Supra-subduction Zone Pyroxenites from San Jorge and Santa Isabel (Solomon Islands)

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Peridotites associated with pyroxenites (with rare olivine and spinel) are exposed on the islands of San Jorge and Santa Isabel in the Solomon Islands. Orthopyroxenite occurs in large outcrops ($\leq 100\ m^2$) whereas websterite and clinopyroxenite occur as layers and veins/dykes in peridotites. The bulk compositions of the pyroxenites are characterized by high $Mg^{2+}/(Mg^{2+} + Fe^{2+})$ (0·78–0·91) and low Al_2O_3 (<2·7 wt %). Low rare earth element abundances are coupled with large ion lithophile element enrichments and positive Sr and Pb anomalies (primitive mantle-normalized) relative to adjacent rare earths. Temperatures of equilibration for the pyroxenites are between 950 and 1050°C. These relatively low temperatures, combined with the occurrence of primary fluid inclusions, suggest that the pyroxenites formed by interaction of peridotite protoliths with an aqueous fluid. Bulk-rock and mineral compositions of the orthopyroxenites are similar to those of mantle-derived pyroxenites, whereas the websterites have closer chemical affinity with crustal arc cumulates. Nevertheless, field relationships plus petrological, textural and geochemical evidence are consistent with formation of all pyroxenite types in supra-subduction zone mantle, resulting from metasomatism of peridotite by subducted Pacific Plate-derived fluid. Such a setting for pyroxenite has not previously been reported in detail. We propose that these processes produce mantle pyroxenite with compositions similar to crustal pyroxenite.

KEY WORDS: *mantle metasomatism; pyroxenite; supra-subduction zone*

INTRODUCTION

Pyroxenites are ultramafic rocks that generally occur in two distinct settings: (1) as veins/dykes in various mantle peridotite lithologies (including xenoliths and ultramafic complexes) on a variety of scales; (2) as cumulates formed in crustal magma chambers. Orthopyroxenite, websterite and clinopyroxenite probably represent <5% of the upper mantle (Fabriès *et al.*, 1991; Hirschmann & Stolper, 1996). Crustal pyroxenites, on the other hand, are dominantly cumulate clinopyroxenites and are commonly found in arc environments (Schiano *et al.*, 2000) as relatively large volumes of cumulates from primitive island arc magmas in crustal magma chambers (DeBari & Coleman, 1989; Schiano *et al.*, 2000). However, little is known about possible pyroxenites in the mantle wedge between the subducted slab and the base of the arc lithosphere. On the basis of peridotite xenoliths found in Lihir and Tubaf (volcanoes located to the rear of the New Britain–Bougainville arc of Papua New Guinea), McInnes *et al.* (2001) suggested that pyroxenites might be common in supra-subduction zone (SSZ) settings, but detailed studies of coherent bodies have not been reported.

Several processes have been proposed for the genesis of these different types of pyroxenite. Field, geochemical and experimental studies have shown that clinopyroxene can be a liquidus phase in various basalt magma types,

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Fig. 1. Geological terrane model of the Solomon Islands after Petterson *et al.* (1999), constructed on the basis of distinctive lithologies, ages and geochemistry of basement sequences, and the existence of arc sequences. At least two active volcanoes (Kana Keoki and Coleman) are located within the area denoted as New Georgia Forearc. Location of Fig. 2 is shown by the rectangle.

and high-pressure mineral segregation from primitive magmas can result in the formation of massif pyroxenite (DeBari & Coleman, 1989; Girardeau & Ibarguchi, 1991; Pearson *et al.*, 1993; Müntener *et al.*, 2001; Parlak *et al.*, 2002; Santos *et al.*, 2002). Alternatively, recycling of elongated slices of subducted oceanic lithosphere through the convective mantle may also result in pyroxenite formation (Polvé & Allègre, 1980; Allègre & Turcotte, 1986; Hamelin & Allègre, 1988; Kornprobst *et al.*, 1990; Kerr *et al.*, 1995). Although these two processes can explain a wide spectrum of occurrence, they are unable to explain crosscutting networks of pyroxenite veins in mantle peridotites. In such settings, many researchers have proposed pyroxenite originates from melt/fluid–rock reaction in the upper mantle (Kelemen *et al.*, 1992; Smith & Riter, 1997; Wilkinson & Stolz, 1997; Garrido & Bodinier, 1999; Smith *et al.*, 1999; Zanetti *et al.*, 1999; McInnes *et al.*, 2001; Wang *et al.*, 2001).

In this paper, we combine a field study with petrological, mineralogical and geochemical investigations in order to characterize pyroxenites exposed on the islands of San Jorge and Santa Isabel (Solomon Islands). The pyroxenites crop out in a boundary zone between the Ontong–Java Plateau (OJP) and the Solomon arc. Our primary aim is to determine whether these pyroxenites formed in an SSZ mantle setting or beneath the OJP. We document the mineralogical and chemical characteristics of the San Jorge and Santa Isabel pyroxenites, and compare these with typical mantle and crustal pyroxenites. We then discuss the processes responsible for the formation and the exhumation of pyroxenite in the Solomon Islands specifically.

GEOLOGICAL SETTING

The location of the Solomon Islands in the SW Pacific is shown in Fig. 1. The Solomon Islands comprise a linear, NW–SE-trending double chain of islands including as major islands in the northeastern chain Choiseul, Santa Isabel and Malaita, and in the southwestern chain the New Georgia Group with the adjacent volcanically active forearc region, Guadalcanal and Makira (Petterson *et al.*, 1999) (Fig. 1). The Solomons are bounded by two trench systems: the fossil Vitiaz Trench to the NE, and the South Solomon Trench System (SSTS) in the SW, marking the

Fig. 2. Simplified geological map of the SE portion of Santa Isabel and the island of San Jorge after Hawkins & Barron (1991). Italicized numbers next to the field locations are samples described in the text. KKKFS, Kia–Korigole–Kaipito Fault System.

present collision zone between the Australian and Pacific plates. The Vitiaz Trench in the Solomons region is now blocked by the OJP, and the complicated geological history of the Solomon Islands is caused, in part, by the collision between the Plateau and the Solomon arc (Petterson *et al.*, 1997; Mann & Taira, 2004).

The Alaska-size OJP is the largest-known oceanic plateau (estimated volume $>5 \times 10^7$ km^3) (Coffin & Eldholm, 1994), and is believed to have formed from a mantle plume (Ely & Neal, 2003). From the time of the OJP–Solomon arc contact in the Pliocene, according to Cowley *et al.* (2004), subduction of the Pacific Plate terminated progressively along the Vitiaz Trench from north of Bougainville through the Solomons and Vanuatu (Mann *et al.*, 1996; Cowley *et al.*, 2004). In response to this collision, subduction of the Indo-Australian Plate northeastwards along the SSTS was initiated (Fig. 1).

On the basis of different lithological assemblages, the Solomon Islands have been subdivided into several provinces (Coleman, 1965) or terranes (Petterson *et al.*, 1999). Those terranes with a major, Cretaceous-aged, component comprise, according to Petterson *et al.* (1999) (Fig. 1): (1) a plume-related (OJP) terrane including Malaita, Ulawa and the northeastern part of Santa Isabel; (2) a mid-ocean ridge basalt (MORB) terrane including much

of Choiseul and Guadalcanal; (3), a hybrid terrane including both MORB and plume components forming Makira. Two stages of Tertiary arc development are recognized: Stage 1 from the Eocene to Miocene, including the Shortland Islands and southern part of Santa Isabel, shown as Terrane 4 in Fig. 1; Stage 2 from the Late Pliocene/Early Pleistocene to the present day, including the active volcanoes of the New Georgia Group (Rohrbach *et al.*, 2005) and adjacent forearc, shown as Terrane 5 in Fig. 1.

As a result of the OJP–Solomon arc collision, the leading edge of the OJP began to be exhumed and portions are now exposed in the islands of Santa Isabel, Malaita, Ulawa and Makira (Petterson *et al.*, 1997). These uplifted, overthrust, largely unmetamorphosed portions of the OJP were termed the Pacific Province by Coleman (1965). Thin fault-bounded slices of peridotite associated with pyroxenite occur on the SW margin of the exposed OJP on the southern part of Santa Isabel and the island of San Jorge (Fig. 2).

THE GEOLOGY OF SANTA ISABEL AND SAN JORGE

The geology of Santa Isabel and San Jorge is shown in Fig. 2. At the southern end of Santa Isabel, a major

portion of the outcrop comprises Sigana Basalts; these are pillow lavas of the OJP. They comprise low-K tholeiites that are petrologically and geochemically similar to lavas recovered by deep-sea drilling of the OJP (Neal *et al.*, 1997). The Sigana Basalts are separated from Terrane 4 by a major suture—the Kia–Korigole–Kaipito Fault System (KKKFS) (Stanton, 1961), which is known through seismic survey to extend offshore (Cowley *et al.*, 2004). Along the KKKFS, thin fault-bound slices of peridotite are distributed on both sides of the fault system (Fig. 2). The ultramafic rocks consist of a series of elongate pods of more or less serpentinized harzburgite and dunite, cut by pyroxenite veins. However, the relationships between these ultramafic rocks and other rock formations (particularly the OJP basalts) are generally not well exposed. Only in two locations (Tanabuli Island and Kolare Island) is a small faulted contact between peridotite and OJP basalts exposed, perhaps indicative of a lack of genetic relationships.

On San Jorge, pillowed and massive flows of basaltic to andesitic lavas (the San Jorge Volcanics) are associated with micro- to medium-grained gabbros (the Kolose'eru Gabbros). All of these rock types are part of the Jajao Igneous Suite (Tejada *et al.*, 1996), with ages in the range 62–46 Ma. They are petrographically, chemically and isotopically distinct compared with the OJP Sigana basalts, and are suggested to represent fragments of uplifted ocean floor formed in an arc–back-arc system (Tejada *et al.*, 1996). In the region of Loguhutu Bay (the southern coast of San Jorge), they are exposed in fault contact with a large body of pyroxenite-veined peridotite (Fig. 2).

The northern part of San Jorge consists of a large domal mass of brecciated and flow-banded serpentine and serpentinous mud, analogous to the diapirs documented along the length of the Izu–Bonin–Mariana forearcs by Fryer *et al.* (1985). This serpentinite dome encloses blocks of very fresh pyroxenite as well as fresh to variably altered harzburgite at Rata Bay (Fig. 2).

PERIDOTITE–PYROXENITE FIELD RELATIONS

The San Jorge and Santa Isabel websterites and clinopyroxenites generally occur as layers and veins/dykes with sharp contacts in host harzburgites and dunites (Fig. 3a). No lherzolites have been found. The harzburgites and dunites are variably serpentinized and some are deformed. However, the pyroxenite veins remain surprisingly fresh, varying from orthopyroxene- to clinopyroxene-rich in composition. The pyroxenite layers are generally subparallel to the foliation of their host ultramafic rocks. Conversely, the veins/dykes are organized in a multi-generational network crosscutting the host peridotites (Fig. 3b). The thickness of these pyroxenites

is variable, ranging from ∼1 cm to ∼30 cm. Veins or dykes are occasionally folded and sheared.

On Santa Isabel, the pyroxenites are exposed on both sides of the KKKFS. The clinopyroxene-rich rocks, including websterites and clinopyroxenites, seem to be more abundant than the orthopyroxene-rich samples. On San Jorge, the pyroxenites occur in two locations: at the southern end of the island, where they are incorporated into an ultramafic massif, and in northern San Jorge, where they crop out in a serpentine diapir. At Johnson Point (Fig. 2), the orthopyroxenite outcrops over an area of about 100 m^2 without any clear relation with the neighbouring peridotite. Websterite is abundant as beach cobbles and pebbles, particularly in the vicinity of the southern ultramafic massif.

ANALYTICAL METHODS
Whole-rock analysis

Whole-rock analyses of 18 representative pyroxenites, sampled from different locations on Santa Isabel and San Jorge, were carried out using X-ray fluorescence (XRF) to determine their major element compositions. Weathered surfaces from samples were removed and then the samples were crushed into fine chips and finally pulverized to a <10 μm powder in an agate mill. Powders were fused into glass discs following the standard procedures employed for XRF analysis (Norrish & Hutton, 1969). All the analyses of Li-borate fusion discs were conducted using a Phillips PW2400 spectrometer for major elements according to the method of Norrish & Hutton (1969). Loss on ignition (LOI) was determined separately by heating at 1100°C for 1 h.

Trace element concentrations were measured with two types of inductively coupled plasma mass spectrometry (ICP-MS) techniques. The solution ICP-MS VG PlasmaQuad (PQ Turbo Plus) of the University of Montpellier (France) has been tuned and calibrated according to Ionov *et al.* (1992) with special care for minimizing memory effects during the determination of very low trace element contents. BIR-1 was the standard used. Sample preparation for the solution ICP-MS was carried out in a 'cleanroom' following the procedures of Barrat *et al.* (1996). A laser ablation (LA) ICP-MS system, consisting of an ArF (193 nm) EXCIMER laser coupled with an Agilent 7500s ICP-MS system, was used to analyse trace element concentrations on fused discs, following the protocol of Eggins *et al.* (1997). Fused discs made for XRF purposes were glued together in stacks of 10, set in epoxy resin blocks, displayed as a sandwich in cross-section, polished, and the laser beam (100 μm in diameter) rastered along the length of each disc. The trace element concentrations obtained using the two types of ICP-MS techniques (solution and laser ablation) are very similar.

Mineral analysis

A Cameca Microbeam electron microprobe analyser (EMPA) and a JEOL 6400 scanning electron microscope (SEM) were used to measure major element concentrations in all silicate and oxide phases. Major elements were routinely analysed with energy-dispersive spectrometry on the SEM, operated at 15 kV acceleration voltage, with a beam current of 1 nA. A focused beam (nominal width of 1 μm) was normally used but a defocused beam (20–30 μm spot size) was also used for pyroxene with exsolution lamellae. Minor elements were additionally analysed with wavelength-dispersive spectrometry on the EMPA using an acceleration voltage of 15 kV and a beam current of 20 nA. Mineral analyses were assisted by backscattered electron-imaging for textural control.

PETROGRAPHY

The 18 pyroxenite samples from San Jorge and Santa Isabel have variable mineralogies and have been classified into three groups with respect to the proportions of orthopyroxene:clinopyroxene as follows: Group 1 (G1), orthopyroxenites (orthopyroxene > 80%); Group 2 (G2), websterites (60% < clinopyroxene < 80%) and (orthopyroxene > 10%); Group 3 (G3), clinopyroxenites (clinopyroxene >80%). Tables 1 and 2 summarize the modes and petrographic features of these pyroxenites.

Primary assemblages

The major phase assemblages are dominated by clinopyroxene and orthopyroxene with minor amounts of olivine and spinel. The fabric of the pyroxenites varies from large grains of approximately equal size in a fine-grained matrix (inequigranular) to complete gradation of fine- to coarse-grained fabric (seriate) (Moore, 1970). The grain boundaries range between straight and lobate. Lobate grain boundaries are consistent with dynamic recrystallization, indicative of relatively high-temperature conditions (Passchier & Trouw, 1996). Exsolution textures in these rocks demonstrate that after initial crystallization, the pyroxenes re-equilibrated to lower temperatures. Intragranular clusters of fluid inclusions are found trapped in clinopyroxene crystals remote from grain boundaries. They are interpreted to represent primary fluid inclusions and are consistent with the presence of fluid during clinopyroxene growth (Van Den Kerkhof & Hein, 2001). Although the mineralogy of the three petrographic groups of pyroxenites seems to be relatively simple, it is nevertheless possible to discern a complex poly-stage history of pyroxene growth and equilibration. For example, within websterites and clinopyroxenites, small (<0·2 mm) relics of optically continuous orthopyroxene occur surrounded by modally predominant clinopyroxene (Fig. 4a). At a macroscopic level, the invasion of a pre-existing peridotitic mantle assemblage by

pyroxene veins is clear (Fig. 3). Figure 4b illustrates some microscopic effects at the contact between a dunite and an invasive clinopyroxenite.

In the orthopyroxenites, the orthopyroxene generally occurs either as large subhedral grains with lobate grain boundaries or as small grains. The large orthopyroxene grains exhibit abundant exsolution lamellae of clinopyroxene, whereas the small orthopyroxene crystals have few exsolution lamellae. Clinopyroxene makes up a maximum 15% modal proportion of these rocks. The clinopyroxene grains are characterized by relatively small sizes (10–400 μm across) with anhedral crystal shapes, and typically have orthopyroxene exsolution lamellae. Spinel is a common minor phase in the orthopyroxenites but with a limited modal abundance (<2%). Spinel forms discrete subhedral rhombs at grain boundaries or is included in large orthopyroxene grains. Olivine occurs rarely except in sample 18.03, where it constitutes 20 modal %.

The websterites are distinct from the orthopyroxenites in having clinopyroxene:orthopyroxene >1. The size of the grains is also generally more homogeneous than in orthopyroxenites (i.e. 1–2 mm across). Exsolution lamellae of orthopyroxene occur in clinopyroxene grains, but are very limited. No exsolution of clinopyroxene has been observed in any orthopyroxene grain. The websterites are devoid of spinel. Olivine appears rarely except in sample 14.04, which contains about 15 modal %, present as interstitial anhedral grains between pyroxene crystals.

The texture of the clinopyroxenites is typically seriate with large subhedral grains (1–5 mm in size) enclosed in a mosaic of fine-grained clinopyroxene (200 μm across). Clinopyroxene grains have small (<10 μm) aligned needles of exsolved spinel, some of which is accompanied by orthopyroxene exsolution blebs. Only small amounts (<10 modal %) of orthopyroxene and/or olivine occur as small interstitial grains or as inclusions in clinopyroxene.

Retrograde assemblages

Amphibole is a common retrograde phase in all the pyroxenites, but rarely exceeds 2% of the mode. Texturally, amphibole occurs as an interstitial phase between interlocking pyroxene crystals, and can appear as blebs, mainly within clinopyroxene. Amphibole streaks or flames are generally very small (<10 μm in length), but can be highly concentrated within clinopyroxene, giving a poikilitic aspect to the pyroxene crystals. Amphibole in pyroxene is typically accompanied by secondary fluid inclusions. These fluid inclusions form trails crosscutting grain boundaries, consistent with entrapment after the growth of the host clinopyroxene, and possibly contemporaneous with growth of amphibole. Interstitial

Table 1: Petrological and mineralogical characteristics of the San Jorge and Santa Isabel pyroxenites

Description						Primary assemblage						Secondary assemblage	
Sample number	Island name	Location name	Sampling type	Rock type	Fabric/texture type	Opx (%)	Cpx (%)	Ol (%)	Sp (%)	Amphibole Hb	Amphibole Trem	Serp (%)	Others Type
14.01	SJ	Johnson Pt	outcrop	Ol–Sp Orthopyroxenite	Seriate–Interlobate	89	3	6	1	Yes	Yes	0·5	—
14.18	SJ	Padeai Pt	pebble	Ol Orthopyroxenite	Inequigranular–Interlobate	86	6	6	—	Yes	No	2	—
16.01	SJ	Salavo Bay	pebble	Sp Orthopyroxenite	Seriate–Interlobate	82	15	—	1	Yes	No	0·5	Fe_2O_3
16.05	SJ	Salavo Bay	outcrop	Sp Orthopyroxenite	Seriate–Interlobate	80	13	—	2	Yes	No	1	Fe_2O_3, Cpx II, Pect
16.11	SJ	Salavo Bay	pebble	Cpx Orthopyroxenite	Inequigranular–Interlobate	79	2	—	1	Yes	Yes	2	—
16.19	SJ	North Jorge	pebble	Ol Orthopyroxenite	Seriate–Interlobate	82	7	9	2	Yes	No	4	Wo
18.03	SI	Kolare Is	pebble	Ol Orthopyroxenite	Seriate–Interlobate	59	5	20	1	Yes	No	5	—
14.33	SJ	Salavo Bay	pebble	Cpx Websterite	Inequigranular–Interlobate/Polyg	20	78	—	—	Yes	No	—	FeS_2
16.09	SJ	Salavo Bay	pebble	Cpx Websterite	Inequigranular–Interlobate	23	74	—	—	Yes	No	1	FeS_2
16.18	SJ	North Jorge	pebble	Cpx Websterite	Inequigranular–Interlobate/Polyg	25	73	—	—	Yes	No	0·2	Fe_2O_3
16.20	SJ	North Jorge	pebble	Cpx Websterite	Inequigranular–Interlobate	20	80	—	—	Yes	No	2	—
15.03	SI	Lepi Bay	outcrop	Ol–Cpx Websterite	Inequigranular–Interlobate	36	70	2	—	Yes	Yes	4	Fe_2O_3
17.01	SI	Lepi Bay	outcrop	Websterite	Inequigranular–Interlobate	40	58	—	—	Yes	Yes	1	FeS_2, Fe_2O_3
14.16	SJ	Padeai Pt	pebble	Websterite	Seriate–Interlobate	47	53	—	—	Yes	No	0·1	—
18.08	SI	Kapika Is	outcrop	Cpx Websterite	Seriate–Interlobate	22	68	—	—	Yes	Yes	3	Pect, Fe_2O_3
15.11	SI	Kako Pt	pebble	Clinopyroxenite	Mosaic seriate–Interlobate	—	90	Trace	Exsol	Yes	No	10	FeS_2, Chl
15.13	SI	M'bani Pt	pebble	Clinopyroxenite	Mosaic seriate–Interlobate	—	96	—	Exsol	Yes	No	2	FeS_2, NiFeS, Fe_2O_3
15.16	SI	M'bani Pt	pebble	Clinopyroxenite	Seriate–Interlobate	7	80	10	Exsol	Yes	No	3	—

Cpx, clinopyroxene; Exsol, exsolution; Hb, hornblende; Ol, olivine; Opx, orthopyroxene; Pect, pectolite; Polyg, polygonal; SI, Santa Isabel; SJ, San Jorge; Serp, serpentine; Sp, spinel; Trem, tremolite; FeS_2, Fe-sulphide; NiFeS, Ni–Fe-sulphide; Fe_2O_3, magnetite; Wo, wollastonite; Chl, chlorite. Numbers in per cent refer to the modal proportion of minerals for each sample, based on the compositions of minerals and bulk-rock for each sample.

Table 2: Mineral assemblages in the San Jorge and Santa Isabel pyroxenites

	Mantle Stage	Pyroxenite crystallisation	Pyroxenite re-equilibration	Amphibolite	Late Retrogression
Cpx		Cpx I	Cpx exsolution in Opx		Cpx II
Opx	Opx relict	Opx	Opx exsolution in Cpx		
Olivine		———————			
Spinel		Cr-Al Spinel – – – – –	– – – – – –		Magnetite
Fl Inc		I	I	II	III ? – –
Amphibole				MgHb→TremHb	Trem
Sulphide				Grains	Veins
Magnetite					———————
Serpentine					———————
Pectolite					———————
Wollastonite					– – – – ———
Chlorite					– – – – ———

Opx, orthopyroxene; Cpx, clinopyroxene; Fl Inc, fluid inclusions; MgHb, magnesio-hornblende; TremHb, tremolitic hornblende; Trem, tremolite. Numbers (I, II, III) refer to different generations of fluid inclusions or clinopyroxenes.

amphibole is associated with grains of Fe-sulphide (pyrite and pyrrhotite) and Ni-sulphide; the sulphide appears to be in textural equilibrium with the amphibole. Anhedral Fe-sulphide is also present in late-stage cracks.

Magnetite only appears in trace amounts, and is often closely associated with amphibole. Tiny anhedral magnetite grains, as small as 1 μm in maximum dimension, have crystallized at the periphery of amphibole blebs and grains. Magnetite also occurs in late cracks crossing all of the minerals. Thin rims of serpentine surround many orthopyroxene and olivine grains, but the majority of the pyroxenites have <1% serpentine. In some samples of orthopyroxenite (samples 16.05, 16.19 and 18.03), serpentine forms veins crosscutting all the phases and is presumably related to late-stage alteration. Wollastonite ($CaSiO_3$) is present in the most altered orthopyroxenite (sample 16.19), and accompanies late-stage vein serpentinization. Chlorite is very limited and only occurs in the serpentine matrix in one clinopyroxenite (sample 15.11). Pectolite [$Ca_2NaH(SiO_3)_3$] is present in samples 16.05 (G1) and 18.08 (G2) in association with serpentine veins (16.05) and late amphibole (18.08), indicating a late-stage alteration paragenesis. Secondary clinopyroxene is also present in orthopyroxenite 16.05, occurring in veins developed syn- to post-serpentinization.

MINERAL COMPOSITIONS

Orthopyroxene

Representative orthopyroxene compositions are given in Table 3. Their CaO contents are relatively low (<1 wt %). The $Mg^{2+}/(Mg^{2+} + \sigma Fe$ as $Fe^{2+})$ (Mg-number) of orthopyroxene in the orthopyroxenites is comparable with accepted mantle values, ranging from 0·88 to 0·91. In contrast, the Mg-number values for orthopyroxene in websterites are lower, between 0·84 and 0·86 (Fig. 5). The Al_2O_3 contents of orthopyroxene are all low, averaging 1·37 wt % in a range from 0·45 to 2·35 wt %; the highest concentration is in orthopyroxene from sample 16.11 (Fig. 5a). Overall, a positive correlation between Cr_2O_3 content and Mg-number is observed in orthopyroxene, although within groups this is not obvious, except possibly in the websterites. The Cr_2O_3 contents reach 0·64 wt % (for Mg-number = 0·91) in the orthopyroxenites, and decreases to 0·04 wt % (Mg-number = 0·79) in the websterites (Fig. 5b).

Fig. 3. Field photographs showing the relationship between pyroxenite veins and host peridotites. (a) Harzburgite boulder crosscut by two parallel veins of websterite at Padeai Point (San Jorge). Length of hammer is 50 cm. (b) Outcrop of harzburgite exposed on the south side of the KKKFS (Sandy Point, Santa Isabel), showing different generations of websterite veins crosscutting each other. Length of hammer head is 18 cm.

Fig. 4. Photomicrographs with crossed polarizers. (a) Replacement texture of a websterite (sample 16.09; Salavo Bay, San Jorge). cpx, clinopyroxene; opx, orthopyroxene; R, relict of mantle orthopyroxene. (b) Contact between clinopyroxenite (vein) and a dunite; sample 15.13 taken at Mbadhambani Pt, Santa Isabel. Sample is also crossed by late-stage serpentine veins.

Clinopyroxene

Primary clinopyroxene

Clinopyroxene has limited compositional variation (Table 4). The Mg-number decreases in the same general order as in orthopyroxene (orthopyroxenite > clinopyroxenite > websterite), consistent with equilibrium between the respective pyroxenes (Fig. 5a). As is the case with orthopyroxene, the Al_2O_3 contents of clinopyroxene remain unchanged with decreasing Mg-number (Fig. 5a). Consistent with the coexisting orthopyroxene, the highest Al_2O_3 content (2·6 wt %) is present in clinopyroxene of orthopyroxenite 16.11; the Al_2O_3 contents of the other clinopyroxenes average 1·5 wt %. Cr_2O_3 correlates positively with Mg-number (Fig. 5b). Whereas the Cr_2O_3 contents of clinopyroxene from orthopyroxenite and clinopyroxenite are relatively high ($Cr_2O_3 > 0·3$) for a given Mg-number (~0·92), those in websterite have both lower Cr_2O_3 and Mg-number (Fig. 5b).

Secondary clinopyroxene

The secondary clinopyroxene (Cpx II) (Table 2) is chemically distinct from the primary clinopyroxene in having lower Al_2O_3 (0·23 wt %), and lower Cr_2O_3 (0·18 wt %) contents compared with primary clinopyroxene with similar Mg-number (Fig. 5b).

Olivine

Olivine grains are compositionally unzoned and homogeneous within individual samples. Representative

Table 3: Representative analyses of orthopyroxene and calculated end-member compositions

Sample:	14.01	14.18	16.01	16.05	16.11	16.19	18.03	14.33	16.09	16.18	16.20	15.03	17.01	14.14	18.08	15.16
Type:	G1	G1	G1	G1	G1	G1	G1	G2	G2	G2	G2	G2	G2	G2	G2	G3
SiO_2	57·22	57·31	56·49	56·17	55·69	56·81	57·18	55·52	56·01	55·55	55·76	54·92	54·75	56·85	55·61	57·27
TiO_2	0·02	0·04	b.d.l.	0·11	0·12	0·12	b.d.l.	0·04	0·04	b.d.l.	0·06	0·04	0·14	b.d.l.	0·05	0·04
Al_2O_3	1·09	1·24	1·74	1·59	2·34	1·16	0·96	1·69	1·57	0·95	1·22	1·37	1·48	1·08	1·46	1·56
Cr_2O_3	0·51	0·46	0·56	0·63	0·54	0·44	0·37	0·41	0·38	0·20	0·14	0·04	0·15	0·26	0·15	0·25
FeO*	6·30	7·84	7·49	6·17	7·72	6·49	6·87	9·42	9·73	10·90	10·81	14·20	12·55	8·99	11·63	7·79
MnO	0·08	0·10	0·16	0·14	0·04	0·12	0·07	0·12	0·07	0·19	0·10	0·17	0·16	0·16	0·26	n.d.
MgO	34·54	33·06	33·10	33·00	32·05	33·18	33·66	31·17	31·18	30·70	30·42	28·53	29·04	32·41	30·70	33·40
CaO	0·70	0·68	0·66	0·91	0·93	1·02	0·61	0·85	0·67	0·77	0·73	0·80	0·97	0·82	0·58	0·52
Na_2O	n.d.	0·23	0·22	0·22	0·21	0·25	0·25	0·26	0·27	n.d.	0·26	0·01	0·22	0·13	0·01	0·21
Total	100·46	100·82	100·16	98·57	99·46	99·24	99·79	99·19	99·77	99·27	99·32	100·08	99·11	100·64	100·44	101·05
% En	90	87	88	89	86	88	89	84	84	82	82	77	79	85	83	88
% Fs	9	11	11	9	12	10	10	14	15	16	16	21	19	13	16	11
% Wo	1	2	1	2	2	2	1	2	1	1	1	2	2	2	1	1
Mg-no.	0·91	0·88	0·89	0·91	0·88	0·90	0·90	0·86	0·85	0·83	0·83	0·78	0·80	0·87	0·84	0·88

*Total iron as FeO.
b.d.l., below detection limit; n.d., not determined; G1, orthopyroxenite; G2, websterite; G3, clinopyroxenite.

Fig. 5. Al_2O_3 and Cr_2O_3 vs Mg-number for the orthopyroxene and clinopyroxene in the different pyroxenites (in this and succeeding figures indicated as: G1, orthopyroxenite; G2, websterite; G3, clinopyroxenite). Secondary clinopyroxene (Cpx II) is indicated by a cross.

Table 4: Representative analyses of clinopyroxene and calculated end-member compositions

Sample:	14.01	14.18	16.01	16.05 (1)	16.05 (2)	16.11	16.19	18.03	14.33	16.09	16.18	16.20	15.03	17.01	14.14	18.08	15.11	15.16
Type:	G1	G1	G1	G1	G1	G1	G1	G1	G2	G2	G2	G2	G2	G2	G2	G2	G3	G3
SiO_2	55.15	54.73	53.81	54.02	54.59	52.46	54.06	54.07	52.71	53.00	53.98	53.08	52.73	52.85	53.58	54.59	54.03	52.50
TiO_2	b.d.l.	0.16	b.d.l.	0.16	0.22	0.13	0.07	0.14	0.11	0.08	0.10	0.17	0.08	0.18	b.d.l.	b.d.l.	0.08	0.15
Al_2O_3	1.15	1.35	1.60	1.23	0.23	2.61	1.52	1.25	1.88	1.69	1.14	1.40	1.56	1.70	1.07	1.84	0.62	2.12
Cr_2O_3	0.74	0.74	0.66	0.83	0.18	0.80	0.77	0.74	0.59	0.58	0.29	0.22	b.d.l.	0.13	0.37	0.16	0.34	0.49
FeO*	2.17	2.72	2.52	2.78	4.76	2.89	2.15	2.47	3.45	3.27	4.01	3.81	4.95	4.90	3.14	4.21	2.41	2.88
MnO	0.08	b.d.l.	0.08	0.09	b.d.l.	0.08	0.10	b.d.l.	0.09	b.d.l.	0.07	b.d.l.	0.09	0.08	b.d.l.	0.05	b.d.l.	0.11
MgO	17.94	17.36	17.42	16.94	16.66	16.60	17.43	17.25	16.57	16.32	17.11	16.29	16.19	16.22	17.15	16.87	17.14	16.94
CaO	24.78	24.45	23.81	24.11	24.54	23.53	23.80	24.14	23.06	23.57	23.22	23.56	23.74	23.00	23.47	24.17	24.78	23.49
Na_2O	0.12	0.33	0.27	0.34	0.17	0.21	0.35	0.29	0.34	0.37	0.26	0.31	b.d.l.	0.38	b.d.l.	0.07	b.d.l.	0.26
Total	102.17	101.85	99.94	100.26	101.34	99.10	100.40	100.13	98.32	98.52	99.87	98.46	99.28	98.84	98.81	102.01	99.15	98.61
% En	49	48	48	47	45	47	49	48	47	46	47	46	45	46	48	46	47	48
% Fs	3	4	4	4	7	5	3	4	6	5	6	6	8	8	5	6	4	5
% Wo	48	48	48	48	48	48	48	48	47	48	46	48	47	47	47	47	49	48
Mg-no.	0.94	0.92	0.92	0.92	0.86	0.91	0.94	0.93	0.90	0.90	0.88	0.88	0.85	0.86	0.91	0.88	0.93	0.91

*Total iron as FeO.
b.d.l., below detection limit; (1), primary clinopyroxene; (2), secondary clinopyroxene (Cpx II); G1, G2 and G3 as in Table 3.

Table 5: Representative analyses of olivine

Sample:	14.01	14.18	14.32	16.19	18.03	14.14	15.03	15.11	15.13	15.16
Type:	G1	G1	G1	G1	G1	G2	G2	G3	G3	G3
SiO_2	40.85	40.72	41.42	40.46	41.26	40.39	39.23	41.34	41.25	40.92
FeO*	10.21	12.90	10.12	9.88	10.16	13.42	21.11	11.79	9.81	11.62
MgO	49.05	46.57	49.12	48.58	48.11	46.71	39.41	47.80	48.79	47.78
NiO	0.44	0.41	n.a.	0.41	n.a.	0.42	0.13	n.a.	0.24	0.24
Total	100.7	101.0	100.7	99.6	99.9	101.1	100.2	101.5	100.4	100.8
Fo	90	87	90	90	89	86	77	88	90	88

*Total iron as FeO.
Single representative point analyses of olivine are reported; n.a., not analysed; G1, G2 and G3 as in Table 3.

olivine compositions are given in Table 5. Within all pyroxenites, forsterite (Fo) contents range from 76·5 to 90·3 mol %, with a high average of 87·4%. A plot of NiO vs Mg-number shows that the olivine in orthopyroxenite has a high Fo (~89) and NiO content (0·42 wt %) (Fig. 6) comparable with olivine from upper mantle peridotites (Girardeau & Ibarguchi, 1991). In contrast, olivine in websterite has much lower Mg-number and NiO concentrations. Although the Fo and NiO contents of olivine in websterite average 86 and 0·39 wt % respectively, olivine from sample 15.03 has markedly low Fo (~76·8) as well as a very low NiO content (0·15 wt %). Olivine in clinopyroxenite is generally similar in terms of major

elements (Fo₈₈) to olivine in orthopyroxenite, but has lower NiO (0·22 wt %). All olivine seems to be in equilibrium with clinopyroxene and orthopyroxene, as the Fo content correlates positively with the Mg-number of clinopyroxene and orthopyroxene (Fig. 7).

Spinel

Representative chromian spinel compositions are presented in Table 6. Only spinel from ortho- and clinopyroxenites is plotted as no spinel was found in the websterites. Spinel is characterized by high molar Cr/ (Cr + Al) (Cr-number) (Fig. 8). In the clinopyroxenites,

Fig. 6. NiO vs mol % forsterite in olivine. Field of the Cabo Ortegal peridotitic olivine (diagonal shading) is from Girardeau & Ibarguchi (1991).

Fig. 7. Mg-number in orthopyroxene, olivine and amphibole vs Mg-number in clinopyroxene in the same rock.

individual spinel grains and exsolution blebs have similar compositions. On a plot of Cr-number vs Mg-number (Fig. 8), the spinel plots at higher Cr-number and lower Mg-number than the abyssal peridotite array (Dick & Bullen, 1984). Spinels in the ortho- and clinopyroxenites overlap in terms of Cr-number but also extend to lower Mg-number than spinel from the SSZ peridotite suite of the Izu–Bonin forearc peridotites (Parkinson & Pearce, 1998).

Amphibole

The composition of amphibole ranges from calcic hornblende to tremolite (Table 7, Fig. 9). Hornblende has low TiO_2 (<0·2 wt %) and K_2O (<0·2 wt %), and intermediate Cr_2O_3 (0·84–0·99 wt %) and Na_2O (0·87–1·5 wt %) (Table 7). The Al_2O_3 content is relatively high, ranging from 5·66 to 11·44 wt % for the magnesio-hornblende and between 1·49 and 6·06 wt % for the tremolitic-hornblende. Some magnesio-hornblende from

Table 6: Representative spinel analyses

Sample:	14.01	16.01	16.05	16.05	16.05	16.11	16.19	18.03	15.13	15.16
Type:	G1	G1	G1	G1	G1	G1	G1	G1	G3	G3
Analysis:	6	I-4	7	11	F2	C1	9	13	G-7	23
TiO_2	0·18	0·18	0·25	0·32	0·21	0·20	0·16	0·17	0·27	0·28
Al_2O_3	14·56	20·83	19·41	19·28	18·16	27·09	16·60	16·37	25·47	34·10
Cr_2O_3	50·60	42·04	49·63	49·68	49·12	36·35	48·16	49·03	36·21	26·67
MgO	9·63	10·86	12·18	12·49	11·67	11·88	10·47	9·53	10·65	11·55
Fe_2O_3*	5·05	6·55	2·36	2·31	3·01	5·10	4·72	3·56	6·72	5·90
FeO	18·83	18·46	16·94	16·42	16·94	17·75	18·21	19·63	19·59	19·20
Total	99·62	98·26	98·38	98·70	98·81	97·86	97·84	97·93	98·24	97·11
Ti^{4+}	0·004	0·004	0·006	0·007	0·005	0·005	0·004	0·004	0·006	0·006
Al^{3+}	0·560	0·780	0·712	0·708	0·681	0·985	0·637	0·633	0·936	1·215
Cr^{3+}	1·307	1·055	1·221	1·223	1·236	0·887	1·240	1·271	0·893	0·638
Mg^{2+}	0·469	0·514	0·565	0·580	0·554	0·547	0·508	0·466	0·495	0·521
Fe^{3+}	0·124	0·156	0·055	0·054	0·072	0·119	0·116	0·088	0·158	0·134
Fe^{2+}	0·514	0·490	0·441	0·428	0·451	0·458	0·496	0·538	0·511	0·486
Mg-no.	0·477	0·512	0·562	0·576	0·551	0·544	0·506	0·464	0·492	0·517
Cr-no.	0·700	0·674	0·731	0·737	0·728	0·474	0·661	0·697	0·488	0·344
$Fe^{3+}/\sigma3+$	0·062	0·079	0·028	0·027	0·036	0·060	0·058	0·044	0·079	0·068

*Fe_2O_3 calculated on the basis of AB_2O_4 stoichiometry.
Single representative point analyses of spinel are reported. G1, G2 and G3 as in Table 3.

spinel-bearing orthopyroxenite is enriched in Cr_2O_3 (2·38–3·12 wt %). Tremolite has a low Al_2O_3 content (2·2–4·3 wt %) as well as low Na_2O (0·37 wt % on average), but relatively high Cr_2O_3 (0·93 wt % on average). In a plot of Mg-number (clinopyroxene vs amphibole), the scatter of the data suggests that only partial equilibrium between amphibole and pyroxene has been achieved, consistent with a retrograde origin for the amphibole (Fig. 7).

CONDITIONS OF FORMATION OF THE PYROXENITES
Temperature
The equilibrium coexistence of clino- and orthopyroxene permits estimation of the temperatures of equilibration of the pyroxenites, using the conventional two-pyroxene thermometer of Wells (1977). The three groups of pyroxenite are estimated to have equilibrated at temperatures between 810°C and 935°C (Fig. 10a). The average temperatures calculated for orthopyroxenite (870°C), websterite (890°C), and clinopyroxenite (900°C) (Fig. 10b) are within error, probably not distinguishable, and provide evidence for recrystallization at 850–900°C. These recrystallization temperatures are consistent with the occurrence of lobate grain boundaries, indicating that deformation has occurred at relatively high temperatures.

There is evidence of previous higher-temperature histories preserved within some of the pyroxenites, given the presence of exsolution textures. Two methods were used to integrate the exsolution lamellae into the pyroxene analyses to estimate original temperatures of crystallization. The first method consisted of analysing large host-lamellae areas by scanning the beam over a $10^3 \mu m^2$ area to restore the bulk composition of the pyroxene prior to exsolution. In the second method, an image corresponding to the area analysed by the first method was taken and analysed using a Geographic Information System technique, using ER Mapper software, to quantify the area that the exsolution lamellae represent relative to the host pyroxene. Then the relative areas were combined with spot analyses of the single phases to recalculate the original pyroxene composition. Both methods indicate that in orthopyroxenite up to 8% of exsolution lamellae are present. Consequently, the recalculated temperature of crystallization is significantly higher, and reaches an average of 990°C (Fig. 10b). A temperature of 950°C was obtained from restored pyroxene analyses from websterites; generally pyroxene in websterite has less exsolution lamellae than in orthopyroxenite, and thus the temperature of original pyroxene formation is closer to the temperature of recrystallization. The clinopyroxenites yield a slightly higher average temperature of 1050°C. A higher temperature of formation

Fig. 8. Cr/(Cr + Al) (Cr-number) vs Mg-number for spinel in the ortho- and clinopyroxenites. The field for abyssal peridotites is from Dick & Bullen (1984). Field for Izu–Bonin (SSZ) peridotites is from Parkinson & Pearce (1998).

for clinopyroxenite is in agreement with the presence of abundant spinel exsolution, a feature not present in orthopyroxenite and websterite.

Pressure

The absence of plagioclase and garnet, together with the presence of spinel, in both pyroxenites and surrounding peridotites constrains the depth of origin of the pyroxenites to be within the stability field of spinel peridotite. Rough estimates of pressure for the pyroxenites can be assessed by the absence of garnet. As garnet is stable to lower pressure in pyroxenite than in peridotite (Ho *et al.*, 2000), these pyroxenites must have equilibrated at pressures <1·7 GPa. Experiments on natural pyroxenites (Irving, 1974; Hirschmann & Stolper, 1996) estimate that 1·3–1·7 GPa corresponds to the low-pressure stability limit for garnet at the solidus.

BULK-ROCK COMPOSITIONS
Major elements

Overall the pyroxenites are characterized by high Mg-number ranging from 0·83 to 0·92 (Tables 1 and 8; Fig. 11a–d). The orthopyroxenites and clinopyroxenites tend to have higher Mg-numbers than the websterites. As might be anticipated, in a plot of CaO vs Mg-number (Fig. 11a), the orthopyroxenites are clearly distinctive compared with the websterites and clinopyroxenites. In addition, the websterites differ from the clinopyroxenites by having a wider range in Mg-number over a similar range of CaO contents.

The pyroxenites are characterized by low Al_2O_3 contents (0·6–2·7 wt %) regardless of Mg-number, and no correlation exists within any group between the Mg-number and Al_2O_3 (Fig. 11b). Orthopyroxenite sample 16.11 has the highest Al_2O_3 (2·7 wt %), whereas other pyroxenites average 1·5 wt % Al_2O_3. In contrast, a positive correlation exists between Mg-number and Cr_2O_3 contents, particularly for the websterites (Fig. 5c). Orthopyroxenites with high modal spinel proportions (samples 16.05, 16.19 and 18.03) (Table 1) are clearly offset from the other samples in terms of higher Cr_2O_3 contents at equivalent Mg-number. The TiO_2 contents are negatively correlated with Mg-number for orthopyroxenite and websterite, ranging from 0·02 wt % (Mg-number = 0·91) to 0·08 (Mg-number = 0·83) (Fig. 11d). In this case, the TiO_2 contents of the clinopyroxenite samples (~0·18 wt %) are much higher than for the other groups.

Trace elements
Rare earth elements

The various pyroxenite groups have distinctive chondrite-normalized (indicated by subscript N) rare earth element (REE) abundance patterns (Fig. 12a) with consistent light (L)REE depletion accompanied by overall low relative REE abundances (Table 9). The orthopyroxenites have very low heavy (H)REE contents (Yb_N = 0·2–0·8) and a steady decrease from the HREE to middle (M)REE (Gd_N = 0·04–0·16). The LREE (La–Eu) contents typically proved difficult to analyse because their concentrations are below the ICP-MS detection limits.

The HREE abundances in the websterites and clinopyroxenites are also strikingly low (Yb_N between 0·4 and 2) (Fig. 12a) despite the high modal proportions of clinopyroxene. These clinopyroxene-rich rocks have convex-upward REE_N patterns with depleted LREE relative to MREE, and relatively flat HREE–MREE gradients [$(Gd/Yb)_N$ ~1]. Two websterites (samples 14.16 and 18.08) are distinctive with unusual bell-shaped REE_N patterns, characterized by enrichment in the MREE

Table 7: Representative analyses of amphibole

Sample:	14.18	16.05	16.11	18.03	16.09	16.18	15.03	17.01	18.08	15.13	15.16
Type:	G1	G1	G1	G1	G2	G2	G2	G2	G2	G3	G3
Analysis:	1	2	1	2	1	1	1	3	2	1	1
SiO_2	51·26	52·61	49·33	53·06	51·31	51·62	49·67	55·64	53·68	50·58	47·53
TiO_2	0·20	0·29	0·39	0·20	0·48	0·13	0·63	0·13	0·13	0·57	0·90
Al_2O_3	7·27	5·71	9·80	5·42	7·38	6·29	9·03	2·38	4·77	8·41	10·37
Cr_2O_3	0·68	1·80	1·63	1·43	1·42	0·69	0·15	0·17	0·11	1·28	1·08
FeO	4·39	2·35	3·89	2·95	4·62	5·92	7·12	5·37	5·27	3·32	4·43
MgO	20·12	20·96	19·12	21·53	19·44	19·63	17·86	21·19	20·57	19·93	19·11
CaO	12·67	13·30	12·69	12·30	12·57	12·99	12·65	12·75	12·60	13·06	12·77
Na_2O	1·80	1·29	1·49	0·97	1·32	1·30	1·43	0·45	0·89	1·45	2·14
Total	98·39	98·31	98·34	97·86	98·54	98·74	98·54	98·08	98·02	98·60	98·33
Si	7·08	7·26	6·81	7·28	7·07	7·14	6·91	7·63	7·38	6·96	6·61
Al^{IV}	0·92	0·74	1·19	0·72	0·93	0·86	1·09	0·37	0·62	1·04	1·39
Al^{VI}	0·26	0·19	0·40	0·16	0·27	0·16	0·39	0·02	0·15	0·33	0·31
Ti	0·02	0·03	0·04	0·02	0·05	0·01	0·07	0·01	0·01	0·06	0·09
Cr	0·07	0·20	0·18	0·16	0·15	0·08	0·02	0·02	0·01	0·14	0·12
Fe^{3+}	0·32	0·01	0·38	0·34	0·34	0·40	0·40	0·44	0·48	0·21	0·39
Fe^{2+}	0·19	0·26	0·06	0·00	0·19	0·28	0·43	0·18	0·13	0·17	0·12
Mg	4·14	4·31	3·93	4·40	3·99	4·05	3·70	4·33	4·22	4·09	3·96
Ca	1·87	1·97	1·88	1·81	1·86	1·92	1·88	1·87	1·86	1·93	1·90
Na	0·48	0·35	0·40	0·26	0·35	0·35	0·39	0·12	0·24	0·39	0·58
Na^A	0·36	0·31	0·27	0·14	0·21	0·27	0·27	0·00	0·09	0·31	0·48
Mg-no.	0·89	0·94	0·90	0·93	0·88	0·86	0·82	0·88	0·87	0·91	0·88

Single representative point analyses of amphibole grains are reported. Cations are calculated on the basis of Sum(cations) − (Na + Ca + K) − 13. Types of amphibole: 1, magnesio-hornblende; 2, tremolitic hornblende; 3, tremolite. b.d.l., below detection limit. K_2O was below detection limit (<0·05 wt %); G1, G2 and G3 as in Table 3.

Fig. 9. (Na + K) in site A vs $Al^{(IV)}$ for amphiboles after the classification of amphiboles by Leake (1978).

relative to both LREE and HREE [$(Gd/Yb)_N$ = 1·4–1·6]. The olivine-rich websterite sample 14.16 is also distinct from other websterites by having extremely low HREE abundances.

The REE_N patterns of the clinopyroxenites are also bell-shaped, similar to those of websterites 14.16 and 18.08. However, the REE concentrations of the clinopyroxenites are higher than those of the orthopyroxenites and websterites with Ce_N = 0·7–1·1, Gd_N = 2·3–3·3 and Yb_N in the range 1·6–2·3 (Fig. 12a).

Other trace elements

Primitive mantle-normalized, multi-element patterns for the three pyroxenite groups are characterized overall by large ion lithophile element (LILE) enrichments relative to REE (except for clinopyroxenites), and in particular, positive Pb and Sr anomalies relative to neighbouring REE (Ce and Nd, respectively) (Fig. 12b). There is no correlation between the LOI and LILE (e.g. Cs, Rb, Ba, Pb, and variably Sr) abundances, indicating that the increases in LILE concentrations relative to the REE are not due to low-temperature alteration. Furthermore, the positive Sr anomalies relative to neighbouring REE elements are not correlated with positive Eu anomalies.

Fig. 10. Calculated temperatures of pyroxene formation and equilibration using the two-pyroxene thermometer of Wells (1977). (a) Calculation of equilibration and crystallization temperatures for individual samples. (b) Average calculated temperatures of equilibration and crystallization. Shaded bars represent the temperatures calculated directly from coexisting pyroxenes whereas the double-headed arrows provide an estimate of the original crystallization temperatures. The latter have been determined using compositions recalculated from grains displaying exsolution lamellae (see text for details).

Table 8: Whole-rock major element compositions

Sample:	14.01	14.18	16.01	16.05	16.11	16.19	18.03	14.33	16.09	16.18	16.20	15.03	17.01	14.16	18.08	15.11	15.13a	15.16
Type:	G1	G1	G1	G1	G1	G1	G1	G2	G2	G2	G2	G2	G2	G2	G2	G3	G3	G3
SiO_2	55·37	56·61	54·05	50·57	54·57	53·14	49·06	53·07	53·52	53·35	53·31	53·01	53·05	54·89	52·56	47·89	46·49	51·69
Al_2O_3	1·16	0·67	1·90	2·07	2·69	1·59	1·26	2·01	1·85	1·23	1·65	1·63	1·77	0·74	2·24	1·77	2·15	2·62
Fe_2O_3	6·88	6·69	7·60	6·74	8·17	7·39	8·39	4·96	5·22	5·99	5·43	8·51	8·69	4·89	6·39	5·04	4·64	3·99
MnO	0·15	0·15	0·16	0·14	0·17	0·15	0·15	0·13	0·13	0·16	0·15	0·20	0·19	0·13	0·14	0·10	0·10	0·09
MgO	33·80	32·85	32·47	31·34	31·29	32·72	35·05	19·85	20·30	20·95	19·40	20·90	21·44	25·03	20·02	23·65	24·55	18·87
CaO	1·53	2·63	1·67	3·85	2·29	2·24	1·55	18·79	17·91	17·26	19·27	15·31	14·25	13·17	17·44	15·43	15·12	21·07
Na_2O	0·02	0·03	0·02	0·10	0·06	0·02	0·02	0·22	0·20	0·15	0·21	0·13	0·13	0·17	0·19	0·39	0·11	0·20
TiO_2	0·02	0·02	0·04	0·05	0·07	0·04	0·03	0·07	0·07	0·06	0·07	0·08	0·07	0·03	0·10	0·13	0·15	0·18
Cr_2O_3	0·75	0·41	0·74	1·44	0·65	1·24	1·20	0·58	0·56	0·32	0·46	0·06	0·15	0·65	0·26	0·56	0·86	0·77
NiO	0·10	0·10	0·09	0·09	0·09	0·10	0·16	0·07	0·08	0·04	0·05	0·02	0·05	0·08	0·04	0·07	0·08	0·04
LOI	0·78	0·20	1·52	4·10	0·67	1·59	3·60	0·43	0·29	0·70	0·46	0·33	0·50	0·55	1·08	5·38	5·89	0·82
Total	100·6	100·4	100·3	100·5	100·7	100·2	100·5	100·2	100·2	100·2	100·4	100·2	100·3	100·3	100·5	100·4	100·2	100·4
FeO*	6·19	6·02	6·83	6·07	7·35	6·65	7·54	4·46	4·70	5·39	4·88	7·66	7·82	4·40	5·75	4·53	4·17	3·59
Mg-no.	0·91	0·91	0·89	0·90	0·88	0·90	0·89	0·89	0·89	0·87	0·88	0·83	0·83	0·91	0·86	0·90	0·91	0·90

All major elements are in wt %; G1, G2 and G3 as in Table 3.
*FeO determined wet chemically.

Fig. 11. Variation of wt % CaO (a), Al$_2$O$_3$ (b), Cr$_2$O$_3$ (c) and TiO$_2$ (d) vs Mg-number for the whole-rock pyroxenites. In (b), the values of Al$_2$O$_3$ and Mg-number (2·7 wt % and 0·88, respectively) for orthopyroxenite 16.11 are off-scale.

Despite some overall similarities, the three pyroxenite groups also differ from each other in terms of specific LILE and high field strength element (HFSE) abundance systematics. For example, the orthopyroxenites are distinctive from the other pyroxenite groups in terms of: (1) marked enrichments of Cs and Ba relative to Rb; (2) strikingly positive Pb and Sr spikes relative to Ce and Nd; (3) positive Ti spikes relative to Eu and Gd; (4) absence of negative Zr and Hf anomalies relative to Nd and Sm. It is also noteworthy that the absolute abundances of Cs, Ba, Pb and Sr are highest in the orthopyroxenites (Table 9, Fig. 12b).

The websterites and clinopyroxenites have similar primitive mantle-normalized multi-element patterns, with particularly variable negative anomalies in Nb relative to La [(Nb/La)$_N$ <1] (Fig. 12b). In more detail, the clinopyroxenites are more enriched in REE, Ti, Zr and Hf than the websterites (Fig. 12b). Whereas positive Pb and negative Ti anomalies are present in the clinopyroxenites, only small positive Sr anomalies exist.

DISCUSSION

Arc- or plume-related pyroxenites?

The occurrence of peridotites associated with pyroxenites on Santa Isabel and San Jorge, at the boundary zone between arc- and OJP-related rock suites, gives no clear evidence as to the provenance(s) of these ultramafic rocks. Based on geochemical and geophysical arguments, it has been speculated that 'hidden' cumulates in the crust of the OJP form the major part of its huge thickness (Neal *et al.*, 1997). Exposures of these rock types have not yet been identified. As the OJP supposedly represents the products of a mantle plume head (Coffin & Eldholm, 1994), its lower cumulative part should be generated from hot primitive magma.

Estimates of temperatures of crystallization indicate that the Santa Isabel and San Jorge pyroxenites were formed at temperatures of about 1000°C (and not exceeding 1050°C) before being partially re-equilibrated at 880–900°C on average. This range of temperatures indicates that the pyroxenites originated in a relatively cold environment that is incompatible with the high temperatures expected for OJP-related pyroxenite cumulates. Whereas the OJP magmas were supposed to be generated in a relatively H$_2$O-poor environment, the presence of primary fluid inclusions in the Santa Isabel and San Jorge pyroxenites may indicate involvement of a H$_2$O-rich fluid at some stage in their development. Together, these features are more consistent with an arc-related origin for the pyroxenites in Santa Isabel and San Jorge.

A subduction-related origin for the pyroxenites is further supported by mineral and bulk-rock compositions. The high Cr-number in spinel has been suggested by many workers (Bonatti & Michael, 1989; Kausar, 1998; Wang *et al.*, 2001) to be characteristic of spinels from a subduction setting. Spinel from the ortho- and clinopyroxenites is similar in terms of Mg-number and Cr-number to spinel formed in SSZ mantle rocks (Fig. 8). Bulk-rock compositions, moreover, are relatively enriched in LILE, Sr and Pb (Fig. 12b). These elements are well known to be mobile in subduction zone environments (Tatsumi *et al.*, 1986; McCulloch & Gamble, 1991; Brenan *et al.*, 1995; Keppler, 1996; Ayers *et al.*, 1997). The San Jorge and Santa Isabel pyroxenites are similar in terms of their trace element characteristics to the

Fig. 12. (a) Chondrite-normalized whole-rock REE patterns for representative pyroxenites from San Jorge and Santa Isabel. Chondrite values are from Sun & McDonough (1989). (b) Primitive mantle-normalized, whole-rock multi-element patterns for representative pyroxenites from San Jorge and Santa Isabel. Primitive mantle values are from Sun & McDonough (1989).

SSZ-related Izu–Bonin peridotites (Fig. 12b) (Parkinson & Pearce, 1998).

All these lines of evidence are consistent with the hypothesis that formation of the San Jorge and Santa Isabel pyroxenites was related to subduction zone processes and not to the genesis of the OJP. The field observation that the pyroxenites occur mainly as veins and dykes within peridotites suggests that they formed in a SSZ mantle wedge rather than as cumulates formed at the base of an island arc crust. This hypothesis is supported by the finding of rare, relict orthopyroxene of mantle affinity within the pyroxenites.

Table 9: Whole-rock trace element compositions

Sample:	14.01*	14.18†	16.01*	16.05*	16.11*	16.19*	18.03†	14.33*	16.09†	16.18†	16.20*	15.03*	17.01*	14.16*	18.08*	15.11†	15.13*	15.16†
Type:	G1	G1	G1	G1	G1	G1	G1	G2	G2	G2	G2	G2	G2	G2	G2	G3	G3	G3
Rb	0·13	0·76	0·14	0·78	b.d.l.	0·05	0·44	0·06	0·13	0·12	0·07	0·04	0·01	0·07	0·06	0·08	0·04	0·07
Sr	2·31	2·21	2·11	16·12	1·85	6·50	24·45	14·61	9·93	16·50	11·89	10·57	6·46	12·15	15·01	12·99	9·34	12·59
Y	0·18	0·35	0·31	0·51	0·74	0·38	0·31	2·72	3·22	2·28	1·93	2·46	1·91	0·80	1·65	3·00	2·85	4·47
Zr	0·03	0·31	0·21	0·33	0·32	0·20	0·26	0·81	1·14	0·98	1·03	1·55	0·80	0·55	2·05	3·23	2·28	2·94
Nb	0·01	0·01	0·01	0·01	b.d.l.	0·01	0·01	0·01	0·03	0·02	0·01	0·02	0·01	0·01	b.d.l.	0·04	0·01	0·04
Cs	1·15	0·20	1·22	5·54	0·08	0·24	0·01	0·16	0·88	1·53	0·16	b.d.l.	b.d.l.	0·11	b.d.l.	0·01	b.d.l.	0·01
Ba	2·67	1·63	3·22	33·46	5·36	308	0·51	2·50	5·54	20·33	9·53	0·57	0·11	2·24	0·14	1·78	0·85	0·22
La	b.d.l.	0·03	0·15	b.d.l.	0·01	0·02	0·02	0·06	0·09	0·10	0·05	0·06	0·02	0·03	0·09	0·33	0·12	0·18
Ce	b.d.l.	0·02	0·42	b.d.l.	0·01	0·02	0·01	0·21	0·22	0·21	0·12	0·25	0·10	0·15	0·36	0·68	0·44	0·52
Pr	b.d.l.	b.d.l.	0·05	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0·06	0·05	0·05	0·03	0·06	0·02	0·03	0·08	0·11	0·11	0·12
Nd	b.d.l.	0·02	0·19	0·02	0·03	0·02	0·01	0·38	0·38	0·31	0·19	0·42	0·17	0·22	0·52	0·65	0·74	0·84
Sm	b.d.l.	0·02	0·03	0·02	0·02	0·02	0·01	0·21	0·23	0·17	0·12	0·22	0·11	0·10	0·23	0·31	0·34	0·45
Eu	b.d.l.	0·01	0·01	0·01	0·01	0·01	0·01	0·09	0·09	0·07	0·05	0·08	0·05	0·04	0·10	0·11	0·14	0·17
Gd	0·01	0·02	0·03	0·05	0·06	0·03	0·02	0·38	0·41	0·30	0·26	0·38	0·25	0·16	0·35	0·46	0·53	0·66
Tb	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0·07	b.d.l.	b.d.l.	0·05	0·07	0·05	0·03	0·06	b.d.l.	0·10	b.d.l.
Dy	0·03	0·05	0·05	0·08	0·12	0·07	0·04	0·54	0·35	0·38	0·39	0·50	0·38	0·17	0·38	0·54	0·69	0·81
Ho	0·01	0·01	0·01	0·02	0·03	0·02	0·01	0·12	b.d.l.	b.d.l.	0·09	0·11	0·08	0·03	0·08	b.d.l.	0·14	b.d.l.
Er	0·03	0·04	0·05	0·08	0·11	0·06	0·03	0·34	0·31	0·26	0·26	0·33	0·25	0·10	0·21	0·31	0·36	0·47
Tm	0·01	b.d.l.	0·01	0·01	0·02	0·01	b.d.l.	0·05	b.d.l.	b.d.l.	0·04	0·05	0·04	0·01	0·03	b.d.l.	0·05	b.d.l.
Yb	0·06	0·05	0·08	0·09	0·13	0·09	0·07	0·31	0·33	0·25	0·24	0·31	0·25	0·09	0·18	0·27	0·31	0·38
Lu	0·01	0·01	0·02	0·02	0·02	0·02	0·01	0·05	0·05	0·04	0·04	0·05	0·04	0·01	0·03	0·04	0·05	0·06
Hf	b.d.l.	0·01	0·01	0·02	0·02	0·01	0·01	0·05	0·06	0·05	0·06	0·08	0·05	0·03	0·12	0·11	0·12	0·14
Pb	0·12	1·06	0·16	0·06	0·06	0·27	1·07	0·23	0·63	0·42	0·11	0·24	0·16	0·06	0·13	0·57	0·23	0·44
Ti	147	134	220	281	369	212	180	440	452	326	390	470	395	162	589	754	941	1079

Abundances in ppm; b.d.l., below detection limit; G1, G2 and G3 as in Table 3.
*Solution ICP-MS.
†Laser ablation ICP-MS (LA-ICP-MS).

Mantle vs crustal arc pyroxenites

We turn now to a comparison of 'mantle' vs 'crustal' pyroxenite characteristics, as outlined in the opening paragraph, with those of the SSZ pyroxenites documented here. It is appropriate to note that distinguishing between crust and mantle on the basis of seismic compressional wave velocities becomes somewhat arbitrary given velocities now acquired in the depth range 20–40 km beneath the Izu–Bonin and Aleutian arcs of ~7·6–8·0 km/s (Suyehiro et al., 1996; Fliedner & Klemperer, 1999). It is possible that these velocities correspond to clinopyroxenite and wehrlite lithologies formed through accumulation from primitive arc magmas, adjacent to the crust–mantle boundary.

Pyroxenites have been described from a number of arc environments such as the Alaskan complexes (Irving, 1974; DeBari et al., 1987; DeBari & Coleman, 1989), North America–Canada ultramafic rocks (Findlay, 1969; James, 1971; Quick, 1981; Snoke et al., 1981), intra-oceanic arc-related cumulates in southern New Zealand (Spandler et al., 2003), the deep crustal cumulates (in part, garnet-bearing) of the Jijal complex (Pakistan) (Jan & Howie, 1981; Kausar, 1998), the pyroxenite-rich peridotites from Cabo Ortegal (Spain) (Girardeau & Ibarguchi, 1991; Santos et al., 2002) and the Beni Bousera massif (Morocco) (Pearson et al., 1993). These are interpreted as medium- to high-pressure, ultramafic cumulates formed in mid- to lower crustal magma chambers, some spanning the crust–mantle boundary at the base of an arc (DeBari & Coleman, 1989).

Given the predominance of olivine and clinopyroxene as early-formed minerals in arc magmas, it follows that arc-related crustal pyroxenites are likely to be dominated by these phases, leading to high CaO at relatively low SiO$_2$ contents in typical cumulate assemblages (Fig. 13a). Mantle-derived pyroxenites differ from arc crustal pyroxenites in having a wider range of CaO (18·5 to <5 wt %) (Hirschmann & Stolper, 1996). Mantle pyroxenites

Fig. 13. Compositional data for whole-rock pyroxenites: (a) SiO_2 vs CaO (wt %); (b) Mg-number vs Al_2O_3 (wt %). Reference fields are for arc-related peridotites (Parkinson & Pearce, 1998), mantle pyroxenites (Kornprobst, 1969; Irving, 1974, 1980; Frey, 1978; Sinigoi *et al.*, 1980; Bodinier *et al.*, 1987, 1988; Griffin *et al.*, 1988; Piccardo *et al.*, 1988; Seyler & Mattson, 1989, 1993; Shervais, 1990; Pearson *et al.*, 1993; Rivalenti *et al.*, 1995; Vaselli *et al.*, 1995; Kumar *et al.*, 1996; Wilkinson & Stolz, 1997; Garrido & Bodinier, 1999; Zanetti *et al.*, 1999; McInnes *et al.*, 2001) and arc-related crustal pyroxenites (Findlay, 1969; James, 1971; Jan & Howie, 1981; Quick, 1981; Snoke *et al.*, 1981; DeBari & Coleman, 1989; Girardeau & Ibarguchi, 1991; Kausar, 1998; Smith *et al.*, 1999; Parlak *et al.*, 2002; Santos *et al.*, 2002; Worden, 2002; Spandler *et al.*, 2003).

generally include a large variety of rock types ranging from orthopyroxenite through websterite to clinopyroxenite. The Santa Isabel and San Jorge pyroxenites are also characterized by mineralogical diversity similar to that of mantle-derived pyroxenites. In a CaO vs SiO_2 diagram (Fig. 13a), the orthopyroxenite field overlaps that of mantle pyroxenite whereas websterites and clinopyroxenites span both the mantle and arc crustal fields. Arc crustal pyroxenite is characterized by a narrow range of Al_2O_3 contents compared with mantle pyroxenite (Fig. 13b). The Santa Isabel and San Jorge pyroxenites have low Al_2O_3 contents, coincident with the arc-related pyroxenite field. Orthopyroxenites and clinopyroxenites overlap the field of low-Al_2O_3 mantle pyroxenite, whereas websterites do not overlap that field (Fig. 13b).

A comparison of Al_2O_3 content vs Mg-number is shown in Fig. 14 between ortho- and clinopyroxene

compositions in the Solomons and peridotites from Izu–Bonin, and the respective fields for mantle pyroxenites. The ortho- and clinopyroxenites of Santa Isabel and San Jorge have strong affinities with mantle-derived pyroxenites whereas the field of websterites overlaps the arc crustal pyroxenite field, and trends towards the more evolved arc-related mafic cumulates. More generally, there is close compositional similarity between the orthopyroxenites and the SSZ Izu–Bonin peridotites.

In a plot of Cr_2O_3 content vs Mg-number of clinopyroxene (Fig. 14c), the more primitive, mantle-like character of the clinopyroxene in the orthopyroxenites is apparent, and is distinct from that of the websterite clinopyroxene, which has a closer affinity to arc crustal pyroxenites. The nominal mantle signature is also documented in the NiO contents of olivine in orthopyroxenites, whereas the olivine in the websterites and clinopyroxenites has both lower Mg-number and lower NiO content than those from mantle-derived peridotites (Fig. 6). Although comparison of Santa Isabel and San Jorge pyroxenites with literature data shows that the orthopyroxenites have a strong affinity with mantle pyroxenites, there are several features, especially the low Al_2O_3 contents, that are similar to arc-related crustal cumulates. In fact, the chemical characteristics of the Santa Isabel and San Jorge pyroxenites are best described by the intersection of the two established pyroxenite fields (Fig. 13a and b). The observation that the websterites, which clearly have a mantle origin, are compositionally similar to crustal pyroxenites indicates that the existing distinctions on chemical arguments are not robust. Some low-Al_2O_3 mantle pyroxenites described in the literature might have formed in a similar way to the rocks described here.

Metasomatic formation of SSZ mantle pyroxenites

Three main processes have been proposed for the formation of pyroxenites. They can form as cumulates in lower crustal magma chambers (DeBari & Coleman, 1989; Schiano *et al.*, 2000), from recycled elongated slices of oceanic crust (Polvé & Allègre, 1980; Allègre & Turcotte, 1986) or from melt/fluid–peridotite interaction (Kelemen *et al.*, 1992; Smith & Riter, 1997; Wilkinson & Stolz, 1997; Garrido & Bodinier, 1999; Smith *et al.*, 1999; Zanetti *et al.*, 1999; McInnes *et al.*, 2001; Wang *et al.*, 2001). The multi-generational network of veins crosscutting massive harzburgites (Fig. 3) and the replacement textures, wherein mantle orthopyroxene relics occur amongst an orthopyroxene–clinopyroxene assemblage (Fig. 4b), exclude an origin as arc cumulates. These features are consistent with origins via reactions between peridotite and a metasomatic agent (aqueous fluid and/or melt). On the basis of the very low REE contents

Fig. 14. Al_2O_3 vs Mg-number for (a) orthopyroxene and (b) clinopyroxene; (c) Cr_2O_3 vs Mg-number in clinopyroxene. Reference fields are for arc-related peridotites (Parkinson & Pearce, 1998), mantle pyroxenites (Kornprobst, 1969; Irving, 1974, 1980; Frey & Prinz, 1978; Sinigoi *et al.*, 1980; Bodinier *et al.*, 1987, 1988; Griffin *et al.*, 1988; Piccardo *et al.*, 1988; Seyler & Mattson, 1989, 1993; Shervais, 1990; Pearson *et al.*, 1993; Rivalenti *et al.*, 1995; Vaselli *et al.*, 1995; Kumar *et al.*, 1996; Wilkinson & Stolz, 1997; Garrido & Bodinier, 1999; Zanetti *et al.*, 1999; McInnes *et al.*, 2001) and arc-related pyroxenites and mafic cumulates (Findlay, 1969; James, 1971; Jan & Howie, 1981; Quick, 1981; Snoke *et al.*, 1981; DeBari & Coleman, 1989; Girardeau & Ibarguchi, 1991; Kausar, 1998; Smith *et al.*, 1999; Parlak *et al.*, 2002; Santos *et al.*, 2002; Worden, 2002; Spandler *et al.*, 2003).

and the presence of mantle relics, the recycling of crustal material alone is an unlikely process to explain the formation of the pyroxenites.

The relatively low temperatures of equilibration for the orthopyroxenites and websterites (950–1000°C), which are below the wet peridotite solidus, indicate that the metasomatizing agent was more probably an aqueous fluid rather than a hydrous silicate melt. This is supported by the widespread occurrence of primary fluid inclusions in the pyroxenites. The orthopyroxenite can be formed by SiO_2-rich aqueous fluid metasomatism, transforming olivine to orthopyroxene (Kelemen *et al.*, 1992; Morishita *et al.*, 2003). The fluid must have contained considerable quantities of fluid-mobile elements such as LILE, Sr, and Pb, which caused the enrichment of these elements with respect to a normal harzburgite mantle. The lower Mg-number of the websterites indicates that

the metasomatic agent responsible for producing the websterites contained substantial quantities of major elements, especially higher amounts of Ca and Fe. The clinopyroxenites are distinct from the orthopyroxenites and websterites in other respects, such as their higher temperature of formation. There is a possibility that the clinopyroxenites formed by interaction between harzburgite and hydrous silicate melt.

A metasomatic origin for pyroxenites with similar major element compositions has been proposed for the Group D-Ronda pyroxenites (Betic Cordillera, Southern Spain) by Garrido & Bodinier (1999). Based on trace element compositions, those workers suggested that some pyroxenites were formed by metasomatic replacement of mafic layers via melt–rock reactions. The REE_N patterns for the San Jorge and Santa Isabel pyroxenites (Fig. 12a) are similar to those of Ronda

Fig. 15. Tectonic model for peridotite and pyroxenite formation and exhumation. (a) The Pacific oceanic lithosphere subducted beneath San Jorge–Santa Isabel (Solomon arc) in pre-Pliocene times. During that time, hydrous melts, released from the descending slab, interacted with the SSZ peridotites. The reaction between the peridotites and the metasomatizing agents forms the pyroxenites (P). (b) The Ontong–Java Plateau (OJP) collided with the Solomon arc (including San Jorge–Santa Isabel) in the Late Pliocene (Cowley *et al.*, 2004). The compressional regime initiated the exhumation of metasomatized upper mantle along the slab via shear zones. During their exhumation, the pyroxenites were still exposed to fluids from the slab. As a result, hydration of clinopyroxene to amphibole occurred in the amphibolite facies. Continued interaction of aqueous fluids with the peridotites resulted in a late retrogression stage.

pyroxenites. Pyroxenites from both locations have depleted LREE relative to HREE and low contents of HREE, consistent with the metasomatic agent being a fluid rather a melt. In the primitive mantle-normalized multi-element pattern (Fig. 12b), additional similarities between pyroxenites from San Jorge, Santa Isabel and Ronda are clear. For example, positive Ti anomalies are characteristic of Ronda D2 orthopyroxenites and those from Santa Isabel and San Jorge, whereas negative Ti anomalies occur in Ronda D3 and Solomons websterites. Pyroxenites from both Ronda and the Solomons are characterized by positive Sr and Pb anomalies relative to neighbouring REE. Whereas we can clearly infer that the Solomon Islands pyroxenite formed in a supra-subduction zone mantle wedge, it has not been possible to reach this conclusion for the Ronda pyroxenite.

Tectonic model of pyroxenite formation and exhumation

Genesis of the pyroxenites

While the Pacific Plate subducted underneath the Solomon arc, dehydration of the descending slab presumably occurred continuously down-dip (Schmidt & Poli, 1998), and large quantities of oxidized hydrous fluids were expelled into the overlying mantle at different depths (Stern, 2002) (Fig. 15a). At deep levels, the slab-derived fluids were responsible for lowering the solidus temperature of the overlying mantle wedge, causing partial melting in the hotter portions of the wedge, generating the Solomon arc lavas.

At shallower levels ($\sim$45 km depth), some slab-derived fluids from the subducted Pacific oceanic crust migrated

via channelized pathways through the mantle wedge, locally changing its bulk composition and mineralogy. The interaction between fluid and peridotite resulted in vein and dyke formation of pyroxenite characterized by enrichments in fluid-mobile elements (Fig. 15a). The chemical characteristics of these pyroxenites share similarities with both upper mantle pyroxenites and arc-related, deep crustal cumulates.

Exhumation of the pyroxenites

The exhumation of the pyroxenites is documented in minor retrograde assemblages (Table 2). As the temperatures and pressures decreased during exhumation, some hydrous metasomatism of the pyroxenites led to local formation of amphibole. The hydration of clinopyroxene to magnesio-hornblende and tremolitic hornblende occurs in the upper amphibolite facies with pyroxene breakdown ($\sim{<}825°C$) and above the hornblende solvus ($T > 600°C$) (Kimball, 1985). Sulphide grains are texturally in equilibrium with the hornblende, and probably formed from the reaction between a hydrous fluid with high SO_2/H_2S and the pyroxenites. Such S-rich fluid arguably originates from the uppermost portion of the oceanic crust (McInnes *et al.*, 2001). As the exhumation continued (Fig. 15b), the pyroxenites partially re-equilibrated with a late retrograde fluid under greenschist-facies conditions. Below 400°C, hornblende became unstable relative to tremolite (Kimball, 1985), and continued interaction of aqueous fluid with the pyroxenites (and associated peridotite) resulted in minor serpentinization of olivine, orthopyroxene and clinopyroxene (Mevel, 1988). The observed retrogression suggests that the pyroxenites were continually exposed to fluids released from the slab that underwent continuous dehydration process (Schmidt & Poli, 1998) (Fig. 15b).

The mechanism of this exhumation is tectonically driven, resulting from the increasing coupling between the OJP and Solomon arc (Petterson *et al.*, 1997, 1999). Portions of the mantle wedge (including the pyroxenites) are therefore exhumed along the slab via shear zones (Fig. 15b). Also, it is possible that buoyant serpentinites played an important role in the exhumation of dense ultramafic rocks. On San Jorge, for example, fresh blocks of peridotite and pyroxenite are enclosed in a large serpentine massif.

CONCLUSIONS

It is proposed that the petrology, mineralogy and geochemistry of the Santa Isabel and San Jorge pyroxenites indicate that they are related to subduction zone processes, and do not represent deep lithologies of the OJP. The pyroxenites are characterized by relatively low temperatures of formation (950–1000°C) consistent with a genesis resulting from interaction between an aqueous fluid and a harzburgitic upper mantle. This process produced a chemical signature in the mantle-derived pyroxenites that is similar, in some respects, to that of arc-related cumulates.

In an arc setting, deep fluids (>120 km) released from the subducted slab are responsible for the partial melting of peridotites in the mantle wedge, generating arc magmas. We suggest that the Santa Isabel and San Jorge pyroxenites represent the products of subduction-related aqueous fluid and melt interaction with the mantle wedge at shallower depths, corresponding to a deep crust–mantle level ($\sim$30–45 km). The prime pressure constraint is the absence of garnet.

A surprising feature of the Solomons pyroxenites, given the relatively low temperatures and pressures of formation and possible involvement of hydrous fluids/melts, is the general absence of primary hydrous phases, such as amphibole and mica; this is a feature shared with the Ronda (Group D) pyroxenites (Garrido & Bodinier, 1999). With respect to this feature, it is worth recalling the pyroxenites are clearly not equivalent in bulk chemistry to ordinary mantle melts (see Table 8), but rather represent replacement-reaction products of peridotite. The vein systems now occupied by pyroxenite probably preserve the channelized flowpaths but not the total flux of fluids and melts. Accordingly, the absence of primary hydrous phases may be the result of insufficient volatile content at any given period in the flux history. Nevertheless, the pyroxenites exposed on the islands of San Jorge and Santa Isabel provide an important opportunity to investigate the nature of fluids released at relatively shallow depths from the subducted slab.

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