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Mantle plume activity and melting conditions: Evidence from olivines in picritic-komatiitic rocks from the Emeishan Large Igneous Province, southwestern China

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Olivine phenocryst compositions and whole-rock chemical compositions are used to identify primitive picritic basalts from the early part of flood basalt successions in the western part of the Emeishan large igneous province (ELIP), southwestern China. The Fo contents of olivine phenocrysts of komatiites with MgO ~21.79 wt% range from 85.2% to 91.4%. The composition of unerupted parental melts is calculated from the compositions of the most Fo-rich olivine phenocrysts. These melts had ~22 wt% MgO, and originated in mantle with a potential temperature of ~1600°C and at about 4.2 GPa pressure, supporting the conclusion that the generation of ELIP can be attributed to melting of a plume head beginning at ~138 km depth.

Introduction

The ~258 Ma Emeishan Province in SW China represents one of the larger Phanerozoic flood basalt events. In recent years it has received considerable international interest because of possible connections between flood volcanism and global mass extinction events (e.g., Wignall, 2001), and a near coincidence of the three major large igneous provinces (LIPs; Emeishan in China, Panjal Traps in India, and Siberian Traps in Russia), which raises important questions concerning global mantle dynamics and the geodynamic setting of the magmatism.

Although the plume model has been widely employed to explain the generation of LIPs, doubts have been raised by some workers about the viability of this model for the ELIP in view of its relatively small exposure (~250,000 km² in comparison with 1,000,000 km² typical of other LIPs; e.g., Coffin and Eldholm, 1994) and the relatively few volcanic cycles of the Emeishan lavas (e.g., Thompson et al., 2001). Anderson (1998) proposed that except for crustal doming immediately preceding eruption or very large volumes of magma (106 km³ of basalt) erupted during a very short time interval (order of 1 Myr), the most convincing argument for a plume would be high magma temperature.

To estimate the magma temperature, it is necessary to place constraints on the primary melt composition. However, few erupted lavas are primary magmas. Highly magnesian volcanic rocks, such as picrites and komatiites, are generally believed to be close to near-

primary mantle-derived melts. Thus, we must seek out such relatively rare lavas. Fortunately, though rare, they are found in many LIPs such as the Karoo (Cox et al., 1984), Deccan (Melluso et al., 1995), West Greenland (Clarke, 1970) and East Greenland (Hansen and Nielsen, 1999).

Recently, we discovered two occurrences of picritic-komatiitic lavas (12–25 wt% MgO, Zhang et al., 2004) in the Lijiang area, Yunnan Province (Figures 1 and 2). Here, we present new results of olivine compositions, and discuss their implications.

Geological setting

The ELIP is a Late Permian continental flood basalt sequence covering large areas in the provinces of Sichuan, Yunnan, and Guizhou along the eastern margin of the Tibetan Plateau to the western margin of the Yangtze Block in southwestern China (Figure 1). The geo-

Figure 1 Sketch map showing outcrops of Emeishan flood basalts (black areas; simplified from Chung & Jahn, 1995). The shaded area marks the location of Figure 2.

Figure 2 Stratigraphic columns showing the picrites-komatiites and basalts exposed at Daju and Shiman in the Lijiang area. Inferred relative stratigraphic locations for samples that were analyzed for major and trace elements (only a few of which were chosen for microprobe analysis) are shown on the right side of each column. Each shaded part represents an individual lava flow or group of flows.

logical setting and occurrence of the Emeishan basalts have been discussed by several researchers (e.g., Chung and Jahn, 1995; Xu et al., 2001; Lo et al., 2002), and we present only a summary here. Magmatism appears to be associated with Late Permian rifting of the Qiangtang terrane from the Yangtze block. The flood basalts were subsequently uplifted, deformed and segmented by block faulting during the closure of the Tethys Ocean in the Jurassic (Chung et al., 1998). The region was reworked further during the collision between the Indian and Eurasian continents (Leloup et al., 2001).

The main volcanic outcrops cover a rhombic-shaped area of 250,000 km² and consist of flood basalts together with lesser volumes of picrites and pyroclastic rocks. Thick flows and tuffs of trachytic and rhyolitic composition are abundant in the upper part of the sequence. Although generally much-reduced by erosion, the lava pile averages 705 m in thickness and locally reaches 5000 m in the western part of the province. The total eruptive volume is estimated to be between 0.3 and 0.5 × 10⁶ km³ (Yin et al., 1992; Jin and Shang, 2000); however, given the large amount of erosion, and by comparison with the other continental flood basalts, this amount may represent less than 50% of the total volume of magma emplaced. The lava successions rest unconformably on the Early Permian Maokou Formation, composed chiefly of marine limestones, and are directly overlain by Late Permian and locally Early Triassic marine strata. A series of layered basic and ultrabasic intrusions, which are associated

with large V-Ti-Fe ore deposits and Cu-Ni-(PGE) ore deposits, are exposed in the central part of the province, apparently underlying the lava pile. These intrusions may represent shallow crustal magma chamber systems that fed the eruptions. Some researchers have determined the age of the ELIP. Lo et al. (2002) concluded from ⁴⁰Ar-³⁹Ar dating that the main phase of flood magmatism occurred at 251-253 Ma, whereas Zhou et al. (2002) obtained U-Pb zircon ages of 256-259 Ma for one of the layered intrusions hosting V-Ti-Fe ores.

Field relations and petrography

Picritic-komatiitic lavas that we discovered recently are located in sections in the southern (Shiman) and northern (Daju) portions of Lijiang county, Yunnan province; the two sections are about 26 km apart. The geological characteristics of the two picrite-komatiite occurrences are similar. In both sections, the picritic-komatiitic lavas are located in the lower part of the flood basalt pile. In the Shiman section, three picritic-komatiitic flows are intercalated with pyroxene-phyric basaltic lavas, whereas four picritic-komatiitic flows occur in the lower part of the Daju section (Figure 2). It is possible, perhaps likely, that the Shiman section corresponds to the middle part of the much thicker Daju section. Massive aphyric basalts and amygdaloidal basalts are dominant in the middle to upper part of the Daju sequence (Figure 2). In the Shiman section, the lowermost and the uppermost picritic-komatiitic flows are 3 to 5 m thick, and the middle picritic-komatiitic layer is 15 to 20 m thick. In the Daju section, the lowermost picritic-komatiitic flow varies from 20 to 50 m in thickness, whereas the thickness of each of the other three flows is similar to those in the Shiman section. Unlike the central and eastern parts of the ELIP, where plagioclase-phyric basalts are widely exposed, plagioclase-phyric basalts are absent in the Lijiang region.

Most of the picritic-komatiitic rocks are highly porphyritic, and contain abundant forsteritic olivine phenocrysts with minor clinopyroxene and Cr-spinel. Olivine phenocrysts are generally subhedral to rounded and sometimes embayed, ranging from 0.2 to 4 mm in size; some appear partly resorbed. Strained, kink-banded crystals are absent. However, olivine is generally completely replaced by serpentine in most picritic-komatiitic samples. Only in individual sample SM-15 are unaltered cores preserved (Figure 3). Olivine crystals usually contain scattered inclusions of glass; some enclose small, equant, euhedral to rounded, dark-brown Cr-spinel crystals, usually tens of micrometers across. Some Cr-spinel crystals are also "free" in

Figure 3 Photomicrograph of komatiite from the Lijiang area. Olivine crystals have rounded or embayed outlines and rims, which are interpreted as a result of resorption. The matrix consists of olivine, optically intergrown clinopyroxene and plagioclase and microcrystalline areas of oxide-rich mesostasis. The microphotograph was taken in plane-polarized light.

the groundmass, which consists of olivine, augite and plagioclase. Quenched olivine groundmass crystals may be tiny and near-equant 'hoppers' but are often elongated, up to 0.5 mm long and 0.1 mm wide, and skeletal to highly skeletal, consisting of parallel-set rods and blebs in optical continuity. Groundmass plagioclase occurs as microcrystals rather than the long platy shapes seen in the basalts, and clinopyroxene is anhedral. In addition, minor orthopyroxene phenocrysts are observed in some highly magnesian picrites (~25 wt% MgO, samples DJ-34-1 and DJ-31, which are strictly komatiites, see the IUGS definition of Le Bas, 2000), which are rarely found in picrites in other LIPs.

The pyroxene-phyric basalts are also porphyritic, with diopside phenocrysts 1–6 mm across; some form glomerocrysts. Phenocrysts account for 5 to 15 % of the volume of the basalts. Groundmass is fine grained with intersertal or intergranular textures, and consists of plagioclase and clinopyroxene with minor iron oxides.

The massive aphyric basalts contain a similar mineral assemblage. Clinopyroxene is the most abundant microphenocryst. The phenocrysts that are present commonly form clots, producing a cumulo-phyric texture. The pyroxene grains are usually small, generally tens of micrometers across. The groundmass comprises plagioclase laths with interstitial sub-ophitic pale-brown clinopyroxene and minor anhedral magnetite.

Composition of minerals

The electron microprobe-analyzed results of representative olivine (core), pyroxene and Cr-spinel in the picrites-komatiites and associated basalts are listed in Table 1.

The olivines are denoted by their *mg-number*, i.e., atomic 100Mg/(Mg+Fe). Their *mg-numbers* span a compositional range from 85.2%–91.4%, with most of the analyses in the range from 89–90 (Table 1). The Mg-rich olivines usually contain inclusions of chromite and glass. They are visually indistinguishable from similar olivine crystals with *mg-number* <90. The Mg-rich olivines normally constitute about 15 percent of a thin section. All olivines contain small amounts of CaO (0.22–0.42 wt%), NiO (0.27–0.44 wt%), CoO (0.01–0.21 wt%) and Cr₂O₃ (0.03–0.17 wt%) besides MgO, FeO and SiO₂.

Olivines with *mg-number* >90 are a common feature of komatiites (e.g., Arndt et al., 1977; Leshner, 1989), and are also reported from picrites in some LIPs (e.g., Larsen and Pedersen, 2000). They are only rarely reported from mid-ocean ridge basalt (MORB) (Donaldson and Brown, 1977), ocean island basalt (OIB) or continental basalts. On Hawaii, the well-analyzed olivine populations do not exceed *mg-number* 90.7 (Maaloe and Hansen, 1982; Yang et al., 1994; Baker et al., 1996).

The compositional range in chromite is very small, with *mg-number* (Mg/(Mg+Fe)) 26.0–45.9, Cr-number (Cr/(Cr+Al)) 71.5–75.2 and TiO₂ >1 wt%, and high contents of total iron (FeO_T=22–33 wt%). They are broadly similar to some Cr-spinels from Archaean komatiites, which have high Fe and Cr and low Mg and Al (Arndt et al., 1977; Nisbet et al., 1977; Leshner, 1989), but the distinct from those from most abyssal peridotites (Dick and Bullen, 1984) or some continental flood basalts, e.g., West Greenland (Larsen and Pedersen, 2000).

The clinopyroxene is characterized by enrichment in Ti, with TiO₂ contents more than 1 wt%, up to 2.85%. The average TiO₂ contents of clinopyroxene in the basaltic rocks and picritic-komatiitic rocks are 1.32 wt% and 1.22 wt%, respectively. Moreover, the TiO₂ contents of

clinopyroxene increase overall with decreasing MgO content. In addition, CaO and MgO content in clinopyroxene are high, belonging to Ti-bearing augite with *mg-number* 0.72–0.86. However, the phenocrysts of the pyroxene-phyric basalts are diopside whereas the clinopyroxene in the basalts from other regions of ELIP is augite (Wang et al., 1993).

Discussion and conclusions

The Mg-rich olivines with *mg-number* >90 contain 0.22–0.39 wt% CaO and 0.05–0.17 wt% Cr₂O₃. This shows that the Mg-rich olivines crystallized from melts. They are not xenocrysts of mantle olivines, which have very low contents of Ca and Cr, less than 0.1 wt% CaO and 0.06 wt% Cr₂O₃, respectively (Gurenko et al., 1996; Thompson and Gibson, 2000). Further, the continuous trends through the whole olivine compositional range strongly suggest that the Mg-rich olivine crystals are cognate phenocrysts, and not accidental xenocrysts.

Here, we focus on the most Mg-rich olivines and their implication. It is widely agreed that rocks with picritic characteristics can arise in a variety of ways; for example, as primitive picritic liquids that are little-modified melts of upper-mantle peridotite, by accumulation of early formed olivine from such primary picritic liquids, or by accumulation of olivine from 'normal' basaltic magmas. Under equilibrium conditions, olivine compositions will reflect the composition of the magma from which they crystallize; thus the composition of the olivine phenocrysts in picrites is a valuable clue to estimate the composition of the parental magma. This approach has been used by many workers (e.g. Larsen and Pedersen, 2000; Thompson and Gibson, 2000; Simkin et al., 1970; Breddam, 2002; Revillon et al., 1999; Green et al., 2001).

In Figure 4, olivine compositions are plotted as a function of the MgO contents of the whole rocks. Also shown are curves representing the composition of olivine in equilibrium with liquids with 8, 9, 10, 11, 12 and 13% FeO. These values correspond to FeO content of likely parental magmas (Figure 4). If the most magnesian olivine from a sample plots on an olivine-liquid equilibrium curve, this indicates that the MgO content of the whole rock is that of the liquid from which the olivine crystallized; if the olivine plots below the curves, this indicates that the rocks have accumulated olivine (Révillon et al., 1999). The sample SM-15, containing the most Fo-rich olivine phenocrysts, has ~21.79 wt% MgO and 12.34 wt% FeO. Assuming Fe³⁺/Fe²⁺=0.15, the FeO content of sample SM-15 is calculated to be ~10.73 wt%.

Both cases are represented by the Emeishan picritic-komatiitic rocks. The most magnesian olivine, containing 91.4% forsterite, plots close to the 10.7 wt% FeO curve. From Figure 4, we estimate

Figure 4 MgO of whole rocks versus Fo contents of olivine in picritic-komatiitic rocks. The curves represent the MgO content of liquids in chemical equilibrium with olivine for a K_d of Fe/Mg between crystal and magma = 0.31 (Roeder and Emslie, 1970) for magma with FeO = 8 to 13 wt%.

Table 1 Representative olivine, pyroxene and Cr-spinel analyses of picrites and associated basalts of the ELIP, SW China (all data in wt%).

Sample	Minerals	SiO ₂	TiO ₂	Al ₂ O ₃	FeOt	MnO	MgO	CaO	Na ₂ O	Cr ₂ O ₃	NiO	CoO	V ₂ O ₅	Sum	Mg#	Cr#	End member
SM-15	Ol Phenocryst	41.04	0.02	0.04	9.08	0.00	50.12	0.28	0.04	0.16	0.43			101.35	0.908		Fo90.8Fa9.2
	Ol Phenocryst	40.57	0.05	0.04	9.40	0.10	49.62	0.30	0.04	0.13	0.44			100.70	0.904		Fo90.4Fa9.6
	Ol Phenocryst	39.99	0.03	0.05	14.14	0.18	45.99	0.30	0.00	0.07	0.38			101.13	0.853		Fo85.3Fa14.7
	Ol Phenocryst	39.16	0.00	0.06	13.92	0.18	45.05	0.31	0.00	0.09	0.42			99.19	0.852		Fo85.2Fa14.8
	Ol Phenocryst	39.97			8.47	0.09	50.46	0.35		0.06	0.41	0.07		99.61	0.914		Fo91.4Fa8.1
	Ol Phenocryst	40.11			8.58	0.08	50.27	0.29		0.05	0.40	0.21		100.00	0.913		Fo91.3Fa8.7
	Ol Phenocryst	40.04			8.49	0.15	50.22	0.31		0.09	0.43	0.01		99.73	0.914		Fo91.4Fa8.6
	Ol Phenocryst	40.19			8.75	0.11	50.02	0.29		0.06	0.37	0.02		99.80	0.911		Fo91.1Fa8.9
	Ol Phenocryst	40.12			8.77	0.06	50.02	0.30		0.17	0.37	0.21		100.01	0.910		Fo91.0Fa9.0
	Ol Phenocryst	40.18			9.88	0.01	48.98	0.28		0.01	0.36	0.13		99.83	0.899		Fo89.9Fa10.1
	Ol Phenocryst	39.70			10.07	0.28	48.77	0.42		0.08	0.28	0.21		99.81	0.897		Fo89.7Fa10.3
	Ol Phenocryst	39.94			10.49	0.21	48.54	0.22		0.10	0.27	0.20		99.98	0.893		Fo89.3Fa10.7
	Ol Phenocryst	39.88			10.79	0.16	48.23	0.30		0.10	0.33			99.78	0.889		Fo88.9Fa11.1
	Ol Phenocryst	39.21			12.21	0.24	47.40	0.33		0.09	0.40			99.88	0.875		Fo87.5Fa12.5
	Ol Phenocryst	39.02			12.70	0.24	46.95	0.39		0.03	0.36	0.05		99.75	0.869		Fo86.9Fa13.1
DJ-31	Ol Phenocryst	37.78	0	0.03	12.45	0.13	48.91	0.25	0	0.06				99.61	0.875		Fo87.5Fa12.5
	Ol Phenocryst	38.12	0	0.05	12.86	0.15	47.58	0.32	0	0.05				99.13	0.868		Fo86.8Fa13.2
	Ol Phenocryst	37.46	0.01	0.04	13.82	0.15	47.08	0.38	0.02	0.06				99.02	0.858		Fo85.8Fa14.2
DJ-26	Ol Phenocryst	37.89	0	0.03	12.09	0.14	49.00	0.27	0	0.06				99.48	0.878		Fo87.8Fa12.2
	Ol Phenocryst	37.97	0.01	0.04	13.3	0.14	47.39	0.29	0.03	0.06				99.23	0.863		Fo86.3Fa13.7
	Ol Phenocryst	37.86	0.02	0.03	14.51	0.15	46.03	0.45	0.02	0.04				99.11	0.85		Fo85.0Fa15.0
DJ-34-1	Ol Phenocryst	38.45	0.01	0.04	10.89	0.11	50.12	0.23	0.05	0.08				99.98	0.891		Fo89.1Fa10.9
	Ol Phenocryst	37.71	0.02	0.08	14.22	0.14	46.25	0.41	0	0.06				98.9	0.852		Fo86.2Fa14.8
	Ol Phenocryst	37.43	0.01	0.04	8.99	0.09	52.46	0.22	0	0.14				99.38	0.912		Fo91.2Fa8.8
	Ol Phenocryst	37.92	0	0.05	10.34	0.12	50.79	0.30	0	0.11				99.63	0.897		Fo89.7Fa10.3
DJ-33	Cpx Phenocryst	49.69	2.00	4.43	7.74	0.19	11.66	22.34	0.18					98.47	0.729		Fs13.5En36.3Wo50.1
	Cpx Phenocryst	50.46	1.41	4.60	6.35	0.07	13.26	22.51	0.27					99.34	0.788		Fs10.8En40.2Wo49.1
	Cpx Phenocryst	52.09	0.81	2.72	4.53	0.00	14.13	23.64	0.18					98.38	0.848		Fs7.5En41.9Wo50.5
	Cpx Phenocryst	47.68	2.85	7.16	7.45	0.09	10.8	22.49	0.12					98.84	0.721		Fs13.4En34.7Wo51.9
	Cpx Phenocryst	50.32	1.17	4.10	5.77	0.11	13.74	23.19	0.06					98.84	0.809		Fs9.6En40.8Wo49.6
	Cpx Phenocryst	51.65	0.94	3.65	5.07	0.11	13.98	22.78	0.20					98.77	0.831		Fs8.6En42.1Wo49.4
	Cpx Phenocryst	51.68	0.86	2.80	4.28	0.03	14.55	23.54	0.12					98.08	0.858		Fs7.1En42.9Wo50.0
DJ-34	Cpx Phenocryst	52.49	0.76	2.02	4.34	0.09	14.25	23.09	0.27					97.56	0.854		Fs7.3n42.8Wo49.9
	Cpx Phenocryst	52.87	1.17	4.89	4.35	0.00	14.43	19.93	0.27					98.14	0.855		Fs7.8En46.2Wo46.0
	Cpx Phenocryst	53.07	1.03	4.12	4.10	0.00	15.42	20.42	0.16					98.70	0.870		Fs7.1En47.6Wo45.3
	Cpx Phenocryst	52.64	1.31	5.36	4.27	0.00	14.57	20.07	0.31					98.86	0.859		Fs7.6En46.4Wo46.0
	Cpx Phenocryst	52.08	1.56	6.25	5.04	0.04	13.31	19.57	0.27					98.41	0.825		Fs9.4En44.0Wo46.6
	Cpx Phenocryst	55.36	0.45	1.82	2.61	0.07	16.98	20.34	0.00					98.15	0.921		Fs4.4En51.3Wo44.2
SM-15	Cpx Quench crystal	54.18	0.69	2.90	5.04	0.13	17.49	17.86	0.42					98.90	0.861		Fs8.5En52.7Wo38.8
	Cpx Quench crystal	53.57	1.02	3.13	8.30	0.16	16.90	16.31	0.00					99.52	0.784		Fs14.5En50.8Wo35.3
	Cpx Quench crystal	53.09	1.12	4.44	5.44	0.11	16.08	17.68	0.16					98.29	0.840		Fs9.6En50.5Wo39.9
	Cpx Quench crystal	53.16	1.06	2.94	9.01	0.32	16.83	15.73	0.09					99.29	0.769		Fs15.2En50.7Wo34.1
	Cpx Quench crystal	50.03	1.20	3.41	6.93	0.18	15.898	21.50	0.23	0.24	0.10	0.18	0.226	100.12	0.803		Fs11.0En45.1Wo43.9
	Cpx Quench crystal	50.16	1.05	3.32	6.96	0.20	16.10	21.58	0.04	0.16	0.08	0.20	0.22	100.06	0.805		Fs11.0En45.3Wo43.7
	Cpx Quench crystal	50.01	1.08	3.54	7.17	0.23	15.87	21.42	0.07	0.26	0.04	0.06	0.17	99.90	0.798		Fs11.4En44.9Wo43.7
	Cpx Quench crystal	48.06	1.68	4.38	9.12	0.28	14.70	20.66	0.44	0.41	0.02	0.09	0.10	99.93	0.742		Fs14.7En42.4Wo42.9
	Cpx Quench crystal	46.98	2.52	5.34	9.92	0.14	14.41	20.02	0.23	0.24	0.06	0.11	0.26	100.23	0.721		Fs16.2En41.9Wo41.9
	Cpx Quench crystal	48.83	1.59	4.52	7.17	0.15	15.65	21.27	0.14	0.28		0.17	0.15	99.91	0.796		Fs11.5En44.7Wo43.8
DJ-31	Opx Phenocryst	52.56	0.00	3.52	12.45	0.20	31.15	1.10	0.00	0.21				99.59			Fs17.9En80.0Wo2.0
	Opx Phenocryst	47.56	0.00	6.97	12.09	0.17	33.92	0.22	0.00	0.11				99.42			Fs16.6En83.0Wo0.4
DJ-34-1	Opx Phenocryst	49.71	0.04	9.26	8.99	0.09	31.38	1.14	0.00	0.39				99.34			Fs13.5En84.3Wo2.2
	Opx Phenocryst	50.62	0.00	6.57	10.34	0.22	30.55	2.26	0.00	0.12				99.58			Fs15.3En80.4Wo4.3
SM-15	Sp Phenocryst	0.07	1.94	10.26	30.36	0.54	8.95	46.39	0.29	0.30	0.25			99.36	0.345	0.752	
	Sp Phenocryst	0.08	1.96	10.41	31.39	0.45	8.42	46.82	0.18	0.10	0.18			99.98	0.323	0.751	
	Sp Phenocryst	0.03	1.95	11.27	32.82	0.64	6.46	45.59	0.09	0.26	0.26			99.37	0.260	0.731	
	Sp Quench crystal	0.10	1.20	13.41	21.80	0.23	13.19	0.00	0.01	50.07	0.14			100.15	0.410	0.715	
	Sp Quench crystal	0.11	1.85	10.95	30.05	0.26	10.37	0.00	0.02	44.76	0.20			100.01	0.459	0.733	

Note: Microprobe analyses were carried out at Cardiff University. Relative analytical uncertainty is <1% for major elements and <16% for minor elements. SM-15, DJ-34-1 and DJ-31 are komatiites, and DJ-34 and DJ-33 are pyroxene-phyric basalts. Mg# = Mg/(Mg+Fe); Cr# = Cr/(Cr+Al)

that the MgO content of the parental liquid was ~22 wt%. Therefore, sample SM-15, with MgO = 21.79 wt% (Zhang et al., 2004), can represent the composition of parental magma. Some komatiitic rocks with MgO contents ~25 wt% (DJ-31 and DJ-34-1) formed through the accumulation of olivine that had crystallized from less magnesian liquids, which is consistent with the fact that they contain some orthopyroxene, which crystallizes relatively late in evolved magmas.

Highly magnesian volcanic rocks, particularly picritic-komatiitic rocks, have been used to estimate mantle melting conditions and mantle compositions because their composition may be close to near-primary mantle-derived melts.

Determining the potential temperature (T_p) can be approached in three main ways: comparison of natural melt compositions with melts produced experimentally from a suitable source under known, realistic, pressure and temperature conditions (e.g., Albarede, 1992); mathematical inversion of erupted melt compositions assuming fractional melting in a polybaric melt column using an assumed mantle source and an experimentally determined pressure- and temperature-dependent sub-solidus assemblage (McKenzie and O'Nions, 1991); and comparison of natural melt compositions and forward melt modelling of idealized mantle sources (e.g., Fram and Leshner, 1993; Thirlwall et al., 1994).

We use the first technique to estimate the potential temperature (T_p) and magma segregation pressure. As sample SM-15 represents a primary magma composition, a hypothetical "eruptive temperature" at 1 atm and 1450°C for such a melt, can be obtained based on the equation of Nisbet et al. (1993) $T(^{\circ}\text{C}) = 17.86 \times \text{MgO}(\%) + 1061^{\circ}\text{C}$. A melt fraction for the 22% MgO komatiite is estimated to be, $X = 31.67\%$ on the basis of experimental evidence (Takahashi et al., 1993; Herzberg and Zhang, 1996; Falloon et al., 1999). Plotting this point on a pressure-temperature diagram for peridotite KLB-1 mantle (Takahashi et al., 1993; Herzberg and Zhang, 1996), the value of T_p of KLB-1 mantle required to generate this melt fraction by anhydrous decompression can be estimated by the method of McKenzie and Bickle (1988). The result is $T = 1600^{\circ}\text{C}$ with dry melting beginning at 4.2 GPa (Figure 5). In other words, the primary magma was generated at 138 km depth. Using the same method, Thompson and Gibson (2000) calculated Hawaiian plume T_p to be 1525°C . Arndt (2000) argued that such a calculation could cause an error of about 100°C (higher or lower), because it is affected by some uncertain factors, e.g., pressure, oxygen fugacity,

Figure 5 Temperature-pressure diagram showing the melting of anhydrous mantle peridotite KLB-1 (after Thompson and Gibson, 2000). The vertical coordinate is the temperature at 1 atm. Dotted line represents melt fraction. The adiabats for representative mantle potential temperatures are marked as solid lines. Estimates for picrites from Horingbaai in Namibia (square) and Mauna Loa (solid circle) in Hawaii are shown for comparison. Open circle is for Emeishan komatiite SM-15.

water, etc. It should be noted that hydrous minerals such as amphibole and biotite are absent in the interstices of coarsely holocrystalline komatiite. In this respect, the samples contrast with the meimechites of the Siberian LIP, which have ~2.5 wt% MgO and Fo93 olivines, but with interstitial hydroxyl-biotite (Arndt et al., 1997). It therefore can be inferred that water is absent in the primary magma or there is little water in the magma, although a small water content would change the calculated temperature. Thus, the error should be lower than 100°C .

The mantle potential temperature that the Mg-rich melt implies ($\sim 1600^{\circ}\text{C}$) is about 75°C more than the present-day values for Hawaii and Iceland plumes (White and McKenzie, 1995; Watson and McKenzie, 1991; Ribe and Christensen, 1999; Shen et al., 1998), and an increase of $\sim 250\text{--}320^{\circ}\text{C}$ above the global normal potential asthenospheric mantle temperature of $1280\text{--}1350^{\circ}\text{C}$ (McKenzie and Bickle, 1988). This suggests that the generation of the primary magma was related to a mantle plume, because normal mantle does not reach such high temperature. A high magmatic temperature is also supported by the high Cr contents of the spinel (cf. Larsen and Pederson, 2000). Hence, we infer that the primary magma was generated by decompression melting of a mantle plume head impinging on the base of the lithosphere at about 138 km depth.

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