**Deformation and seismic anisotropy of the subcontinental lithospheric mantle in NE Spain: EBSD data on xenoliths from the Catalan Volcanic Zone**

Mercè Fernández-Roig1, Gumer Galán1, Elisabetta Mariani2

1 Departament de Geologia, Universitat Autònoma de Barcelona, Av. dels Til·lers s/n, 08193 Bellaterra, Barcelona, Spain, e-mail: merce.fernandez@uab.cat, gumer.galan @uab.cat

2 School of Environmental Sciences, University of Liverpool, 4 Brownlow Street, L60 3GP Liverpool, England, UK, e-mail: mariani@liverpool.ac.uk

**Abstract**

Mantle xenoliths in Neogene-Quaternary basaltic rocks related to the European Cenozoic Rift System serve to assess the evolution of the subcontinental lithospheric mantle beneath the Catalan Volcanic Zone in NE Spain. Crystallographic preferred orientations, major element composition of minerals, and temperature and pressure estimates have been used to this end. The mantle consists of spinel lherzolites, harzburgites and subordinate websterites. Protogranular microstructures are found in all peridotites and websterites, but lherzolites also display finer-grained porphyroclastic and equigranular microstructures. The dominant olivine deformation fabric is [010] fiber, but subordinate orthorhombic and [100]-fiber types are also present, especially in porphyroclastic and equigranular lherzolites. The fabric strength (J index=10.12-1.91), equilibrium temperature and pressure are higher in xenoliths with [010]-fiber fabric and decrease in those with orthorhombic and [100]-fiber type. Incoherence between olivine and pyroxene deformation fabric is mostly found in porphyroclastic and equigranular lherzolites. Seismic anisotropy, estimated from the crystal preferred orientations, also decreases (AVp= 10.2- 2.60%; AVs max= 7.95- 2.19%) in porphyroclastic and equigranular lherzolites. The olivine [010]-fiber fabric points to deformation by simple shear or transpression which is likely to have occured during the development of late-Hercynian strike-slip shear zones, and to subsequent annealing during late Hercynian decompression, Permian and Cretaceous rifting. Also, it cannot be excluded that the percolation of mafic magmas during these extensional events provoked the refertilization of the lithospheric mantle. However, no clear relationship has been observed between fabric strength and mineral mode and composition. Later transtensional deformation during late Alpine orogenesis, at higher stress and decreasing temperature and pressure, transformed the earlier fabric into orthorhombic and [100]-fiber type.

Comparison of seismic anisotropy estimates with the available SKS-wave splitting data suggests that most of the measured seismic anisotropy would be explained by the lithospheric contribution, if the lithospheric mantle fabrics record mainly transpression and transtensional deformation.

**Key words:** Lithosphere; mantle; olivine; pyroxene; deformation fabric; seismic anisotropy

**1. Introduction**

Mantle xenoliths enclosed in volcanic rocks provide key insights into the evolution of the lithospheric mantle. These xenoliths within Neogene-Quaternary alkaline basaltic rocks from the Catalan Volcanic Zone (CVZ) in NE Spain have been previously studied in attempts to determine the composition and geochemical evolution of the subcontinental lithospheric mantle (SCLM) in this area (Bianchini et al., 2007; Galán et al., 2008; 2011; Galán and Oliveras, 2014). However, little is known about this mantle petrophysical properties and deformation history. The CVZ is part of the European Cenozoic Rift System (ECRIS) which also lies along the Iberian Mediterranean coast (Fig. 1a). The deformation affecting the SCLM could well be older though, taking into account that the overlying crust registered the Variscan and Alpine orogenesis.

Experimental studies demonstrate that plastic deformation related to mantle flow processes at different scales causes strong crystal preferred orientation (CPO) of olivine and of other mantle minerals (Karato et al., 2008 and references therein). In addition, numerical modelling shows that different deformation conditions lead to varying olivine CPO (Tommasi et al., 1999). Variables include temperature (T), pressure (P) (Couvy et al., 2004; Jung et al., 2009; Ohuchi et al., 2011), stress and H2O concentration (Jung and Karato, 2001; Jung et al., 2006; Katayama et al., 2004; Katayama and Karato, 2006). CPO can also be affected by factors such as partial melting and melt percolation (Vauchez and Garrido, 2001; Holtzman et al., 2003; Rampone et al., 2010; LeRoux et al, 2008). Furthermore, CPO due to plastic deformation has been demonstrated to be the major cause of seismic anisotropy (Nicolas and Christensen, 1987; Mainprice and Silver, 1993; Mainprice et al., 2000; Mainprice, 2007). The comparison between mantle seismic anisotropy estimated from mineral CPO (Mainprice and Humbert, 1994) and regional anisotropy deduced from geophysical methods is therefore particularly useful for interpreting geodynamic and geochemical processes affecting the SCLM (Vonlanthen et al., 2006; Falus et al., 2008; Bascou et al., 2011; Palasse et al., 2012; Baptiste and Tommasi, 2014; Park et al., 2014; Park and Jung., 2015).

"Anhydrous" spinel lherzolites and harzburgites are the lithologies most frequently found in mantle xenoliths from lavas of the CVZ. Smaller numbers of websterites are also found (Bianchini et al., 2007; Galán et al., 2008). Lherzolites display variable microstructures, from protogranular (descriptive terminology by Mercier and Nicolas, 1975) or coarse granular to porphyroclastic and equigranular, whereas harzburgites and websterites display only protogranular microstructures. In this study, we evaluate the xenolith microstructures, major element compositions of minerals, T and P estimates, olivine and pyroxene CPO and resulting anisotropic seismic properties, in order to determine the deformation history, relationships with crustal tectonothermal events and thermal state of the SCLM beneath the CVZ. The results serve to explain the deformation mechanisms, geodynamic setting and evolution of this part of the European SCLM and are compared with SKS-wave splitting measurements for this area to assess the SCLM contribution to the measured seismic anisotropy.

**2. Geological Setting**

The Catalan Volcanic Zone (CVZ) is part of the Neogene-Quaternary volcanism of the Iberian Peninsula (Ancochea, 2004), and the result of late-Alpine extensional fractures that extended the ECRIS along the Iberian Mediterranean coast (Martí, 2004 and references therein) from 35 to 0 Ma (Figs. 1a-b). This rifting produced extensional basins in the western Mediterranean off-shore (viz., the Valencia Trough, the Gulf of Lion, the Liguro-Provençal basin) and in the Catalan margin on-shore (Roca and Desegaulx, 1992). However, as the extensional direction in this area is NW-SE instead of W-E, some authors (Barruol and Granet, 2002; Barruol et al., 2004) have argued that it is not related to the ECRIS event at all, but rather to back-arc spreading episodes induced by a roll-back of the Tethys slab subduction beneath the Corsica-Sardinia lithospheric block towards the NW, between 22-17 Ma.

Before the Neogene rifting, the lithosphere of this area was mainly structured by the Hercynian orogenesis and, to a lesser degree, by the Alpine orogenesis which led to the formation of the Catalan Coastal Ranges. Lithospheric extensive episodes with injections of mafic magmas in between the two orogenesis took place during the Permian (Martínez-Poza et al., 2014), when calc-alkaline to alkaline lamprophyres were intruded (Losantos et al., 2000; Enrique Gisbert, 2009; Ubide et al., 2010), and during the Cretaceous, with intrusion of alkaline lamprophyres (Solé et al., 2003; Ubide et al., 2012; Esteve et al., 2014).

The CVZ is divided into three subzones (Fig. 1b): L’Empordà, La Selva and La Garrotxa, which are Neogene-Quaternary basins, mainly limited by normal NW-SE faults related to neotectonic activity (11 to 0 Ma) (Lewis et al., 2000). The xenoliths studied were collected in alkaline basaltic necks, lava flows, and pyroclasts from three volcanoes: Sant Corneli in La Selva subzone (3.5-2Ma; Donville, 1973c), and Canet d’Adri and La Banya del Boc in La Garrotxa subzone (0.1-0.01Ma; Donville, 1973b) (Fig. 1b). The volcanic activity is linked to NW-SE Neogene normal faults, such as the Llorà fault (Fig. 1b), which has a transtensional component (Goula et al., 1999; Olivera et al., 2003). According to Bolos et al. (2014), this fault may extend down to the lower crust and lithospheric mantle controlling the eruption of alkaline mafic magmas that include lower-crust cumulates and mantle-derived xenoliths. The presence of these xenoliths is restricted to the volcanoes related to this fault.

Previous thermobarometric estimates on the mantle xenoliths from the CVZ (Galán et al., 2011), using the two-pyroxene thermometer of Brey and Köhler (1990), provide higher equilibrium T for harzburgites (1062± 29 ºC) and websterites (up to 1257 ºC) than for lherzolites (972± 89 ºC). Pressure estimates, using the clinopyroxene-olivine thermobarometer of Köhler and Brey (1990), range from 1.41 GPa to 0.70 GPa for lherzolites, while values of 1.02± 0.19 GPa are obtained for harzburgites. The lowest P in lherzolites is consistent with the existence of plagioclase in coronae around spinel in several samples. The CVZ xenoliths are classified as the off-craton type, with major and trace element compositions similar to those of other xenoliths from neighbouring European SCLM areas, for instance to those from the southern part of the Massif Central, France (Galán and Oliveras, 2014). According to these authors, lherzolites and harzburgites from NE Spain resulted from melt depletion and subsequent cryptic metasomatism. However, they do not exclude the possibility that some lherzolites, with clinopyroxene showing light Rare Earth Elements (LREE) depletion and Depleted MORB Mantle (DMM) isotopic compositions, could have been formed by refertilization of refractory, but isotopically enriched, harzburgites, via percolation and reaction with N-MORB type basalts, a theory proposed for lherzolites from the neighbouring ultramafic massif of Lherz (Eastern Pyrenees) (Le Roux et al., 2007). On the other hand, harzburgites from the CVZ underwent multi-stage metasomatism (Galán and Oliveras, 2014). An earlier episode was caused by subduction-related hydrous fluids or melts. A second metasomatic episode was mainly cryptic and related to the percolation of alkaline silicate melts and carbonatite derivatives. The latter metasomatism also affected a few lherzolites. The Sr-Nd isotopic compositions of the metasomatizing alkaline silicate melts and carbonatite derivatives were close to that of websterites (Galán and Oliveras, 2014), and within the range of the Continental Mantle Reservoir (Lustrino and Wilson, 2007). Websterites are interpreted as cumulates from the mafic alkaline silicate melts that caused the latter cryptic metasomatism (Galán et al., 2008).

**3. Methods**

The mineral mode (% volume) of the xenoliths was determined using the petrographic microscope with a precision stepping stage and  an automatic James Swift point counter (Table 3.1E in electronic supplementary material). Microstructures were established through the optical analysis of thin sections cut in random orientation since the foliation and lineation could not be determined in most macroscopic hand specimens.

Major element mineral compositions of the samples were analyzed using an Electron Probe Microanalyser (EPM) (Cameca SX50) at the Scientific-tecnic Service of the University of Barcelona on polished thin sections. Beam current and acceleration voltage were 15 nA and 20 kV respectively, standards were both natural and synthetic, and the counting time was 10 s for all elements, except for Ca in olivine, which was 30 s. The beam width was enlarged to 50 μm in order to resolve, within uncertainty, pre-exsolved compositions of both pyroxenes in a few xenoliths. Tables 3.2E-3.3E, in supplementary electronic material, include average compositions of orthopyroxene and clinopyroxene used for thermometric estimates. Composition of both pyroxenes and other phases are summarized in Table 1.

The EPM thin sections were further prepared for quantitative crystallographic analyses, using electron backscatter diffration (EBSD) (eg., Ben Ismail and Mainprice, 1998; Prior et al., 2009) by performing chemo-mechanical polishing with a 0.05 µm colloidal silica solution, and then coating with 3 nm thick carbon layer to avoid charging issues in the scanning electron microscope (SEM). EBSD was carried out in the EBSD-SEM Laboratory in the School of Environmental Sciences at the University of Liverpool, using two different SEMs: a XL30 tungsten filament SEM and a CamScan X500 CrystalProbe field emission gun (FEG) SEM, both equipped with EBSD-EDS systems from Oxford HKLTechnology. Automatic orientation mapping was carried out over the entire thin section area (~ 30 mm × 25 mm) of three samples (SC.11.16, BB.08.20, BB.08.101), and over an area of ~ 15 mm × 20 mm for the remaining thin sections, with sampling steps ranging from 30 to 50 μm depending on the sample’s grain size. Accelerating voltage and working distance were 20 kV and 25 mm respectively. The Channel 5 software was used for automated indexing of all minerals present and for processing EBSD datasets. Indexation rates in the raw maps were ~ 65% and with generally better results for olivine than for pyroxenes, mainly due to the difficulty of polishing several coexisting mineral phases. Non-indexed areas resulted from fractures, alteration, basalt zones, grain boundaries and polishing defects. Mean angular deviation (MAD), which shows the match between lattice planes in a calculated orientation and measured lattice planes in EBSD patterns was commonly< 1° (mainly ≈ 0.6°). Post-acquisition data processing allowed us to improve the indexation rate by using grey-scale values of band contrast images (these provide a visual report of kikuchi pattern quality) to constrain areas where non-indexed pixels could be filled with the average orientation of neighbouring pixels. During this process, isolated single pixels representing points misorientated ≥ 10° from the average orientation of surrounding pixels were removed first (ie., wild spikes). Then, empty pixels were replaced with the most common neighbouring orientation, if they had up eight identical neighbours. This process was repeated using sequentially seven, six, five and four identical neighbours. At each step great care was taken by using both band contrast images and micrographs to ensure that no artifacts were introduced into datasets. This method minimizes the over-counting of individual grains when extracting one orientation point per grain for the construction of pole figures (section 8). Raw and processed EBSD maps of representative samples are included in supplementary electronic material (Fig. 3.1E).

Olivine and pyroxene CPO data are displayed in pole figures, presented as lower hemisphere stereographic projections, using the PFctf programme by Mainprice (ftp://www.gm.univ-montp2.fr/mainprice//CareWare\_Unicef\_Programs/). To avoid over-representation of coarse grains, pole figures were plotted by considering one point per grain. To facilitate comparison of CPO between different samples, the lineation and foliation of most of which could not be identified clearly in hand specimen, data were rotated to a standard position, using the ROTctf programme by Mainprice, with the maximum concentration of olivine [100] and [010] axes parallel to the E–W and the N–S direction, respectively (see section 7 here below). This orientation would correspond to that expected in a sample with a horizontal E–W lineation (structural X direction) and a vertical foliation plane (normal to the structural Z direction), if olivine had been deformed by dislocation creep with dominant activation of the high-temperature, low-stress (010)[100] slip system (Ben Ismaïl and Mainprice, 1998). Therefore, when we refer to "lineation" and "folitation" from now on, it must be understood they are inferred from CPO data. Finally, the rotation axes that accommodate all low-angle misorientations (2-15 °) or subgrains are analysed within all olivine crystals of the EBSD maps and within selected olivine crystals of specific xenoliths and displayed on inverse pole figures (section 8).

The fabric strength (Bunge, 1982) was estimated from the dimensionless number J index= ∫ f(g)2 dg, where f(g) is the density in the orientation distribution function at orientation g, and dg = dφ1dФdφ2sinФ/8π2, where φ1, Φ and φ2 are the Euler angles that define the rotations which allow for comparison between the crystallographic and external reference frames (Bunge, 1982). The J indices for all samples were calculated from the mean orientation of each grain using the SuperJctf software by Mainprice: ftp://www.gm.univmontp2.fr/mainprice//CareWare\_Unicef\_Programs/, using a 10 ° Gaussian half-width, 1 ° cells, and truncation of the orientation distribution function at 22 °.

 Further analysis of the mineral CPO was performed using the point (P), girdle (G) and random (R) fabric-type indices for olivine [100], [010] and [001] axes, and clinopyroxene (100), (010) and [001] distributions. These indices were calculated from the three eigenvalues (λ1, λ2, λ3) of the normalized orientation matrix (Woodcock and Naylor, 1983; Humbert et al., 1996) for each principal crystallographic axis as: P = λ1–λ3, G = 2(λ2–λ3) and R = 3 λ3 (Vollmer, 1990) and can be combined for a more thorough analysis of CPO symmetry. The BA index= 0.5 ∗ (2.0 − (P010 / (G010 + P010)) − (G100 / (G100 + P100))) (Mainprice et al., 2015) and the LS index = 0.5 ∗ (2.0 − (P010 / (G010 + P010)) − (G001 / (G001 + P001))) (Ulrich and Mainprice, 2005) were used for classifying olivine and clinopyroxene CPO, respectively. As these numerical classifications do not require knowledge of the orientation of the specimen reference frame in absolute terms, they are convenient for the analysis of mantle xenoliths.

Seismic properties were calculated for ambient T and P conditions, using single crystal elastic constants, mineral density, CPO and main mineral (olivine, orthopyroxene and clinopyroxene) mode (Table 3.1E) following Mainprice and Humbert (1994). The Anis\_ctf and Poly\_Adobe\_VpVs software by D. Mainprice(ftp://www.gm.univmontp2.fr/mainprice//CareWare\_Unicef\_Programs/) was used in the computation, with the Voigt-Reuss-Hill average scheme for experimentally determined single crystal elastic constants (Abramson et al., 1997; Chai et al., 1997; Collins and Brown, 1998). In addition, the EMATRIX\_average software also by D. Mainprice served to estimate the average seismic properties of the local SCLM. Calculated seismic properties are summarized in Table 3.4E (in supplementary electronic material) and displayed in lower hemisphere projections plotted following the same conventions used to represent the CPO data in pole figures (ie., one point per grain and data rotation) as described in the above paragraphs (see section 9).

**4. Petrography of xenoliths**

A total of eighteen xenoliths were studied. Seventeen were new and one was previously studied (CA.44.05) by Galán et al. (2011). Selection was based on size, with those of ca., 2×4 cm or larger and less affected by basalt impregnation being regarded as the most suitable. These xenoliths are representative of the main lithologies found in the SCLM under the CVZ (Galán et al., 2008) and include thirteen spinel lherzolites, one plagioclase-spinel lherzolite, three spinel harzburgites, and one olivine websterite (Fig. 2). Accessory amphibole and phlogopite are present only in lherzolite BB.08.59 and in the sole websterite, respectively. Amphibole forms scarce disseminated anhedral crystals (≈100 µm in size) which can grow around spinel. Phlogopite crystals are very sparse and subhedral (500-1000 µm at their long axis). Both lherzolites and harzburgites were taken from the Canet d'Adri and Banya del Boc volcanoes, whereas only lherzolites were found in the Sant Corneli neck. The websterite is from the Banya del Boc volcano (Fig. 1b).

The three harzburgites, four lherzolites and the websterite show protogranular or coarse granular microstructure (Figs. 3a-c) with average grain size ≥ 2mm (Harte, 1977). Among these samples, one of the harzburgites (BB.12.04) and two of the lherzolites (BB.08.101, BB.08.98) show a preferred orientation of elongated olivine crystals, which gives rise to a coarse tabular microstructure. Although protogranular is the dominant microstructure, lherzolites also display finer-grained porphyroclastic and equigranular microstructures, with protogranular-porphyroclastic transitions in a few of them (Figs. 3d-f).

Lherzolites represent 60% of the SCLM rocks in this area, of which 15% are porphyroclastic and 9% are equigranular (Fernández-Roig and Galán, 2015). The mineral mode of the protogranular lherzolites is 7 to 14% clinopyroxene, 26-37% orthopyroxene, 50-65% olivine and < 5% spinel. Olivine and orthopyroxene crystals are the largest (2-5mm at the long axis), whereas clinopyroxene and spinel crystals are < 2mm. All minerals form anhedral crystals with straight or slightly curved olivine-olivine grain boundaries, occasionally at 120°. However, more interlobate grain boundaries are seen when olivine is next to other minerals, eg., bordering orthopyroxene crystals, which can include occasionally small rounded olivine crystals in random orientation (BB.08.98, BB.12.01; Fig. 4a). Spinel forms interstitial brownish-green or green crystals with frequent embayed grain boundaries (up to 1700 µm in size). These crystals can also have round shaped olivine and orthopyroxene inclusions and be orientated. Other forms of spinel are lamellae within both pyroxenes. Clinopyroxene crystals (0.5< grain size< 2mm) show either straight or gently curved grain boundaries. Olivine (and more rarely orthopyroxene) show sporadic evidence of intracrystalline deformation and recovery, such as undulose extinction, deformed exsolved lamellae, deformation bands and widely spaced subgrain boundaries (Figs. 4a and 3.1Eb in supplementary electronic material).

Two lherzolites with transitional protogranular-porphyroclastic microstructure (BB.08.59, BB.12.13) have similar mineral mode range as the protogranular lherzolites (Table 3.1E in supplementary electronic material). These transitional samples include protogranular and finer grained (0.3 <grain size< 2 mm) areas without any specific arrangement (Fig. 3d). In the former, there is more evidence of intracrystalline deformation in olivine and orthopyroxene crystals and more frequent exsolved clinopyroxene lamellae in orthopyroxene than in the protogranular lherzolites. In the finer-grained areas, recrystallized strain-free neocrystals, or with rare subgrains, and frequent grain boundaries at 120 ° dominate. Spinel appears either as isolated interstitial crystals (≈50 µm) or more rarely as larger amoeboid crystals (up to 3 mm at their long axis), intergrown with orthopyroxene as “clusters” (Fig. 4b).

Porphyroclastic lherzolites (SC.11.16, SC.11.49, SC.11.52, BB.08.80, CA.12.03; Table 3.1E) (Fig. 3e) have a mineral mode range overlapping that of protogranular lherzolites. SC.11.49 stands out for showing plagioclase along with olivine, clinopyroxene and chromite forming coronae around spinel crystals. These lherzolites display mainly orthopyroxene porphyroclasts (3-11 mm at their long axis). Less frequently they display olivine, clinopyroxene and spinel porphyroclasts in a much finer matrix (0.2< grain size< 1mm) made up of strain-free neocrystals, or with sporadic subgrains, of the same minerals. Orthopyroxene porphyroclasts display evident features of intracrystaline deformation (Fig. 4c), such as frequent bent clinopyroxene and spinel lamellae at the core, undulose extinction, kinks bands and more closely spaced subgrain boundaries than in protogranular lherzolites. They also display embayment by strain-free olivine neocrystals that can be also seen as rare inclusions (eg., CA.12.03). Spinel porphyroclasts are amoeboid crystals intergrown mainly with orthopyroxene. They can be very stretched (aspect ratios from 1:3 to 1:7) and along with orthopyroxene porphyroclasts define a clear orientation (eg., in SC.11.16; Figs. 4d and 3.1Ed in electronic supplementary material). Spinel neocrystals are smaller (≈ 60 µm) rounded and interstitial between silicates. Clinopyroxene is mainly present as isolated and elongated clustered (aspect ratio 1:2) neocrystals (Fig.3.1E in supplementary electronic material). Spinel lamellae are frequent in both clinopyroxene neocrystals and rare porphyroclasts. Two types of matrix were distinguished according to the shape of the grain aggregates: one is polygonal as in SC.11.49, SC.11.52, SC.11.16, whereas the other has slightly interlobate grain boundaries (BB.08.80). Finally, in two cases (SC.11.16, CA.12.03), elongated olivine neocrystals define a tabular matrix (Fig. 3e).

Most equigranular lherzolites (BB.08.38, BB.08.57, SC.11.54) are also tabular because of elongated olivine neocrystals. Those included in Table 3.1E show the highest clinopyroxene mode (12-18%), 20-37% orthopyroxene, 42-62% olivine and <5% spinel. They show less than 5% porphyroclasts (grain size up to 5 mm), mostly of orthopyroxene, in a finer grained matrix (0.3< grain size< 1mm), where spinel is usually present as interstitial rounded crystals, or elongated crystals with lobate grain boundaries. Grain boundaries are generally straight or slightly curved among neocrystals, with frequent triple junctions (Fig. 3f). Intracrystalline deformation or recovery features are either absent or very sporadic.

 Protogranular harzburgites (BB.08.20, CA.12.02, BB.12.04) represent 36% of the total mantle xenoliths sampled so far. Those in Table 3.1E have 71-76 % olivine, 21-27% orthopyroxene, 4-1% clinopyroxene and <1% spinel. Microstructure features are similar to those of protogranular lherzolites but they differ in the following aspects: (1) crystals of olivine and orthopyroxene are more coarse grained (2-6 mm at their long axis); (2) spinel crystals (100< grain size< 700 µm) are generally darker brown anhedral and interstitial (Fig. 4e) and, more rarely, amoeboid; and (3), intracrystalline deformation or recovery features (subgrains) are sparser than in lherzolites. Interpenetrating olivine-orthopyroxene grain boundaries, with both minerals demonstrating rounded inclusions are occasionally observed in sample CA.12.02 (Fig. 4f). Moreover, it has been noted that BB.12.04 is protogranular tabular due to the orientation of elongated olivine crystals.

The websterite BB.12.17 is also protogranular, with average grain size of 2-3 mm, straight or gently curved grain boundaries and few deformation effects: only olivine crystals display undulose extinction and deformation bands.

**5. Mineral chemistry**

Average chemical analyses of major elements and formula unit (fu) of both pyroxenes are included in Tables 3.2E-3.3E of electronic supplementary material and a brief summary of all phases in Table 1. Compositional range of periodotite minerals are comparable to those provided by Galán and Oliveras (2014). Crystal zoning was not observed and compositions of minerals forming different types of crystals (porphyroclasts, neocrystals, lamellae) in a single sample are comparable. However, analyses of a few orthopyroxene and clinopyroxene porphyroclasts both showing mutual and spinel lamellae, performed with an enlarged electron beam (ie., unfocused analyses), indicate higher Mg number [Mg#= 100\*Mg/(Mg+Fet) in atoms per fu] and lower Cr number [Cr#= 100\* Cr/(Cr+Al)] in pre-exsolved compositions. The forsterite (Fo) component in olivine, Mg# and Cr# for both pyroxenes are seen to decrease from harzburgites to lherzolites (Table 1). Spinel *sensu stricto* is present in lherzolites and Mg-Al chromite in harzburgites. Cr# decreases for spinel while, by contrast, Mg# increases from harzburgites to lherzolites. These general compositional variations are consistent with both melt depletion trends and with trends resulting from the refertilization of harzburgites via the percolation of basaltic melts to give lherzolites. The websterite is of particular note because olivine and pyroxenes show similar or higher Mg# and Cr# values than do harzburgites. This is in contrast to results of a previously studied websterite which has more evolved compositions (Galán et al., 2008).

Covariation diagrams for lherzolite minerals (Fig. 5) serve to establish the relationships between their chemical compositions and the microstructure types. Harzburgites were excluded from these diagrams because they represent a more refractory system and only show protogranular microstructure.

Mg# vs. Cr# for lherzolite clinopyroxene and spinel show overlap between the three microstructural types of lherzolites (Fig. 5a-b) and a broad overall positive correlation between the two numbers only for clinopyroxene. Also Al(M1) vs. Na(M2) diagram for clinopyroxene (Fig. 5c) illustrates a positive correlation, with both elements decreasing from protogranular (-porphyroclastic) lherzolites to porphyroclastic and equigranular forms. That is to say, the jadeite substitution (M2NaM1AlM2Ca-1M1Mg-1) is found in decreasing amounts as microstructure changes from coarse towards finer grained lherzolites (Fernández-Roig and Galán, 2015).

**6. Temperature, pressure and deviatoric stress estimates**

Equilibrium T were estimated with the two-pyroxene (TBK) and the Ca-in-orthopyroxene (TCa-in-opx) thermometers (Brey and Kölher, 1990), both based on the transfer of the enstatite component between co-existing clinopyroxene and orthopyroxene. P of 15 kb was assumed for all estimates.Several pairs of analyses from neighbouring clinopyroxene and orthopyroxene crystals were used for the average values in Table 2. Where possible, estimates from porphyroclasts and neocrystals have been presented separately. Uncertainty (1s) is < 50 °C for TBK estimates and lower for the T Ca-in-opx values. Although estimates from both thermometers show a good positive correlation, those of T Ca-in-opx are somewhat higher than TBK values in most samples (Fig. 6). However, most T Ca-in-opx are within ± 50 ºC of TBK estimates. Bias is under 9%, except in two cases: orthopyroxene porphyroclasts in lherzolite CA.12.03 (22%) and the equigranular lherzolite BB.08.57 (18%). In the former case, bias for the neocrystals is much lower (5%) than for the porphyroclasts suggesting disequilibrium between the two different types of crystals (Fig. 6). The highest bias corresponds to TBK≤ 900 °C, as was noted by Brey and Köhler (1990). These T are mainly registered by porphyroclastic and equigranular lherzolites.

The highest T values are registered by protogranular harzburgites and the websterite (1171 to 1006 °C; TBK values). TBK  and TCa-in-opx estimates for the protogranular lherzolites range from 1057 down to 924 °C and from 1038 to 975 °C, respectively. These ranges for lherzolites with transitional protogranular-porphyroclastic (TBK : 996-931 °C, TCa-in-opx: 992-969 ºC) and porphyroclastic microstructures (TBK : 942-856 °C, TCa-in-opx: 1155-922 ºC) are somewhat lower. Where porphyroclast and neocrystal T can be calculated separately, they do not exhibit significant differences, except for lherzolite CA.12.03 where the porphyroclast TCa-in-opx value is significantly higher (1155 ºC, Table 2). For two of these samples, where analyses of orthopyroxene and clinopyroxene were performed with an enlarged beam for integrating mutual and/or spinel lamellae, TBK and especially T Ca-in-opx estimates are higher, and approximate those registered by protogranular peridotites and orthopyroxene porphyroclasts in CA.12.03 (Fig. 6; Table 2). Finally, two equigranular lherzolites register slightly lower TBK (959-766 °C) estimates than do porphyroclastic lherzolites, although TCa-in-opx (970-917 ºC) estimates are similar.

P estimates based on the Ca-exchange reaction between co-existing clinopyroxene and olivine (Köhler and Brey, 1990) were hampered by the dependence of this barometer on both temperature and the uncertainty of Ca analysis in olivine by EPM (O'Reilly et al., 1997). Only two harzburgites (BB.12.04: 1.13±0.10 GPa; BB.08.20: 0.83±0.17GPa), one protogranular lherzolite (BB.08.98: 1.51±0.21GPa), one porphyroclastic lherzolite ( SC.11.49: 0.67±0.10GPa) and one equigranular lherzolite (BB.08.38: 0.60±0.32 GPa) show consistent P values and low uncertainty (1s). SC.11.49 is a plagioclase-spinel lherzolite. Decreasing P from protogranular to porphyroclastic and equigranular lherzolites is consistent with the decreasing jadeite component of their clinopyroxene (Fig. 5c).

Finally, the average grain size of neocrystals in protogranular-porphyroclastic, porphyroclastic and equigranular lherzolites, measured using the EBSD maps, has been used to estimate the deformation stress based on the equation: $D\_{g}=Aσ^{-n}$ (Karato et al., 1980; Van der Wal et al., 1993), where Dg is the recrystallized grain size, σ is the differential stress, and A and n are empirical constants [Karato et al. (1980): A=8300; n=1.18; Van der Wal et al. (1993): A=15000; n=1.33]. The average grain size of neocrystals ranges from 960 µm in the protogranular-porphyroclastic lherzolite BB.12.13 to 360 µm in the porphyroclastic sample SC.11.16, which gives 6.1-14.1 MPa (Karato et al., 1980) or 8.0-16.7 MPa (Van der Wal et al. 1993) stress values.

In summary, more porphyroclastic and equigranular lherzolites were equilibrated at lower T and P and recrystallized during higher stress deformation than was the case for protogranular harzburgites, lherzolites and the sole websterite.

**7. Crystal preferred orientations**

*7.1. Olivine*

All the harzburgites, nine of the lherzolites and the websterite show the same deformation fabric for the olivine: they are characterized by point concentration of [010] parallel to Z and girdles of [100] and [001] axes normal to each other within the structural XY ("foliation") plane (Fig. 7). The main slip system is (010)[100]. BA indices for these samples range from 0.09 to 0.39 (Table 3; Fig. 8a). The highest main density (MD) corresponds mostly to [010] axis and [100] axis shows higher MD than [001] axis, whereas the reverse holds true in lherzolite CA.44.05 and websterite BB.12.17. This fabric is classified as [010] fiber or AG-type (Ben Ismaïl and Mainprice, 1998; Mainprice, 2007). Four other lherzolites are characterized by orthogonal point concentrations of the three olivine axes. The dominant slip system is also (010)[100]. BA indices range from 0.42 to 0.48. The highest MD corresponds to [010] axis in two samples (BB.08.59, SC.11.49) and to [100] and [001] in the other two cases (BB.08.57 and CA.12.03, respectively). This fabric is classified as orthorhombic (Tommasi et al., 1999) or A-type (Jung and Karato, 2001). Finally, only one lherzolite shows strong point concentration of [100] axis parallel to the structural X direction and girdles of [010] and [001] axes parallel to each other and normal to [100] (Fig. 7). This is characteristic of the [100]-fiber fabric or D-type (Karato et al., 2008). In this case, the dominant slip system is {0kl}[100], BA index is 0.78, [100] axis shows the highest MD and the MD of [001] axis is slightly higher than that of [010] axis. In spite of this numerical classification of deformation fabrics based on the BA index, there are transient types between AG-, A- and D-types in some lherzolites (Fig. 7, Table 3). For instance, CA.44.05 can be considered transitional AG-A type because both [100] and [001] axes tend to point concentrations. Other cases are transitional AG-D type (BB.08.38, SC.11.52, SC.11.54) since [001] axis tends to be normal to the structural XY ("foliation") plane instead of being within it (Fig. 7). Finally, lherzolite BB.08.57 can be considered transitional A-D type for the same reason: [001] axis tends to be normal to the "foliation" plane. It should be noted that the distribution of [100] and [001] axes is bimodal in two lherzolites with AG-D type fabric (BB.08.38, SC.11.52). In these two samples, [100] distribution is characterized by two mutually orthogonal maxima within the "foliation" plane, whereas [001] distribution is characterized by one maximum aligned with [100] axis and the other with [010] axis, that is to say normal to the "foliation" plane (Fig. 7). Finally, only in harzburgite CA.12.02 the number of grains (<100) per thin section was not sufficient for reliable statistical analysis of the fabric strength (Ben Ismaïl and Maimprice, 1998). This sample is not contoured in Fig. 7. For the remaining samples, the fabric strength or J index is variable. The highest values (10.12-5.43) correspond to the harzburgites and lherzolites which have well defined AG-type fabric. The websterite also follows this pattern. J indices for the other lherzolites range from 4.69 (A-type fabric) to 1.91 (transitional AG-D-type fabric) (Table 3).

J indices vs. BA indices display scatter, although negative correlation is observed overall (Fig. 8a). No clear relationship exists between BA indices and the microstructure types (not represented), although most BA values> 0.40 correspond to porphyroclastic lherzolites. However, a better relationship is displayed between J indices and the microstructure types: J indices decrease from protogranular harzburgites and lherzolites towards porphyroclastic and equigranular lherzolites (Fig. 8b), which overlap both. The relationships between J indices and other variables are as follows: (1) there is a positive correlation with the olivine mode in peridotites (Fig. 9a) and a negative correlation with the clinopyroxene mode (not represented), although porphyroclastic and equigranular lherzolites display scatter; (2) correlation with Mg# of clinopyroxene is also positive for the protogranular (-porphyroclastic) lherzolites and harzburgites, whereas most equigranular and porphyroclastic lherzolites do not follow this trend (Fig. 9b); (3) in contrast, correlation with Na (Fig. 9c) (and Al) atoms per formula unit in clinopyroxene is broadly negative for protogranular (-porphyroclastic) lherzolites and harzburgites, but among lherzolites the concentration of both elements drops significantly from those with protogranular miscrostructure to most porphyroclastic and equigranular lherzolites; and (4), there is a roughly positive correlation with equilibrium T whether estimated with the TBK or TCa-in-Opx thermometers (Brey and Köhler, 1990) (Fig. 9d); porphyroclastic and equigranular lherzolites register the lowest J indices and equilibrium temperatures.

*7.2. Orthopyroxene*

In six lherzolites, one harzburgite and the websterite (Table 3), the analysis of orthopyroxene CPO was hindered by the paucity of grains (<100 grains per thin section). These are not contoured in Fig. 7. Nevertheless, the deformation fabric is well defined and coherent with olivine fabric in eight xenoliths (44% of the total): all are characterized by the alignment of orthopyroxene [001] and olivine [100] axes. Among these orthopyroxene CPO, three different types were identified according to Jung et al. (2010): AC-type in lherzolites SC.11.52, BB.12.01 and harzburgite CA.12.02, characterized by the alignment of [100] axis subnormal to the "foliation" plane and of [001] axis subparallel to the "lineation"; BC-type in lherzolites BB.08.98, BB.08.59, SC.11.49 and harzburgite BB.08.20, differing from the previous type by the alignment of the [010] axis subnormal to the "foliation" plane; and ABC-type in lherzolite BB.08.101, defined by girdles of both [100] and [010] axes aligned subnormal to the "foliation" plane. According to the nomenclature of Jung et al. (2010), the first and last capital letters represent the slip plane and slip direction, respectively (Table 3). In one more lherzolite (BB.12.13) and the websterite (BB.12.17) the orthopyroxene fabric is less well-defined, but these samples do show coherent fabrics with olivine. In seven other lherzolites (CA.12.03, CA.44.05, BB.08.57, BB.08.80, BB.08.38, SC.11.16, SC.11.54) and one harzburgite (BB.12.04) (44% of the total) (Table 3), the olivine and orthopyroxene fabrics are incoherent. Among these samples, lherzolite CA.44.05 stands out because orthopyroxene [010] and olivine [100] axes align, therefore, the slip direction for orthopyroxene would be [010]. Lherzolite SC.11.16 and harzburgite BB.12.04 also stand out because [100] axes of both minerals are subparallels; [001] axis of both minerals are also aligned and normal to the olivine [100] axis in SC.11.16 (Fig. 7). For those samples where the analysis of orthopyroxene CPO was statistically reliable (Fig. 7; Table 3), values of J indices are similar to, or higher than, those found in lherzolite olivine, whereas the reverse holds true in harzburgites.

*7.3. Clinopyroxene*

The number of clinopyroxene crystals per thin section is <100 in all harzburgites and four of the lherzolites, the pole figures of which are not contoured (Fig. 7). Moreover, clinopyroxene of one harzburgite with rare crystals of this mineral was not indexed. In spite of this paucity of grains, 44% of the samples have clinopyroxe [010] and [001] axes normal and parallel, respectively, to olivine [100] axis, ie., both minerals show coherent deformation fabrics (Table 3). The dominant slip system would be {110}[001] (Bascou et al., 2002; Zhang et al., 2006). Moreover, in these samples (lherzolites BB.08.101, BB.12.13, BB.08.98, BB.12.01, SC.11.49, SC.11.52, harzburgite BB.08.20 and the websterite BB.12.17) both pyroxenes are coaxial (Table 3; Fig. 7). There are nine (50%) incoherent samples (lherzolites BB.08.59, CA.12.03, CA.44.05, BB.08.57, BB.08.80, BB.08.38, SC.11.16, SC.11.54 and harzburgite BB.12.04), which are most of them porphyroclastic and equigranular lherzolites. It is worth noting that in five of these incoherent samples, the [100] axes of both olivine and clinopyroxene align (Fig. 7). The slip direction of clinopyroxene in these samples would be [100]. However, very few dislocations have Burgers vector b=[100] (Zhang et al., 2006). For instance, the (010)[100] and (001)[100] slip systems have been rarely observed (Philippot and Van Roermund, 1992; Godard and Van Roermund, 1995; Palasse et al., 2012). Clinopyroxene fabric types (Table 3) have been classified numerically according to LS index values as follows: S-type or flattening fabric (0.19-0.27) for one harzburgite (BB.08.20) and one lherzolite (BB.12.01), where [001] axis tends to form a girdle normal to the (010) pole maximums; L-type or constriction fabric (0.71-0.86) for three lherzolites (CA.12.03, BB.08.57, SC.11.52), where [001] axis tends to concentrate and (010) poles tend to form a girdle orthogonal to [001] axis; and the intermediate SL-type (0.32-0.64) for most lherzolites and the websterite (Helmstadt et al., 1972; Zhang et al., 2006) (Table 3). In samples where the number of both pyroxene crystals per thin section is > 100, J values for clinopyroxene are higher than those of orthopyroxene in each sample (Table 3). The highest values of J index of both pyroxenes are always associated with protogranular and protogranular-porphyroclastic rocks and the lowest with porphyroclastic lherzolites. Finally, LS index is broadly negatively and positively correlated to olivine J index and BA index, respectively (Fig. 10a-b).

**8. Misorientations**

Deformation by dislocation creep causes not only characteristic CPO but also distinct misorientations, ie., changes in crystallographic orientation between two points. Misorientations can occur within grains or across low-angle grain boundaries (ie., subgrains). Since the orientations of subgrain boundaries and the rotation axes of subgrains depend on the dislocations which accumulate within the grain or form the boundaries (Frank, 1950; Amelinckx and Dekeyser; 1959), they can be used to infer the active slip systems during deformation (Lloyd et al., 1991). To this end, inverse pole figures (IPFs) of all rotation axes (<15 °) within all olivine crystal of the EBSD maps were considered (Fig. 7), along with the orientation of the trace of subgrain boundaries of selected olivine crystals (Fig. 11) in specific xenoliths. For instance, the isolated olivine crystal from the protogranular lherzolite BB.08.101 (Figs. 11a-c and 3.1Eb in supplementary electronic material) has subgrain boundaries with a trace subnormal to the [100] axis. This implies that the subgrain walls are mainly tilt boundaries composed of dislocations slipping in the [100] direction. Accordingly, the corresponding IPF (Fig. 11d) illustrates that the rotation axes are distributed between [001] and [010] axes, which implies that the subgrain walls were mainly formed by dislocations of the {0kl}[100] slip systems or "pencil glide". Also, stronger cluster of rotation axes around [010] than around [001] axis (Fig. 11d) suggests the predominance of the (001)[100] slip system over the (010)[100]. Therefore, if deformation was mainly accommodated by "pencil glide" dislocations, the assumption that olivine [100] maxima are parallel to the lineation would be corroborated. In the remaining xenoliths, the activation of the {0kl}[100] slip systems is consistent with the distribution of the rotation axes between [001] and [010] axes in the IPFs (Fig. 7). There are three exceptions (lherzolites SC.11.49, SC.11.52, SC.11.54) with olivine rotation axes very scattered (they are not contoured in Fig. 7). In the other samples, we observe similar densities for the distribution of [001] and [010] rotation axes in three samples (BB.08.59, BB.08.38, CA.12.02), rare [001] maxima (lherzolite BB.08.98) and [010] maxima in eleven samples (61% of the total). The last distribution suggests the predominance of the (001)[100] slip system over the (010)[100] in more than half of the xenoliths, contradicting the dominant slip system deduced from the AG- and A-type olivine deformation fabrics. This partial disagreement has been observed in other peridotites (Soustelle et al., 2010; Falus et al., 2011; Kaczmarek and Tommasi, 2011; Zaffarana et al., 2014) which leads to the conclusion that dislocations in subgrain boundaries may not be completely representative of the slip system activity. Alternatively, the predominance of the olivine [010] rotation axes could also imply the activation of the (100)[001] slip system (de Kloe, 2001), the presence of which is supported by the occasional presence of traces of subgrain boundaries normal to [001] maxima (Fig. 8.1E in supplementary electronic material).

**9. Seismic properties**

Calculated seismic properties [velocity and anisotropy of the P (Vp, A) and S-waves (Vs1, Vs2, AVs)] are summarized in Table 3.4E in supplementary electronic material and some of them illustrated in Fig. 12.

Vp patterns are similar in most samples: Vp distribution displays a well-defined maximum aligned with the maximum of olivine [100] axis and a minimum subparallel to the maximum of olivine [010] axis. Vp maximum ranges from 8.78 to 8.33 km/s, Vp minimum from 8.21 to 7.84 and AVp from 10.2 to 2.60%. AVp> 6% are found in protogranular harzburgites, protogranular lherzolites and the websterite, whereas lower values correspond to porphyroclastic and equigranular lherzolites, although there are a few exceptions (eg., CA.12.03).

S-wave birrefringence patterns are more heterogeneous. In most samples, the highest AVs values tend to form one maximum (AVs max) (eg., CA.12.03, CA.44.05, BB.08.20), two maxima (eg., BB.08.98, BB.08.57, BB.08.38, CA.12.02), or a girdle (BB.08.101, BB.12.13, BB.12.01, BB.08.80, SC.11.54, CA.12.02, BB.12.04, SC.11.49) within the structural XY plane ("foliation" defined from the CPO). In the first and the third case, the maximum is normal to the olivine [100] axis ("lineation" defined from the CPO), whereas in the second case, the two maxima are at 45° of the "lineation".Three other lherzolites (BB.08.59, SC.11.16, SC.11.52) are noteworthy because AVs maximum values tend to form a girdle normal to the "foliation" plane andwith AVs max only normal to the "lineation".AVs maximum values range from 7.95 to 2.19% (Table 3.4E), with values> 4% being registered by most protogranular harzburgites and lherzolites and lower values by porphyroclastic, equigranular lherzolites and the websterite. The lowest AVs value (AVs min) is generally at ca. 45 ° of both olivine [100] and [010] maxima, and more rarely, parallel (SC.11.16) or normal (BB.08.38, BB.12.17) to the olivine [100] axis.

Finally, S1 tends to be polarized in a plane that contains the propagation direction and the olivine [100] maxima except for SC.11.16 where S1 is polarized in a direction subnormal to the olivine [100] axis (Fig. 12).

Fig. 12 and Table 3.4E also include the estimates for the average seismic properties of the local SCLM using samples weighted according to their abundance [viz., 40% harzburgites, 34% protogranular (-porphyroclastic) lherzolites, and 26% porphyroclastic and equigranular lherzolites, according to section 3], their respective simplified mode, spinel excluded, and the CPO data in a common standard orientation (the one used for displaying the CPO in Fig. 7). This method would provide a maximum estimate for the the local SCLM anisotropy since coherent deformation (similar orientation of foliation and lineation) for the whole mantle column is assumed.The Vp and S-wave birrefringence patterns are similar to those of most protogranular xenoliths.

**10. Discussion**

*10.1. The origin of olivine CPO: relationships with pyroxene CPO, microstructures and deformation*

The results of this study indicate that the dominant olivine deformation fabric in mantle xenoliths from NE Spain is [010] fiber (or AG-type) (present in 72% of the total studied samples; Fig. 7; Table 3). This CPO type also shows the highest fabric strength (Fig. 8a, 10b). This fact, along with the coherent orthopyroxene CPO in most of these xenoliths (62%; Table 3), points to deformation by dislocation creep at high T, dry conditions and low stress (Carter and Avé Lallemant, 1970; Tommasi et al., 1999). The AG-type fabric is thought to be less common than the A- and D-types in mantle rocks (Mainprice, 2007) but it has been found in different samples of both orogenic peridotitic massifs (Le Roux et al., 2008; Soustelle et al., 2009; Kaczmareck and Tommasi, 2011; Higgie and Tommasi, 2014) and mantle xenoliths: cratonic (Baptiste et al., 2012), off-craton (Tommasi et al., 2008; Hidas et al., 2007; Kovacs et al., 2012) and in oceanic within-plate context (Bascou et al, 2008). In some of these examples, AG-type is the dominant deformation fabric, as in Ronda (Betic orogenic belt, S Spain) and Lanzo massifs (western Alps; Italy) (Soustelle et al., 2009; Higgie and Tommasi, 2014), in the mantle xenoliths of Tok (SE Siberia) (Tommasi et al., 2008) and in the Kerguelen islands (Indian ocean) (Bascou et al., 2008). However the AG-type fabric may be subordinate to other types, as in the nearby peridotitic massif of Lherz (Eastern Pyrenees, France), where the D-type is dominant (Le Roux et al., 2008). The other olivine fabrics in the studied SCLM are A- (22%) and D-type (6%) (Fig. 10b) which share [100] as the main glide direction with the AG-type.

Hypotheses for the origin of AG-type fabric are: (1) deformation by axial shortening or transpression (Nicolas et al., 1973; Tommasi et al., 1999; Vauchez et al., 2000); (2) melt assisted shear deformation at high T and P (Holtzman et al., 2003; Le Roux et al., 2008; Higgie and Tommasi, 2014); (3) simultaneous activation of the (010)[100] and (010)[001] slip systems under high stress and/or high water content or high P conditions (Tommasi et al., 2000; Mainprice et al., 2005; Vauchez et al., 2005; Hidas et al., 2007; Bascou et al., 2011; Jung et al., 2014); and (4), static recrystallization (annealing) (Tommasi et al., 2008; Zaffarana et al., 2014). Hypothesis (1) is characterized by girdle distribution of both olivine [100] and orthopyroxene [001] axes (Tommasi et al., 2006; Le Roux et al., 2008; Bascou et al. , 2008; Higgie and Tommasi, 2014), which is only observed here in three lherzolites (BB.08.101, BB.08.98, BB.12.01) and two harzburgites (CA.12.02, BB.08.20) (Fig. 7). However, it is worth noting that deformation by axial shortening and/or simple shear is consistent with the values of the clinopyroxene LS index for most protogranular (-porphyroclastic) lherzolites, harzburgites and the sole websterite, all of them with AG-type fabric and coherent clinopyroxene fabrics (Fig. 10b). Hypothesis (2) posits that melt assisted shear deformation, either during mantle partial melting (Vauchez and Garrido, 2001; Soustelle et al., 2009) or during percolation of basaltic melts (Le Roux et al., 2008; Higgie and Tommasi, 2014), would account for AG-type deformation fabric. In the former case, the olivine mode would be increased due to incongruent melting of orthopyroxene. Local orthopyroxene replacement could also result from olivine crystallization via reaction of the peridotite with an orthopyroxene-undersaturated percolating melt (Kelemen et al., 1998; Dijkstra et al., 2003). Opposite, orthopyroxene would be increased at the expense of clinopyroxene and/or olivine if the percolating melt were SiO2 (or orthopyroxene) saturated (Kelemen et al., 1992; 1998; Dijkstra et al., 2003), or olivine would be decreased and both pyroxenes±spinel, garnet or plagioclase be increased, if refertilization reactions had taken place during the basaltic percolation (Lenoir et al., 2001; Dijkstra et al., 2003; Le Roux et al., 2007; Bodinier et al., 2008; Kaczmarek and Müntener, 2008; Rampone et al., 2010; Johanesen and Platt, 2015). However, the following findings do not confirm the hypothesis that pervasive refertilization would have enhanced the development of the AG-type fabric in the protogranular (-porphyroclastic) lherzolites and harzburgites: (i) positive correlations of the J index with respect to both the olivine mode and the Mg# of clinopyroxene (Figs. 9a-b); (ii) negative correlations of the J index vs. the clinopyroxene mode (not represented) and the Al and Na (Fig. 9c) content of this mineral; and finally (iii), the scattered projection of the J index vs. orthopyroxene mode (not represented). Other microstructural features evidence of melt-rock reaction are: (i) incoherence between olivine and clinopyroxene CPO (Soustelle et al., 2010; Zaffarana et al., 2014); (ii) interpenetrating olivine-pyroxene boundaries (Dijkstra et al., 2003; Soustelle et al., 2009; Higgie and Tommasi, 2014; Zaffarana et al., 2014); and (iii), olivine inclusions in pyroxenes (Rampone et al., 2010; Soustelle et al., 2010; Zaffarana et al., 2014). Evidence (i) is present here in four lherzolites (CA.44.05, BB.08.80, BB.08.38, SC.11.54), three of which show transitional AG-A and AG-D olivine deformation fabrics (Fig. 7, Table 3). Nevertheless, these incoherent samples also show the lowest J indices (Figs. 8a-c, 10a), suggesting that if refertilization had assisted shear deformation, this process would have decreased the J index and triggered the transformation of the AG-type fabric into the other types, which contradicts hypothesis (2). Finally, sporadic evidence of (ii) and (iii) can be seen in two lherzolites (BB.08.98, BB.12.01) and one harzburgite (CA.12.02) with olivine AG-type fabric; all of them show occasional orthopyroxene crystals with olivine embayment and/or olivine inclusions (Figs. 4a, f). The fact that both microstructural features (ii and iii) are seen in the same sample (Fig. 4f) suggests that the percolating melt was at or just below orthopyroxene saturation (ie., the reaction olivine+ high-Si melt <=> orthopyroxene+ low-Si melt would be close to equilibrium). Nevertheless, these occasional microstructures in a few samples do not prove that reactive percolation was widespread and responsible for the dominant olivine AG-type fabric in most xenoliths. Refertilization of earlier refractory harzburgites giving rise to spinel lherzolites via the percolation of MORB-type basaltic melts, was posited for the neighbouring Lherz massif (Le Roux et al., 2007). These refertilized lherzolites show clinopyroxene with LREE depletion and DMM mantle Sr and Nd isotopes, all of which are geochemical features of most lherzolite xenoliths from the CVZ (Galán and Oliveras, 2014). However, it is worth noting that there are significant differences between these two mantle sections: websterites are less frequent in the SCLM of NE Spain than in the Lherz massif; harzburgite xenoliths do not show porphyroclastic microstructure as often as Lherz harzburgites do; lherzolite xenoliths rarely display cm-scale websterite layering such as that seen in the Lherz massif examples; and lastly, olivine AG-type deformation fabric characterizes most protogranular harzburgite and lherzolite xenoliths from the CVZ, whereas both the weakly percolated harzburgites and highly deformed refertilized lherzolites from Lherz have olivine [100]-fiber or D-type fabric and only weakly deformed refertilized lherzolites show [010]-fiber or AG-type fabric (Le Roux et al., 2008). As regards hypothesis (3), simultaneous activation of the slip systems (010)[100] and (010)[001] is not consistent with the P -T equilibrium conditions of these rocks within the spinel lherzolite field. These are considerably lower than those considered responsible for the activation of [001] glide in experimental studies (Couvy et al., 2004; Mainprice et al., 2005; Jung et al., 2009; Ohuchi et al., 2011). However, the presence of H2O could have also favoured the simultaneous activation of both slip systems at lower P (Jung et al., 2014), but unless H2O were present in anhydrous minerals, this option would be unlikely since hydrous minerals are rare or absent. Predominance of the (010)[100] slip system would be also consistent with the fact that the MD of olivine [100] axis is higher than that of [001] axis in most xenoliths with AG-type fabric. Finally, static recrystallization or annealing (hypothesis 4) seems to be compatible with: (i) the coarse-grained microstructure characteristic of protogranular peridotites and the websterite, most of them with AG-type fabric (Figs. 3a-c, 7); (ii) the straight or slightly curved grain boundaries and interfacial angles of 120 ° among olivine crystals (Figs. 3a,c); and (iii), higher T estimates for these rocks (Figs. 6 and 9d). Grain growth by grain boundary area reduction (GBAR), most likely favoured by diffusion processes, would have taken place subsequent to other accommodating mechanisms of deformation based on dislocation creep, such as subgrain rotation (SR) and GBM recrystallization. In summary, the annealing hypothesis (4) is the most credible explanation for the origin of olivine AG-type deformation fabric in the case study. However the possibility of the AG-type fabric being inherited from earlier deformation stage(s) caused by axial shortening or transpression (hypothesis 1) cannot be completely ruled out, since annealing does not necessarily modify previous fabric (Vauchez and Garrido, 2001; Tommasi et al., 2008). Also, the presence of melt (hypothesis 2) could have been a factor in deformation and/or annealing events. Taking into account that AG-type olivine fabric exists in harzburgites, lherzolites and the websterite, and since the websterites were interpreted as cumulates from percolating alkaline mafic silicate melts causing cryptic metasomatism in most harzburgites and a few lherzolites (Galán et al., 2008), it is likely that annealing was stimulated by the percolation of low fractions of these metasomatic melts and derivatives.

The other olivine deformation fabrics, A-(orthorhombic) and D- ([100]-fiber) types, are mainly seen in finer-grained protogranular-porphyroclastic, porphyroclastic and equigranular lherzolites (Figs. 7, 8a, 10b), all of them with low fabric strength (Figs. 8a, Table 3) and, except for one sample, with incoherent clinopyroxene fabrics (Fig. 10b). According to experimental data (Jung and Karato, 2001), both fabrics are developed by simple shear deformation at dry conditions, low to moderate stress (A-type), high stress (D-Type), and high T, although lower for D-type than for A-type. Numerical simulations of CPO by Tommasi et al. (1999) also indicate deformation by pure shear or plane transpression for the A-type and transtension for the D-type, which agrees with deformation regimes deduced from most clinopyroxene LS index (Fig. 10b). Taking into account that both deformation fabrics are found in finer-grained peridotites, most of them equilibrated at lower T and P than those with AG-type (Figs. 5c, 6, 9d, Table 2), and that they grade to the AG- type in four samples (Table 3; Fig. 7), we deduce that the AG-type was an earlier deformation fabric later transformed into A- and D-types by changes in the deformation regime that happened at higher stress and decreasing T and P. The same conclusion was after the study of mantle xenoliths displaying similar characteristics found in the southeastern Carpathians and San Quintin (Baja California) (Falus et al., 2008; Palasse et al., 2012 and references therein) and in extensional shear zones of orogenic peridotitic massifs (Kaczmarek and Tommasi, 2011). Deformation at decreasing T and P is consistent with: (i) higher T estimates from original compositions of orthopyroxene and clinopyroxene porphyroclasts than from neocrystals (Table 2, Fig. 9c); (ii) clinopyroxene with lower jadeite component (Fig. 5c, 9c) found in porphyroclastic and equigranular lherzoliteswith A- and D-type fabrics**;** and (iii), possible activation of the olivine (100)[001] slip system along with the dominant {0kl}[100], as deduced from misorientations in lherzolites with A- and D-type fabrics and transitional of these to AG-type (eg., BB.08.59, CA.12.03, CA.44.05, BB.08.57, SC.11.16 in Figs. 7 and Fig. 8.1E in electronic supplementary material) (Carter and Avé Lallemant, 1970; Tommasi et al., 2000 and references therein; Kaczmarek and Reddy, 2013). It is worth noting that clinopyroxene and olivine fabrics in most of these samples are incoherent because olivine [100] axis and clinopyroxene [001] axis are orthogonal and both olivine and clinopyroxene [100] and/or [001] are coaxial (eg., BB.08.59, CA.12.03, BB.08.57, SC.11.16, Table 3). Therefore, the clinopyroxene slip direction may have been [100]. Although very few dislocations in this mineral have Burgers vector b= [100], they have been observed in omphacite from eclogites deformed at low T (Philippot and Van Roermund, 1992; Godard and Van Roermund, 1995) and in clinopyroxene from mantle xenoliths of San Quintin (Baja California), where these dislocations were also related to low T conditions (Palasse et al. 2012). Alternatively, it is likely that during plastic deformation at increasing stress and decreasing T, the olivine [001] slip direction had been also activated (Kaczmarek and Reddy, 2013), as it is documented from the trace of subgrain boundaries (Fig. 8.1E in supplementary electronic material) and from stronger cluster of olivine rotation axes around [010] in most of these lherzolites (Figs. 7). Accordingly the MD of olivine [001] axis is only slightly lower or even higher than the MD of [100] axis in these samples (eg., CA.12.03, CA.44.05; Fig. 7). Therefore, if the olivine [001] slip direction had been also activated, the incoherence between olivine and pyroxene CPO could be apparent**.** Deformation based on dislocation creep would have been accommodated mainly by SR and GBM recrystallization most likely assisted by diffusion processes. However, microstructural evidence of accommodation by grain boundary sliding has not been observed, eg., four-grain junctions (Lee et al., 2002). On the other hand, it has been suggested that decreasing J indices and incoherent olivine and pyroxene CPO are due to late to post-kinematic melt-rock reaction, causing olivine consumption and the formation of new pyroxenes±spinel±garnet± plagioclase via refertilization processes (Soustelle et al., 2009; 2010; Zaffarana et al., 2014). The presence of reactive-melt flow would have enhanced diffusion processes that contribute to weaken the olivine CPO (Soustelle et al., 2009). In the samples discussed here, when pyroxene J index in porphyroclastic and equigranular lherzolites could be estimated, it is even higher than that of olivine and both are positively correlated (Table 3). These results seem to contradict the late to post-kinematic character of the melt-rock reaction process. Moreover, microstructural evidence of olivine being replaced by clinopyroxene was not observed: grain boundaries between neocrystals of both phases are straight or slightly curved; only porphyroclastic lherzolite CA.12.03 shows occasional orthopyroxene porphyroclasts with both olivine embayment and rare rounded olivine inclusions. It is worth noting that both porphyroclastic lherzolite CA.12.03 (with A-type fabric) and equigranular lherzolite BB.08.57 (A-D type fabric) provide the lowest T estimates, the highest bias between TBK and TCa-in-opx values (Fig. 6) and the highest difference between porphyroclast and neocrystal T in the case of CA.12.03. These thermometric results suggest chemical disequilibrium and cast doubt on the existence of more effective diffusion enhanced by the presence of significant amount of melt during the development of the olivine A- and D-type fabrics. In spite of these drawbacks, the fact that some of these lherzolites are among the most fertile (Fig. 5a-b), with the highest clinopyroxene and the lowest olivine mode (eg., BB.08.57, SC.11.49 in Table 3.1E; Fig. 9a), points in favour of likely melt-rock reaction processes leading to refertilization. Nevertheless, these processes could have been pre-kinematic with respect to the development of olivine A- and D-type fabrics and difficult to recognize from the major element composition of minerals ("stealth metasomatism" of O'Reilly and Griffin, 2013): clinopyroxene of these samples does not show the lowest Mg# but a wide range of values overlapping that of the other lherzolites (Fig. 9b).

*10.2. CPOs, seismic anisotropy and modal compositions*

The seismic anisotropy of the upper mantle is mainly a function of intrinsic elastic anisotropy of the rock forming minerals, mineral mode, deformation fabric types and strength (Nicolas and Christensen, 1987; Ben Ismaïl and Mainprice, 1998; Mainprice and Silver, 1993; Mainprice et al., 2000).

Relationships between seismic anisotropy patterns and olivine deformation fabrics of the studied xenoliths are as follows: AG-and A-type fabrics share similar seismic anisotropy patterns; the slowest P-waves are normal to the structural XY ("foliation") plane, while the AVs min is at ≈45 °of the maximum concentration of olivine [100] axis ("lineation"). However, maximum values of Vp and AVs within the "foliation" plane tend to be more dispersed for AG-type (eg., BB.12.01, BB.08.38) than for A-type fabric (eg., CA.12.03) (Fig. 12). SC.11.16, with olivine D-type symmetry (Fig. 12), shows highly distinctive seismic properties, where the highest AVs values and the S1 polarization plane are orthogonal to the "foliation" (XY) plane, whereas in the other samples they are parallel to it. This is explained by the fact that the olivine [001] axis clearly defines a girdle orthogonal to the olivine [100] axis in this xenolith (Fig. 7).

Moreover, there are positive correlations of AVp and AVs with respect to J indices (Figs. 10.2E a-b in electronic supplementary material), which means that the dynamic recrystallization observed in porphyroclastic and equigranular lherzolites probably decreased seismic anisotropy. Finally, there are general positive and negative correlations of AVp vs. olivine mode and orthopyroxene mode (Figs. 10.2E c-d in supplementary electronic material), respectively. No relationship has been observed with clinopyroxene mode. Variation of AVs with respect to mineral mode is similar, but there is a broadly negative correlation with the clinopyroxene mode (not shown).

In summary, our results indicate that the average seismic anisotropy of the SCLM beneath the CVZ is within the range of less fertile lherzolites and refractory harzburgites (Fig. 10.1E) which have most of them protogranular microstructure and stronger olivine deformation fabric than dynamically recrystallized porphyroclastic and equigranular lherzolites.

*10.3. Seismic anisotropy from CPO versus seismic data for the SCLM in the NE Spain*

Contrasting the average seismic anisotropy of the SCLM estimated from the studied mantle xenoliths (Table 3.1E) with seismic data may help to assess the thickness of the lithosphere and the deformation regime in the area. Shear-wave splitting using SKS phases is one of the most effective seismic methodologies to explore the uppermost mantle anisotropy. However, this method does not provide information about the depth distribution of the anisotropy. The fast polarization directions (FDPs) deduced from SKS-wave splitting are expected to be parallel to the mantle flow direction (Silver and Chan, 1988) and related to the CPO of olivine, which is the main cause of mantle anisotropy (Nicolas and Christensen, 1987). Barruol et al. (1998) and more recently Diaz et al. (2015) deduced the FPDs around N 120 °E and delay time (δt) range of 0.6-1.1 s for stations nearby La Garrotxa sub-zone.

Below the CVZ, AVs estimates range from 2.19 to 7.95%, therefore, the SCLM cannot be discarded in the interpretation of the SKS-wave splitting. However, assessing the SCLM contribution is hindered by the fact that mantle xenoliths provide no information on their original position and orientation. Therefore, we have assumed that Vp max, the highest AVs values and the S1 polarization plane of most samples match the FPD deduced from SKS-wave splitting in the area. That is to say, the olivine [100] axis is generally oriented ≈ N120 °E as the FPD, defining a "lineation" within the "foliation" plane, the orientation of which is unknown.

In order to constrain the orientation of this "foliation" plane and to estimate the thickness of the anisotropic layer that cause the observed delay times, we have used the following equation by Mainprice and Silver (1993), following the method used by Baptiste and Tommasi (2014):

$δt=L×\frac{AVs}{<Vs>}$ (1)

where δt is the local S-wave splitting delay time, L is the SKS-wave effective path length, AVs=[Vs max-Vs min]/<Vs> in a particular direction and <Vs>=[Vs max+Vs min]/2 for the average SCLM. Assuming that δt is mainly due to the SCLM anisotropy (Mainprice and Silver, 1993), with a smaller contribution made by the crust (0.1 s per 10 km; Barruol and Mainprice, 1993), the thickness of which is 30-26km in this area (Gallart et al., 1984, 1991; Dañobeitia et al., 1992; Galán et al., 2011), then L is the SKS effective path length within the SCLM, ie., its thickness. For the estimates, two end-member orientations were considered: (i) vertical "foliation" and horizontal olivine [100] maxima or "lineation" and (ii) horizontal "lineation" and "foliation" (Fig. 13). SKS-wave splitting δt range (0.6-1.1 s) was taken from Díaz et al. (2015). L estimates so determined (Fig. 13) were compared to existing data on the lithosphere-asthenosphere limit (LAL) in this area. Setting aside the crustal thickness, the thickness average is 51 km for the SCLM in case (i) (Fig. 13a). This is consistent with the depth estimate for the lithosphere asthenosphere limit (LAL) (70-75 km), using gravity and geoid modelling in the shoreline of the Valencia Trough (Ayala et al., 2003) and thermobarometric data in mantle xenoliths from the CVZ (Galán et al., 2011). By contrast, the SCLM thickness average in case (ii) (Fig. 13b) is much thicker (251 km). Therefore, it is inferred that most of the measured seismic anisotropy would be explained by the lithospheric contribution, if the SCLM fabrics record mainly transpression and transtensional deformation. This inference means that the fabrics have the olivine [100] maxima ("lineation") horizontal and orientated ca. N120 °E, like that of the FPD, and the olivine [001] maxima (structural Y direction) vertical. If this were not the case, the seismological data may also be explained by other orientations of the SCLM fabric and a larger contribution of an asthenospheric deformation fabric (Diaz et al., 2015), as has been suggested for other European zones (Klébetz et al., 2015).

*10.4. Deformation of the lithospheric mantle in NE Spain*

Mineral CPO and resulting anisotropy in the SCLM could be the result of the last significant tectonic event, of the present asthenosphere dynamic flow, or of a combination of both factors (Savage, 1999; Vauchez et al., 2012). The interpretations of the data in the area studied are contradictory. Firstly, based on SKS-wave splitting measurements, Barruol et al. (1998) proposed that the local N120 °E FPD was mainly of lithospheric origin: subverticalE-W to NW-SE Hercynian fabrics at the base of the lithosphere would have been overprinted by similarly orientated but shallower and moderately dipping foliation related to the opening of the ECRIS, during the Miocene. For Barruol et al. (1998), the ECRIS fabrics would not have dominated the SKS signal. Secondly, Barruol et al. (2004) reconsidered the FPD in NE Spain (ie., the mantle flow) in the context of those observed in the South of the Massif Central and SE France, conflating both groups and interpreting their similar characteristics as due to a common Neogene horizontal asthenospheric flow (17-10 Ma) from the hotspot in the South of the Massif Central towards the SE France. This flow would have been induced by the previous sinking of the Apenninic slab (20-17 Ma) towards the NWbeneath the Corsica-Sardinia lithospheric block, and its subsequent eastward roll-back causing the opening of the Liguro-Provençal basin (17-10 Ma). As a result, the vacuum left behind by the retreat of the slab would have been filled by the regional asthenospheric flow from the Massif Central that may be preserved since the Neogene. The origin of the anisotropy would, therefore, be related to a present-day frozen mantle flow. However, more recently, Díaz et al. (2015) proposed that the general E-W FPD in the Iberian mantle was mainly due to the current asthenospheric flow related to the absolute plate motion (APM). In spite of this general conclusion, these authors admit the existence of local FPD anomalies which is due to (i) previous "frozen-in" anisotropy caused by Hercynian and/or Alpine deformation and to (ii) the influence of the sinking and roll-back of the Apenninic slab in NE Spain**,** following Barruol et al. (2004).

Both hypothesis by Barruol et al. (2004) and Diaz et al. (2015) are consistent with subhorizontal foliation in the SCLM of NE Spain. However if this were to be the case, the resulting low vertical anisotropy would require a much thicker lithosphere (Fig. 13b) than that predicted by available data from gravity and geoid modelling (Ayala et al., 2003). Taking these anomalies and also the fact that a mantle is usually deformed coherently with its crust, especially by strike slip faults of lithospheric scale and in transform plate boundaries (Vauchez et al., 2012), we must conclude that the results of the current study are more consistent with the first model by Barruol et al. (1998). On the one hand, the Hercynian orogenesis, which is the most significant in the area, is characterized by late W-E to NW-SE orientated strike-slip faults, compatible with transpressional deformation, in the nearby Pyrenees (Carreras and Capellà, 1998; Carreras 2001). These faults could have deformed not only the crystalline crustal basement, but also caused subvertical foliation in the underlying SCLM which would have resulted in AG-type olivine deformation fabric. Subsequent annealing, probably aided by asthenospheric upwelling during Late Hercynian decompression and Permian and Cretaceous rifting episodes, would have preserved this fabric unchanged. The presence of mantle derived rocks, such as Late Hercynian hornblendites and hornblende gabbros (Butjosa et al., 2013), Permian calcoalcaline to alkaline lamprophyres (Losantos et al., 2000; Enrique, 2009; Ubide et al. 2010) and Cretaceous alkaline lamprophyres (Ubide et al., 2012; Esteve et al., 2014) demonstrate that mantle partial melting and melt percolation through the SCLM took place at those times. Therefore, if melt-rock reaction processes had modified the composition of the SCLM in this area, they would have likely happened at those times. Later on, the Alpine orogenesis was less intense in the Catalan Coastal Ranges than in the Pyrenees, but late Alpine extensional movements did cause the formation of Neogene basins during or post-dating the ECRIS formation. Mantle exhumation during Neogene-Quaternary periods may have taken place through extensional shear zones developed at decreasing T and P (Vauchez et al., 2012). This change in the deformation regime, from dominant transpressive to mainly transtensive, would have reactivated the earlier mantle foliation (Vauchez et al., 2000) and transformed the olivine AG-type fabric into A- and D-types, mainly through SR and GBM dynamic recrystallization most likely accompanied by diffusion processes. These mantle extensional shear zones are probably related to NW-SE Neogene normal faults, such as the Llorà fault (Fig. 1b) with a transtensional component, along which the most recent CVZ volcanoes are aligned (Bolós et al., 2014). The influence of these younger extensive shear zones in the mantle anisotropy would be minor, as suggested by Barruol et al. (1998). This is in agreement with the fact that porphyroclastic and equigranular lherzolites with olivine A- and D-type fabrics are found less frequently, implying that the Neogene deformation affected smaller volumes of the SCLM, and hence has a minor, more local contribution, to the measured SKS-wave splitting. In addition, as suggested by Barruol et al. (1998), the new foliation would be less dipping, but with the lineation orientated (W-E, NW-SE) as in the earlier foliation.

**11. Conclusions**

The SCLM of NE Spain is mainly formed of spinel lherzolites, harzburgites and pyroxenites (websterites), in decreasing order, as inferred from the mantle xenoliths found. All these rocks are protogranular or coarse-granular, but among lherzolites there are also subordinate porphyroclastic and equigranular microstructures, with gradual transition between them.

Both Mg# and Cr# increase from lherzolite to harzburgite silicates, whereas the reverse is true in spinel. Mineral compositions of websterite silicates are similar to those of harzburgites. No compositional differences were observed between the three microstructural types of lherzolites.

T and P estimates both decrease from protogranular peridotites and the websterite towards pophyroclastic and equigranular lherzolites recrystallized at higher stress.

Harzburgites, the websterite and most lherzolites have olivine AG-type or [010]-fiber deformation fabric. The remaining lherzolites show A-type or orthorhombic fabric, except in one case which shows D-type or [100]-fiber fabric. Transitions between all types are observed. No relationship exists between microstructure types and olivine deformation fabrics, although the highest BA indices correspond to porphyroclastic lherzolites. There are overall positive correlations of olivine mode and equilibrium T with respect to the fabric strength: the J index decreases from protogranular harzburgites, clinopyroxene-poor lherzolites and websterite towards porphyroclastic and equigranular lherzolites equilibrated at lower T and P. In most xenoliths, orthopyroxene and clinopyroxene CPO are coherent with olivine AG- type deformation fabric~~s~~. Incoherence between olivine and pyroxene CPO is more frequent in porphyroclastic and equigranular lherzolites with AG-, A- and D-type deformation fabrics. Olivine AG-type fabric is related to deformation by simple shear or transpression, probably associated with late-Hercynian strike-slip shear zones, and to subsequent annealing during late Hercynian decompression, Permian and Cretaceous rifting episodes. All these decompressional events favoured the percolation of mafic magmas through the SCLM causing metasomatism and possible refertilization. Olivine AG-type fabric was transformed later on into A- and D-types through SR and GBM dynamic recrystallization, taking place at higher stress and decreasing T and P. This would have been due to changes in the deformation regime, from simple shear or traspression to mainly transtension, during late Alpine extensional movements that led to the formation of Neogene basins.

Calculated seismic properties indicate that Vp distribution displays a well-defined maximum aligned with the maximum of olivine [100] axis. In most xenoliths, maximum values of AVs are also within the "foliation" plane and the direction of S1 polarization planes tends to be aligned with the olivine [100] axis. Refractory harzburgites, less fertile lherzolites and the websterite show higher anisotropy than the recrystallized porphyroclastic and equigranular lherzolites.

Comparison with available SKS-wave splitting data and assuming that estimates for Vp max, the highest AVs values and the direction of S1 polarization of the average mantle composition match the local FPD (N120 °E), it is inferred that most of the measured seismic anisotropy would be explained by the lithospheric contribution, if the SCLM fabrics record mainly transpression and transtensional deformation.

**Acknowledgments**

Financial support for this study was provided by the MINECO research project CGL2011-26700 and the MINECO grants BES-2012-051860 and EEBB-I-14-07880 to MFR. We are much indebted to two anonymous reviewers who helped to improve previous versions of the manuscript. C. Reed is thanked for the English edition of the manuscript.

**References**

Abramson, E.H., Brown, J.M., Slutsky, L.J., Zang, J., 1997. The elastic constants of San Sarlos olivine to 17 Gpa. J. Geophys. Res. 102, 12253–12263.

Amelinckx, S., Dekeyser, W., 1959. The structure and properties of grain boundaries. Solid State Phys. 8, 325–499.

Ancochea, E., 2004. El vulcanismo neógeno peninsular. In: Vera J.A., (Ed) Geología De España. Sociedad Geológica de España- Instituto Geológico y Minero de España, Madrid, 671–672.

Ayala, C., Torne, M., Pous, J., 2003. The lithosphere–asthenosphere boundary in the western Mediterranean from 3D joint gravity and geoid modelling: tectonic implications. Earth Planet. Sci. Lett. 209, 275–290.

Baptiste, V., Tommasi, A., 2014. Petrophysical constraints on the seismic properties of the Kaapvaal mantle root. Solid Earth 5, 1–19.

Baptiste, V., Tommasi, A., Demouchy, S., 2012. Deformation and hydration of the lithospheric mantle beneath the Kaapvaal Craton, South Africa. Lithos 149, 31–50.

Barruol, G., Mainprice, D., 1993. A quantitative evaluation of the contribution of crustal rocks to the shear wave splitting of teleseismic sks waves. Phys. Earth Planet. Inter. 78, 281–300.

Barruol, G., Souriau, A., Vauchez, A., Diaz, J., Gallart, J., Tubia, J., Cuevas, J., 1998. Lithospheric anisotropy beneath the Pyrenees from shear wave splitting. J. Geophys. Res. 103, 30039–30054.

Barruol, G., Granet, M., 2002. A tertiary asthenospheric flow beneath the southern French Massif Central related to the west Mediterranean extension evidenced by upper mantle seismic anisotropy. Earth Planet. Sci. Lett. 202, 31–47.

Barruol, G., Deschamps, A., Coutant, O., 2004. Mapping upper mantle anisotropy beneath SE France by sks splitting indicates Neogene asthenospheric flow induced by apenninic slab roll-back and deflected by the deep alpine roots. Tectonophysics 394, 125–138.

Bascou, J., Tommasi, A., Mainprice, D., 2002. Plastic deformation and development of clinopyroxene lattice preferred orientations in eclogites. J. Struct. Geol. 24, 1357–1368.

Bascou, J., Delpech, G., Vauchez, A., Moine, B.N., Cottin, J.Y., Barruol, G., 2008. An integrated study of microstructural, geochemical, and seismic properties of the lithospheric mantle above the Kerguelen Plume (Indian Ocean). Geochem. Geophys. Geosyst. 9, Q04036. http:// dx.doi.org:10.1029/2007GC001879.

Bascou, J., Doucet, L.S., Saumet, S., Ionov, D.A., Ashchepkov, I.V., Golovin, A.V., 2011. Seismic velocities, anisotropy and deformation in Siberian Cratonic Mantle: EBSD data on xenoliths from the Udachnaya Kimberlite. Earth Planet. Sci. Lett. 304, 71–84.

Ben Ismaïl, W., Mainprice, D., 1998. An olivine fabric database: an overview of upper mantle fabrics and seismic anisotropy. Tectonophysics 296, 145–157.

Bianchini, G., Beccaluva, L., Bonadiman, C., Nowell, G., Pearson, G., Siena, F., Wilson, M., 2007. Evidence of diverse depletion and metasomatic events in harzburgite– lherzolite mantle xenoliths from the Iberian Plate (Olot, NE Spain): implications for lithosphere accretionary processes. Lithos 94, 25–45.

Bodinier, J.L., Garrido, C.J., Chanefo, I., Bruguier, O., Gervilla, F., 2008. Origin of pyroxenite– peridotite veined mantle by refertilization reactions: evidence from the Ronda Peridotite (Southern Spain). J. Petrol. 49, 999–1025.

Bolós, X., Barde-Cabusson, S., Pedrazzi, D., Martí, J., Casas, A., Lovera, R., Nadal-Sala, D., 2014. Volcano structural-analysis of La Garrotxa volcanic Field (NE Iberia): Implications for the plumbing system. Tectonophysics 642, 58–70.

Brey, G.P., Köhler, T., 1990. Geothermobarometry in four-phase lherzolites II. New thermobarometers, and practical assessment of existing thermobarometers. J. Petrol. 31, 1353–1378.

Bunge, H.J., 1982. Texture analysis in materials sciences. Buttleworth, London, 593 p.

Butjosa, L., Enrique, P., Galán. G., 2013. Las hornblenditas, gabros y dioritas del macizo del Montnegre (Barcelona, Cordilleras Costero Catalanas). Geogaceta 54, 35–38.

Carreras, J., 2001. Zooming on northern Cap de Creus shear zones. J. Struct. Geol.  23, 1457–1486.

Carreras, J., Capella, I., 1994. Tectonic levels in the palaeozoic basement of the Pyrenees: a review and a new interpretation.  J. Struct. Geol. 16, 1509–1524.

Carter N.L., Avé Lallemant, H.G., 1970. High temperature flow of dunite and peridotite. Geol. Soc. Am. Bull. 81, 2181–2202.

Chai, M., Brown, J.M., Slutsky, L.J., 1997. The elastic constants of an aluminous orthopyroxene to 12.5 Gpa. J. Geophys. Res. 102, 14779–14786.

Collins, M.D., Brown, J.M., 1998. Elasticity of an upper mantle clinopyroxene. Phys. Chem. Miner. 26, 7–13.

Couvy, H., Frost, D.J., Heidelbach, F., Nyilas, K., Ungár, T., Mackwell, S., Cordier, P., 2004. Shear deformation experiments of forsterite at 11 Gpa–1400 °C. in the multianvil apparatus. Eur. J. Mineral. 16, 877–889.

Dañobeitia, J.J., Arguedas, M., Gallart, J., Banda, E., Makris, J., 1992. Deep crustal configuration of the Valencia Trough and its balearic borders from extensive refraction and wide-angle reflection seismic profiling. Tectonophysics 203, 37–55.

De Kloe, R., 2001. Deformation mechanisms and melt nano-structures in experimentally deformed olivine–orthopyroxene rocks with low melt fractions. An electron microscopy study. Ph.D. Thesis. Utrecht University, 173 p.

Díaz, J., Gallart, J., Morais, I., Silveira, G., Pedreira, D., Pulgar, J. A., Dias, N.A., Ruiz, M., González-Cortina, J. M., 2015. From the Bay of Biscay to the High Atlas: completing the anisotropic characterization of the upper mantle beneath the westernmost Mediterranean region.Tectonophysics 663, 192–202.

Dijkstra, A.H., Barth, M.G., Drury, M.R., Mason, P.R.D., Vissers, R.L.M., 2003. Diffuse porous melt flow and melt–rock reaction in the lithospheric mantle at a slow-spreading ridge: a structural petrology and LA-ICP-MS study of the Othris Peridotite Massif (Greece). Geochem. Geophys. Geosyst. 4, 8613. http://dx.doi.org/10.1029/2001GC000278.

Donville, B., 1973b. Ages potassium-argon des vulcanites du Bas Ampurdan (Nord-Est de L’Espagne). CR Acad. Sci. Paris 276, 3253–3256 .

Donville, B., 1973c. Ages potassium-argon des roches volcaniques de la depresion de La Selva (Nord-Est de L’Espagne). CR Acad. Sci. Paris 277, 1–4.

Downes, H., 1987. Relationship between geochemistry and textural type in spinel lherzolites, Massif central and Languedoc, France. In: Nixon, P.H. (Ed.),Mantle Xenoliths, pp. 125–133.

Droop, G.T.R., 1987. A general equation for estimating Fe3+ in ferromagnesian silicates and oxides from microprobe analysis using stoichiometry criteria. Mineral. Mag. 51, 2417–2423.

Enrique, P., 2009. Las espesartitas, camptonitas y bostonitas del complejo intrusivo de Aiguablava (Cadenas Costeras Catalanas): cartografía y composición. Geogaceta 47, 125–128.

Esteve, S., Enrique, P., Galán, G., 2014. The camptonites in the multiple intrusion of Platja Fonda (Girona, NE Spain): mechanisms of intrusion and geochemistry.J. of Geosciences 59, 23–40.

Falus, G., Tommasi, A., Ingrin, J., Csaba, S., 2008. Deformation and seismic anisotropy of the lithospheric mantle in the Southeastern Carpathians: a mantle xenolith study. Earth Planet. Sci. Lett. 272, 50–64.

Falus, G., Tommasi, A., Soustelle, V., 2011. Effect of dynamic recrystallization on olivine crystal preferred orientations in mantle xenoliths deformed under varied stress conditions. J. Struct. Geol. 33, 1528–1540.

Fernández-Roig, M., Galán, G., 2015. Microestructuras y química mineral de lherzolitas en xenolitos mantélicos de Cataluña (NE España). Geogaceta. 57, 3–6.

Frank, F.C., 1950. Report on the Conference on Plastic Deformation of Crystalline Solids. Carnegie Institut of Technology and Office of Naval Research, Washington, 150 p.

Galán, G., Oliveras, V., Paterson, B.A., 2008. Types of metasomatism in mantle xenoliths in Neogene–Quaternary alkaline mafic lavas from Catalonia (NE Spain). In: Coltorti, M., Grégoire, M. (Eds.), Metasomatism in oceanic and continental lithospheric mantle. Geological Society, London, Special Publication, 293, 121–153.

Galán, G., Oliveras, V., Paterson, B.A., 2011. Thermal and redox state of the subcontinental lithospheric mantle of NE Spain from thermobarometric data on mantle xenoliths. Int. J. Earth Sci. 100, 81–106.

Galán, G., Oliveras, V., 2014. Melting and metasomatism in the lithospheric mantle of NE Spain: Geochemical and Sr-Nd isotopic characteristics. Chem. Geol. 366: 75–89

Gallart, J., Olivera, C., Correig, A.M., 1984. Aproximación geofísica a la zona volcánica de Olot (Girona). Estudio local de sismicidad. Rev. Geofis. 40, 205–226.

Gallart, J., Pous, J., Boix, F., Hirn, A., 1991. Physical constraints on the crustal structure of the Olot volcanic area, Northeastern Iberian Peninsula. J. Volcanol. Geotherm. Res. 47, 33–44.

Godard, G., Van Roermund, H.L.M., 1995. Deformation-induced clinopyroxene fabrics from eclogites. J. Struct. Geol., 17, 1425-1443.

Goula, X., Olivera, C., Fleta, J., Grellet, B., Lindo, R., Rivera, L.A., Cisternas, A., Carbon, D., 1999. Present and recent stress regime in the Eastern part of the Pyrenees. Tectonophysics 308, 487–502.

Harte, B., 1977. Rock nomenclature with particular relation to deformation and recrystallisation textures in olivine-bearing xenoliths. J. Petrol. 85, 279–288

Helmstaedt, H., Anderson, O.L., Gavasci ,A.T., 1972. Petrofabric studies of eclogite, spinel-websterite, and spinel-lherzolite xenoliths from kimberlite-bearing breccia pipes in southeastern Utah and northeastern Arizona. J. Geophys. Res. 77, 4350-4365.

Hidas, K., Falus, G., Szabó, C., Szabó, P.J., Kovács, I., Földes, T., 2007. Geodynamic implications of flattened tabular equigranular textured peridotites from the Bakony–Balaton Highland volcanic field (Western Hungary). J. Geodyn. 43, 484–503

Higgie, K., Tommasi, A., 2014. Deformation in a partially molten mantle: constraints from plagioclase-lherzolites from Lanzo, Western Alps. Tectonophysics 615–616, 167–181.

Holtzman, B.K., Kohlstedt, D.L., Zimmerman, M.E., Heidelbach, F., Hiraga, T., Hustoft, J., 2003. Melt segregation and strain partitioning; implications for seismic anisotropy and mantle flow. Science 301, 1227–1230.

Humbert, M., Cey, N., Mullier, J., Esling, C., 1996. Determination of mean orientation from a cloud of orientations. Application to electron backscatterring pattern measurements. J. Appl. Crystallog. 29, 662–666.

Johanesen, K.E., Platt, J.P., 2015). Rheology, microstructure, and fabric in a large scale mantle shear zone, Ronda Peridotite, Southern Spain. J. Struct. Geol., 73, 1–17.

Jung, H., Karato, S.I., 2001. Water-induced fabric transitions in olivine. Science 293, 1460–1464.

Jung, H., Katayama, I., Jiang, Z., Hiraga, T., Karato, S., 2006. Effect of water and stress on the lattice-preferred orientation of olivine. Tectonophysics 421, 1–22.

Jung, H., Mo, W., Green, H.W., 2009. Upper mantle seismic anisotropy resulting from pressure-induced slip transition in olivine. Nat. Geosci. 2, 73-77.

Jung, S., Jung, H., Austrheim, H., 2014. Characterization of olivine fabrics and mylonite in the presence of fluid and implications for seismic anisotropy and shear localization, Earth, Planets Space, 66:46.

Kaczmarek, M.-A., Müntener, O., 2008. Juxtaposition of melt impregnation and high- temperature shear zones in the upper mantle; field and petrological constraints from the Lanzo Peridotite (Northern Italy). J. Petrol. 49, 2187–2220.

Kaczmarek, M.-A., Tommasi, A., 2011. Anatomy of an extensional shear zone in the mantle (Lanzo Massif, Italy). Geochem. Geophys. Geosyst. 12, Q0AG06. http://dx.doi.org/ 10.1029/2011gc003627.

Kaczmarek, M.A., Reddy, S.M., 2013. Mantle deformation during rifting: Constraints from quantitative microstructural analysis of olivine from the East African Rift (Marsabit, Kenya). Tectonophysics, 608, 1122-1137.

Karato, S.-I., Toriumu, M., Fujii, T., 1980. Dynamic recrystallization of olivine single crystals during high-temperature creep. Geophys. Res. Lett. 7, 649–652.

Karato, S.I., Jung, H., Katayama, I., Skemer, P., 2008. Geodynamic significance of seismic anisotropy of the upper mantle: new insights from laboratory studies. Annu. Rev. Earth Planet. Sci. 36, 59–95.

Katayama, I., Jung, H., Karato, S.I., 2004. New type of olivine fabric from deformation experiments at modest water content and low stress. Geology 32, 1045–1048.

Katayama, I., Karato, S., 2006. Effects of temperature on the B- to C-type fabric transition in olivine. Phys. Earth Planet. Inter*.* 157, 33–45

Kelemen, P.B., Dick, H.J.B. And Quick, J.E., 1992. Formation of harzburgite by pervasive melt/rock reaction in the upper mantle. Nature 358, 635–41.

* Kelemen, P.B., Hart, S.R., Bernstein, S., 1998. Silica enrichment in the continental upper mantle lithosphere via melt/rock reaction. Earth Planet. Sci. Lett.164, 387–406.

Klébesz, R., Gráczer, Z., Szanyi, G., Liptai, N., Kovács, I., Patkó, L., Pintér, Z., Falus, G., Wesztergom, V., Szabó, C., 2015. Constraints on the thickness and seismic properties of the lithosphere in an extensional setting (Nógrád-Gömör Volcanic Field, Northern Pannonian Basin). Acta Geod. Geophys. 50, 133–149.

Köhler, T.P., Brey, G.P., 1990. Calcium exchange between olivine and clinopyroxene calibrated as a geothermobarometer for natural peridotites from 2 to 60 Kb with applications. Geoch. Cosmoch. Acta 54, 2375–2388

Kovács, I., Falus, G., Stuart, G., Hidas, K., Szabó, C., Flower, M. F. J., Hegedus, E., Posgay, K., Zilahi-Sebess, L., 2012. Seismic anisotropy and deformation patterns in upper mantle xenoliths from the central Carpathian-Pannonian Region: asthenospheric flow as a driving force for Cenozoic extension and extrusion? Tectonophysics 514–517, 168–179

Kretz, R., 1983. Symbols for rock forming minerals. Am. Mineral. 68, 277–279

Le Maitre, R.W., 2002. Igneous rocks; a classification and glossary of terms. Cambridge. University Press, Cambridge. 236 p.

Le Roux, V., Bodinier, J.L., Tommasi, A., Alard, O., Dautria, J.M., Vauchez, A., Riches, A.J.V., 2007. The Lherz spinel lherzolite: refertilized rather than pristine mantle. Earth Planet Sci. Lett. 259, 599–612.

Le Roux, V., Tommasi, A., Vauchez, A., 2008. Feedback between deformation and melt percolation in an exhumed lithosphere–asthenosphere boundary. Earth Planet. Sci. Lett. 274, 401–413.

Lee, K., Jiang, Z., Karato, S., 2002. A scanning electron microscope study of the effects of dynamic recrystallization on lattice preferred orientation in olivine. Tectonophysics 351, 331-341.

Lenoir, X., Garrido, C., Bodinier, J.-L., Dautria, J.-M., Gervilla, F., 2001. The recrystallization front of the Ronda peridotite: evidence for melting and thermal erosion of lithospheric mantle beneath the Alboran Basin. J. Petrol. 42, 141–158.

Lewis, C. J., Vergés, J., Marzo, M., 2000. High mountains in a zone of extended crust: insights into the Neogene-Quaternary topographic development of northeastern Iberia. Tectonics 19, 86–102.

Lloyd, G. E., Schmidt, N.-H., Mainpricc, D., Prior, D.J., 1991. Crystallographic textures. Mineral. Mag. 55, 331–345.

Losantos, M., Montaner, J., Solà, J., Mató, E., Sampsó, J.M., Picart, J., Calvet, F., Enrique, P., Ferrés, M. Y Solé, J., 2000. Mapa geològic de Catalunya 1:25.000, Palafrugell 335- 1-1 (79-25). ICC (Servei Geològic de Catalunya).

Lustrino, M., Wilson, M., 2007. The circum-mediterranean anorogenic cenozoic igneous province. Earth Sci. Rev. 81, 1–65.

Mainprice, D., 2007. Seismic anisotropy of the deep earth from a mineral and rock physics perspective. In: Schubert, G. (Ed.), Treatise on Geophysics. Oxford: Elsevier Ltd. Volume 2, 437–492.

Mainprice, D. Silver, P.G., 1993. Interpretation of sks-waves using samples from the subcontinental lithosphere. Phys. Earth Planet. In. 78, 257–280,

Mainprice, D., Humbert, M., 1994. Methods of calculating petrophysical properties from lattice preferred orientation data. Surv. Geophys. 15, 575–592.

Mainprice, D., Barruol, G., Ben Ismaïl, W., 2000. The seismic anisotropy of the Earth’s Mantle: from single crystal to polycrystal, In: AGU Geophysical Monograph, Edited By: Karato, A.M.F.S., Libermann, R. C., Masters, G., Stixrude, L., AGU, Washington, DC, 117, 237–264.

Mainprice, D., Tommasi, A., Couvy, H., Cordier, P., Frost, D.J., 2005. Pressure sensitivity of olivine slip systems and seismic anisotropy of Earth's upper mantle. Nature 433, 731–733.

Mainprice, D., Bachmann, F., Hielscher, R., Schaeben, H., 2015. Descriptive tools for the analysis of texture projects with large datasets using MTEX — strength, symmetry and components. In: Faulkner, D. R., Mariani, E., Mecklenburgh, J. (Eds.), Rock deformation from field, experiments and theory. A volume in honour of Rutter. Geological Society, London, Special publication 409: 251–271.

Martí, J., 2004. La región volcánica de Gerona. In: Vera, J.A. (Ed.), Geología de España. Sociedad Geológica de España. Instituto Geológico y Minero de España. Ed. pral, Madrid, 672–675.

Martí, J., Pujadas, A., Ferrés, D., Planagumà, L., Mallarach, J.M., 2001. El vulcanisme. Guia de camp de la Zona Volcànica de la Garrotxa, 2nd edn. Parc Natural de la Zona Volcànica de la Garrotxa, Generalitat de Catalunya, Departament de Medi Ambient.

Martínez-Poza, A. I., Druguet, E., Castaño, L. M., Carreras, J., 2014. Dyke intrusion into a pre-existing joint network: the Aiguablava lamprophyre dyke swarm (Catalan Coastal Ranges).Tectonophysics 630, 75–90.

Mercier, J.C.C., Nicolas. A., 1975. Textures and fabrics of upper mantle peridotites as illustrated by xenoliths from basalts. J. Petrol. 16, 454–487.

Morimoto, N., Fabries, J., Ferguson, A.K., Ginzburg, I.V., Ross, M., Seifert, F.A., Zussman, J., Aoki, K., Gottardi, G. 1988. Nomenclature of pyroxenes. Am. Mineral. 73, 1123–1133.

Nicolas, A., Boudier, F., Boullier, A.M., 1973. Mechanisms of flow in naturally and experimentally deformed peridotites. Am. J. Sci. 273, 853–876.

Nicolas, A., Christensen, N.I., 1987. Formation of anisotropy in upper mantle peridotites—A review. In: Fuchs, K., Froidevaux, C. (Eds.), Composition, structure and dynamics of the lithosphere–asthenosphere system, Geodyn. Ser. AGU, Washington, DC, 111–123.

Ohuchi, T., Kawazoe, T., Nishihara, Y., Nishiyama, N., Irifune, T., 2011. High pressure and temperature fabric transitions in olivine and variations in upper mantle seismic anisotropy. Earth Planet. Sci. Lett. 304, 55-63.

Olivera, C., Fleta, J., Susagna, T., Figueras, S., Goula, X., Roca, A., 2003. Seismicity and recent deformations in the northeastern of the Iberian Peninsula. Fís. Tierra 15, 111–114.

Oliveras, V., 2009. Xenòlits mantèl·lics lligats al vulcanisme neògeno-quaternari de Catalunya: caracterització, processos i condicions d'equilibri. Tesis Doctoral. Universitat Autònoma de Barcelona. 210 p.

O’Reilly, S., Griffin, W. L., 2013. Mantle metasomatism. In: Harlov, D. E., Austrheim, H. (eds) Metasomatism and the Chemical Transformation of Rock. Berlin Heidelberg: Springer, 471–533.

Palasse, L.N., Vissers, R.L.M., Paulssen, H., Basu, A.R., Drury, M.R., 2012. Microstructural and seismic properties of the upper mantle underneath a rifted continental terrane (Baja California): an example of sub-crustal mechanical asthenosphere? Earth Planet. Sci. Lett. 345–348, 60–71.

Park, M., Jung, H., Kil, Y., 2014. Petrofabrics of olivine in a rift axis and rift shoulder and their implications for seismic anisotropy beneath the Rio Grande rift. Island Arc, 23, 299-311.

Park, Y. Jung, H., 2015. Deformation microstructures of olivine and pyroxene in mantle xenoliths in Shanwang, eastern China, near the convergent plate margin, and implications for seismic anisotropy. Int. Geol. Rev. 57, 629-649.

Philippot, P., Van Roermund, H.L.M., 1992. Deformation processes in eclogitic rocks: evidence for the rheological delamination of the oceanic crust in deeper levels of subduction zones. J. Struct. Geol. 14, 1059-1077.

Prior, D.J., Mariani, E., Wheeler, J., 2009. EBDS in earth Sciences: Applications, common practice, and challenges. In: Schwartz, A.J., Kumar, M., Adams, B.L., Field, D.P. (Eds.), Electron Backscatter Diffraction in Materials Science. Springer, 345–360.

Rampone, E., Vissers, R.L.M., Poggio, M., Scambelluri, M., Zanetti, A., 2010. Melt migration and intrusion during exhumation of the Alboran lithosphere: the Tallante mantle xenolith record (Betic Cordillera, SE Spain). J. Petrol. 51, 295–325.

Roca, E., Desegaulx, P., 1992. Analysis of the geological evolution and vertical movements in the València Trough area, western Mediterranean. Mar. Petrol. Geol. 9, 167–185.

Savage, M.K., 1999. Seismic anisotropy and mantle deformation: what have we learned from shear wave splitting? Rev. Geophys. 37, 65–106.

Silver, P.G.,  W.W. Chan, 1988. Implications for continental structure and evolution from seismic anisotropy. Nature 335, 34-39.

Solé Sabaris, L., 1962. Observaciones sobre la edad del volcanismo gerundense. Mem. Real Acad. Cienc. Artes Barc. 34, 359–372.

Solé, J., Pi, T., Enrique, P., 2003. New geochronological data on the late Cretaceous alkaline magmatism of the northeast Iberian Peninsula. Cretac. Res. 24, 135–140.

Soustelle, V., Tommasi, A., Bodinier, J.-L., Garrido, C.J., Vauchez, A., 2009. Deformation and reactive melt transport in the mantle lithosphere above a large-scale partial melting domain: the Ronda Peridotite Massif, Southern Spain. J. Petrol. 50, 1235–1266.

Soustelle, V., Tommasi, A., Demouchy, S., Ionov, D.A., 2010. Deformation and fluid‚ a rock interaction in the supra-subduction mantle: microstructures and water contents in peridotite xenoliths from the Avacha Volcano, Kamchatka. J. Petrol. 51 363–394.

Tommasi, A., Tikoff, B., Vauchez, A., 1999. Upper mantle tectonics: three-dimensional deformation, olivine crystallographic fabrics and seismic properties. Earth Planet. Sci. Lett. 168, 173–186.

Tommasi, A., Mainprice, D., Canova, G., Chastel, Y., 2000. Viscoplastic self-consistent and equilibrium-based modeling of olivine lattice preferred orientations: implications for the upper mantle seismic anisotropy. J. Geophys. Res. Solid Earth 105, 7893–7908.

Tommasi, A., Vauchez, A., Godard, M., Belley, F., 2006. Deformation and melt transport in a highly depleted peridotite massif from the Canadian Cordillera: implications to seismic anisotropy above subduction zones. Earth Planet. Sci. Lett. 252, 245–259.

Tommasi, A., Vauchez, A., Ionov, D., 2008. Deformation, static recrystallization, and reactive melt transport in shallow subcontinental mantle xenoliths (Tok Cenozoic Volcanic Field, SE Siberia). Earth Planet. Sci. Lett. 272, 65–77.

Tournon, J. 1968. Le volcanisme de la province de Gérone (Espagne). Études de basaltes quaternaires et de leurs enclaves. Thèse 3e Cycle, Université de Paris. 120 p.

Ubide, T., Lago, M., Arranz, E., Galé, C., Larrea, P., 2010. The lamprophyric sub-vertical dyke swarm from Aiguablava (Catalonian Coastal Ranges): petrology and composition. Geogaceta 49, 75–78.

Ubide, T., Arranz, E., Lago, M., Galé, C., Larrea, P., 2012. The influence of crystal settling on the compositional zoning of a thin lamprophyre sill: a multi-method approach. Lithos 132–133, 37–49.

Ulrich, S., Mainprice, D., 2005. Does cation ordering in omphacite influence development of lattice-preferred orientation? J. Struct. Geol. 27, 419–431.

Van Der Wal, D., Chopra, P., Drury, M., Fitzgerald, J., 1993. Relationships between dynamically recrystallized grain size and deformation conditions in experi- mentally deformed olivine rocks. Geophys. Res. Lett. 20, 1479–1482.

Vauchez, A., Tommasi, A., Barruol, G., Maumus, J., 2000. Upper mantle deformation and seismic anisotropy in continental rifts. Phys. Chem. Earth 25, 111–117.

Vauchez, A., Garrido, C., 2001. Seismic properties of an asthenospherized lithospheric mantle: constraints from lattice preferred orientations in peridotite from the Ronda Massif. Earth Planet. Sci. Lett. 192, 235–249.

Vauchez, A., Dineur, F., Rudnick, R.L., 2005. Microstructure, texture and seismic anisotropy of the lithospheric mantle above a mantle plume. Insights from the Labait Volcano xenoliths (Tanzania). Earth Planet. Sci. Lett. 232, 295–314.

Vauchez, A., Tommasi, A., Mainprice, D., 2012. Faults (shear zones) in the Earth's Mantle. Tectonophysics 558–559, 1–27.

Vollmer, F.W., 1990. An application of eigenvalue methods to structural domain analysis. Geol. Soc. Am. Bull. 102, 786–791.

Vonlanthen, P., Kunze, K., Burlini, L., Grobety, B., 2006. Seismic properties of the upper mantle beneath Lanzarote (Canary Islands): model predictions based on texture measurements by EBSD.Tectonophysics 428, 65–85.

Woodcock, N. H., Naylor, M. A., 1983. Randomness testing in three-dimensional orientation data. J. Struct. Geol. 5, 539–548.

Zaffarana, C., Tommasi, A., Vauchez, A., Grégoire, M., 2014. Microstructures and seismic properties of south Patagonian mantle xenoliths (Gobernador Gregores And Pali Aike). Tectonophysics 621, 175–197.

Zhang, J., Green, H.W., Bozhilov, K., 2006. Rheology of omphacite at high temperature and pressure and significance of its lattice preferred orientations. Earth Planet. Sci. Lett. 246, 432–443.

**Figure captions**

**Fig. 1**.Location of the Catalan volcanic zone (CVZ). (a) The CVZ within the European Cenozoic Rift System. Lighter grey represents orogenic belts. VT stands for the Valencia Trough. Modified after Martí et al. (2001). (b) Simplified geological map of the CVZ with the location of La Banya del Boc (BB), Canet d’Adri (CA) and Sant Corneli volcanoes (SC). Map modified after Solé Sabaris (1962) and Tournon (1968). Geographic ETRS89 coordinates are indicated.

**Fig. 2.** Olivine-orthopyroxene-clinopyroxene diagram for ultramafic rocks (Le Maitre, 2002) with the projection of the studied mantle xenoliths classified by lithological and microstructural features

**Fig. 3**. Microstructures of the mantle xenoliths. (a) Protogranular harzburgite (CA.12.02) with sporadic subgrain boundaries (SGB) in olivine crystals. (b) Protogranular websterite (BB.12.17) with olivine crystals partially altered to iddingsite. (c) Protogranular lherzolite (BB.08.98) with triple junction boundaries between olivine crystals. (d) Lherzolite (BB.08.59) with transitional protogranular-porphyroclastic microstructure. (e) Elongated orthopyroxene porphyroclasts, one of them with clinopyroxene lamellae, in a finer-grained tabular matrix (lherzolite SC.11.16). (f) Equigranular lherzolite (SC.11.54) with grain boundaries at 120 º. Mineral abbreviations after Kretz (1983).

**Fig. 4**. Microstructural features of peridotites. (a) Subgrain boundaries in olivine crystals and olivine inclusion in orthopyroxene of lherzolite BB.08.98. (b) Amoeboid spinel crystals intergrown with orthopyroxene forming a cluster (lherzolite BB.12.13). (c) Orthopyroxene porphyroclast with bent clinopyroxene lamellae (lherzolite SC.11.52). (d) Stretched spinel crystal in porphyroclastic lherzolite SC.11.16. (e) Interstitial anhedral spinel crystals in harzburgite BB.08.20 (f) Orthopyroxene crystal embayed by olivine and with an olivine inclusion (harzburgite CA.12.02).

**Fig. 5.** Variation of mineral chemistry in lherzolites. (a) Cr# vs. Mg# in clinopyroxene. (b) Cr# vs Mg# in spinel. (c) Na in M2 site vs. Al in M1 site of clinopyroxene. Unfocused clinopyroxene analyses are related to focused analyses by dashed lines.

**Fig. 6.** TCa-in-Opx vs. TBK estimates for the different types of rocks. Solid lines joint porphyroclast (p) with neocrystal (n) temperatures of a single sample. Values from unfocused porphyroclasts (u) are joined to other estimates of the same sample by dashed lines. Dotted and dashed lines represent ±50 ºC of TBK.

**Fig. 7**. Stereograms for olivine, orthopyroxene and clinopyroxene CPO rotated to match olivine [100] and [010] axes with structural X and Z directions, respectively. Stereograms were not contoured when the number of indexed grains is lower than 100. Inverse pole figures for all olivine rotation axes accommodating misorientations of less than 15 º are also included (contours at 1.0 multiples of a uniform distribution intervals). Values of J index (Ben Ismaïl and Mainprice, 1998) and deformation fabric types of olivine according to values of BA index (Mainprice et al., 2015) are also included. Xenoliths are ordered according to fabric strength.

**Fig. 7.** continued

**Fig. 8.** (a) J index vs. BA index. Encircled symbols stand for xenoliths with incoherent olivine and clinopyroxene deformation fabrics. See text for further explanation. (b) Relationships between J index and microstructures.

**Fig. 9**. (a) J index vs. olivine mode. (b) J index vs. Mg# of clinopyroxene (c) J index vs. Na (atoms per fu) in clinopyroxene. (d) T estimates with the TCa-in-opx thermometer of Brey and Kölher (1990) vs. J index. Arrows with a solid line join T values corresponding to porphyroclasts (p) and neocrystals (n) of a single sample. Arrows with a dashed line join T estimates corresponding to unfocused (u) and focused analyses of a single sample. Encircled symbols represent xenoliths with incoherent olivine and clinopyroxene deformation fabrics. Solid line field in Figs a-c: protogranular and protogranular-porphyroclastic peridotites; dashed line field: porphyroclastic and equigranular lherzolites.

**Fig. 10**. (a) Clinopyroxene LS index (Ulrich and Mainprice, 2005) vs. Olivine J index. (b) Olivine BA index (Mainprice et al., 2015) vs. Clinopyroxene LS index. Deformation regimes for olivine fabrics after Tommasi et al. (1999) and for clinopyroxene fabrics after Ulrich and Mainprice (2005). Encircled symbols as in Figs. 8.

**Fig. 11.** (a) Large olivine crystal from protogranular lherzolite BB.08.101 with traces of subgrain boundaries subnormal to [100] indicating [100] slip direction. (b) EBSD map of the selected olivine crystal indicating the grain boundaries with thick black lines and subgrain boundaries with finer black lines. (c) Stereograms for the olivine CPO of lherzolite BB.08.101 in the original position (not rotated). (d) Inverse pole figure for the rotation axes of the selected olivine crystal accommodating misorientations of less than 15 º.

**Fig. 12**. Calculated seismic properties for mantle xenoliths, ordered by rock type and decreasing olivine fabric strength (J index) as in Fig. 7. 3-D distributions of P wave velocities (Vp), S-wave polarization anisotropy (AVs %) and the orientation of the fast shear-wave polarization plane (S1) are displayed from left to right for each sample. Calculated seismic properties for the average SCLM beneath the CVZ are also included.

**Fig. 13**. Estimated SKS effective path length within the SCLM (L) or thickness of the mantle anisotropic layer (Mainprice and Silver, 1993; Baptiste and Tommasi, 2014) in two different end-members for the orientation of the foliation and lineation. (a) Vertical foliation and horizontal lineation. (b) Horizontal foliation and lineation. Filled and empty squares stand for maximum and minimum values of AVs. SKS splitting δt range and average for the estimates by Díaz et al. (2015) setting aside crustal contribution (Barruol and Mainprice, 1993).

**Fig. 3.1E**. (a-b) Raw and processed EBSD maps of protogranular lherzolite BB.08.101; yellow crystals: olivine; blue crystals: orthopyroxene; green crystals: clinopyroxene; red crystals: spinel; traces of subgrain boundaries are represented by thin red lines. (c-d) Idem for porphyroclastic lherzolite SC.11.16. (e-f) Idem for equigranular lherzolite BB.08.57. (g-h) Idem for protogranular harzburgite CA.12.02. EBSD maps for xenoliths BB.08.101 and SC.11.16 were acquiered with the CamScan X500 CrystalProbe field emission gun (FEG) SEM, whereas those for xenolihs BB.08.57 and CA.12.02 were acquired with the XL30 tungsten filament SEM.

**Fig. 8.1E.** (a) Processed EBSD map of the equigranular lherzolite BB.08.57, with different colours standing for different minerals as in Fig. 3.1E

 and traces of subgrain boundaries in thin red lines. Two types of subgrain boundaries are displayed in olivine crystals: one type highlighted with red arrows is normal to olivine [100] axis, which points to [100] slip direction; the other type highlighted with white arrows is parallel to [100] axis (dislocations slipping in the [001] direction). (b) Stereograms for the olivine CPO of the whole sample in the original position (not rotated).

**Fig. 10.2E.** (a) Olivine J index vs. AVp%. (b) Olivine J index vs. AVs%. (c) AVp % vs. olivine mode. (d) AVp% vs. orthopyroxene mode. Values for the average SCLM of the CVZ are also included.