1185B. Positive correlations exist between Ni and Cr v. Ir and Ru (Fig. 4a–d), with the primitive Kroenke-type basalts containing higher abundances. The pattern breaks down, however, if basalts external to this site are also included. Primitive-mantle-normalized plots of the PGEs (Fig. 3a–d) exhibit a generally positive slope. Slight negative Pd anomalies also occur in some samples, but most have a flat–slightly positive transition from Pt to Pd to Y (Fig. 3a–d). The Leg 130 Kwaimbaita-type basalts have slightly lower Ru and Rh abundances than the Leg 192 basalts, but have similar Pt and Pd, although both Leg 130 basalts also exhibit Pd depletions that produce negative anomalies in normalized profiles (Fig. 3b).

The PGE compositions of Leg 130 and Leg 192 OJP basalts are similar to those from the Solomon Islands (Fig. 5) (Ely & Neal 2003), regardless of age. The Leg 192 Kwaimbaita-type basalts generally have lower PGE abundances relative to the Kroenke-type basalts, but they overlap with PGE abundances of Kwaimbaita- and Singgalo-type basalts from the Solomon Islands (Ely & Neal 2003). The clinopyroxene-plagioclase-titanomagnetite cumulates ML-475 and ML-476 (Malaita; Ely & Neal 2003) contain PGE abundances that are similar to the Leg 192 basalts, but contain twice the Pd content. While

Table 1. *Platinum-group (ppb), major- (wt%), and trace-element (ppm) abundances for the ODP Leg 130 and Leg 192 basalts analysed in this study**

Leg	Site	Hole	Core	Section	Piece	Interval (cm)	Unit	Type	Ir	Ru	Rh	Pt	Zn	Pd	2 σ	Sr	
192	1185	A	10R	3	1A	0-20	5B	Kroenke	0.5	0.9	0.4	8.5	0.5	13.4	0.5	2 σ	
192	1185	B	5R	5	2C	143-149	2	Kroenke	2.4	1.5	1.3	13.6	2.0	18.2	0.1	2.3	
192	1185	B	6R	4	2	101-120	5	Kroenke	1.5	1.3	1.2	11.7	1.3	13.9	0.3	4.5	
192	1185	B	11R	1	3A	13-17	9	Kroenke	1.4	1.3	1.1	11.3	1.9	11.5	0.1	2.1	
192	1187	A	6R	6	2B	92-97	3B	Kroenke	0.6	1.0	0.9	11.3	1.9	11.5	0.1	0.8	
192	1187	A	13R	2	1B	50-55	6	Kroenke	1.1	0.8	1.2	9.4	1.0	11.7	0.3	5.2	
192	1187	A	13R	2	1B	50-55	6	Kroenke	1.7	1.1	0.9	13.8	2.9	14.3	0.2	1.6	
192	1187	A	16R	3	3A	34-39	12	Kroenke	BDL	1.4	1.0	11.2	5.2	24.8	0.1	3.0	
192	1183	A	59R	1	3B	40-45	5B	Kwaibaita	BDL	1.7	1.0	28.8	3.6	18.9	0.3	2.1	
192	1183	A	65R	1	5D	112-128	7	Kwaibaita	1.1	0.9	1.3	19.1	3.4	16.7	0.8	6.6	
192	1185	B	17R	3	3	39-45	10	Kwaibaita	1.4	0.9	0.9	18.7	4.3	13.3	0.1	4.3	
192	1185	B	21R	4	1B	45-50	11	Kwaibaita	1.0	1.2	0.4	10.1	0.7	10.9	0.2	1.2	
192	1185	B	24R	2	6	87-93	12	Kwaibaita	0.9	1.1	BDL	12.5	0.4	9.1	0.8	3.1	
192	1186	A	32R	2	1	101-106	1	Kwaibaita	BDL	1.2	0.1	16.3	6.9	8.5	0.3	1.6	
192	1186	A	39R	1	10	42-58	4	Kwaibaita	1.4	1.5	1.1	13.6	0.7	11.4	0.1	0.8	
130	803	D	70R	3	7	93-98	9B	Kwaibaita	BDL	0.7	0.1	17.3	2.0	7.3	0.1	0.8	
130	807	C	93R	3	1	16-19	4G	Kwaibaita	BDL	0.6	0.1	18.4	0.4	8.8	0.04	0.9	
								UMT-1 (<i>n</i> = 13)	9.4	9.7	1.3	113	12	95	1.4	11	
								UMT-1 (cert.)	8.8	10.9	1.5	129	5	106	1.5	3	
192	1185	A	10R	3	1A	0-20	5B	Kroenke	8.90	10.99	40	248	404	51	189	101	81
192	1185	B	5R	5	2C	143-149	2	Kroenke	10.44	10.92	41	543	397	56	238	103	101
192	1185	B	6R	4	2	101-120	5	Kroenke	10.91	10.86	42	260	443	63	250	102	96
192	1185	B	11R	1	3A	13-17	9	Kroenke	8.93	10.65	42	294	403	51	177	100	88
192	1187	A	6R	6	2B	92-97	3B	Kroenke	9.69	10.92	40	236	371	49	199	96	71
192	1187	A	13R	2	1B	50-55	6	Kroenke	9.96	10.72	40	307	469	51	207	100	95
192	1187	A	16R	3	3A	34-39	12	Kroenke	8.03	10.67	42	n.a.	494	52	212	100	n.a.
192	1183	A	59R	1	3B	40-45	5B	Kwaibaita	7.59	12.07	48	212	212	51	135	149	0.55
192	1183	A	65R	1	5D	112-128	7	Kwaibaita	7.64	12.26	46	459	282	48	131	150	128
192	1185	B	17R	3	3	39-45	10	Kwaibaita	7.40	12.60	45	265	142	49	127	149	83
192	1185	B	21R	4	1B	45-50	11	Kwaibaita	7.61	12.58	47	295	240	49	129	141	89
192	1185	B	24R	2	6	87-93	12	Kwaibaita	7.29	11.05	49	n.a.	230	51	122	136	n.a.
192	1186	A	32R	2	6	101-106	1	Kwaibaita	7.34	12.36	47	n.a.	218	51	131	243	0.16
192	1186	A	39R	1	10	42-58	4	Kwaibaita	7.96	12.23	49	425	294	50	137	126	112
130	803	D	70R	3	7	93-98	9B	Kwaibaita	6.53	12.29	59	297	275	n.a.	142	n.a.	n.a.
130	807	C	93R	3	1	16-19	4G	Kwaibaita	7.22	12.62	58	348	178	n.a.	117	n.a.	n.a.

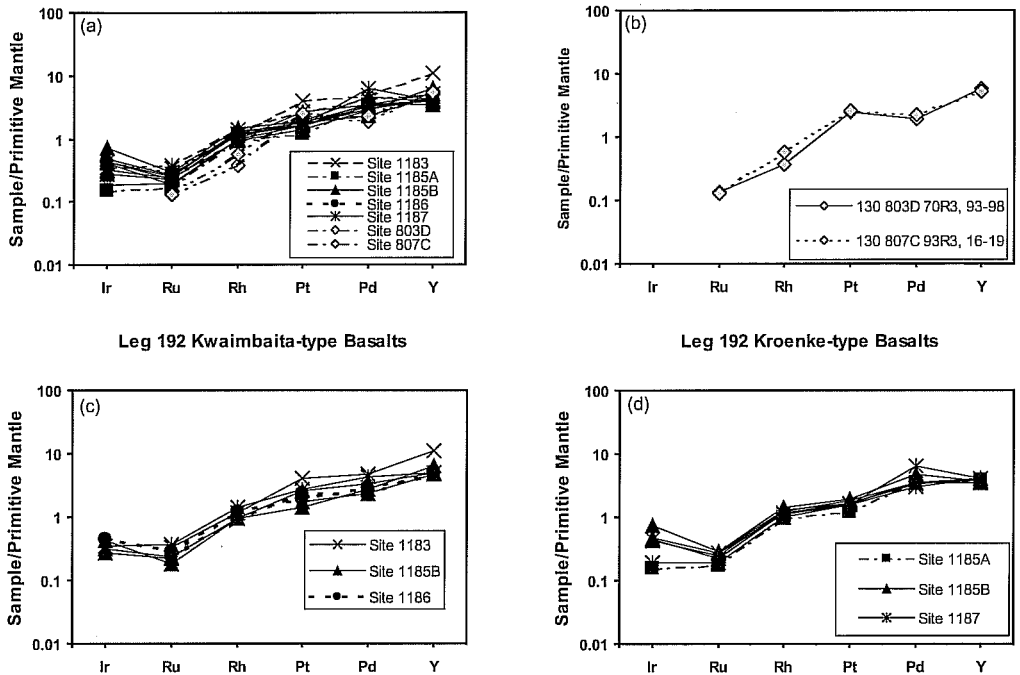


Fig. 3. Primitive-mantle-normalized plots of PGE abundances for (a) all of the OJP basalts analysed in this study, (b) the Leg 130 basalts, (c) the Leg 192 Kwaimbaita-type basalts and (d) the Leg 192 Kroenke-type basalts. Only the Leg 130 samples indicate any loss of Pd via low-temperature alteration ($[\text{Pd}/\text{Pt}]_{\text{pm}} < 1$). Normalization values are from McDonough & Sun (1995).

the Leg 192 basalts generally contain more Ir and Rh than those from the Solomon Islands, it is evident that a continuum of compositions exists.

Discussion

Low-temperature alteration and the effects of sulphide immiscibility

Thin-section examination revealed that all samples were subject to at least small degrees of secondary (low-temperature) alteration. Most alteration took place near veins and miarolitic cavities, but mild, incomplete replacement of olivine and the mesostasis by clay minerals was also present. Ely & Neal (2003) reported negative Pd anomalies in several samples from the Solomon Islands, and suggested it may result from the preferential mobilization of Pd during secondary alteration processes. This negative Pd anomaly is also found more commonly in Singgalo-type basalts, which are stratigraphically above the more ubiquitous Kwaimbaita-type basalts and have been exposed to

secondary alteration–weathering processes for a longer period. However, while this feature is present in both samples analysed from Leg 130 (Fig. 3b), it is generally absent from the Leg 192 basalts (Fig. 3c, d). The presence of a limestone unit (Unit B) between the Singgalo- and Kwaimbaita-type basalts at Site 807 and the lack of any Singgalo-type basalt flows at Site 803 suggests that the Kwaimbaita-type basalts at these two sites were similarly exposed to secondary alteration–weathering processes. Plots of MgO v. Pt and Pd (Fig. 6a, b) demonstrate that several OJP samples from the Solomon Islands have similar Pt abundances to those from Leg 192, but Pd data for the former are scattered and are sometimes as low as mid-ocean ridge basalt (MORB) values. While the separation of immiscible sulphide could have this effect on Pd for the now subaerially exposed basalts from the Solomon Islands, this would affect all of the PGEs, not just Pd (Ely & Neal 2003). Basalts, such as MORBs, that have experienced the separation of immiscible sulphide are lower in overall PGE abundance, and have significant vertical scatter for Pt and Pd at a given MgO abundance (Fig. 6a, b). The lack of negative Pd

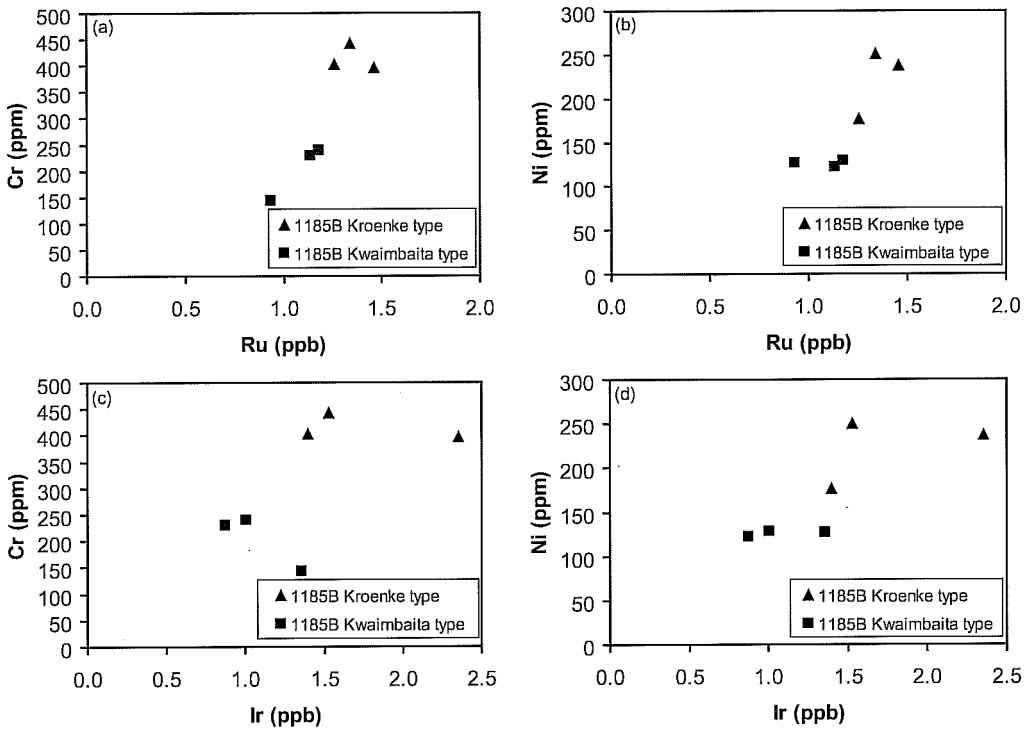


Fig. 4. (a–d) Plots of Ir and Ru v. Cr and Ni. Positive correlation between these elements demonstrates that these PGEs are exhibiting lithophile behaviour and are being controlled by olivine and/or Cr-spinel crystallization. See text for discussion.

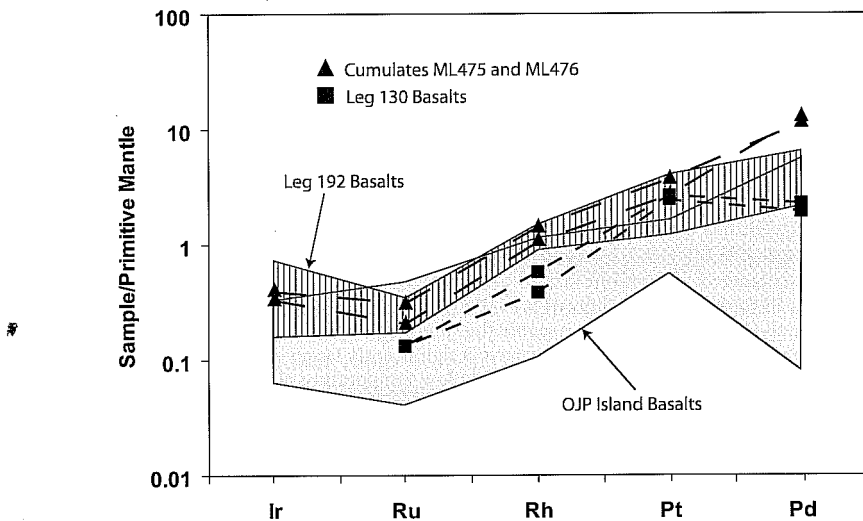


Fig. 5. Platinum-group element profiles, normalized to primitive mantle, for all OJP basalts analysed to date. The Leg 192 Kroenke- and Kwaimbaita-type basalts make up the high end of the range of PGEs found in the OJP, but are not completely distinct from Kwaimbaita-type basalts from elsewhere on the plateau or the Singgalo-type basalts. A continuum of compositions is evident, with Singgalo-type basalts tending towards lower PGE abundances, and Kroenke-type basalts tending towards higher. Data for the Solomon Island basalts are from Ely & Neal (2003). Primitive-mantle-normalization values are from McDonough & Sun (1995).

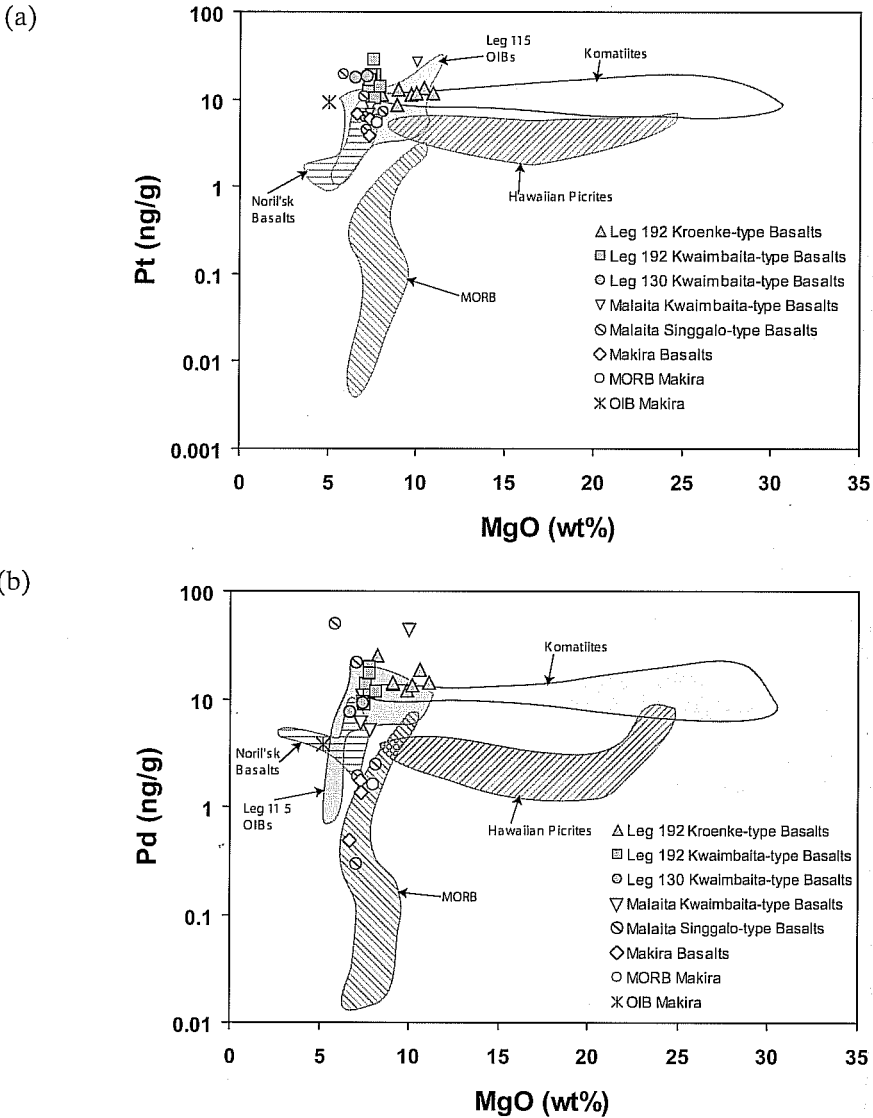


Fig. 6. (a) MgO (wt%) v. Pt (ppb) and (b) MgO (wt%) v. Pd (ppb) for various types of basalts and komatiites. **#** Platinum abundances for all OJP samples plot in roughly the same region, and are similar to values for komatiites. Palladium abundances for the OJP basalts have a significantly larger range than those of Pt, with many Solomon Islands basalts having experienced Pd loss via low-temperature alteration; sulphide removal would have affected all of the PGEs, as it has in MORB (see text for discussion). The Pd values for the Leg 192 basalts appear unaffected by low-temperature alteration. Data are from Arndt & Nesbitt (1984), Arndt (1986), Lightfoot *et al.* (1990), Fryer & Greenough (1992), Brüggmann *et al.* (1993), Devey *et al.* (1994), Puchtel *et al.* (1996), Norman & Garcia (1999), Rehkämper *et al.* (1999), Tatsumi *et al.* (1999), Bennett *et al.* (2000), Puchtel & Humayun (2001), Tejada *et al.* (1996, 2002), Ely & Neal (2003) and Fitton & Godard (2004).

anomalies in the Leg 192 basalt PGE profiles indicates that the data reflect primary igneous processes and are not the result of low-temperature alteration.

The PGEs partition strongly into sulphides, with partition coefficients between sulphide and silicate melts of greater than 1×10^4 (e.g. Barnes *et al.* 1985; Peach *et al.* 1990, 1994; Fleet *et al.*

1991, 1996; Bezmen *et al.* 1994). If a silicate magma were to reach saturation with respect to sulphur under typical mantle oxygen fugacities (e.g. Lee *et al.* 2003 and references therein), an immiscible sulphide liquid would form, removing chalcophile elements such as the PGEs, Fe, Ni and Cu. However, compositional data indicate that the OJP basalts were sulphur-undersaturated at the time of eruption (cf. Michael & Cornell 1996; Roberge *et al.* 2004). Indeed, thin-section examination reveals only traces of primary sulphide, which is present as micron-size interstitial blebs. This could indicate that immiscibility has occurred and practically all sulphur has been removed or that immiscibility was only reached once the bulk of the basalt flow had crystallized. Ratios such as Cu/Pd are similar to those in primitive mantle (7300) for the Kroenke-type basalts (*c.* 4100–8350) and are generally higher for all Kwaimbaita-type basalts (8000–28 500). Singgalo-type basalts from the Solomon Islands vary dramatically, but both Cu and Pd may have been affected by alteration in these samples (Ely & Neal 2003). Primitive-mantle-normalized PGE profiles (Fig. 3a–d) demonstrate the difference between the OJP basalts and MORB, where the latter has experienced sulphur saturation, followed by immiscibility and separation of a sulphide-rich liquid (e.g. Keays 1995; Roy-Barman *et al.* 1998; Rehkämper *et al.* 1999). During crystal fractionation, Pd and Y should have similar incompatibilities in a sulphur-undersaturated system, resulting in [Pd/Y]_{pm} ratios of approximately 1 (Brügmann *et al.* 1993; where pm stands for ‘normalized to primitive mantle’). This is the case for the OJP basalts, whereas in MORB [Pd/Y]_{pm} ≪ 1. Experimental observations of the sulphur concentration at sulphide saturation (SCSS) by Mavrogenes & O’Neill (1999) and Holzheid & Grove (2002) indicate that many magmas may initially be close to sulphur saturation, but a strong negative correlation between SCSS and pressure suggests that most should arrive at the surface sulphur-undersaturated. Only large degrees of crystal fractionation, rapid temperature decrease due to assimilation, or assimilation of a relatively sulphur-rich component (such as continental crust) would push the system towards sulphur-saturation. Although Kwaimbaita- and Singgalo-type basalts have experienced relatively large degrees of crystal fractionation, the [Pt/Y]_{pm} and [Pd/Y]_{pm} ratios have not been overly fractionated. This demonstrates that the PGEs are behaving as lithophile elements, and indicates a lack sulphur saturation and subsequent removal of a sulphide-rich immiscible melt.

On the basis of the above discussion, temperature estimates of the OJP source (our assumption is that this is a plume head) can now be made. Using major- and trace-element models, pressure estimates for the initial melt of the OJP are *c.* 3.5–4.0 GPa (Mahoney *et al.*, 1993; Farnetani & Richards, 1994; Neal *et al.* 1997; Tejada *et al.* 2002). Estimates for sulphur in primitive mantle are about 250 ppm (Keays 1995), and with the average degree of partial melting estimated to be approximately 30% (Fitton & Godard 2004), the models of Mavrogenes & O’Neill (2002) and Holzheid & Grove (2002) can be used to estimate the minimum temperature of the plume head required to generate the OJP basalts. One assumption in this temperature calculation is that the sulphur content of the initial melt can be projected from that of the source and the degree of partial melting. Sulphide in the source is assumed to be exhausted at no more than approximately 20–23% partial melting (e.g. Rehkämper *et al.* 1999). Therefore, at 30% partial melting, sulphur is exhausted, and the maximum concentration of sulphur in the melt is approximately 830 ppm. Using equation (10) from Mavrogenes & O’Neill (1999):

$$\ln[S(\text{ppm})] = \frac{A}{T} + B + \frac{CP}{T} + \ln a_{\text{FeS}}^{\text{sulphide}}$$

and assuming $a_{\text{FeS}}^{\text{sulphide}} = 1$ (where $a_{\text{FeS}}^{\text{sulphide}}$ is the activity of FeS in the sulphide liquid), and the basaltic values for the parameters A (enthalpy; –6684K), B (entropy; 11.52), and C (volume; –0.047K/bar) along with the pressure (*P*) estimates outlined above, the minimum temperature (*T*) required for sulphur to remain undersaturated at a given pressure can be calculated. Pressure estimates for the Ontong Java Plateau based on phase equilibria suggest melting in a range of 1.0–6.0 GPa, but with the majority of data indicating pressures between 1.0 and 4.0 GPa (Tejada *et al.* 2002). Other methods used to estimate pressure of melting (e.g. Farnetani & Richards 1994) are consistent with these values (i.e. *c.* 0.6–4.0 GPa). For melts originating at 1.0 and 6.0 GPa, the minimum calculated temperatures are approximately 1170°C and 1650°C, respectively. Trace-element modelling seems to constrain much of the melting to between 3.5 and 4.0 GPa (Mahoney *et al.* 1993; Neal *et al.* 1997; Tejada *et al.* 2002), and calculated temperatures for melts originating at these pressures are approximately 1465–1515°C. However, variation in the sulphur content of the source could cause these temperature estimates to change significantly. At 200 ppm sulphur the calculated

temperatures are *c.* 1385–1435°C, while at 300 ppm sulphur they would be *c.* 1535–1585°C to prevent saturation. Assuming a potential temperature for ambient mantle of 1280°C (McKenzie & Bickle 1988), the model temperatures for 250 ppm sulphur at 3.5–4.0 GPa are lower than the ΔT of +350°C excess temperature estimates (without lithospheric extension) of Farnetani & Richards (1994) that were required to produce magma volumes of 5×10^7 km³ at 35% partial melting, but overlap with those of Griffiths & Campbell (1990) who estimated a 100–200°C initial excess temperature for a plume head that is beginning to impinge on the rigid lithosphere. Farnetani & Richards (1994) also use an ambient mantle temperature of 1280°C, making their temperature estimates for large volume melts ($\geq 1 \times 10^7$ km³) in excess of 1600°C. The model of Farnetani & Richards (1994) takes account of the presence of lithosphere, which necessitates large temperature differentials to produce high-volume melts. While the constants A, B and C (and thus the SCSS) can be affected by variations in magma composition, especially Fe and Ni, and would therefore need to be experimentally determined for OJP basalts to ensure more accurate temperature estimates, the +185–+235°C initial excess temperatures of this study are comparable to the those for the OJP (>1500°C; Fitton & Godard 2004) and other plumes (e.g. Iceland; 1480–1520°C or a ΔT of 180–240°C; Maclennan *et al.* 2001). The temperatures may thus be underestimated considering the high degree of partial melting for the OJP (*c.* 30%; Fitton & Godard 2004) relative to other plumes. However, it must be remembered that these estimates are *minimum* temperatures because they are the minimum required for sulphur to remain undersaturated.

One possible problem with the model is that melting is polybaric. In a triangular melting regime, where the degree of partial melting decreases to zero at the base, it is possible that sulphur saturation may be reached. Assuming that the degree of partial melting varies linearly with depth, that the initial sulphur content of the mantle is constant through the melt column and the maximum degree of partial melting is 35%, then at 4.0 GPa the degree of partial melting would be 14%, and the resulting magma concentration for a source containing 250 ppm sulphur would be 1786 ppm (assuming sulphide exhaustion), requiring unrealistically high temperatures in excess of 1800°C to remain undersaturated. However, there are several problems with this analysis. First, at 14% partial melting, the sulphides in the source may or may not have been exhausted. As noted above, it has

been estimated that the *maximum* degree of partial melting at which sulphides are exhausted in a mantle source is approximately 20–23%, although the exact point at which this occurs will be model dependent (i.e. source composition, modal mineralogy, etc.). If we assume that sulphide is exhausted at 20% and all of the sulphur is in the sulphides, then sulphur concentrations could be much lower. Another factor to take into account is that, due to the relative homogeneity of the OJP basalts, the magma(s) that produced them must have been well mixed. If the resulting basalts indicate that the *average* degree of partial melting was 30% (Fitton & Godard 2004), then the maximum degree of melting in a triangular melting regime would have to be much higher (*c.* 40–50%). Thus, at 4.0 GPa, assuming partial melting is linear over the melting depths, the degree of partial melting would be 16–20%, the sulphur concentration would be 1250–1563 ppm for a 250 ppm source (assuming source exhaustion of sulphides) and would require temperatures of approximately 1680–1780°C for sulphur to remain undersaturated. These temperatures exceed even mantle potential temperature of Farnetani & Richards (1994). In addition, the maximum degree of partial melting would be unreasonably high. Finally, we envision the source of the OJP to have risen while it melted, such that a given part of the source actually melts over a range of depths and pressures. Thus, at melt initiation, the degree of partial melting may be smaller than the average, but during ascent it is envisaged that the melt produced would have been well mixed, thus producing basalts with compositions consistent with derivation by an average of 30% partial melting.

Supporting evidence for the calculated temperature excess comes from the fossil geotherm preserved by megacrysts and peridotite xenoliths found in pipe-like intrusives of alnöite cropping out on the island of Malaita, Solomon Islands (Allen & Deans 1965; Nixon & Coleman 1978). The geotherm thus recorded gives a temperature at 4.0 GPa of 1320–1350°C (Nixon & Coleman 1978; Nixon and Boyd 1979). Phase relations in the peridotite xenoliths are consistent with a cooling event in the mantle prior to inclusion in the erupting alnöite. For example, spinel in some of the peridotite xenoliths has a corona of garnet and such textures have been interpreted as indicating a cooling of the mantle (Neal & Nixon 1985; Nixon & Neal 1987), which occurred after the dissipation of heat that produced the large degrees of partial melting needed to form the OJP. Hence, while the temperatures recorded in the fossil

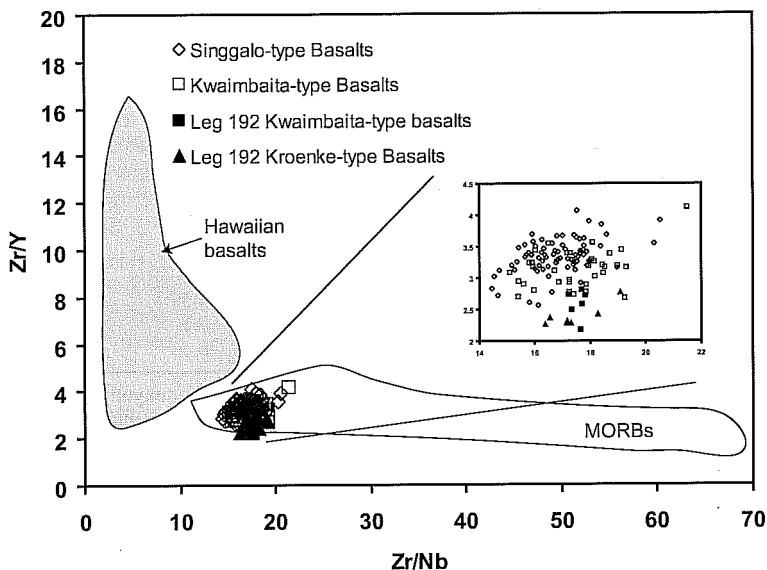


Fig. 7. Zr/Y v. Zr/Nb for the OJP basalts. The inset is an expanded view of the OJP sample set. New data from Leg 192 are similar to those of previous studies. Leg 192 basalts, like other OJP basalts, exhibit relatively limited variability in Zr/Y and Zr/Nb , which can be generated by initial melting in the garnet stability field followed by continued melting in the spinel stability field as the plume head continued to rise. See text for discussion. Data are from Tejada *et al.* (1996, 2002) and Fitton & Godard (2004). The MORB field represents East Pacific Rise MORB (J. Sinton unpublished data). The Hawaiian basalt field is generated by data from references too numerous for them all to be included here.

geotherm are about 150°C cooler than the estimates presented here, they are consistent with a large temperature perturbation associated with OJP formation.

Source modelling

Previous source modelling used a hybrid source composed of lower- and upper-mantle components to generate the OJP basalts (Mahoney *et al.* 1993; Farnetani & Richards 1994; Farnetani *et al.* 1996; Neal *et al.* 1997; Tejada *et al.* 2002; Ely & Neal 2003). New estimates indicate that the maximum degree of partial melting that formed OJP basalts may have been as high as 35%, although the average is still about 30% (Fitton & Godard 2004). The new samples from Leg 192 can now be used to refine the overall mineralogy of the proposed hybrid source. For example, Leg 192 basalts exhibit relatively limited variability in Zr/Y and Zr/Nb (Fig. 7). Neal *et al.* (1997) showed that such limited ranges in these ratios could be generated by initial melting in the garnet stability field of a composite source (primitive (lower) mantle \pm MORB source) followed by melting of this source in the spinel stability field as the plume head surfaced.

Following this, Ely & Neal (2003) successfully modelled the PGE abundances and patterns in OJP basalts exposed on the Solomon Islands using this composite source but with the addition of 0–0.5% of outer core material. Recently, it has been argued that there is no isotopic evidence for the presence of a depleted MORB source component in the OJP basalts (Tejada *et al.* 2002). However, if the plume model is to be used to explain OJP origin, entrainment of some upper-mantle material into the plume head would have occurred (cf. Campbell & Griffiths 1990), even though it may not be depleted MORB source. Therefore, the model presented here assumes an average degree of partial melting of 30% (cf. Fitton & Godard 2004), which is achieved through 10% melting of a garnet peridotite of primitive mantle composition, which then rises through the garnet–spinel transition and melts a further 20%. In addition to this plume component, an average fertile upper-mantle spinel lherzolite (Rehkämper *et al.* 1999) representing entrained upper mantle is melted to 30% and mixed into the plume melt at a ratio of 1: 19. The primitive mantle composition of McDonough & Sun (1995) is used. Source mineralogy for the garnet lherzolite is

Table 2. Partition coefficients and concentrations (ppb) for the OJP basalt modelling

	Ir	Ru	Rh	Pt	Pd	Reference*
Olivine	0.77	1.7	1.8	0.08	0.03	1
OPX	1.8	1.9	0.8	2.2	0.3	2
CPX	1.8	1.9	0.8	0.8	0.3	3
Cr-spinel	100	151	63	3.3	1.6	4
Sulphide	4400	2400	3000	6900	6300	5
Plagioclase	0.3	0.3	0.4	0.3	0.2	6
Ti-mag	500	300	130	3	1.1	7
Spinel lherzolite	2.46	4.29	0.8	4.34	2.48	8
Primitive mantle	3.2	5.0	0.9	7.1	3.9	9
Outer core	188	669	134	776	572	10

* References for data are as follows: 1, Capobianco *et al.* (1994), Puchtel & Humayun (2001); 2, Ely & Neal (2003); 3, Capobianco *et al.* (1990, 1991), Capobianco & Drake (1994); 4, Capobianco *et al.* (1990), Puchtel & Humayun (2001); Capobianco *et al.* (1990); 5, Bezmen *et al.* (1994), Peach *et al.* (1994), Tredoux *et al.* (1995); 6, Capobianco *et al.* (1990, 1991), Capobianco & Drake (1994); 7, Capobianco *et al.* (1994), Neal *et al.* (2002); 8, Rehkämper *et al.* (1997); 9, McDonough & Sun (1995); 10, Snow & Schmidt (1998).
OPX, orthopyroxene; CPX, clinopyroxene; Ti-mag, titanomagnetite.

taken to be olivine, orthopyroxene, clinopyroxene, sulphide and garnet (60: 14.94: 20: 0.06: 5). For both compositions of spinel lherzolite, the mineralogy was taken to be olivine, orthopyroxene, clinopyroxene, sulphide and Cr-spinel (60: 14.94: 20: 0.06: 5). The modal mineralogies are similar to the peridotite xenoliths brought to the surface in the alnöite pipes on Malaita, Solomon Islands (Neal 1986; Neal & Davidson 1988). The model employs non-modal batch melting for all sources, and initial melting ratios for garnet peridotite (36.9: 20: 27.5: 0.6: 15) and spinel lherzolite (34.4: 20: 25: 0.6: 20) both reflect sulphide completely entering the melt at 10% melting. It should be noted that partition coefficients are not available for all of the PGEs in orthopyroxene or garnet. For orthopyroxene, it is assumed that partition coefficients are comparable to clinopyroxene (cf. Ely & Neal 2003). For modelling purposes, the PGEs are assumed to be perfectly incompatible in garnet ($K_D = 0$; where K_D is the bulk distribution coefficient). While this may not be strictly true, Mitchell & Keays (1981) suggested that the PGEs partition very poorly into garnet, and therefore that any contribution to the bulk distribution coefficient would be swamped by those of other phases. A complete list of partition coefficients used in the modeling can be found in Table 2.

PGE partitioning and fractionation

The three different groups of basalts from the OJP have different phenocryst assemblages. The Kroenke-type basalts from Leg 192 are

olivine-phyric, with olivine phenocryst abundances of up to 10 modal% (Sano & Yamashita 2004). These basalts also contain small Cr-spinels that are commonly included in the olivine phenocrysts. Kwaimbaita-type basalts from Leg 192 have significantly less olivine, ranging from 1–3 modal%, but also have occurrences of clinopyroxene and plagioclase phenocrysts, usually less than 2 modal% each. Cr-spinel is typically absent. Singgalo-type basalts, although not recovered by Leg 192, contain only rare olivine phenocrysts, but more commonly are sparsely clinopyroxene–plagioclase-phyric (Petterson 1995; Ely & Neal 2003). Published partition coefficients (Table 2) indicate that Ru and Rh are slightly–moderately compatible in both olivine and clinopyroxene (Table 2). Positive correlations of Ru with Cr and Ni (Fig. 4a, b) support such partition coefficients. Similar positive correlations are also seen with Ir (Fig. 4c, d), and we suspect this is due to the influence of Cr-spinel (cf. Puchtel & Humayun 2001; Richter *et al.* 2004).

The ODP Leg 192 Kroenke-type basalts, with 8–11 wt% MgO and approximately 10 modal% olivine, are obviously less evolved than other OJP basalt types and confirm that spinel was an early fractionating phase. Experimental results indicate that the major-phase fractionation sequence for the leg 192 basalts was olivine, olivine + plagioclase, olivine + clinopyroxene + plagioclase (Sano & Yamashita 2004). Clinopyroxene–plagioclase–titanomagnetite cumulates from the Solomon Islands indicate that additional stages of crystallization may have occurred for some parts of the OJP (Neal *et al.*

1997; Ely & Neal 2003). Thus, a combination of these models was employed in this study. Differentiation was assumed to have been accomplished in five stages of fractional crystallization, each crystallizing 10% of the magma volume (cf. Neal *et al.* 1997). The stages proceed sequentially in the following manner: olivine (100%); olivine (95%) + Cr-spinel (5%); olivine (50%) + plagioclase (50%); olivine (20%) + clinopyroxene (60%) + plagioclase (20%); clinopyroxene (60%) + plagioclase (39%) + titanomagnetite (1%).

Model results

Figure 8a shows the results for the mixed source parental magma and the effect of the fractionation sequence on the PGEs. Variation in the proportions of silicate phases during crystal fractionation has only a minor effect on the PGE model results; the opposite is true for the oxide minerals (Table 2). The primary melt generated by our model falls within the field of the OJP basalts, but cannot account for the range of abundances or Pt/Ru and Pt/Ir ratios, even if the degree of partial melting is varied by 10–15%. Given that the Kroenke-type basalts were fractionating olivine and spinel immediately before eruption, the model clearly underestimates Ir, Ru, Rh and Pt for these lavas. The model Pd abundances are within error of the abundances determined in the Leg 192 samples.

When 1% of outer core material (cf. Snow & Schmidt 1998; see Table 2) is included (Fig. 8b), the primary magma thus generated has PGE abundances of similar magnitude to the most enriched Leg 192 samples. The slight overestimate for Ru is inconsequential, as at least some crystal fractionation must have occurred before eruption of the Kroenke-type samples in order to reduce the MgO abundance to 11 wt% (the highest MgO value; sample 192-1185B-6R-4, 101–120 cm, Unit 5). It is likely that Cr-spinel fractionation will decrease $[Ru/Ir]_{pm}$, based on published partition coefficients (Puchtel & Humayun 2001; see Table 2), which results in a characteristic ‘kick-up’ in Ir on the normalized PGE profiles. It is possible, however, that the ‘kick-up’ on Ir is a result of the sample concentration being close to the LOD for the analysis, resulting in larger uncertainties than for the other PGEs. The fractionated magmas properly estimate the highest Pt abundances for the OJP basalts. Late-stage fractionation of titanomagnetite can also help increase $[Ru/Ir]_{pm}$, resulting in the relatively flat transitions from Ir to Ru seen in some OJP samples (i.e. compare the Kroenke-type basalts to the Kwaimbaita-type

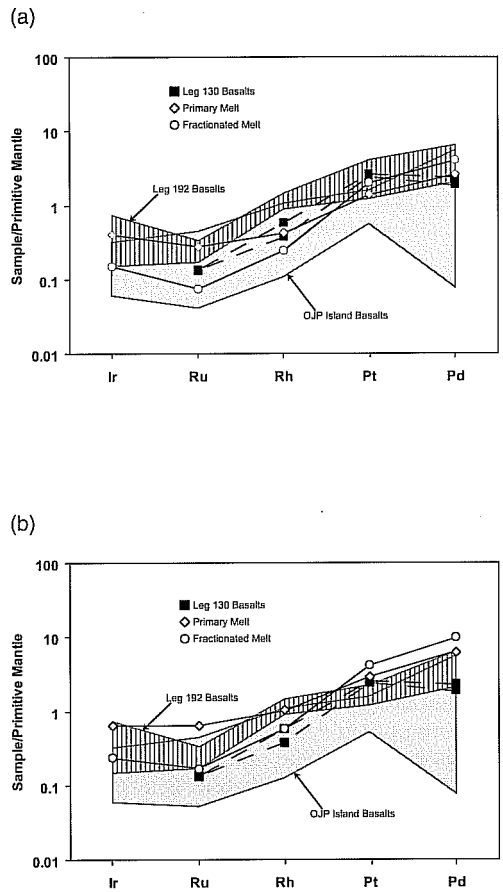


Fig. 8: (a) Modelling results for the PGE using a composite source. The parental melt contains a mixture of melts from a plume component (95%) and a fertile upper-mantle component (5%). The plume component is a 10% melt of a garnet peridotite, which then passes into the spinel stability field and is melted a further 20%. Added to this is a 30% melt of spinel lherzolite of fertile upper-mantle composition. This models a surfacing plume head. Crystal fractionation is accomplished in five stages of 10% of each of the following: olivine; olivine + Cr-spinel; olivine + plagioclase; olivine + clinopyroxene + plagioclase; clinopyroxene + plagioclase + titanomagnetite. For full details, see text. This model approximates the compositions of basalts with relatively low PGE abundances, but fails to generate the basalts with higher abundances. (b) Model using the same parameters as in (a), except for an addition of 1% outer core material to the sources with otherwise primitive mantle composition. All of the PGE abundances in the OJP basalts are successfully modelled, assuming that the outer core material varies from 0 to 1%, and that the degree of fractionation may also vary; the Leg 192 basalts, for instance, only involve three of the five stages of fractional crystallization.

basalts at Site 1185, Hole B; Fig. 3c, d). While we recognize the model-dependence of our results, our modelling is consistent with the conclusion of Ely & Neal (2003): a small amount of an outer core component in the OJP source is necessary to account for the observed PGE abundances in some, but not all, of the erupted basalts. The maximum amount of outer core (1%) is double that proposed in the model of Ely & Neal (2003), as a result of the generally greater PGE abundances of the Leg 192 basalts, especially the Kroenke-type. As only some basalts require this component to account for PGE abundances, it is likely to have been heterogeneously distributed in the otherwise relatively homogeneous OJP source. Inclusion of 1% outer core material (assuming it to be approximately 100% Fe metal) is unlikely to have a definitive effect on Fe abundances in the OJP basalts because, at its maximum, it would contribute about 1 wt% FeO to the magma; OJP basalts analysed as part of this study have Fe (as Fe₂O₃) ranging from 10.3 to 12.6 wt% (Table 1). The model developed above is also consistent with the Re–Os isotopic analyses of Leg 192 basalts. Parkinson *et al.* (2001) concluded that in the limited number of samples they analysed a core component was not evident, at least from Re–Os isotopes. Our modelling demonstrates that *some* of the basalts do not require this outer core component to account for the PGE abundances. While we recognize the dependence of our model on the parameters chosen, it is evident that, overall, that this model successfully accounts for all measured PGE abundances in the OJP basalts from ODP Leg 192, it is consistent with other models that replicate major- and trace-element abundances for the OJP (e.g. Neal *et al.* 1997; Tejada *et al.* 2002), and it does not violate isotopic or petrological considerations. Basically, the PGE concentrations from ODP Legs 130 and 192 basalts are consistent with a heterogeneously distributed outer core component in the OJP source region. The sample from Site 803, which is the only sample analysed from the 90 Ma-event, is in the middle of the range for the OJP and can be adequately modelled without the use of a core component (Fig 8a). While it is difficult to draw the conclusion that the source for the 90 Ma-event contained no core component based on one sample, it may be a worthwhile avenue for future investigation. For example, is the 90 Ma-event the result of a separate plume or is it the result of remelting of the ‘fossil’ plume head beneath the OJP (cf. Richardson *et al.* 2000; Klosko *et al.* 2001)? Within the constraints of

the modelling presented above, the core component within the original plume material would have been exhausted during the c. 122 Ma melting event.

Summary and conclusions

- The OJP basalts form a continuum of PGE compositions and, as such, no discrete groups can be defined on the basis of PGE abundances or ratios as is possible with radiogenic isotope ratios and incompatible trace elements.
- Although the PGE compositions of the OJP basalts are similar even if they are of different ages, the 90 Ma sample from ODP Site 803 can be modelled without the incorporation of a core component.
- Unlike the OJP basalts from ODP Leg 130 and the Solomon Islands, low-temperature alteration does not appear to have affected the Leg 192 basalt Pd abundances.
- The melt that produced the OJP basalts was not saturated with respect to sulphur; therefore, the PGEs exhibit lithophile behaviour with oxide phases exerting the greatest influence.
- At the time of melting, the minimum initial excess temperature of the OJP plume head was +185–+235°C (above ambient), with a minimum temperature range of 1465–1515°C at 3.5–4.0 GPa. These results are consistent with other estimates for OJP plume-head temperatures, including that of Fitton & Godard (>1500°C, 2004). The results of this study are also broadly consistent with the fossil geotherm defined by megacrysts and peridotite xenoliths from alnöite pipes on Malaita, Solomon Islands.
- In order to generate the observed range of PGE abundances in ODP Leg 192 basalts, a small outer core component is required in some but not all samples, suggesting that this component is heterogeneously distributed in the OJP basalt source.

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