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Formation and Modification of the Shallow Sub-continental Lithospheric Mantle: a Review of Geochemical Evidence from Ultramafic Xenolith Suites and Tectonically Emplaced Ultramafic Massifs of Western and Central Europe

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The petrology and geochemistry of shallow continental lithospheric mantle (SCLM) can be studied via (1) tectonically emplaced ultramafic massifs and (2) mantle xenoliths entrained in alkaline magmas. Data from these two separate sources are used to identify processes that have formed and modified the SCLM. In western and central Europe where the continental crust consolidated in Phanerozoic times, both sources of information are available for study. Rock types found in ultramafic massifs in Europe are generally similar to those found in ultramafic xenolith suites. The most frequent lithology is anhydrous spinel lherzolite, grading towards harzburgite. Massifs reveal pyroxenite layering, harzburgite bands and cross-cutting mafic and ultramafic dykes. The Phanerozoic European SCLM xenoliths and massifs show broad mineralogical and chemical similarities to Phanerozoic continental spinel peridotites world-wide. The main process that controls the geochemistry of the SCLM is depletion by removal of basaltic melt. Differences from this norm reflect significantly different processes in the SCLM, such as interaction with melts and fluids. Such processes probably gave rise to hornblendite veins and pyroxenite layers, although the latter have also been interpreted as recycled oceanic crust. Rare earth element data for whole-rock peridotites and their constituent clinopyroxenes show a variety of patterns, including light rare earth element (LREE) depletion as a result of removal of basaltic melt, LREE enrichment caused by metasomatism, and U-shaped REE patterns that are probably due to interaction with carbonatite melts.

Extended mantle-normalized incompatible trace element patterns for whole rocks show enrichment in Rb and Ba in peridotites considered to have been subduction-metasomatized, whereas those considered to be carbonate-metasomatized have strong negative anomalies in Zr, Nb and Hf. Mantle amphiboles are strongly enriched in LREE when found in veins, but can be LREE depleted if they are interstitial. Radiogenic isotope ratios for xenoliths and massifs largely overlap, although the xenoliths show a significant clustering around a 'plume-component' identical to the Neogene alkaline magmatism of Europe. This component is lacking in the massifs, most of which were emplaced into the crust before the onset of Neogene plume activity. Infiltration of carbonatite melts is observed petrographically in some xenoliths and evidenced by low Ti/Eu ratios in bulk rocks, but is very rare. The effect of passage of hydrous fluids from subducting slabs is also seen in some suites and massifs, being exhibited mainly as unusual Sr and Pb isotope ratios, although enrichment in K, Rb and Ba, and the presence of modal phlogopite, may also point to subduction-metasomatism.

KEY WORDS: peridotites; xenoliths; orogenic massifs; Europe

INTRODUCTION

The shallow continental lithospheric mantle (SCLM) can be studied via (1) tectonically emplaced ultramafic massifs

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and (2) ultramafic xenoliths entrained in mafic alkaline magmas. Massifs provide field relations between different mantle lithologies and an estimate of the proportions of rock-types, whereas xenoliths provide samples away from orogenic belts and in a single vertical section beneath a single locality. Both sources are considered to provide samples of typical SCLM (Wilshire & Pike, 1975; Griffin *et al.*, 1998). As such, they can yield information about geochemical processes that occur in the shallow lithospheric mantle. Such processes could include removal of melt, addition of melt, and/or interaction with fluids derived from subduction or asthenospheric upwelling.

The lithosphere of western and central Europe was consolidated in Phanerozoic times, although evidence for crustal events as old as 2 Ga can be found within the region. Major Phanerozoic events that have affected western Europe include subduction and collision that occurred during both the Hercynian and Alpine orogenies. Upwelling of a plume beneath the European plate during Tertiary times (Granet *et al.*, 1995; Hoernle *et al.*, 1995; Wilson & Patterson, 2001) gave rise to widespread mafic alkaline magmatism. Phanerozoic Europe contains numerous mantle xenolith localities and ultramafic massifs, the latter being emplaced into the crust during the Hercynian and Alpine orogenies. This paper summarizes the main results of geochemical studies of the SCLM as represented by xenoliths and massifs within Europe. It builds on the work of Menzies & Bodinier (1993) but differs in that xenolith suites from cratonic northern Europe are omitted and ultramafic massifs are included. Xenoliths from pre-Tertiary alkali basalts have also been omitted, but scope remains for a comparison of pre-Tertiary and Neogene mantle xenoliths in Europe.

LOCALITIES AND LITHOLOGICAL VARIATIONS OF EUROPEAN SCLM

Localities of ultramafic xenoliths are abundant in the Neogene mafic alkaline volcanics in western and central Europe (Fig. 1). The best examples are documented mainly from France and Germany. Mantle xenoliths have also been studied in Poland, Hungary, Austria and Romania. Italian and Sardinian xenolith suites are also known but data from Spain and Bulgaria are sparse. The localities in Spitsbergen (Ionov *et al.*, 1993) have been omitted from this review. Ultramafic massifs occur in all of Europe's orogenic belts, i.e. the Pyrenees, Alps, Betics and Apennines. They were emplaced during the Alpine or Hercynian orogenies and hence represent material that was detached from the *in situ* SCLM at an earlier time than the mantle material that occurs as xenoliths in the Neogene alkaline magmas. The following localities have been omitted from this review: ultramafic massifs of the Northern Apennines (External Ligurides) and

Central Alps (Val Malenco), which have been shown to have undergone tectonic denudation during passive extension of the continental lithosphere (Rampone *et al.*, 1995; Müntener & Hermann, 1996; Hermann *et al.*, 1997), and the Internal Liguride massifs, which are considered to represent sub-oceanic mantle (Rampone *et al.*, 1996). The Alpe Arami garnet peridotite massif in the western Alps is also not discussed here, as few geochemical data are yet available (Ottonello *et al.*, 1984).

Ultramafic massifs in Europe contain a wide variety of lithologies and adjacent massifs can show entirely different mineralogical features (Fabriès *et al.*, 1991, 1998). Those in the Pyrenees are mainly formed of spinel lherzolite and harzburgite, and often display layering, with both well-defined and diffuse spinel and garnet pyroxenite layers a few millimetres to 1 m in thickness. Harzburgite bands up to 20 m thick are also present. Veins, sills or dykes of hornblendite, amphibole pyroxenite and garnet-rich amphibole pyroxenite cross-cut some massifs. These are generally a few centimetres to 30 cm thick. The presence of amphibole or phlogopite in the peridotites is generally related to the proximity of veins, i.e. they occur in metasomatic aureoles around veins. One eastern Pyrenean massif (Caussou) is composed of amphibole-bearing lherzolites and clinopyroxene-rich peridotites without veins (Fabriès *et al.*, 1989).

Among the Alpine massifs, Baldissero is almost entirely composed of spinel lherzolite, with very rare pyroxenite (Hartmann & Wedepohl, 1993). The Balmuccia massif is composed of clinopyroxene-poor spinel lherzolite, with rare dunites and harzburgites (Shervais & Mukasa, 1991). Pyroxenites form ∼5% of the outcrop and have very variable mineralogies, including clinopyroxenites, orthopyroxenites and websterites. Dykes cutting the massif include spinel pyroxenites, gabbronorites and hornblendites. In contrast, the Finero massif is dominantly harzburgitic and contains abundant pervasive phlogopite and amphibole. The Lanzo massif consists of three bodies composed of plagioclase lherzolite separated by shear zones (Bodinier *et al.*, 1991). It contains a few pyroxenite layers, refractory peridotites and gabbro veins. Recrystallization of the massif in the plagioclase-peridotite facies indicates that it has been derived from very shallow mantle, perhaps transitional between sub-oceanic and sub-continental (Bodinier *et al.*, 1991).

The Ronda massif in SE Spain is mainly lherzolite with ∼5% garnet pyroxenite, spinel pyroxenite or olivine gabbro layers (Obata *et al.*, 1980). In addition to spinel, some lherzolites contain either plagioclase or garnet. This could be interpreted to indicate that the Ronda massif contains material from the shallowest (plagioclase lherzolite), intermediate (spinel lherzolite) and deepest parts (garnet lherzolite) of the SCLM. However, Schubert (1982) suggested that the garnet peridotites result from
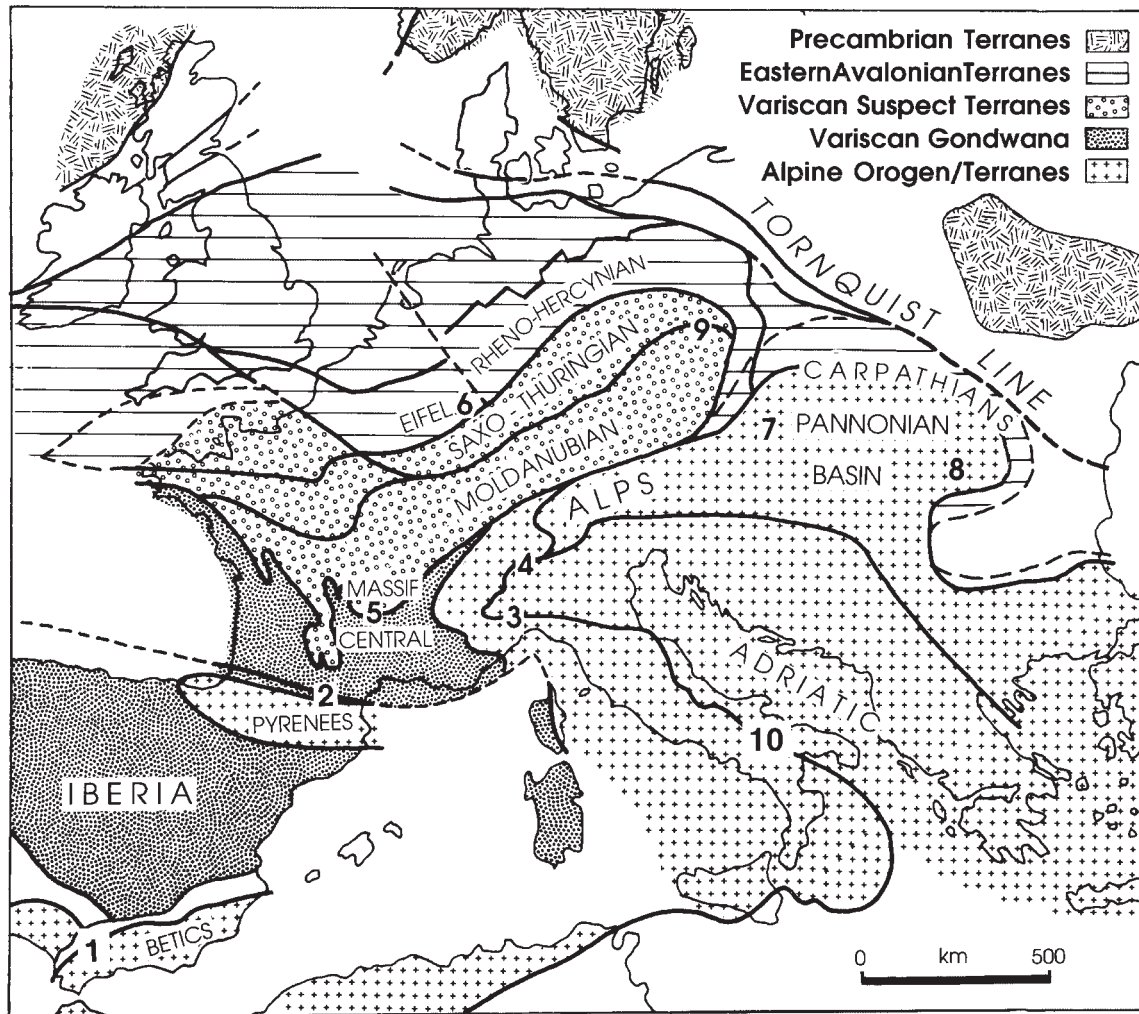
Fig. 1. Sketch map of geology of Europe, showing main localities of mantle xenoliths and ultramafic massifs. Massifs: 1, Ronda; 2, Pyrenees (France); 3, External Ligurides (Italy); 4, western Alpine massifs ('Ivrea zone'), Italy. Xenolith localities: 5, Massif Central (France); 6, Eifel (Germany); 7, Pannonian Basin (Hungary–Austria); 8, Romania; 9, Poland; 10, Mte Vulture (Italy). Individual massifs have been the subject of the following studies: Suen & Frey (1978), Obata *et al.* (1980), Zindler *et al.* (1983), Frey *et al.* (1985), Reisberg & Zindler (1986), Bodinier *et al.* (1987a, 1987b, 1988, 1990, 1991), Voshage *et al.* (1987), Bodinier (1988), Fabriès *et al.* (1989, 1991, 1998), Reisberg *et al.* (1989), Downes *et al.* (1991), Mukasa *et al.* (1991), Shervais & Mukasa (1991), Hartmann & Wedepohl (1993), Obermiller (1994), Rivalenti *et al.* (1995), McPherson *et al.* (1996), Woodland *et al.* (1996), Zanetti *et al.* (1996, 1999) and Mukasa & Shervais (1999). Xenolith localities include those studied by Kurat *et al.* (1980, 1991), Stosch & Seck (1980), Stosch *et al.* (1980), Stosch (1982), Mengel *et al.* (1984), Dietrich & Poultidis (1985), Cabanes & Briqueu (1986), Stosch & Lugmair (1986), Downes & Dupuy (1987), Fabriès *et al.* (1987), Morten (1987), Nicolas *et al.* (1987), Witt & Seck (1987, 1989), Kempton *et al.* (1988), Beccaluva *et al.* (1989), Capedri *et al.* (1989), Edgar *et al.* (1989), Embey-Isztin *et al.* (1989), Conticelli & Peccerillo (1990), Hartman & Wedepohl (1990), Lloyd *et al.* (1991), Downes *et al.* (1992), Witt-Eickschen (1993), Witt-Eickschen *et al.* (1993, 1998), Blusztajn & Shimizu (1994), Witt-Eickschen & Harte (1994), Szabó *et al.* (1995), Vaselli *et al.* (1995, 1996), Alard *et al.* (1996), Chalot-Prat & Bouiller (1997), Rosenbaum *et al.* (1997), Witt-Eickschen & Kramm (1997), Zangana *et al.* (1997, 1998), Xu *et al.* (1998), Dobosi *et al.* (1999), Jones *et al.* (2000) and Lenoir *et al.* (2001).

incorporation of garnet pyroxenite layers into spinel peridotite by deformation. The spinel–plagioclase peridotites may result from low-pressure re-equilibration during emplacement.

SCLM xenoliths in the Neogene volcanics of Europe are mainly spinel peridotites that grade from undepleted lherzolite through clinopyroxene-poor lherzolites to harzburgites. Garnet peridotites are not found as xenoliths,

presumably because of the relatively shallow depth of segregation of the host alkali basaltic magmas. Rarer xenolith types include dunites, spinel pyroxenites, wehrlites and hornblendites. On the basis of percentages of xenoliths found, such rocks probably constitute only 1–5% of the SCLM, although they may be more abundant in certain regions. Some of these may be analogous to the mafic layers found in ultramafic massifs (Wilshire

& Pike, 1975). Very rare rock types (probably ≪1% of the SCLM) include garnet pyroxenites, amphibole pyroxenites and phlogopite pyroxenites. Extremely rare plagioclase peridotites (Brousse & Idelfonse, 1970) and garnet–spinel peridotites (e.g. Cheval et al., 1989) have been reported from France. Some ultramafic xenoliths are composite, with veins of hornblendite, pyroxenite or wehrlite (e.g. Embey-Isztin, 1976; Witt-Eickschen et al., 1993, 1998; Vaselli et al., 1995; Zangana et al., 1997). In rare locations, e.g. Gees (Eifel, Germany), the majority of xenoliths are orthopyroxene-free phlogopite wehrlites, phlogopite dunites and phlogopite clinopyroxenites, with subordinate harzburgites and very few lherzolites (Lloyd et al., 1991). Such an assemblage is extremely unusual and must indicate unusual processes of metasomatism in the SCLM beneath this area.

Thus, the sub-continental lithospheric mantle, as exemplified by material available for study in xenolith suites and ultramafic massifs in Phanerozoic Europe, is largely composed of spinel peridotite with only a few percent of other lithologies. Garnet peridotites are undoubtedly present in the thick lithosphere under the Alps (Menzies & Bodinier, 1993) and are known to be present in the lithosphere beneath the Baltic Shield, but are not sampled by Neogene alkaline magmas. They are rare among the ultramafic massifs, possibly because of the difficulties of tectonic exhumation from the great depths they represent.

The European SCLM presents a very narrow range of common lithologies dominated by an anhydrous spinel lherzolite-to-harzburgite series (Downes, 1997). This world-wide observation is probably the result of ancient depletion of the mantle as a result of extraction of basalts (Ottonello et al., 1984; Frey et al., 1985; Hartmann & Wedepohl, 1993); however, this view has been questioned by Kelemen et al. (1992), who argued that harzburgites may result from the interaction of mantle wall-rock and melt, and by Burnham et al. (1998), who developed a model of refertilization of a highly depleted mantle residue. The next most common rock-type in the SCLM is pyroxenite, considered by many workers to be the product of passage of asthenosphere-derived magmas (Bodinier et al., 1987a, 1987b; Girardeau & Gil Ibarguchi, 1991). However, Allègre & Turcotte (1986) argued that pyroxenites are the remains of oceanic crust returned to the mantle by subduction. Hornblendite veins and amphibole peridotites or pyroxenites are generally considered to be formed by the passage of mafic alkaline magmas (e.g. Wilshire et al., 1980; Vaselli et al., 1995; Witt-Eichschen et al., 1998). Wehrlites, which might represent mantle that has interacted with carbonatite melts (Yaxley et al., 1998), are generally very subordinate and spatially restricted. Phlogopite-rich peridotites, which could represent mantle that has experienced subduction flux, are abundant in Europe only as xenoliths from Germany (Kempton et al., 1988; Hartmann & Wedepohl, 1990;

Lloyd et al., 1991), Italy (Conticelli & Peccerillo, 1990), central Spain (Ancochea & Nixon, 1987) and in the Finero massif (Zanetti et al., 1999).

GEOCHEMISTRY OF THE EUROPEAN SCLM
Major elements in bulk rocks
In Fig. 2, key oxide abundances and ratios in bulk-rock spinel peridotites from the European SCLM have been plotted. Data from xenolith suites form very tight arrays from fertile lherzolite to refractory harzburgite, i.e. Al_2O_3, TiO_2 and CaO all correlate negatively with MgO. The arrays are usually considered to be due to removal of a basaltic component from the fertile upper mantle (Frey & Green, 1974) and are identical to those found by McDonough (1990) and Griffin et al. (1998) in studies of world-wide spinel peridotite xenoliths. These arrays are sufficiently robust to enable us to determine whether xenolith suites or massif samples show significant deviations from the normal SCLM. Major-element data from ultramafic massifs follow very similar trends to those shown by the xenoliths (Fig. 2). Only a few peridotites fall outside the arrays and these must have therefore experienced some unusual process within the mantle. In the case of the samples that have higher CaO and Al_2O_3, together with lower MgO, the simplest explanation is that they are mechanical mixtures between peridotites and pyroxenite layers. Such layering occurs on very small scales in many massifs and it may not be possible to remove all of the pyroxenite material before bulk-rock analysis. Mantle pyroxenites plot away from the peridotite major element trends (Fig. 2), having lower MgO contents and variable but generally high concentrations of Al, Ca and Ti. Pyroxenites probably formed as cumulates from mafic melts, and their compositions are a reflection of (1) the nature of the magma from which they were formed, and (2) the abundances of their constituent cumulate minerals.

CaO/Al_2O_3 ratios in spinel peridotites are generally uniform at ∼0·87 (Fig. 2). A few have CaO/Al_2O_3 ratios as low as 0·05; these are mainly harzburgites and their low CaO/Al_2O_3 ratios probably result from the complete melting of clinopyroxene. High CaO/Al_2O_3 ratios (>1·0) are found mainly in wehrlites and, if not due to superficial calcitic alteration, such values could be related to carbonatite metasomatism (Yaxley et al., 1991). Pyroxenites show a wide range of CaO/Al_2O_3 ratios, from 0·5 to 4·5, reflecting their more varied mineralogy, and possibly sea-floor alteration if the hypothesis of their origin as ocean crust is correct.
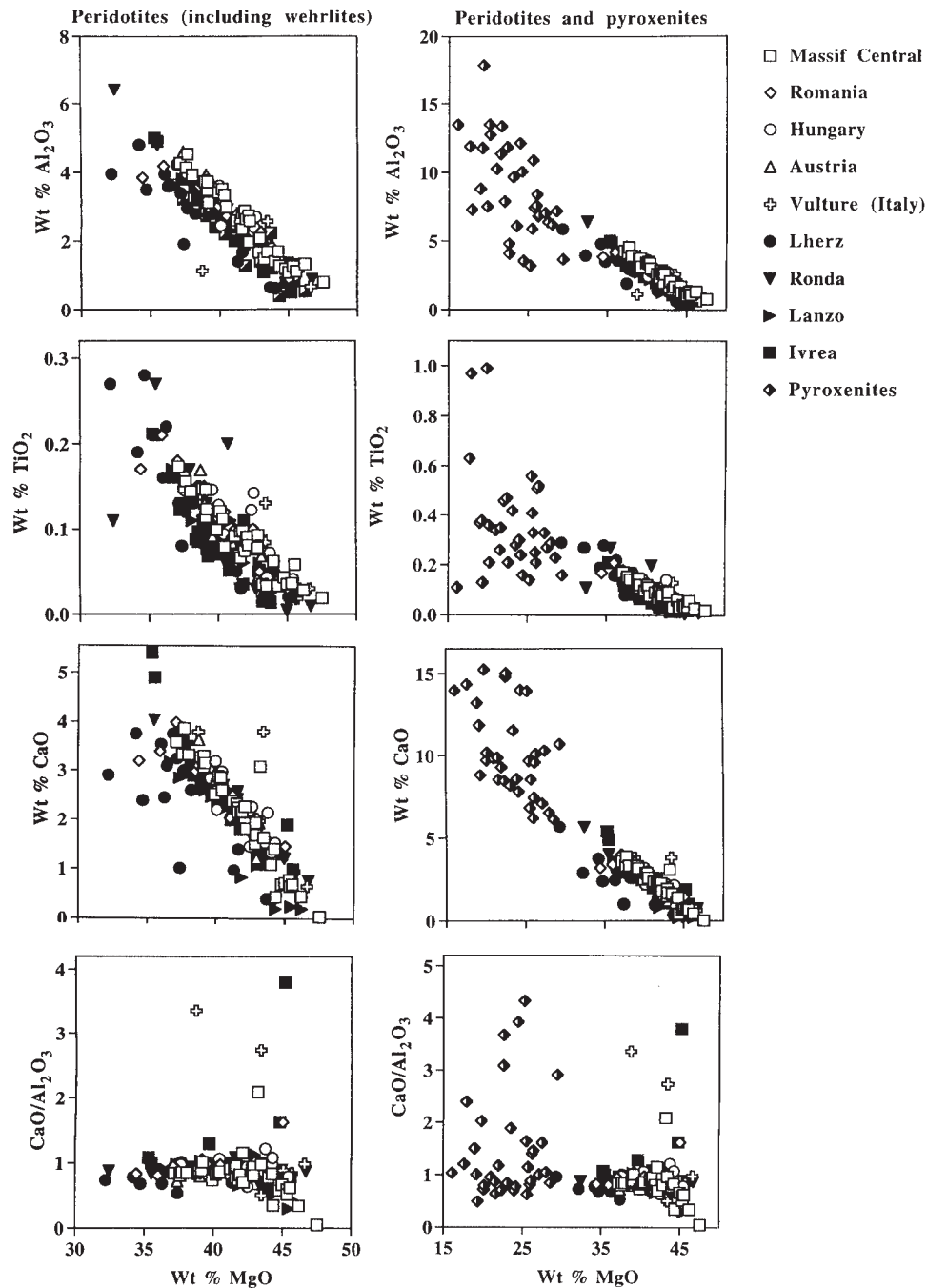
Fig. 2. Left-hand column: Al_2O_3, TiO_2, CaO and CaO/Al_2O_3 ratio vs MgO in European SCLM peridotites, including wehrlites. Open symbols, xenolith suites; filled symbols, ultramafic massifs. Data sources: Frey *et al.* (1985), Downes & Dupuy (1987), Bodinier (1988), Downes *et al.* (1992), Hartmann & Wedepohl (1993), Rivalenti *et al.* (1995), Vaselli *et al.* (1995, 1996), Zangana *et al.* (1998) and H. Downes & T. Kostula, unpublished data (1999). Right-hand column: samples as in left-hand column, but including data from mantle pyroxenites (both massifs and xenolith suites). Pyroxenite data from Bodinier *et al.* (1987*b*), Rivalenti *et al.* (1995) and Vaselli *et al.* (1995).

Incompatible trace elements in bulk SCLM

The rare earth elements (REE) are incompatible with anhydrous spinel peridotite, and removal of a basaltic component from the mantle strongly affects their relative abundances [see review by McDonough & Frey (1989)]. As the light REE (LREE) are more incompatible with mantle assemblages than the heavy REE (HREE), they become strongly depleted in the refractory mantle. The

ubiquitous LREE depletion is thus readily explained as the result of extraction of basalt from the mantle, i.e. the same process that controls the major element trends. Even xenoliths that are apparently undepleted in major-element terms still show LREE depletion. The majority of whole-rock lherzolite samples from ultramafic massifs are LREE depleted, as are many lherzolite xenoliths (Fig. 3). Harzburgites additionally show HREE depletion, probably as a result of extensive melting of clinopyroxene, which is the major host of all REE in the anhydrous SCLM. Layered pyroxenites in peridotite massifs tend to show LREE depletion; this may be due to the original chemistry of the magma that formed the pyroxenite [e.g. streaked out layers of mid-ocean ridge basalt (MORB)-type oceanic crust] or because of subsequent equilibration with the surrounding LREE-depleted peridotites. Over-printing this depletion is a variable LREE enrichment, particularly shown by xenoliths (Fig. 3). Both LREE-enriched and LREE-depleted xenoliths can occur in the same volcanic vent (Zangana *et al.*, 1997). Massif peridotites tend to show LREE enrichment only when they are the host rock to amphibole-rich veins, which Bodinier *et al.* (1990) showed is due to chromatographic fractionation. Rare U-shaped patterns [showing middle REE (MREE) depletion relative to HREE and LREE] are found in some peridotite xenoliths and have been interpreted as resulting from carbonatite metasomatism.

Trace-element signatures in bulk rocks can be used to some extent to fingerprint mantle processes. A range of incompatible trace elements can be plotted on extended mantle-normalized diagrams, in a manner analogous to the REE diagrams. Figure 4 shows a range of mantle-normalized peridotite compositions from the European SCLM. In Fig. 4a, the LREE-depleted examples show smoothly decreasing concentrations from right to left in the diagram, as incompatibility of the elements increases. This indicates that the depleted mantle has lost its highly incompatible elements, as well as its LREE (Hartman & Wedepohl, 1993). Increases in Nb, Ba and Rb may be due to incipient metasomatism. Mantle that has interacted with a silicate melt would be expected to show increases in all incompatible elements (Fig. 4b), although the effect on elements such as Rb, Ba and Nb, which have been previously strongly depleted, may not be as obvious as in the LREE. Samples that may have been fluxed by subduction fluids would be expected to show strong enrichment in incompatible elements that are fluid soluble, and may be phlogopite bearing. Such samples tend to have negative anomalies at Nb, Zr and Hf relative to adjacent elements, and strong enrichment in Rb and Ba (Fig. 4c). Carbonatite melt enrichment would also result in LREE enrichment with extremely low concentrations of Nb, Zr and Hf (Ionov *et al.*, 1993). Figure 4d shows the bulk-rock trace-element composition of a peridotite xenolith that has been postulated to be carbonatite enriched. Lenoir *et al.* (2001) have used incompatible element ratios such as Zr/Hf and Nb/Ta in bulk-rock peridotite xenoliths to define two separate domains within the SCLM of the French Massif Central, one of which is more ancient and has experienced a long history of metasomatism, whereas the other is more juvenile and less strongly metasomatized. It remains to be seen whether such domains can be traced elsewhere in the mantle of Phanerozoic Europe.

REE geochemistry of clinopyroxene and amphibole

REE patterns of SCLM clinopyroxenes are usually very similar to those of the host peridotite, unless the peridotite contains another phase with a high distribution coefficient for REE. As the content of REE in clinopyroxene is about 7–10 times higher than in the bulk rock, analysis of clinopyroxene is much easier than analysis of bulk peridotites or pyroxenites. Figure 5 shows typical REE patterns for mantle minerals in European SCLM. Clinopyroxenes from many mantle peridotites are LREE depleted (Fig. 5a and b). Those from pyroxenites have variable REE patterns (Fig. 5c). Some are very similar to those from LREE-depleted peridotites and may have equilibrated with host LREE-depleted peridotite mantle. Others are additionally HREE depleted, as the pyroxene is in equilibrium with garnet in the rock, which preferentially takes up the HREE. Many mantle peridotites contain clinopyroxene that shows strong enrichment in LREE as a result of the effect of the passage of LREE-enriched fluids or magmas (Fig. 5d and e). However, there are several varieties of LREE-enriched patterns in mantle clinopyroxene, some of which actually show a depletion in La and Ce relative to Nd and Sm, as in clinopyroxene from the amphibole-metasomatized peridotite massif of Caussou (Fig. 5f), or show strong depletion in the MREE, as in peridotite xenoliths from Poland (Fig. 5g) interpreted by Blusztajn & Shimizu (1994) as being metasomatized by a carbonatite fluid or melt. A similar pattern, but showing extreme depletion in Ce and Nd ($0.1 \times$ chondrite), is shown by clinopyroxene in xenoliths from Montboissier (France) (Fig. 5h). Xenoliths from Mte Vulture and peridotites from the Finero massif (both interpreted as representing subduction-metasomatized SCLM) tend to show S-shaped patterns, with flat LREE–MREE sections and MREE enrichment over HREE (Fig. 5i).

With the advent of secondary ion mass spectrometry (SIMS) and laser–inductively coupled plasma mass spectrometry (ICPMS) techniques, very large numbers of REE analyses of clinopyroxenes from the SCLM are now available. Attempting to show differences between
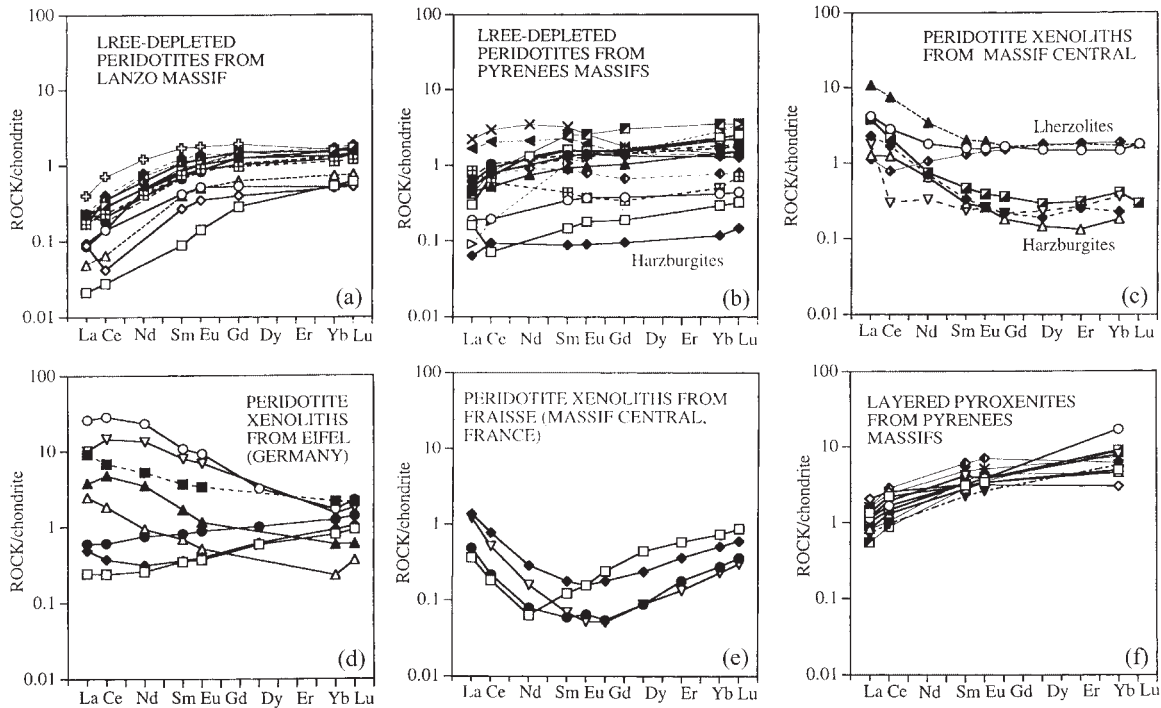
Fig. 3. Chondrite-normalized REE diagrams for selected whole-rock peridotites and pyroxenites. Normalization coefficients from Nakamura (1974). (a) LREE-depleted peridotites from Lanzo massif, Italy (Bodinier, 1988); (b) LREE-depleted peridotites from Pyrenees peridotite massifs, France (Bodinier *et al.*, 1988); (c) LREE-depleted and LREE-enriched peridotite xenoliths from Ray Pic, Massif Central, France (Zangana *et al.*, 1998); (d) LREE-enriched peridotite xenoliths from Eifel, Germany (Stosch & Seck, 1980); (e) U-shaped REE patterns from peridotite xenoliths from Fraisse, Massif Central, France (Lenoir *et al.*, 1997); (f) layered pyroxenites in Pyrenees ultramafic massifs (Bodinier *et al.*, 1987*a*, 1987*b*).

suites by plotting all these data on conventional chondrite-normalized REE diagrams, as in Fig. 5, becomes problematic and it is useful to represent each REE pattern as a single data point. REE data for mantle clinopyroxenes can be plotted by representing the LREE/MREE slope by the La/Nd ratio and the MREE/HREE slope by the Sm/Yb ratio. Figure 6 shows such a plot of results from the European SCLM. LREE-depleted (lower half of the diagram) and LREE-enriched (upper half of the diagram) patterns are clearly differentiated. Clinopyroxenes from peridotite xenoliths show a much wider variation in the two parameters than do peridotites from ultramafic massifs, which tend to be LREE depleted. Within the LREE-enriched half of the diagram (La/Nd >0·522), there are two important trends, one towards high La/Nd at moderate Sm/Yb, and one towards high Sm/Yb at moderate La/Nd. The latter group are mainly clinopyroxenes from harzburgites that have experienced HREE depletion as well as LREE enrichment, although clinopyroxenes from phlogopite-metasomatized peridotites from Finero also show this trend, as do those from xenoliths in Vulture volcano (Italy). A few rare samples from the Massif Central and Germany contain clinopyroxene with extremely low Sm/Yb and extremely high La/Nd, i.e. strongly U-shaped patterns. These

samples have undergone depletion (giving rise to low Sm/Yb) followed by enrichment only in LREE. In contrast, only a few peridotites contain clinopyroxenes with N-shaped patterns (depletion in both LREE and HREE), although pyroxenes from garnet-bearing pyroxenites plot in this area, probably as a result of subsolidus redistribution of REE with the coexisting garnet.

Metasomatism can affect the mineralogy of the SCLM, with the development of amphibole and phlogopite, e.g. as amphibole peridotites or hornblendites (Wilshire & Trask, 1971; Wilshire *et al.*, 1980). Examples from the European SCLM include xenoliths from Germany (Witt-Eickschen *et al.*, 1993, 1998) and Romania (Vaselli *et al.*, 1995) and Pyrenean peridotite massifs (Downes *et al.*, 1991; McPherson *et al.*, 1996; Woodland *et al.*, 1996; Zanetti *et al.*, 1996). Amphiboles in veins and in amphibole-rich peridotites have high REE contents and are usually enriched in LREE (Fig. 7a). The effect of an amphibole-rich vein on the REE pattern of the adjacent mantle is very clear, in that an increase in LREE enrichment can be observed in clinopyroxenes near the vein, compared with those farther away. In the Caussou massif (Pyrenees) no enriched veins are present, but instead the entire massif is rich in strongly LREE-enriched clinopyroxene and amphibole. This has been interpreted by Fabriès *et al.* (1989) and Downes *et al.*
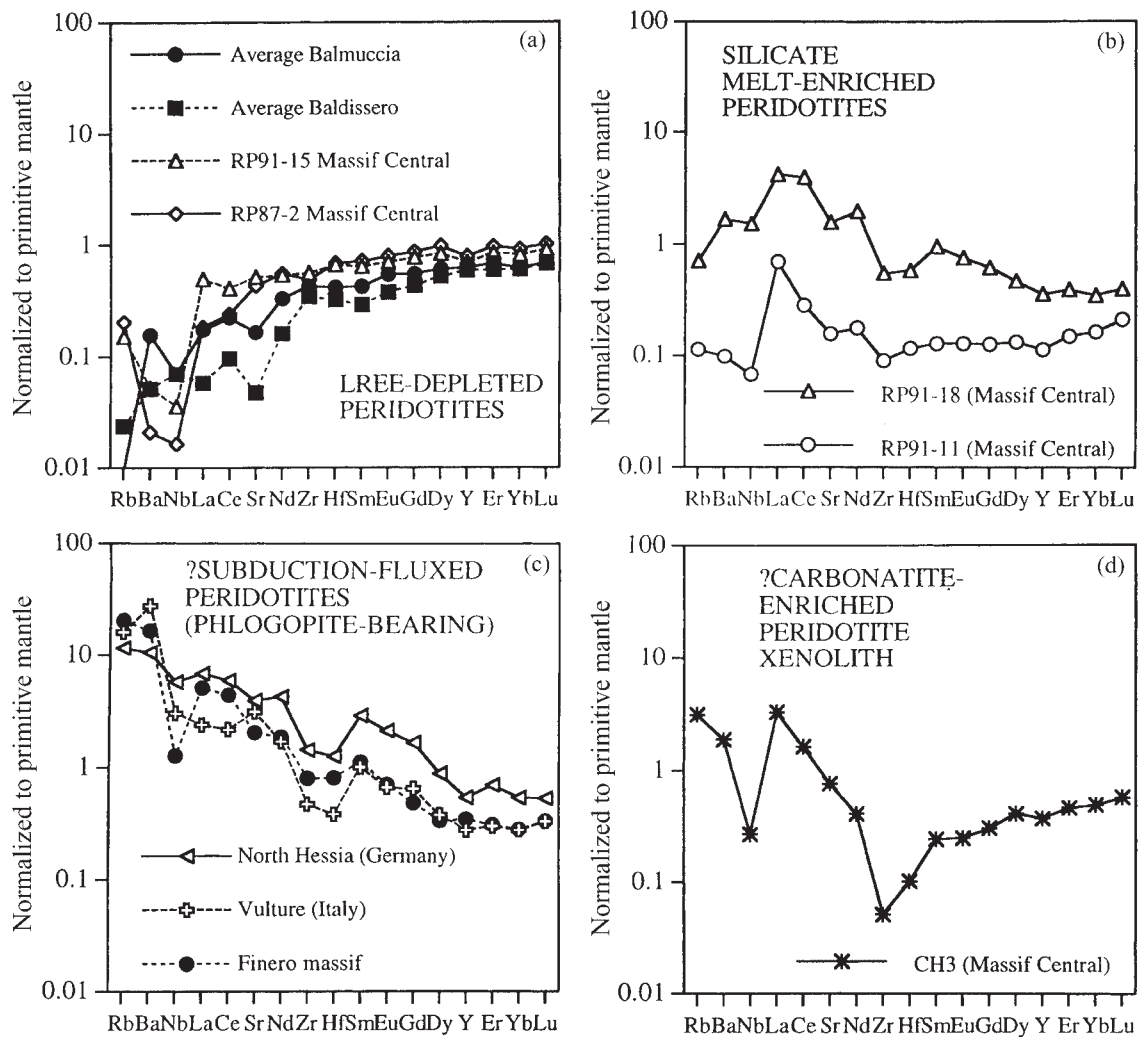
Fig. 4. Examples of mantle-normalized incompatible trace element diagrams for whole-rock peridotite samples from western Europe. (a) LREE-depleted peridodites from Balmuccia and Baldissero massifs (Alps) and Massif Central xenoliths (data from Hartmann & Wedepohl, 1993; Zangana *et al.*, 1998); (b) silicate-melt enriched peridotite xenoliths, Massif Central (H. Downes & J.-L. Bodinier, unpublished data, 1999, ICP-MS, University of Montpellier); (c) possible subduction-fluxed peridotite xenoliths from North Hessen and Vulture and massif peridotite from Finero (Alps) (Hartmann & Wedepohl, 1990, 1993; H. Downes & J.-L. Bodinier, unpublished data, 1999, ICP-MS, University of Montpellier); (d) possible carbonatite-enriched peridotite xenolith, Massif Central (Lenoir *et al.*, 1997). Normalizing coefficients from Sun & McDonough (1989).

(1991) as being due to interaction of originally depleted mantle with mafic alkaline magmas, in which the magma/mantle ratio was very high. In contrast, disseminated or interstitial amphiboles in ultramafic massifs and xenoliths are commonly LREE depleted (Fig. 7b). Vannucci *et al.* (1995) considered that this is due to diffusive equilibration with the surrounding LREE-depleted mantle.

RADIOGENIC ISOTOPE GEOCHEMISTRY OF THE EUROPEAN SCLM

Wide variations in $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic compositions are present in clinopyroxenes from mantle

xenoliths of Europe, ranging from ratios similar to those of depleted mantle to values more enriched than Bulk Earth (Fig. 8a). Xenoliths from a single locality can cover this entire range, as can samples from a single massif. The depleted mantle signature is the time-integrated effect of extraction of basalt leaving an LREE- and large ion lithophile element (LILE)-depleted residue, which has evolved to high $^{143}Nd/^{144}Nd$ and low $^{87}Sr/^{86}Sr$. The time-integrated effect of ancient enrichment in the SCLM will manifest itself in high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$. Values around Bulk Earth could also be due to enrichment by passage of fluids and melts through the mantle. The array of data from ultramafic massifs (Fig. 8b) is similar to that of the xenoliths but not quite
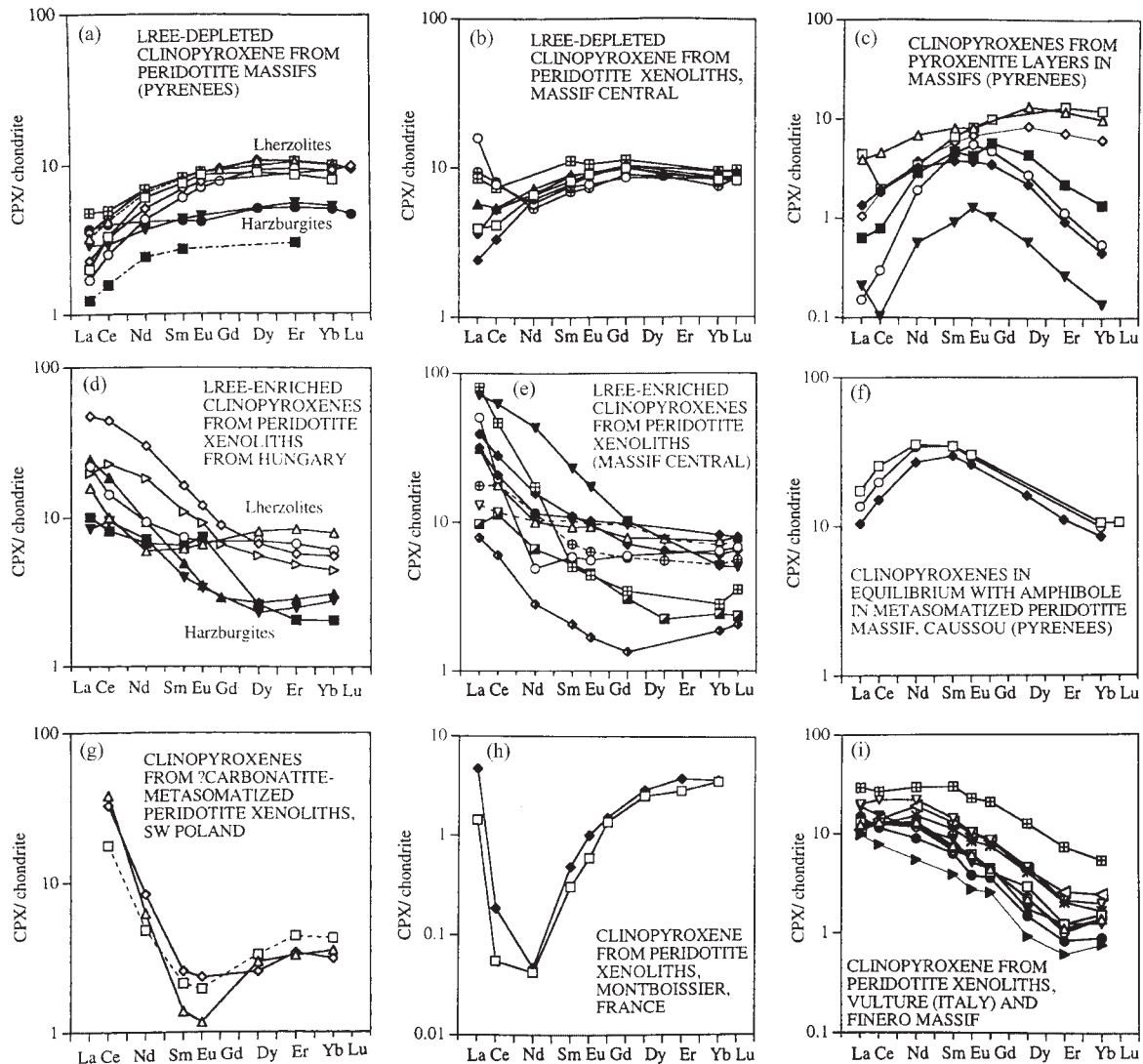
Fig. 5. Chondrite-normalized REE diagrams for separated minerals from SCLM peridotites: (a) LREE-depleted clinopyroxenes from ultramafic massifs, Eastern Pyrenees (Downes *et al.*, 1991); (b) LREE-depleted clinopyroxenes from spinel peridotite xenoliths, Massif Central (Downes & Dupuy, 1987); (c) clinopyroxenes from pyroxenite layers in ultramafic massifs, Pyrenees; open symbols: spinel pyroxenites; filled symbols: garnet-bearing pyroxenites (Downes *et al.*, 1991); (d) LREE-enriched clinopyroxenes from spinel peridotite xenoliths, Hungary (Downes *et al.*, 1992); (e) LREE-enriched clinopyroxenes from spinel peridotite xenoliths, Massif Central (Downes & Dupuy, 1987); (f) LREE-enriched clinopyroxenes in equilibrium with amphibole from metasomatized peridotite massif, Caussou, Pyrenees (Downes *et al.*, 1991); (g) clinopyroxenes with U-shaped REE patterns from carbonatite-metasomatized xenoliths, Poland (Blusztajn & Shimizu, 1994); (h) clinopyroxenes with extreme depletion in Ce and Nd from peridotite xenoliths from Montboissier (Massif Central) (Vannucci *et al.*, 1994). Chondritic normalization coefficients from Nakamura (1974). (i) LREE-enriched clinopyroxenes from Vulture (Italy) peridotite xenoliths (H. Downes & J.-L. Bodinier, unpublished data, 1999) and Finero ultramafic massif (Zanetti *et al.*, 1999). [Note change of scale in plots (c), (h) and (i).]

identical. Among the xenoliths, a large number of samples fall within the field for European Neogene alkali basalts (Wilson & Downes, 1991); data from the peridotite massifs in general do not fall within this field. This may indicate that some of the geochemical enrichment in the xenoliths may be due to infiltration of melts from the Neogene mantle plume beneath Europe (Hoernle *et al.*, 1995; Granet *et al.*, 1995; Wilson & Patterson, 2001). Some SCLM samples (e.g. xenoliths from Hungary and Italy)

deviate from the main field towards high $^{87}Sr/^{86}Sr$ values. This may be the influence of subduction-derived fluids, as both these areas are situated above Neogene subduction zones. Subduction fluids would contain relatively high concentrations of Sr but very little Nd, and therefore their effect on the mantle would be greater in $^{87}Sr/^{86}Sr$.

Sm–Nd isochron diagrams for European SCLM samples (Fig. 9) generally show positive correlations. These could be interpreted as having age significance,
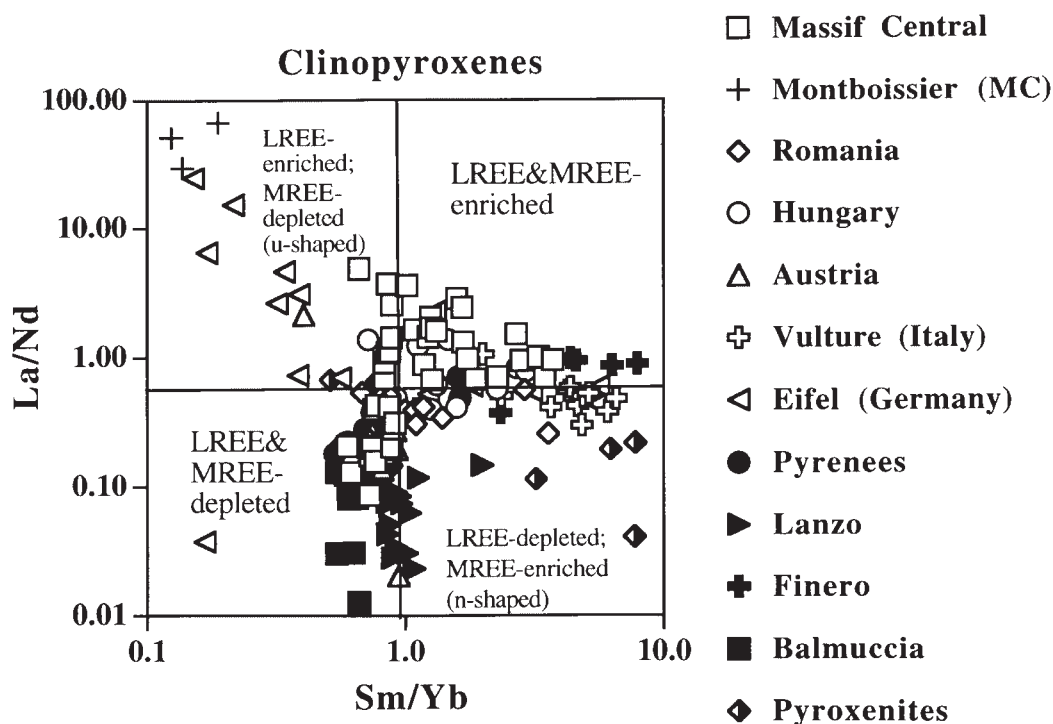
Fig. 6. La/Nd vs Sm/Yb diagram for clinopyroxenes from peridotite xenoliths (open symbols) and massifs (filled symbols) in Europe. Data sources: Pyrenees and pyroxenites, Downes *et al.* (1991); Hungary, Downes *et al.* (1992); Massif Central, Downes & Dupuy (1987) and Zangana *et al.* (1997); Pyrenees and Lanzo, Bodinier *et al.* (1998, 1990, 1991); Pyrenees, Zanetti *et al.* (1996); Montboissier, Vannucci *et al.* (1994); Romania, Vaselli *et al.* (1995); Austria, Vaselli *et al.* (1996); Balmuccia, Rivalenti *et al.* (1995); Vulture, T. Kostoula & H. Downes (unpublished data, 1999); Germany, Stosch & Lugmair (1986); Finero, Zanetti *et al.* (1999).

in which case the European SCLM is generally of late Proterozoic to Palaeozoic age. However, more probably this correlation is due to mixing between ancient depleted mantle with high $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios and an enriched end-member with lower $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios. This may be the plume component (Witt-Eickschen *et al.*, 1998); peridotite massifs show this component to a much smaller degree. The effects of ancient enrichment can be seen in the low $^{143}Nd/^{144}Nd$ of some massifs. Other samples fall away from the correlation, as a result of processes that have occurred subsequently. Recent depletion will cause an increase in $^{147}Sm/^{144}Nd$ (LREE depletion) without affecting $^{143}Nd/^{144}Nd$, whereas recent melt enrichment could cause a decrease in $^{147}Sm/^{144}Nd$ (LREE enrichment) and will alter the $^{143}Nd/^{144}Nd$ ratio towards that of the metasomatizing melt. Subduction-zone fluid contamination of the mantle should have no effect on Nd isotope systematics.

Pb isotope data for clinopyroxenes from much of the European SCLM fall on an array parallel to the Northern Hemisphere Reference Line (Hart, 1984), indicative of mixing between depleted mantle with unradiogenic Pb isotope compositions and a plume-type mantle with more radiogenic Pb (Fig. 10). The latter component may

relate to asthenosphere upwelling during Tertiary times (Hoernle *et al.*, 1995; Wilson & Patterson, 2001). In contrast, SCLM beneath Hungary shows a trend towards high $^{207}Pb/^{204}Pb$ at a given $^{206}Pb/^{204}Pb$. This trend is more extreme than in other European SCLM samples and probably indicates a component of subducted sediment derived from ancient continental crust (Rosenbaum *et al.*, 1997). The Lherz massif shows this feature to a limited extent (Fig. 10), perhaps as a result of Hercynian subduction. Further high-precision Pb isotope data are required to determine whether similar high $^{207}Pb/^{204}Pb$ mantle underlies other regions of Europe that have experienced subduction.

Several Os-isotope studies of SCLM material from western Europe have been published in the past decade. The Re–Os isotope system differs from Sr, Nd and Pb in that the daughter product Os is highly compatible in the residual mantle during partial melting, whereas the parent Re is incompatible. Thus, melting lowers the Re/Os ratios in the residue, so highly depleted mantle may in effect retain the Os isotope composition of the mantle at the time of melting. The Os-isotope system may therefore be used to determine the age of depletion of the lithospheric mantle. Reisberg *et al.* (1991) suggested an age of 1·2 Ga for the time at which the Ronda
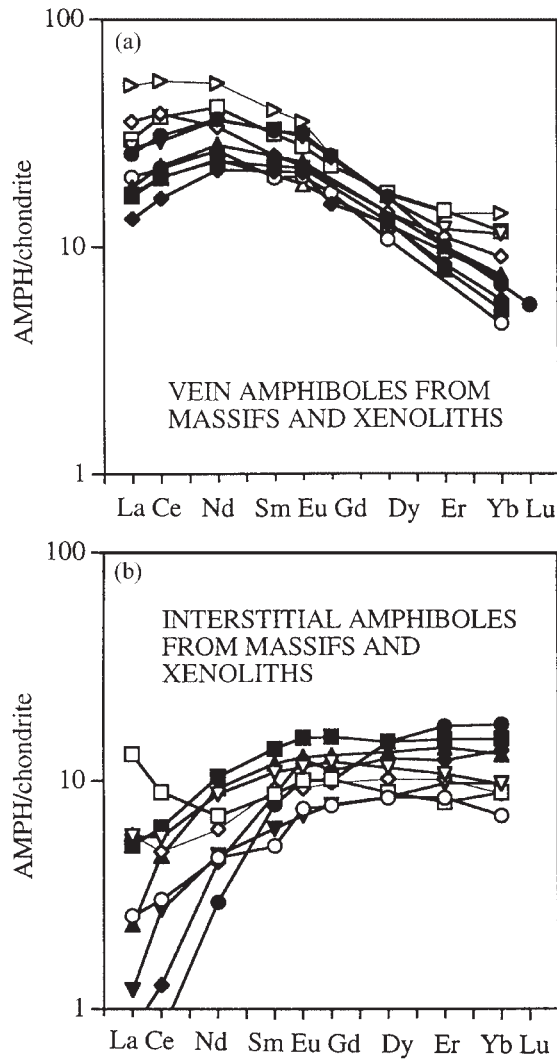
Fig. 7. (a) LREE-enriched vein amphiboles in xenoliths from Germany, Hungary and Romania (Ionov & Hofmann, 1995; Chalot-Prat & Bouiller, 1997; Witt-Eickschen et al., 1998) and in Finero and Lherz ultramafic massifs (Zanetti et al., 1996, 1999). (b) LREE-depleted amphiboles in xenoliths from Romania and in Alps and Pyrenees massifs (Vannucci et al., 1995; Vaselli et al., 1995). Chondritic normalization coefficients from Nakamura (1974).

Fig. 8. (a) Sr–Nd radiogenic isotope variations in shallow SCLM xenoliths from Europe; (b) ultramafic massifs in western Europe. Data sources: (a) Massif Central, Downes & Dupuy (1987) and Zangana et al. (1997); Hungary, Downes et al. (1992); Eifel, Stosch & Lugmair (1986) and Witt-Eickschen et al. (1998); Romania, Vaselli et al. (1995); Austria, Vaselli et al. (1996); Poland, Blusztajn & Shimizu (1994); Mte Vulture, T. Kostoula & H. Downes (unpublished data, 1999); (b) ultramafic massifs from Pyrenees, Downes et al. (1991) and Mukasa et al. (1991); Balmuccia, Shervais & Mukasa (1991) and Mukasa & Shervais (1999); Lanzo, Bodinier et al. (1991); Ronda, Reisberg & Zindler (1986) and Reisberg et al. (1989); Finero, Obermiller (1994). Plume component from Wilson & Downes (1991).

massif left the convecting asthenosphere and became incorporated into the lithosphere. Such an age is not to be confused with the time at which the massif became emplaced into the crust, dated at 22 Ma. In a study of eastern Pyrenean ultramafic massifs, Burnham et al. (1998) obtained an Os model age for mantle depletion of ∼1·9 Ga, somewhat younger than the 2·3 Ga obtained by Reisberg & Lorand (1995) for the same area, but in agreement with the model age of peridotite xenoliths from Germany (Cohen et al., 1996) and still significantly older than the apparent age of Ronda. Further studies
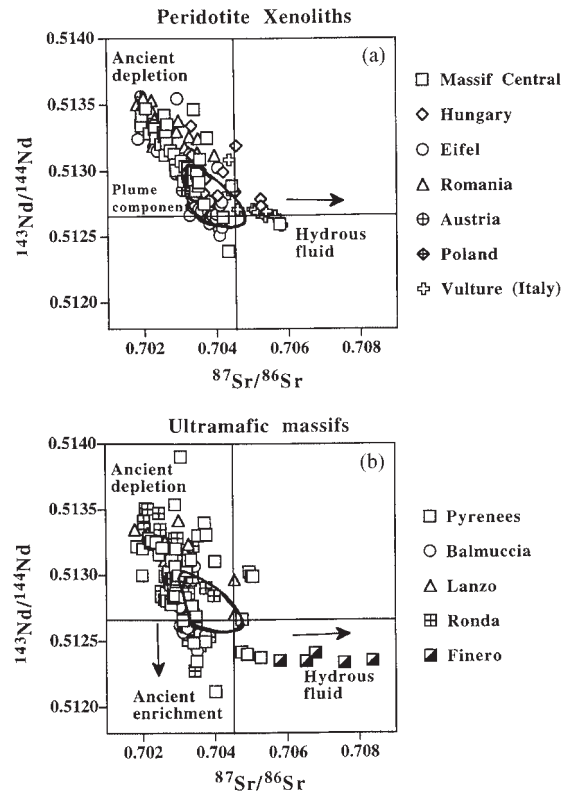
are needed to establish if the differences in Os model ages are consistent across the European continent and whether lithospheric domains can be traced using Os-isotopes.

Finally, a study by Dunai & Baur (1995) of the He, Ne and Ar systematics of European peridotite xenoliths showed that their noble gases are dominated by the signature of the magmatic host rocks, and therefore these results are more important for studies of the astheno-spheric mantle rather than the lithospheric mantle. Nevertheless, it was shown that the source of the Neogene alkaline magmas of western Europe was enriched by subduction of crustal material during the Variscan oro-geny. Contamination of the SCLM by passage of such magmas would enrich the lithosphere in this subducted component.
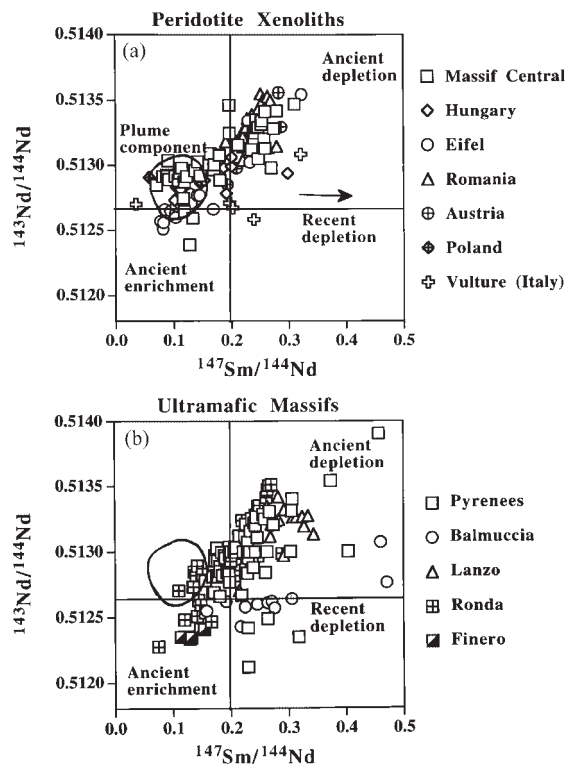
Fig. 9. $^{143}Nd/^{144}Nd$–$^{147}Sm/^{144}Nd$ isochron diagrams for shallow SCLM xenoliths from Europe and ultramafic massifs in western Europe. Data sources as for Fig. 8.
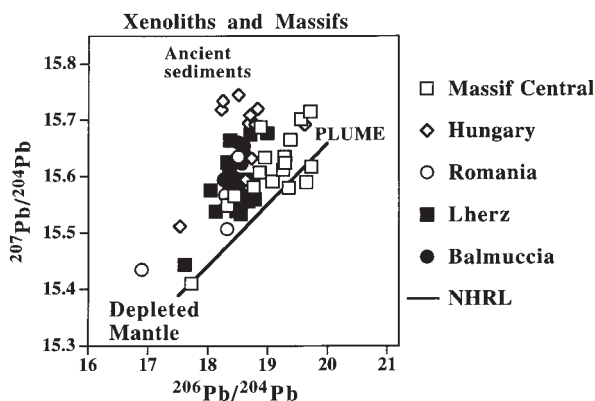

Fig. 10. $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ isotope variation in Pyrenean ultramafic massifs (Mukasa *et al.*, 1991); Balmuccia (Mukasa & Shervais, 1999); Massif Central xenoliths (Zangana *et al.*, 1997; Wilson *et al.*, in preparation); xenoliths from Hungary and Romania (Rosenbaum *et al.*, 1997). NHRL, Northern Hemisphere Reference Line (Hart, 1984). Plume component from Wilson & Downes (1991).

DISCUSSION AND INTERPRETATION

Major processes that have been identified as causing compositional variations in the SCLM are: (1) ancient depletion as a result of extraction of basalts, as discussed by Frey & Green (1974), Ottonello *et al.* (1984) and Frey

et al. (1985); (2) enrichment because of passage of silicate magmas [see review by Nielson & Noller (1987)]; (3) enrichment as a result of influx of highly mobile carbonate magmas (e.g. Yaxley *et al.*, 1991, 1998; Ionov *et al.*, 1993); (4) enrichment caused by hydrous subduction-related fluids, suggested by Hartmann & Wedepohl (1990) and Zanetti *et al.* (1999). Each process has a unique signature. These processes and their chemical signatures will now be considered in turn. Finally, the origin of pyroxenites will be briefly discussed.

Mantle depletion by extraction of basaltic melts

This is readily observed in major-element variations, as it is the process that governs the large-scale composition and mineralogy of the SCLM. Linear trends in Al_2O_3, CaO and TiO_2 vs MgO (Fig. 2) are the result of the removal of basalt magma from undepleted mantle. This depletion eventually converts the SCLM from lherzolite to harzburgite. LREE depletion (Fig. 3) is also caused by the same process, but occurs even in peridotites that have undepleted major element compositions. Sr, Zr, Hf and Y tend to behave coherently with REE of similar incompatibility (Fig. 4) whereas the strongly incompatible elements Nb, Rb and Ba are often very strongly depleted in xenoliths that have experienced melt removal. If partial melting has been extreme and clinopyroxene is melted completely, neither HREE nor CaO can be accommodated in the remaining mineralogy and are released into the silicate melts. The resulting harzburgitic residual mantle is HREE depleted (Fig. 3) and has low CaO/Al_2O_3 ratios (Fig. 2). The time-integrated effects of LREE and LILE depletion are high $^{143}Nd/^{144}Nd$, low $^{87}Sr/^{86}Sr$ (Fig. 8), and unradiogenic Pb isotope ratios similar to those of the MORB source (Fig. 9). The age of depletion can be estimated from Os-isotope characteristics and ranges from 1·2 to 2·0 Ga in different regions of Europe. Nd isotope model ages for LREE-depleted xenoliths from western Europe also fall in the same range.

Enrichment by passage of silicate magmas

Much of the enrichment in the European SCLM represented by xenoliths is related to the Neogene asthenospheric mantle plume and is commonly accompanied by the formation of LREE-enriched amphiboles and clinopyroxenes (Figs 5 and 7). This component is largely absent from massifs. Mechanisms of enrichment by the passage of silicate magmas through the mantle such as reactive porous flow and chromatographic fractionation (Navon & Stolper, 1987; Kelemen, 1990; Takazawa *et al.*, 1992) have been widely studied in the

SCLM of Phanerozoic Europe (Bodinier *et al.*, 1990; McPherson *et al.*, 1996; Woodland *et al.*, 1996; Zanetti *et al.*, 1996). Modal metasomatism often occurs as hornblende- or pyroxene-rich veins, aureoles around veins, or as hornblende- and clinopyroxene-rich peridotites. The metasomatizing magmas are considered to be derived from the deeper asthenospheric mantle (Vaselli *et al.*, 1995; Witt-Eickschen *et al.*, 1998). The effect of their passage on the major-element composition of the SCLM is limited, but incompatible trace elements are strongly affected, with enrichment in LREE, Sr, Zr, Hf and often Nb in the bulk rock (Fig. 3). LREE enrichment in clinopyroxene can also result from cryptic metasomatism, when no new phases are formed (Fig. 4). Metasomatic hornblende in veins is enriched in LREE, Nb and other incompatible elements (Fig. 7) and this enrichment often decreases towards the edge of the vein and into hornblende in the host peridotite (e.g. Witt-Eickschen *et al.*, 1998). Changes in isotope composition of the SCLM towards that of the invading magma are observed (Vaselli *et al.*, 1995). In massifs, dykes of anhydrous pyroxenite and gabbro are also considered to represent several generations of addition of asthenospheric melts (Downes *et al.*, 1991; Mukasa & Shervais, 1999) but are not related to the Neogene plume magmatism.

Enrichment by passage of carbonate magmas

Carbonatite enrichment of the SCLM can be detected from petrographic features, trace element patterns in clinopyroxene, and bulk-rock geochemistry in a few European xenolith localities. Mineralogical evidence for interaction between carbonate magmas and the SCLM can lead to the formation of wehrlites (Yaxley *et al.*, 1991) or primary mantle carbonates, which may be present as pockets of granular dolomite or as patches and veins of quenched carbonate melt (Ionov *et al.*, 1993). This latter type of interaction has been suggested from petrographic studies of SCLM xenoliths from Hungary (Bali *et al.*, 1999), Italy (Jones *et al.*, 2000) and Romania (Chalot-Prat & Arnold, 1999). Interaction between carbonate magmas and the SCLM should increase CaO/Al_2O_3 ratios in whole rocks, as occurs in a wehrlite xenolith from the Massif Central (Zangana *et al.*, 1997). Yaxley *et al.* (1998) have suggested that carbonatite metasomatism in the SCLM will reduce the Ti/Eu ratio in the bulk peridotites. Ti/Eu is generally considered to remain unchanged by different degrees of partial melting, so lherzolites and harzburgites should carry the primitive mantle Ti/Eu ratio (7740), but carbonatites have significantly lower Ti/Eu ratios. In Fig. 11, Ti/Eu is plotted vs Ti concentration for European SCLM peridotites. Most have Ti/Eu ratios between 2400 and 32 000 (Fig.
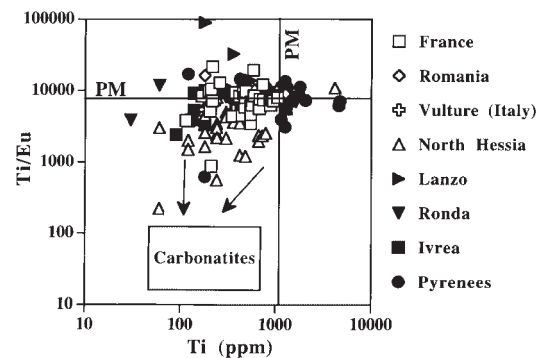

Fig. 11. Log Ti/Eu vs log Ti ppm in bulk-rock peridotites from the European SCLM. Open symbols, xenoliths; filled symbols, samples from ultramafic massifs. Primitive mantle values from Sun & McDonough (1989). Data sources: France, Zangana *et al.* (1998); Romania, Chalot-Prat & Bouiller (1997); Vulture, T. Kostoula & H. Downes (unpublished data, 1999); North Hessen, Hartmann & Wedepohl (1991); Lanzo, Bodinier (1988); Ronda, Frey *et al.* (1985); Ivrea, Hartmann & Wedepohl (1993) and Rivalenti *et al.* (1995); Pyrenees, Bodinier *et al.* (1987*a*, 1987*b*).

11), but a few (notably from North Hessen, Germany) have substantially lower Ti/Eu ratios, i.e. <1000. From the bulk-rock evidence, carbonatite metasomatism appears to have had limited impact on the European SCLM.

Xenoliths from Poland were identified by Blusztajn & Shimizu (1994) as having undergone carbonatite metasomatism, as their clinopyroxenes have unusual U-shaped REE patterns (Fig. 3). Lenoir *et al.* (1997) suggested that xenoliths from Forez (NE Massif Central) show carbonatite metasomatism, as they are extremely depleted in Zr and show fractionation of Zr from Hf (Fig. 4). None of the ultramafic massifs show any evidence for carbonatite metasomatism, although if the carbonatites are related to the Tertiary mantle plume beneath western Europe, then it is not surprising that this component is missing from the massifs, which were physically detached from the SCLM before the plume became active. Tertiary carbonatites from southern Germany have isotopic characteristics similar to those of Neogene alkali basalts from western Europe (Wilson & Downes, 1991) and hence interaction with carbonatites would be isotopically indistinguishable from interaction with Neogene silicate melts.

Effect of interaction with subduction-related hydrous fluids

Enrichment by subduction-related hydrous fluids would be anticipated in areas of Alpine and/or Hercynian subduction. Hydrous fluids do not carry REE to any great extent and hence should not cause changes in REE or Nd isotopes within the SCLM. Mineralogical

consequences of hydration of the lithosphere could include formation of amphibole and phlogopite. Major elements other than the alkalis are unlikely to be strongly affected by hydrous fluids, but some increase in highly soluble LILE (Rb, Sr, K, Ba) at the expense of insoluble high field strength elements (HFSE; Zr, Hf, Nb) would be expected. This type of enrichment is seen in phlogopite metasomatized xenoliths from Hessen (Germany). However, caution should be exercised, as these same xenoliths have low Ti/Eu contents, considered to be a signature of carbonatite enrichment. Xenoliths from Vulture (central Italy) also come from a region of SCLM situated above a recent subduction zone. They contain phlogopite and have very similar trace element patterns to those of the Hessian xenoliths (Fig. 4) but again show petrographic evidence for carbonatite metasomatism (Jones *et al.*, 2000). The Finero massif is also considered to be subduction fluxed (Zanetti *et al.*, 1999) and Finero peridotites also have high Rb, Sr and Ba contents and low Nb, Zr and Hf (Hartmann & Wedepohl, 1993).

Subduction-zone fluids are derived largely from seawater-altered basalts and sea-floor sediments and may affect the Sr isotopic ratio of the SCLM into which they migrate. Increases in $^{87}Sr/^{86}Sr$ in xenoliths from Hungary may relate to fluids released by Neogene subduction (Downes *et al.*, 1992; Rosenbaum *et al.*, 1997). Xenoliths from Vulture (Italy) also show high $^{87}Sr/^{86}Sr$ ratios, as do peridotites from the Finero massif (Fig. 8). Subduction-related fluids may also have distinctive Pb isotope signatures if the sea-floor sediments have been derived from old crustal rocks, giving rise to high $^{207}Pb/^{204}Pb$ ratios at a given value of $^{206}Pb/^{204}Pb$. This causes a trend on a $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram at a high angle to the NHRL (Fig. 10). Such effects have been observed in SCLM xenoliths from Hungary (Rosenbaum *et al.*, 1997), whereas xenoliths from regions where subduction has not occurred since Hercyanian times (e.g. Massif Central) do not show this effect.

Origin of pyroxenites

There are two schools of thought concerning the origin of layered pyroxenites found in ultramafic massifs and as xenoliths: (1) they are formed by passage of silicate magmas through the lithosphere (see above), or (2) they are remnants of oceanic crust, subducted into the asthenosphere and then streaked out by mantle convection until they were incorporated into the lithosphere, together with their host peridotites (Allègre & Turcotte, 1986). Figure 2 shows that pyroxenites are very variable in their bulk-rock compositions, with high CaO contents. This feature could be due either to accumulation from mafic magma, in which Ca-rich clinopyroxene and Ca-rich plagioclase were cumulus phases, or from seawater alteration of oceanic gabbros. Graphite in pyroxenite layers may be a product of subducted organic material. Many pyroxenites in ultramafic massifs are LREE depleted (Fig. 3), consistent with being a product of MORB; however, positive and negative Eu anomalies are not observed in their bulk-rock compositions, as would be expected if fractionation or accumulation of plagioclase had occurred in magma chambers in the ocean crust. Several spinel pyroxenite xenoliths from Romania are LREE enriched; magmas in equilibrium with these pyroxenites must also have been LREE enriched, unlike MORB (Vaselli *et al.*, 1995).

Radiogenic ingrowth in pyroxenite layers contributes to some of the isotope heterogeneity observed in ultramafic massifs. Both the highest and lowest present-day Nd isotope ratios occur in layered pyroxenites from the Lherz massif (Downes *et al.*, 1991); the Hf isotope composition of pyroxenites from the Beni Bousera massif (Morocco) is likewise very variable (Blichert-Toft *et al.*, 1999). Thus, if pyroxenite layers represent oceanic crust, they must be very ancient to have produced such isotopic heterogeneity in response to relatively minor fractionation of Sm/Nd ratios. In a study of Os-isotopes in the ultramafic massifs of Baldissero, Lanzo and Lherz, Roy-Barman *et al.* (1996) concluded that the pyroxenites were possibly formed 1 by ago, and may represent old delaminated ocean crust. However, the question of how such subducted material could be incorporated into sub-continental lithosphere which was detached from the convecting asthenosphere at 2 Ga remains to be answered. It is probable that there are two different origins for pyroxenites in the upper mantle.

SOME UNRESOLVED QUESTIONS

Despite the significant increase in our understanding of mantle processes over the past few decades, many important questions remain unresolved. For example, is the enrichment seen in mantle xenoliths a result of sampling bias in the areas of Neogene magmatism or would such a signature be found in the present-day SCLM in areas remote from magmatic activity? Could some of the plume signature be a result of very shallow level contamination of the xenoliths by their host mafic alkaline magmas, and could the petrographic evidence of carbonatite interaction be a result of the evolution of carbonate fluids from alkaline melts during or even after eruption? Can the existence of former subduction zones be mapped by using mantle xenoliths that show phlogopite enrichment? Do lithospheric terrane boundaries exist in the mantle and can they be mapped by careful examination of mantle xenolith compositions, as has been suggested by Lenoir *et al.* (2001)? These questions can

be addressed by a combination of small-scale petrographic and mineralogical studies on individual xenoliths and massifs, and regional-scale analysis of variations in mantle compositions and relationships with geodynamic features.

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