Melt extraction and enrichment processes in the New Caledonia lherzolites: Evidence from geochemical and Sr–Nd isotope data

Arianna Secchiarì, Alessandra Montanini, Delphine Bosch, Patrizia Macera, Dominique Cluzel

1. Introduction

Ophiolites have attracted much scientific interest since their recognition as surface exposures of fossil oceanic lithosphere accreted to continental margins (Dilek and Furnes, 2011 and references therein). The accessibility of ophiolitic sequences makes them an excellent natural laboratory for the study of oceanic lithosphere. Hence, field-based geochemical investigation of ophiolitic peridotites may prove valuable in understanding melt extraction and melt/rock interaction processes in the upper mantle. The genesis of oceanic lithosphere differs with geodynamic setting, i.e., mid-ocean ridge (MOR), ocean–continent-transition (OCT) and supra-subduction zone (SSZ) have distinct structural and chemical properties that permit their identification (Dilek and Furnes, 2011; Miyashiro, 1973; Pearce et al., 1984). The coexistence of peridotites with contrasting geochemical signatures indicative of MOR-type and SSZ-type melting regimes has been documented in other ophiolitic sequences (e.g., Lycian and Antalya ophiolitic complexes from Turkey, Aldanmaz et al., 2009, and Coast Range ophiolites, Choi et al., 2008). Geochemical and mineralogical characteristics of mantle residua may be an effective tool to identify the geodynamic setting of ophiolite formation.

The New Caledonia Peridotite Nappe is one of the largest ophiolite complexes in the world. The ophiolite is dominated by mantle lithologies (mainly refractory harzburgites, together with minor lherzolites), locally capped by mafic–ultramafic cumulates. Although the New Caledonia mantle bodies have been the subject of several petrological and geochemical investigations starting from the 1980s (Dupuy et al., 1981; Nicolas and Dupuy, 1984; Prinzhofer and Allegre, 1985; Prinzhofer...
et al., 1980), the ultra-depleted nature of these rocks hampered their geochemical characterisation. In particular, isotopic data for the peridotites are still lacking, save for some Pb isotope data (Hamelin et al., 1984). Over the past decade, several petrological and geochemical investigations were undertaken, with a view to understanding the origin of the harzburgites and associated intrusive rocks (Marchesi et al., 2009; Pirard et al., 2013; Ulrich et al., 2010). These studies converged on the consensus that the harzburgites are of a SSZ affinity. In contrast, a relatively limited geochemical dataset is available for the lherzolites (Ulrich et al., 2010) meaning that their evolution and geodynamic affinity are poorly understood. Another debated issue concerns the existence of a genetic link between the New Caledonia Iherzolites and the basaltic unit of Poya that tectonically underlies the Peridotite Nappe (Eissen et al., 1998; Marchesi et al., 2009; Whattam, 2009).

In order to bridge these gaps in knowledge, a comprehensive petrographic and geochemical characterisation of the main lherzolite bodies from New Caledonia are presented, in which we investigate melt extraction and melt–peridotite interaction processes. To this end, major and trace elements were determined in minerals and bulk rocks for selected samples of spinel and plagioclase lherzolites. In addition, Sr–Nd isotope data were obtained on bulk rocks and clinopyroxene separates to fingerprint the geochemical signatures of the New Caledonia lherzolites in the context of mantle endmembers. Finally, we evaluate the relationship between the lherzolites and the oceanic basalts from the Poya terrane.

2. Geological and petrological background

New Caledonia is located in the southwest Pacific Ocean. The island represents the northern extension of the Norfolk Ridge (Fig. 1) and is composed of a complex association of volcanic, sedimentary and metamorphic terranes ranging in age from Permian to Miocene (Aitchison et al., 1995a; Cluzel et al., 1994, 2001, 2012; Lagabrielle et al., 2013). In particular, New Caledonia can be subdivided into four main geological domains (Cluzel et al., 2001): (i) the Basement units (pre-Late Cretaceous basement and Late Coniacian-to-Late Eocene sedimentary cover), (ii) the Cenozoic high-pressure metamorphic belt, (iii) the basaltic Poya Terrane and (iv) a large slab of obducted peridotites, the Peridotite Nappe.

The Peridotite Nappe is an allochthonous sheet of oceanic lithosphere from the Loyalty Basin (Fig. 1) thrust from NNE to SSW over the continental basement of the island in the Late Eocene (Collot et al., 1987; Prinzhofer et al., 1980). Its emplacement has been related to the failed subduction of the New Caledonia continental basement in a NE-dipping subduction zone, which culminated with the obduction of the Loyalty oceanic lithosphere (Cluzel et al., 1998). The Peridotite Nappe overlies the basaltic Poya Terrane, separated by a sub-horizontal thrust fault and a thick (50–200 m) sole of serpentinite mylonites, locally associated with high-temperature amphibolite lenses formed during subduction initiation in the Early Eocene (Cluzel et al., 2012).

The Peridotite Nappe is largely exposed in the Massif du Sud (Fig. 1), a thick harzburgite–dunite sequence with kilometre-scale lenses of maﬁc and ultramaﬁc cumulates at the top (Pirard et al., 2013). The cumulates crystallised from ultra-depleted melts, most likely in a nascent arc environment (Marchesi et al., 2009; Pirard et al., 2013). Large klippen of peridotites also occur along the west coast (Fig. 1), in the Isle of Pines (south) and Belep Islands (north). Smaller klippen are exposed in the central chain, and a few kilometre-scale lenses of peridotite are included in the HP–LT metamorphic complex.

Fig. 1. Simplified geological map of New Caledonia (modified after Cluzel et al., 2001 and Vitale Brovarone and Agard, 2013). Location of the northern lherzolite massifs and the main sampling areas are also shown.
The New Caledonia harzburgites are highly depleted peridotites, as evidenced by the absence of primary clinopyroxene and exceedingly low whole-rock trace element contents (Prinzhofer and Allegre, 1985). Recent geochemical studies (Marchesi et al., 2009; Ulrich et al., 2010) invoke a complex history in the formation of these rocks, in which melting, melt–rock interaction and re-melting, gave rise to their ultra-depleted compositions.

Lherzolitic rocks occur only in the northern part of New Caledonia, where they form discrete ultramafic bodies (Poum and Tiebaghi massifs, Fig. 1). These bodies are mostly composed of spinel and plagioclase lherzolites, with minor cpx-bearing harzburgites and dunite–harzburgites (Leblanc, 1995; Nicolas and Dupuy, 1984; Sécher, 1981). Centimetre-thick pyroxenite dikes may be common, especially in the plagioclase lherzolites from the Poum Massif (Leblanc, 1995; Nicolas and Dupuy, 1984).

In the lherzolites, the widespread development of superimposed structures testifying evolution from high-temperature ductile to fragile deformation, has been related to the presence of a major shear zone that runs along the northeastern side of Belep Islands and the Poum and Tiebaghi massifs. This high-strain zone, characterised by a steeply dipping foliation, has been interpreted as a palaeo-transform fault (Nicolas, 1989; Sécher, 1981) that was reactivated at shallower conditions during and after obduction (Titus et al., 2011). In the Poum Massif, plagioclase lherzolites with mylonitic to protomylonitic textures grade into relatively undeformed plagioclase and spinel lherzolites and, finally, cede to harzburgites, moving away from the shear zone (Sécher, 1981). According to Leblanc (1995), the harzburgites represent more depleted asthenospheric material channelled along the transform fault. The Poya Terrane is mainly composed of massive and pillow basalts intruded by rare dolerite dikes. Tholeiitic basalts with enriched-MORB affinity predominate, and are associated with minor occurrences of back-arc basin basalts (BABB) and ocean island basalts (OIB) (Cluzel et al., 1997, 2001; Eissen et al., 1998). These variable geochemical signatures suggest derivation from a variety of mantle sources. The Poya basalts are tectonically interlayered with thin (1–3 m) radiolarian-bearing pelagic sediments, which constrain the timing of basin eruption between Campanian and Late Paleocene-Early Eocene (Aitchison et al., 1995b; Cluzel et al., 2001).

3. Sampling and petrography

Due to widespread serpentinisation, fresh peridotite exposures are rare. After extensive exploration and careful sampling, nine samples from the Babouillat (Tiebaghi Massif) and Poum outcrops (Fig.1) were chosen for this study, together with two basaltic lavas from the Poya Terrane.

In the peridotites, olivine is partially replaced by serpentine + magnetite assemblages exhibiting mesh structures, whereas ortho-and clinopyroxenes are not affected by serpentinisation. Other secondary minerals are represented by small amounts of chlorite, talc and tremolitic amphibole as rare overgrowths on pyroxenes. Spinellilherzolites are clinopyroxene-poor (up to 7–8 vol%), medium- to fine-grained rocks, characterised by a porphyroclastic texture (Fig. 5a) with local protomylonite development.

Olivine occurs as kinked and elongated porphyroclasts, forming the main foliation. Orthopyroxene porphyroclasts are coarse-grained crystals (sometimes up to 5–6 mm across) with a rounded habit. They also occur as elongate grains parallel to the foliation, and show evidence of ductile deformation (undulose extinction and kink bands) and exsolution lamellae of clinopyroxene. Large orthopyroxenes frequently display embayments filled by undeformed olivine crystals (Fig. 5a). Small, undeformed and exsolution-free orthopyroxene grains are found in association with recrystallised olivine, spinel and clinopyroxene. Exsolved clinopyroxene porphyroclasts occur as smaller (up to 1–2 mm), light green-brownish crystals in apparent equilibrium with the primary assemblage (Fig. 5b). Yellow-brown spinels display irregular or rounded shapes. Holly-leaf shaped grains are sometimes found between olivine and pyroxene porphyroclasts.

Plagioclase lherzolites are characterised by medium- to coarse-grained porphyroclastic textures. Plagioclase occurs as interstitial blebs, irregular patches or rims around spinel (Fig. 5c) and are often sericitised. Clinopyroxene porphyroclasts show very irregular shapes, frequent embayments and partial replacement by orthopyroxene + plagioclase aggregates. Olivine may be rimmed by secondary orthopyroxene.

4. Analytical techniques

Major element mineral analyses were performed at the Physics and Earth Sciences Department in Parma using a JEOl-6400 electron microprobe equipped with a LINK-ISIS energy dispersive microanalytical system. The electron beam was produced at an accelerating voltage of 15 kV and probe current of 0.25 nA, and both natural minerals and synthetic compounds were used as standards.

The samples for bulk rock analyses were sawn into thin slabs and altered parts were carefully removed before crushing. In order to obtain representative major and trace element analyses, rock powders were prepared from at least 1 kg of material and crushed to powders in an agate mill.

Whole-rock major-element oxides and Ni, Co, Cr, V, and Zn were measured by conventional XRF techniques at the Department of Chemical and Geological Sciences in Modena using a wavelength dispersive X-ray fluorescence spectrometer (Philips PW 1480).

Whole-rock and in situ trace element analyses were performed at the AETE technical platform (Trace Element Analysis in the Environment, Université de Montpellier). Bulk trace element concentrations (excluding Ni, Co, Cr, V, and Zn) were determined using a ThermoFinnigan ELEMENT XR (HR-ICP-MS: high resolution inductively coupled plasma mass spectrometer) after acid digestion with HNO₃–HF–HClO₄, as described by Ionov et al. (1992). The replicate analyses of international standards and three investigated samples (BA1, BAB2A, and POU2B) are reported in Table S1.

Trace element compositions of clinopyroxenes were measured in-situ on polished thick sections (thickness = 120 μm) with a Geolas Lambda Physik CompEx 102 excimer. Signals were acquired in time resolved acquisition, wherein the background was measured for 2 min compared with 1 min for the sample. The laser was fired at an energy density of 15 J/cm² and a frequency of 8–10 Hz, using a spot size of 102 μm. Si was used as the internal standard and analyze concentrations were calibrated against the NIST 612 rhodochrosite glass, according to the values of Pearce et al.’s (1997) study. Data were subsequently reduced using the GLITTER software by inspecting the time-resolved analysis and selecting time intervals in which signal intensities were homogeneous. Our data for the international standard reference material BIR-1G shows good agreement with published values.

For isotope analyses, powdered samples were weighed to obtain approximately 100–200 ng of Sr and Nd. A 1 hour 6 N HCl leaching step at 95 °C was performed before acid digestion. After leaching, residues were rinsed three times in Milli-Q H₂O. The samples were subsequently dissolved during 72 h on a hot plate with a mixture of HF and HNO₃. After evaporation to dryness, 2 ml of HNO₃ was added to the residue and kept at 95 °C for 24–48 h.

Strontium isotopes were separated using Sr Eichrom resin (Pin et al., 1994), while neodymium isotopes were separated in two steps, a first using AG50W-X12 cation exchange resin followed by a second to purify REE using HDEEH conditioned Teflon columns.

Sr isotopic ratios were measured by thermal ionisation mass spectrometry using a Triton Finnigan Mat spectrometer from the Labogis from the Nimes University. ⁸⁷Sr/⁸⁶Sr isotopic ratios were normalised using a value of 0.1194 for the ⁸⁶Sr/⁸⁸Sr ratio. The NBS987 standard, analysed during the course of this study, yielded an average ⁸⁷Sr/⁸⁶Sr of 0.710243.
$^{143}$Nd/$^{144}$Nd isotopic composition was determined by multi-collector inductively coupled plasma mass spectrometry (Finningan Neptune Plus) at the École Normale Supérieure de Lyon (ENS). The average values were $0.511971 \pm 0.000012$ ($n = 10$) for the AMES-Rennes (Chauvel and Blichert-Toft, 2001) Nd standard.

5. Major element mineral chemistry

The forsterite content of olivine in the spinel and plagioclase lherzolites is rather homogenous (88.5–90.0 mol%).

The representative major element mineral compositions of clino- and orthopyroxenes are reported in Table S2. Orthopyroxene is enstatitic with an average composition $\text{WO}_{0.01}\text{En}_{0.84}\text{Fs}_{0.11}\text{Fo}_{0.14}$ and $\text{Mg}^\# \left[\text{Mg}/(\text{Mg} + \text{Fe}^{2+})\right]$ between 0.88–0.90. Orthopyroxene mirrors the behaviour of olivine, with no significant compositional difference between spinel and plagioclase lherzolites in terms of $\text{Mg}^\#$. $\text{TiO}_2$ contents are generally very low (<0.10 wt.%). The cores of orthopyroxene porphyroclasts in the spinel lherzolites are $\text{Al}_2\text{O}_3$-rich (4.9–5.9 wt.%) and have $\text{Cr}_2\text{O}_3$ contents varying between 0.63 and 0.90 wt.%. $\text{Al}_2\text{O}_3$ and $\text{Cr}_2\text{O}_3$ slightly decrease from core to rim, whereas distinctly lower values are observed in neoblastic orthopyroxenes ($\text{Al}_2\text{O}_3 = 2.4–3.2$ wt.% and $\text{Cr}_2\text{O}_3 = 0.57–0.19$ wt.%, respectively), most likely as a result of synkinematic spinel crystallisation. As a whole, polymineralic and neoblastic orthopyroxenes from plagioclase lherzolites show decreasing $\text{Al}_2\text{O}_3$ contents between porphyroclast cores and neoblasts (5.6–3.2 wt.% and 2.3–1.8 wt.%, respectively), whereas no substantial $\text{Cr}_2\text{O}_3$ variation is observed. The orthopyroxene crystals in plagioclase–orthopyroxene aggregates have low $\text{Al}_2\text{O}_3$ (1.8–2.3 wt.%) and $\text{Cr}_2\text{O}_3$ concentrations ranging between 0.44 and 0.90 wt.%, in agreement with the concomitant crystallisation of plagioclase.

Orthopyroxene is diopsidic with relatively high $\text{Na}_2\text{O}$ and $\text{Al}_2\text{O}_3$ contents and $\text{Mg}^\#$ varying between 0.90 and 0.92. Porphyroclastic clinopyroxene cores from spinel and plagioclase lherzolites have similar $\text{Al}_2\text{O}_3$ contents (5.3–6.7 and 5.0–6.4 wt.%, respectively). Spinel-facies clinopyroxene porphyroclasts have slightly higher $\text{Na}_2\text{O}$ contents (0.48–0.58 wt.%) and lower $\text{Cr}_2\text{O}_3$ (0.80–1.0 wt.%) than those from plagioclase lherzolites ($\text{Na}_2\text{O} = 0.25–0.40$ wt.% and $\text{Cr}_2\text{O}_3 = 1.2–1.4$ wt.%). A moderate decrease of $\text{Al}_2\text{O}_3$ and $\text{Na}_2\text{O}$ is observed in porphyroclast rims and in the neoblastic clinopyroxene of spinel lherzolites.

The spinel compositions (Table S3) show significant differences between spinel and plagioclase lherzolites. In spinel lherzolites, $\text{Cr}^\#$ and $\text{Mg}^\#$ vary between 0.132–0.167 and 0.732–0.759, respectively. Plagioclase-bearing lherzolites are characterised by considerably higher $\text{Cr}^\#$ (0.484–0.624) and lower $\text{Mg}^\#$ (0.375–0.551). High $\text{Cr}^\#$ of spinel from plagioclase lherzolites is associated with elevated $\text{TiO}_2$ (0.2–0.5 wt.%), whereas $\text{TiO}_2$ is generally <0.15 wt.% in spinel lherzolites (Fig. 2c).

$\text{Mg}^\#–\text{Cr}^\#$ relationships for spinel are shown in Fig. 2a. Spinel compositions of the New Caledonia harzburgites have been also reported for comparison. Spinel from spinel lherzolites display compositions similar to those of abyssal (i.e. mid-ocean ridge) peridotites. In contrast, spinels from plagioclase lherzolites are characterised by lower $\text{Mg}^\#$ and higher $\text{Cr}^\#$, similar to the refractory harzburgites.

The compositional relationships between olivine and chromian spinel are illustrated in Fig. 2b. The compositions of spinel from the spinel lherzolites fall in the range of passive margin lherzolites, very close to the field of abyssal peridotites. In contrast, $\text{Mg}^\#$ and $\text{Cr}^\#$ composition for plagioclase lherzolites plot off the mantle array.

Ca-rich plagioclase (anorthite – 88 mol%) has been analysed in one sample (POU1A) where tiny relict patches of fresh plagioclase have been preserved.

6. Geothermometry

Equilibrium temperatures for the lherzolites were calculated on the basis of mineral composition data obtained by microprobe analyses.

![Fig. 2. Compositional variations of spinel: (a) Cr#–Mg#. Field for abyssal and fore-arc peridotites are taken from Dick and Bullen (1984) and Ishii et al. (1992), respectively; (b) Average olivine $\text{Mg}^\#$ and spinel $\text{Cr}^\#$ for spinel and plagioclase lherzolites. The olivine–spinel mantle array (OSMA) is shown by dashed lines. Field for subduction zone, abyssal and passive margins peridotites are after Dick and Bullen (1984); Arai (1994) and Pearce et al. (2000). Data for the New Caledonia harzburgites are from Secchiari (unpublished PhD thesis data); (c) $\text{Cr}^\#–\text{TiO}_2$ (wt.%). Spinel data from impregnated ophiolitic plagioclase lherzolites (Lower Platta: Müntener et al., 2010; Lanzo: Piccardo et al., 2007; External Ligurides: Rampone et al., 1995) are also shown for comparison.](image-url)
and neoblastic assemblages of spinel lherzolites, for an assumed confining pressure of 1.5 GPa. To avoid the uncertainties related to Fe\(^{3+}\) recombinations, all iron in pyroxenes was assumed to be Fe\(^{2+}\). This assumption has negligible effects on two-pyroxene thermometry, as discussed by Canil and O’Neill (1996).

The cores of spinel–facies orthopyroxene porphyroclasts yield \(T(Ca-In_{Opk})\) between 1030 and 1095 °C, whereas two-pyroxene porphyroclast temperatures vary over a wider range (940–1080 °C). The lower \(T\) values obtained using two-pyroxene geothermometry are related to pyroxene exsolution during cooling. Preservation of Ca-rich domains in orthopyroxene, on the other hand, is a consequence of lower diffusivity of Ca in orthopyroxene compared to high Ca–Mg diffusivity in clinopyroxene (Lenoir et al., 2001; Montanini et al., 2006). Core to rim compositional variations in the orthopyroxene porphyroclasts locally reflect cooling of ca. 50–100 °C. The temperatures obtained with pyroxene thermometry for neoblastic assemblage of spinel lherzolites are shifted towards lower values (890–920 °C).

We have also applied the Ca-in-orthopyroxene geothermometer to the plagioclase lherzolites for an assumed pressure of 0.8 GPa. The slightly higher \(T\) obtained for porphyroclastic (980–1120 °C) and neoblastic assemblages (960–1070 °C) are possibly related to the melt impregnation process (see Section 8.2). Pressure is difficult to constrain owing to the lack of appropriate geobarometers in garnet-absent systems. The pressure of the garnet–spinel transition in peridotitic systems is mainly controlled by the Cr content of the spinel (Klemme, 2004; O’Neill, 1981; Webb and Wood, 1986). In particular, spinel is stabilised to higher pressures with increasing Cr/(Cr + Al); A maximum pressure estimate for the spinel facies can be determined by using the formulation of Webb and Wood (1986). Using the composition of spinel porphyroclast cores not rimmed by plagioclase, the obtained maximum pressure is 2.2 GPa.

7. Geochemistry

7.1. Major and trace element bulk rock composition

The whole-rock major- and trace element contents of lherzolites are reported in Table 1. The LOI values, ranging between 6.39 and 7.28 wt.%, reflect moderate degrees of serpentinisation. Nevertheless, projection of the data onto a MgO/SiO\(_2\)-Al\(_2\)O\(_3\)/SiO\(_2\) diagram (Fig. 5) shows that the analysed peridotites plot close to the terrestrial array defined by fresh mantle rocks. This indicates that their major element composition was not affected by a significant chemical exchange related to seawater alteration.

Plagioclase and spinel lherzolites show similar major element compositions, except for their Al\(_2\)O\(_3\) and CaO contents, which are shifted to higher values in plagioclase lherzolites (Table 1, Fig. 3). Al\(_2\)O\(_3\) and CaO vary between 1.30–2.83 wt.% and 1.26–2.20 wt.% respectively. Both lherzolite types display variable degrees of depletion, with MgO ranging from 36.7 to 40.8 wt.%. The negative correlations of Al\(_2\)O\(_3\), SiO\(_2\) and CaO vs. MgO are observed (Fig. 3).

The chondrite-normalised REE patterns are plotted in Fig. 4a. The analysed samples are characterised by higher MREE–HREE abundances than the LREE, where Nd\(_0\)/Sm\(_0\) = 0.19–0.27 and Nd\(_0\)/Yb\(_0\) = 0.03–0.07 for Yb\(_0\) = 0.99–1.30. They broadly exhibit positively-sloping patterns, where HREEs are more abundant than MREEs (Tb\(_0\)/Yb\(_0\) = 0.59–0.70). On the whole, these patterns closely resemble those of modern abyssal peridotites (e.g. Pearce et al., 2000).

Spinel and plagioclase lherzolites have REE patterns that are subparallel to one another, though the plagioclase-bearing varieties are typified by higher MREE–HREE contents. A slight enrichment in La–Ce, particularly in the spinel lherzolites, is also observed (Fig. 4a).

Extended multi-element diagrams normalised to Primitive Mantle (PM) are reported in Fig. 4b. Patterns are subparallel for spinel and plagioclase lherzolites from Ce. Cs is enriched only in plagioclase lherzolites, especially in the samples from Babouillat (BAB1B and BAB2B). Rb and Ba are commonly enriched with respect to adjacent elements. Positive Pb and Sr anomalies are present in all samples, as are pronounced negative Zr anomalies.

Whole rock compositions and chondrite-normalised REE patterns of two Poya basalts are reported in Table 1 and Fig. S3. The basalts display nearly flat REE patterns, with abundances between 15 and 20 times chondritic values.

7.2. Clinopyroxene trace element composition

The trace element concentrations and chondrite-normalised REE patterns for the analysed clinopyroxenes are reported in Table S4 and Fig. 5.

Clinopyroxene from spinel and plagioclase lherzolites have subparallel patterns. Slightly higher REE contents are observed in the plagioclase-bearing rocks (Fig. 5a). Clinopyroxenes have strongly depleted LREE abundances but relatively flat HREE (Tb\(_0\)/Yb\(_0\) = 0.72–1.01) and MREE/HREE fractionation (Sm\(_0\)/Yb\(_0\) = 0.27–0.50). The lherzolite patterns closely resemble those of clinopyroxenes from abyssal peridotites (Brunelli et al., 2006; Hellebrand et al., 2002; Johnson et al., 1990; Warren et al., 2009). Clinopyroxenes from spinel lherzolites are also characterised by a small La–Ce enrichment. Small negative Eu anomalies (Eu/Eu\(^*\) = 0.77–0.92), occur in the clinopyroxene from plagioclase lherzolites, as a result of equilibration with plagioclase.

The absolute concentrations of the most incompatible trace elements (from Cs to Pb) are generally higher in the clinopyroxene from the plagioclase lherzolites (Fig. 5b). Negative anomalies for Ba and Pb enrichment are observed. Clinopyroxene from spinel and plagioclase lherzolites display similar and convex upwards patterns from Sr to Lu, with higher absolute concentrations for plagioclase lherzolites. Sr invariably shows negative anomalies, as well as Zr, which is decoupled from the neighbouring Hf.

7.3. Sr–Nd isotope composition

The Sr and Nd isotopic ratios are reported in Table 2. \(^{87}\)Sr/\(^{86}\)Sr and \(^{143}\)Nd/\(^{144}\)Nd have been determined for six lherzolites, one clinopyroxene fraction from the lherzolite sample BA1 and two basaltic lavas from the Poya unit. All the samples have been corrected for radiogenic ingrowth assuming an age of 53 Ma, i.e. the inferred timing of subduction inception (Cluzel et al., 2006).

The lherzolites display initial \(^{143}\)Nd/\(^{144}\)Nd ratios between 0.512930 and 0.513134 and \(^{87}\)Sr/\(^{86}\)Sr between 0.705782 and 0.707527. The corresponding values of \(\varepsilon_{Nd(i)}\) vary from +6.98 to +10.97, while \(\varepsilon_{Sr(i)}\) is more variable, ranging from +19.19 to +43.85. A slight difference can be detected for \(^{87}\)Sr/\(^{86}\)Sr values between the Babouillat and Poum localities, with more radiogenic values obtained for Poum (\(\Delta^{87}\)Sr/\(^{86}\)Sr = 0.00123).

The clinopyroxene separate from the spinel lherzolite BA1 yields initial values of 0.705791 for \(^{87}\)Sr/\(^{86}\)Sr and 0.512982 for \(^{143}\)Nd/\(^{144}\)Nd, in good agreement with the initial value obtained for the bulk rock (0.706128 and 0.513005 for \(^{87}\)Sr/\(^{86}\)Sr and \(^{143}\)Nd/\(^{144}\)Nd, respectively). Sr isotopic ratios of bulk rock and coexisting clinopyroxenes are significantly more radiogenic than DMM values, whereas Nd isotope compositions are more akin to a depleted asthenospheric mantle source (e.g. Hofmann, 2003).

In a Sr–Nd isotope diagram (Fig. 6) the lherzolite samples define a subhorizontal trend, falling outside the fields of both abyssal peridotites and Pacific arc volcanics, as a result of highly radiogenic Sr isotope compositions.

The two basalts from the Poya Terrane have similar Sr and Nd initial isotopic ratios (\(^{87}\)Sr/\(^{86}\)Sr = 0.70407 and 0.70442, \(^{143}\)Nd/\(^{144}\)Nd = 0.51285 and 0.51284 for the samples KO1 and BOU1, respectively). They plot in the Pacific arc domain and fall close to the compositions of Poya basalts with E-MORB affinity (Cluzel et al., 2001). As a whole, the
**8. Discussion**

**8.1. Evaluation of partial melting processes: garnet involvement**

The New Caledonia lherzolites have a depleted chemical signature and show remarkable similarity to abyssal peridotites. To assess whether major element variation in the lherzolites is brought about by partial melting, trends are compared with the anhydrous melting curves calculated by Niu (1997). The fractional melting curves in the interval 2.5–0.8 GPa (fractional melting degree ≤0.25) and 1.5–0.8 GPa (fractional melting degree ≤0.15) are shown in Fig. 3. Isobaric batch melting curves for 1 and 2 GPa are also plotted for comparison. The spinel lherzolite samples fall on the polybaric fractional melting curves between 2.5 and 0.8 GPa in the Al₂O₃–MgO and CaO–MgO plots, and plot close to the curves in the SiO₂–MgO and FeO₀₉₋₀–MgO diagrams. Their major element compositions are therefore consistent with partial melting residues. On the other hand, the plagioclase lherzolites depart from the melting curves, being characterised by higher Al₂O₃ contents and lower CaO and SiO₂ concentrations. Hence, the compositions of the plagioclase lherzolites are not compatible with products of partial melting, rather they reflect additional processes that will be discussed in Section 8.2.

The observed major element trends discussed above suggest that the widespread occurrence of orthopyroxene embayments filled by olivine in spinel lherzolites may be ascribed to incongruent melting of orthopyroxene, characteristic of decomposition melting beneath oceanic ridges (Niu, 1997). These textures could be also related to reactive...
Fig. 3. Bulk rock abundances of CaO, Al₂O₃, SiO₂ and FeO₄ vs. MgO (data on anhydrous basis in wt.%), for New Caledonia spinel and plagioclase peridotites compared with calculated trends for residual peridotites using the model of Niu (1997). Source end-member is from McDonough and Sun (1995). Dashed lines: melting curves for isobaric batch melting; thick curves: near fractional melting.

Fig. 4. (a) Chondrite-normalised REE patterns for spinel and plagioclase lherzolites. Normalising values after Sun and McDonough (1989); (b) Primitive mantle-normalised trace element patterns of New Caledonia spinel and plagioclase lherzolites. Normalising values after Sun and McDonough (1989).
porous flow of pyroxene-undersaturated melts (e.g. Piccardo et al., 2007; Rampone et al., 2008). Nevertheless, the absence of chemical evidence, such as the decrease of SiO$_2$ and increase of FeO against MgO, resulting from olivine addition during melt migration, argues against this interpretation.

A rough estimate of the degree of partial melting for the spinel lherzolites can be obtained adopting the method of Hellebrand et al. (2001), which is based on the Cr$#$ value of spinel ($F = 10 \cdot \ln$ Cr$#$ + 24). Applying this equation to the spinel porphyroclast cores of spinel lherzolites (Table S3), low degrees of partial melting were inferred (4%–5% and 5%–6% for the samples BA1 and POU2, respectively). This method, however, may underestimate the degree of bulk melting of the peridotites, as Cr$#$ values can be strongly affected by subsequent refertilisation. To provide further constraints on the partial melting

![Graph](image)

**Fig. 5.** (a) Chondrite-normalised REE compositions of clinopyroxenes from spinel and plagioclase lherzolites. Shaded field represent clinopyroxene from abyssal peridotites (Johnson et al., 1990). Normalising values after Sun and McDonough (1989); (b) Primitive mantle-normalised trace element patterns of clinopyroxene from spinel and plagioclase lherzolites.

**Table 2**
Sr and Nd isotopic composition for New Caledonia spinel and plagioclase lherzolites.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithology</th>
<th>Type</th>
<th>Outcrop</th>
<th>Measured ratio $^{143}$Nd/$^{144}$Nd</th>
<th>Nd (ppm)</th>
<th>$^{143}$Sm/$^{144}$Nd</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>$^{147}$Sm/$^{144}$Nd</th>
<th>Age corrected ratio (53 My) $^{143}$Nd/$^{144}$Nd</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>$^{147}$Sm/$^{144}$Nd</th>
<th>$^{87}$Rb/$^{86}$Sr</th>
<th>$^{87}$Rb/$^{86}$Sr</th>
</tr>
</thead>
<tbody>
<tr>
<td>BA1</td>
<td>Sp lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513384 ± 10 0.513407 ± 6 0.513402 ± 8 0.513230 ± 13</td>
<td>0.02 0.02 0.03 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>BA1 Bpx</td>
<td>Sp lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513407 ± 10 0.513407 ± 6 0.513402 ± 8 0.513230 ± 13</td>
<td>0.02 0.02 0.03 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>BA2</td>
<td>PI lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.02 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>POU2</td>
<td>PI lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.02 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>POU1A</td>
<td>PI lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.02 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>POU2B</td>
<td>BA1 Bpx</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513407 ± 10 0.513407 ± 6 0.513402 ± 8 0.513230 ± 13</td>
<td>0.02 0.02 0.03 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513398 ± 13 0.513398 ± 8 0.513398 ± 8</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>KO1</td>
<td>Sp lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513407 ± 10 0.513407 ± 6 0.513402 ± 8 0.513230 ± 13</td>
<td>0.02 0.02 0.03 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
<tr>
<td>BOU1</td>
<td>PB lhz</td>
<td>WR</td>
<td>Babouillat</td>
<td>0.513407 ± 10 0.513407 ± 6 0.513402 ± 8 0.513230 ± 13</td>
<td>0.02 0.02 0.03 0.02</td>
<td>1.093 0.772 0.867 0.706704 ± 5</td>
<td>0.706024 ± 5</td>
<td>1.225 0.772</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>0.513408 ± 7 0.513408 ± 7</td>
<td>1.046 1.046</td>
<td>0.02 0.02</td>
<td>0.23 0.17</td>
</tr>
</tbody>
</table>

All errors are 2σ and relate to the last significant digits. Decay constants for $^{87}$Rb = 1.42 × 10$^{-11}$y$^{-1}$, for $^{144}$Sm = 6.54 × 10$^{-12}$y$^{-1}$. 


process, we carried out trace element modelling based on whole rock and clinopyroxene trace element compositions. We initially applied a non-modal fractional melting model, using the equations of Albarede (1996) and assuming that melting occurred entirely in the spinel stability field. Details of the model and adopted partition coefficients are reported in Table S5.

The results, shown in Fig. 7a, indicate that HREE patterns of spinel lherzolites are broadly consistent with low melting degrees (6%–8%) of a depleted mantle source, whereas MREE would require higher degrees of melting. Similar discrepancies have been also observed applying fractional melting model to reproduce the clinopyroxene REE composition of spinel lherzolites (Fig. 7b). For the clinopyroxene, the estimated melting degrees vary between 4% and 7%. Moreover, the La–Ce enrichments are not consistent with melting processes and they will be discussed later (see Section 8.3).

On the whole, fractional melting under spinel facies conditions fail to reproduce the REE patterns of spinel lherzolites. In particular, relatively low MREE/HREE ratios are coupled with high YbN in clinopyroxene and reproduce the REE patterns of spinel lherzolites. In particular, relatively high Na2O content of clinopyroxenes from spinel lherzolites (Table S2) compared with their very low Nd abundances (NdN = 0.24–0.33) are also consistent with a polybaric melting scenario. In fact, Na compatibility during partial melting increases with increasing pressure whereas Nd behaviour is nearly constant, as discussed by Müntener et al. (2010). The Na2O and Nd contents of the Poum and Babouillat clinopyroxenes (samples POU2 and BA1, respectively) plot on polybaric melting curves calculated by Müntener et al. (2010), which include 4% melting in the garnet solidus (Takahashi and Kushiro, 1983). On the basis of the estimated pressure for the garnet–spinel transition in the investigated lherzolites (–2.2 GPa, see Section 6) the garnet melting stage may be tentatively constrained to the 2.5–2.2 GPa interval.

Hence, polybaric fractional melting of a depleted mantle source starting in the garnet stability field, followed by additional melting in the spinel field, was also computed for REE and other incompatible trace element data (Sr, Zr, Hf, Y, and Ti). The obtained patterns (Fig. 8) faithfully replicate the observed whole rock and clinopyroxene compositions. In particular, 4% of melting in the garnet stability field followed by 4–5% melting in the spinel stability field is required to reproduce the REE patterns of clinopyroxenes from the Poum (sample POU2) and Babouillat massifs (sample BA1).

A good agreement is also found for the other trace element data, with the exception of Zr, which is invariably lower than the predicted values. Although the occurrence of Zr depletion in mantle clinopyroxenes has been documented by various authors (Johnson et al., 1990; McDonough et al., 1992; Rampone et al., 1991; Salters and Shimizu, 1988), its origin remains uncertain. Negative Zr anomalies in clinopyroxene may be balanced by corresponding positive anomalies in orthopyroxene as a result of different partitioning behaviour between mantle minerals (McDonough et al., 1992; Rampone et al., 1991). However, this cannot produce the observed bulk rock Zr anomalies (Figs. 4, 5). During the spinel field partial melting, Zr and Hf do not behave coherently, i.e. Zr is more incompatible than Hf (Dbulk, Zr/Dbulk, Hf = –0.5, see also Table S5 and Niu, 2004), leading to strong Zr depletion in the residuum. In a garnet-bearing assemblage, however, the difference of compatibility is negligible, significantly reducing the Zr anomaly (see Fig. 8). Alternatively, the occurrence of such prominent Zr/Hf fractionation in both residual whole rocks and clinopyroxenes could be the consequence of chromatographic effects related to melt percolation at low melt/rock ratios, as proposed by Niu (2004).

The onset of melting in the garnet stability field indicates that the melting region extended to great depths (possibly exceeding 70–80 km). This deepening of the melting column may be related to high potential mantle temperatures or the degree of mantle fertility (e.g. see Brunelli et al., 2006). On the other hand, the estimated bulk degree of melting does not exceed 8%–10%, and the melting in the spinel stability field is limited, which suggests that conductive cooling from above prevented protracted decompression melting to shallower levels. Such
a melting regime is typical of slow-spreading ridges (e.g. Hellebrand et al., 2002) or may indicate proximity of the investigated peridotites to a transform fault (Leblanc, 1995; Nicolas, 1989).

8.2. Origin of the plagioclase lherzolites

In peridotitic rocks, plagioclase rims around spinel may form during subsolidus evolution from spinel-to-plagioclase facies conditions or as a result of a reaction involving a melt. Metamorphic equilibria governing the spinel-to-plagioclase-facies transition (enstatite + diopside + spinel = anorthite + forsterite and enstatite + Cat-Tsch = anorthite + forsterite, Rampone et al., 1993) imply formation of olivine together with plagioclase during subsolidus transformations. In contrast, we observe formation of a new generation of orthopyroxene together with plagioclase, at the expense of clinopyroxene and olivine. Textural evidence thus argues against a closed system subsolidus evolution from spinel- to plagioclase-facies conditions, suggesting that plagioclase crystallisation may be related to a reaction with (SiO$_2$-saturated) melt.

The spilite composition (Fig. 2) provides further support for this hypothesis. In fact, the plagioclase lherzolites contain spinels with significantly higher Cr# than spinel lherzolites, partially overlapping those of refractory harzburgites in New Caledonia, although at lower Mg#. Remarkably, these spinels plot outside the partial melting trend defined by the Cr#–Mg# mantle array. High and variable Cr# values are also associated with distinctly higher TiO$_2$ contents (0.21–0.49 wt.%) compared to spinel lherzolites (Fig. 2c). Similar compositional variations of spinel are commonly observed in plagioclase peridotites that were affected by melt/rock interaction processes (e.g. Cannat et al., 1990; Dick and Bullen, 1984; Kaczmarek and Mün tenere, 2008; Seyler and Bonatti, 1997).

The formation of plagioclase lherzolites is also accompanied by Al$_2$O$_3$ and CaO increase and incompatible trace element enrichment with respect to the spinel lherzolites (Figs. 3–5). The chemical modifications associated with plagioclase crystallisation may derive from simple melt entrapment and trace element re-equilibration of minerals with the melt (Mün tenere et al., 2010) or may be related to a process where the infiltrating melt reacts with the solid to variable extents (e.g. Rampone et al., 2008).

![Fig. 7. Results of fractional melting modelling in the spinel stability field. (a) Chondrite-normalised REE patterns for whole rock residues formed by fractional melting of a spinel peridotite (light grey). Numbers next to grey lines are the degrees of melting. DMM = depleted MORB mantle (Sal ters and Stracke, 2004). Melting modes and crystal/liquid partition coefficients are reported in Table S5; (b) Chondrite-normalised REE patterns for residual clinopyroxenes after different degrees of fractional melting. DMM clinopyroxene after Hirschmann and Stolper (1996).](image-url)
To provide further constraints on the impregnation process we have first attempted to simulate the chemical effect of simple melt entrapment in a spinel-bearing residual lherzolite. We have calculated the trace element composition (REE, Zr, Y, and Ti) of clinopyroxene equilibrated with a plagioclase-bearing assemblage produced from mixing between a spinel lherzolite protolith and small fractions of different liquids with highly depleted compositions. The liquids used for the simulation were a “boninite”-like melt and an incremental melt fraction derived from a depleted asthenospheric mantle source under plagioclase-facies conditions. The resulting clinopyroxene compositions are reported in Fig. 9. For both types of liquid, the models fail to reproduce the convex-up REE patterns and distinctly higher MREE-Y contents in the observed clinopyroxenes. On the other hand, assuming higher proportions of trapped melt would overestimate the calculated bulk rock compositions. Models involving more enriched liquids, e.g. aggregated MORB melts generated by variable degrees of partial melting (5%–15%), are not applicable because they would produce impregnated rocks with significantly higher incompatible element enrichments (up to an order of magnitude), even for very small amounts of melt (1%–2%).

Textural evidence (e.g. widespread occurrence of embayed clinopyroxene porphyroclasts and the local presence of orthopyroxene + plagioclase aggregates) suggests that the formation of plagioclase-bearing assemblages may be related to a reactive process, whereby a silica-oversaturated liquid crystallised plagioclase + orthopyroxene, while dissolving clinopyroxene and olivine. To test this hypothesis, we applied an AFC model, following the approach adopted by Rampone et al. (2008) for the impregnated peridotites from the Corsican ophiolites. The compositions of clinopyroxenes in equilibrium with the reacting melts with decreasing melt mass are plotted in Fig. 10. The same depleted melt is shown to have similar geochemical characteristics (e.g. Piccardo and Guarnieri, 2012; Rampone et al., 2008). In general, the formation of impregnated plagioclase lherzolites in the New Caledonia mantle section reflects in situ melt entrapment at relatively shallow levels after polybaric melt extraction starting in the garnet stability field. The impregnation process at plagioclase-facies conditions most likely involves incremental melt fractions that survived unmixed before aggregation. In some alpine ophiolitic mantle sections, the impregnating melt is shown to have similar geochemical characteristics (e.g. Piccardo and Guarnieri, 2012; Rampone et al., 2008). The close association of the plagioclase lherzolites from the Poum Massif with a major mylonitic shear zone, interpreted as an oceanic paleo-transform fault (Leblanc, 1995; Nicolas, 1989), may be consistent with a scenario of melt focusing along the shear zone, leading to melt stagnation and melt-rock reaction in the undeformed peridotites nearby.

8.3. Late-stage enrichment processes in bulk rock and clinopyroxene

The New Caledonia residual lherzolites display enrichments in the most incompatible trace elements (from Cs to Sr in the extended diagrams normalised to PM, Figs. 4–5) in both the bulk rock and constituent clinopyroxenes that cannot be accounted for by melting processes. These selective enrichments require different processes, possibly related to melt-rock or fluid–rock interaction.

In abyssal and ophiolitic residual peridotites the enriched signature of clinopyroxene may result from inefficiently extracted melt (Brunelli et al., 2006; Hellebrand et al., 2002; Niu, 2004). Nevertheless, the lack of textural evidence, i.e. interstitial, secondary clinopyroxene or thin films around primary mantle minerals (Seyler et al., 2001), does not

Fig. 8. Results of fractional melting modelling in the garnet and spinel stability fields for clinopyroxene. All data are chondrite-normalised. DMM Cpx after Hirschmann and Stolper (1996).
support this hypothesis. Alternatively, the cryptic enrichment shown by clinopyroxene can be driven by chromatographic processes related to percolation of small melt fractions into already depleted peridotites (Godard et al., 1995; Navon and Stolper, 1987; Vernières et al., 1997). Percolation may have occurred either during the sub-ridge evolution or during the subsequent involvement of the New Caledonia lherzolites in a convergent setting during Eocene (Aitchison et al., 1995a; Cluzel et al., 2001, 2012). Very low Th/Nb ratios (<0.05) recorded by clinopyroxenes cores, however, tend to rule out an origin related to slab-derived melts in subduction setting (Dijkstra et al., 2009).

Bulk rock compositions are more enriched in Ba, Pb, Sr, La and Ce than clinopyroxenes (Figs. 4, 5). As no textural evidence for modal metasomatism is discernible, the incompatible elements in “excess” may be stored along grain boundaries, in small amounts of fluid/melt inclusions trapped in minerals (Garrido et al., 2000) or in secondary alteration minerals.

Addition of incompatible elements by the highly depleted melts involved in the formation of the plagioclase lherzolites (Section 8.2) is unlikely. Overall, the lack of meaningful correlation with HFSE abundances rules out the possibility that selective increases of bulk-rock Pb, Pb, Sr, La and Ce contents are related to melt–rock interaction (Paulick et al., 2006). On the other hand, these elements are well known to be mobile during seafloor hydrothermal alteration (Deschamps et al., 2010; Paulick et al., 2006) and in subduction-related fluids (Bebout, 2014). We therefore propose that the fluid-mobile element enrichments, and the associated displacement of \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios towards radiogenic values is a result of low to moderate degrees of serpentinisation (Fig. 6), likely related to hydrothermal seafloor alteration. Although the role of subduction-related fluids cannot be ruled out from existing data, direct evidence for their involvement is lacking.

### 8.4 Isotopic constraints on the mantle sources

As shown in Fig. 6, most lherzolite samples have initial Sr isotopic compositions well above 0.705 and display a subhorizontal trend in a Sr–Nd plot, showing increasing \(^{87}\text{Sr}/^{86}\text{Sr}\) against slightly variable \(^{143}\text{Nd}/^{144}\text{Nd}\). Similar trends are commonly observed in peridotites affected by hydrothermal alteration (Bosch et al., 2004; Cipriani et al., 2004; Rampone and Hofmann, 2012), which shifts the Sr isotopic composition towards the radiogenic values of seawater. In contrast, Nd is unaffected by these alteration processes so that its isotope composition may be considered representative of its mantle source, even in highly serpentinised rocks.
Initial Nd isotope ratios of both spinel and plagioclase lherzolites are similar to the DMM composition (Fig. 6). The analysed samples have relatively homogeneous Nd isotope composition compared to the heterogeneity displayed by modern abyssal peridotites (Fig. 6). In addition, somewhat uniform $^{143}$Nd/$^{144}$Nd ratios coupled with variable $^{147}$Sm/$^{144}$Nd ratios (Fig. 11), suggest that the New Caledonia lherzolites represent a residual mantle source of asthenospheric origin that experienced a recent (MORB-producing?) depletion event (see also Rampone and Hofmann, 2012). Likewise, the young Nd model ages obtained for New Caledonia lherzolites ($T_{DM} < 80$ Ma, Table 2) indicate recent melting of a depleted asthenospheric source. We therefore propose that this event may be related to decompression melting in a marginal basin before the onset of Eocene subduction (Aitchison et al., 1995a, 1995b; Cluzel et al., 2001, 2012).

8.5. Crust–mantle relationships: assessment of a genetic link between Poya Terrane basalts and northern lherzolites

Based on a batch melting model, Ulrich et al. (2010) recently proposed that the melts extracted from the lherzolites were similar to the E-MORB basalts of the Poya Terrane. In the light of our new trace element and Nd isotopic data, we re-examine this hypothesis. To this end, we have computed the REE composition of aggregated liquids derived from melting of a depleted mantle source (Table S5). The calculated liquids were performed using the garnet-spinel fractional melting model which successfully explains the residual spinel lherzolite compositions (see Section 8.1). The obtained liquids (Fig. 12a) were then compared with the Poya E-MORB. These basalts display flat LREE patterns and nearly flat to slightly positive MREE/HREE ratios, which argue for the involvement of garnet in their sources (e.g. Hirschmann and Stolper, 1996). The calculated liquids formed after 4% of melting in the garnet stability field followed by 4%–6% of melting in the spinel peridotite field have REE patterns characterised by flat to slightly depleted LREE and weakly fractionated MREE/HREE, subparallel to those of Poya E-MORBs, but at lower absolute concentrations. On the other hand, a smaller degree of garnet involvement of garnet in their sources (e.g. Hirschmann and Stolper, 1996). The calculated liquids formed after 4% of melting in the garnet stability field followed by 4%–6% of melting in the spinel peridotite field have REE patterns characterised by flat to slightly depleted LREE and weakly fractionated MREE/HREE, subparallel to those of Poya E-MORBs, but at lower absolute concentrations. On the other hand, a smaller degree of garnet involvement of garnet in their sources (e.g. Hirschmann and Stolper, 1996). 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The calculated liquids formed after 4% of melting in the garnet stability field followed by 4%–6% of melting in the spinel peridotite field have REE patterns characterised by flat to slightly depleted LREE and weakly fractionated MREE/HREE, subparallel to those of Poya E-MORBs, but at lower absolute concentrations. On the other hand, a smaller degree of garnet involvement of garnet in their sources (e.g. Hirschmann and Stolper, 1996). The calculated liquids formed after 4% of melting in the garnet stability field followed by 4%–6% of melting in the spinel peridotite field have REE patterns characterised by flat to slightly depleted LREE and weakly fractionated MREE/HREE, subparallel to those of Poya E-MORBs, but at lower absolute concentrations. On the other hand, a smaller degree of garnet involvement of garnet in their sources (e.g. Hirschmann and Stolper, 1996). 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On the whole, the modelling indicates that these basalts require a source more enriched in incompatible elements with respect to a typical MORB-type mantle.

The lherzolites from the Babouillat and Poum massifs and the Poya basalts display distinct Nd isotope compositions (Fig. 6). In particular, the lherzolites are characterised by age-corrected Nd isotope compositions significantly more radiogenic ($\varepsilon_{Nd(i)}$ from $+6.98$ to $+10.97$) than those shown by the Poya basalts ($\varepsilon_{Nd(i)}$ from $+2.77$ to $+5.52$, Cluzel et al., 2001 and this work).

In summary, geochemical trace element modelling and Nd isotopes do not seem to support a genetic mantle–crust link between the Poya Terrane basalts and the New Caledonian lherzolites that underlie them.

Mantle–crust relations at oceanic settings is a debated issue. Radiogenic isotope studies have shown that abyssal and ophiolitic peridotites have more heterogeneous isotopic compositions than the crust to which they are related (e.g. Mallick et al., 2014; Rampone and Hofmann, 2012). Our findings are also consistent with the notion that abyssal peridotites are, on average, isotopically more depleted than MORB. Nd isotopic variability of the crustal rocks may reflect different proportions of variably depleted peridotite and the presence a low solidus component with a more enriched isotopic signature in the subridge melting region (Salters and Dick, 2002; Sobolev et al., 2007). We can therefore speculate that melting of mafic heterogeneities stored in the convecting mantle could have contributed to the enriched geochemical and Nd isotopic signature of the Poya E-MORB (Montanini et al., 2012; Sobolev et al., 2007). Hence, the lack of a simple residue-melt relationship does not preclude that the lherzolites were involved in the basalt petrogenesis, if a more enriched, non-peridotitic component was present in the source. We conclude that geochemical data cannot unambiguously determine whether the lherzolites and the Poya basalts represented the oceanic lithosphere of a single marginal basin.

9. Summary and conclusions

In this work, we have carried out a geochemical investigation on the New Caledonia lherzolites, providing a comprehensive geochemical data set (major, trace element and Sr-Nd isotopes) on these rocks. Our...
study reconstructs the evolution of the peridotites from a recent event of melt extraction starting in the garnet stability field, to shallow level melt impregnation. Mantle–crust relationships in light of geochemical modelling and Nd isotope compositions were also discussed. Mineral and bulk rock geochemistry indicate that the New Caledonia lherzolites represent moderately depleted residual peridotites with abyssal-type affinity. Bulk rock and clinopyroxene compositions of spinel lherzolites are consistent with moderate degrees of fractional melting of a DMM source (8%–9%), including a first stage of deep melting in the presence of residual garnet. According to the Sm–Nd isotope systematics, the lherzolites are derived from a relatively homogeneous asthenospheric reservoir that experienced a recent depletion event. Plagioclase lherzolites are formed by post-melt extraction evolution of the depleted spinel lherzolites. Geochemical modelling suggests that plagioclase crystallised following trapping of highly depleted, incremental melt fractions of a MORB-type source.

The New Caledonia lherzolites record a history of partial melting in the asthenosphere, and subsequent melt–rock interaction at shallow, lithospheric depths. This evolution was likely related to seafloor spreading in a marginal basin predating Eocene subduction. Geochemical trace element modelling and Nd isotopes do not support a single genetic mantle–crust link between the northern massif lherzolites and basaltic rocks with E-MORB affinity from the Poya Terrane. These basalts require the presence of isotopically enriched material, possibly non-peridotitic, in their mantle source.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.lithos.2016.04.030.

Acknowledgments

The authors are grateful to B. Galland, C. Douchet, O. Bruguier, P. Verdoux and P. Telouk for their expertise in chemical room management, ICP-MS, TIMS and MC-ICP-MS analyses, respectively. We are indebted to Daniele Brunelli (University of Modena) and Roberto Braga (University of Bologna) for the XRF analyses. We thank P. Sossi (Institute de Physique du Globe de Paris) and S.L. Williams (University of Parma) for the revision of the English text. This paper is part of the PhD work of A.S. which was supported by a Vinci grant (Italian–French University).

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