Ultra-refractory Domains in the Oceanic Mantle Lithosphere Sampled as Mantle Xenoliths at Ocean Islands

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Many peridotite xenoliths sampled at ocean islands appear to have strongly refractory major element and modal compositions. To better constrain the chemistry, abundance and origin of these ultra-refractory rocks we compiled a large number of data for xenoliths from nine groups of ocean islands. The xenoliths were filtered petrographically for signs of melt infiltration and modal metasomatism, and the samples affected by these processes were excluded. The xenolith suites from most ocean islands are dominated by ultra-refractory harzburgites. Exceptions are the Hawaii and Tahiti peridotites, which are more fertile and contain primary clinopyroxene, and the Cape Verde suite, which contains both ultra-refractory and more fertile xenoliths. Ultra-refractory harzburgites are characterized by the absence of primary clinopyroxene, low whole-rock Al2O3, CaO, FeO/MgO and heavy rare earth element (HREE) concentrations, low Al2O3 in orthopyroxene (generally <5 wt %), high Cr-number in spinel (93–0.8) and high forsterite contents in olivine (averages >91.5). They are therefore on average significantly more refractory than peridotites dredged and drilled from mid-ocean ridges and fracture zones. Moreover, their compositions resemble those of oceanic forearc peridotites. The formation of ultra-refractory ocean island harzburgites requires potential temperatures above those normally observed at modern mid-ocean ridges, and/or fluid fluxed conditions. Some ultra-refractory ocean island harzburgites give high Os model ages (up to 3300 Ma), showing that their formation significantly pre-dates the oceanic crust in the area. A genetic relationship with the host plume is considered unlikely based on textural observations, equilibration temperatures and pressures, inferred physical properties, and the long-term depleted Os and Sr isotope compositions of some of the harzburgites. Although we do not exclude the possibility that some ultra-refractory ocean island harzburgites have formed at mid-ocean ridges, we favor a model in which they formed in a process spatially and temporally unrelated to the formation of the oceanic plate and the host plume. As a result of their whole-rock compositions, ultra-refractory harzburgites have a very high solidus temperature at a given pressure, low densities and very high viscosities, and will tend to accumulate at the top of the convecting mantle. They may be preserved as fragments in the convecting mantle over long periods of time and be preferentially incorporated into newly formed lithosphere.

KEY WORDS: mantle xenoliths; ocean islands; recycled oceanic mantle lithosphere; ultra-depleted mantle

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INTRODUCTION

Detailed information about chemical composition is essential in understanding the formation and evolution of different parts of the Earth’s mantle. Small variations in the major element composition of the lithospheric mantle may strongly influence its physical properties and its behavior in response to stress (e.g. Karato, 1986; Hirth & Kohlstedt, 1996; Afonso et al., 2007; Simon & Podladchikov, 2008). It is therefore necessary to obtain reliable chemical compositions of different mantle domains to develop realistic geodynamic models and correctly interpret geophysical data.

Mantle xenoliths sampled in ocean islands (OI) cover a wide range in compositions. Some islands are dominated by spinel harzburgites that are more refractory than most peridotites sampled along mid-ocean ridges and fractures zones (MORP: mid-ocean ridge peridotites). The xenoliths in a second group of islands are relatively fertile, whereas in a third group of ocean islands the xenoliths range from strongly refractory to fertile. In the third group the fertile compositions are to a large extent the results of metasomatism associated with ocean island formation and/or melt–rock interaction during ascent, and many of the more fertile peridotites have strongly refractory pre-OI protoliths (e.g. Ryabchikov et al., 1993; Wulff-Pedersen et al., 1996; Grégoire et al., 1997, 2000a, 2000b; Moine et al., 2000; Neumann et al., 2002, 2004; Delpech, 2004; Bonadiman et al., 2005; Shaw et al., 2006).

The presence and abundance of ultra-refractory mantle not only in continental keels, but also in oceanic domains, is an important observation. Most of this evidence for ‘depleted compositions’ is based on isotope and trace element data on single minerals, samples and/or locations. However, to use compositional variations in geodynamic modeling, whole-rock major element compositions and modal mineralogy averaged over some larger area are actually the most important variables. This study represents a first step in providing such constraints for more petrologically consistent quantitative geodynamic modeling. Moreover, the mechanism for the generation of these samples is far from clear and this review will serve as a solid basis for further work.

Studies of mantle xenoliths in ocean islands have so far mainly been focused on mantle metasomatism and melt–rock interaction during ascent. These are recent processes in the evolution of the peridotites. We are interested in the pre-ocean island composition of ocean island peridotites and particularly in the most refractory xenoliths. The aim of this study is (1) to establish the distribution of strongly refractory peridotites in ocean islands, (2) to constrain their origin and mode of formation, and (3) to discuss the implications of strongly refractory mantle material for geodynamics with respect to physical and chemical properties.

We have compiled available (published and unpublished) major element whole-rock and mineral composition data for 294 harzburgite and lherzolite xenoliths and their minerals from the Canary Islands (Neumann, 1991; Neumann et al., 1995, 2002, 2004; Wulff-Pedersen et al., 1996; Abu El-Rus et al., 2006), Madeira (Munha et al., 1990), the Azores (Johansen, 1996), Cape Verde (Ryabchikov et al., 1993; Bonadiman et al., 2005; Shaw et al., 2006), Kerguelen (Grégoire et al., 1997, 2000a, 2000b; Delpech, 2004), Grande Comore (Coltorti et al., 1999), Samoa (Hauri et al., 1993; Hauri & Hart, 1994), Hawaii (e.g. Jackson & Wright, 1970; Sen, 1987; Goto & Yokoyama, 1988; Sen & Leeman, 1991; Bizimis et al., 2003) and Tahiti (Tracy, 1980; Qi et al., 1994). We refer to samples from the same island as a xenolith suite. The data are presented in an Electronic Appendix (available for downloading at http://www.petrology.oxfordjournals.org). The database also gives the major element compositions of 450 MORP used for comparison.

Our study is based mainly on the most abundant major elements ($\text{SiO}_2$, $\text{MgO}$, $\text{FeO}$, $\text{Al}_2\text{O}_3$ and $\text{CaO}$) because these are less easily reset by fertilization processes than incompatible trace elements. In addition to spinel harzburgites and lherzolites, mantle xenoliths from OI include dunites, wehrlites and pyroxenites. Dunites, wehrlites, pyroxenites, etc. are not discussed here as these rock-types are closely associated with fertilization processes (e.g. Frey, 1980; Sen, 1988; Neumann, 1991; Sen & Leeman, 1991; Kelemen et al., 1992, 1998; Grégoire et al., 1997, 2000b; Neumann et al., 2004). However, data on some of those samples are included in the database.

COMPOSITIONAL VARIATIONS

Classification of samples

Because the focus of this study is on pre-ocean-island (pre-OI) compositions we apply petrographic criteria to recognize OI peridotites that have been least affected by melt–rock interaction processes. The effects of such processes range from very minor, reflected only in mild enrichment in strongly incompatible trace elements and resetting of radiogenic isotopes, to extensive petrographic changes such as those involving the formation of hydrous minerals, reactions involving the growth of olivine±clinopyroxene±spinel (±ol±cpx±sp) at the expense of orthopyroxene (cpx), formation of poikilitic orthopyroxene and clinopyroxene, glass-rich reaction zones on primary minerals (e.g. orthopyroxene and clinopyroxene) and sieve-textured spinel and clinopyroxene (Fig. 1), in addition to significant enrichment in incompatible trace elements. Peridotites showing these textures are classified as OI2, indicating the presence of a second generation of minerals related to melt–rock reaction (Table 1).

Those xenoliths that show no petrographic evidence of metasomatism (as listed above and summarized
in Table 1) are classified as OI1 (Fig. 2). These xenoliths typically show porphyroclastic to protogranular textures with deformed porphyroclasts of olivine and exsolved orthopyroxene; in some xenolith series the porphyroclast assemblage comprises exsolved clinopyroxene (e.g. from Hawaii and Tahiti) and/or spinel. The matrix consists of less deformed or undeformed neoblasts of olivine and exsolved orthopyroxene. In some cases the published petrographic descriptions do not give enough information to classify a given sample; such samples are assigned to the OI2 group. Because this study is aimed at investigating the composition of the upper mantle before the onset of ocean island magmatism the processes that have led to the petrographic changes recorded in the OI2 samples are of no interest here, and are not discussed. We therefore make no effort to distinguish between in situ metasomatism in the mantle and interaction with the host magma during ascent. Data points in the figures are restricted to OI1 xenoliths. However, as it is of interest to the discussion to know the extent to which melt–rock
Table 1: Summary of textural and geochemical characteristics of the peridotite groups

<table>
<thead>
<tr>
<th></th>
<th>Ultra-refractory (OI-u)</th>
<th>Fertile (OI-f)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>OI1-u</strong></td>
<td></td>
<td></td>
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<tr>
<td>porphyroclastic to protogranular</td>
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<td></td>
</tr>
<tr>
<td>no petrographic evidence of melt-rock interaction</td>
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<td></td>
</tr>
<tr>
<td>deformed porphyroclasts of ol + opx</td>
<td></td>
<td></td>
</tr>
<tr>
<td>matrix: less deformed or undeformed neoblasts of ol + opx + sp ± cpx</td>
<td></td>
<td></td>
</tr>
<tr>
<td>primary cpx appears absent</td>
<td></td>
<td></td>
</tr>
<tr>
<td>high WR MgO: 39–48 wt %</td>
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<td></td>
</tr>
<tr>
<td>low WR CaO: 0.4–1.2 wt %</td>
<td></td>
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<tr>
<td>low WR Al2O3: 0.2–1.3 wt %</td>
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<td></td>
</tr>
<tr>
<td>[Al2O3]opx: 0.4–4.3 wt %</td>
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<td></td>
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<tr>
<td>main range 3–0 wt %</td>
<td></td>
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<tr>
<td>high Cr-no sph: 0.3–0.9; main range 0.4–0.7</td>
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<tr>
<td>narrow range in high Fos: 89.7–92.6</td>
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<tr>
<td>low HREEopx Yb: 0.04–0.04</td>
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<tr>
<td>LREE-depleted to U-shaped REEopx patterns</td>
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<tr>
<td><strong>OI2-u</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>porphyroclastic/protoporphyroclastic/poikilitic</td>
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<tr>
<td>petrographic evidence of melt-rock reactions, modal</td>
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<tr>
<td>metasomatism, etc. (Fig. 1)</td>
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<tr>
<td>primary cpx appears to be absent</td>
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<tr>
<td>lower WR MgO: 39–48 wt %</td>
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<tr>
<td>wide range in WR CaO: 0.3–3.6 wt %</td>
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<tr>
<td>wide range in WR Al2O3: 0.3–3.7 wt %</td>
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<tr>
<td>wide range in [Al2O3]opx: 0.1–4.0 wt %</td>
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<tr>
<td>wide range in Cr-no sph: 0.2–0.9</td>
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<tr>
<td>wide range to low Fos: 84.6–91.5</td>
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<tr>
<td>higher HREEopx Yb: 0.1–2</td>
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<tr>
<td>LREE-depleted to U-shaped REEopx patterns</td>
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<tr>
<td><strong>OI1-f</strong></td>
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<td>textures as in OI1-u (Fig. 2)</td>
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<tr>
<td>no petrographic evidence of melt-rock interaction</td>
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<tr>
<td>deformed porphyroclasts of ol + opx + sp ± cpx</td>
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<tr>
<td>matrix: less deformed or undeformed neoblasts of ol + opx + sp ± cpx</td>
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<tr>
<td>up to 20 vol. % cpx</td>
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<tr>
<td>primary clinopyroxene common</td>
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<tr>
<td>low WR MgO: 38.45 wt %</td>
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<tr>
<td>high WR CaO: 0.9–4.6 wt %</td>
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<tr>
<td>high WR Al2O3: 0.6–3.7 wt %</td>
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<td></td>
</tr>
<tr>
<td>higher [Al2O3]opx: 1.7–6.0 wt %</td>
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<tr>
<td>main range 2–6 wt %</td>
<td></td>
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<tr>
<td>low Cr-no sph: 0.1–0.5; main range 1–0.4</td>
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<tr>
<td>wide range in Fos: 83.6–92.0</td>
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<tr>
<td>low HREEopx Yb: 0.2–0.04</td>
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<tr>
<td>LREE-depleted to U-shaped REEopx patterns</td>
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<tr>
<td><strong>OI2-f</strong></td>
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<tr>
<td>petrographic evidence of melt-rock reactions, modal</td>
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<td></td>
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<tr>
<td>metasomatism, etc. (Fig. 1)</td>
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<tr>
<td>low WR MgO: 37.45 wt %</td>
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<td></td>
</tr>
<tr>
<td>high WR CaO: 0.8–3.9 wt %</td>
<td></td>
<td></td>
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<tr>
<td>high WR Al2O3: common: 0.6–3.9 wt %</td>
<td></td>
<td></td>
</tr>
<tr>
<td>high [Al2O3]opx: 2.9–6.0 wt %</td>
<td></td>
<td></td>
</tr>
<tr>
<td>low Cr-no sph: 0.03–0.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>wide range to low Fos: 83.6–91.7</td>
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</table>

WR, whole-rock.

interaction has changed pre-OI peridotite compositions, we show the ranges covered by OI2 xenoliths in some diagrams.

In addition to the petrographic distinction into OI1 and OI2, the xenoliths from different islands fall in two categories with respect to major element compositions (Table 1). One category comprises ultra-refractory harzburgites (OI-u) with low contents of clinopyroxene (<3 vol. %) and other fusible components; the other category consists of more fertile xenolith series (OI-f), in which lherzolites with primary clinopyroxene are common. Essential petrographic and chemical differences between the two groups are outlined in Table 1. The focus of this study is on the OI-u xenoliths.

We emphasize that we distinguish between refractory/fertile, which refers to major element compositions and ultimately solidus temperatures, and the terms depleted/enriched, which refer to trace element and isotope characteristics. These chemical characteristics may be decoupled. A refractory harzburgite may, for example, be enriched in certain incompatible elements, but still retains its high solidus temperature and related physical properties. Most samples in the OI-u group show some evidence of cryptic metasomatism [i.e. U-shaped rare earth element (REE) patterns]. We have, however, found a clear positive correlation between the extent of melt–rock interaction processes, as indicated by petrographic observations, and enrichment in strongly incompatible trace elements. On the basis of numerical modeling, Herzberg et al. (2007) concluded that a small fraction of trapped melt in a residual rock does not significantly affect the most abundant major elements. For the rocks discussed here this conclusion is supported by an excellent correlation between whole-rock major element chemistry and parameters such as cr-number [cation proportion Cr/(Cr + Al)] and mg-number [cation proportion Mg/(Mg + Fe3+)] in spinel, and the concentration of Al2O3 and FeO content in cores of orthopyroxene and olivine porphyroclasts, respectively.

In addition to chemical compositions we compare equilibrium temperatures (Table 2) for the various peridotite suites. OI1-u harzburgites lack primary clinopyroxenes, and secondary clinopyroxenes in these rocks are commonly too small to analyze. OI2 xenoliths show two or more stages of mineral growth, and it is difficult to ascertain which mineral compositions represent equilibrium. We therefore chose to use the single-mineral CaO-in-opx geothermometer of Sachtleben & Seck (1991), assuming that the presence of clinopyroxene exsolved from orthopyroxene satisfies the four-phase requirements, also for harzburgites without primary clinopyroxene. This is in agreement with the conclusions of Neumann (1991) and Neumann et al. (1995) based on the application of different geothermometers for OI1-u harzburgites from the Canary Islands. The most important characteristics of the xenolith suites from each ocean island are summarized below.

Canary Islands

Xenoliths from the Canary Islands belong to the OI-u group (Table 1). They are dominated by spinel harzburgites with protogranular, porphyroclastic or poikilitic textures
The OI1-u group is restricted to porphyroclastic to protogranular harzburgites with porphyroclasts of olivine (ol1) and orthopyroxene (opx1); in some samples large equidimensional spinels may also represent porphyroclasts (sp1). The groundmass consists of equant to irregular grains of olivine, orthopyroxene and spinel, and minor amounts (generally <3 vol. %) of clinopyroxene. Clinopyroxene is present as exsolution lamellae in orthopyroxene and as small interstitial grains, as reaction rims on orthopyroxene (Fig. 1a and c), in opx±sp±opx symplectites along grain boundaries and/or as part of reaction textures (formation of clinopyroxene±olivine at the expense of orthopyroxene; Fig. 1a and c), or as poikilitic grains enclosing round grains of olivine, embayed grains of orthopyroxene (Fig. 1d) and small spinel grains. Many xenoliths from Fuerteventura exhibit fibrous orthopyroxene. Some OI2 samples, particularly from La Palma and Tenerife, carry trace amounts of phlogopite.

No primary clinopyroxene porphyroblasts (cpx1) have been observed in the Canary Islands xenoliths; the clinopyroxene present (cpx2) is believed to have formed partly by exsolution from orthopyroxene porphyroclasts (opx1) and subsequent recrystallization (Fig. 2a), partly from interaction with infiltrating melts (Fig. 1a, d and e; Neumann et al., 1995, 2004). The samples with the highest modal proportions of clinopyroxene also show the strongest textural evidence of metasomatism.

Whole-rock major element relationships for Canary Islands OI1-u are plotted in Fig. 3. We have also plotted orthopyroxene–olivine (opx–ol) tie-lines, as defined by OI1-u xenoliths from Lanzarote. Major elements other than Na2O plot close to the opx–ol tie-line for most OI1-u xenoliths, consistent with the absence of primary clinopyroxene and confirming their strongly refractory composition. A few samples scatter at a moderate distance away from the opx–ol tie-line, for example in the SiO2–CaO diagram. We interpret this scatter as the result of minor metasomatism and/or melt–rock interaction. The wide range in Na2O is probably due to enrichment processes or...
Table 2: Equilibration temperatures estimated for ocean island xenoliths using the Sachtelen & Seek (1983) geothermometer based on CaO in orthopyroxene

<table>
<thead>
<tr>
<th>Island</th>
<th>Group</th>
<th>n</th>
<th>Range (°C)</th>
<th>Average (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Canary Islands</td>
<td>OI1-u</td>
<td>48</td>
<td>873–1209</td>
<td>1024</td>
</tr>
<tr>
<td></td>
<td>OI2-u</td>
<td>41</td>
<td>862–1215</td>
<td>1068</td>
</tr>
<tr>
<td>Madeira</td>
<td>OI1-u</td>
<td>7</td>
<td>891–1017</td>
<td>936</td>
</tr>
<tr>
<td></td>
<td>OI2-u</td>
<td>5</td>
<td>892–1182</td>
<td>1037</td>
</tr>
<tr>
<td>Azores</td>
<td>OI2-u</td>
<td>17</td>
<td>1003–1322</td>
<td>1135</td>
</tr>
<tr>
<td>Cape Verde</td>
<td>OI1-u</td>
<td>1</td>
<td>1163</td>
<td></td>
</tr>
<tr>
<td></td>
<td>OI1-f</td>
<td>4</td>
<td>1047–1233</td>
<td>1190</td>
</tr>
<tr>
<td></td>
<td>OI2-f</td>
<td>2</td>
<td>1028–1227</td>
<td>1127</td>
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<tr>
<td>Kerguelen</td>
<td>OI1-u</td>
<td>16</td>
<td>933–1104</td>
<td>1014</td>
</tr>
<tr>
<td></td>
<td>OI2-u</td>
<td>25</td>
<td>960–1193</td>
<td>1062</td>
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<td>Grande Comore</td>
<td>OI2-u</td>
<td>8</td>
<td>965–1239</td>
<td>1109</td>
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<tr>
<td>Samoa</td>
<td>OI1-u</td>
<td>6</td>
<td>983–1142</td>
<td>1063</td>
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<tr>
<td></td>
<td>OI2-u</td>
<td>13</td>
<td>1006–1228</td>
<td>1072</td>
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<tr>
<td>Hawaií islands</td>
<td>OI1-f</td>
<td>9</td>
<td>914–1205</td>
<td>1004</td>
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<td>OI2-f</td>
<td>23</td>
<td>881–1161</td>
<td>1005</td>
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<tr>
<td>Loihi seamount</td>
<td>OI1-u</td>
<td>4</td>
<td>1139–1182</td>
<td>1163</td>
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<tr>
<td>Tahiti</td>
<td>OI1-f</td>
<td>19</td>
<td>996–1108</td>
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<tr>
<td></td>
<td>OI2-f</td>
<td>1</td>
<td>1093</td>
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</tbody>
</table>

contamination (Neumann, 1991; Siena et al., 1991; Neumann et al., 1995, 2004). The strongly melt-depleted nature of the OI1-u harzburgites is also reflected in low Al2O3 in orthopyroxene (0.3–3.4 wt %), Cr-rich spinels (cr-number = 0.29–0.95, with the majority of the samples in the range 0.53–0.71), and relatively high Fo-content [100Mg/(Mg + Fe3+)] in olivine (90.7–92.5; Fig. 4).

Many OI2-u xenoliths also plot close to the opx–ol duplicate line (Fig. 3), although the majority have lower SiO2 and MgO, and higher CaO–FeO+ contents than the OI1-u xenoliths and show wide ranges in TiO2 and Na2O. Compared with the OI1-u rocks, the OI2-u rocks show a somewhat wider range in \( \text{Al}_2 \text{O}_3 \) in orthopyroxene, a similar range in cr-number, but tend towards lower mg-number (Fig. 4). Estimated equilibration temperatures for OI2-u xenoliths are 873–1209°C with a mean of 1024°C (Table 2). The OI2-u xenoliths cover a similar range (862–1215°C) but have a significantly higher average, 1068°C.

Cores of OI1-u orthopyroxene porphyroclasts (opx1) are strongly depleted in REE \( \langle \text{Yb}_N = 0.09–0.41, \text{La}_N = 0.002–1.2 \rangle \) and in light relative to heavy REE (LREE and HREE, respectively; Neumann et al., 1995, 2004; Fig. 5), which is typical for mantle rocks that represent residues after extensive partial melting (e.g. Johnson et al., 1990; Neumann et al., 2004). A tendency for enrichment in LREE relative to the middle REE (MREE) is most probably due to contamination (Neumann et al., 2002).

In contrast to the LREE-depleted opx1 in the OI1-u xenoliths, opx2 in the same rocks shows a wide range in REE patterns and concentrations, ranging from LREE-depleted in the clinopyroxenes with the lowest HREE, through U-shaped and flat patterns, to LREE-enriched patterns in clinopyroxenes with the highest HREE \( \langle \text{Yb}_N = 0.54–1.1, \text{La}_N = 0.22–128 \rangle \) (Fig. 6).

**Madeira**

Mantle xenoliths from Madeira comprise refractory deformed harzburgites and lherzolites (OI-u; Fo90–91) and less deformed dunites, websterites, wehrlites and clinopyroxenites (Fo77–96; Munha et al., 1990). Harzburgites and lherzolites are granular (most common) to porphyroclastic with deformed olivine and exsolved orthopyroxene porphyroclasts in a fine-grained matrix of ol + opx + sp ± cpx; minor amounts of plagiophylos may be present interstitially or as inclusions in clinopyroxene (in lherzolite).

OI1-u xenoliths from Madeira (Munha et al., 1990) have lower \( \langle \text{Al}_2 \text{O}_3 \rangle_{\text{opx}} \) (1.5–3.6 wt %) than orthopyroxene in MORB, but fall within the depleted range of MORB with respect to spinel compositions (cr-number: 0.26–0.53) and \( \text{Fool} (90.4–91.2) \) (Fig. 4). Temperature estimates for the OI1-u xenoliths are 891–1017°C with a mean of 936°C (Table 2); OI2-u xenolith give the wider range of 892–1182°C, with a mean of 1037°C.

**The Azores**

Mantle xenoliths from the Azores comprise Mg-Cr-rich, porphyroclastic spinel harzburgites (<2 vol. % clinopyroxene) and lherzolites (≤5 vol. % clinopyroxene) as well as granular Fe–Ti-rich spinel wehrlites, clinopyroxenites and dunites (Johansen, 1996; E. R. Neumann, unpublished data, 2007). Harzburgites predominate over lherzolites. The oldest generation of minerals in the harzburgites and lherzolites is irregular, deformed porphyroclasts of olivine and orthopyroxene with clinopyroxene exsolution lamellae, and occasional large spinels. The matrix appears to consist of two generations of neoblasts. The older neoblast generation consists of equigranular domains of polygonal neoblasts of olivine (occasionally with undulatory extinction) and rare unexsolved orthopyroxene; the younger generation is made up by fine-grained, glass-rich domains of irregular neoblasts of olivine, orthopyroxene, clinopyroxene, and spinel, without undulatory extinction (Johansen, 1996). Clinopyroxene occurs partly as small interstitial grains, partly as poikilitic grains. All Azores xenoliths are classified as OI2.

The Azores xenoliths are dominated by fertile whole-rock major element compositions, but are depleted with respect to \( \langle \text{Al}_2 \text{O}_3 \rangle_{\text{opx}} \) (2.1–2.3 wt %) and cr-number, sp (0.51–0.72; Fig. 4). \( \text{Fool} \) falls within the range 90.0–91.2 in all samples, with the exception of one with \( \text{Fool} \) of 86.5.
Equilibration temperatures are high (1003–1232°C, mean: 1135°C; Table 2). The textures of these rocks suggest that the contrast between whole-rock and mineral chemistry is the result of recent melt infiltration into a refractory peridotite protolith that in general has not had time to affect the cores of porphyroclasts and large spinels.

**Cape Verde**

Xenoliths from Sal in the Cape Verde archipelago have been described by Ryabchikov et al. (1995), Bonadiman et al. (2005) and Shaw et al. (2006). The xenoliths have protogranular to porphyroclastic and mylonitic textures and comprise refractory harzburgites without primary...
clinopyroxene (≤3 % cpx) as well as mildly depleted spinel lherzolites with primary clinopyroxene (≤18 vol. % cpx).

Many samples show extensive evidence of reaction (e.g. glass-bearing sieve-textured rims on clinopyroxenes and spinels) and rims of ol + sp + cpx + glass on orthopyroxene.

Rybachikov et al. (1995) also reported phlogopite in some samples. Glass is common, both interstitially and as small inclusions in olivine and orthopyroxene. Bonadiman et al. (2005) proposed that the strongly refractory spinel harzburgites and the mildly depleted spinel lherzolites represent two unrelated suites. Both suites include OI1 and OI2 samples (Fig. 2b, d and Fig. 1f, respectively). The presence of two distinct xenolith suites is supported by the relationship between modal proportions of clinopyroxene and olivine (Fig. 7). Within overlapping ranges of olivine the lherzolites (OI-f) define a trend of strongly decreasing clinopyroxene with increasing olivine, whereas the harzburgites (OI-u)
show roughly constant, low clinopyroxene contents. Equilibration temperatures are given in Table 2.

The OI-u harzburgites are strongly refractory with whole-rock major element contents falling close to the opx–ol tie-line (e.g., MgO, CaO, Al2O3; Fig. 3). Somewhat higher TiO2 in a few samples is probably due to minor contamination.

Mineral major element data are available only for one OI1-u harzburgite, which has Cr-rich spinel (cr-number = 0.55), Fo-rich (91.3) olivine and Al2O3-poor orthopyroxene (2.6 wt %; Fig. 4). Cores of orthopyroxene porphyroclasts in OI1-u have low HREE contents and show strongly LREE-depleted to U-shaped REE patterns (YbN = 0.04–0.29, LaN = 0.02–0.35; Fig. 5).

Clinopyroxene in OI1-u shows LREE-depleted to LREE-enriched patterns (YbN = 0.7–4.5, LaN = 0.3–12.2; Fig. 6).

The Cape Verde OI-f lherzolites are relatively fertile with MgO, CaO and Al2O3 (Fig. 8); cr-numbersp is ~0.3, Fool = 87–89 and (Al2O3)opx = 44–60 wt % (Fig. 4). Orthopyroxene porphyroclasts in OI1-u are depleted in HREE and have moderately LREE-depleted patterns (YbN = 0.23–0.36, LaN = 0.03–0.5; Fig. 5), whereas OI1-u clinopyroxene have upward-convex REE patterns (YbN = 0.9–1.5, LaN = 1.7–2.0; Fig. 6).

Kerguelen

Mantle xenoliths from Kerguelen belong to the OI-u group and the suite is dominated by protogranular (OI1-u, Fig. 2a) and poikilitic harzburgites (OI2-u, Fig. 1c). Some OI1-u samples are cut by thin mafic veinlets (of, e.g. pyroxenite, glimmerite, hornblendite), which may show reactions against spinel or orthopyroxene (e.g. Mattielli et al., 1996; Gregoire et al., 1997, 2000a, b; Moine et al., 2000; Delpech, 2004). In OI1-u, clinopyroxene is generally present as small dispersed, interstitial neoblasts (cpx2), sometimes associated with opx and vermicular spinels. In OI2, clinopyroxene also occurs in symplectites, in or along veinlets, and in reaction zones around orthopyroxene and spinel. A few Poikilitic OI2 samples contain phlogopite and amphibole. Trace amounts of apatite, feldspar, rutile, chrome, ilmenite and armalcolite have been observed in glass-bearing domains or veinlets.

All the OI1-u and most of the OI2-u xenoliths fall close to the opx–ol tie-line in most major element diagrams (Fig. 3). OI1-u have orthopyroxenes with low Al2O3 contents (1.8–2.4 wt %), high cr-numbersp (0.42–0.61) and high, uniform Fool of 91.0–91.8 (Fig. 4). Minerals in OI2-u tend towards more fertile compositions. OI1-u xenoliths give lower equilibration temperatures (933–104°C,
average 1014°C; Table 2) than OI2-u (960–1193°C, average 1062°C).

Pyroxene REE patterns indicate minor interaction with REE-enriched melts or fluids (e.g. Mattielli et al., 1996, 1999; Grégoire et al., 1997, 2000a, 2000b; Delpech, 2004). Most OI1-u orthopyroxenes show low concentrations in HREE and MREE/HREE ≪1, but some have irregular REE patterns (e.g. YbN = 0.03–0.23, LaN = 0.011–0.93; Fig. 5). Clinopyroxenes in OI1-u xenoliths show a range in REE patterns from LREE-depleted, through U-shaped, to LREE-enriched, with YbN = 0.13–0.97 and LaN = 0.013–2.1 (Fig. 6).

**Grande Comore**

Mantle xenoliths from Grande Comore comprise spinel lherzolites, dunites and wehrlites; no harzburgites have been reported (Coltorti et al., 1999). The lherzolites have protogranular to porphyroclastic textures, and consist...
of large, deformed porphyroclasts of olivine and orthopyroxene and small, strain-free grains of clinopyroxene and spinel, which occur interstitially or as inclusions in orthopyroxene (Ludden, 1977; Coltorti et al., 1999). The primary textures have been overprinted by pyrometamorphic textures (e.g. melt infiltration), resulting in reactions leading to formation of significant amounts of clinopyroxene, olivine and spinel at the expense of orthopyroxene. Melt–rock interaction processes in the Grande Comore lherzolites are also recorded in grain-to-grain compositional differences within single samples, and zoning (Hauri et al., 1993; Hauri & Hart, 1994). All Grande Comore xenoliths are classified as OI2.

Despite the reaction textures, the Grande Comore OI2 xenoliths have low \((\text{Al}_2\text{O}_3)_{\text{opx}}\) (0.8–3.4 wt %), high cr-number\(_{\text{cr}}\) (0.4–0.7; Fig. 4) and moderate Fo\(_{\text{d}}\) (90–91.3), classifying them as OI2-u. Estimated equilibration temperatures fall in the range 965–1139°C, with a high mean temperature of 1109°C (Table 2).

**Samoa**

Mantle xenoliths from Sava'i (Samoa) are dominated by strongly refractory spinel harzburgites with protogranular to porphyroclastic (most common) textures (Hauri et al., 1993; Hauri & Hart, 1994) and belong to the OI-u group. Symplectites are common in the porphyroclastic rocks. Clinopyroxene occurs mostly along the edges of orthopyroxene + spinel symplectites, but is also found as small interstitial grains rimming orthopyroxene and olivine. In some rocks, melt–rock interaction led to interstitial glass + clinopyroxene ± apatite-bearing patches in various stages of reaction with adjacent porphyroclasts (Hauri et al., 1993). Three harzburgites with porphyroclastic textures (≤0.5% cpx) have been classified as OI1-u.

Whole-rock major element data are available only for OI2-u harzburgites; these fall close to the opx–ol tie-line (not shown). Mineral compositions are typical of OI2-u (1.3–2.6 wt % \((\text{Al}_2\text{O}_3)_{\text{opx}}\); high cr-number\(_{\text{cr}}\) 0.49–0.70; Fo\(_{\text{d}}\) 90.5–91.4; Fig. 4). OI2-u xenoliths tend towards more fertile compositions, particularly less Mg-rich spinel and olivine. OI1- and OI2-u rocks give roughly similar equilibration temperatures (983–1142°C with average 1063°C and 1045–1097°C with average 1072°C, respectively; Table 2).

Samoa OI1-u clinopyroxenes show a wide range in REE contents \((\text{Yb}_N = 0.045–8.6; \text{La}_N = 0.011–2.5); \) Hauri et al., 1993; Hauri & Hart, 1994; Fig. 6) both between and within samples. Hauri & Hart (1994) found a strong correlation between REE\(_{\text{opx}}\) and textures.

**Hawaii**

The Hawaiian mantle xenoliths belong to the mildly depleted OI-f group. They range in rock types from spinel- and spinel–plagioclase-bearing harzburgites and lherzolites, to dunites, wehrlites, websterites and pyroxenites, some of which contain garnet (e.g. Jackson & Wright, 1970; Sen, 1987; Goto & Yokoyama, 1988; Sen & Leeman, 1991; Bizimis et al., 2003). Textures range from porphyroclastic to coarse-grained and fine-grained granular. Clinopyroxene-rich spinel lherzolites are common (up to 24 vol. % cpx; e.g. Jackson & Wright, 1970; Sen, 1987, 1988; Goto & Yokoyama, 1988; Sen & Leeman, 1991; Bizimis et al., 2003). Primary clinopyroxene porphyroclasts have spinel or orthopyroxene exsolution lamellae (e.g. Sen, 1988).

The whole-rock major element compositions of all the Hawaiian harzburgites and lherzolites plot far away from the opx–ol tie-lines in Fig. 8; Hawaiian OI1 xenoliths include some of the lowest MgO (37–43 wt %), and highest CaO (14–3.9 wt %), TiO\(_2\) (≤0.43 wt %), Al\(_2\)O\(_3\) (≤4 wt %) and Na\(_2\)O (≤0.7 wt %) contents reported from ocean islands. OI1-f mineral compositions are also relatively fertile [1.7–4.9 wt % \((\text{Al}_2\text{O}_3)_{\text{opx}}\); cr-number\(_{\text{cr}}\) 0.09–0.55; Fo\(_{\text{d}}\) 88.5–92.0; Fig. 4]. There is a strong tendency for lower Fo\(_{\text{d}}\) in the OI2-f compared with the OI1-f. Equilibration temperatures are similar for OI1-f (914–1205°C with a mean of 1004°C) and OI2-f (881–1161°C, mean 1005°C; Table 2).

A large amount of clinopyroxene REE data from Hawaiian peridotites has been published, but no orthopyroxene OI2-f data. Clinopyroxene in one OI1-f plagioclase–spinel harzburgite and four OI1-f spinel lherzolites from Oahu exhibit smooth REE patterns with strong LREE depletion \((\text{Yb}_N = 3.4–4.7; \text{La}_N = 0.02–1.0); \) Yang et al., 1998; Sen et al., 2003; Fig. 6). Minor metasomatism is indicated by increasing La and Ce from cores to rims in a couple of samples.

In contrast to the OI1-f xenoliths from the Hawaiian islands, mantle xenoliths from the Loihi Seamount comprise dunites and spinel harzburgites, but no lherzolites (Clague, 1988). The harzburgites have cataclastic textures. OI1-u compositional characteristics are reflected in low \((\text{Al}_2\text{O}_3)_{\text{opx}}\) (1.5–2.0 wt %), high cr-number\(_{\text{cr}}\) (0.33–0.75; Fig. 4) and high Fo\(_{\text{d}}\) (90.6–92.3; Fig. 4). Estimated temperatures for the OI1-u are even higher than in Hawaiian OI1-f, 1139–1182°C, with a mean of 1163°C (Table 2).

**Tahiti**

Tahiti mantle xenoliths are porphyroclastic to equigranular, mildly refractory OI-f spinel lherzolites (most abundant) and harzburgites (Tracy, 1980; Qi et al., 1994). Clinopyroxene occurs partly as large porphyroclasts, partly in vermicular intergrowths with spinel and olivine. Some clinopyroxenites have sponge glass-bearing rims, interpreted as being a result of metasomatism (Qi et al., 1994). No other evidence of metasomatism has been given by Tracy (1980) or Qi et al. (1994). Samples with sponge glassy rims have been classified as OI2-f, and the rest of the Tahiti samples as OI1-f.

Like the Hawaii xenoliths, most OI1-f from Tahiti have lower MgO and higher SiO\(_2\), CaO, FeO\(^\text{t}\) and Al\(_2\)O\(_3\) than OI1-u harzburgites, but they have moderate TiO\(_2\) and
Na₂O contents (≤0.12 and ≤0.05 wt. %, respectively; Fig. 3). They also have relatively fertile mineral compositions with ($\text{Al}_2\text{O}_3)_{\text{opx}}$ 2.3–5.6 wt. %, $\text{cr}-\text{number}_{\text{sp}}$ 0.12–0.41 and $\text{Fo}_{\text{ol}}$ 84.5–92.7 (Fig. 4). OI1-f equilibration temperatures are 996–1108°C with a mean of 1058°C (1093°C for the single OI2-f sample; Table 2). Clinopyroxenes in OI1-f have uniform, mildly LREE-enriched REE patterns ($\text{Yb}_N = 2.2–4.0$, $\text{La}_N = 1.6–3.4$; Fig. 6).

DISCUSSION

Different ocean islands show significant contrasts in their OI1 mantle xenolith populations. Xenoliths from the Canary Islands, Madeira, Kerguelen, Samoa and Loihi seamount are strongly refractory (OI1-u and OI2-u), whereas those from Hawaii and Tahiti are mildly refractory (OI1-f and OI2-f). The textural and compositional characteristics of each group are summarized in Table 1. Cape Verde xenoliths appear to comprise both categories. Although no data on OI1 xenoliths are available for the Azores and Grande Comore, the low ($\text{Al}_2\text{O}_3)_{\text{opx}}$, high $\text{cr}-\text{number}_{\text{sp}}$, and high $\text{Fo}_{\text{ol}}$ in many OI2 samples suggest that ultra-refractory OI1-u domains may also be present in the upper mantle beneath these islands. Below we summarize the characteristics of OI-u and OI-f xenoliths, compare the OI-u with MORP, and speculate on the origin and evolution of OI1-u. Furthermore, we provide some estimates of the proportion of the uppermost mantle that might consist of ultra-refractory peridotites and discuss some implications for geodynamics.

Ultra-refractory OI peridotites (OI-u)

As shown in Fig. 3, the whole-rock major element compositions of OI1-u and OI2-u xenoliths (the Canary Islands, Madeira, Cape Verde, Kerguelen, Samoa and Loihi) essentially reflect different modal proportions of olivine and orthopyroxene; clinopyroxene generally makes up <2 vol. % and occurs as secondary neoblasts (Figs 1 and 2). OI1-u have high $\text{cr}-\text{number}_{\text{sp}}$, high $\text{Fo}_{\text{ol}}$, low ($\text{Al}_2\text{O}_3)_{\text{opx}}$, and low HREE concentrations in pyroxenes (Table 1; Figs 4–6). Their ultra-refractory compositions strongly suggest that they are residues after partial melting beyond the stability of primary clinopyroxene. This is in agreement with the degrees of partial melting proposed, for example, for Samoa (33–42% melting; Hauri & Hart, 1994) and the Canary Islands (25–30% melting; Neumann et al., 2004). Furthermore, the OI1-u compositions are similar to residues formed in two-stage melting experiments (Green et al., 2004; Fig. 3) in which a residual, dry, chromite-bearing harzburgite with a composition close to MORP was melted by an additional 10% under $\text{H}_2\text{O}$-$\text{CO}_2$-fluxed conditions at 1.5 GPa and 1320–1350°C.

The secondary origin of clinopyroxene in OI1-u and OI2-u xenoliths from the Canary Islands is reflected in their REE characteristics. Two OI1-u samples with mildly LREE-depleted cpx patterns have opx/cpx REE ratios close to the range of published equilibrium opx/cpx REE ratios in spinel peridotites (Fig. 9), supporting petrographic evidence (Fig. 2) that these clinopyroxenes (cpx2) formed through exsolution from orthopyroxene porphyroclasts (opxl), followed by recrystallization. Many OI2-u, in contrast, have opx/cpx REE ratios that fall significantly below published REE partition coefficients (Fig. 9), implying disequilibrium between the pyroxenes. We interpret these low ratios as the result of a much stronger influence of LREE-enriched fluids or melts on clinopyroxene than on orthopyroxene; some of the most LREE-enriched clinopyroxenes may have formed directly from LREE-rich infiltrating melts.

The less refractory nature of the OI2-u lherzolites does not imply that they represent pristine mantle or mantle domains that have undergone lower degrees of partial melting than the OI1-u harzburgites. Instead, we interpret the OI2-u as being formed by interaction between melts and OI1-u protoliths. Evidence that interaction between melts and refractory mantle is an important mode of formation of lherzolites comes from abyssal peridotites, peridotite massifs, ophiolites and mantle xenoliths (e.g. Hellebrand et al., 2002; Grégoire et al., 2003; Münhener & Piccardo, 2003; Simon et al., 2003; Le Roux et al., 2007; Godard et al., 2008), and from experimental work (Yaxley & Green, 1998; Yaxley, 2000; Bulatov et al., 2002).

Mildly refractory OI peridotites (OI-f)

The OI-f xenoliths (Hawaii, Tahiti, the lherzolite suite from Cape Verde) are significantly more fertile than the
OII-u series [lower whole-rock MgO and higher CaO–FeO–TiO₂–Al₂O₃ contents, higher (Al₂O₃)ᵦ₋₂,opt lower cr-numberᵦ and lower Fe₀₋₄, common presence of primary clinopyroxene; Figs 4, 6 and 8; Table 1]. A primary origin of clinopyroxenes in many xenoliths from Hawaii is supported by their strongly LREE-depleted REE patterns (Yang et al., 1998; Fig. 6). The OII-f peridotites appear to have undergone a lower degree of partial melting than OII-u. Yang et al. (1998) estimated about 2–8% melting of a depleted mantle source for a group of spinel peridotites from Oahu. The OII-f xenolith suites from Hawaii, Tahiti and Cape Verde indicate the presence of mildly refractory domains beneath these islands; the upper mantle beneath Hawaii–Loihi and Cape Verde includes both strongly refractory and more fertile domains.

Comparison between OII-u and MOR peridotites

Peridotites that are dredged and drilled at ‘normal’ segments along mid-ocean ridges and at fracture zones (MORP) make up a large part of the available oceanic peridotite database and are often regarded as representative of all oceanic lithospheric mantle. We therefore explore here how MORP compare with OII-u peridotites. We restricted the MORP samples (recalculated to 100% dry) presented in Fig. 10 to harzburgites and lherzolites (using modal compositions as criteria). However, as modal data are not available for all samples, it is likely that some of the rocks with the lowest SiO₂ contents (<42 wt %) are in fact dunites, even though they are classified as harzburgites in the literature. The samples that Seyler & Bonatti (1997) and Tararotti et al. (2002) described as strongly infiltrated by melts are excluded from Fig. 10.

MORP show wide ranges in whole-rock and mineral major element compositions. Most samples have 35–49 wt % MgO, 40–49 wt % SiO₂, <0–1–4.5 wt % CaO and 0.3–4–3 wt % Al₂O₃ (Fig. 10). (Al₂O₃)ᵦ₋₂,₂,opt falls in the range 1.7–6 wt %, cr-numberᵦ is 0.12–0.57, and in most MORP Fo₀₋₄ is 90–91 (Fig. 4). It is generally accepted that MORP do not represent pristine residues after extraction of mid-ocean ridge basalts, but have been chemically modified by open-system melt migration and crystallization, melt–rock interaction by reactive porous flow, metamorphism, serpentinization and marine weathering (e.g. Elthon, 1992; Hekinian et al., 1993; Cannat et al., 1997; Casey, 1997; Niu, 1997, 2004; Niu & Hekinian, 1997a; Asimow, 1999; Baker & Beckett, 1999; Tararotti et al., 2002). The combination of these processes explains much of the strong compositional heterogeneity observed in MORP. The low CaO/Al₂O₃ ratios exhibited by many MORP (ranging from about 10 to <3–0; Fig. 11) are probably due to leaching of Ca during serpentinization (e.g. Bach et al., 2004; Palandri & Reed, 2004). OII-u xenoliths have CaO/Al₂O₃ ratios close to 1–0, which is typical of harzburgites and lherzolites that have not been subjected to serpentinization (Boyd, 1996).

Because of post-melting processes, the origin of clinopyroxene in MORP has been a subject of debate. Elthon (1992) claimed that MORP experienced partial melting to the exhaustion of clinopyroxene. He proposed that ≲2% clinopyroxene may be explained by granular exsolution from orthopyroxene, whereas proportions >2% most probably formed from infiltrating melt. Other workers (e.g. Dick & Bullen, 1984; Edwards et al., 1996; Seyler & Bonatti, 1997; Tararotti et al., 2002; Seyler et al., 2003; Niu, 2004; Brunelli et al., 2006) have presented extensive petrographic and geochemical evidence that a significant proportion of the clinopyroxene found in MORP is primary, implying melting within the stability field of ol+opx+cpx. However, these researchers have acknowledged that some clinopyroxene has formed through melt infiltration.

On average OII-u harzburgites have higher MgO, lower (Al₂O₃)ᵦ₋₂,₂,opt and higher cr-numberᵦ than MORP and are thus significantly more refractory (Fig. 4). Furthermore, OII-u has no residual clinopyroxene, whereas residual clinopyroxene is present in many MORP. Orthopyroxenes in MORP (Tararotti et al., 2002; Brunelli et al., 2003; Fig. 5) are, on average, less depleted in REE than those in OII-u (e.g. Ybₓ <0–5–0–41 in OII-u). Olivine in MORP has very uniform compositions with average Fₒ contents between 89–6 (Central Indian Ridge) and 90–7 (Mid-Atlantic Ridge; Fig. 12), whereas most cores of olivine porphyroclasts in OII-u have Fₒ contents between 91 and 92 (e.g. Hierro, Lanzarote, Kerguelen, Loihi). OII-u xenolith suites that include a significant proportion of samples with hydrous minerals or reaction textures (OII2-u) show a wider range in Fₒ contents and lower Fₒ averages (e.g. La Palma, Tenerife, Cape Verde) than OII-u suites. The tails towards lower Fₒ can be explained by metasomatic overprinting with some addition of Fe in OII2-u. Clear peaks at Fₒ contents between 92 and 93 in some islands with heterogeneous olivine compositions (e.g. Gran Canaria and Samoa) reflect the ultra-refractory nature of the protolith (OII-u). Olivines in OII-f from Hawaii and Tahiti show wider ranges in Fₒ and lower averages than OII-u.

The proportion of ultra-refractory peridotites among ocean island xenoliths is high. Out of 241 OI peridotites 68% have Fₒ contents ≥90–5; that is, Fₒ at least as high as or higher than average MORP. If the OI-f xenoliths from Hawaii and Tahiti are excluded, 81% of 191 xenoliths have Fₒ ≥90–5 (Fig. 12). These numbers illustrate that a significant proportion of mantle sampled by ocean island volcanism is more refractory than MORP peridotites. In spite of chemical changes induced by post-melting processes in MORP, the observed differences in whole-rock and mineral compositions (Figs 4, 5, 6 and 12) suggest that,
on average, OI1-u peridotites have experienced higher degrees of partial melting than MORP.

However, although MORP, on average, are significantly less refractory than OI1-u, some MORP are as refractory as OI1-u (Fig. 10). This is, for example, true for domains along the East Pacific Rise and Mid-Atlantic Ridge (Niu & Hékinian, 1997a, 1997b). Niu & Hékinian (1997b) proposed that the extent of mantle melting beneath normal mid-ocean ridges increases with increasing spreading rate. This may explain the presence of ultra-refractory peridotites along the East Pacific Rise. However, some of the most refractory MORP known have been collected along one of the slowest-spreading segments of the Mid-Atlantic Ridge (north and south of the Fifteen-Twent’y Fracture Zone; Harvey et al., 2006; Godard et al., 2008). The formation of these ultra-refractory MORP is clearly not related to fast spreading.

The presence of OI1-u peridotites along mid-ocean ridges and the compositional overlap between OI1-u and the most refractory MORP (Figs 4, 5, 10 and 11) suggest that OI1-u may form by mid-ocean ridge processes although this would require potential mantle temperatures of at least 1400°C (Fig. 13). However, available Re–Os data (see Table 3) on some OI1-u (and some ultra-refractory MORP) are not compatible with a simple ridge-melting model.
Alternative models for the formation of OI1-u are discussed below.

The origin of ultra-refractory peridotites

If some (or all) OI1-u did not form by ridge processes, what is the origin of these peridotites and how did they acquire their highly refractory nature? Different sets of data that throw light on these problems are discussed in the following sections.

Old Re–Os isotope model ages in OI peridotites

There is increasing evidence that many peridotite xenoliths collected from ocean islands are older than the adjacent oceanic crust in the area. Unradiogenic Os isotope compositions and old Os model ages are found for some of the most refractory OI1-u harzburgites, and in OI2 type xenoliths from the Kerguelen Islands (<1.6 GPa; Hassler & Shimizu, 1998; Delpech, 2004), a group of lherzolites (OI-f) from Sal, Cape Verde (33.2 ±0.15 Ga; Bonadiman et al., 2005), OI-u and OI2-u xenoliths from the Canary Islands (<650 Ma; Table 2), and in spinel lherzolite xenoliths (OI-f) from various localities in Oahu, Hawaii (90 Ma to 2 Ga; Griselin & Lassiter, 2001; Bizimis et al., 2004, 2007). Evidence of an old depleted mantle component has also been found in Azores lavas (2.5 Ga; Schaefer et al., 2002).

Unradiogenic Os isotope ratios and high and heterogeneous Os model ages (up to ≥2 Ga) have also been reported for MORB from various ridge segments (e.g. Brandon et al., 2000; Alard et al., 2006; Harvey et al., 2006). These data include some of the ultra-refractory peridotites collected along the Mid-Atlantic Ridge (Harvey et al., 2006). According to Harvey et al. (2006), a possible explanation of the lack of evidence of such reservoirs in the Re–Os chemistry of MORB may be that ultra-refractory segments of the upper mantle are resistant to further melting (i.e. sterile).

The old-age peridotites sampled as ocean island xenoliths and along mid-ocean ridges may represent recycled continental or oceanic lithosphere, or sub-arc mantle wedge material. In neither case are they complementary to local MORB, but represent material ‘exotic’ in the local oceanic mantle lithosphere. Alternatively, the ultra-refractory harzburgites could be residues of large melting events episodically occurring over the history of the Earth. This interpretation seems to be supported by a correlation between peaks in Re–Os age distribution of Os-rich alloy grains measured in peridotite massifs and peaks in crustal ages (Pearson et al., 2007). The tectonic setting of these large melting events, however, remains poorly constrained but could be related to large mantle upwellings following slab avalanches into the lower mantle (Condie, 1998). In any case, the degree of melt depletion observed in OI1-u exceeds the degree of melting at normal mid-ocean ridges. Hence, the formation of OI1-u requires either significantly higher potential mantle temperatures or the availability of a fluid flux during melting to lower the solidus.

Brandon et al. (2000) and Harvey et al. (2006) interpreted the old ages as evidence that ancient melt depletion may be preserved over time in the convecting mantle. Old, ultra-refractory peridotites may thus be present in the convecting mantle and may be incorporated into the oceanic lithosphere, either during passive advection along mid-ocean ridges or by freezing to its base as the mantle cools and the lithosphere thickens away from the mid-ocean ridge. Alternatively, fragments of ultra-depleted peridotites may be brought to the base of the abyssal lithosphere from the convecting mantle by rising mantle plumes. However, the presence of old, ‘exotic’ OI1-u (and OI1-f) type material in the oceanic lithospheric mantle does not exclude the possibility that some highly refractory OI1-u type peridotites may form by partial melting along fast-spreading mid-ocean ridges, as proposed by Niu & Hékínian (1997a, 1997b).

Fig. 11. Variations in CaO/Al2O3 ratios of (a) MORP and (b) OI1-u. Data sources as in Figs 3 and 10. MORP show a strong tendency for lower CaO/Al2O3 ratios than OI1-u, probably as a result of serpentinization.
A plume origin?

It has been suggested previously that refractory peridotites with old Re–Os ages are transported from the deeper mantle to lithospheric depths by rising plumes (Bizimis et al., 2007). Such a mechanism supports the model by Parman (2007) that the unradiogenic He isotope signature of many ocean island lavas originates from an old, depleted peridotite component that is stored somewhere in the deep mantle, probably close to the core mantle boundary. Because P–T estimates can give important constraints on the evolution of OI-u we investigate these and other models by looking at the temperatures and pressures at which the ocean island xenoliths last equilibrated.

Available data imply that OI-u xenoliths last equilibrated at relatively low pressures and temperatures. Minimum pressures obtained from geothermometry combined with densities of CO₂ inclusions in mantle xenoliths from the Canary Islands, Samoa, Tahiti, Kerguelen and Grand Comore fall in the range 0.7 to GPa (Hansteen et al., 1991, 1998; Schiano et al., 1992, 1994, 1998;
for OI-u (Table 2). Grey fields show the ranges of pressure and temperature estimates taken as 550°C geotherm at a mid-oceanic ridge. The temperature at the Moho is significantly higher than the pressures for OI1-u. Also shown is the simplified pressure-temperature diagram showing the stability fields for OI1-u peridotite. The calculated garnet-in boundary is shifted to further than estimated by the simple extrapolation used here. The OI1-u have lost all clinopyroxene, which will increase solidus temperatures and probably gives minimum temperatures for the onset of melting as OI1-u have experienced in total 30% melt extraction from a pyrolite source. The resulting OI1-u solidus probably gives minimum temperatures for the onset of melting as OI1-u have lost all cpx, which will increase solidus temperatures further than estimated by the simple extrapolation used here. The diagram shows that the calculated garnet-in boundary is shifted to significantly higher pressures for OI1-u. Also shown is the simplified geotherm at a mid-oceanic ridge. The temperature at the Moho is taken as 350°C and \( T_{\text{atm}} \) at the base of the lithosphere is 1330°C. Grey fields show the ranges of pressure and temperature estimates for OI1-u (Table 2).

Fig. 13. Pressure-temperature diagram showing the stability fields of plagioclase- spinel- and garnet-peridotites for different bulk compositions (Hawaiian pyrolite, average MORB, and average OI1-u; Table 4). Spinel stability extends into the garnet and plagioclase fields. Phase boundaries are computed with Perple_X including the recent addition of Cr-bearing end-members to the Holland and Powell database (Holland & Powell, 1998; Connolly & Petreini, 2002; Klemme et al., in preparation). Ultra-refractory OI1-u peridotite is assumed to contain 0.05 wt % Na2O (Table 4). The solids are taken from Green & Ringwood (1997) for wet and dry pyrolite and extrapolated to more refractory compositions using the approximation of Robinson & Wood (1998; increase of 7°C per degree of melt extraction) and assuming that OI1-u have experienced in total 30% melt extraction from a pyrolite source. The resulting OI1-u solidus probably gives minimum temperatures for the onset of melting as OI1-u have lost all cpx, which will increase solidus temperatures further than estimated by the simple extrapolation used here. The diagram shows that the calculated garnet-in boundary is shifted to significantly higher pressures for OI1-u. Also shown is the simplified geotherm at a mid-oceanic ridge. The temperature at the Moho is taken as 350°C and \( T_{\text{atm}} \) at the base of the lithosphere is 1330°C. Grey fields show the ranges of pressure and temperature estimates for OI1-u (Table 2).

Table 3: Bulk-rock Re–Os isotope data for Canary Island harzburgite and lherzolite xenoliths (E.-R. Neumann & E. Widom, unpublished data)

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Os (ppb)</th>
<th>Re (ppb)</th>
<th>(^{187}\text{Os}/^{188}\text{Os} (\pm 2\sigma))</th>
<th>(^{187}\text{Re}/^{188}\text{Os})</th>
<th>(T_{\text{ma}}) (Ma)</th>
<th>(T_{\text{rd}}) (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LAB-10(^{a})</td>
<td>OI1</td>
<td>2.363</td>
<td>0.067</td>
<td>0.12409(19)</td>
<td>0.136</td>
<td>654.52</td>
<td>434.57</td>
</tr>
<tr>
<td>LA1-7</td>
<td>OI1</td>
<td>1.251</td>
<td>0.003</td>
<td>0.12613(20)</td>
<td>0.010</td>
<td>134.60</td>
<td>131.30</td>
</tr>
<tr>
<td>LA1-13</td>
<td>OI1</td>
<td>7.576</td>
<td>0.003</td>
<td>0.12423(11)</td>
<td>0.002</td>
<td>415.75</td>
<td>413.81</td>
</tr>
<tr>
<td>PAT2-68</td>
<td>OI1</td>
<td>1.271</td>
<td>0.036</td>
<td>0.12554(13)</td>
<td>0.134</td>
<td>329.10</td>
<td>219.17</td>
</tr>
<tr>
<td>H1-5(^{a})</td>
<td>OI2</td>
<td>3.631</td>
<td>0.002</td>
<td>0.12652(18)</td>
<td>0.002</td>
<td>73.58</td>
<td>73.14</td>
</tr>
<tr>
<td>PAT2-75</td>
<td>OI2</td>
<td>1.750</td>
<td>0.030</td>
<td>0.12406(16)</td>
<td>0.082</td>
<td>402.09</td>
<td>320.28</td>
</tr>
<tr>
<td>TF14-38</td>
<td>OI2</td>
<td>2.189</td>
<td>0.011</td>
<td>0.12303(27)</td>
<td>0.025</td>
<td>630.61</td>
<td>591.55</td>
</tr>
<tr>
<td>TF14-47</td>
<td>OI2</td>
<td>1.206</td>
<td>0.012</td>
<td>0.12678(12)</td>
<td>0.049</td>
<td>39.13</td>
<td>34.34</td>
</tr>
<tr>
<td>TF14-48</td>
<td>OI2</td>
<td>2.273</td>
<td>0.021</td>
<td>0.12486(32)</td>
<td>0.045</td>
<td>360.77</td>
<td>320.87</td>
</tr>
</tbody>
</table>

\(^{a}\)Measured July 1995. All others June 1993 at Department of Terrestrial Magnetism, Carnegie Institution of Washington. \(T_{\text{ma}}\), Re–Os mantle model age; \(T_{\text{rd}}\), Re-depletion age, assuming all Re was extracted during melting.
(Bizimis et al., 2007) and is supported by some geophysical data (Li et al., 2004) and certain numerical models (e.g. Moore et al., 1999).

A plume origin, however, is hard to reconcile with the low equilibration temperatures estimated for the OI1-u xenoliths, 860–1200°C (Table 1). These temperatures would require plume material to stagnate at shallow depths for sufficiently long times to allow thermal and chemical re-equilibration and phase changes during cooling to the low (lithospheric) temperatures recorded in the xenoliths. Moore et al. (1999) have shown numerically that significant lithosphere erosion happens only if the plume is at least 140°C hotter than the background temperature. Such very high temperatures should leave a trace in the peridotites (e.g. resetting of the Hf isotopic clock), which is not observed (Bizimis et al., 2007). Furthermore, if the OI1-u were part of the rising plume and transported to pressures below 1.3 GPa, they should not only be hot, but also show more sign of melt infiltration and deformation. Even though the OI1-u harzburgites probably would not melt because of their high solids temperatures, more fertile plume components would definitely be partially molten at 1.3 GPa. We would therefore expect plume melts to have invaded stretched and folded OI1-u fragments (Erneta et al., 2002). Because of their strong depletion, OI1-u are highly susceptible to reaction with or contamination by enriched melts and would react quickly. We do not observe this. Furthermore, OI2-u xenoliths, which do show melt-peridotite reactions, tend to give higher average equilibration temperatures than OI1-u xenoliths from the same island (Table 2). This suggests local heating of cold, lithospheric peridotites by plume melts rather than regional cooling of accreted plume material.

Another argument against transport of OI1-u peridotites from the deep mantle is their low densities (Simon & Podladchikov, 2008). Because the ultra-refractory OI1-u peridotites are strongly buoyant such material is not expected to be stored for long periods of time deep in the lower mantle, but rather to float to the top of the convecting mantle. OI1-u material is therefore unlikely to be present in large quantities in the source region of deep thermally activated processes. However, some plumes may themselves consist of low-density residual mantle that has been subducted to some depth and has become buoyant relative to the surrounding more fertile peridotite when heated to ambient temperatures (Oxburgh & Patenier, 1977; Pressnall & Helsley, 1982; Lee & Chen, 2007).

Evidence against a plume origin for OI1-u xenoliths is also provided by Sr isotopic data for xenoliths from the Canary Islands. In general, Sr, Nd, Pb and Hf isotopic data for OI mantle xenoliths indicate isotopic signatures similar to those of the local ocean island basalts (e.g. Vance et al., 1989; Whitehouse & Neumann, 1995; Mattielli et al., 1996, 1999; Sen et al., 2003; Neumann et al., 2004).

Sr–Nd–Pb isotopic data on whole-rock samples and clinopyroxene separates from mantle xenoliths from the Canary Islands follow this general pattern (Whitehouse & Neumann, 1995; Neumann et al., 2004; Fig. 14). However, in situ laser ablation analyses of $^{87}$Sr/$^{86}$Sr in clinopyroxenes show a range in isotope ratios (Neumann et al., 2004). Clinopyroxenes in two OI1-u xenoliths have $^{87}$Sr/$^{86}$Sr isotope ratios of 0.70266 and 0.70276, which are significantly below those of Canary Islands basalts and require long-term evolution in a depleted, low Rb/Sr environment. The ultra-refractory nature of the Canarian OI1-u peridotites must thus be an old feature caused by processes unrelated to the present plume. Only clinopyroxenes in OI2-u xenoliths gave ratios of 0.7031–0.7033, which, like those of the whole-rock and clinopyroxene separates, is within the range of Canary Islands basalts field. One OI2-u sample gave an intermediate value of 0.70288. The original old depleted isotopic signature has therefore recently been overprinted by metasomatism by melts or fluids originating in the Canary Islands plume (Neumann et al., 2004).

In conclusion, we consider a recent plume origin for the ocean island xenoliths highly unlikely. Based on the $P$–$T$ estimates, the inferred physical properties (high viscosity and low density) and the in situ Sr isotopic data, we favor the idea that the OI peridotites were part of the oceanic background.
lithosphere before the onset of the plume processes that brought them to the surface.

**Did the OI1 peridotites acquire their highly depleted compositions by plume processes?**

Although we concluded above that the OI-u peridotites are not brought up by the associated plume, the lithospheric mantle beneath ocean islands is clearly affected by plume-related processes. The most obvious one is metasomatism by plume-derived melts and fluids, which is reflected in OI2 xenoliths, and which has also caused some enrichment in incompatible elements and radiogenic isotopes in OI1 xenoliths (e.g. Figs 1–6 and 14). However, plume-related processes have also been proposed to be a cause of more refractory mantle compositions by the following processes: (1) reactions between plume melts or fluids and peridotite wall-rock may extract lithophile components from the wall-rock minerals and cause orthopyroxene to form at the expense of olivine and clinopyroxene (e.g. Kelemen et al., 1992); (2) the plume may lead to thermal erosion of the base of the oceanic lithosphere and trigger partial melting in its remaining parts (e.g. Moore et al., 1999; Li et al., 2004; Bizimis et al., 2007).

Microtextural evidence of melt–wall-rock reactions is common in OI2 peridotites. They show the reaction (opx + melt → ol + cpx). This is opposite to the reaction described by Kelemen et al. (1992), but is compatible with infiltration by transitional to strongly alkaline ocean island melts saturated in clinopyroxene. Infiltration of ocean island peridotites by alkaline melts typically leads from harzburgite (OI1-u) to lherzolite (OI2-u) to dunite–wehrlite. In addition, the Ti–Al–Fe-rich nature of the clinopyroxenes in some OI dunites, wehrlite and clinopyroxenites suggests crystallization directly from alkali basaltic magmas (e.g. Neumann et al., 2004). Such reactions are by definition absent from OI1 xenoliths. Additionally, interaction between peridotites and melts is expected to reset the Sr isotope ratio in the peridotites to those of the melts. As reported above, this is seen in all whole-rock samples and clinopyroxene separates of OI-u xenoliths from the Canary Islands, as well as in in situ analyses of clinopyroxenes in OI2-u rocks (Neumann et al., 2004). However, the low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios obtained from in situ analyses of clinopyroxenes in Canarian OI1-u xenoliths (Fig. 14) are robust evidence against interaction with plume melts.

The Sr isotope ratios in Canarian OI1-u rocks do not exclude the possibility that heating by a thermal plume caused partial melting of the lithospheric mantle. However, the low equilibration temperatures obtained for the OI1-u (and OI2-u) xenoliths make this possibility highly unlikely. In Fig. 13, the estimated $P$–$T$ conditions for the ultra-refractory xenoliths are compared with estimated soli for peridotites of average OI1-u and more fertile compositions. The equilibration temperatures obtained for the OI-u xenoliths are far below their estimated solidus temperatures. Heating and melting related to the present plumes therefore seems an implausible explanation for the formation of the OI1-u harzburgites. As neither plume-induced melting nor reaction with plume melts appears to be responsible for the ultra-refractory compositions of the OI1-u peridotites, we conclude that this composition was acquired a significant period of time before the onset of plume processes. This is in agreement with the conclusions that the old Os model ages obtained for some ultra-refractory OI peridotites and MORP reflect ancient melting events (e.g. Hassler & Shimizu, 1998; Brandon et al., 2000; Griselin & Lassiter, 2001; Bizimis et al., 2004, 2007; Delpech, 2004; Bonadiman et al., 2005; Alard et al., 2006; Harvey et al., 2006).

**Comparison of OI1-u with refractory peridotites from other settings**

Although some OI-u peridotites may have formed along mid-ocean ridges, we are left with the hypothesis that many OI1-u peridotites represent old, refractory material that has been passively accreted to the oceanic lithosphere along mid-ocean ridges or during cooling and thickening of the lithosphere as it ages. By which kind of process did OI1-u peridotites acquire their ultra-refractory composition if they were not formed in their host plume? We compare OI1-u compositions with those of sub-continent mantle xenoliths and peridotites from oceanic sub-arc mantle (Figs 15 and 16) to determine if there are similarities that might suggest a related origin. Peridotite suites from both types of setting include highly refractory samples, whose large degree of melt extraction has been ascribed to either higher mantle temperatures or flux by hydrous fluids, or both (see, e.g. review by Pearson et al., 2003).

The most refractory sub-continental mantle xenoliths are cratonic peridotites from South Africa (Boyd & Mertzman, 1987; Cox et al., 1987; Nixon et al., 1987; Winterburn et al., 1990; Boyd et al., 1999; Simon et al., 2003, 2007) and East Greenland (Bernstein et al., 1998). However, these xenoliths differ significantly from the OI1-u xenoliths by lower FeO, and higher SO$_2$ and MgO (Fig. 15). Figure 15 also shows the comparison of OI1-u xenoliths with some refractory off-craton continental mantle xenoliths, which are characterized by large scatter in major elements. Some off-craton xenoliths fall within the OI1-u field, but the majority are more fertile than OI1-u. We did not, however, filter the continental peridotite data for metasomatized samples, which might explain a large part of the scatter. We cannot therefore exclude the possibility that OI1-u represent fragments of continental lithospheric mantle that was eroded from the base of the continental lithosphere or sheared off during rifting and
passive margin formation. The latter scenario has been proposed as the ‘raft’ model by Muñtener & Manatschal (2006) to explain the occurrence of some strongly refractory spinel peridotites in the apparently oceanic domain at the Newfoundland margin. Such a model, however, does not explain how the presumed continental lithospheric mantle reached its high degree of depletion. Muñtener & Manatschal (2006) proposed that the original melt depletion event took place in an arc setting. Extensive partial melting in an arc setting has also been proposed as a mechanism to form Archean cratonic lithosphere (e.g. Kesson & Ringwood, 1989; De Wit et al., 1992; Westerlund et al., 2006; Simon et al., 2007).

In Fig. 16 we compare OI1-u with some mantle peridotite series from modern oceanic sub-arc settings. As a group, the arc samples also show considerable

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**Fig. 15.** Comparison between OI1-u xenoliths (grey fields) and continental mantle xenolith suites with a high proportion of harzburgites. Data from The Thumb, New Mexico (Ehrenberg, 1982), Patagonia, South America (Skews & Stern, 1979; Laurora et al., 2001), NE Brazil (Rivalenti et al., 2000), Sardinia (Dupuy et al., 1987; Beccaluva et al., 2001), Namibia (Boyd et al., 2004), East Greenland (Bernstein & Brooks, 1996) and South Africa (Boyd & Mertzman, 1987; Cox et al., 1987; Nixon et al., 1987; Winterburn et al., 1990; Boyd et al., 1999; Simon et al., 2003, 2007). The off-craton mantle xenolith series show wide compositional ranges, including fertile compositions.
compositional variations, but the peridotite suites from the Izu–Bonin–Mariana forearc, the Torishima Forearc Seamount (data from Ishii et al., 1992; Parkinson & Pearce, 1998; Zanetti et al., 2006) and Lihir (Papua New Guinea; data from McInnes et al., 2001) are strikingly similar to OII-u. This similarity suggests that OII-u type peridotites may form by similar processes. A more detailed comparison between OII-u peridotites and sub-arc mantle will be given elsewhere (Neumann & Simon, in press).

In conclusion, the highly refractory OII-u xenoliths most probably represent residues of shallow melting processes at elevated temperatures or water contents, or both. Following melting these residues were recycled within the upper part of the convecting mantle and became accidentally trapped in the oceanic lithosphere before they were sampled by the ocean island basalts. However, some may have formed along hot mid-ocean ridges and been directly incorporated in the oceanic mantle lithosphere.
CONCLUSIONS AND IMPLICATIONS FOR GEODYNAMICS

The major element compositions of oceanic island peridotites have been reviewed. After petrographically filtering for the effects of mantle metasomatism and melt–rock interaction during transport, a large proportion of the peridotite xenoliths sampled in ocean islands proved to be ultra-refractory harzburgites. These samples are indicative of partial melting beyond the cpx-out reaction and are on average significantly more refractory [higher $Fo_{94}$ and $cr$-number$_{sp}$, lower CaO and $(Al_2O_3)_{opx}$] than abyssal peridotites sampled by dredging and drilling along mid-ocean ridges and at oceanic fracture zones. We have shown that the strong melt depletion in the OI1-u harzburgites is not related to melt–rock interaction, and that the OI1-u are not genetically related to their associated ocean island plume. These peridotites must therefore be residues of large-degree melting events at temperatures significantly higher than present-day normal mantle potential temperatures, or they are residues formed during extensive fluid-fluxed melting, possibly in an arc setting. The last possibility is supported by their compositional similarity to modern oceanic forearc peridotites. If the melt depletion of the OI1-u is caused by fluid-fluxed melting any textures associated with fluid–rock interaction, which are commonly observed in recent arc-related peridotites, must have been erased by complete textural and chemical re-equilibration.

We propose that ultra-refractory OI1-type harzburgites represent the end-product of mantle differentiation, in the sense that they have reached a maximum degree of melt extraction. Although we cannot provide an estimate on how much of the mantle consists of strongly refractory harzburgite, sterile OI1-u material makes up more than two-thirds of all harzburgites and lherzolites sampled by ocean island magmas (Fig. 12) and is thus common in the mantle beneath ocean islands. This is in accordance with a model of a strongly heterogeneous mantle, as supported by recent geochemical and geophysical observations and theoretical models (e.g. Phipps Morgan & Morgan, 1999; Brandon et al., 2000; Helffrich & Wood, 2001; Meibom et al., 2002; Anderson, 2006; Harvey et al., 2006; Helffrich, 2006; Armienti & Gasperini, 2007; Sobolev et al., 2007). The presence of such material has consequences for geodynamics because of its distinct geochemical and physical properties.

Strong depletion in fusible components (and increase in forsterite content), as in OI1-u peridotites, leads to high solidus temperatures and decreased water contents, which cause higher viscosities and greater resistance to deformation (Karato, 1986; Hirth & Kohlstedt, 1996). Manga (1996) has shown numerically that highly viscous material (10–100 times more viscous that the surrounding mantle, such as strongly melt-depleted peridotite; Karato, 1986; Hirth & Kohlstedt, 1996) can persist relatively undisturbed for many mantle overturns in the convecting mantle and has a tendency to aggregate to form larger heterogeneities from smaller blobs. Because of their high solidus temperatures such rocks will be ‘sterile’ and are not expected to participate in further partial melting. These rocks are therefore not expected to lend their geochemical signatures to partial melts formed either along mid-ocean ridges or in ocean islands, except if temperatures are extraordinarily high (Fig. 13).

Because of their high Mg/Fe ratios and low Al contents, the OI1-u harzburgites have relatively low densities and therefore strong buoyancy, and will tend to rise above less refractory, denser mantle material. The accumulation of refractory mantle at the top of the convecting mantle over time is confirmed by 3-D mantle convection simulations (Tackley & Xie, 2002). Highly refractory material may therefore be abundant in regions where new plates form (i.e. mid-ocean ridges) and may preferentially be incorporated into the cooling and growing oceanic lithosphere. We thus expect the proportion of highly refractory harzburgite to be significantly higher in the oceanic mantle lithosphere and underlying asthenosphere than in the deeper parts of the convecting mantle, in agreement with the model of Phipps Morgan & Morgan (1999). An important consequence of a high proportion of low-density OI1-u material in the oceanic lithosphere is that it will be difficult to subduct, particularly in the initial stages of subduction. Some preliminary calculations with PerpleX [http://www.perplex.ethz.ch/perplex.html; see Connolly & Petrini (2002) for methods] show that a mantle column (100 km thick) with the average composition of the Canary Island OI1-u xenoliths overlain by a 6.5 km thick basaltic crust ($\rho_{vucu} = 2950 \text{ kg/m}^3$) will have a density of 3200 kg/m$^3$ close to the ridge, compared with 3220 kg/m$^3$ for the average composition of oceanic peridotite (Table 4). Thickening of the lithosphere by simple half-space cooling leads to a 100 km thick, old, thermally equilibrated lithosphere with average column densities of 3245 kg/m$^3$ and 3263 kg/m$^3$ for Canarian OI1-u and average MORP, respectively. Afonso et al. (2007) developed a more sophisticated model for the calculation of oceanic lithosphere density structure and buoyancy over time. Their model results in a 106 km thick mature lithosphere, where the upper 60–80 km consist of variably melt-depleted material and the lower 20–40 km are made up of fertile mantle, depending on the melt model used. Afonso et al. (2007) concluded that such a lithosphere develops smaller negative buoyancies ($\sim 40 \text{ kg/m}^3$) than commonly assumed, and thus that the role of intrinsic negative buoyancy may not be as critical for the onset of subduction. If the oceanic lithosphere contains a significant amount of ultra-refractory peridotite,
as proposed here, buoyancy will be further enhanced, making initiation of subduction even more difficult. This is, of course, assuming that the asthenosphere on average consists of more fertile, denser mantle, which may not be realistic given the presumably large amount of buoyant recycled residues in the convecting mantle. If the asthenosphere is rich in ultra-refractory material the cooling oceanic lithosphere may become negatively buoyant because its upper to middle part is made up of denser (relative to OI1-u) MORP. Lenses of material with different density in the lithosphere will also enhance the formation of gravitational instabilities and small-scale convection. Better constraints on models for the composition and stratification of the oceanic lithosphere may be derived from comparison of calculated geophysical properties (e.g. gravity, geoid anomalies, topography, seismic velocities) with measured geophysical data (e.g. Afonso et al., 2007).

ACKNOWLEDGEMENTS
Dante Canil generously gave us a copy of his database on mantle rocks, which was the foundation of our own database. The work has profited from discussions with a number of colleagues, in particular Y. Podladchikov, L. Rupke, J. Phipps Morgan, O. Alard, J. C. Afonso and E. Hellebrand. H. Downes, G. Pearson and two anonymous reviewers are thanked for their comments that

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**References:**
SUPPLEMENTARY DATA

Supplementary data for this paper are available at Journal of Petrology online.

REFERENCES


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