Metasomatized Mantle Xenoliths as a Record of the Lithospheric Mantle Evolution of the Northern Edge of the Ahaggar Swell, In Teria (Algeria)

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ABSTRACT

Mantle-derived xenoliths hosted by melilitite lavas from In Teria (Ahaggar, SE Algeria) include garnet and spinel peridotites, pyroxenite and phlogopite megacrysts. The spinel and garnet peridotites record an early deformation event, which formed porphyroclastic microstructures and olivine crystal preferred orientations, followed by static infiltration of hydrous alkaline melts. This metasomatic stage (stage 1) is characterized by the crystallization of phlogopite in the garnet and spinel peridotites, amphibole in the spinel peridotites and clinopyroxene in the garnet peridotite, which record chemical equilibration with an alkaline silicate melt. These early events were largely overprinted by carbonatitic metasomatism (stage 2), which is observed only in the spinel peridotites. Spinel peridotite major and trace element compositions, as well as the compositions of newly formed minerals, are characteristic of interaction with carbonate melt, associated with strong enrichment in incompatible trace elements in clinopyroxene. This second stage was followed by crystallization of pyroxenites (stage 3) in vein conduits, probably segregated from alkaline melts. We propose a scenario in which the different metasomatic imprints record successive stages of interaction between lithospheric mantle and sublithospheric melts throughout the Cenozoic. In Sr–Nd isotope space, the host melilitites and several xenoliths are clustered and plot close to the HIMU mantle end-member. However, some peridotite xenoliths are shifted towards more radiogenic 87Sr/86Sr values. In 207Pb/204Pb–206Pb/204Pb and 208Pb/204Pb–06Pb/204Pb space the In Teria samples define a relatively large domain characterized by high 206Pb/204Pb and 208Pb/204Pb, consistent with a contribution of an HIMU component, considered to represent a sublithospheric signature. The highest 87Sr/86Sr values are comparable with those ascribed to the EM1 mantle end-member, representing the signature of the lower continental lithosphere, and are probably inherited from the pre-metasomatic lithospheric mantle beneath In Teria. Numerical modelling of porous percolation of melt of sublithospheric origin through an EM1-like lithospheric mantle protolith reproduces the In Teria peridotite compositions, using moderately sub-chondritic Sr/Nd values for the peridotite (e.g. In Teria garnet peridotite) and moderately super-chondritic Sr/Nd values in the melt (approximately ocean island basalt values). A few spinel peridotites require a component characterized by
plate since the Paleocene (Piromallo & Faccenna, 2004). Central Atlantic province and dragged by the African material emplaced during the Cretaceous beneath the North Africa has been attributed to plume-head spot tracks and the persistence of a plume signature been related to marginal effects of Ahaggar plume activity (Davies & Bunge, 2006), or to more recent mantle flow beneath North Africa (e.g. Ebinger & Sleep, 1998; Piromallo & Faccenna, 2004; Forte et al., 2010).

To understand the above observations we have conducted a detailed study of the In Teria peridotite and pyroxenite xenoliths, phlogopite megacrysts, and their host melilitite. Microstructural analysis based on the measurement of the crystallographic preferred orientations (CPO) of constituent minerals allows constraints to be made on the deformation processes affecting the lithospheric mantle beneath In Teria and on the relative timing of deformation and metasomatism, as well as determination of possible toptaxial relationships during melt–rock reaction. New isotopic data (Sr–Nd–Pb) for the mantle xenoliths and host lavas, together with major and trace element analyses of whole-rocks and constituent minerals, allow definition of the composition of the Ahaggar mantle lithosphere and description of its evolution in response to different metasomatic events. The Sr–Nd–Pb isotopic data indicate that the percolating melt has a mantle affinity with an EM1 signature, which has not yet been observed from the Ahaggar, nor in Morocco or Libya (Beccaluva et al., 2007, 2008; Raffone et al., 2010; Natali et al., 2013). Together these data allow discussion of the nature and evolution of the lithospheric mantle at the northern edge of the Ahaggar Swell and the relationship of this evolution to the different geodynamic events that affected the region.

**INTRODUCTION**

North Africa is characterized by the existence of several volcanic swells (e.g. Ahaggar, Tibesti, Darfur) associated with regional topographic uplift, which suggests interactions between lithospheric architecture and mantle plume activity. However, the relationships between volcanism and large-scale mantle circulation beneath North Africa remain poorly understood. Several models have proposed a plume origin for the recent volcanism and topography in North Africa, but they differ on the proposed location of the plumes. Other models highlight the importance of pre-existing lithosphere architecture. Ebinger & Sleep (1998) proposed the feeding of northern African hotspots (Ahaggar, Tibesti and Darfur) by asthenospheric material rising from the Afar plume and channelled along zones of thinned lithosphere. Another view is the development of splash plumes and channelled along zones of thinned lithosphere. The formation of splash plumes in North Africa might be due to the accumulation of subducted oceanic slabs during closure of the Tethys Ocean in the Mesozoic (Lustrino & Wilson, 2007). Another model proposes that the volcanism is due to local mantle upwellings linked to reactivation of older lithospheric structures during the early stages of collision of the African plate with Eurasia (Liègeois et al., 2005). Edge-driven convection at craton boundaries was also proposed to explain some of the recent volcanism in North Africa (King & Anderson, 1995, 1998; Missenard & Cadoux, 2012). The absence of hot-spot tracks and the persistence of a plume signature beneath North Africa has been attributed to plume-head material emplaced during the Cretaceous beneath the Central Atlantic province and dragged by the African plate since the Paleocene (Piromallo & Faccenna, 2004).

This study focuses on the Ahaggar (Hoggar) Massif, which is the most extensively studied volcanic swell in North Africa, and more specifically on its northeastern termination. The In Teria district, near Illizi, is located at the northeastern edge of the Ahaggar volcanic swell and at the western edge of the Saharan Metacraton, to the south of the Sahara Basins (Fig. 1a and b). The lithospheric mantle beneath In Teria was probably extensively modified during major lithospheric–asthenospheric interactions that produced tholeiitic and alkaline magmatism in central Ahaggar (Aït-Hamou et al., 2000) (Fig. 1b). The occurrence of garnet peridotite among the suite of In Teria mantle xenoliths (Dautria et al., 1992) suggests that relics of older thicker lithosphere may have been preserved beneath In Teria (Beccaluva et al., 2007). Paradoxically, the In Teria district belongs to a domain of anomalously high heat flow (>100 mW m–2) (Takherist & Lesquer, 1989; Lesquer et al., 1990), and the exhumed mantle xenoliths display evidence of extensive interaction with silica-undersaturated silicate and carbonate melts (Dautria et al., 1992). Both the high heat flow and the metasomatism might be related to the so-called effect of Ahaggar plume activity (Davies & Bunge, 2006), or to more recent mantle flow beneath North Africa (e.g. Ebinger & Sleep, 1998; Piromallo & Faccenna, 2004; Forte et al., 2010).

**GEOLOGICAL SETTING**

The main structures of the Ahaggar Massif are inherited from the Pan-African orogeny, which resulted from
continental collision between the West African Craton and an East African block between 580 and 760 Ma (e.g. Caby et al., 1981; Lesquer et al., 1984; Bertrand et al., 1986; Liégeois et al., 1994). Most of the uplift occurred from Late Eocene to Quaternary times (Guiraud & Bellion, 1995; Guiraud et al., 2005), but fission-track age data suggest that the regional uplift of the Ahaggar basement started at least as early as the Cretaceous (Khaldi et al., 2006). Volcanic activity in the Ahaggar Massif began during the late Eocene (c. 34 Ma; Remy, 1959; Rossi et al., 1979; Aït-Hamou et al., 2000), peaked during the Miocene and continued episodically through the late Pliocene and into the late Quaternary (Girod, 1971).

The In Teria district is situated within an anomalously ‘hot’ east-west-trending zone characterized by heat flow higher than 100 mW m⁻², which has been attributed to anomalously high temperatures in the underlying asthenosphere (Fig. 1a; Takherist & Lesquer, 1989; Lesquer et al., 1990). Significantly lower heat flow is recorded to the south of the study area in the Ahaggar Massif (53 mW m⁻² on average), suggesting a stable lithosphere 100 km in thickness (Lesquer et al., 1989). However, free-air gravity data led Crough (1981) to suggest that the Central Ahaggar is underlain by anomalously low-density upper mantle. Similar conclusions have been reached based on seismic tomography models (Ayadi et al., 2000; Sebai et al., 2006; Zhao, 2007; Begg et al., 2009).

The In Teria volcanic district includes about 20 diatreme craters, each no larger than 2 km in diameter (Megartsi, 1972). These cross-cut horizontal Carboniferous strata and correspond to pipe-like structures. The ejecta are cemented by calcite and include lava blocks up to 1 m in size, as well as lapilli, olivine and phlogopite megacrysts, and various mantle and crustal xenoliths. Ultramafic xenoliths were collected by J.-M. Dautria from the ejecta within a 50 km² area. The studied samples come from two diatremes, No. 14 and No. 17 (Fig. 1c); No. 14 is the larger structure (2 km in diameter), formed of hydromagmatic tuff containing pyroxenite and mafic granulite xenoliths, phlogopite megacrysts, peridotite xenoliths, and the only garnet peridotite xenolith sampled in the In Teria district. The second diatreme (No. 17) is 500 m in diameter with a 5 m high ejecta rim characterized by block and ash deposits. The ejecta include granite, pyroxenite and peridotite xenoliths, phlogopite megacrysts, and pyroclastic breccia. All xenolith samples analyzed measured between 5 and 12 cm. Pyroxenite xenoliths represent the majority of the xenolith population. Most pyroxenite xenoliths have a black coating and may contain centimetre-sized amphibole crystals. The xenoliths are embedded in a porphyritic melilitite host lava (Megartsi, 1972), which formed at 1000–1200°C at a pressure greater than 20 kbar (Girod, 1971; Kechid & Megartsi, 2005). Although the age of the volcanic activity is not well established,
the freshness of some tuff rings suggests a Quaternary age (Conrad, 1969; Megarits, 1972).

ANALYTICAL METHODS
Crystal preferred orientations
Olivine, pyroxene, and amphibole CPO were measured in 19 samples by indexing of electron back-scattered diffraction (EBSD) patterns using a JEOL JSM 5600 scanning electron microscope at Géosciences Montpellier. Data collection, indexing and analysis of electron backscatter diffraction patterns (EBSP) were performed with the CHANNEL 5.10 software by Oxford Instruments. For each sample, we obtained crystallographic orientation maps covering most of the thin section (maps are usually 20 mm × 35 mm) with a regular grid step ranging from 35 to 55 μm depending on the mean grain size. Average indexation rates were ~70%. The EBSD data were noise-reduced using a ‘wildspike’ correction to remove isolated erroneous points, followed by a six-neighbour zero solution extrapolation following standard procedures (Prior et al., 2002; Bestmann & Prior, 2003). At each of these steps, the resulting orientation maps were compared with band contrast maps to ensure that the data treatment did not compromise the data. The crystallographic orientation data are presented as lower hemisphere pole figures prepared using the Pf_ctf software of D. Mainprice (http://www.gm.univ-montp2.fr/PERSO/mainprice/W_data/CareWare_Unicef_Programs/). Average Euler angles for each grain (one point per grain) were used to avoid over-representation of large grains in thin section. (The complete dataset is available in Supplementary Data as Fig. A1; the supplementary data are available for downloading at http://www.petrology.oxfordjournals.org.) In those samples in which the foliation could not be identified, the CPO data are presented with the maximum concentration of [100] axes of olivine parallel to x (east–west) and the maximum concentration of [010] axes of olivine parallel to z (north–south) for easy comparison between the samples.

Whole-rock and mineral compositions
Whole-rock major element compositions were analyzed at Nancy SARM by inductively coupled plasma optical emission spectroscopy (ICP-OES). Whole-rock trace element compositions were determined by inductively coupled plasma mass spectrometry (ICP-MS) using a Quadrupole VG-PQ2 system at Montpellier University (France) following the procedure described by Ionov et al. (1992). Major elements in minerals were analyzed by electron microprobe using the CAMECA-SX100 at the Microsonde Sud facility (Université de Montpellier II) equipped with five wavelength-dispersive spectrometers (Supplementary Data Table A2). Operating conditions comprised an acceleration voltage of 20 kV and a 10 nA beam current. K and Na were counted for 20 s with a 10 s background and the other elements were counted for 30 s with a 15 s background. Mineral trace element data were obtained by laser ablation (LA)-ICP-MS using the facility available at the ATEF platform (Supplementary Data Table A3). The GeoLas Q+ laser system used at Géosciences Montpellier is an Excimer (Compex 102) operating in the deep UV (193 nm). Ablations were performed in a pure He atmosphere (~0.6 l min⁻¹) using a beam diameter ranging between 30 and 120 μm, with an energy density of c. 15 × 10⁻³ J cm⁻². The LA platform is linked to an extended range Element 2 ICP-MS system operated in low-resolution mode at 1350 W. The ICP-MS system was daily tuned to maximum sensitivity while keeping oxide production to its minimum level (ThO/Th ≤ 1%). The NIST612 glass was used as external standard (Pearce et al., 1997) and SiO₂ or CaO contents determined by electron probe for each mineral were used as an internal standard. Data were processed using the GLITTER software package (Van Achterbergh et al., 2001).

Sr–Nd–Pb isotopes
Sr, Nd and Pb isotopic compositions were obtained from 400 mg of powdered whole-rock samples. After a step of leaching with 6N HCl at 80°C for 30 min and three cycles of rinsing with purified milli-Q H₂O, samples were then dissolved for 48 h on a hot plate in a mixture of HF 48% (1:2) and 13 N HNO₃ (1:2). After evaporation to dryness (120°C), 2.5 ml of 13 N HNO₃ was added to the residue and kept at about 100°C for 24–48 h and evaporated. A last cycle of dissolution and evaporation was performed with 2 ml of 13 N HNO₃ (1:2). For Pb separation, after complete evaporation, ~100 μl of 8 N HBr was added to the sample and kept at 90°C for 5 h before another complete evaporation. The chemical separation of Pb was carried out using 50 μl of anion exchange resin (AG1X8, 200–400 mesh) using 0.5 N HBr and 6 N HCl as eluants. Strontium isotopes were separated using Sr Eichrom resin (Pin et al., 1994). For nine samples we conducted a stronger leaching to compare Sr isotope compositions. The first step of leaching was with 2.5 N HCl at 95°C for 1 h 30 min followed by three cycles of rinsing with purified Milli-Q H₂O; samples were then leached with 6 N HCl at 95°C for 1 h 30 min followed by three cycles of rinsing with purified Milli-Q H₂O. After leaching an identical dissolution procedure was applied. Neodymium isotopes were separated during three steps: first a rare earth element (REE) separation using AG50WX12 cation exchange resin, followed by two steps of Nd purification using HDEHP columns. Total blank contents for Pb, Sr and Nd were less than 30, 60 and 30 pg, respectively. Lead and neodymium isotopic compositions were measured on a VG Plasma 54 and a Nu 500 multicollector (MC)-ICP-MS system at the Ecole Normale Supérieure de Lyon. Pb isotopic compositions were measured with an external precision of c. 100–150 ppm for ²⁰⁶Pb/²⁰⁴Pb using the TI normalization method.
described by White et al. (2000). Every two samples were bracketed between the NIST 981 standard. For Nd isotopic measurements, every two samples were bracketed between the Lyon 'in-house' Nd standard (a 500 ppb dilution of the JMC commercial solution of ICP-MS Nd standard, batch 801149A; Luais et al., 1997), with an average 143Nd/144Nd = 0.710254 ± 0.000006 (2σ) (n = 20). Strontium isotopic compositions were measured on a Finnigan Triton TI mass spectrometer at the ‘Laboratoire de Géochimie GIS’ of Nîmes. Results on NBS 987 Sr standards yielded a mean value of 87Sr/86Sr = 0.710254 ± 1 (2σ) (n = 18). Results on NBS 981 Pb standards yielded a mean value of 206Pb/204Pb = 15.4874 ± 4 (2σ), 207Pb/204Pb = 15.9331 ± 4 (2σ) (n = 14). Given the inferred Quaternary age of the In Teria magmatic district (Conrad, 1969; Megartsi, 1972), no age correction was performed on the measured isotopic ratios.

PETROGRAPHY AND TEXTURE OF THE XENOLITHS

Garnet peridotite
Sample Int-14-7 is a phlogopite-bearing garnet lherzolite with a weakly foliated porphyroclastic texture (Table 1, Fig. 2a). Although the petrography of this sample was previously described by Dautria et al. (1992), we summarize here the key characteristics of this rock. It has a centimeter-scale compositional layering marked by alternate olivine and pyroxene enrichment. Garnet has a spheroidal shape with a diameter between 4 and 8 mm. Brown Al-spinel locally forms coronas around garnet (Fig. 2a). Primary clinopyroxene (Cpx-I) is located around garnet or is associated with orthopyroxene; it has a grain size up to 1 mm. Orthopyroxene porphyroclasts (Opx-I) have grain sizes smaller than 1 mm. Olivine (Ol-I) has rounded or elongated shapes (aspect ratios 4:1) with a grain size up to 1 mm. Elongated olivine grains show low-angle boundaries or undulose extinction perpendicular to the grain elongation, which is in general parallel to the layering. The primary assemblage composed of olivine, orthopyroxene, clinopyroxene and garnet is overprinted by metasomatic phlogopite (Phl-II). The occurrence of Phl-II is irregular; it occurs either around garnet or as an interstitial phase with a poikilitic texture (up to 2.5 mm in size) (Fig. 2b).

Spinell peridotites
Lherzolite, harzburgite and Ol-websterite
In Teria spinel peridotites are characterized by a porphyroclastic texture formed by primary minerals that
are overprinted by two metasomatic assemblages. The primary assemblage (I) comprises olivine, orthopyroxene and spinel ± clinopyroxene. Erem-2 is the only spinel-bearing sample containing primary clinopyroxene (Cpx-I) (Table 1, Fig. 2c). Cpx-I grains from sample Erem-2 are anhedral, up to 0.5 cm wide, and contain thin exsolution lamellae of orthopyroxene.

The orthopyroxenes (Opx-I) in all peridotites are 2–5 mm in size and have irregular shapes (Fig. 2d and e). Olivine (Ol-I) occurs as elongated crystals (up to 8 mm long) that define the foliation (Fig. 2f). Olivine grains display undulose extinction and subgrains oriented perpendicular to the grain elongation (Fig. 2f). Spinel, commonly replaced by chromite, is up to

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**Fig. 2.** Photomicrographs in plane-polarized light (PPL) and cross-polarized light (XPL) and backscattered electron (BSE) images illustrating the typical microtextures of the In Teria peridotites. (a) Garnet peridotite with a layer rich in orthopyroxene; PPL. (b) Garnet with a corona essentially composed of spinel and clinopyroxene; PPL. Phlogopite (Phl-II) is poikilitic. (c) Xenolith Erem-2 with primary clinopyroxene (Cpx-I), orthopyroxene (Opx-I) and olivine (Ol-I), showing reaction zones with former melt pockets; PPL. (d) Primary orthopyroxene (Opx-I) in IL-17-8 is partially replaced by amphibole-II and phlogopite-II; PPL. (e) Opx-I in the IL-17-4 is ‘dismembered’ and partially replaced by crystallized melt pockets; PPL. (f) Spinel peridotite IL-17-20 characterized by tabular olivine crystals and subgrain boundaries; XPL. Dashed line indicates foliation. (g) Coarse granular spinel peridotite IL-17-10 displays foliation marked by spinel and amphibole + phlogopite. The green areas correspond to former melt pockets with clinopyroxene (Cpx-III) enrichment; PPL. Dashed line indicates foliation. (h) BSE image of IL-17-10 showing a melt pocket composed of clinopyroxene (Cpx-III), olivine (Ol-III), interstitial glass and holes (black areas) generated during polishing. (i) Melt pocket with partially oriented Cpx-III in IL-17-11; PPL. It should be noted that amphibole is replaced by Cpx-III. (j) BSE image of an amphibole–clinopyroxene symplectite in IL-17-16. grt, garnet; cpx, clinopyroxene, opx, orthopyroxene, ol, olivine, amph, amphibole, gl, glass.
400 µm in size and has a skeletal shape, but with alignment parallel to the foliation when the latter is observed (Fig. 2g). The Ol-websterite, Int-14-6, has a texture and grain size similar to those of the lherzolites and harzburgites, but is characterized by a higher proportion of orthopyroxene (~50%, Table 1).

The second mineral assemblage (II), previously referred to as ‘metasomatic hydrous mineral phases’ by Dautria et al. (1992), corresponds to the occurrence of phlogopite (Phl-II) and amphibole (Amph-II) in variable proportions (Table 1, Fig. 2b–d). Amphibole and phlogopite often replace Opx-I or spinel (Fig. 2d and g). The spinel–phlogopite–amphibole aggregates are aligned with the foliation, but the elongation of single amphibole and phlogopite grains (up to 1 mm) is not always parallel to the foliation (Fig. 2g) and these minerals do not display intracrystalline deformation features.

The third assemblage (III) consists of reaction aggregates or patches (possible former melt-pockets), which are present in variable proportions in the spinel peridotites (10–40% of the sample volume; Table 1). These reaction aggregates are composed of microgranular clinopyroxene, olivine, chromite, Al-spinel, sulphides, K-feldspar and glass (Fig. 2c and e–i) and were interpreted by Dautria et al. (1992) as the products of carbonatitic melt percolation through the peridotite. Dark green clinopyroxene (Cpx-III) and associated minerals partially replace primary orthopyroxene (Fig. 2e) and secondary amphibole (Fig. 2i); Cpx-III is euhedral and has a grain size between 50 and 200 µm (Fig. 2h and i); it may show elongated shapes parallel to orthopyroxene or amphibole cleavages (Fig. 2i). Ol-III crystallizes as neoblasts with subhedral shapes and displays a grain size between 50 and 150 µm (Fig. 2h).

**Wehrlite and Ol-clinopyroxenite**

Wehrlite and Ol-clinopyroxenite form the Cpx-III-rich peridotite xenoliths that are common at In Teria (5–10% of collected xenoliths). Three samples were studied: Il-17-21, Int-17-14 and Il-17-16. Wehrliites Int-17-14 and Il-17-21 contain primary olivine (7 and 20%, respectively; Table 1) with a grain size up to 1 mm and anhedral shapes. The clinopyroxene is similar to Cpx-III observed in the spinel peridotites. It has euhedral shapes with a grain size between 50 and 200 µm. Ol-III grain size varies between 50 and 150 µm. Amphibole and phlogopite are occasionally found and have anhedral shapes with grain sizes of up to 0.3 mm.

The Ol-clinopyroxenite Il-17-16 is a peculiar sample, characterized by the lack of primary olivine (Ol-I). This sample contains clinopyroxene megacrysts with numerous thin orthopyroxene exsolution lamellae and clinopyroxene–amphibole symplectites (Fig. 2j). The clinopyroxene megacrysts are embedded in a fine-grained reaction matrix, similar to that which composes...
the reaction patches in the spinel peridotites, characterized by a large proportion of Cpx-III and Ol-III (Table 1).

Pyroxenites
Pyroxenite xenoliths are more abundant (~70%) than peridotite xenoliths at In Teria (~20%; Kechid & Megartsi, 2005). They are characterized by a cumulative texture, the absence of olivine and garnet, and no reaction aggregates or patches like those observed in the spinel peridotite xenoliths. In some samples subtle layering is defined by variations in grain size. They are mainly clinopyroxenite, composed of clinopyroxene and phlogopite, with or without amphibole; websterites are rare. Clinopyroxenes are zoned xenomorphic crystals or patches like those observed in the spinel peridotite xenoliths. In some samples subtle layering is defined by variations in grain size. They are mainly clinopyroxenite, composed of clinopyroxene and phlogopite, with or without amphibole; websterites are rare. Clinopyroxenes are zoned xenomorphic crystals, measuring between 3 and 4 mm in size. Phlogopite has a grain size of 1–3 mm. It is generally interstitial and associated with clinopyroxene. When present, the amphibole is millimeter-sized and is interstitial between clinopyroxene and phlogopite.

CRYSTAL PREFERRED ORIENTATIONS
Crystal preferred orientations (CPO) of olivine, orthopyroxene, clinopyroxene and amphibole were analyzed in the garnet peridotite and 18 spinel peridotite xenoliths. Results for the garnet peridotite and six spinel peridotites are shown in Fig. 3. The complete dataset is available as Supplementary Material Fig. A1.

The garnet peridotite (Int-14-7) has an olivine CPO characterized by a strong maximum orientation of [010] axes close to the pole of the foliation (z) (maximum density MD ≈ 7 86, Fig. 3a). The [001] and [100] axes form a girdle roughly parallel to the foliation plane (x). Both axes have weak maxima subparallel to each other, but the [001] maximum is stronger than the [100] maximum. This distribution characterizes a fiber-[010] olivine CPO. Orthopyroxene and clinopyroxene have similar CPO with a strong alignment of [001] axes in the foliation plane and subparallel to the olivine [100] maxima. The [100] and [010] axes of both pyroxenes form incomplete girdles in the foliation plane. The [010] axes form a maximum at high angle to the foliation.

Most spinel peridotites show a weak preferred orientation of olivine grains with a point concentration of [100] axes parallel to the foliation plane, probably marking the lineation. The olivine [010] axes form either a maximum perpendicular to the foliation plane or a girdle normal to the [100] maximum (Fig. 3b–d and Supplementary Material Fig. A1). In all spinel peridotites olivine [001] axes are more dispersed than the other two, but they form a weak maximum in the foliation plane perpendicular to the [100] maximum.

In spinel peridotites, orthopyroxene has a weak CPO, which is characterized nevertheless by a weak [001] maximum subparallel to the olivine [100] maximum (e.g., II-17-17 and Int-17-10; Fig. 3b). The multiple maxima present in the orthopyroxene CPO are probably due to over-representation of some former orthopyroxene porphyroclasts, which were corroded and dislocated during subsequent metasomatism, now being present as separate grains. This effect was corrected for as much as possible during data reduction.

Clinopyroxene and especially amphibole CPO in spinel peridotites are strongly dispersed, but distributions of the [100], [010] and [001] axes are consistently parallel between these minerals. They display two types of microstructural relationships with olivine and orthopyroxene from the primary mineral assemblage: (1) Type A, where despite the much more dispersed CPO, the clinopyroxene and amphibole [001] axes are parallel to the [001] maximum of orthopyroxene and to the [100] maximum of olivine (Fig. 3b); (2) Type B, where clinopyroxene and amphibole display similar CPO orientations, which are discordant to orthopyroxene CPO (Fig. 3d). Six samples show an intermediate organization between Type A and B (Fig. 3c and Supplementary Data Fig. A1). The high dispersion of the CPO, as well as the Type B and intermediate textural relationships are consistent with petrographic observations, which imply that all clinopyroxene and amphibole in the spinel peridotites are products of metasomatism.

WHOLE-ROCK CHEMISTRY
The previous study of Dautria et al. (1992) on In Teria xenoliths reported major and trace element data for peridotite and pyroxenite xenoliths, as well as host melilitites samples. Major elements were analyzed by atomic absorption and trace elements (Hf, Ta, Th, U, and REE) were determined by neutron activation (samples marked with asterisks in Tables 2 and 3). In this study we report major element analyses for 10 new samples by ICP-OES and new trace element analyses for all samples by solution ICP-MS. Overall, 17 peridotite and 11 pyroxenite xenoliths, one phlogopite megacryst and five host melilitites were analysed in this study for major and trace elements. Analytical results are given in Tables 2 and 3 and illustrated in Figs 4 and 5.

Garnet peridotite
The garnet peridotite has low MgO (35.7 wt%) and high Al₂O₃ and Fe₂O₃ contents (3.3 and 11.4 wt%, respectively; Fig. 4, Table 2). Such a composition overlaps the field of off-cratonic mantle xenoliths (Fig. 4a; Canil, 2004). However, comparison with the mantle melting array shows that this garnet peridotite does not represent a melting residue (Fig. 4b and c). This sample has a relatively flat REE pattern with (La/Lu)ₚ = 2.81 (the sub-
Fig. 3. Crystal preferred orientations (CPO) of olivine, orthopyroxene, clinopyroxene and amphibole from In Teria garnet peridotite xenolith (a) and spinel peridotite xenoliths (b–d). Spinel peridotites are organized in three groups defined by the relations between the orthopyroxene, clinopyroxene and amphibole CPO. (b) Type A, where olivine, orthopyroxene, clinopyroxene and amphibole are coherent; (c) intermediate type between Types A and B, characterized by a partial correlation between the orthopyroxene and the clinopyroxene and amphibole CPO; (d) Type B, where olivine and orthopyroxene CPO are coherent but uncorrelated to the clinopyroxene and amphibole CPO. Two samples of each type are illustrated. (The complete dataset is given in Supplementary Data, Fig. A1.) Equal-area lower hemisphere stereographic projections in the structural reference frame. N is the number of grains analyzed. Contours are at 0.5 multiples of uniform distribution intervals. Orthopyroxene and amphibole stereoplots were contoured when more than 100 grains could be measured in the thin section. MD is maximum density; PfJ is a scalar measure of the strength of the axis orientation.
Fig. 3. Continued.
Table 2: Major (wt %) whole-rock analyses of peridotite, pyroxenite xenoliths and melilitites

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<tr>
<td>SiO₂</td>
<td>46.1 44.7 43.0 43.0 44.9 44.9 44.0 44.2 41.3 42.6 44.4 43.5 45.13 49.8 49.3 50.4 48.1</td>
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<tr>
<td>Al₂O₃</td>
<td>3.28 2.15 1.77 2.13 2.40 2.02 2.04 1.90 1.37 0.60 2.18 2.08 2.54 0.98 3.19 3.55 5.01</td>
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<tr>
<td>FeO</td>
<td>10.3 7.57 8.38 8.21 7.70 7.45 7.60 7.60 8.51 8.01 7.47 7.96 7.98 8.60 4.88 4.82 6.61</td>
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<tr>
<td>MnO</td>
<td>0.15 0.12 0.15 0.16 0.16 0.13 0.13 0.12 0.15 0.14 0.12 0.13 0.12 0.17 0.11 0.11 0.13</td>
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<tr>
<td>MgO</td>
<td>35.9 41.1 40.6 40.4 39.9 39.4 41.1 42.4 44.2 45.7 42.2 40.4 39.4 37.3 28.1 25.5 21.2</td>
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<tr>
<td>CaO</td>
<td>2.23 2.26 2.74 2.69 2.63 2.89 2.18 1.73 1.82 1.37 1.85 2.93 2.79 1.02 8.50 9.46 12.3</td>
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<tr>
<td>Na₂O</td>
<td>0.27 0.33 1.08 1.21 1.10 1.21 0.69 0.39 0.90 0.49 0.73 1.13 0.86 0.72 2.76 3.14 2.72</td>
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<tr>
<td>K₂O</td>
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<tr>
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<tr>
<td>P₂O₅</td>
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<td>Total</td>
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<tr>
<td>Mg#</td>
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Pyroxenites

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<tr>
<td>MgO</td>
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<tr>
<td>CaO</td>
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<td>TiO₂</td>
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Melilitites

*Major element whole-rock analyses from Dautria et al. (1992) and unpublished data from J.-M. Dautria. (-1-) corresponds to sample numbering from Dautria et al. (1992).
### Table 3: Trace element (ppm) whole-rock analyses of peridotite, pyroxenite xenoliths and melilitites by solution ICP-MS

| Sample numbering as in Table 2. |

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**Spinel peridotite**

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**Pyroxenite**

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**Melilitite**

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Sample numbering as in Table 2.
The whole-rock Ca/Al ratio is highly variable and reaches values as high as 3.1 (sample IL-17-4). The majority of the In Teria lherzolites and harzburgites form a group within the off-cratonic mantle array (Fig. 4a) with compositions comparable with those of the most fertile Cap Verde peridotite xenoliths (Bonadiman et al., 2005) (Fig. 4b and c). Only three samples show distinct compositions. Two samples (Int-17-1 and Int-17-6), are more refractory and plot within or close to the field of Canary Island peridotite xenoliths (Neumann et al., 2004; Fig. 4b and c); the Ol-websterite (sample Int-14-6) has low Al2O3, CaO and MgO contents compared with the other peridotites (Fig. 4).

Lu varies from 0.012 to 0.13 ppm (Table 3), and is positively correlated with Al2O3 and negatively with MgO. Although showing variable REE contents and distributions, all spinel peridotites are more enriched in light REE (LREE) than the garnet peridotite (Fig. 5a and c). On the basis of normalized REE patterns three subgroups can be distinguished (Fig. 5). The first group (group 1) comprises most of the lherzolites and harzburgites and is characterized by almost straight positive slope from heavy REE (HREE) to LREE (Fig. 4c). The other two groups (groups 2 and 3) are distinguished by concave-upwards REE patterns. The second group comprises the lherzolites Erem-2 and Int-15-1; Erem-2 is the sample containing relicts of primary clinopyroxene. These two samples have a roughly flat to slightly middle REE (MREE)-depleted MREE–HREE segment [0.67 < (Eu/Lu)N < 0.74] and are enriched in LREE relative to MREE [(La/Sm)N > 8]. The third group includes the lherzolite samples IL-17-18 and Int-17-13 and the Ol-websterite sample Int-14-6. These samples are distinguished from the second group by positive (Eu/Lu)N ratios, between 1.5 and 3.3, and lesser LREE enrichment [6 < (La/Sm)N < 8]. On Primitive Mantle (PM)-normalized diagrams group 1 peridotites are mostly characterized by negative anomalies of Zr, Hf, and Ti relative to the MREE. Conversely, several samples show positive anomalies of Nb and Ta relative to Th, U, and LREE; the anomalies tend to be marked in samples with lower Th, U, and LREE contents. Except for a positive spike in Th in one sample, group 1 peridotites also tend to be depleted in Th and U relative to Ba and LREE. Group 2 samples show markedly distinct trace element patterns, characterized by a strong negative anomaly in Nb and Ta, whereas Zr, Hf, and Ti do not show significant anomalies. Group 2 samples are also distinguished by selective enrichment of Th, U, Pb, and Sr. In contrast, group 3 is more akin to group 1, particularly in showing negative anomalies of Zr, Hf, and Ti, and lacking substantial anomalies of Nb and Ta.

**Spinel peridotite**

*Lherzolite, harzburgite and Ol-websterite*

Lherzolite, harzburgite and one orthopyroxene-rich peridotite (sample Int-14-6) have Mg# varying from 88-90 to 91.0, whereas Al2O3 and CaO contents vary between 0.98 and 2.54 wt %, and 1.02 and 2.7 wt %, respectively. The whole-rock Ca/Al ratio is highly variable and reaches values as high as 3.1 (sample IL-17-4). The majority of the In Teria lherzolites and harzburgites form a group within the off-cratonic mantle array (Fig. 4a) with compositions comparable with those of the most fertile Cap Verde peridotite xenoliths (Bonadiman et al., 2005) (Fig. 4b and c). Only three samples show distinct compositions. Two samples (Int-17-1 and Int-17-6), are more refractory and plot within or close to the field of Canary Island peridotite xenoliths (Neumann et al., 2004; Fig. 4b and c); the Ol-websterite (sample Int-14-6) has low Al2O3, CaO and MgO contents compared with the other peridotites (Fig. 4).

Lu varies from 0.012 to 0.13 ppm (Table 3), and is positively correlated with Al2O3 and negatively with MgO. Although showing variable REE contents and distributions, all spinel peridotites are more enriched in light REE (LREE) than the garnet peridotite (Fig. 5a and c). On the basis of normalized REE patterns three subgroups can be distinguished (Fig. 5). The first group (group 1) comprises most of the lherzolites and harzburgites and is characterized by almost straight positive slope from heavy REE (HREE) to LREE (Fig. 4c). The other two groups (groups 2 and 3) are distinguished by concave-upwards REE patterns. The second group comprises the lherzolites Erem-2 and Int-15-1; Erem-2 is the sample containing relicts of primary clinopyroxene. These two samples have a roughly flat to slightly middle REE (MREE)-depleted MREE–HREE segment [0.67 < (Eu/Lu)N < 0.74] and are enriched in LREE relative to MREE [(La/Sm)N > 8]. The third group includes the lherzolite samples IL-17-18 and Int-17-13 and the Ol-websterite sample Int-14-6. These samples are distinguished from the second group by positive (Eu/Lu)N ratios, between 1.5 and 3.3, and lesser LREE enrichment [6 < (La/Sm)N < 8]. On Primitive Mantle (PM)-normalized diagrams group 1 peridotites are mostly characterized by negative anomalies of Zr, Hf, and Ti relative to the MREE. Conversely, several samples show positive anomalies of Nb and Ta relative to Th, U, and LREE; the anomalies tend to be marked in samples with lower Th, U, and LREE contents. Except for a positive spike in Th in one sample, group 1 peridotites also tend to be depleted in Th and U relative to Ba and LREE. Group 2 samples show markedly distinct trace element patterns, characterized by a strong negative anomaly in Nb and Ta, whereas Zr, Hf, and Ti do not show significant anomalies. Group 2 samples are also distinguished by selective enrichment of Th, U, Pb, and Sr. In contrast, group 3 is more akin to group 1, particularly in showing negative anomalies of Zr, Hf, and Ti, and lacking substantial anomalies of Nb and Ta.

**Wehrlite and Ol-clinopyroxenites**

The wehrlite and Ol-clinopyroxenites (II-17-21, Int-17-14 and II-17-16) have low bulk-rock MgO contents (21.2–28.1 wt %), relatively high Al2O3 (3.19–5.01 wt %), and elevated CaO contents (8.50–12.3 wt %; Fig. 4a, Table 2).
Their normalized REE and trace element patterns (Fig. 5c and d) are almost indistinguishable from peridotite group 1 patterns, except for higher concentrations of most elements.

**Pyroxenite**

The pyroxenites contain between 6.95 and 11.6 wt % \(\text{Al}_2\text{O}_3\) and between 6.0 and 14.9 wt % \(\text{CaO}\), and their \(\text{MgO}\) ranges between 11.5 and 16.9 wt % (Table 2). They
show convex-upward REE patterns, characterized by positive slopes from the HREE to the MREE \([5.6 < \text{Eu/Lu}]_\text{N} < 7.4\) and a roughly flat LREE segment \([\text{La/Sm}]_\text{N} = 0.77-1.56\); Fig. 5g]). The PM-normalized trace element patterns are distinguished by positive HFSE anomalies in Nb, Ta, and Ti in all samples, and Zr-Hf in two samples. The pyroxenites are also distinctly enriched in Rb and Ba relative to Th and U (Fig. 5h).

**Melilitite**

The compositions of the five analysed host-rock melilites are homogeneous (Table 2). The samples are undersaturated in SiO\(_2\) (~33.5 wt %) and have low total alkali \((K_2O + Na_2O)\) contents varying from 3.0 to 4.7 wt %.

The high Mg \((\text{MgO} = 14.2-15.3\text{ wt %})\), high Ni and Cr contents indicate that these lavas have not been extensively modified by fractional crystallization (Dautria et al., 1992). The chondrite-normalized REE patterns show a smooth positive slope from HREE to LREE \([48 \leq \text{La/Yb}]_\text{N} \leq 70\); Fig. 5a). The PM-normalized trace element patterns are convex upwards, with high but relatively unfractonated abundances of the most incompatible elements (Rb to LREE), except for Pb, which shows a negative anomaly relative to the LREE \([\text{Pb}/\text{Ce}]_\text{N} = 0.28\). Among the HFSE, only Zr and Hf show a noticeable negative anomaly.

**MINERAL CHEMISTRY**

**Major elements**

**Garnet peridotite**

Garnet has a homogeneous, pyrope-rich composition \((\text{Mg}\# = 0.80; \text{Cr}\# = 0.04, \text{Supplementary Data Table A2})\).

Clinopyroxenes have high TiO\(_2\) contents (1.6 wt %) and show some compositional similarities to secondary clinopyroxenes from Manzaz in the Ahaggar district (Beccaluva et al., 2007) and primary clinopyroxenes from Cape Verde (Bonadiman et al., 2005). The orthopyroxene porphyroclasts (Opx-I) are zoned; for instance, \(\text{Al}_2\text{O}_3\) increases from core to rim \((3.10 \text{ to } 4.09 \text{ wt %})\), respectively; \text{Supplementary Data Table A2}), as does Mg\# \((0.87 \text{ to } 0.89)\). Olivine has a homogeneous composition with low Mg\# \(= 0.86\). The phlogopite (Phil-II) has Mg\# \(= 0.86\) and a high TiO\(_2\) content \((5.50 \text{ wt %})\, \text{Supplementary Data Table A2}).

Using garnet-pyroxene geothermo-geobarometry (Nickel & Green, 1985; Brey & Köhler, 1990) core equilibrium temperature and pressure have been estimated at 1050–1100 °C and 2.5–2.7 GPa, respectively (Dautria et al., 1992). However, pyroxene and garnet rim compositions yield temperatures and pressures of 1210–1240 °C and 2.6–2.7 GPa.

**Spinel peridotite**

Lherzolite, harzburgite and Ol-websterite. Primary clinopyroxenes (Cpx-I) from lherzolite Erem-2 have high \(\text{Al}_2\text{O}_3\) \((6.8 \text{ wt %})\) and \(\text{Na}_2\text{O} \((1.74 \text{ wt %})\, \text{Supplementary Data Table A2})\) contents and resemble primary clinopyroxene from the Ahaggar district (Beccaluva et al., 2007) (Fig. 6a). The primary orthopyroxene (Opx-I) Mg\# varies from 0.87 to 0.91, and its \(\text{Al}_2\text{O}_3\) content varies from 0.53 to 4.86 wt %, \(\text{O}-\text{I}\) have Mg\# between 0.89 and 0.94 and low \(\text{CaO}\) contents \((0.2 \text{ wt %})\). In the secondary mineral assemblage (II), the amphibole (Amph-II) is parasitic to edenite with very variable Mg\# \((0.89 \text{ and } 0.96)\) and \(\text{Na}_2\text{O} \((3.30–5.10 \text{ wt %})\) contents, and low contents of \(\text{TiO}_2 \((0.15–0.35 \text{ wt %})\) and \(\text{K}_2\text{O} \((0.06–0.60 \text{ wt %})\). The phlogopite (Phil-II) shows less variability \((\text{Mg}\# = 0.92–0.93)\) and is \(\text{TiO}_2\)-poor \((0.27–0.28 \text{ wt %})\) and \(\text{K}_2\text{O}\)-rich \((6.8–7.0 \text{ wt %})\). Clinopyroxenes from the third assemblage (Cpx-III, Fig. 6a) have Mg\# between 0.92 and 0.93, high \(\text{CaO} \((18.5–23.5 \text{ wt %})\), low but variable \(\text{Al}_2\text{O}_3 \((0.83–3.09 \text{ wt %})\), and low \(\text{TiO}_2\) \((0.08–0.25 \text{ wt %})\) contents. These compositions are similar to those of finely disseminated secondary clinopyroxene replacing orthopyroxene in Cape Verde mantle xenoliths (Bonadiman et al., 2005) and secondary large poikilitic clinopyroxene from the Canary Islands (Neumann et al., 2002; Fig. 6a). Ol-III displays a large variation in Mg\# \((0.89–0.94)\), encompassing the composition of Ol-I; however, Ol-III has on average a more Mg-rich composition \((\text{Mg}\# = 0.92)\).
than Ol-I (Mg# 0.90, Fig. 6b). Olivine neoblasts within reaction patches have high CaO contents (0.01–0.03 wt %) compared with Ol-I CaO contents (0.01–0.03 wt %). The glass analysed within the melt pockets is siliceous (up to 68 wt % SiO$_2$) and peraluminous (up to 25 wt % Al$_2$O$_3$) in composition, and has low MgO, FeO and CaO contents, resembling feldspar in its composition. This glass may have formed at the expense of feldspar (Dautria et al., 1992). Its composition is similar to that observed in many mantle xenolith suites (e.g. Frey & Green, 1974; Ionov et al., 1993b; Coltorti et al., 2000).

Using the Brey & Köhler (1990) geothermometer calibration on Opx-I and Cpx-I (sample Erem-2), Dautria et al. (1992) obtained a core equilibrium temperature between 770 and 870 °C and a rim temperature increasing up to 1080 °C. According to the regional geotherm (Lesquer et al., 1990), such temperatures would yield a pressure range between 1.6 and 2.1 GPa. However the absence of garnet indicates that pressure should be lower than 1.9 GPa (Green & Ringwood, 1967; Wallace & Green, 1988).

**Wehrlite and Ol-clinopyroxenite.** Primary olivine in wehrlite has the same composition as Ol-I in spinel peridotite with Mg# 0.89–0.90 and 0.01–0.03 wt % CaO (Supplementary Data Table A2). The amphibole (Amph-II) has a similar composition to Amph-II in the spinel peridotites with Mg# 0.89, Na$_2$O of 3.87 wt %, and low TiO$_2$ (0.27 wt %) and K$_2$O (0.35 wt %) contents.

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**Fig. 7.** Primitive Mantle normalized trace element patterns of clinopyroxene and phlogopite in the garnet peridotite, spinel peridotites and pyroxenite xenoliths from In Teria. Values are normalized to the Primitive Mantle of McDonough & Sun (1995). Each pattern represents an average of several analyses in different grains (three on average). Error bars represent one standard deviation and are reported for representative samples. In (f) the phlogopite megacryst was analysed by solution ICP-MS.
Fig. 8. Primitive Mantle normalized trace element patterns of orthopyroxene from garnet and spinel peridotites and amphibole from spinel peridotites and pyroxenites. Values are normalized to the Primitive Mantle (PM) values of McDonough & Sun (1995). Each pattern represents the average of several analyses (three on average). Error bars represent one standard deviation and are reported for representative samples.

Orthopyroxene
Garnet peridotite (a)

Amphibole
Spinel peridotite (c)

Spinel peridotite (b)

Pyroxenite (d)

Ol-clinopyroxenite Il-17-16 contains clinopyroxene phenocrysts and clinopyroxene in symplectites, which have low Al₂O₃ (0.80–1.13 wt %) and high CaO (22.8–23.4 wt %) contents (Fig. 2; Supplementary Data Table A2). Cpx-III forming the fine-grained reaction matrix displays a similar composition to that in other spinel peridotites. This sample contains olivine neoblasts (Ol-III) with a high CaO content (0.23 wt %) as observed in Ol-III from spinel peridotites, but with lower Mg# (0.89) (Fig. 6b).

Pyroxenite
The clinopyroxene porphyroclast cores are homogeneous in composition with Mg# from 0.77 to 0.78. Phlogopite has Mg# varying between 0.73 and 0.74, and TiO₂ between 2.77 and 6.18 wt %, which is the typical composition for upper mantle phlogopite (e.g. Delaney et al., 1980; Ionov et al., 1997). Two distinct phlogopite zoning trends from core to rim were observed: the first shows a decrease in TiO₂ (6.18–5.58 wt %) associated with an increase in Al₂O₃, Na₂O and Mg# (e.g. sample Int-17-89-100), and the second an increase in TiO₂ from 2.77 to 4.76 wt % (sample Il-17-27). The amphiboles have Mg# 0.70–0.72 and high K₂O contents from 2.34 to 2.56 wt %. Temperature estimates for crystallization of the pyroxenites range between 1100 and 1180°C (Boissière & Megartsi, 1982).

Trace elements
Garnet peridotite
Garnet has a high HREE content (Lu_N = 9.27) and a strong LREE depletion relative to the HREE and MREE [(La/Lu)_N = 0.002; Supplementary Data Table A3]. Clinopyroxene has a convex PM-normalized REE pattern, in which the MREE are enriched relative to the HREE [(Eu/Lu)_N = 0.3; Fig. 7a] and LREE [(La/Sm)_N = 0.56]. It is characterized by relatively unfractio-nated abundances of LILE except for Pb, which shows a negative anomaly [(Pb/Ce)_N = 0.11]. The HSFE show barely detectable positive anomalies [e.g. (Nb/ Ce)_N = 0.16]. Orthopyroxene REE patterns display a relatively flat HREE to MREE segment [(Eu/Lu)_N = 1.01] and a negative slope from MREE to LREE [(La/Sm)_N = 0.13] with well-marked Zr–Hf and Ti anomalies [(Zr/Sm)_N < 3; Fig. 8a]. Olivine displays enrichment from Dy to Lu [(Dy/Lu)_N = 0.19]. Phlogopite is characterized by low REE abundances [(Lu_N = 0.010; La_N = 0.028], but prominent positive HFSE anomalies [(Nb/La)_N = 422; (Zr/Sm)_N = 278; Fig. 7d] and strong Cs, Rb and Ba enrichment, but low U–Th content (Supplementary Data Table A3).

Spinel peridotite
Lherzolite, harzburgite and Ol-websterite. The primary clinopyroxene from sample Erem-2 displays a LREE-
Table 4: Pb-Sr–Nd isotopic compositions of whole-rock garnet and spinel peridotites, pyroxenites, melilitites and phlogopite megacrysts

<table>
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<tr>
<th>Garnet peridotite</th>
<th>Spinel peridotite</th>
<th>Erm-2</th>
<th>Il-17-4</th>
<th>Il-17-5</th>
<th>Il-17-13</th>
<th>Il-17-14</th>
<th>Il-17-18</th>
<th>Int-15-1</th>
<th>Int-17-1</th>
<th>Int-17-1r</th>
<th>Int-17-6</th>
<th>Int-17-7</th>
<th>Int-17-12</th>
<th>Int-17-13</th>
<th>Int-14-6</th>
<th>Ol-clinopyroxenite</th>
<th>Wehrlite</th>
<th>Int-17-21</th>
<th>Pyroxenite</th>
<th>Int-17-22</th>
<th>Int-17-24</th>
<th>Int-17-7</th>
<th>Int-17-100</th>
<th>Melilitite</th>
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<th>Phlogopite</th>
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Like clinopyroxene, amphibole exhibits positive Zr–Hf [(Zr/Sm)$_N$ = 2.33–4.03] and Ti anomalies. The opposite was observed for the peridotite amphibole and clinopyroxene. Amphibole showing both RB–Ba enrichment and depletion has been observed in the same pyroxene sample (Fig. 8d). Phlogopites have low REE contents and show an overall positive slope from MREE to LREE. Their extended trace elements patterns are similar to those described for peridotite xenoliths (especially for the garnet peridotite; Fig. 7d and f) and are marked by prominent positive HFSE anomalies (e.g. 160 < (Nb/ La)$_N$ < 6300), 27 < (Zr/Sm)$_N$ < 237) and high Cs, Rb, Ba and Sr abundances (e.g. Sr and Ba up to ~170 ppm and ~2287 ppm, respectively). In contrast to phlogopite in spinel peridotites, that in the pyroxenites shows low U and Th abundances, as in the garnet peridotites. A phlogopite megacryst analyzed by solution ICP-MS shows a similar pattern (Fig. 7f; Supplementary Data Table A2), except for the HREE. We note that $D^{\text{phl/cpx}}$(Na,Ta) and $D^{\text{phl/cpx}}$(Zr, Hf) are in the range of 0.6–2 and 0.04–0.09, respectively, as reported by Grégoire et al. (2000) and Moine et al. (2001) (and references therein). Similarly, the partitioning of Rb (9–22), Ba (7–20), and Sr (0.1–0.5) between these two phases is also broadly similar to that reported in previous studies.

**Sr, Nd AND Pb ISOTOPIC DATA**

Whole-rock isotopic compositions of mantle xenoliths, phlogopite megacrysts and host melilitites are reported in Table 4. The $^{87}\text{Sr} / ^{86}\text{Sr}$ of the In Teria xenoliths varies widely from 0.70327 to 0.70503 and does not correlate with $^{143}\text{Nd} / ^{144}\text{Nd}$ (Fig. 9a). Two spinel peridotites (samples Erem-2 and Int-15-1) have higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values of 51309 and 51310, respectively (Table 4). In Sr–Nd isotope space, 11 mantle xenoliths show higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values at nearly constant $^{143}\text{Nd} / ^{144}\text{Nd}$ (Fig. 9a). Two spinel peridotites (samples Erem-2 and Int-15-1) have higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values of 0.51309 and 0.51310, respectively (Table 4). In Sr–Nd isotope space, 11 mantle xenoliths show higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values at nearly constant $^{143}\text{Nd} / ^{144}\text{Nd}$ (Fig. 9a). Two spinel peridotites (samples Erem-2 and Int-15-1) have higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values of 0.51309 and 0.51310, respectively (Table 4). In Sr–Nd isotope space, 11 mantle xenoliths show higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values at nearly constant $^{143}\text{Nd} / ^{144}\text{Nd}$ (Fig. 9a). Two spinel peridotites (samples Erem-2 and Int-15-1) have higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values of 0.51309 and 0.51310, respectively (Table 4). In Sr–Nd isotope space, 11 mantle xenoliths show higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values at nearly constant $^{143}\text{Nd} / ^{144}\text{Nd}$ (Fig. 9a). Two spinel peridotites (samples Erem-2 and Int-15-1) have higher $^{143}\text{Nd} / ^{144}\text{Nd}$ values of 0.51309 and 0.51310, respectively (Table 4).
Fig. 9. Variation of $^{143}$Nd/$^{144}$Nd vs $^{87}$Sr/$^{86}$Sr (a), $^{207}$Pb/$^{204}$Pb vs $^{206}$Pb/$^{204}$Pb (b) and $^{208}$Pb/$^{204}$Pb vs $^{206}$Pb/$^{204}$Pb (c) for in Teria melilitites, pyroxenites, spinel peridotites, the garnet peridotite (Int-14-7) and phlogopite megacrysts. The following fields are shown for comparison: Ahaggar alkali basalts (Allègre et al., 1981; Dupuy et al., 1993); Ahaggar tholeiitic basalts (Aït-Hamou, 2000; Aït-Hamou
Manzaz and Libyan peridotites (Beccaluva et al., 1993; Beccaluva et al., 2007, 2010; Natali et al., 2013) show significantly lower 206Pb/204Pb and 208Pb/204Pb values (Beccaluva et al., 2007, 2008). The garnet-bearing peridotite and four spinel peridotite xenoliths from the Manzaz district of central Ahaggar (Aït-Hamou et al., 2000). Compared with In Teria whole-rock xenoliths, some of the separated clinopyroxenes from the xenoliths from Libya and Ahaggar (Manzaz district) show significantly higher 142Nd/144Nd values (Beccaluva et al., 2007, 2008).

The Pb isotopic compositions of the In Terra xenoliths are in the range of 18.36–19.98 for 206Pb/204Pb, 15.60–15.69 for 207Pb/204Pb, and 38.17–39.51 for 208Pb/204Pb (Table 4). In 207Pb/204Pb–206Pb/204Pb and 208Pb/204Pb–206Pb/204Pb space, the samples define a relatively large range (Fig. 9b and c). Seven spinel peridotites, all the pyroxenites and the melilitites are characterized by high 206Pb/204Pb and 208Pb/204Pb ratios trending towards the HIMU end-member. This group of samples overlaps the fields defined by alkali basalts from Ahaggar and Libya (Allègre et al., 1981; Dupuy et al., 1993; Beccaluva et al., 2008) and the fields of Manzaz and Libyan peridotites (Beccaluva et al., 2007, 2008). The garnet-bearing peridotite and four spinel peridotites show significantly lower 206Pb/204Pb and 208Pb/204Pb values (<18.76 and <38.50, respectively) associated with high 207Pb/204Pb ratios (>15.60) (Fig. 9b and c) and partly overlap the field of Libyan peridotites (Beccaluva et al., 2008). The three phlogopite megacrysts display intermediate 206Pb/204Pb values (18.68–19.09), a large variation in 208Pb/204Pb and 207Pb/204Pb, and plot between the two xenolith groups identified previously (Fig. 9b and c).

**DISCUSSION**

**Mantle deformation**

The association of porphyroclastic microstructures, marked by elongated olivine crystals with common subgrains, and the well-developed olivine and orthopyroxene CPO indicate that the studied xenoliths are deformed. In the garnet peridotite, CPO are characterized by the parallelism of clinopyroxene and orthopyroxene [001] maxima and the olivine [100] maxima (Fig. 3a). This suggests that all the minerals accommodated the same deformation, consistent with petrographic observations indicating that the clinopyroxene is primary (Fig. 2a). In all spinel peridotites, the parallelism of olivine [100] and orthopyroxene [001] maxima implies that the primary phases (Ol-I and

\[ ^{87}\text{Sr} / ^{86}\text{Sr} \] values are not a secondary weathering effect. The three phlogopites analysed yield low 142Nd/144Nd values (+1.4 < εNd < +3) at relatively high \[ ^{187}\text{Sr} / ^{86}\text{Sr} \] (0.7038–0.7042) and plot in the field of tholeiitic basalts from central Ahaggar (Aït-Hamou et al., 2000).

Compared with In Teria whole-rock xenoliths, some of the separated clinopyroxenes from the xenoliths from Libya and Ahaggar (Manzaz district) show significantly higher 142Nd/144Nd values (Beccaluva et al., 2007, 2008).

Chondrite-normalized (Ce/Yb)N vs. (Ti/Eu)N for (a) clinopyroxene and (b) orthopyroxene (Opx-I) from In Teria garnet peridotite, spinel peridotites and pyroxenites. These diagrams are based on those proposed by Cottrell (1999). Chondrite-normalizing values are from McDonough & Sun (1995). Opx-I, primary clinopyroxene; Cpx-III, clinopyroxene from the third metasomatic assemblage in the spinel peridotites (stage 3).

Opx-I have recorded the same deformation event. In contrast to garnet peridotite, in most spinel lherzolites, the lack of correlation between clinopyroxene and amphibole CPO, together with petrographic observations, suggests that clinopyroxene and amphibole are secondary and that this metasomatic addition post-dated the main deformation episode. An orthopyroxene crystallographic control on the growth of clinopyroxene and amphibole might explain the partial correlation between clinopyroxene and amphibole CPO and the olivine and orthopyroxene CPO in the Type A spinel peridotites (Fig. 3b). An implication of these observations is that prior to the metasomatic event, the shallow lithospheric mantle beneath In Teria had a refractory harzburgitic composition, whereas deeper levels had more fertile compositions.

Analysis of the olivine CPO patterns also single out the garnet peridotite. The latter has a fiber-[010] olivine (Fig. 3a), which implies simultaneous activation of the [100][010] and [001][010] slip systems or a
tholeiites, transitional basalts, basanites and nephelinites from dominant activation of the (010)[100] slip system, sug-

rhombic or fiber-[100] symmetries, consistent with Sidamo, Ethiopia (Bedini, 1994); orogenic lherzolite massifs (Lherz, France, Bodinier

lished data); the intrusive pyroxenite–carbonatite body of Orberger Ronda, Spain, Garrido & Bodinier, 1999); ophiolites (Thailand, some of the In Teria pyroxenites and the amphibole-bearing 1989) should be noted. Shaded lava fields: (1) South African al-
to carbonatitic melts from Morocco (Wagner

Teria xenoliths, as follows. Based on petrographic evidence and geochemical data, Metasomatic imprints

2. The second stage (stage 2) of metasomatism is akin to the percolation of carbonatitic melt and has been well documented by Dautria et al. (1992). It is re-

stricted to the spinel peridotites, where it is responsible for the crystallization of Cr-rich diopside at the expense of orthopyroxene and amphibole. It results

in the formation of microgranular aggregates (for-

mer melt pockets) composed of secondary clino-

pyroxene, olivine, Al- and Cr-spinel, sulphides, K-feldspar and glass, which extensively replace the primary phase assemblages (Fig. 2h and i). Most of the major and trace element characteristics of these samples and their minerals are typical of mantle metasomatism by carbonate melt (Woolley & Kempe, 1989; Yaxley et al., 1991; Ionov et al., 1993a; Hauri & Hart, 1994; Blundy & Dalton, 2000; Chakhmouradian, 2006). These characteristics not-

ably include a strong enrichment of highly incompatible trace elements in whole-rocks and pyroxenes (Opx-I and Cpx-III), negative HFSE (Ta, Zr, Hf and Ti) anomalies in Cpx-III, and negative Zr and Hf anoma-

lies in Opx-I (Figs 5 and 7). OI-III have high Mg# and are enriched in CaO compared with OI-I, and Cpx-III are enriched in Cr (Fig. 6). Moreover, secondary clino.pyroxene (Cpx-III) and coexisting orthopyrox-

ene (Opx-I) show evolutionary trends in Fig. 10 marked by a strong increase in (Ce/Nb)N at relatively low (Ti/Eu)N, confirming interaction with a carbonatitic melt. The possibility that the melt pockets were formed by interaction with the host lava during the exhumation of the peridotite xenoliths is excluded because only spinel peridotites show this interaction (i.e. it is not observed in the garnet peri-

dotite and pyroxenites).

3. The metasomatic history of the In Teria xenolith suite concludes with the crystallization of pyroxen-

ites (stage 3). These rocks, which are predominant among the xenolith suite, crystallized in the spinel stability field, lack the carbonate-melt imprint (stage 2) observed in all spinel peridotites and resemble the amphibole-pyroxenite veins found in Pyrenean peridotites (Bodinier et al., 1987, 2004) with respect to their major and trace element characteristics (Figs 6 and 11). Trace element similarities between the In Teria and Pyrenean pyroxenites include

transpressional deformation (Tommasi et al., 1999, 2000). However, the strong [001] maximum of the pyr-

oxene CPO is not consistent with transpressional de-

formation. We suggest therefore that this olivine CPO records deformation at high stress or under high-pres-

sure conditions (e.g. Jung & Karato, 2001; Mainprice et al., 2005; Demouchy et al., 2013). The spinel perido-

tites, on the other hand, show olivine CPO with ortho-

rhombic or fiber-[100] symmetries, consistent with dominant activation of the (010)[100] slip system, sug-


gest deformation at low-stress, high-temperature, low-pressure conditions (Tommasi et al., 2000).

Metasomatic imprints

Based on petrographic evidence and geochemical data, three metasomatic imprints are recognized in the In Teria xenoliths, as follows:

1. The garnet and spinel peridotites were first affected by crystallization of amphibole and/or phlogopite at the expense of orthopyroxene, a feature that was ascribed by Dautria et al. (1992) to reactive porous percolation of a hydrous alkaline melt. Although it shows a lesser extent of modal metasomatism (crystallization of phlogopite; Fig. 2b), the garnet

peridotite (sample Int-14-7) is characterized by a con-

vex-upward normalized REE pattern in clinopyroxene (Fig. 7a), which suggests equilibration with a LREE-

enriched melt (Irving, 1980; Irving & Frey, 1984; Bodinier et al., 1987). Moderate LREE enrichment without significant negative HFSE anomalies (or with positive anomalies) in clinopyroxene (or whole-rocks) is widely considered as evidence for mantle meta-

somatism involving alkaline silicate melts (e.g. Downes, 2001; Bodinier et al., 2004; Powell et al., 2004; Kaeser et al., 2006). Together the observations on the garnet and spinel peridotites suggest that the lithospheric mantle beneath In Teria was pervasively metasomatized by hydrous alkaline melts.

Fig. 11. Variation of CaO vs SiO2 for the In Teria pyroxenite xenoliths and host melilitites, compared with literature data for pyroxenites and basaltic to carbonatitic mantle melts (shaded fields). The pyroxenite dataset includes samples from the fol-

lowing xenolith suites: Hawai (Frey, 1980); Australia (Griffin & O’Reilly, 1986; O’Reilly & Griffin, 1987; O’Reilly et al., 1989); Sidamo, Ethiopia (Bedini, 1984); o rogenic lherzolite massifs (Lherz, France, Bodinier et al., 1987, 1990; McPhail et al., 1990; Ronda, Spain, Garrido & Bodinier, 1999); ophiolites (Thailand, Orberger et al., 1995; New Caledonia, J. L. Bodinier, unpublished data); the intrusive pyroxene–carbonatite body of Tamazeght, Morocco (Marks et al., 2008). The overlap between some of the In Teria pyroxenites and the amphibole-bearing pyroxenite veins from Lherz (Bodinier et al., 1987; Bodinier, 1989) should be noted. Shaded lava fields: (1) South African al-
akaline basalts and olivine melilitites (Janney et al., 2002); (2) theo
dite veins, compared with literature data for mantle pyroxenites and amphibole pyroxenite veins from Lherz (Bodinier et al., 1987; Bodinier, 1989) should be noted. Shaded lava fields: (1) South African al-
akaline basalts and olivine melilitites (Janney et al., 2002); (2) theo
dite veins, compared with literature data for mantle pyroxenites

2. The second stage (stage 2) of metasomatism is akin to the percolation of carbonatitic melt and has been well documented by Dautria et al. (1992). It is re-

stricted to the spinel peridotites, where it is responsible for the crystallization of Cr-rich diopside at the expense of orthopyroxene and amphibole. It results

in the formation of microgranular aggregates (for-

mer melt pockets) composed of secondary clino-

pyroxene, olivine, Al- and Cr-spinel, sulphides, K-feldspar and glass, which extensively replace the primary phase assemblages (Fig. 2h and i). Most of the major and trace element characteristics of these samples and their minerals are typical of mantle metasomatism by carbonate melt (Woolley & Kempe, 1989; Yaxley et al., 1991; Ionov et al., 1993a; Hauri & Hart, 1994; Blundy & Dalton, 2000; Chakhmouradian, 2006). These characteristics not-

ably include a strong enrichment of highly incompatible trace elements in whole-rocks and pyroxenes (Opx-I and Cpx-III), negative HFSE (Ta, Zr, Hf and Ti) anomalies in Cpx-III, and negative Zr and Hf anoma-

lies in Opx-I (Figs 5 and 7). OI-III have high Mg# and are enriched in CaO compared with OI-I, and Cpx-III are enriched in Cr (Fig. 6). Moreover, secondary clino.pyroxene (Cpx-III) and coexisting orthopyrox-

ene (Opx-I) show evolutionary trends in Fig. 10 marked by a strong increase in (Ce/Nb)N at relatively low (Ti/Eu)N, confirming interaction with a carbonatitic melt. The possibility that the melt pockets were formed by interaction with the host lava during the exhumation of the peridotite xenoliths is excluded because only spinel peridotites show this interaction (i.e. it is not observed in the garnet peri-

dotite and pyroxenites).

3. The metasomatic history of the In Teria xenolith suite concludes with the crystallization of pyroxen-

ites (stage 3). These rocks, which are predominant among the xenolith suite, crystallized in the spinel stability field, lack the carbonate-melt imprint (stage 2) observed in all spinel peridotites and resemble the amphibole-pyroxenite veins found in Pyrenean peridotites (Bodinier et al., 1987, 2004) with respect to their major and trace element characteristics (Figs 6 and 11). Trace element similarities between the In Teria and Pyrenean pyroxenites include
convex-upward normalized REE patterns and positive HFSE anomalies in whole-rocks and amphibole (Nb, Ta, Zr, Hf, Ti) and, to a lesser degree, in clinopyroxene (Zr, Hf, Ta). In Fig. 10a, the pyroxenites show a trend defined by a strong increase in the (Ti/Eu)N ratio of clinopyroxene correlated with a moderate increase in (Ce/Yb)N. This evolutionary trend diverges markedly from that defined by the secondary clinopyroxene in the spinel peridotites. The In Teria pyroxenites probably represent crystal segregates from hydrous alkaline melts crystallized in vein conduits—rather than in large intrusions.

Metasomatism: several events or multi-stage melt–rock interaction

Microstructural observations indicate that all metasomatic events postdate the last deformation event, recorded by olivine in both garnet and spinel peridotites, that affected the lithospheric mantle in this region. The alignment of the plagiogpate-amphibole aggregates in the foliation, but the undeformed character of the crystals and the dispersion of crystallographic axes of Amph-II and Cpx-III (e.g. Type B in Fig. 3d) indicate that although melt infiltration postdated the deformation, it was at least partially controlled by the pre-existing structure of the peridotites. A stronger structural control of the pre-existing microstructure on the reaction products is also indicated by the similarity of Amph-II and Cpx-III CPO in Type A spinel lherzolites (Fig. 3b), which suggests totopaxial growth of at least part of the metasomatic products.

Our geochemical data show striking similarities between the mineral compositions of peridotites affected solely by stage 1 metasomatism and those of stage 3 pyroxenites. This may be exemplified by comparing the trace element signatures of clinopyroxene and plagioclase in the garnet peridotite sample and in the pyroxenites (Figs 7 and 10a). Moreover, our data show the clear geochemical affinity of the pyroxenite xenoliths (stage 3) with the melilitite host lavas. As illustrated in Fig. 11, the In Teria pyroxenites differ from other mantle pyroxenites (including, to some degree, the Pyrenean veins) in their lower SiO2 contents and as such are compositionally akin to strongly undersaturated and carbonatic magmas. The equilibrium melt calculated from the trace element composition of clinopyroxene in pyroxenite is comparable with the host melilitites—as well as with the garnet peridotite equilibrium melt [Fig. 12; partition coefficients (Kp) from Hart & Dunn (1993)]. The only significant differences are the positive HFSE anomalies (particularly for Zr and Hf) that are observed in the pyroxenite equilibrium melt (but not in the garnet peridotite equilibrium melt). These anomalies probably reflect the inadequacy of available experimental clinopyroxene/melt Kp values for clinopyroxenes crystallized from strongly SiO2-undersaturated alkaline melts.

The microstructural observations and the geochemical affinity of the xenoliths with the melilitite host lavas favour a scenario whereby the different metasomatic imprints record successive stages of interaction between lithospheric mantle and sublithospheric melts. The evolution from stage 1 to stage 2 (silicate/hydrous to carbonate melt) may result from a differentiation process involving melt ingress down a lithospheric thermal gradient associated with melt–rock reactions at decreasing melt mass (Bedini et al., 1997; Bodinier et al., 2004).
Gradual solidification of the melt, also referred to as ‘percolative fractional crystallization’ (Harte et al., 1993), would be responsible for marked changes in its composition, culminating in the individualization of volatile-rich and/or carbonated small melt fractions that are expected to be strongly enriched in highly incompatible elements (McKenzie, 1989). Equilibrium melts calculated from the trace element composition of In Teria Cpx-III are enriched in incompatible elements and display negative HFSE anomalies in their PM-normalized patterns [Fig. 12; Klemme et al. (1995)]. This signature is similar to that obtained by numerical simulations of reactive porous flow at decreasing melt mass (Bedini et al., 1997; Ionov et al., 2002a).

Trace element modelling predicts that the combination of chromatographic effects related to melt transport and ‘source’ effects related to melt–rock reactions results in spatial decoupling between melt–rock reactions and their trace element expression (Godard et al., 1995). This may be responsible for transient depletion of several incompatible elements in porous-flow systems, owing to buffering of the percolating melt by the depleted peridotite protolith (see, e.g. fig. 13 of Bodinier et al., 2008). In contrast to other pyroxenes from the In Teria xenolith suite, Cpx-I and Opx-I of the spinel peridotite Erem-2 are depleted in LREE (Figs 7, 8, 10 and 12). The unique trace element signature of this sample may represent a remnant of the pre-metasomatic, depleted composition of the lithospheric mantle beneath In Teria. Alternatively, it may also be explained by transient buffering of the percolating melt by the peridotite protolith.

Melt compositions appear to have varied with percolative differentiation from siliceous and hydrous, but locally impoverished in highly incompatible trace elements relative to the infiltrated melt (e.g. sample Erem-2 equilibrium melt with cpx-I; Fig. 12), to carbonated and strongly enriched in highly incompatible trace elements, but with negative HFSE anomalies (other spinel peridotites equilibrium melt in Fig. 12). Clinopyroxene trace element compositions in the garnet peridotite and pyroxenites (Fig. 7) and the similarity between the equilibrium melt for the garnet peridotite and the host melilitite (Fig. 12) indicate that the garnet peridotite has interacted with a slightly evolved silicate melt. This may have a deeper origin than the spinel peridotites and/or a
The garnet peridotite records higher pressure-temperature (P–T) conditions than the spinel peridotites and is devoid of any carbonatitic metasomatic overprint, suggesting a deeper origin. The multi-stage metasomatic event that affected the In Teria xenoliths was probably coupled with a transient thermal event; that is, heating of subcontinental lithosphere followed by thermal relaxation. Heating is recorded by the compositional zoning of the primary minerals, indicating a core to rim increase of equilibrium temperatures from 1050–1100 °C to 1210–1240 °C for the garnet peridotite sample Int-14-7 and from 770–870 °C to about 1080 °C for the spinel peridotite sample Erem-2 (Fig. 13). Combined with pressure estimates, these rim temperatures define a rather steep thermal gradient roughly consistent with the high heat flux measured in the In Salah–Illizi district (>100 mW m−2; Lesquer et al., 1990). The rim P–T estimates lie in the interval between the hydrous and anhydrous peridotite solidus (Fig. 13) and may therefore record lithospheric P–T conditions during the first metasomatic stage. The intergranular porous flow of hydrous alkaline melt inferred for this stage requires ambient temperatures in excess of (at least) the hydrous peridotite solidus. The observation that phlogopite is the only metasomatic mineral formed in the garnet peridotite is consistent with the high ‘rim’ temperature of this sample, well beyond the stability field of pargasite (Niida & Green, 1999; Fig. 13). Compared with amphibole, phlogopite is stable at higher temperatures (Wendlandt & Egglar, 1980; Mengel & Green, 1989). Conversely, the predominance of pargasite over phlogopite in spinel peridotites would reflect the lower temperatures of the shallower lithosphere during stage 1, overlapping the amphibole stability field (Fig. 13).

In a scenario in which carbonate melt is produced by reactive differentiation of percolating silicate melt down a thermal gradient, the silicate and carbonate melt metasomatism are coeval but disconnected in space (Bedini et al., 1997; Bodinier et al., 2004). During conductive heating of the lithosphere, of an asthenospheric origin, silicate melt metasomatism would occur in deeper and/or hotter lithosphere whereas carbonate melt metasomatism would affect shallower and/or colder domains (Fig. 14a). Such an arrangement of metasomatic aureoles related to a thermal gradient has been observed on a small scale in vein wall-rocks within the Lherz peridotite (Bodinier et al., 2004). In the In Teria suite, the lack of a carbonate melt imprint in the deeper garnet peridotite, whereas it is pervasive in the shallower spinel peridotites, provides evidence for a similar, thermally controlled arrangement of metasomatic domains at lithospheric scale.

In the proposed scenario, overprinting of stage 1 metasomatism in spinel peridotites by stage 2 metasomatism is explained by downwards subsidence of the metasomatic domain owing to lithospheric thermal relaxation (Fig. 14b). The lack of any carbonate-melt imprint in the In Teria pyroxenites, in contrast to the spinel...
peridotites, which are extensively metasomatized, suggests that the pyroxenite veins were late stage. The emplacement of a dense vein network—judging from the predominance of pyroxenites within the xenolith suite—might be associated with further lithospheric cooling favouring extensive hydraulic fracturing (Fig. 14c). Based on their similarity to phlogopite within the pyroxenites, the phlogopite megacrysts may represent deep crystal segregates from the infiltrating alkali silicate melt (Fig. 7).

Based on the time constraints necessary to allow successive heating and thermal relaxation of the lithospheric mantle, the metasomatic evolution proposed in our model probably lasted for the whole Cenozoic era. Stage 1 heating was possibly related to Eocene tholeiitic magmatism, for which a lithospheric contribution has been suggested (Maza et al., 1998), implying substantial heating. The resemblance of In Teria phlogopite megacrysts to Ahaggar tholeiitic volcanism in terms of their Nd–Sr isotope composition (Fig. 9) lends support to this hypothesis and suggests that the phlogopites are related to stage 1. The suggested subsequent evolution would have lasted until the recent melilitite volcanism, based on the observation that stage 3 pyroxenites probably represent mantle segregates from melilitite magmas. The relationship between the suggested lithospheric thermal evolution and the higher heat flow observed in the In Teria region, compared with the rest of the Ahaggar Massif (Lesquer et al., 1989), is poorly understood. The presence of garnet peridotites in a region of higher heat flow is paradoxical. The origin of the high heat flux is probably related to the situation of In Teria at the southern border the Sahara Basins (Fig. 1), characterized by a regional-scale heat flow anomaly (Takherist & Lesquer, 1989; Lesquer et al., 1990), rather than to the mantle upwelling beneath Ahaggar.

**Origin of $^{87}$Sr enrichment**

Among the In Teria xenoliths, the enrichment in $^{87}$Sr without any concomitant decrease of $^{143}$Nd/$^{144}$Nd is restricted to the spinel peridotites and one pyroxenite (sample Int-17-100). This enrichment is not observed in the host melilitites, which show a narrow range of isotopic compositions comparable with those of the Ahaggar alkaline basalts (Fig. 9). Together with the lack of correlation between $^{87}$Sr/$^{86}$Sr and Sr concentration, this does not support a hypothesis of late-stage introduction of radiogenic Sr into the studied samples. The enrichment in $^{87}$Sr is not related to the degree of metasomatism either, nor to the metasomatic enrichment in incompatible trace elements. Although extensively re-equilibrated with alkali-silicate and carbonate melts, respectively, the garnet peridotite and the Ol-clinopyroxenite (sample Il-17-21) are not significantly enriched in radiogenic Sr ($^{87}$Sr/$^{86}$Sr < 0.7035; Fig. 9a). Conversely, the spinel peridotite Erem-2, which is the least metasomatized sample with LREE-depleted pyroxenes (Fig. 10), and the phlogopite megacrysts are relatively enriched in radiogenic Sr.
Fig. 16. Variation of $^{87}$Sr/$^{86}$Sr vs $^{143}$Nd/$^{144}$Nd for the studied in Teria melilitites, pyroxenites and peridotite xenoliths, and phlogopite megacrysts, compared with results of one-dimensional percolation–diffusion modelling showing the effect of compositional variations in Sr and Nd (see text). The short-dashed line represents a reference experiment calculated with Primitive Mantle (PM) Sr–Nd contents as the protolith (McDonough & Sun, 1995) and the average Sr–Nd contents of the studied melilitites for the infiltrated melt. The long-dashed line represents a mixing curve calculated with the same Nd–Sr contents. (a–d) illustrate the effects of varying the following: (a) the Sr–Nd concentrations in the peridotite protolith, from 0 to 3, the average Sr–Nd contents in the melts; (b) the Sr–Nd concentrations in the melt, from 0 to 3, to 3 in the melt, from 0 to 3, to 3 in the melt; (c) the Sr/Nd ratios in the peridotite protolith and (d) in the melt, from 10 to 20. The other parameter values are the same as in Fig. 15 and are given in Supplementary Data Table A5. The symbols for the In Teria samples and the data sources for the Ahaggar basalt fields and mantle EM1 and HIMU end-members are the same as in Fig. 9.

$^{87}$Sr/$^{86}$Sr > 0.704 and > 0.7038, respectively). Elevated $^{87}$Sr/$^{86}$Sr values found in a few limestone-hosted, Ca-rich carbonatites from Tamazert (Morocco) are clearly related to crustal contamination (Bouabdellah et al., 2010). To test the origin of high $^{87}$Sr/$^{86}$Sr values in the In Teria xenoliths we performed a stronger leaching on seven spinel peridotites, the garnet peridotite and one pyroxenite (Table 4). The new $^{87}$Sr/$^{86}$Sr compositions for most samples are less radiogenic and closer to HIMU than previously (Supplementary Data Fig. 4). These results suggest that the high $^{87}$Sr/$^{86}$Sr values could be related to the crystallization of interstitial micro-phases from small-volume melts during the waning stages of metasomatism. Detailed studies of trace element distribution in metasomatized mantle xenoliths have shown that a significant proportion of the whole-rock budget of large ion lithophile elements (LILE, including Sr) is hosted by such intergranular components (Bedini & Bodinier, 1999). Recently, Kourim et al. (2014) have revealed the existence of such a LILE-enriched interstitial component in mantle xenoliths from southwestern Ahaggar. Chromatographic theory predicts that downstream in a reactive percolation column (i.e. at very low melt/rock ratio), the isotopic signature of the percolating melt may be decoupled from its trace element signature: the melt may be isotopically equilibrated with the peridotite protolith although strongly enriched in highly incompatible elements as a result of reactions at decreasing melt mass (Bodinier et al., 2004). Owing to their ability to percolate as small-volume melts through relatively cold peridotite (McKenzie, 1989) and precipitate LILE-enriched micro-phases, carbonate melts such as those invoked for the In Teria stage 2 metasomatism are probably the best candidates to generate such enrichments. This may explain why $^{87}$Sr enrichment is observed in xenolith suites that have experienced carbonate-melt metasomatism (e.g. In Teria and the Canary Islands; Neumann et al., 2002, 2015).
The high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios recorded by the In Teria xenoliths are most probably inherited from the pre-metasomatic composition of the lithospheric mantle. The highest values are comparable with the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios ascribed to the EM1 mantle end-member, a composition that is considered to represent the signature of the lower continental lithosphere (Hawkesworth et al., 1986; Mahoney et al., 1991; Milner & Le Roex, 1996).

**Fig. 17.** Variation of $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ for the studied In Teria melilitites, pyroxenite and peridotite xenoliths, and phlogopite megacrysts, compared with results of the one-dimensional percolation–diffusion modelling showing the effect of varying the $(\text{Pb/Sr})_{\text{melt}}/(\text{Pb/Sr})_{\text{peridotite}}$ ratios (see text). The model runs a–g in (a) were calculated with $(\text{Pb/Sr})_{\text{melt}}/(\text{Pb/Sr})_{\text{peridotite}}$ values in the range 1.4–0.6; the model runs a–d in (b) were obtained with a more restricted range of $(\text{Pb/Sr})_{\text{melt}}/(\text{Pb/Sr})_{\text{peridotite}}$ values, from 1.1 to 0.7. The other parameter values are given in Supplementary Data Table A5. The symbols for the In Teria samples, and the references for the fields of North African mantle xenoliths and Ahaggar Cenozoic volcanism, as well as for the mantle DMM, EM1 and HIMU end-members, are the same as in Fig. 9.
The Ahaggar intra-plate tholeiites contain an EM1-like mantle component, generally ascribed to the involvement of lower lithosphere during their genesis (Maza et al., 1998; Aït-Hamou et al., 2000). Conversely, the HIMU signature found in North African alkali basalts, and in In Teria melilitites, pyroxenites and several spinel peridotites (Fig. 9), is widely considered to represent the signature of upwelling asthenosphere or a mantle plume (e.g. Maza et al., 1998; Aït-Hamou et al., 2000; Beccaluva et al., 2008). In this respect, it is worth noting that the three analysed phlogopite megacrysts plot between HIMU–DMM and EM1 compositions, within the field of the Ahaggar tholeiites, hence providing further evidence for the existence of an EM1 mantle signature beneath In Teria (Fig. 9).

However, although enriched in radiogenic Sr, the isotopic composition of the In Teria spinel peridotites differs markedly from the end-member EM1 signature. Their $^{143}$Nd/$^{144}$Nd ratios are too high and their $^{208}$Pb/$^{204}$Pb ratios too low at a given $^{206}$Pb/$^{204}$Pb value (Fig. 9). Among the analysed phlogopites, one sample shows Pb isotopic compositions consistent with a contribution from EM2 (i.e. a high $^{208}$Pb/$^{204}$Pb value and a composition comparable with that of the Ahaggar tholeiites) whereas the two other phlogopites are isotopically comparable with the In Teria peridotites (Fig. 9c). These features are reminiscent of previous observations of isotopic decoupling in mantle rocks that have been ascribed to daughter element fractionation as a result of melt–rock interactions (e.g. Ionov et al., 2002b; Bodinier et al., 2004; Le Roux et al., 2009). The ‘instantaneous’ effects of daughter element fractionation as a result of melt–rock interactions (as opposed to the ‘time-integrated’ effects of parent–daughter element fractionation) might account for a significant part of the isotopic variability observed in mantle rocks. Porous melt flow and related diffusional or reaction processes may generate isotopic covariation trends between peridotite and melt end-members that diverge markedly from mixing lines.

Percolation model

We used the one-dimensional percolation–diffusion model of Vasseur et al. (1991), modified to include isotopic homogenization (Bodinier et al., 2004), to evaluate a process involving percolation of melt through a lithospheric mantle column to generate the Nd–Sr–Pb isotopic compositions observed in the In Teria mantle xenolith suite. We consider in the model a protolith with an EM1 signature, even if the mantle was probably heterogeneous in composition, varying between DMM and EM1. The model incorporates the effects of melt infiltration velocity, distance of percolation, critical distance of isotopic homogenization along the column, chemical diffusional exchange between melt and minerals, and mineral grain sizes. It assumes instantaneous solid–liquid equilibrium at the surface of mineral grains, considered to be spherical, and chemical diffusion within the grains. Isotopic equilibrium between melt and minerals is governed by the mass-balance equation of isotopic homogenization at the scale of critical volumes. The critical volume of isotopic homogenization is defined as the volume of percolated peridotite in which melt and minerals have reached isotopic equilibrium during the critical time of percolation (i.e. the time it takes for the melt to reach the top of the column).

The parameters used for modelling are compiled in Supplementary Data Table A5 and the results are shown in Figs 15–17. The only parameters that are common to all model runs are the isotopic compositions of the melt and the peridotite protolith. The melt isotopic composition was fixed to the composition of In Teria melilitites for Nd and Sr. For Pb, we took slightly higher values in the range of the Ahaggar and other North African alkali basalts (Allègre et al., 1980; Dupuy et al., 1993; Beccaluva et al., 2008). For the protolith we took the EM1 values of Eisele et al. (2002) for Nd and Pb, and a slightly higher value for Sr, although still within the range of literature values (e.g. Hofmann, 2002).

The results of models to evaluate the influence of compositional parameters (daughter element contents in peridotite and melt, and mineral/melt partition coefficients) on the $^{143}$Nd/$^{144}$Nd vs $^{87}$Sr/$^{86}$Sr covariation are shown in Figs 15 and 16. We first explored parameter values retrieved from the composition of the peridotite xenoliths, their host lavas and clinopyroxene equilibrium melts (Fig. 15). For the peridotite protolith, we tried three Sr–Nd compositions reflecting the compositional range observed in the studied peridotites: (1) the less metasomatized spinel peridotite Erem-2 (run a); (2) the garnet peridotite Int-14-7, metasomatized by alkaline silicate melt but devoid of carbonate-melt imprint (runs b–d); (3) the spinel peridotites (average composition), metasomatized by carbonate melt (runs e–h). For the melt, we used either the average composition of the analysed melilitites (runs a–g) or the theoretical melt in equilibrium with clinopyroxene from the spinel peridotites (run h). Because the composition of the percolating melt evolved along with metasomatism, from silicate (stage 1) to carbonate melt (stage 2), we tried two different sets of $K_D$. We used the experimental values of Hart & Dunn (1993) for minerals/silicate melt (runs a–c) and minerals/clinopyroxene partitioning, and those of Klemme et al. (1995) for clinopyroxene/carbonate melt partitioning (runs d–h). Among the other parameters, only the protolith modal composition and mineral grain sizes are slightly variable (Supplementary Data Table A5). The modes vary from that of an amphibole-free spinel lherzolite (runs a and b) to the amphibole-bearing, average composition of the In Teria spinel peridotites (runs f–h). Mineral grain size is constant and moderately coarse in most of the runs (Cpx radius = 0.15 mm), except for runs g and h, for which we used the average grain size of the spinel peridotites, characterized by very fine clinopyroxene grains (Cpx radius = 0.01 mm). The other parameters (length of the percolation column, porosity, melt velocity, critical distance of isotopic
homogenization, and chemical diffusivity in minerals and melt) are constant in all runs. Their effects were assessed separately (Supplementary Data Fig. A6).

The results in Fig. 15 show a wide range of $^{143}\text{Nd}^{/}^{144}\text{Nd}$ vs $^{87}\text{Sr}^{/}^{86}\text{Sr}$ covariation patterns, from deeply concave to convex upward. All of these patterns differ markedly from the mixing curves between peridotite protoliths and melts, which are almost linear. The shape of the covariation patterns and the degree of decoupling are correlated with the difference in Sr/Nd between melt and peridotite. Low Sr/Nd in the melt and/or high Sr/Nd in the protolith (i.e. $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} < 1$) generate concave variation patterns indicating that $^{87}\text{Sr}^{/}^{86}\text{Sr}$ in the protolith is re-equilibrated with incoming melt at lower melt/rock ratio than $^{143}\text{Nd}^{/}^{144}\text{Nd}$. This is well illustrated by the strongly concave covariation patterns obtained in run h ($(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} = 0.805$) and run a ($(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} = 0.285$). Conversely, high Sr/Nd ratio in the melt and/or low Sr/Nd ratio in the protolith (i.e. $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} > 1$) generate convex patterns indicating that $^{143}\text{Nd}^{/}^{144}\text{Nd}$ re-equilibrates at a lower melt/rock ratio than $^{87}\text{Sr}^{/}^{86}\text{Sr}$. This pattern was obtained with the experiments b–d where $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} = 1.072$. This confirms that the Sr/Nd ratios in the protolith and melt play a major role in the decoupling of Nd–Sr isotopic compositions during porous melt flow (Ionov et al., 2002b; Le Roux et al., 2009). Compared with the Sr/Nd values, the other parameters that were explored through the numerical experiments shown in Fig. 15 have only limited effects. Runs c and d illustrate the subtle effect of mineral/melt partition coefficients as they were calculated with the same set of parameters, except for distinct mineral/melt K0 values (for silicate- and carbonate-melt, respectively). Experiments b and c illustrate the limited effect of adding 6% amphibole to the peridotite mineralogy, whereas experiments e and f show the effect of reducing the mean radius of clinopyroxene grains from 0.15 to 0.01 mm.

Despite their capability to decouple the $^{143}\text{Nd}^{/}^{144}\text{Nd}$ and $^{87}\text{Sr}^{/}^{86}\text{Sr}$ ratios, the experiments shown in Fig. 15 do not fit the whole range of data for the In Teria xenoliths. The samples that are moderately enriched in $^{87}\text{Sr}$ ($^{87}\text{Sr}^{/}^{86}\text{Sr} < 0.704$) can be accounted for by the experiments b–d (run with a relatively low Sr/Nd value for the protolith, based on the composition of the garnet peridotite). However, the other samples are not fitted by these experiments and would require even lower Sr/Nd ratio in the protolith, or a higher Sr/Nd value in the percolating melt. As a consequence, we ran further experiments to explore the effect of compositional variations in Sr and Nd (Fig. 16). For this approach, we first calculated a reference model (short-dashed lines) using the Sr–Nd composition of Primitive Mantle (McDonough & Sun, 1995) for the peridotite protolith and that of the In Teria melanitites for melt ($(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}} = 0.956$). Then we simulated variations of the Sr–Nd contents from 0.33 to three times the concentrations of the reference model (Fig. 16a and b), and variations of the Sr/Nd ratios from 10 to 20 (Fig. 16c and d), by varying the Nd contents (Supplementary Data Table A5). The other parameters are the same as for experiment g in Fig. 15, except for the minerals/melt partition coefficients for which we used the silicate-melt values, instead of the carbonate-melt values.

As shown by the similarity between the reference model and the mixing line, these experiments confirm that the decoupling of Nd and Sr isotopic compositions is virtually non-existent when the peridotite protolith and the incoming melt have similar Sr/Nd ratios, and this result remains true within a wide range of Sr and Nd absolute concentrations (Fig. 16a and b). Conversely, the results shown in Fig. 16b and c indicate that even moderate variations of the $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}}$ ratio generate significant isotopic decoupling. Almost the whole range of the Nd–Sr isotopic variations observed in the In Teria mantle xenoliths can be accounted for by experiments involving $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}}$ values in the range 1.15–1.30. This value may result from a sub-chondritic Sr/Nd ratio (10–12) in the protolith or from a super-chondritic ratio (~20) in the percolating melt. The latter alternative is more likely, as mantle rocks rarely show such low Sr/Nd values, whereas elevated Sr/Nd values can be found in carbonate melts (e.g. ~58 in carbonatites from Morocco; Wagner et al., 2003). Alternatively, the In Teria Nd–Sr isotopic compositions can be explained by a combination of moderately sub-chondritic Sr/Nd values in the peridotite (e.g. ~14.2 as in the garnet peridotite Int-14-7) and moderately super-chondritic Sr/Nd values in melt (e.g. ~17-14, as in ocean island basalt (OIB); McDonough & Sun, 1995). Moreover, the decoupling can be amplified by a short distance of isotopic homogenization, thus requiring lower $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}}$ values. The influence of parameters on Nd–Sr decoupling is discussed in the Supplementary Material (Fig. A6).

It is worth noting that the peculiar Sr–Nd isotopic signature of the In Teria phlogopite megacrysts can be explained by the same two-component porous-flow model as for the spinel peridotites, but requires either lower $(\text{Sr/Nd})_{\text{melt}}/(\text{Sr/Nd})_{\text{peridotite}}$ values (Fig. 16c and d) or a longer distance of isotopic homogenization (Supplementary Data Fig. A6f), or a combination of these. This result lends support to the hypothesis that the phlogopites and the metasomatized spinel peridotites are related to a single event of interaction between the lithospheric mantle and sublithospheric melts, but the former record interaction at relatively high temperature involving silicate melt, whereas the spinel peridotites record lower temperature interaction with more evolved, carbonate melt (Fig. 14). However, two spinel peridotite samples (Erem-2 and In-15-1) remain unexplained by the model, showing elevated $^{143}\text{Nd}^{/}^{144}\text{Nd}$ at a given $^{87}\text{Sr}^{/}^{86}\text{Sr}$ value. The isotopic composition of these samples requires a component characterized by a $^{143}\text{Nd}^{/}^{144}\text{Nd}$ signature higher that both the EM1 signature and the composition of the Ahaggar basalts. Mantle xenoliths with Nd isotopic compositions recording long-term depletion in LREE have been reported from Ahaggar and other North African
localities (Beccaluva et al., 2007, 2008). In addition, the $^{208}\text{Pb}^{204}\text{Pb}$ compositions of several samples also point to the involvement of a depleted component with a DMM-like signature (Fig. 9c).

To evaluate how much of the Pb isotopic systematics can be explained by the two-component porous-flow model, and how much requires a third, DMM component, we carried out experiments simulating covariations of $^{207}\text{Pb}^{204}\text{Pb}$ and $^{208}\text{Pb}^{204}\text{Pb}$ vs $^{87}\text{Sr}^{86}\text{Sr}$ (Fig. 17). We ran several experiments involving various combinations of Pb/Sr ratios in the protolith and melt (Supplementary Data Table A5) by varying the Pb content in the protolith and melt ($^{207}\text{Pb}^{204}\text{Pb}$ vs $^{87}\text{Sr}^{86}\text{Sr}$, Fig. 17a), or in the melt only ($^{208}\text{Pb}^{204}\text{Pb}$ vs $^{87}\text{Sr}^{86}\text{Sr}$, Fig. 17b). The other parameters, including the Sr content in the protolith (≈ PM after McDonough & Sun (1995)) and melt (≈ In Teria melilitite), are the same as in the reference model in Fig. 16. The results for the $^{207}\text{Pb}^{204}\text{Pb}$ vs $^{87}\text{Sr}^{86}\text{Sr}$ covariation (Fig. 17a) show that the whole In Teria array can be accounted for by the two-component model within a narrow range of (Pb/ Sr)$_{\text{melt}}$(Pb/Sr)$_{\text{peridotite}}$ values between 0.7 and 0.9 (runs c–f). For Pb/Sr values in melt consistent with the In Teria melilitites (Pb/Sr = 0.00371 ± 0.00015) and OIB values ($0.0048$; McDonough & Sun, 1995), this constrains the Pb/Sr values of the protodite protolith in the In Teria melilitites. However, the existence of an EM1 signature both in the Ahaggar continental tholeiites and in the In Teria xenoliths suggests that this signature was widely distributed in the lithospheric mantle beneath the Tuareg Shield and adjacent areas. For Beccaluva et al. (2007), the absence of this signature in the Ahaggar (Manzaz) xenoliths is thus ‘perplexing’. Those researchers interpreted the mantle beneath the Ahaggar Swell as newly accreted material, owing to the lateral displacement and replacement of old lithospheric mantle by upwelling asthenosphere. They also suggested that the In Teria xenoliths could ‘plausibly represent the (EM1-metasomatized) older cratonic lithospheric mantle’.

In fact, the In Teria mantle xenoliths differ from true cratonic mantle in several respects (e.g. their relatively low Mg# values compared with kimberlite-hosted peridotite xenoliths and their dominantly fertile, lherzolite compositions) and are much more similar to mantle xenoliths from younger lithospheric domains (Menzies, 1990; O’Reilly et al., 2001, and references herein). This lends support to the hypothesis that the destabilization of the Saharan Metacraton in the Neoproterozoic was associated with extensive rejuvenation of the lithosphere, which occurred either regionally as a result of its delamination and thermo-mechanical erosion after thickening (Ashwal & Burke, 1989; Black & Liégeois, 1993; Abdelsalam et al., 2011) or more locally along mega-shear zones (Liégeois et al., 2003; Fezaa et al., 2010). Despite their strong metasomatic imprint, the In Teria xenoliths have nevertheless preserved some mineralogical and geochemical features that may be ascribed to the original Pan-African lithosphere, prior to the Cenozoic events. These include remnants of the EM1 lithospheric isotopic signature, as predicted by Beccaluva et al. (2007). The EM1 signature was possibly inherited from subduction processes that occurred during the Pan-African orogeny in conjunction with the
amalgamation of lithospheric terranes (Black et al., 1994; Liégeois et al., 1994; Caby, 2003).

Apart from the probable existence of a widespread lithospheric EM1 signature, the nature of the lithospheric mantle beneath the Tuareg Shield and adjacent areas prior to the Cenozoic asthenosphere–lithosphere interactions is otherwise virtually unknown. In particular, the lateral variations in lithospheric thickness and composition that would be expected from the juxtaposition of lithospheric blocks of different provenances and ages (e.g. Liégeois et al., 2005) are poorly constrained. In addition to the scarcity of data, the lack of information also reflects the modifications that have affected the lithospheric mantle during the Cenozoic. These modifications were particularly severe in the central part of the Ahaggar Swell where the lithosphere was strongly modified by extensive interaction with plume melts (Dautria & Lesquer, 1989) or, possibly, rejuvenated by removal of old lithospheric material by upwelling asthenosphere (Beccaluva et al., 2007). The modification of the lithosphere was less intense in the In Teria district where melt–rock interactions culminated in the pervasive infiltration of small-volume carbonate melts. The persistence of a relatively thick thermal boundary layer beneath In Teria would have facilitated the formation of carbonate melts by incipient lithospheric melting during transient heating and/or through the evolution of primary silicate melts via melt–rock reactions (Wallace & Green, 1988; Dalton & Wood, 1993; Sweeney, 1994; Yaxley & Green, 1996).

In scenarios involving the impingement of a mantle plume head centred on the Ahaggar Swell, or any alternative involving upper mantle diapiric instabilities (Davies & Bunge, 2006; Lustrino et al., 2007) or shallow asthenospheric upwelling (Beccaluva et al., 2007), the evolution recorded by the In Teria mantle xenolith suite may simply reflect the attenuated effects of this process owing to the distal location of In Teria, to the NE of the Ahaggar Swell (Fig. 1). As noted by Aït Hamou & Dautria (1994), this scheme is supported for the Ahaggar Swell by the spatio-temporal evolution of the volcanism, varying from volumetrically predominant Eocene to Oligocene plateau tholeiites in the central part of the Swell (Taharaq district) to smaller volumes of Miocene to Quaternary undersaturated alkaline lavas in the outer parts. The isotopic signature of the In Teria melilitites is virtually indistinguishable from that of the Ahaggar alkaline lavas (Fig. 9) and thus this recent volcanic district might be viewed as the most distal manifestation of the asthenosphere–lithosphere interactions related to the Ahaggar Swell. However, available heat flow data (Lesquer et al., 1989, 1990; Takherist & Lesquer, 1989) do not support this model. In Teria is situated on an east–west high heat flow axis that trends from Libya towards the Canary Islands, between the Sahara Basins to the north and the Ahaggar Swell and the West African Craton to the south (Lesquer et al., 1990). This anomaly is a prominent feature of the heat flow pattern in northwestern Africa, showing its maximum values (100–120 mW m$^{-2}$) in southern Algeria. As suggested by Lesquer et al. (1990), the In Teria volcanism might be related to the mantle processes that are responsible for this anomaly. In this respect, it may be worth noting that mantle xenoliths comparable with those from In Teria—and also brought to the surface by dominantly undersaturated lavas—are found along the western extension of the anomaly, in southern Morocco (Jbel Sagho, Ibhi et al., 2002) and in the Canary Islands (e.g. Neumann et al., 2002). The similarities not only include the presence of abundant reaction aggregates containing microgranular secondary phases and glass, and evidence for interaction with carbonate melts. Alkaline pyroxenites and wehrlites are common, sometimes accompanied by carbonated peridotites or carbonatites. The significance of the North Saharan thermal anomaly is poorly understood, however. Edge-driven convection (Missenard & Cadoux, 2012) was recently proposed for southern Morocco, where the western termination of the anomaly runs along the northern border of the West African Craton. In Algeria, however, this model is consistent with the observation that the anomaly cross-cuts at a high angle the north–south lithospheric structures resulting from terrane amalgamation and shear-zone activity during the Pan-African orogeny (Black et al., 1994).

CONCLUSIONS

The In Teria xenolith suite records three different metamorphic imprints, which all postdate the last deformation episode in the lithospheric mantle of this region. Stage 1 is a diffuse infiltration of hydrous alkaline melts affecting the spinel and garnet peridotites. Melt percolation and the crystallization of the reaction products during this episode are controlled by the pre-existing deformation microstructure of the peridotite. This first metamorphic stage was probably coupled with a heating of the deep subcontinental lithosphere by melts of an asthenospheric origin. Stage 2 is a carbonate melt metasomatic imprint restricted to the spinel peridotites, which largely overprints stage 1. Finally, stage 3 corresponds to the crystallization of pyroxenites, which are probably crystal segregates from hydrous alkaline melts crystallized in vein conduits. The three metasomatic imprints represent successive stages of interaction between the lithosphere and sublithospheric melts.

The In Teria peridotites with enriched $^{87}\text{Sr}/^{86}\text{Sr}$ compositions, as well as phlogopite megacrysts, plotting within the field of the Ahaggar tholeiites, show evidence for the involvement of an EM1-like component, probably inherited from the lithospheric mantle beneath In Teria. However, the isotopic composition of the In Teria spinel peridotites differs from the end-member EM1 signature, which might be accounted for by isotopic decoupling in mantle rocks. Assuming that the Ahaggar alkaline basalts and the In Teria melilitites represent melts of essentially sublithospheric origin, numerical modeling simulating porous melt percolation through an
EM1 lithospheric mantle can reproduce the high $^{87}$Sr/$^{86}$Sr values observed in some peridotite xenoliths. These high $^{87}$Sr/$^{86}$Sr values may be related to the crystallization of interstitial micro-phases from small-volume melts and chromatic effects. Conversely, the HIMU signature observed in the In Teria melilitites, pyroxenites and several spinel peridotites, as well as North African alkali basalts, is widely considered to represent a mantle plume signature. Additionally, the $^{208}\text{Pb}/^{204}\text{Pb}$ composition of several samples points to the involvement of a depleted component with a DMM signature. The lithospheric mantle beneath In Teria was probably not uniformly EM1-like; it probably included a DMM peridotite component as well as some peridotites with elevated $^{143}\text{Nd}/^{144}\text{Nd}$ values recording long-term LREE depletion.

The In Teria mantle xenoliths differ from North African and Ahaggar Swell mantle xenoliths by the presence of garnet peridotite, the extensive carbonate melt metasomatic imprint, and their high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The last are considered to be inherited from an EM1-like lithospheric mantle protolith. The similar EM1 signature in Ahaggar continental tholeiites suggests that this signature extends over the entire Tuareg Shield and adjacent areas and that the generation of the tholeiites involved melting of the base of the lithosphere. The EM1 signature was possibly inherited from subduction processes that occurred during the Pan-African orogeny. In Teria mantle xenoliths therefore have geochemical characteristics similar to those of young lithospheric mantle rather than cratonic mantle domains, supporting the hypothesis that the destabilization of the Saharan Metacraton in the Neoproterozoic was associated with extensive lithosphere rejuvenation.

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SUPPLEMENTARY DATA

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

REFERENCES


