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FLUID FOCUSING IN MANTLE SHEAR ZONES  
DURING RIFTING: PETROSTRUCTURAL  
CONSTRAINTS FROM THE ZABARGAD  
PERIDOTITES, RED SEA**

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1 **PROGRESSIVE STRAIN LOCALIZATION AND FLUID FOCUSING IN MANTLE**  
2 **SHEAR ZONES DURING RIFTING: PETROSTRUCTURAL CONSTRAINTS**  
3 **FROM THE ZABARGAD PERIDOTITES, RED SEA**

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9

10 **ABSTRACT**

11 This article documents the evolution of pressure and temperature conditions and the  
12 successive influence of hydrous melts and aqueous fluids on the operation of extensional  
13 shear zones, which produced exhumation of mantle slivers from deep lithospheric or  
14 asthenospheric depths, nearly reaching the surface, in a rift-to-drift setting. These results are  
15 based on a re-analysis of 40 samples from the three peridotite massifs of Zabargad island in  
16 the northern Red Sea. By integrating high-resolution mapping of the microstructure by  
17 electron backscattered diffraction with recent developments in barometry for plagioclase-  
18 bearing peridotites and in thermodynamic modelling of peridotitic compositions, this study:  
19 (1) constrains the temporal and spatial evolution of petrological and tectonic processes in the  
20 shallow mantle during rifting and (2) documents the presence of melts or aqueous fluids  
21 throughout the functioning of the shear zones, unravelling substantial feedback between  
22 petrological and tectonic processes. Thermobarometry and thermodynamic modelling  
23 constrained by the microstructural observations document progressive strain localization  
24 during shearing, which occurred under decreasing pressure and temperature from near solidus  
25 conditions at >1 GPa (north and central peridotite massifs) or ~0.7 GPa (southern massif) to <  
26 600°C and <0.3 GPa in all three massifs. The data substantiate local aqueous fluid saturation  
27 in the shear zones with increasing hydration during their operation, indicating fluid focusing,  
28 with seawater ingress extending to 10 km depth. The presence of melts or fluids enabled  
29 concurrent dislocation and dissolution-precipitation creep, which resulted in weakening in the  
30 shear zones. However, fluid supply was spatially heterogeneous and likely intermittent, with  
31 equilibrium achieved only locally in the ultramylonites. The present study documents  
32 therefore how progressive strain localization and fluid-focusing in extensional shear zones  
33 may contribute to the thinning and exhumation of the mantle during continental rifting and  
34 the rift-to-drift transition.

35

36 **INTRODUCTION**

37 The processes thinning the lithosphere and more specifically its mantle section until  
38 complete breakup of a plate and creation of a new divergent plate boundary are still poorly

39 understood. Traditionally, continental rifting models range between symmetric pure shear and  
40 asymmetric simple shear end-members (McKenzie 1978; Wernicke 1981). Yet, independently  
41 of the symmetry or asymmetry of the rift system as a whole, geological and geophysical data  
42 on continental margins, in particular those exposing mantle rocks at the ocean bottom, favour  
43 deformation accommodated by large-scale extensional shear zones as the most effective  
44 process for thinning the lithosphere (e.g., Boillot et al. 1987; Whitmarsh et al. 2001; Dean et  
45 al. 2015). Numerical models that successfully reproduce the geometry of continental margins  
46 also generally show mantle thinning and asthenospheric upwelling in response to localized  
47 shear deformation in translithospheric extensional shear zones (e.g., Weinberg et al. 2007;  
48 Brune et al. 2023). However, we lack knowledge on the actual geometry of these shear zones  
49 in the mantle and on the processes controlling their nucleation and growth. The role of melts  
50 and fluids on the thinning of the lithosphere and exhumation of the deep levels of the  
51 lithosphere during continental rifting and at the transition from rifting to drifting also remains  
52 a widely debated question.

53 The three peridotite massifs that outcrop in Zabargad (23° 36' 35" N, 36° 11' 45" E), a  
54 small (3 km<sup>2</sup>) island located within the Zabargad fracture zone, 50 km east of the axis of the  
55 Red Sea (Figure 1), offer an exceptional opportunity to document the evolution of the  
56 deformation and petrological processes, such as partial melting, fluid ingression, and fluid-rock  
57 reactions in the mantle during the last stages of continental rifting and the rift-to-drift transition.  
58 These peridotite massifs preserve a complete record of the deformation processes and  
59 metamorphic and metasomatic reactions involved in the exhumation of mantle slices to very  
60 shallow, near seafloor depths (Bonatti et al. 1986; Nicolas et al. 1987; Piccardo et al. 1988).  
61 The central and northern massifs are mainly composed of spinel lherzolites, whereas the  
62 southern massif is dominantly composed of plagioclase lherzolites (Bonatti et al. 1983, 1986;  
63 Nicolas et al. 1987), indicating evolution along different pressure-temperature paths.

64 Structural mapping and microstructural analysis document that all three massifs are  
65 characterized by pervasive foliations and lineations formed under high-temperature conditions,  
66 which are overprinted by lower temperature deformation focused in discrete shear zones tens  
67 of meters to tens of centimetre wide (Bonatti et al. 1986; Nicolas et al. 1987). The mylonitic  
68 foliations in the shear zones are in general slightly oblique to the high-temperature ones (Figure  
69 1). Both the pervasive and mylonitic foliations have in all three massifs a general NNW to  
70 WNW trend, but dip steeply towards the SW in the central and northern massifs and variably  
71 towards the NE in the southern massif (Figure 1). Lineations also have contrasted orientations  
72 in the central and northern massifs, where they plunge shallowly towards the E, and in the  
73 southern massif, where they have variable plunges towards the NW (Figure 1). These studies  
74 also documented the development of amphibole-rich peridotites within the shear zones.  
75 Moreover, the Central and North peridotite massifs (Figure 1) are in direct contact with  
76 amphibolite to granulite facies gneisses interpreted as slivers of the thinned continental lower  
77 crust and, at the southern limit of the Central massif, the gneisses and peridotites were co-  
78 deformed (Boudier et al. 1988; Boullier et al. 1997; Seyler and Bonatti 1988).

79 Petrological and geochemical studies (Agrinier et al. 1993; Bonatti et al. 1986; Dupuis et  
80 al. 1993; Piccardo et al. 1998) corroborated evolution under decreasing pressures and

81 temperatures. These studies also documented: (i) petrological evidence for melt impregnation,  
82 but depleted LREE patterns in plagioclase peridotites of the southern massif outside the shear  
83 zones, (ii) diffuse websteritic layering, flat REE patterns in peridotites, and rare Ti-rich  
84 pargasite in both peridotites and pyroxenites indicating variable degrees of metasomatism by  
85 alkalic basalts in coarse-grained peridotites of the central and northern massifs, and (iii) marked  
86 enrichment in amphiboles with variable major elements and isotopic compositions and strong  
87 LREE enrichment of the peridotites indicating metasomatism by aqueous fluids in the shear  
88 zones in all three massifs, with increasing contribution of sea-water in the later stages of  
89 deformation.

90 In the present study, we make use of the major advances in the analysis of microstructures  
91 brought by the development of automated high-resolution cartography of crystallographic  
92 orientations by Electron BackScattered Diffraction (EBSD), as well as of recent developments  
93 in barometry of plagioclase-bearing peridotites (Fumagalli et al. 2017) and in thermodynamic  
94 modelling of anhydrous and hydrous melting and subsolidus phase relations in peridotitic  
95 compositions (Holland et al. 2018, Jennings et al. 2015, Tomlinson and Holland 2021), to better  
96 constrain the deformation processes and feedbacks between deformation, melts, and fluids at  
97 the different stages of mantle thinning and exhumation recorded in the Zabargad peridotites.

98

## 99 **METHODS**

100 We performed a detailed petrostructural study of 40 samples derived from the three massifs  
101 (Figure 1 and Table 1). The samples are part of the collection of Geosciences Montpellier. They  
102 were collected by Françoise Boudier and Adolphe Nicolas during field trips in 1985 and 1986,  
103 who also performed the first microstructural analyses of the entire collection (Nicolas et al,  
104 1985; 1987). Previous studies have analysed the amphiboles composition, whole-rock Sr, H,  
105 and O isotopes (Agrinier et al. 1993), and whole-rock trace-element compositions (Dupuy et  
106 al. 1991) of part of this collection, including some of the samples studied here.

107 The microstructures of all samples were analysed by optical microscopy and by  
108 cartography of the crystal orientations by electron backscattered diffraction (EBSD) at the thin  
109 section scale. For 12 samples, we also performed simultaneous high-resolution EBSD and  
110 chemical mapping by dispersive electron spectroscopy (EDS) of selected areas. Chemical  
111 compositions of the major rock-forming minerals were measured by electron microprobe  
112 analysis (EPMA) for 18 samples representative of the different microstructures and modal  
113 compositions in the three massifs.

### 114 **EBSD data acquisition and processing**

115 Mapping of crystallographic orientations of olivine, pyroxenes, plagioclase, spinel and  
116 amphibole was performed by indexing backscattered electron diffraction (EBSD) patterns in  
117 the Geosciences Montpellier SEM-EBSD facility. EBSD mapping was performed on entire  
118 thin sections (at least 90% of the section surface was analyzed) with steps ranging from 7 to 44  
119 microns depending on the average grain size (in all cases, most grains are covered by at least

120 3 data points) using a JEOL5600 scanning electron microscope equipped with a Nordlys II  
121 EBSD detector and the AZtecHKL software from Oxford Instruments. Data were acquired with  
122 a working distance of 24 mm, a beam current of 10 nA, and an acceleration voltage of 17 kV.  
123 For 12 samples representative of the different microstructural groups, higher spatial resolution  
124 EBSD and energy-dispersive X-ray spectroscopy (EDS) chemical maps with step sizes of 1 to  
125 12.5 microns were acquired using a field-emission scanning electron microscope – the  
126 Camscan Crystal Probe X500 FE, equipped with the Oxford Symmetry EBSD detector and an  
127 UltimMax 100 EDS detector. The phases declared for EBSD indexation were: olivine,  
128 orthopyroxene, clinopyroxene, spinel, plagioclase, and amphibole (pargasite). In the high-  
129 resolution analyses, antigorite, chlorite, calcite, and talc were added to the mineral phase list.  
130 The EBSD data was cleaned by removing wild spikes (isolated measurement points) and  
131 replacing the unindexed pixels with the average orientation of their neighbours if at least 6  
132 neighbours belonged to the same phase and had a consistent orientation.

133 EBSD data treatment was performed using the MTEX toolbox v.5.10.2 in Matlab  
134 (<http://mtex-toolbox.github.io/>; Bachmann et al., 2010, 2011). Grains were reconstructed using  
135 a Voronoi tessellation constrained by the indexed points and a misorientation threshold of 15°  
136 to define a grain boundary. Subgrain boundaries are defined with a misorientation threshold of  
137 2°, considering a maximum angular uncertainty in the EBSD data of 1°. Modal compositions  
138 were approached as the area fraction occupied by each mineral on the EBSD maps. For better  
139 visualization of the microstructures, modal composition and orientation maps are presented in  
140 the following with all indexed data, but statistical analysis of the microstructure and  
141 crystallographic orientation was performed after removing grains composed of less than three  
142 pixels.

143 Orientation distribution functions (ODF) describing the crystal-preferred orientations of  
144 all minerals were calculated using a “de la Vallée Poussin” kernel with a half-width of 10°.  
145 They are displayed as pole figures, presented as lower hemisphere stereographic projections.  
146 The orientation of the main crystallographic directions: [100], [010] and [001] of each mineral  
147 phase is plotted relative to the principal axes of the strain ellipsoid X, Y, and Z. In some samples  
148 the orientation of the foliation and lineation could not be defined.

149 The concentration (intensity) of the CPO of the different minerals was quantified using the  
150 dimensionless J-index, which is the volume-averaged integral of the squared orientation  
151 densities in the ODF (Bunge, 1982). The J-index varies from 1 for a uniform orientation  
152 distribution, to infinity for a single crystal. In naturally deformed peridotites, the J-index  
153 calculated using the same parameters as in the present study typically ranges between 1  
154 (random CPO) and 18, with a median at 4.8 (Tommasi and Vauchez, 2015). The symmetry of  
155 the olivine CPO is characterized using the BA-index, which varies between 0 and 1 (Mainprice  
156 et al, 2014). The BA-index allows a classification of the olivine CPO symmetry into 3 types:  
157 axial-[010], characterized by a point concentration of [010] and a girdle distribution of [100]  
158 (BA-index<0.35), orthorhombic, characterized by point concentrations of both [100] and [010]  
159 (0.35<BA-index<0.65), and axial-[100], characterized by a point concentration of [100] and a  
160 girdle distribution of [010] (BA-index>0.65).

161 Intragranular misorientations, which result essentially from the accumulation of  
162 geometrically necessary dislocations, are quantified using the Kernel Average Misorientation  
163 (KAM, which is a measure of the mean misorientation around a measurement point), the local  
164 misorientation relative to the mean orientation of the grain (Mis2Mean), and the Grain  
165 Orientation Spread (GOS, which is the average of the local misorientations relative to the mean  
166 orientation of the grain). These data, together with grain sizes and shapes, are the basis for the  
167 quantitative analysis of the microstructures. Grains were classified as porphyroclasts (grains  
168 predating the deformation) or neoblasts (newly formed grains) based on both their size and  
169 internal deformation. The justification for these criteria is that both recrystallization and  
170 reactions create new grains (neoblasts) with initially small sizes and very low dislocation  
171 densities. Neoblasts are therefore defined as having a GOS  $< 1^\circ$  and an equivalent grain  
172 diameter (the diameter of a circle with the same area as the grain)  $< 400 \mu\text{m}$  (Figure 2). This  
173 double threshold assumes that coarse grains with low GOS values are porphyroclasts poorly  
174 oriented to deform by dislocation creep. Amphiboles being a secondary phase in the present  
175 rock assemblages, they were classified as neoblasts independently of their GOS or grain size.

176 To establish a quantitative characterization of the microstructure, we calculated, for each  
177 mineral, average values at the thin section scale of the orientation data (J- and BA-indexes,  
178 KAM, Mis2Mean, and GOS), as well as area-weighted averages of the equivalent grain size  
179 (diameter of a circle with the same area as the grain), aspect ratio (the ratio between the long  
180 and short axis of the ellipse that best fits the grain), and shape factor (the perimeter of the grain  
181 divided by the perimeter of a circle with the same area as the grain, which increases with  
182 increasing sinuosity of the grain boundaries). These values were calculated for the entire grain  
183 population in the EBSD map and separately for the porphyroclasts and neoblasts. Paleostresses  
184 associated with the different microstructures were estimated based on the olivine recrystallized  
185 grain size using the Van de Wal et al. (1993) piezometer based on the arithmetic mean of the  
186 equivalent neoblast size of each sample using the Python script developed by M. Sanchez-  
187 Lopez with correction for the difference in stress regime between experiments and nature  
188 ([https://github.com/marcoalopez/GrainSizeTools/blob/master/DOCS/\\_Paleopizometry.md](https://github.com/marcoalopez/GrainSizeTools/blob/master/DOCS/_Paleopizometry.md)).

## 189 **Mineral chemistry analysis and geothermometry**

190 Major element composition of olivine, orthopyroxene, clinopyroxene, spinel, plagioclase,  
191 and amphibole were analysed in 14 samples representative of the different microstructural  
192 groups and modal composition range in the three massifs using a Cameca SX100 electron  
193 microprobe analyser (EPMA) at the University of Montpellier. Analytical conditions and the  
194 full dataset are presented in the Supporting Information Table S1.

195 Optical observation of microstructures and EBSD maps were used to identify the  
196 equilibrium assemblages associated with the different stages of deformation. EDS maps have  
197 been used to identify spatial variations in chemical composition and correlate them with the  
198 deformation microstructures.

199 Equilibrium temperatures and pressure for different microstructural assemblages have been  
200 calculated using iteratively the Ca in Opx thermometer of Brey and Köhler (1990) revised by  
201 Nimis and Grütter (2003) and the FACE barometer of Fumagalli et al (2017). For samples that

202 do not contain plagioclase or clinopyroxene, the temperature calculation was performed using  
203 an average pressure of 0.5 GPa. Temperatures were also estimated using the Al-Cr in  
204 orthopyroxene thermometer (Witt-Eickschen and Seck, 1991). The uncertainty in the  
205 equilibrium pressure and temperature conditions due to spatial variations in the minerals  
206 composition within a sample was estimated by considering all possible combinations between  
207 core, intermediate position, rim and matrix analyses for the different minerals in each sample  
208 and presenting the results as the median with error bars indicating the dispersion (minimum  
209 and maximum values) in each petrographic class.

## 210 **Thermodynamical modelling**

211 Phase equilibria and mineral compositions were computed by Gibbs energy minimization  
212 using the PerpleX algorithm version 7.0.9 (Connolly 2005, 2009). Forward thermodynamic  
213 modelling was performed in the KNCFMASHTOCr system using oxide and O<sub>2</sub> as  
214 thermodynamic system components (K<sub>2</sub>O-Na<sub>2</sub>O-CaO-FeO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O-TiO<sub>2</sub>-O<sub>2</sub>-  
215 Cr<sub>2</sub>O<sub>3</sub>). Thermodynamic data for end members are defined based on Holland and Powell  
216 (2011). The considered solid solutions models were: olivine (Ol), orthopyroxene (Opx),  
217 clinopyroxene (Cpx), garnet (Gt), spinel (Spl) from (Tomlinson and Holland, 2021),  
218 plagioclase (Pl) from Holland et al. (2021), clinoamphibole from Green et al. (2016), chlorite  
219 (Chl) from White et al. (2014) with the fclin3 endmember excluded, peridotitic dry melt from  
220 Jenning and Holland (2015), and generic hydrous melt from Holland et al. (2018).

221

## 222 **RESULTS**

### 223 **Microstructural groups**

224 Based on field observations, Bonatti et al. (1986) and Nicolas et al. (1987) proposed that  
225 the three massifs recorded deformation with progressive strain localization, which allowed the  
226 preservation of a record of deformation under decreasing pressure and temperature conditions.  
227 To describe this evolution, we classified the microstructures of the peridotites into four groups  
228 (Figure 3). The first: **coarse-porphyroclastic peridotites**, is composed of peridotites with  
229 microstructures typical of deformation at high temperature and low stress conditions. It  
230 corresponds to the pervasive structuration of the massifs. The three other microstructural  
231 groups record increasing reworking of the primary coarse-porphyroclastic microstructure  
232 within the shear zones under lower temperature conditions: **protomylonites**, **mylonites**, and  
233 **ultramylonites**.

234 Deformation facies are usually defined based on the proportion of neoblasts of all minerals  
235 in a rock. However, in the present study the classification of the microstructures of the  
236 peridotites has been defined solely based on the microstructure of olivine (Figure 3), because,  
237 as discussed later, the evolution of the microstructure of the other major rock-forming minerals  
238 was affected by synkinematic reactions with melts or fluids. Peridotites in which more than  
239 50% of the olivine surface area is recrystallized are classified as **ultramylonites** and between



240 50 and 20 % as **mylonites**. When less than 20 % of the olivine is recrystallized, they are  
241 classified as **coarse-porphyroclastic peridotites** if the area-weighted equivalent grain  
242 diameter of the olivine neoblasts is larger than 100 microns, or **protomylonites**, if the area-  
243 weighted equivalent grain diameter of neoblasts is lower than 100 microns, suggesting incipient  
244 recrystallization under higher stress conditions. The pyroxenites were classified using the  
245 recrystallized fraction of all minerals in the thin section.

## 246 **Modal compositions**

247 Very fertile compositions, characterized by > 20 area % of Al-rich phases (clinopyroxene,  
248 plagioclase, spinel, or amphibole), predominate in all three massifs independently of the  
249 deformation microstructure (Figure 4, Supplementary Material Figure S1). Modal  
250 compositions vary almost continuously between peridotites and pyroxenites (Figure 4). Some  
251 of the studied samples are heterogeneous, containing diffuse pyroxenitic or troctolitic layers at  
252 the mm to cm scale. Yet, some samples with compositions intermediate between that of a  
253 peridotite and an igneous rock are homogeneous at the thin section scale, the heterogeneity in  
254 composition occurring at the outcrop (tens of cm to meters) scale.

255 In the North and Center massifs, the pyroxenites are spinel-websterites and the Al-rich  
256 phases in both peridotites and pyroxenites are clinopyroxene and amphibole. Plagioclase is  
257 only occasionally observed rimming spinel. High contents of plagioclase (up to 14 area %,   
258 Table 1) are only observed in the southern massif, where plagioclase is often the dominant Al-  
259 rich phase (Figures 4, 5a,e, 6a). In this massif, the mafic layers have gabbroic to troctolitic  
260 compositions (Figure 6d).

261 Amphibole-rich peridotites have dominantly mylonitic to ultramylonitic microstructures  
262 (Figures 4 and 5g,i, Table 1). Coarse-porphyroclastic peridotites have in general very low  
263 amphibole contents (Figure 4, Table 1). High amphibole contents can be, nevertheless,  
264 observed in the vicinity (within a few mm or even cm) of fractures that crosscut at variable  
265 angles the ductile foliation in samples from all microstructural groups (Figure 6d).

## 266 **Microstructures**

267 We describe below the main characteristics of each microstructural group, but the evolution  
268 of the microstructure is continuous (Figure 3) and many samples have microstructures  
269 intermediate between those described.

270

### 271 *Microstructural groups*

272

273 **Coarse-porphyroclastic peridotites** are dominantly composed of coarse olivine,  
274 orthopyroxene, clinopyroxene, variable amounts of plagioclase (in the southern massif only),  
275 and minor spinel and amphibole (Figure 5a, 6a). Olivine and orthopyroxene porphyroclasts are  
276 up to 5 mm, but most often 1-2 mm wide (Figures 5a,b, 6a), while clinopyroxene and  
277 plagioclase are usually smaller than 2 mm. Coarse-porphyroclastic peridotites from the  
278 southern massif show a clear foliation and lineation marked by a shape preferred orientation

279 (SPO) of elongated olivine and plagioclase grains (Figure 5a, 6a). Coarse-porphyroclastic  
280 peridotites from the center and northern massifs have a much weaker olivine SPO (more  
281 polygonal shapes and lower aspect ratios, Figure 6b). They have, by consequence, a weaker  
282 foliation, but the latter is often underlined by websterite layers with diffuse contacts with the  
283 peridotite and thicknesses ranging from a few mm to tens of cm (Figure 6b). Olivine-olivine  
284 grains boundaries are more sinuous in the southern massif (compare Figures 5a and 6a to 6b).  
285 In coarse-porphyroclastic peridotites from all three massifs, olivine grains have well-developed  
286 sub-structures, characterized by widely-spaced subgrain boundaries with strong  
287 misorientations at high angle to the grain elongation (Figure 7a,b). Olivine grain sizes have a  
288 multimodal distribution (Supplementary Material Figure S3) and the neoblasts, which are on  
289 average coarser than 100  $\mu\text{m}$  (Figure 3), may only be separated from porphyroclasts based on  
290 their lower intragranular misorientation.

291 In coarse-porphyroclastic peridotites of all three massifs, coarse orthopyroxene grains have  
292 variable sizes, very irregular shapes, with embayments filled by olivine, and no shape-preferred  
293 orientation (Figures 5a,b, 6a,b). Coarse orthopyroxene grains often display undulose extinction  
294 (Figures 5b and 7a,b), but their intragranular misorientation is systematically lower than that  
295 of olivine. Smaller orthopyroxene grains have interstitial shapes with cusp-like terminations or  
296 film-like habitus along olivine-olivine grains boundaries (Figures 5a,b and 6a,b). Sometimes  
297 the film-like orthopyroxene grains are in continuity with coarse ones (Figures 6a,b).  
298 Clinopyroxenes have smaller sizes, but similar shapes as the orthopyroxenes (Figures 5a and  
299 6a,b). Both ortho- and clinopyroxene grain boundaries are usually more sinuous in coarse-  
300 porphyroclastic peridotites from the southern massif (compare Figures 6a and 6b). In coarse-  
301 porphyroclastic peridotites from the southern massif, irregularly-shaped clinopyroxene grains  
302 tend to form trails parallel to the foliation (Figure 5a), whereas in coarse-porphyroclastic  
303 peridotites from the center and northern massif, clinopyroxene is more evenly distributed  
304 (Figure 6b).

305 In coarse-porphyroclastic peridotites of the southern massif, plagioclase occurs mainly as  
306 elongated grains, 1 or 2 mm long, with interstitial shapes, but a well-defined SPO marking the  
307 foliation and lineation, or in aggregates of smaller grains with rounded shapes rimming spinel  
308 (Figures 5a, 6a). In coarse-porphyroclastic peridotites of the northern and center massifs,  
309 plagioclase is usually absent, but it may occur locally as tiny grains rimming spinel (Figure  
310 6b). Spinel occurs as coarse light-brown grains in association or not with pyroxenes. In the  
311 southern massif, it is often elongated parallel to the lineation, despite its irregular shapes, and  
312 systematically rimmed by plagioclase aggregates (Figure 5a). In coarse-porphyroclastic  
313 peridotites of the northern and center massifs, spinel has more rounded shapes and is usually  
314 associated with pyroxenes (Figure 6b). Amphibole is always minor. In coarse-porphyroclastic  
315 peridotites of the northern and center massifs, it occurs as mm-sized grains associated with  
316 clinopyroxene in the peridotites and as interstitial grains in the pyroxenite layers (Figure 6b).  
317 In coarse-porphyroclastic peridotites of the southern massif, amphibole is rare, but may occur  
318 rimming plagioclase (Figure 6a).

319 The **protomylonites** (external domains of the localized shear zones) are characterized by  
320 the development of seams of a very fine-grained matrix forming a discontinuous losangular to

321 anastomosed network at  $\pm 15^\circ$  on average to the elongation of the olivine porphyroclasts  
322 (Figures 5c,d and 6c). The fine-grained matrix is composed mainly of neoblasts of olivine with  
323 average grain sizes between 25 and 100  $\mu\text{m}$  (Figure 3), but it also locally contains significant  
324 amounts of fine-grained orthopyroxene or clinopyroxene neoblasts (Figure 6c). Olivine  
325 porphyroclasts are on average smaller and display more sinuous grain boundaries than in  
326 coarse-porphyroclastic peridotites (Figures 5c,d and 6c). They have a well-developed sub-  
327 structure, characterized by strong undulose extinction (Figure 5d) and a high density of  
328 subgrain boundaries with strong misorientations at high angle to the grain elongation (Figure  
329 7c,d), and sutured grain boundaries (Figures 5c,d, 6b). As in coarse-porphyroclastic peridotites,  
330 orthopyroxene neoblasts have interstitial or film-like shapes, but they are significantly smaller  
331 (60-80  $\mu\text{m}$ , Figure 3b) and tend to form asymmetric trails departing from the porphyroclasts  
332 (Figure 5c, 6c). Clinopyroxene is smaller than orthopyroxene, but has a similar habitus (Figures  
333 5c, 6c). Plagioclase is on average smaller, but displays a similar habitus than in the coarse-  
334 porphyroclastic peridotites. Both clinopyroxene and plagioclase aggregates form trails that  
335 contribute to define the foliation and lineation. Spinel occurs both as coarse grains rimmed by  
336 spinel and tiny grains dispersed in the matrix (Figures 5c, 6c). Amphibole is rare, but may  
337 locally replace clinopyroxene and plagioclase in the fine-grained matrix network (Figure 5c).  
338 In the present sampling, this microstructural group is best expressed in the southern massif  
339 (Figures 1 and 2).

340 The **mylonites** (Figures 5e, f, g, h) are characterized by an increase in the fine-grained  
341 matrix fraction (Figure 3) and a decrease in the size of the porphyroclasts. The average size of  
342 the olivine neoblasts also tends to decrease, but it shows no clear correlation with the olivine  
343 neoblast fraction (Figure 2). The fine-grained matrix forms a continuous anastomosed network  
344 that encloses polycrystalline lenses up to 5 mm long, which preserve the microstructural record  
345 of the previous deformation steps (Figures 5e, f, g, h, and 9a). The anastomosed matrix network  
346 together with the SPO of olivine porphyroclasts, as well as olivine and plagioclase aggregates,  
347 define a strong foliation and lineation (Figures 5e, f, g, h, 7e-h). The fine-grained matrix is  
348 dominantly formed by intermixed olivine and orthopyroxene neoblasts (Figures 8a,c) with  
349 average grain sizes between 25 and 100  $\mu\text{m}$  (Figure 3), but it also contains neoblasts of other  
350 phases in variable proportions. Neoblasts of clinopyroxene and plagioclase tend to form  
351 asymmetric trails extending from the porphyroclasts (Figures 5e, f, g, h, and 8a,c). The matrix  
352 contains higher proportions of amphibole than in the protomylonites. Orthopyroxene neoblasts  
353 have interstitial, film-like shapes along olivine-olivine grain boundaries even far away from  
354 any orthopyroxene porphyroclast; these film-like orthopyroxenes are best developed along  
355 olivine grain boundaries normal to the compression direction (Figure 8a). All phases, except  
356 amphibole, have a clear bimodal grain size distribution (Figures 5e, f, g, h, 8a,c). In mylonites  
357 of the southern massif, plagioclase has clear core and mantle structures (Figure 8c). The  
358 plagioclase rims and the neoblasts are enriched in Ca relative to the cores of the porphyroclasts,  
359 which are more sodic (Figure 8d). In mylonites of the northern and central massifs, plagioclase  
360 occurs as small grains rimming coarse spinels and as exsolutions in clinopyroxene (Figures  
361 8a,d). Spinel also occurs as small grains within the matrix with no plagioclase rims. In all  
362 mylonites, clinopyroxene and plagioclase are partially replaced by amphibole; for a similar

363 recrystallized fraction, the amount of amphibole tends to be higher in the mylonites of the  
364 northern and central massifs (compare Figures 5e, g and 8a). Some mylonites of the northern  
365 massif display two generations of amphibole, which differ in grain size and Ti-content: the  
366 coarser amphiboles are enriched in Ti (Figure 8b). One mylonite of the northern massif –  
367 85ZA51 - has a particular microstructure: it is almost entirely composed of strongly  
368 recrystallized olivine (82%), amphibole (11%), and orthopyroxene neoblasts (6%), but  
369 neoblasts of olivine and amphibole are coarser grained than in the other mylonites.

370 The **ultramylonites** are characterized by almost total recrystallization – the fine-grained  
371 matrix occupies on average 60-70 % of thin section surface and  $\geq 50$  % of the olivine fraction  
372 is composed of neoblasts (Figure 3). The matrix is essentially composed of olivine, amphibole,  
373 and orthopyroxene neoblasts, with minor clinopyroxene and spinel, and traces of plagioclase  
374 (Figures 5i,j). Most plagioclase as well as the majority of the clinopyroxene have been replaced  
375 by amphiboles. The very strong foliation is underlined by variations in the olivine and  
376 amphibole contents, with local development of mm to cm wide amphibolite layers. This fine-  
377 grained matrix encloses rather coarse, millimetric olivine porphyroclasts with ellipsoidal  
378 shapes, which have a clear SPO, well-developed subgrain structures, and recrystallization tails  
379 (Figures 5i,j 7i,h). Olivine porphyroclasts are locally rimmed or present fracture-like features  
380 filled by fine-grained orthopyroxene (Figure 5j). Orthopyroxene porphyroclasts are rare.

381 A single ultramylonite sample – 85ZA71-13 – displays a microstructure indicating  
382 deformation at the brittle-ductile transition. This sample contains coarse olivine porphyroclasts  
383 that are crosscut by conjugated or extensional fractures sealed by either recrystallized olivine  
384 or a very fine-grained orthopyroxene matrix (Figure 8d). These porphyroclasts are enclosed in  
385 a fine-grained matrix-rich composed of olivine-, orthopyroxene-, amphibole-, and scapolite-  
386 rich layers or lenses. The mylonitic foliation defined by these almost monomineralic  
387 recrystallized layers is anharmonically folded. The amphibole-rich layers have a bimodal grain  
388 size, enclosing locally crystals hundreds of  $\mu\text{m}$  wide. EDS analyses document two types of  
389 scapolite with variable Cl contents, the Cl-rich scapolites also being S-rich, which form  
390 independent layers, both containing amphibole and plagioclase clasts (Figure 8f). Plagioclase  
391 clasts are locally observed in the S-rich scapolite layers. Biotite occurs locally within the fine-  
392 grained matrix, but chlorite is not observed.

393

#### 394 *Pyroxenites*

395 In the central and northern massifs, **pyroxenites** are spinel-websterites, which occur as  
396 diffuse layers of variable thickness. They have in general coarse-granular microstructures and  
397 are composed of millimetric, equiaxed, but irregularly shaped, intermixed ortho- and  
398 clinopyroxenes, with variable proportions of olivine crystals with very irregular shapes, minor  
399 spinel and amphibole (replacing clinopyroxene or as isolated grains with interstitial shapes),  
400 and occasionally very minor plagioclase and amphibole rimming spinel (Figure 6b). The  
401 peridotite in the vicinity of the websterite layers display common film-like orthopyroxene  
402 grains along olivine-olivine grain boundaries (Figure 6b). Although amphibole is a primary  
403 accessory phase in most spinel-websterites, in some, like 85ZA42b, the amphibole fraction is  
404 enhanced in the vicinity of ghost conjugated fractures that crosscut the coarse-granular

405 microstructure. In the central massif, at the contact with the gneisses (Figure 1), pyroxenite  
406 layers show ultramylonitic microstructures and have higher plagioclase or amphibole contents  
407 (Figure 3 and Table 1). In these ultramylonites, all phases, except spinel, are largely  
408 recrystallized, forming a very fine-grained matrix (grain sizes  $\sim 25 \mu\text{m}$ ) that encloses rare  
409 orthopyroxene and clinopyroxene porphyroclasts. Olivine and plagioclase are always fine-  
410 grained. Amphibole replaces clinopyroxene and to a lesser extent plagioclase and  
411 orthopyroxene.

412 Samples with gabbroic to troctolitic compositions from the southern massif, like 85ZA26c  
413 or 85ZA71-12, display coarse porphyroclastic microstructures and a rough compositional  
414 layering, marked by variations in the olivine content, parallel to the foliation (Figure 6d).  
415 Plagioclase crystals have interstitial habitus. Coarse clinopyroxene crystals ( $>5 \text{ mm}$  long) have  
416 thick exsolutions of orthopyroxene and plagioclase. Coarse orthopyroxene has finer  
417 clinopyroxene exsolutions. In the olivine-bearing layers, large polycrystalline lenses of  
418 plagioclase rim coarse spinel grains, with both the spinel grains and the plagioclase lenses  
419 elongated parallel to the lineation. Static replacement of clinopyroxene and plagioclase by  
420 amphibole in the vicinity of straight millimetric scale ultramylonite bands forming conjugate  
421 pairs or brittle fractures is observed in peridotites and pyroxenites with variable microstructures  
422 in all three massifs (Figure 6d).

423

#### 424 **Microstructure quantification**

425 The evolution of the microstructure, illustrated above by representative examples, may be  
426 quantified for the entire dataset by the analysis of the area-weighted mean of the equivalent  
427 grain diameter, aspect ratio, shape factor, and mean intragranular misorientation of the olivine  
428 and orthopyroxene porphyroclasts. We based this analysis on the porphyroclasts, rather than  
429 the neoblasts, because the latter were in many cases mapped by too few measurements to be  
430 correctly described. Olivine and orthopyroxene were chosen because they are the dominant  
431 mineral phases in the peridotites. Moreover, as it will be discussed later, they represent end-  
432 members in terms of deformation processes: olivine microstructures are typical of dislocation  
433 creep and dynamic recrystallization, whereas orthopyroxene microstructures point to reactions  
434 with a liquid phase (melt or aqueous fluid).

435 With rare exceptions, the olivine porphyroclasts average size decreases from millimetric in  
436 the coarse-porphyroclastic peridotites to  $<250 \mu\text{m}$  in the ultramylonites (Figure 9a). In contrast,  
437 the average size of the orthopyroxene porphyroclasts, which ranges from  $> 2 \text{ mm}$  to  $<100 \mu\text{m}$ ,  
438 shows a large variation from sample to sample, but no systematic decrease from coarse-  
439 porphyroclastic peridotites to protomylonites. Orthopyroxene porphyroclasts sizes decrease,  
440 however, from the protomylonites to the mylonites and ultramylonites. In most samples, the  
441 orthopyroxene porphyroclasts are coarser than the olivine porphyroclasts. There are no  
442 systematic variations between peridotites from different massifs.

443 Comparison of the average aspect ratios of olivine and orthopyroxene porphyroclasts  
444 indicates that olivine is systematically more elongated than orthopyroxene (Figure 9b).  
445 However, neither olivine nor orthopyroxene shows a systematic variation of the porphyroclast

446 average aspect ratio between the different microstructures. Olivine in the coarse-  
447 porphyroclastic peridotites of the central and northern massifs tends to be less elongated than  
448 in the southern massif. Analysis of the EBSD maps of samples representative of the different  
449 microstructures (Figure 5) shows that the elongation of olivine porphyroclasts is associated  
450 with a shape-preferred orientation (SPO), which materializes the lineation and foliation in all  
451 samples, but that in the coarse-porphyroclastic and protomylonitic peridotites, the  
452 orthopyroxenes, despite being anisometric, do not display a SPO consistent with the foliation  
453 and lineation. They have very irregular shapes and, when they show a weak SPO, the latter is  
454 oblique to the foliation.

455 Orthopyroxene porphyroclasts have usually more sinuous grain boundaries than olivine  
456 porphyroclasts, quantified by their higher shape factor values (Figure 9c). There is no  
457 systematic variation of the olivine porphyroclasts average shape factor between the different  
458 microstructures. There is no systematic variation of the sinuosity of orthopyroxene  
459 porphyroclasts grain boundaries between the coarse-porphyroclastic peridotites to the  
460 protomylonites, but a clear decrease from the protomylonites to the ultramylonites.

461 The mean average intragranular misorientation is also systematically stronger in the olivine  
462 porphyroclasts than in the orthopyroxene porphyroclasts (Figure 9d). Although there is no  
463 systematic variation of the intragranular misorientation, quantified by the GOS, between the  
464 coarse-porphyroclastic peridotites to the protomylonites, the intragranular misorientation of  
465 both olivine and orthopyroxene tends to decrease from the protomylonites to the  
466 ultramylonites. This evolution is also visible in the intragranular misorientation maps of  
467 selected samples (Figure 7), which show more saturated colour scales and more closely spaced  
468 subgrain boundaries, in particular for olivine, in the ultramylonites. In addition, within each  
469 microstructural group, the intragranular misorientation in the olivine porphyroclasts is higher  
470 in the southern massif peridotites.

471

## 472 **Crystal preferred orientations**

473 **Olivine** crystal preferred orientations (CPO) have dominantly an orthorhombic symmetry  
474 (BA-indexes between 0.35 and 0.65, Figure 10a). Coarse-porphyroclastic peridotites have,  
475 nevertheless, a stronger tendency towards axial-010 patterns (BA-indexes <5, cf. red symbols  
476 in Figure 10a). On average, the intensity of the olivine CPO (J index) decreases from the coarse-  
477 porphyroclastic peridotites to the ultramylonites. Separate analysis of the CPO of olivine  
478 porphyroclasts and neoblasts illustrates that the porphyroclasts CPO is systematically more  
479 concentrated (higher J indexes, Figure 10b) than the neoblasts CPO, indicating that  
480 recrystallization produces dispersion of the CPO. It also shows that, with few exceptions, the  
481 neoblasts CPO has lower BA-indexes than the porphyroclasts CPO, indicating that  
482 recrystallization tends to disperse more the [100] axes than the [010] ones. Comparison of pole  
483 figures of the CPO of olivine porphyroclasts and neoblasts for a representative sample  
484 illustrates this evolution (cf. inset in Figure 10b). It also highlights the strong correlation  
485 between the porphyroclasts and neoblasts orientations, consistently with inheritance of  
486 orientation from the parent grains during dynamic recrystallization.

487

488 Figure 11 illustrates the CPO of all major rock-forming minerals in the structural reference  
489 frame for representative samples of the different microstructural facies. Despite the decrease  
490 in intensity of the CPO from the coarse-porphyroclastic peridotites to the ultramylonites  
491 (Figure 10), in all microstructural facies, the **olivine** CPO is characterized by a maximum of  
492 [100] aligned with the lineation (materialized by the elongation of olivine and plagioclase  
493 porphyroclasts) and a maximum of [010] orthogonal to the foliation plane, with, in the rare  
494 samples that have a axial-[100] CPO, some dispersion of [010] in the YZ plane. In the coarse-  
495 porphyroclastic peridotites crosscut by cm-scale diffuse spinel-websterite (northern and central  
496 massif) or gabbroic/troctolitic (southern massif) layers, the [010] maximum is normal to this  
497 layering. [001] is more dispersed and forms either a weak maximum normal to the lineation  
498 within the foliation or rough girdles in the foliation plane or normal to the lineation. Solely rare  
499 ultramylonites show a different olivine CPO, which is characterized by similar distributions of  
500 [100] and [001], with both axes displaying a weak maximum parallel to the lineation (Figure  
501 11).

502 **Orthopyroxene CPO** is highly variable (Figure 11). In many samples, it is characterized  
503 by a maximum of [001] aligned with the lineation and the [100]<sub>ol</sub> maximum and a maximum  
504 of [100] perpendicular to the foliation. However, very weak orthopyroxene CPO, with  
505 secondary concentrations of [001] oblique to the lineation is also common. In addition, some  
506 mylonites and ultramylonites show weak, but clear orthopyroxene CPO characterized by a  
507 maximum of [001] normal to the foliation.

508 **Clinopyroxene CPO** is usually weaker than the olivine and orthopyroxene CPO (Figure  
509 11). Most coarse-porphyroclastic peridotites and protomylonites display, nevertheless, some  
510 alignment of [001] parallel to the lineation and of [010] or the pole to (100) normal to the  
511 lineation. In the mylonites and ultramylonites, the clinopyroxene CPO is poorly defined, with  
512 multiple maxima due to the presence of a few porphyroclasts. In some samples, one of the  
513 [001] maxima is aligned with the lineation and the [100]<sub>ol</sub> maximum, but in most cases, there  
514 is no clear relation between the clinopyroxene CPO and the structural reference frame. In  
515 coarse-porphyroclastic peridotites crosscut by cm-scale diffuse spinel-websterite layers from  
516 the north and central massif, clinopyroxene in the websterite layers is characterized by either a  
517 girdle distribution of [001] in the layering plane, whereas the gabbroic or troctolitic layers in  
518 the southern massif are characterized by a concentration of [001]<sub>cpx</sub> at high angle to the  
519 layering.

520 In coarse-porphyroclastic peridotites and protomylonites, the **plagioclase** CPO is  
521 characterized by either a [001] maximum slightly oblique to the lineation and a (100) maximum  
522 parallel to the foliation or the opposite (Figure 11). In the mylonites, plagioclase CPO are  
523 weaker and characterized by a poor alignment of the normal of (010) with the lineation and of  
524 [100] normal to the foliation. The plagioclase fraction in the mylonites of the northern and  
525 central massifs and in the ultramylonites, in general, is too low for a quantitative analysis of  
526 the CPO.

527 In coarse-porphyroclastic peridotites and protomylonites, the **amphibole** content is often  
528 low and its CPO is not clearly defined, displaying multiple maxima. It often mimics the cpx

529 CPO (Figure 11). In mylonites containing significant amounts of amphibole, the amphibole  
530 CPO, which is most often weak, is characterized by a [001] maximum at a low angle or parallel  
531 the lineation and the [100] olivine maximum and concentration of (100) at a low angle to the  
532 foliation. However, in some mylonites, amphibole has a CPO characterized by a girdle  
533 distribution of [001] in the XZ structural plane, which mimics, in more dispersed, the  
534 clinopyroxene CPO (Figure 11). In amphibole-rich ultramylonites, amphibole displays a strong  
535 CPO characterized by alignment of [001] with the lineation and the [100]<sub>ol</sub> maximum and of  
536 (100) with the foliation (Figure 11).  
537

## 538 Mineral chemistry

539 **Olivine** has  $X_{Mg}$  (Mg/(Mg+Fe) apfu, atoms per formula unit) varying from 0.88 to 0.91 and  
540 Ni contents ranging from 0.0035 to 0.01 apfu in the peridotites and cm-scale pyroxenite layers  
541 in the composite samples (Figure 12a). The thicker pyroxenite 85ZA63 has significantly lower  
542  $X_{Mg}$  (0.85-0.86) and Ni content (0.0017-0.004 apfu). Compositions, in particular Ni contents,  
543 are highly variable both among samples and within each sample, but no systematic variation  
544 between core, rims, or neoblasts within a sample or correlation with the neither the  
545 microstructural facies of the sample nor its provenance (massif) are observed. The dispersion  
546 in olivine compositions documented in the present study is consistent with previous data by  
547 Bonatti et al. (1986) and Piccardo et al. (1988).

548 **Orthopyroxene** is enstatite with  $X_{Mg}$  ranging between 0.88 and 0.91 in the peridotites and  
549 cm-scale pyroxenite layers in the composite samples and between 0.83 and 0.87 in the  
550 pyroxenite 85ZA63 (Figure 12b). As for olivine, there are no systematic variations in  $X_{Mg}$  of  
551 orthopyroxene as a function of the microstructure or provenance. In contrast, Ca contents  
552 systematically decrease from core to rim in the porphyroclasts and the neoblasts within most  
553 samples (Figure 12b). An exception is pyroxenite 85ZA63, where orthopyroxene has rather  
554 homogeneous low Ca contents, that are slightly higher in some neoblasts. Another exception  
555 is an orthopyroxene neoblast in an amphibole-domain in ultramylonite 85ZA11, which displays  
556 a high Ca content (>0.06 apfu), whereas most orthopyroxenes in this sample have Ca contents  
557 <0.01 apfu. Ca content in orthopyroxene also varies systematically with provenance: the  
558 highest Ca contents (>0.06 apfu) were measured in the cores of orthopyroxene porphyroclasts  
559 of peridotites from the southern massif. Al contents in orthopyroxene are also highly variable  
560 (from 0.005 to 0.3 apfu) and decrease rather systematically from core to rim in the  
561 porphyroclasts and the neoblasts, except for two neoblasts in ultramylonite 85ZA11 that have  
562 rather high Al contents (Figure 12c). However, in contrast to the Ca contents, the highest Al  
563 contents and most marked core-rim-neoblasts variations were measured in peridotites and  
564 pyroxenites of the Central and Northern massifs.

565 **Clinopyroxene** is diopside with  $X_{Mg}$  ranging between 0.88 and 0.95. The Al content, which  
566 varies from 0.05 to 0.3 apfu, is anticorrelated with  $X_{Mg}$ , defining three trends: a first, flatter,  
567 trend for the peridotites of the central and northern massif, a second, steeper, trend for the  
568 peridotites of the southern massif, and a third for the intergranular variability in pyroxenite  
569 85ZA63 (Figure 12d). In the peridotites, these trends result in a systematic increase in  $X_{Mg}$  and



570 a decrease in Al content from core to rim in the porphyroclasts and even more in the neoblasts  
571 within the individual samples. The decrease in  $X_{Mg}$  is more marked in the peridotites of the  
572 southern massif. There is no systematic variation of Cr and Ti contents in clinopyroxene with  
573 the sample provenance or microstructure; most samples have Cr contents in clinopyroxene  
574 between 0.15 and 0.04 apfu and Ti contents below 0.025 apfu (Supporting Information Table  
575 S1). However, in most samples, the Cr content in clinopyroxene decreases from core to rim in  
576 the porphyroclasts and even more in the neoblasts (Supporting Information Table S1). The  
577 present analyses are consistent with previous data by Bonnatti et al. (1986) and Piccardo et al.  
578 (1988), including the contrast in compositions between the southern massif and the central and  
579 northern massifs.

580 **Plagioclases** display an extreme variation in Al content and  $X_{Ca}$  (Ca/(Ca+Na) apfu). They  
581 range from andesine ( $X_{Ca}$  of 0.4-0.5) in peridotite and pyroxenite layers of protomylonite  
582 86Z30 from the central massif, labradorite ( $X_{Ca}$  0.5-0.7) in plagioclases rimming spinel in  
583 mylonites and ultramylonites in the central and northern massifs (85ZA63, 85ZA50 and  
584 85ZA48) and cores of plagioclase porphyroclasts in protomylonites from the southern massif  
585 (85ZA37 and 85ZA36), bytownite ( $X_{Ca}$  between 0.7-0.9) in cores and rims of plagioclase  
586 porphyroclasts coarse porphyroclastic peridotites 85ZA1, protomylonites 85ZA37 and  
587 85ZA36, and mylonite 85ZA7b from the southern massif as well as neoblasts in mylonites of  
588 the central and northern massifs, to anorthite ( $X_{Ca}>0.9$ ) in rims and matrix grains of mylonite  
589 85ZA7b and ultramylonite 85ZA71-13 and in coarse grains in the pyroxenitic layer in  
590 protomylonite 85ZA31c from the southern massif (Figure 12e). Both coarse porphyroclastic  
591 peridotite 85ZA1 and mylonite 85ZA7b show enrichment in  $X_{Ca}$  from core to rim (Figure 12e,  
592 Figure 6d=EDS 7b). The present analyses are consistent with previous data by Bonnatti et al.  
593 (1986) and Piccardo et al. (1988), but some plagioclases analyzed by Piccardo et al. (1988)  
594 display significantly lower Ca contents than our data (Figure 12e).

595 **Spinel** displays extreme variations in both  $X_{Mg}$  and  $X_{Cr}$  (Cr/(Cr+Al+Fe<sup>3+</sup>) apfu) (Figure 12f).  
596 Coarse porphyroclastic peridotites and protomylonites from the northern and central massifs  
597 display the highest  $X_{Mg}$  (~0.75) and lowest  $X_{Cr}$  (~0.1; Figure 12f). The lowest  $X_{Mg}$  (~0.25) and  
598 highest  $X_{Cr}$  (~1) is displayed by ultramylonite 85ZA51 from the northern massif. Peridotites  
599 from the southern massif show intermediate values, but also present a trend of decrease in  $X_{Mg}$   
600 and increase in  $X_{Cr}$  with increasing localized deformation. In addition, all samples show  
601 enrichment from cores to rims to matrix in Cr and Fe. The most extreme variation ( $X_{Cr}$  0.1-0.6,  
602  $X_{Mg}$  0.7-0.2) is displayed by mylonite 85ZA50 from the northern massif. Relative to the  
603 peridotites, spinels in the pyroxenite 85ZA63 are enriched in Al for a given  $X_{Mg}$ . The present  
604 analyses are consistent with those from Bonnatti et al. (1986), but we sampled a higher  
605 variability in spinel composition. Some spinels analyzed by Piccardo et al. (1988) have higher  
606  $X_{Mg}$ .

607 As observed in previous studies (Bonnatti et al., 1986; Piccardo et al., 1988; Agrinier et al.  
608 1993), **amphiboles** have a wide range of compositions, from Ti-rich pargasite to tremolite  
609 (Figure 13a = K+Na vs. Al), based on structural formula calculated following Li et al. (2022).  
610 Pargasite with < 0.2 apfu Ti occurs in coarse-porphyroclastic peridotites from the three massifs  
611 (Figure 13a, b). Synkinematic amphiboles in mylonites of the three massifs are also dominantly

612 pargasite, but with lower Ti contents, or magnesio-hornblende, and those in ultramylonites  
613 range from Ti-poor pargasite to magnesio-hornblende with rare tremolite, marking a trend of  
614 decreasing Ti, Al<sup>IV</sup> and Na+K contents with increasing low-temperature deformation (Figure  
615 13a,b). The lowest Al<sup>IV</sup> and Na+K contents are observed in amphiboles that statically replace  
616 clinopyroxenes in the vicinity of a fracture in protomylonites 85ZA41 and 85ZA31c, but on  
617 average amphiboles in 85ZA31c have higher Al<sup>IV</sup> and Na+K contents. It is noteworthy that the  
618 static amphiboles in these samples show a wide range of compositions covering most of the  
619 Al<sup>IV</sup> vs Na+K trend of the Zabargad data (Figure 13a). The static amphiboles in both samples  
620 have very low Ti and high Si contents, with on average lower Si contents in 85ZA31c (Figure  
621 13b). Mylonites 85ZA48 and 85ZA45, which have two petrographically distinct generations  
622 of amphibole (cf. Figure 8b), show a strong variation in Ti (Figure 13b) and Na+K contents  
623 (Figure 13a). Synkinematic amphiboles from mylonite 85ZA51 are slightly offset from the  
624 major trend, being enriched in Na+K relative to their Al<sup>IV</sup> content (Figure 13a). These  
625 amphiboles are enriched in Cr (0.35-0.4 apfu) relative to the main Zabargad amphibole  
626 population (Figure 13c). A similar offset is displayed by some analyses of amphiboles of the  
627 northern massif by Agrinier et al. (1993) and to a lesser extent by Bonatti et al. (1986) and  
628 Piccardo et al. (1988). Amphiboles from ultramylonitic pyroxenite 85ZA63, which were also  
629 previously analyzed by Agrinier et al. (1993), have highly variable compositions (Figure  
630 13a,b,c).

631 **Amphiboles** in coarse-porphyroclastic peridotites 86Z30 and 86Z31a from the center massif  
632 and 85ZA37 from the southern massif have very low Cl contents, but coarse-porphyroclastic  
633 peridotite 85ZA1 has Ti-rich pargasite (Figure 5) with Cl contents up to 0.08 apfu (Figure 13d).  
634 In most mylonites and ultramylonites the amphiboles contain Cl (Figure 13d). In the southern  
635 massif, with increasing strain localization, the amphibole Cl content first increases to > 0.13  
636 apfu in mylonite 85ZA7b and then decreases to 0.05-0.1 apfu in ultramylonites 85ZA11 and  
637 85ZA71-13. Mylonites from the northern and center massifs display highly variable Cl  
638 contents, covering the entire range of observations, and no simple relation between the Cl and  
639 Si content. Cl contents in amphiboles that statically replace clinopyroxene in the vicinity of  
640 fractures are low and independent of the Si content in protomylonite 85ZA41, but high and  
641 anticorrelated with the Si content in protomylonite 85ZA31c (Figure 13d).

## 642 **Equilibrium pressures et temperatures**

643 Equilibrium temperatures have been estimated using the Al-Cr and the Ca in Opx  
644 thermometers (Figures 14a,b). Both thermometers predict a decrease in equilibrium  
645 temperature from core to rim to matrix within individual samples. They also predict a consistent  
646 decrease in equilibrium temperature with increasing localized deformation within the entire  
647 dataset, with the highest temperatures predicted for the cores of Opx porphyroclasts in the  
648 coarse-porphyroclastic peridotites and the lowest temperatures for Opx neoblasts in the matrix  
649 of ultramylonites. However, the plot of the Al content in Opx vs. the temperatures estimated  
650 using the Al-Cr in Opx thermometer shows distinct trends for the southern and central and  
651 northern massifs (Figure 14a). For comparable Al contents lower equilibrium temperatures are  
652 predicted for the central massif. This highlights a possible effect of pressure on this system. In

653 contrast, temperatures predicted using the Ca in Opx thermometer vs. Ca contents neatly  
654 materialize the exponential relation used to adjust the experimental data for the thermometer,  
655 indicating a lower pressure sensitivity (Figure 14b).

656 Despite a high variability of compositions within each sample, in most peridotites from the  
657 northern and central massifs and in coarse-porphyroclastic peridotite 85ZA37 from the  
658 southern massif, the Al<sup>IV</sup> in clinopyroxene displays a rough negative correlation to the X<sub>Ca</sub> of  
659 plagioclase (Figure 14c). The lowest X<sub>Ca</sub> of plagioclase and the highest Al<sup>IV</sup> contents in  
660 clinopyroxene were measured in cores from coarse-porphyroclastic peridotites from the central  
661 massif and the lowest values are displayed by matrix grains in mylonite 85ZA50. It is  
662 noteworthy that the cores plagioclase and clinopyroxene of coarse-porphyroclastic peridotite  
663 85ZA37 from the southern massif have similar compositions to rims in mylonite 85ZA51 from  
664 the northern massif (Figure 14c). In contrast, in coarse-porphyroclastic peridotite 85ZA1 and  
665 mylonite 85ZA7b of the southern massif as well as mylonite 85ZA45 of the northern massif,  
666 the Al<sup>IV</sup> in clinopyroxene varies strongly at roughly constant X<sub>Ca</sub> of plagioclase (Figure 14c).  
667 Within each sample, despite the strong dispersion in the measurements, the median Al<sup>IV</sup> in  
668 clinopyroxene decreases from core, to rims, and to matrix grains. The present data correlate  
669 well with Fumagalli et al. (2017) data for fertile compositions, which show an increase Al<sup>IV</sup> in  
670 clinopyroxene (from 0.06 to 0.16) and a decrease in X<sub>Ca</sub> of plagioclase (from 0.83 to 0.5) with  
671 increasing pressure from 0.3 to 0.9 GPa. The present analyses are consistent with previous data  
672 by Bonnatti et al. (1986) and Piccardo et al. (1988), including the contrast in compositions  
673 between the southern massif and the central/northern ones.

674 The pressures calculated using the FACE barometer of Fumagalli et al. (2017) based on the  
675 temperatures estimates with the Ca in Opx thermometer imply a progressive decrease in  
676 equilibration conditions in all three massifs (Figure 14d). In the southern massif, we document  
677 both cooling and decompression associated with the localized deformation. The composition  
678 of the cores in coarse-porphyroclastic, protomylonitic, and mylonitic peridotites, as well as  
679 previous data by Bonnatti et al. (1986), points to initial pressures > 0.7 GPa and temperatures  
680 ~1200 °C, which evolve during the localized deformation to 900-800 °C and 0.5-0.3 GPa based  
681 on rim and matrix neoblasts data. Intermediate pressures and temperatures (<1000 °C at ~0.6  
682 GPa) are recorded by the compositions of partially reequilibrated cores. The matrix data of  
683 mylonite 85ZA7b results in equilibration temperatures ~700°C, but negative pressures.

684 In the central and northern massifs, the thermobarometric data point to almost isothermal  
685 decompression at 750-850°C, starting at >0.8 GPa (core and rim data in coarse porphyroclastic  
686 peridotite 86ZA30) and ending at ~0.2 GPa (matrix data in mylonites 85ZA45 and 85ZA63).  
687 Higher temperatures (>900°C) at ~0.8 GPa are predicted based on previous analyses by  
688 Bonnatti et al. (1986) and Piccardo et al. (1988). This evolution is consistent with the  
689 equilibration conditions inferred for the gneisses co-deformed at the contact with mylonitic  
690 peridotites of the Central massif (Figure 14d), which evolved from 1.05 GPa and 850-800°C  
691 to 700-800°C and 0.3 GPa (Boudier et al. 1988, Seyler and Bonnatti 1988). In these gneisses,  
692 analysis of fluid inclusions implies further exhumation and cooling up to 0.15 GPa and 450°C  
693 (Boullier et al. 1997).

694

695

696 **DISCUSSION**697 **Evolution of deformation conditions and deformation processes**

698 The present petrostructural data document the evolution of deformation processes and  
699 reactions during the exhumation of the mantle along extensional shear zones during the  
700 formation of the Red Sea rift. In the three peridotite massifs, strain localization into  
701 progressively smaller volumes (discrete shear zones) preserved a complete record of the  
702 evolution of deformation at decreasing temperature and pressure conditions. The last stages of  
703 deformation recorded in all three massifs correspond to very shallow depths (<0.3 GPa), but  
704 still elevated temperatures (>700°C, constrained by the absence of chlorite, Figures 14d and  
705 15). However, the petrostructural record in the southern and central/northern massifs implies  
706 markedly different initial conditions. The pervasive deformation recorded in the southern  
707 massif occurred in a shallow (< 30 km depth) lithosphere-asthenosphere boundary (Figure 14),  
708 whereas the pervasive deformation in the central and northern massifs is associated with  
709 thinning of the subcontinental lithospheric mantle. Despite this difference in initial conditions,  
710 the active deformation processes and the role of melts and fluids in the different stages of  
711 mantle exhumation are similar.

712

713 *Pervasive deformation – coarse-porphyroclastic microstructures*

714

715 Analysis of the microstructures and CPO of the coarse-porphyroclastic peridotites indicate  
716 that the pervasive deformation in the three massifs occurred by melt-assisted dislocation creep  
717 under near solidus conditions. The coarse-porphyroclastic peridotites have a clear foliation and  
718 lineation materialized by the elongation and shape preferred-orientation of olivine crystals with  
719 strong intragranular deformation (subgrains and undulose extinction, Figures 5a,b, 9b,d) and  
720 CPO (Figures 10 and 11). Thus olivine, which is the volumetrically dominant phase forming a  
721 continuous stress-bearing framework in peridotites, deforms essentially by dislocation creep.  
722 However, (1) the parallelism between the diffuse pyroxenitic or gabbroic layering and the  
723 foliation (Figure 6) and (2) the irregular shapes of coarse orthopyroxene crystals and cusp-  
724 shaped terminations and film-like habitus of orthopyroxene crystals along olivine grain  
725 boundaries (Figures 5a,b and 6a), which record dissolution and precipitation from a liquid  
726 phase, respectively, indicate that the pervasive deformation occurred under melt-present, near  
727 solidus conditions, in all three massifs. In the southern massif peridotites, the high contents of  
728 plagioclase, which cannot result from subsolidus re-equilibration of spinel-peridotites, further  
729 point to refertilization by percolating melts (Figures 4 and 5a). The interstitial, but elongated  
730 shape of these plagioclases as well as clinopyroxene crystals and aggregates and the fact that  
731 they show clear CPO consistent with the olivine CPO despite having low intragranular  
732 misorientations (Figures 8a,b and 11) further points to an origin by crystallization of melts  
733 coeval with the deformation. The undulose extinction and rare subgrains, but the absence of

734 recrystallization of these melt-derived phases indicate limited pervasive deformation under  
735 subsolidus conditions.

736 The composition of the mafic layers indicates that the high-temperature, near-solidus  
737 deformation that formed the coarse-porphyroclastic foliation occurred in the spinel stability  
738 field in the central and northern massif, but in the plagioclase stability field in the southern  
739 massif. Thermobarometric estimates derived from cores data from coarse-porphyroclastic  
740 peridotites of the southern massif (Figure 13, 1100-1150 °C, 0.6-0.7 GPa) record equilibration  
741 at near solidus conditions at low pressure, consistently with the inferred synkinematic  
742 conditions. In contrast, thermobarometric estimates obtained using cores data from coarse-  
743 porphyroclastic peridotites of the central massif (Figure 13, ~800°C, 0.8 GPa) correspond to  
744 subsolidus conditions, implying chemical re-equilibration of these peridotites at lower  
745 temperatures and pressures after the deformation that produced the pervasive coarse-  
746 porphyroclastic fabric and pyroxenitic layering. The less sinuous pyroxene and olivine grain  
747 boundaries in the coarse-porphyroclastic peridotites and pyroxenites of the central and northern  
748 massifs (Figure 6b) further point to subsolidus re-equilibration (annealing). Both data suggest  
749 a time gap between the pervasive and localized deformations in the central and northern massif,  
750 during which the mantle volume that presently outcrops was maintained at temperatures high  
751 enough to allow for effective solid-state diffusion. This contrasts with the continuous  
752 petrological and microstructural record in the southern massif.

753 The olivine CPO in coarse-porphyroclastic peridotites from all three massifs records  
754 dominant activation of the [100](010) slip system (Figure 11). The moderate axial-[010]  
755 tendency (BA-index between 0.15-0.4, Figure 10) may be explained by deformation in the  
756 presence of low melt fractions (Higgie and Tommasi, 2012, 2014). The CPO of pyroxenes,  
757 which is often is characterized by multiple maxima, with one coherent and others uncorrelated  
758 with the olivine CPO, implies that part of the pyroxenes formed prior to the deformation and  
759 part result from reactive melt percolation during or after the deformation. The intragranular  
760 deformation of the coarse orthopyroxene crystals corroborates their pre-deformation origin.  
761 The higher dispersion of the clinopyroxene CPO may be explained by a higher fraction of  
762 secondary, melt-derived crystals. The plagioclase CPO, which contrasts with the low  
763 intracrystalline deformation of the grains, may result from oriented crystallization in a  
764 deforming rock (Higgie and Tommasi, 2012, 2014).

765 Stresses associated with the pervasive deformation, based on the arithmetic mean of the  
766 equivalent neoblast sizes of olivine and the Van der Wal et al. (1993) piezometer, are on the  
767 range 50-85 MPa, with a mean at 67 MPa (Figure 15).

768

#### 769 *Localized deformation – Protomylonites, mylonites and ultramylonites*

770

771 The early stages of localized deformation, which are preserved in the protomylonites that  
772 compose the external domains of the shear zones, are characterized by the development of low  
773 fractions of a very fine-grained matrix that forms a discontinuous losangular to anastomosed  
774 network at  $\pm 15^\circ$  on average to the elongation of the olivine porphyroclasts (Figure 5c,d, 6c).  
775 This matrix is produced by dynamic recrystallization of olivine, as documented by the core-

776 and-mantle structure of the olivine porphyroclasts (Figure 5c,d, 6c), but also by dissolution-  
777 precipitation processes, which are attested by the shapes of the orthopyroxene porphyroclasts  
778 and the asymmetric tails composed of intermixed fine-grained olivine-orthopyroxene that  
779 depart from them (Figure 5c,d, 6c). These dissolution-precipitation microstructures imply  
780 deformation in presence of aqueous fluids (e.g., Hidas et al. 2016) or water-rich evolved melts  
781 (e.g., Frets et al. 2013, Tholen et al. 2023). The nature of the fluid – hydrous melt or aqueous  
782 Si-bearing fluid – cannot be determined based on the petrological observations alone and will  
783 be further discussed on the next section, based on thermodynamical modelling. The  
784 protomylonites record no first order change in modal composition relative to the coarse-  
785 porphyroclastic peridotites (Figure 4), but display evidence for remobilization of the existing  
786 phases, in particular orthopyroxene (and olivine), and crystallization of minor amounts of  
787 amphibole in the matrix (Figure 5c,d, 6c). The decrease in recrystallized olivine grain sizes  
788 (Figure 3) suggests deformation under higher stresses (75-112 MPa with an average at 93 MPa,  
789 Figure 15, Supplementary Material Table S1) relative to that involved in the pervasive and is  
790 consistent with the lower equilibration temperatures (800-900°C) recorded by the rims of  
791 orthopyroxene crystals in protomylonitic peridotite 85ZA37 (Figure 13d).

792 Progression of the localized deformation is recorded by a gradual increase in the matrix  
793 fraction in the mylonites and ultramylonites (Figure 2, 3, 5e-i, 8). Dislocation creep and  
794 dynamic recrystallization play an essential role in the deformation of olivine, as documented  
795 by the progressive replacement of the porphyroclasts by recrystallized grains (decrease in size  
796 and number of porphyroclasts, Figure 8a) and by the increasing dispersion of the CPO (Figure  
797 10). In the mylonites of the southern massif, plagioclase also dynamically recrystallizes, with  
798 an increase in  $X_{Ca}$  of the neoblasts relative to the porphyroclasts (Figures 8c,d), which records  
799 the decrease in confining pressure (Figure 13d). Evidence for dissolution-precipitation and  
800 hydration reactions is omnipresent in the mylonites and ultramylonites. It encompasses: (1)  
801 decrease in the size and increase in the shape factor of orthopyroxene porphyroclasts (Figure  
802 8c), (2) increasing fractions of fine-grained orthopyroxene, and to a lesser extent  
803 clinopyroxene, with interstitial or film-like habitus in the fine-grained mylonitic matrix and  
804 along recrystallized planes that crosscut obliquely, forming C' planes, olivine porphyroclasts  
805 (Figures 2, 5e-i, 8a,c,e), and (3) increasing amphibole fractions that progressively replace  
806 clinopyroxene and plagioclase in the recrystallized matrix (Figures 2, 5e-i, 8a,c,e). Dissolution  
807 and precipitation are also active during deformation at the brittle-ductile transition as attested  
808 by the fractures in olivine filled by fine-grained orthopyroxene in ultramylonite 85ZA71-13  
809 (Figure 8e). Plagioclase remains nevertheless a common phase both as porphyroclasts and  
810 neoblasts in the matrix in mylonites from the southern massif (Figures 3, 5e, 8c, Table 1).  
811 Plagioclase also occurs as small grains rimming spinel in most mylonites from the central and  
812 northern massifs (Figures 3, 5g, 8a, Table 1). In contrast, the ultramylonites have almost no  
813 plagioclase and only relicts of clinopyroxene (Figures 3, 5i, 8a, Table 1).

814 Evolution from protomylonites to mylonites is accompanied by a minor, but not systematic  
815 decrease in the average recrystallized grain sizes of olivine (Figure 2). Recrystallized olivine  
816 in the ultramylonites has also similar average sizes as in the mylonites. This implies that,  
817 despite the decrease in temperature and pressure conditions of the deformation, recorded by

818 the data on rims and matrix grains of mylonites and ultramylonites (<500-850°C, 6.5-1.5 GPa,  
819 Figure 13d), and increasing strain localization, stresses did not increase, remaining below 110  
820 MPa (Figure 15).

821 Olivine crystal preferred orientations in the mylonites and most ultramylonites (Figure 11)  
822 corroborate that, despite the presence of fluids along grain boundaries, the activation of  
823 dissolution-precipitation processes, and the decreasing temperature conditions, olivine  
824 deformed by dislocation creep with dominant activation of the [100](010) slip system. The  
825 higher dispersion of the olivine CPO relative to that in the coarse-porphyroclastic peridotites  
826 and protomylonites (Figure 10) is consistent with the increase in recrystallized fraction (e.g.,  
827 Falus et al. 2011). Only four amphibole-rich samples show CPO suggesting similar  
828 contributions of [100] and [001] glide to olivine deformation: mylonite 85ZA51 and  
829 ultramylonites 85ZA43b and 85ZA31b, in which olivine is almost completely recrystallized,  
830 and ultramylonite 85ZA71-13 in the southern massif, which displays evidence for deformation  
831 at the brittle-ductile transition (Figure 8e). Ultramylonites 85ZA31b and 85ZA71-13 are  
832 characterized by maxima of both [100] and [001] parallel to the lineation and to the maximum  
833 of [001] of amphibole (Figure 11), suggesting activation of both [100](010) and [001](010).  
834 Ultramylonite 85ZA43b has [100] and [001] maxima normal to the amphibole [001] maximum  
835 (Figure 11). Note that the lineation in this ultramylonite is poorly defined. An increase in the  
836 contribution of [001] glide to the deformation may result from an increase in stress during  
837 deformation during decreasing temperature conditions (e.g., Durham and Goetze 1977,  
838 Demouchy et al. 2013). Incorporation of H<sup>+</sup> in olivine may also favour [001] glide (e.g.,  
839 Katayama et al., 2004), but at much higher pressures and stresses than those inferred for the  
840 deformation of the Zabargad peridotites. Moreover, the H<sup>+</sup> solubility in olivine at low pressures  
841 and temperature conditions is very low (Demouchy and Bolfan-Casanova 2016, Padrón-  
842 Navarta and Hermann, 2017). A possible explanation, already proposed to explain [001]-glide  
843 olivine CPO in ultramylonites with similar microstructures (Hidas et al. 2016), development  
844 of axial-[010] olivine CPO in peridotites deformed in the presence of melts (Higgie and  
845 Tommasi 2012, 2014), or of [001] glide CPO in experimental deformation of olivine+basalt  
846 systems (Holtzman et al 2003) is that the presence of a fluid phase along grain boundaries may  
847 induce additional rotation components and favour orientation of the olivine [001] axis in the  
848 flow direction. The fact that in the present study such olivine CPO are only observed in  
849 ultramylonites with high amphibole contents and almost no plagioclase suggests that  
850 instantaneous fluid fractions have to be rather high for this process to be activated.

851 The pyroxenes CPO are more variable (Figure 11). Parallelism between the pyroxenes [001]  
852 and the olivine [100] maxima suggests deformation by dislocation glide, consistently with the  
853 observed undulose extinction and, for the orthopyroxenes, elongation of the porphyroclasts in  
854 some mylonites. However, some mylonites and ultramylonites show pyroxene CPO with a  
855 concentration of [001], which is the only possible glide direction in pyroxenes (cf. Frets et al.  
856 2012 and references therein), at high angle to the foliation. These CPO are similar to those  
857 described in peridotite mylonites and ultramylonites deformed in presence of aqueous fluids  
858 by Hidas et al. (2016). In dissolution-precipitation creep, CPO development may be controlled  
859 by epitaxy. However, our data do not show any consistent orientation relation between olivine

860 and pyroxenes CPO. The plagioclase CPO in the mylonites and ultramylonites is much weaker  
861 and in most cases different from that in the coarse-porphyroclastic peridotites. The change in  
862 CPO pattern is associated with the development of a marked substructure and dynamic  
863 recrystallization, suggesting that it is due to deformation by dislocation creep with the  
864 activation of multiple slip systems.

865 In conclusion, analysis of the microstructures and CPO in the mylonites and ultramylonites  
866 indicates that dislocation creep played an essential role in deformation in the localized shear  
867 zones. It also documents that, except in the final stages of this deformation recorded in the  
868 ultramylonites, olivine deformed by activating the same slip systems that during deformation  
869 in the asthenosphere of deep lithospheric mantle. Evidence for dissolution-precipitation  
870 processes, indicating local fluid saturation, is, nevertheless, omnipresent. Fluids play a major  
871 role in the mechanical behaviour of the peridotites in the shear zone by (1) allowing  
872 deformation of the pyroxenes by dissolution-precipitation creep, (2) creating, via dissolution-  
873 precipitation, a fine-grained polymineralic matrix in which grain growth will be hindered by  
874 pinning, and (3) lubricating the grain boundaries within this matrix. These fluid-assisted  
875 processes allow for deformation to continue at decreasing temperature conditions and faster  
876 strain rates (localization implies accommodating the same displacement gradient in a smaller  
877 volume) without a major increase in stress. However, the fact that the olivine recrystallized  
878 fraction continuously increased until it represented ~50 % of the olivine volume implies that  
879 dislocation creep remained important even in the ultramylonites. This suggests that  
880 instantaneous fluid fractions remained low and that grain size reduction did not suffice to  
881 produce a switch to a deformation dominated by grain boundary sliding. One may also  
882 speculate that dissolution-precipitation creep did not play a more important role in the  
883 deformation because of the limited solubility of olivine and pyroxenes in aqueous fluids at  
884 temperatures < 900° C (Newton and Manning 2002, Macris et al. 2020).

885

## 886 **Thermodynamical modelling: constraints on the evolution of the pT conditions and** 887 **melts or free fluids contents during deformation**

888 To constrain the composition and amount of fluids involved in the different stages of  
889 deformation, we run thermodynamical models, which estimate the sequence of mineral  
890 reactions and mineral compositional changes in a system with a whole-rock composition  
891 representative for the Zabargad peridotites (cf. Supplementary Material Figure S3). These  
892 models were run under the assumption of thermodynamic equilibrium along the decompression  
893 paths inferred for: (1) the southern massif and (2) the central and northern massifs. The  
894 challenge was to model an open system relative to the H<sub>2</sub>O component. Indeed, except for the  
895 water responsible for low amounts of high-Ti pargasite observed in the coarse-porphyroclastic  
896 peridotites from the central and northern massifs, the petrostructural data in this study imply  
897 that externally derived aqueous fluids (here treated as an external H<sub>2</sub>O influx) have infiltrated  
898 the deforming peridotites either continuously or discontinuously during the decompression.  
899 Depending on the PT conditions, this external H<sub>2</sub>O influx should lead to reactions producing  
900 amphibole and consuming plagioclase and clinopyroxene until hydrous melting or saturation



901 with an aqueous fluid was reached. However, the strong assumption of complete  
902 thermodynamic equilibrium with the external fluid is certainly not fulfilled in the natural  
903 system. The preservation of core-to-rim compositional gradients, bimodal to multimodal grain  
904 size distributions, and contrasting microstructures certificate that complete equilibrium was  
905 never achieved, except in the amphibole-rich ultramylonites that record the most localized  
906 deformation. The thermodynamic simulations predict therefore minimum local instantaneous  
907 fluid contents. Moreover, the thermodynamic modelling was performed using a single  
908 representative bulk whole-rock composition. It predicts therefore evolution trends for mineral  
909 reactions and mineral compositional changes rather than attempting to reproduce the individual  
910 evolution of the diversity of compositions observed in Zabargad peridotites.

911 The bulk rock composition of sample 85ZA37 was estimated based on modal abundances  
912 from EBSD mapping and average mineral chemical compositions from EPMA analyses (cf.  
913 Supplementary Material Table 2). Thermodynamic modelling was conducted using oxides with  
914 additional O<sub>2</sub> to account for potential redox reactions involving ferric iron. The amount of O<sub>2</sub>  
915 component was estimated based on the ferric iron content in spinel inferred by stoichiometry  
916 and the spinel modal abundance in 85ZA37, leading to a bulk oxidation state of iron rather  
917 conservative ( $\text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Fe}^{2+}) = 0.033$  in atom proportions) when compared with that reported  
918 by Bonatti et al. (1986) and Piccardo et al. (1988), which ranges from 0.121 and 0.410 with an  
919 average of 0.213.

920 To account for the contrasting PT paths inferred for the southern massif and the central  
921 and northern massifs (Figure 14d), pseudosections were computed along two different PT  
922 linear decompression trajectories (vertical axis in Figure 16) for a system with variable H<sub>2</sub>O  
923 content: from strictly dry to 1.0 wt. % bulk H<sub>2</sub>O added to the bulk composition of sample  
924 85ZA37 (horizontal axes in Figure 16). Predictions for the evolution of modal and mineral  
925 compositions at equilibrium are then presented for nominally dry (200 ppm wt. H<sub>2</sub>O)  
926 decompression paths (full red and blue lines close to 0 in XH<sub>2</sub>O in Figure 17) and two slightly  
927 different hydrous decompression paths (red and blue thick lines with a central white line in  
928 Figure 16). These hydrous paths diverge from the nominally dry path at 1170 °C for the  
929 northern and central massifs and 1100 °C for the southern massif (first appearance of amphibole  
930 at near water-undersaturated solidus conditions) and are thereafter constrained by the water-  
931 saturated solidus down to ca. 950 °C and 975 °C for northern and central and southern massifs,  
932 respectively, and by the presence of a free aqueous fluid at lower temperatures (aqueous fluid  
933 saturation). In addition, for the southern massif, we also computed modal and mineral  
934 compositions at equilibrium for a melt/aqueous fluid undersaturated path along the near-zero  
935 mode of plagioclase (dashed red line in Figure 16) to estimate the plagioclase composition just  
936 before its complete replacement.

937 The nominally dry PT paths correctly reproduce, for instance, the persistence of spinel  
938 through the entire decompression path in the central and northern massifs and the occurrence  
939 of very coarse (mm-sized) spinel with plagioclase rims in the southern massif peridotites,  
940 suggesting that the southern history initially started in the spinel stability field at 0.9 GPa and  
941 temperatures around 1150°C (Figure 16). In addition, the very high plagioclase content (up to  
942 15%, but in most samples 5-8%) with clear synkinematic crystallization habitus implies

943 nominally dry initial conditions in the southern massif, since an external influx of only 0.1  
944 wt.% at these conditions would have caused the complete disappearance of plagioclase (Figure  
945 16). In the central and northern massifs, the synkinematic crystallization of websterites, which  
946 implies conditions close to the hydrous solidus, and the assemblages of ol + opx + cpx + sp +  
947 Ti-rich amphibole, with neither garnet nor garnet pseudomorphs, such as aggregates of  
948 opx+cpx+sp, in the coarse-porphroclastic peridotites and websterites constrain initial  
949 conditions for the pervasive deformation of at least 1170°C and 1.7 GPa and a bulk H<sub>2</sub>O content  
950 in the system lower than 0.2 wt. %. The occurrence of orthopyroxene and amphibole in  
951 websteritic layers in the north and central massif support the hydrous nature of the melts.

952 Departure from the nominally dry path during further decompression are however  
953 constrained by the microstructural evidence for fluid-assisted deformation. Thermodynamical  
954 simulations predict that, at equilibrium, addition of external H<sub>2</sub>O to the system systematically  
955 results in crystallization of amphibole, with complete replacement of plagioclase (in the north)  
956 or of Al-rich spinel (in the central and northern massifs) by amphibole, before reaching  
957 conditions at which a free melt or aqueous fluid may exist (Figure 16). At temperature below  
958 840 °C, fluid-saturation conditions further require complete replacement of clinopyroxene  
959 leaving a mineral assemblage consisting on olivine + amphibole + Cr-rich spinel in both  
960 massifs. The hydrous melt/fluid-saturated decompression paths would have produced  
961 essentially the same mineral assemblage in both massifs if equilibrium had been reached  
962 continuously during cooling and fluid infiltration, in contrast to the observations. Equilibrium  
963 seems, nevertheless, to have been achieved in the ultramylonite 85ZA11, which contains 25  
964 area % of amphibole, almost no clinopyroxene, and has spinels with high Cr contents (compare  
965 Figure 16 to Figures 4 and 13). The appearance of a Cr-rich spinel in the ultramylonites of the  
966 southern massif constrains the temperatures and pressures in the last deformation stages to  
967 < 845 °C and 0.4 GPa and requires instantaneous H<sub>2</sub>O contents of at least 0.6 wt %. The absence  
968 of chlorite even in the most hydrated samples constrain the minimum temperature of  
969 deformation to 700 °C (Figure 16). The preservation of coarse spinels rimmed by plagioclase  
970 (in all massifs) and the persistence of plagioclase in the southern massif mylonites requires a  
971 spatially heterogeneous fluid distribution and probably heterogeneity of the fluid flux through  
972 time, which prevented the reactions to go to completion while still allowing for widespread  
973 deformation by dissolution-precipitation.

974 The observed mineral chemistry trends are also in relatively good agreement with the  
975 thermodynamic modelling predictions given the provision that the latter represent the  
976 equilibrium compositions towards which the evolution should trend, ignoring any kinetics  
977 effects. The comparison between the modelled and observed mineral chemical compositions  
978 (Figure 17) may be only made in a qualitative way, because, in addition to the limitation of  
979 partial local re-equilibration, a single bulk rock composition was used as a representative  
980 composition to model the mineral composition, whereas the actual Zabargad peridotites show  
981 a relatively large compositional range (see Supplementary Material Figure S3).

982 Under nominally dry conditions, the evolution of orthopyroxene composition is strongly  
983 dependent on the decompression and cooling path followed. In contrast, if hydrous  
984 decompression paths were followed under complete equilibrium, orthopyroxenes from all three

985 massifs should display similar compositional evolutions (Figure 17a). The fact that different  
986 trends in Ca/Al ratios can be still observed (mostly in cores) – with higher ratios in the southern  
987 massif and an almost constant Ca content for a range of Al contents in the central and northern  
988 massifs – supports initial low H<sub>2</sub>O contents and limited re-equilibration of the orthopyroxene  
989 porphyroclasts. The similar compositions of the rims and matrix orthopyroxenes in the three  
990 massifs are consistent with hydrated systems. The weak decrease in X<sub>Mg</sub> with decompression  
991 and hydration predicted by the models is probably masked in the observations by the variability  
992 in X<sub>Mg</sub> among and within the samples (Figure 17b). Both dry and water-saturated models do  
993 consistently predict, however, the increase in and decrease in Al content in clinopyroxenes  
994 along the decompression and cooling paths (Figure 17c). It is worth noting, however, that the  
995 most recent subsolidus models for orthopyroxene (clinopyroxene) from Tomlinson and  
996 Holland (2021) systematically overestimated (underestimated) the Al contents at all conditions  
997 relative to the present observations and also through direct comparison with the experimental  
998 data of Fumagalli et al. (2017). The same discrepancy is observed if previous suprasolidus  
999 models from Jennings and Holland (2015) and Holland et al. (2018) are used.

1000 The observed progressive enrichment in the anorthite component in plagioclase during  
1001 decompression and cooling (see also Fumagalli et al. 2017) is also correctly modelled, but  
1002 under nominally dry conditions X<sub>Ca</sub> never reaches values higher than 0.78 (Figure 17d).  
1003 Hydrous models predict, however, significant additional enrichment in X<sub>Ca</sub> under isothermal-  
1004 isobaric conditions in response to fluid influx (cf. Supplementary Material Figure S4).  
1005 Anorthitic plagioclases similar to that observed in the matrix of mylonites of the southern  
1006 massif are predicted to be in equilibrium at temperatures of ~700 °C and ~0.6 wt % H<sub>2</sub>O  
1007 (Figures 16 and 17d). However, under water-saturated conditions, at equilibrium, plagioclase  
1008 should be fully replaced by amphibole for the entire central/northern path and between 1100-  
1009 700 °C for the southern path, leading to a gap in the predicted plagioclase anorthite content  
1010 (Figure 17d). This contrasts with the almost continuous increase in X<sub>Ca</sub> in plagioclase observed  
1011 in the Zabargad peridotites, which can be modelled if a melt/aqueous fluid undersaturated path  
1012 along the near-zero mode of plagioclase (dashed red line in Figure 17) is considered. The  
1013 composition of the last plagioclase in equilibrium with amphibole under fluid-undersaturated  
1014 conditions for the southern pT path (cf. Figure 17d). Finally, the X<sub>Ca</sub> values lower than 0.60  
1015 observed in coarse-porphyroclastic peridotites from the central massif (Figure 17d) can only  
1016 be modelled under nominally dry conditions if whole rock compositions with higher Na<sub>2</sub>O/CaO  
1017 ratios than that of sample 85ZA37 are considered (such as the composition HNaFLZ in  
1018 Fumagalli et al. 2017). This change in whole rock composition would have the consequence to  
1019 displace the plagioclase maximum pressure stability up to 0.9 GPa at 1150 °C.

1020 The thermodynamic models predict of a stability gap and sharp increase in the Cr-content  
1021 and decrease in X<sub>Mg</sub> in spinel along the H<sub>2</sub>O saturated decompression and cooling paths  
1022 (Figures 16 and 17e). This compositional evolution is consistent with the observations from all  
1023 three massifs (Figures 13f and 17e). The evolution in the spinel composition is variably  
1024 recorded in the three massifs, but all samples show a consistent enrichment of Cr-content  
1025 towards the rims that requires partial re-equilibration with amphibole and therefore external  
1026 fluid infiltration. An outstanding case is recorded by sample 85ZA50, which displays the most

1027 extreme compositional core-rim range (Figure 13f). The absence of the spinel stability gap in  
1028 the observations may be explained by only partial re-equilibration of the samples during the  
1029 decompression and cooling paths. Note that the spinel model from Tomlinson and Holland  
1030 (2021) used in the present thermodynamic models successfully reproduces the Cr evolution  
1031 trend, but fails to quantitatively reproduce the observed  $X_{Mg}$  range with the discrepancy  
1032 enhanced at the highest Cr-content. This failure might suggest a poor extrapolation of the model  
1033 to the low temperature conditions inferred for these spinels (<840 °C).

1034 The H<sub>2</sub>O saturated decompression and cooling models also reproduce the general observed  
1035 trend from pargasite to tremolite-rich compositions reflected in the steady decrease in the  
1036 amphibole occupancy of the A site (Na + K) and the associated linear decrease in Al<sup>IV</sup> with  
1037 increasing hydration during decompression and cooling (Figure 17f). However, the model of  
1038 Green et al (2016) used in the present study systematically overestimates the Al<sup>IV</sup> content  
1039 (Figure 17f). The hydrous models also consistently predict a decrease in Ti content in  
1040 amphibole with decreasing temperature, but the actual Ti contents are strongly dependent on  
1041 the Ti content in the whole rock and on the amphibole modal content (Figure 17g). Moreover,  
1042 the amphiboles in Zabargad peridotites show first decrease in Ti with only minor increase in  
1043 Si content and then increase in Si (i.e., in the tremolite component) under constant a low Ti  
1044 content (compare Figures 14 and 17g). The amphibole model of Green et al. (2016), which, in  
1045 contrast to the other models used in this work, was calibrated for mafic systems, is not able to  
1046 reproduce this two-stage change in Ti vs. Si contents and predicts lower Si contents than the  
1047 usual ones in pargasite (Figure 17g) calling for new amphibole models for ultramafic  
1048 compositions.

1049

## 1050 **Implications for the evolution of the Red Sea**

1051 The initial deformation in the central/northern and southern massifs records different  
1052 stages of the rifting process. In both cases, the initial deformation occurred in the presence of  
1053 melts and is pervasive at the hundreds of meters to the kilometer scale. However, the peridotites  
1054 of the central and northern massifs record thinning of the subcontinental lithospheric mantle,  
1055 with deformation initially assisted by reactive percolation of hydrous melts (probably related  
1056 to the Afar plume) and later by aqueous fluids. The pervasive deformation occurred at  
1057 temperatures of 1150-1180 °C and pressures of 1.45-1.65 GPa (Figure 16) that is, at depths of  
1058 50-70 km in the subcontinental lithosphere, if one considers a 35 km-thick crust. Subsequent  
1059 thinning and exhumation of this subcontinental lithospheric mantle by shearing in localized  
1060 shear zones followed a steep, almost isothermal decompression path (Figure 14d). This steep  
1061 decompression path is consistent with that recorded by the lower crust gneisses in contact with  
1062 the peridotites of the central massif (Boudier et al. 1988, Boullier et al. 1997, Seyler and Bonatti  
1063 1988). The equilibrated textures of the coarse-porphyroclastic peridotites in these massifs may  
1064 indicate a time gap between the pervasive and localized deformations, but it could also be  
1065 explained by fast grain boundary migration, allowing for effective textural re-equilibration  
1066 during the deformation in presence of hydrous melts (Tommasi et al. 2017). In contrast, the  
1067 peridotites of the southern massif record mantle exhumation in in a more mature system, at the

1068 rift-to-drift transition. The pervasive deformation occurred in a shallow <30 km deep  
1069 lithosphere-asthenosphere boundary, in the presence of melts produced by decompression  
1070 upwelling in the asthenosphere. The subsequent exhumation of this lithosphere-asthenosphere  
1071 boundary occurred along a path in which decompression is accompanied by fast cooling.

1072 The final stages of mantle exhumation recorded in the shear zones in all three massifs are  
1073 marked by an increasing contribution of fluids to the deformation, with increasing amounts of  
1074 sea water (Figures 13d and 16). This channelling of surface fluids indicates that these shear  
1075 zones crosscut the entire lithosphere, being connected to brittle faults that attained the surface.  
1076 The equilibrium conditions recorded in the ultramylonites, including that preserving  
1077 microstructures typical of deformation at the brittle to ductile transition, indicate that high  
1078 geothermal gradients and hence that ductile deformation extended up to very shallow depths  
1079 (700°C at < 0.2 GPa, that is, ~6 km depth). These conditions are equivalent to those expected  
1080 under an active ridge. The presence of fluids and activation of fluid-assisted deformation  
1081 processes allowed for increasingly localized deformation at decreasing temperature at roughly  
1082 constant stresses (Figure 15). The presence of scapolites in the sample deformed at the brittle-  
1083 ductile transition (Figure 8f) suggests the presence of an evaporitic layer between the  
1084 peridotites and the sea water during these last stages of deformation. Such a layer might have  
1085 protected the exhumed peridotites from extensive reaction with seawater, explaining their  
1086 exceptional freshness (almost complete absence of serpentines).

1087

## 1088 CONCLUSION

1089 The present study documents thinning and exhumation of the mantle during continental  
1090 rift and rift-to-drift transition accommodated in extensional shear zones. The petrostructural  
1091 data in the Zabargad peridotites records both progressive strain localization and fluid-focusing  
1092 in these extensional shear zones. Coupling of microstructural and petrological analyses enables  
1093 documenting the full exhumation history of mantle sections up to subsurface conditions at  
1094 different stages of the rifting process. The central and northern massifs record an evolution  
1095 starting at depths of 50-70 km in the subcontinental lithospheric mantle (1150-1180 °C, 1.45-  
1096 1.65 GPa) followed by a steep, almost isothermal decompression path. The southern massif  
1097 represents a more mature system, at the rift-to-drift transition. Deformation started in a shallow  
1098 <30 km deep lithosphere-asthenosphere boundary (1100-1200°C, 0.78-0.90 GPa), which was  
1099 exhumed along a path in which decompression is accompanied by fast cooling.

1100 The association of high-resolution cartography of the microstructure by EBSD with  
1101 thermodynamic modelling documents the deformation of the olivine load-bearing framework  
1102 of the peridotites essentially by dislocation creep, despite the evidence for the presence of, first,  
1103 melts, and, later, aqueous fluids within the extensional shear zones. The presence of fluids  
1104 enabled activation of additional deformation processes, such as dissolution-precipitation,  
1105 which: (1) contributes to deformation by advective transport of matter along stress gradients,  
1106 (2) reduces strain incompatibility by dissolving ortho- and clinopyroxene, which require higher  
1107 stresses to deform by dislocation creep than olivine, (3) effectively reduces grain sizes by  
1108 producing a well-mixed polymineralic fine-grained matrix. In addition, melts and fluids

1109 lubricate grain boundaries, enhancing grain boundary mobility and probably, locally, grain  
1110 boundary sliding.

1111 The evolution of olivine recrystallized grain sizes in the mylonites and ultramylonites  
1112 indicates that stresses varied little despite the decrease in temperature conditions and the  
1113 localization of the deformation in smaller and smaller volumes. This points to the progressive  
1114 weakening of the rocks within the shear zones, which we interpret as due to increasing fluids  
1115 implication in the deformation. However, the fact that the olivine recrystallized fraction  
1116 continuously increased implies that dislocation creep remained important even in the  
1117 ultramylonites, suggesting that fluid-assisted processes alone did not suffice to accommodate  
1118 the imposed deformation.

1119 The thermodynamic models constrain that decompression and cooling is accompanied by  
1120 infiltration of increasing amounts of externally derived aqueous fluids. However, the  
1121 preservation of core-to-rim compositional gradients and, specially, the preservation of  
1122 plagioclase and clinopyroxene in the mylonites and to a lesser extent ultramylonites suggests  
1123 that complete thermodynamic equilibrium with the external fluid was never achieved, except  
1124 in a few ultramylonites. This implies spatially and probably temporally heterogeneous fluid  
1125 distribution. Despite the limitations of representing the diversity of compositions with a single  
1126 representative bulk composition and assuming instantaneous equilibrium, the models for H<sub>2</sub>O-  
1127 saturated decompression and cooling paths successfully predict the general evolution trends  
1128 for the orthopyroxene, clinopyroxene, plagioclase, and amphibole compositions shedding light  
1129 on the hydration process during decompression and cooling. Systematic discrepancies between  
1130 modelled and observed compositions highlight, however, limitations in the presently available  
1131 models for pyroxenes, spinel, and amphiboles in ultramafic systems.

## 1132 **DATA AVAILABILITY**

1133 The data underlying this article are available in the article and in its online supplementary  
1134 material.

## 1135 **SUPPLEMENTARY MATERIAL**

1136 Supplementary files are available at Journal of Petrology online.

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1148

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- 1296

## 1297 **FIGURE CAPTIONS**

1298 Figure 1. Map of the three massifs showing the spatial distribution of studied samples and the  
 1299 orientation of foliations and lineations for both pervasive and localized deformations (after  
 1300 Nicolas et al, 1987). Squares, diamonds, and circles correspond to samples from the southern,  
 1301 central, and northern massifs, respectively. The red-to-blue colour scale reflects the variations  
 1302 in deformation microstructure (cf. Figure 3). Shades of red indicate samples with coarse  
 1303 porphyroclastic microstructures, associated with the initial pervasive deformation. Shades of  
 1304 blue indicate samples from the localized deformation (metric to cm-scale shear zones) with  
 1305 lighter shades indicating higher recrystallized fractions: dark purple symbols mark  
 1306 protomylonites, medium blues, mylonites and light blues, ultramylonites. Point 85ZA71 in the

1307 southern massif is a scree, in which loose rocks with highly variable microstructures were  
1308 collected.

1309

1310 Figure 2. (a) EBSD phase map and (b) map illustrating the separation of porphyroclasts  
1311 (purple) from neoblasts (orange) based on the double criteria of a Grain Orientation Spread  
1312 (GOS)  $\geq 1^\circ$  and an equivalent grain diameter  $\geq 400 \mu\text{m}$  for a domain of mylonitic peridotite  
1313 85ZA45.

1314

1315 Figure 3. (a) Olivine neoblasts area-weighted mean equivalent grain size vs. neoblasts area  
1316 fraction (relative to the total olivine area fraction). (b) Olivine neoblasts area-weighted mean  
1317 equivalent grain size vs. orthopyroxene area-weighted mean equivalent grain size. Squares,  
1318 diamonds, and circles correspond to samples from the southern, central, and northern massifs,  
1319 respectively.

1320

1321 Figure 4. Modal composition as a function of the bulk (all phases) recrystallized fraction (cf.  
1322 Figure 2).

1323

1324 Figure 5. Phase maps derived from EBSD data (a,c,e,g,i) and photomicrographs (b,d,f,h)  
1325 illustrating typical (a,b) coarse-porphyroclastic, (c,d) protomylonitic, (e,f,g,h) mylonitic, and  
1326 (i, j) ultramylonitic microstructures.

1327

1328 Figure 6. Phase maps derived from EBSD data illustrating details of the (a) coarse-  
1329 porphyroclastic microstructure in plagioclase-rich peridotite 85ZA1 from the southern massif,  
1330 (b) cm-scale pyroxenitic layering and coarse-porphyroclastic microstructure spinel-peridotite  
1331 86Z30 from the central massif, (c) protomylonitic microstructure in plagioclase and  
1332 clinopyroxene peridotite 85ZA37 from the southern massif, and (d) sample 85ZA26C from the  
1333 southern massif with gabbroic to troctolitic composition displaying a coarse-porphyroclastic  
1334 microstructure crosscut by a amphibole-rich vein.

1335 Figure 7. (a,c,e,g,i) Misorientation relative to the mean orientation of the grain and (b,d,f,h,j)  
1336 Kernel Average Misorientation (KAM) maps depicting the internal deformation of the grains  
1337 in the different microstructural groups. High KAM values materialize the subgrain  
1338 boundaries. Tones of red = olivine, blue = orthopyroxene, green = clinopyroxene, yellow =  
1339 plagioclase, purple = spinel, and cyan = amphibole. Grain boundaries (misorientations  $> 15^\circ$ )  
1340 in black.

1341 Figure 8: Phase maps derived from EBSD data (a,c,e) and compositional maps derived from  
1342 EDS data (b,d,f) illustrating microstructurally controlled variations in mineralogic  
1343 composition in mylonites (a-d) and ultramylonites (e,f). (b) Ti map documenting two  
1344 generations of amphibole replacing clinopyroxene: coarse grains with high Ti contents  
1345 surrounded by finer-grained Ti-poor amphiboles, in mylonite 85ZA45 of the northern massif.  
1346 (d) Ca+Na maps with red to blue tones documenting the increase in anorthite content in  
1347 plagioclase from the core of porphyroclasts to the recrystallized grains in mylonite 95ZA7b

1348 of the southern massif. (e) Cl map in which cyan tones document the occurrence of two  
1349 generations of scapolite with different Cl-contents in ultramylonite 85ZA71-13 of the  
1350 southern massif.

1351 Figure 9. Statistical microstructural data for olivine and orthopyroxene porphyroclasts: (a)  
1352 grain size, defined by the equivalent grain diameter, (b) elongation, defined by the aspect ratio,  
1353 (c) shape, defined by the shape factor, and (d) internal deformation, defined by the mean grain  
1354 orientation spread (GOS). The values displayed are area-weighted averages over the entire  
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1356 small grains that occupy in most cases a limited surface of the thin section. Legend as in Figure  
1357 1: squares = Southern massif, circles = Central massif; diamonds = Northern massif; redish  
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1360 dashed line marks the 1:1 relation.

1361  
1362 Figure 10. Olivine crystal preferred orientation symmetry (quantified by the BA index) vs  
1363 intensity (quantified by J index): (a) Average values over the entire olivine population and (b)  
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1365 same sample are linked by a line. Inset in (b) illustrates the difference in crystal preferred  
1366 orientation of olivine between porphyroclasts and neoblasts (mean orientation of the grain) for  
1367 a representative sample. Stereographic projections in the structural reference frame; contours  
1368 at 1 multiple of a uniform distribution intervals.

1369  
1370 Figure 11. Crystal preferred orientations of the main rock-forming minerals in samples  
1371 representative of the different deformation facies in the (a) central and northern massifs and (b)  
1372 southern massif. Contoured lower hemisphere stereographic projections in the structural  
1373 reference frame (X is the lineation and Z the normal to the foliation) of all measurements in  
1374 the map. Contours at 1 multiple of a uniform density intervals. Pie diagrams on the right show  
1375 the modal composition of the samples (color legend as in the phase maps in Figure 6: olivine  
1376 = red; orthopyroxene = blue; clinopyroxene = green; plagioclase = yellow; amphibole = cyan;  
1377 spinel = purple) with the hatched portion representing the porphyroclasts' area fraction for each  
1378 mineral.

1379  
1380 Figure 12. Compositions of olivine, orthopyroxene, clinopyroxene, plagioclase, and spinel in  
1381 atoms per unit formula. Symbols indicate the type of analysis: core, intermediate, or rims of  
1382 porphyroclasts, neoblasts within the fine-grained matrix, or inclusion within another mineral  
1383 phase. Warm colors indicate samples from the southern massif, greens, samples from the  
1384 central massif and blues, samples from the northern massif. Lighter tones indicate more  
1385 recrystallized samples.

1386  
1387 Figure 13. Compositions of amphiboles in atoms per unit formula. Symbols indicate the type  
1388 of analysis: coarse or fine grains in the mylonitic matrix, rims around clinopyroxene or  
1389 plagioclase, or crystals formed by static replacement in the vicinity of fractures. Warm colors

1390 indicate samples from the southern massif, greens, from the central massif and blues, from the  
1391 northern massif. Lighter tones indicate more recrystallized samples.

1392

1393 Figure 14. Equilibrium temperature and pressure conditions estimated for the different  
1394 microstructural assemblages in each sample: cores, intermediate positions, and rims of  
1395 porphyroclasts or neoblasts from the recrystallized matrix. (a) Al in orthopyroxene (in atoms  
1396 per formula unit, apfu) vs temperatures estimated using the Al in Opx thermometer, the  
1397 difference in temperature estimates for similar Al contents between samples from the Central  
1398 and North massifs and the Southern massif denote the effect of pressure on this thermometer.  
1399 (b) Ca in orthopyroxene (apfu) vs temperatures estimated using Ca in Opx thermometer. (c)  
1400 Al<sub>IV</sub> in clinopyroxene vs XCa in plagioclase; the experimental data of Fumagalli et al. (2017)  
1401 used to calibrate the FACE barometer is presented for reference. d. Ca in Opx temperatures vs  
1402 pressures estimated using the FACE barometer for all plagioclase-bearing microstructural  
1403 assemblages. In (a) and (b) results of individual analyses are presented (errorbars are smaller  
1404 than the symbols). In (c) and (d) the uncertainty due to spatial variations in the minerals  
1405 composition within a sample was estimated by considering all possible combinations between  
1406 cores, intermediate position, rim and matrix analyses for the different minerals in each sample;  
1407 it is represented by plotting the median with errorbars representing the dispersion of the  
1408 estimated values for each microstructural class.

1409

1410 Figure 15. Paleostresses determined based on the mean olivine neoblast equivalent grain size  
1411 using the piezometer of Van de Wal et al. (1993) for all studied peridotites.

1412

1413 Figure 16. Evolution of the system composition as a function of the water content (XH<sub>2</sub>O wt.  
1414 %) for the pT paths inferred for the northern/central (left, blue) and southern (right, red) massifs  
1415 for the bulk rock composition of sample 85ZA36, chosen as representative of the average  
1416 composition of the massif. The full lines on the pseudosections indicate the paths used for  
1417 calculating the evolution of mineral compositions during exhumation: the simple line marks an  
1418 almost anhydrous path (100 ppm H<sub>2</sub>O), the composite line, a path following the melt or fluid  
1419 saturation surface, and the dashed line, a melt/aqueous fluid undersaturated path along the near-  
1420 zero mode of plagioclase. Coloured squares represent the stable modal compositions of the  
1421 solid matrix: Ol and Opx are not represented because they are stable over the entire field, for  
1422 the other phases colors as in the phase maps: green = Cpx, yellow = Plg, cyan = Amph, gray =  
1423 Sp. Garnet, which is not observed in the studied peridotites is indicated in brown. The panels  
1424 below each pseudosection represent the stability field and modal contents of the four Al-  
1425 bearing phases. Grey contours in the bottom left of the plagioclase content panel for the  
1426 southern massif document the stability field and modal content of chlorite.

1427

1428 Figure 17. Comparison between the mineral compositions predicted by the thermodynamical  
1429 calculations for the four pT-hydration paths and those measured in the Zabargad peridotites.  
1430 For the calculated compositions, the filling of the symbols indicates the temperature and their  
1431 contours the nature of the path: black= almost anhydrous, blue=melt or fluid-saturated, with

1432 squares for the northern/central pT path and circles for the southern pT path. Gray circles in  
1433 Figure 17d indicate the evolution of  $X_{Ca}$  along a melt/aqueous fluid undersaturated path along  
1434 the near-zero mode of plagioclase. For the measured data, symbols as in Figures 11 and 12  
1435 with data from the southern massif in black, central massif in gray, and northern massif in  
1436 white.

1437

1438 Table 1 : Samples provenance (massif + lat/lon, modal composition (area fraction, determined  
1439 from the EBSD maps covering at least 90% of the thin section with an uncertainty > 1%),  
1440 deformation facies, total recrystallized fraction, and pT equilibrium conditions.

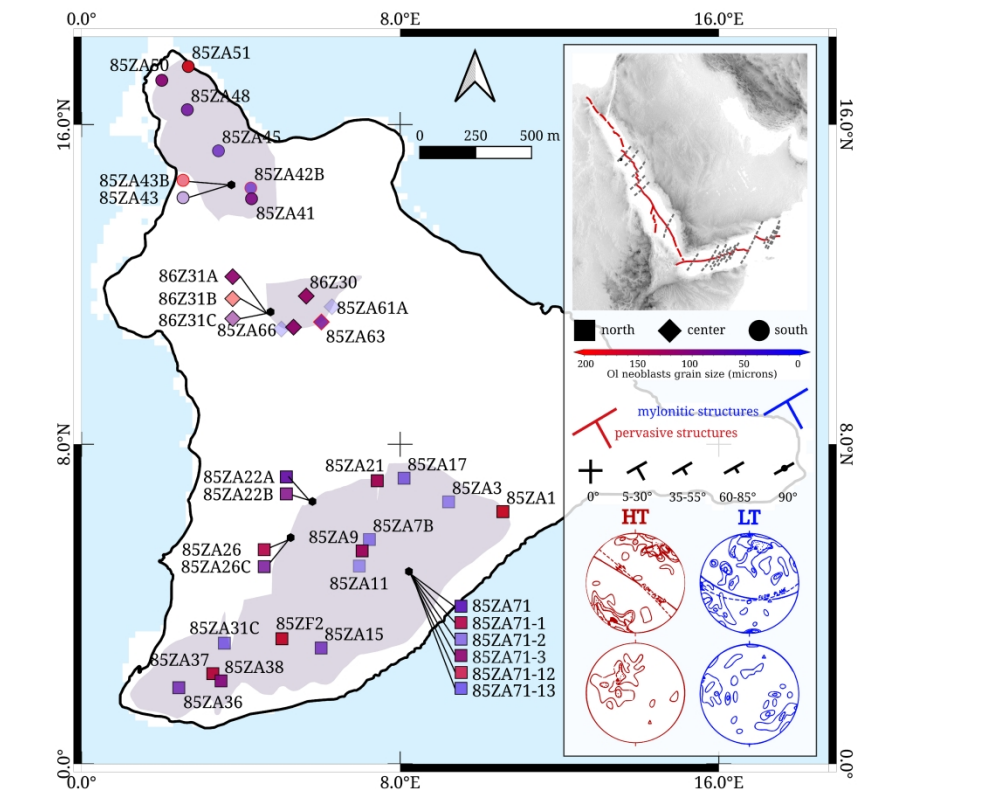


Figure 1. Map of the three massifs showing the spatial distribution of studied samples and the orientation of foliations and lineations for both pervasive and localized deformations (after Nicolas et al, 1987). Squares, diamonds, and circles correspond to samples from the southern, central, and northern massifs, respectively. The red-to-blue colour scale reflects the variations in deformation microstructure (cf. Figure 3). Shades of red indicate samples with coarse porphyroclastic microstructures, associated with the initial pervasive deformation. Shades of blue indicate samples from the localized deformation (metric to cm-scale shear zones) with lighter shades indicating higher recrystallized fractions: dark purple symbols mark protomylonites, medium blues, mylonites and light blues, ultramylonites. Point 85ZA71 in the southern massif is a scree, in which loose rocks with highly variable microstructures were collected.

432x373mm (300 x 300 DPI)

SAMPLES				COORDINATES		COMPOSITIONS (%)						OLIVINE NEOBLASTS	
massif	lithology	facies	sample	longitude	latitude	OI	Opx	Cpx	Plg	Sp	Amph	area fraction (%)	grain size ( $\mu\text{m}$ )
center	peridotite	coarse porphyroclastic	86Z30	36°11'35"	23°36'54"	45	29	17	0	0	5	3	104
center	peridotite	coarse porphyroclastic	85ZA65	36°11'32"	23°36'49"	51	19	24	0	1	2	6	102
center	peridotite	coarse porphyroclastic	86Z31A	36°11'29"	23°36'52"	45	34	11	0	1	6	6	115
center	peridotite	mylonite	86Z31C	36°11'29"	23°36'52"	61	17	0	0	0	19	47	98
center	pyroxenite	ultramylonite	86Z31B	36°11'29"	23°36'52"	0	33	49	1	0	13	56	197
center	pyroxenite	ultramylonite	85ZA63	36°11'37"	23°36'50"	3	8	42	11	2	31	77	82
center	pyroxenite	ultramylonite	85ZA61A	36°11'39"	23°36'53"	0	42	1	0	1	54	99	18
center	pyroxenite	ultramylonite	85ZA66	36°11'31"	23°36'49"	0	13	26	0	0	59	100	21
north	peridotite	mylonite	85ZA51	36°11'16"	23°37'30"	81	6	0	0	0	11	15	168
north	pyroxenite	mylonite	85ZA42B	36°11'26"	23°37'11"	0	28	57	2	0	10	18	62
north	peridotite	mylonite	85ZA50	36°11'12"	23°37'28"	69	16	4	0	1	7	22	96
north	peridotite	mylonite	85ZA48	36°11'16"	23°37'24"	66	13	1	0	1	16	35	82
north	pyroxenite	mylonite	85ZA43	36°11'23"	23°37'12"	10	33	0	0	0	55	47	173
north	peridotite	mylonite	85ZA45	36°11'21"	23°37'17"	60	20	2	0	2	15	49	68
north	peridotite	protomylonite	85ZA41	36°11'26"	23°37'10"	61	15	2	0	2	17	4	91
north	peridotite	ultramylonite	85ZA43B	36°11'23"	23°37'12"	60	9	0	0	0	29	66	71
south	peridotite	coarse porphyroclastic	85ZA21	36°11'46"	23°36'25"	66	19	7	4	0	1	6	114
south	peridotite	coarse porphyroclastic	85ZA26	36°11'32"	23°36'16"	73	19	2	1	0	1	6	143
south	peridotite	coarse porphyroclastic	85ZA1	36°12'6"	23°36'20"	62	18	2	13	1	1	7	155
south	peridotite	coarse porphyroclastic	85ZA71-1	36°11'51"	23°36'11"	62	19	6	9	0	1	7	143
south	peridotite	coarse porphyroclastic	85ZA71-3	36°11'51"	23°36'11"	53	0	33	8	0	3	7	113
south	peridotite	coarse porphyroclastic	85ZF2	36°11'31"	23°36'0"	63	24	6	4	0	0	13	152
south	pyroxenite	coarse porphyroclastic	85ZA9	36°11'43"	23°36'14"	35	10	30	6	0	16	14	108
south	peridotite	mylonite	85ZA3	36°11'57"	23°36'22"	45	19	21	9	0	3	16	40
south	peridotite	mylonite	85ZA22B	36°11'35"	23°36'22"	54	24	11	6	0	2	16	98
south	pyroxenite	mylonite	85ZA71-12	36°11'51"	23°36'11"	22	6	30	38	1	1	18	154
south	peridotite	mylonite	85ZA37	36°11'20"	23°35'55"	57	23	9	7	0	0	19	120
south	pyroxenite	mylonite	85ZA26C	36°11'32"	23°36'16"	15	19	32	25	0	7	20	87
south	peridotite	mylonite	85ZA38	36°11'21"	23°35'54"	64	17	7	4	0	5	22	94
south	peridotite	mylonite	85ZA36	36°11'14"	23°35'53"	44	31	11	9	0	1	22	72
south	peridotite	mylonite	85ZA7B	36°11'44"	23°36'16"	52	24	8	9	0	4	39	46
south	peridotite	mylonite	85ZA71-2	36°11'51"	23°36'11"	54	15	0	0	0	29	46	37
south	peridotite	mylonite	85ZA71-13	36°11'51"	23°36'11"	64	24	0	0	0	9	48	36
south	peridotite	protomylonite	85ZA15	36°11'37"	23°35'59"	63	22	7	5	0	0	4	69
south	peridotite	protomylonite	85ZA31C	36°11'22"	23°35'60"	59	13	11	6	0	9	6	54
south	peridotite	protomylonite	85ZA22A	36°11'35"	23°36'22"	72	22	2	0	0	1	8	73
south	peridotite	protomylonite	85ZA71	36°11'51"	23°36'11"	53	27	3	0	0	15	12	61
south	peridotite	protomylonite	85ZA17	36°11'50"	23°36'26"	60	18	7	9	0	3	13	55
south	peridotite	ultramylonite	85ZA11	36°11'43"	23°36'12"	67	8	1	0	0	21	52	35



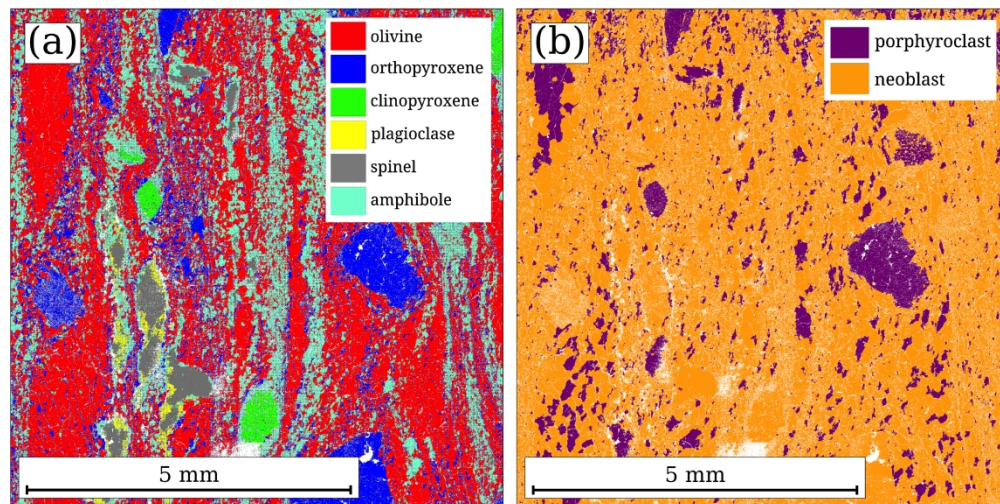


Figure 2. (a) EBSD phase map and (b) map illustrating the separation of porphyroclasts (purple) from neoblasts (orange) based on the double criteria of a Grain Orientation Spread (GOS)  $\geq 1^\circ$  and an equivalent grain diameter  $\geq 400 \mu\text{m}$  for a domain of mylonitic peridotite 85ZA45.

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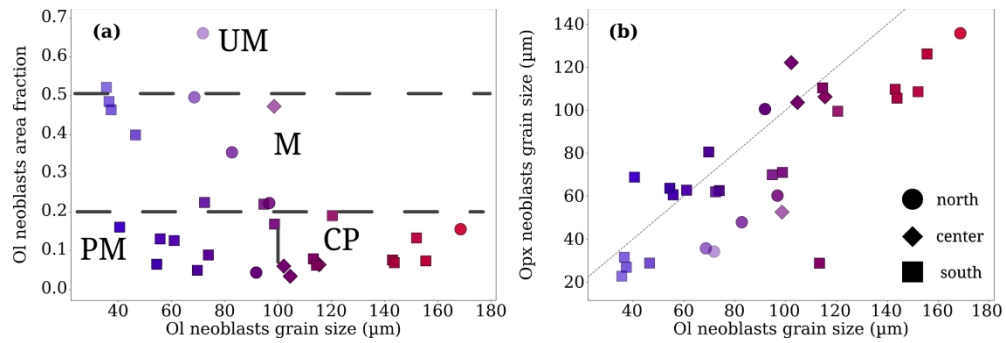


Figure 3. (a) Olivine neoblasts area-weighted mean equivalent grain size vs. neoblasts area fraction (relative to the total olivine area fraction). (b) Olivine neoblasts area-weighted mean equivalent grain size vs. orthopyroxene area-weighted mean equivalent grain size. Squares, diamonds, and circles correspond to samples from the southern, central, and northern massifs, respectively.

161x53mm (600 x 600 DPI)

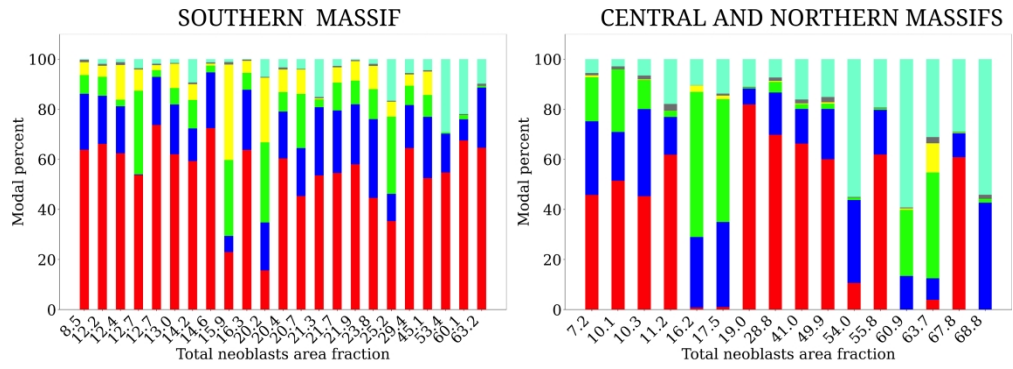


Figure 4. Modal composition as a function of the bulk (all phases) recrystallized fraction (cf. Figure 2).

181x64mm (600 x 600 DPI)

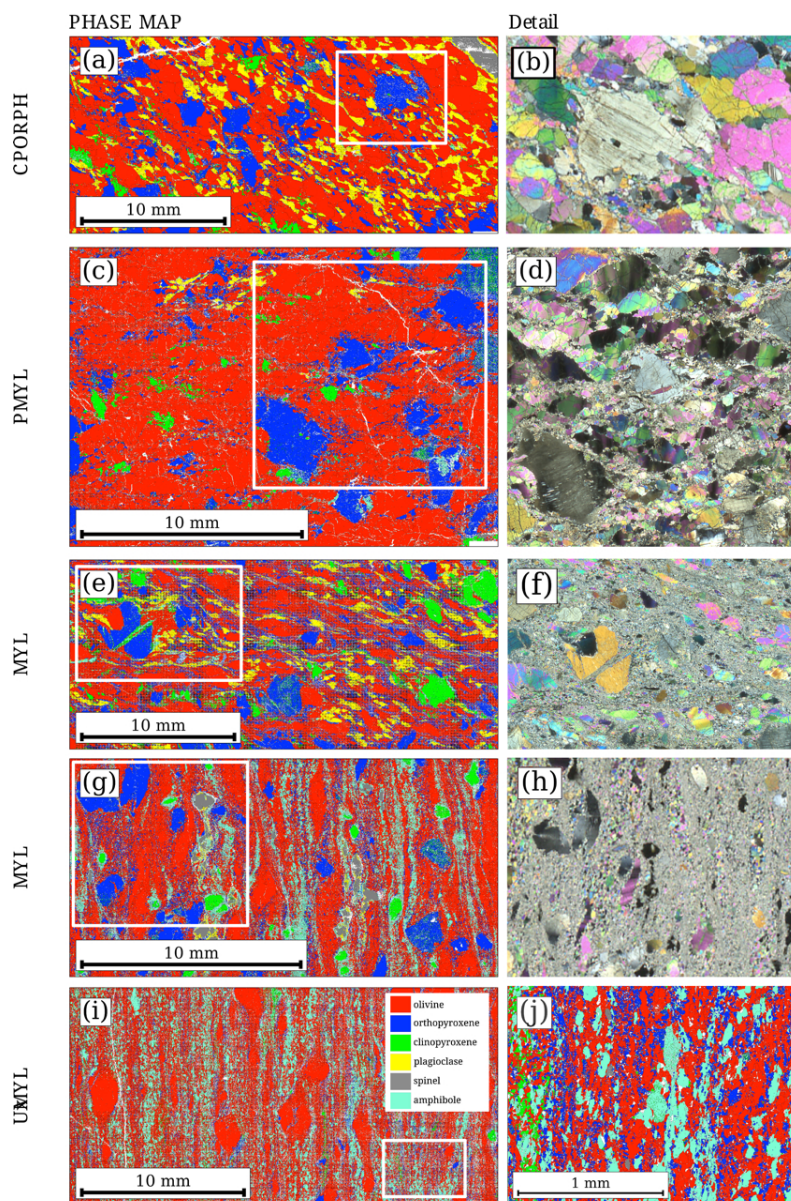


Figure 5. Phase maps derived from EBSD data (a,c,e,g,i) and photomicrographs (b,d,f,h) illustrating typical (a,b) coarse-porphroclastic, (c,d) protomylonitic, (e,f,g,h) mylonitic, and (i, j) ultramylonitic microstructures.

297x451mm (72 x 72 DPI)

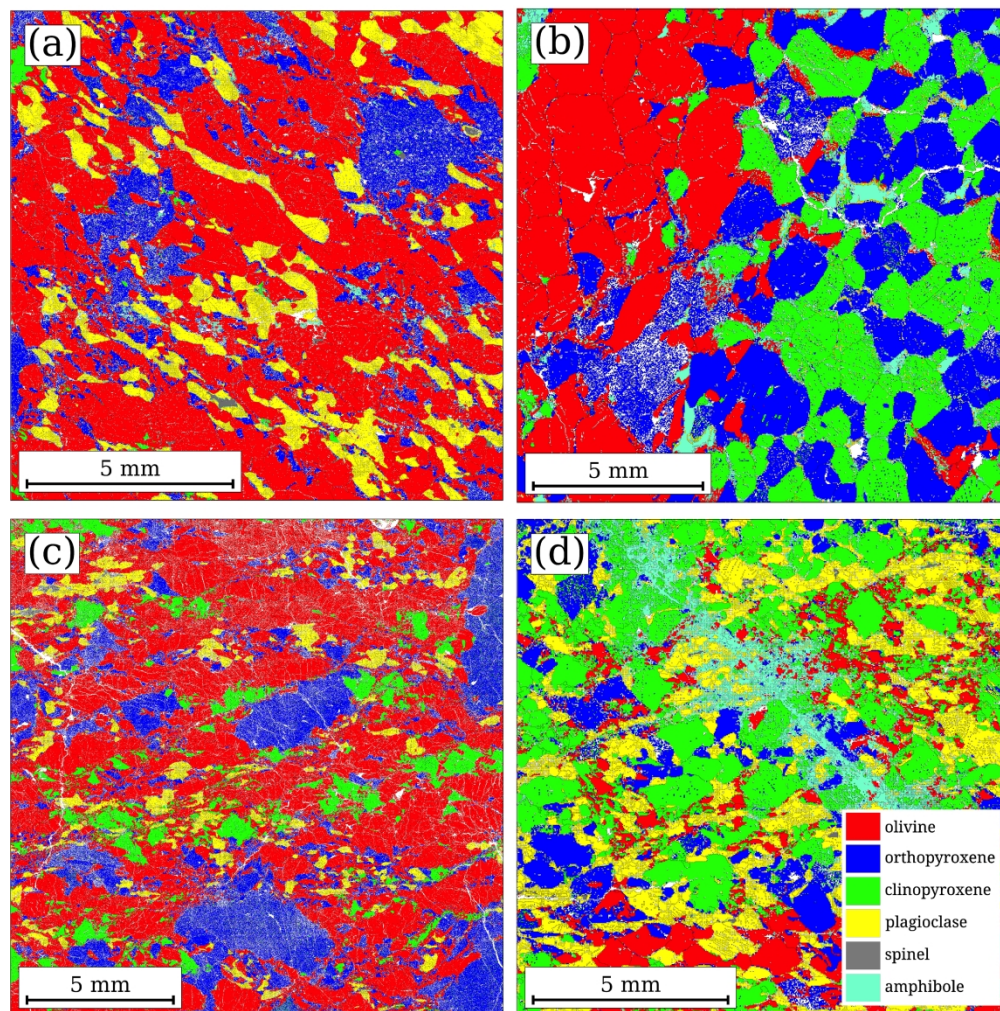


Figure 6. Phase maps derived from EBSD data illustrating details of the (a) coarse-porphyroclastic microstructure in plagioclase-rich peridotite 85ZA1 from the southern massif, (b) cm-scale pyroxenitic layering and coarse-porphyroclastic microstructure spinel-peridotite 86Z30 from the central massif, (c) protomylonitic microstructure in plagioclase and clinopyroxene peridotite 85ZA37 from the southern massif, and (d) sample 85ZA26C from the southern massif with gabbroic to troctolitic composition displaying a coarse-porphyroclastic microstructure crosscut by an amphibole-rich vein.

161x164mm (600 x 600 DPI)

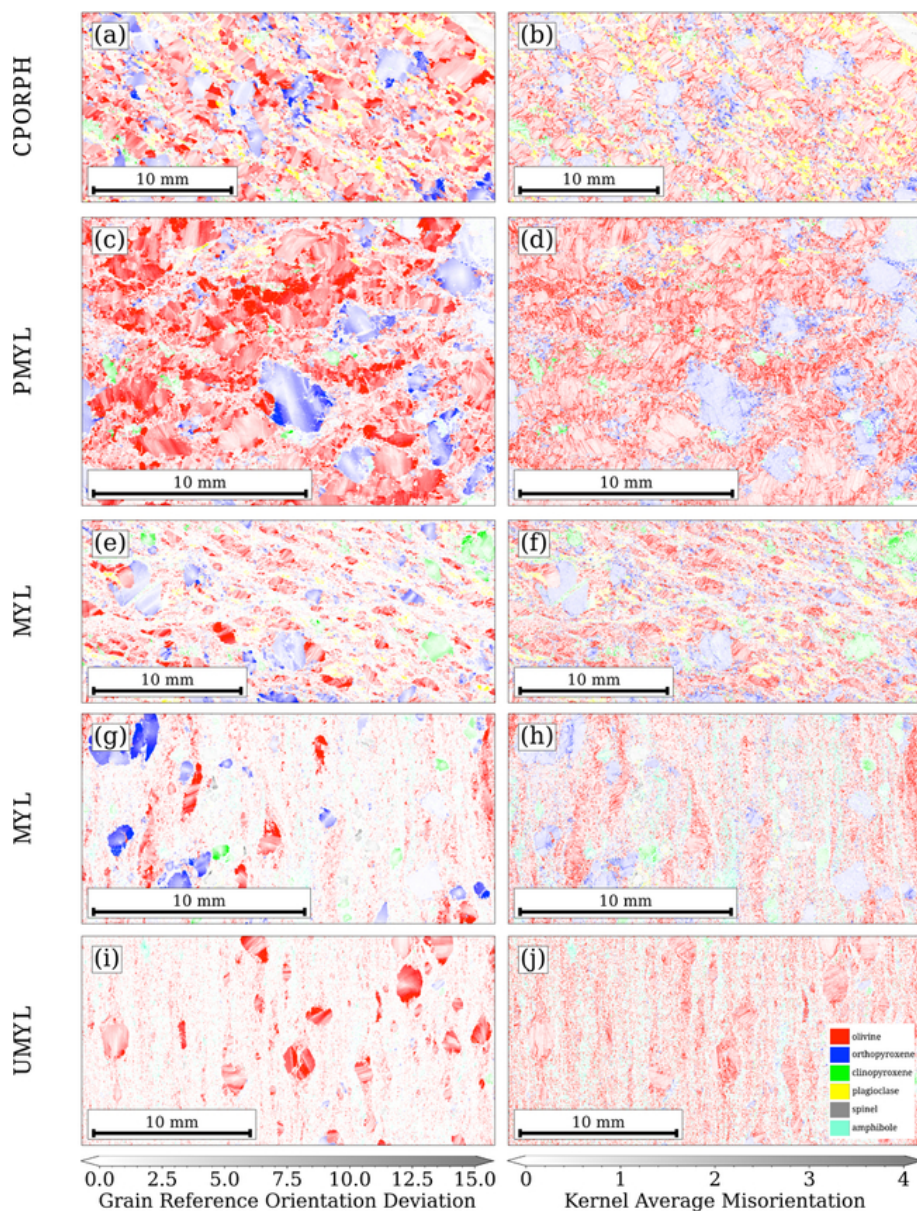


Figure 7. (a,c,e,g,i) Misorientation relative to the mean orientation of the grain and (b,d,f,h,j) Kernel Average Misorientation (KAM) maps depicting the internal deformation of the grains in the different microstructural groups. High KAM values materialize the subgrain boundaries. Tones of red = olivine, blue = orthopyroxene, green = clinopyroxene, yellow = plagioclase, purple = spinel, and cyan = amphibole. Grain boundaries (misorientations > 15°) in black.

27x36mm (600 × 600 DPI)

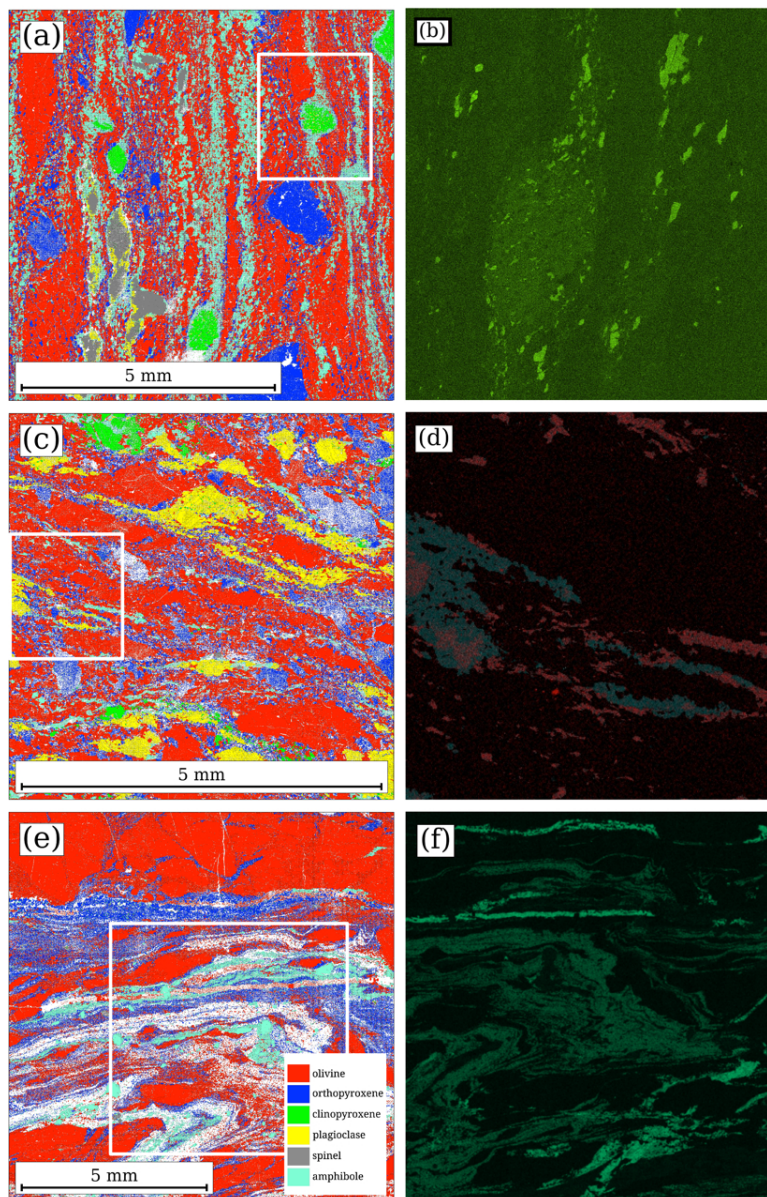


Figure 8: Phase maps derived from EBSD data (a,c,e) and compositional maps derived from EDS data (b,d,f) illustrating microstructurally controlled variations in mineralogic composition in mylonites (a-d) and ultramylonites (e,f). (b) Ti map documenting two generations of amphibole replacing clinopyroxene: coarse grains with high Ti contents surrounded by finer-grained Ti-poor amphiboles, in mylonite 85ZA45 of the northern massif. (d) Ca+Na maps with red to blue tones documenting the increase in anorthite content in plagioclase from the core of porphyroclasts to the recrystallized grains in mylonite 95ZA7b of the southern massif. (e) Cl map in which cyan tones document the occurrence of two generations of scapolite with different Cl-contents in ultramylonite 85ZA71-13 of the southern massif.

293x451mm (72 x 72 DPI)

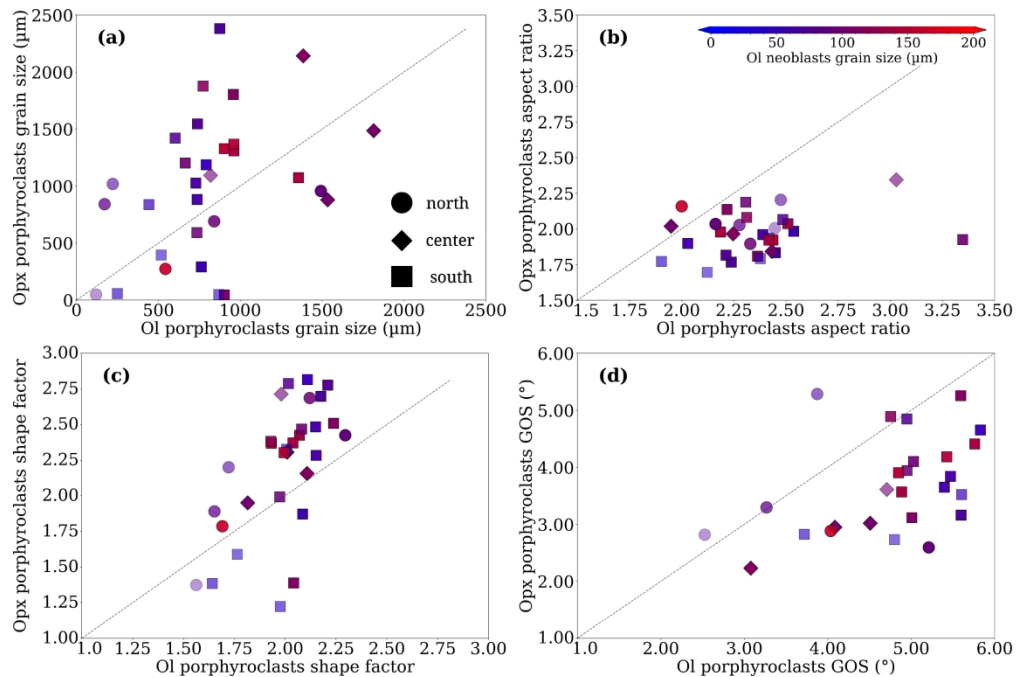


Figure 9. Statistical microstructural data for olivine and orthopyroxene porphyroclasts: (a) grain size, defined by the equivalent grain diameter, (b) elongation, defined by the aspect ratio, (c) shape, defined by the shape factor, and (d) internal deformation, defined by the mean grain orientation spread (GOS). The values displayed are area-weighted averages over the entire EBSD map scale. This averaging was chosen to avoid over-representation of the numerous small grains that occupy in most cases a limited surface of the thin section. Legend as in Figure 1: squares = Southern massif, circles = Central massif; diamonds = Northern massif; redish symbols = coarse-porphyroclastic samples; bluish symbols = LT protomylonites to ultramylonites, with the transparency increasing with increasing neoblasts proportion. The dashed line marks the 1:1 relation.

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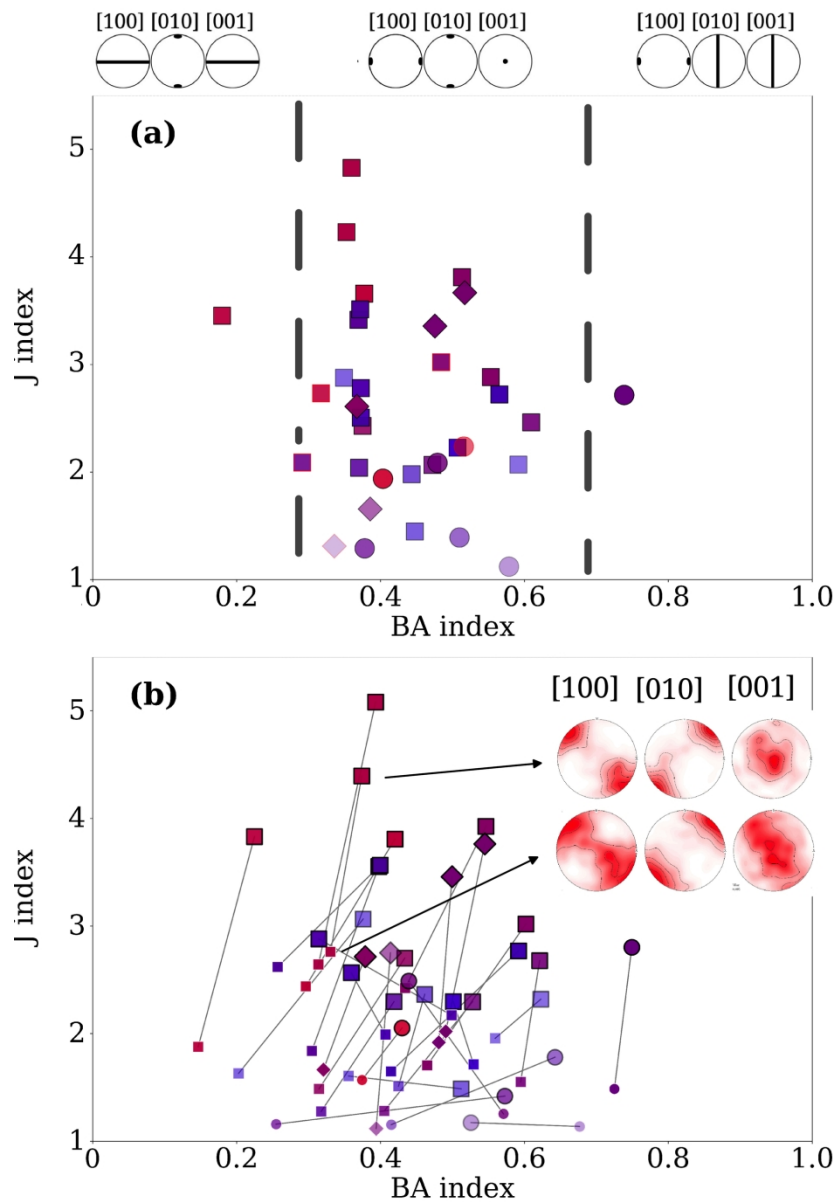


Figure 10. Olivine crystal preferred orientation symmetry (quantified by the BA index) vs intensity (quantified by J index): (a) Average values over the entire olivine population and (b) average values for porphyroclasts (coarse symbols) and neoblasts (small symbols); data for the same sample are linked by a line. Inset in (b) illustrates the difference in crystal preferred orientation of olivine between porphyroclasts and neoblasts (mean orientation of the grain) for a representative sample. Stereographic projections in the structural reference frame; contours at 1 multiple of a uniform distribution intervals.

80x117mm (600 x 600 DPI)

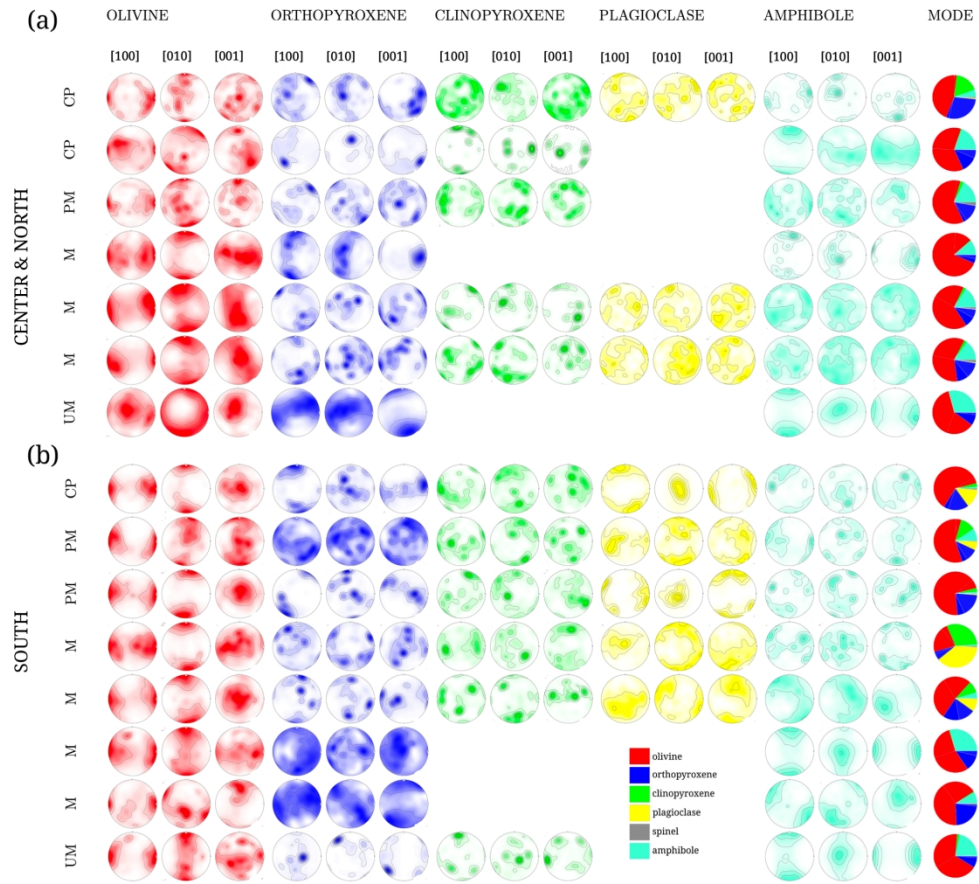


Figure 11. Crystal preferred orientations of the main rock-forming minerals in samples representative of the different deformation facies in the (a) central and northern massifs and (b) southern massif. Contoured lower hemisphere stereographic projections in the structural reference frame (X is the lineation and Z the normal to the foliation) of all measurements in the map. Contours at 1 multiple of a uniform density intervals. Pie diagrams on the right show the modal composition of the samples (color legend as in the phase maps in Figure 6: olivine = red; orthopyroxene = blue; clinopyroxene = green; plagioclase = yellow; amphibole = cyan; spinel = purple) with the hatched portion representing the porphyroclasts' area fraction for each mineral.

811x724mm (72 x 72 DPI)

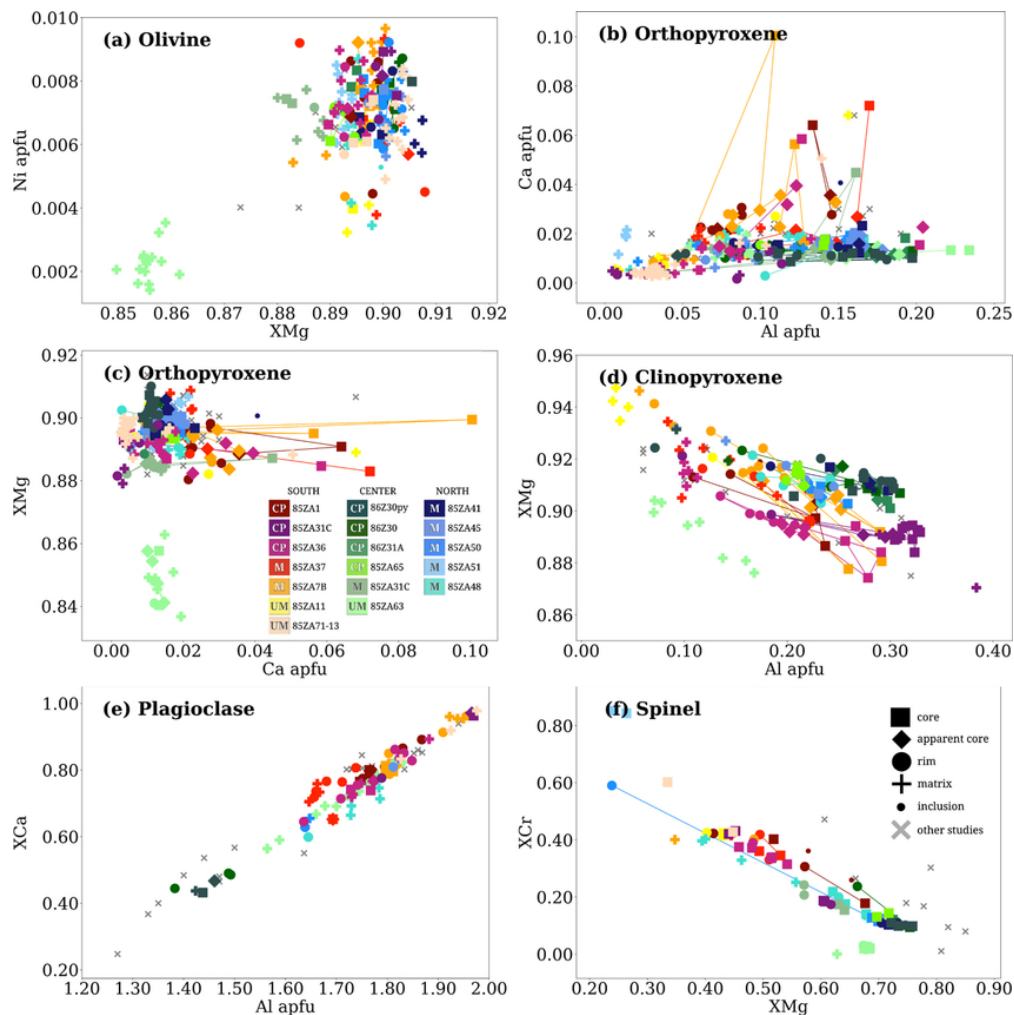


Figure 12. Compositions of olivine, orthopyroxene, clinopyroxene, plagioclase, and spinel in atoms per unit formula. Symbols indicate the type of analysis: core, intermediate, or rims of porphyroclasts, neoblasts within the fine-grained matrix, or inclusion within another mineral phase. Warm colors indicate samples from the southern massif, greens, samples from the central massif and blues, samples from the northern massif. Lighter tones indicate more recrystallized samples.

35x36mm (600 x 600 DPI)

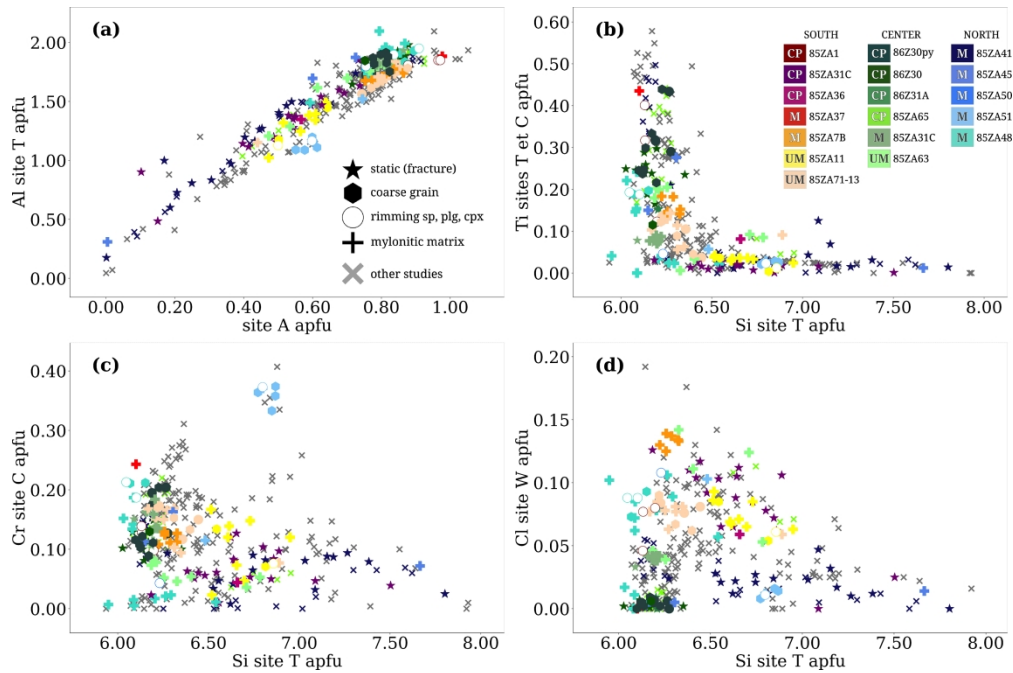


Figure 13. Compositions of amphiboles in atoms per unit formula. Symbols indicate the type of analysis: coarse or fine grains in the mylonitic matrix, rims around clinopyroxene or plagioclase, or crystals formed by static replacement in the vicinity of fractures. Warm colors indicate samples from the southern massif, greens, from the central massif and blues, from the northern massif. Lighter tones indicate more recrystallized samples.

182x121mm (600 x 600 DPI)

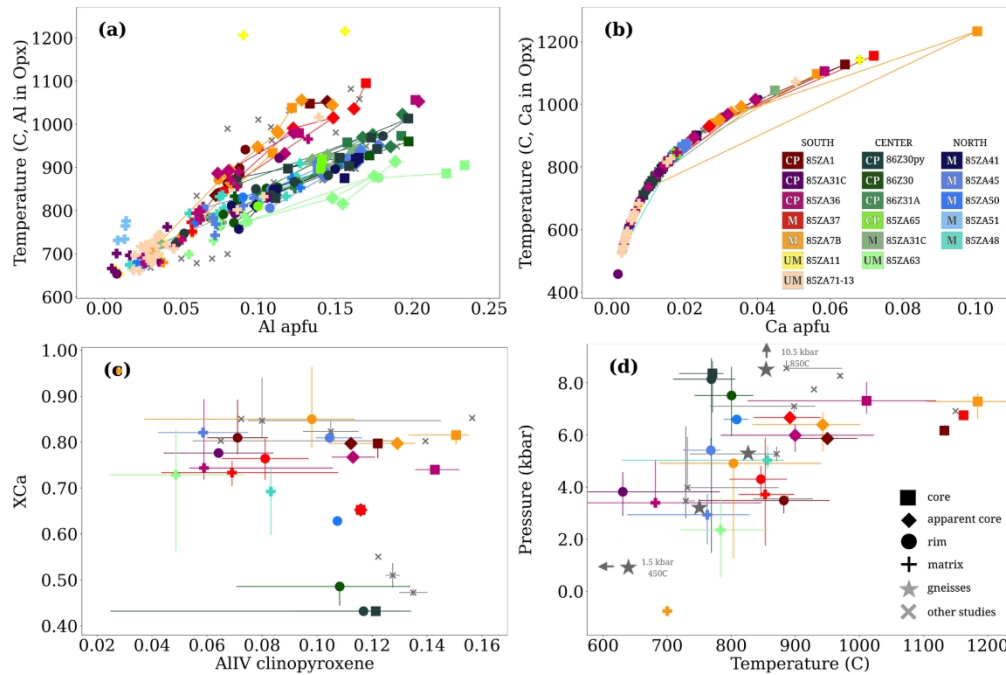


Figure 14. Equilibrium temperature and pressure conditions estimated for the different microstructural assemblages in each sample: cores, intermediate positions, and rims of porphyroclasts or neoblasts from the recrystallized matrix. (a) Al in orthopyroxene (in atoms per formula unit, apfu) vs temperatures estimated using the Al in Opx thermometer, the difference in temperature estimates for similar Al contents between samples from the Central and North massifs and the Southern massif denote the effect of pressure on this thermometer. (b) Ca in orthopyroxene (apfu) vs temperatures estimated using Ca in Opx thermometer. (c) AlIV in clinopyroxene vs XCa in plagioclase; the experimental data of Fumagalli et al. (2017) used to calibrate the FACE barometer is presented for reference. d. Ca in Opx temperatures vs pressures estimated using the FACE barometer for all plagioclase-bearing microstructural assemblages. In (a) and (b) results of individual analyses are presented (errorbars are smaller than the symbols). In (c) and (d) the uncertainty due to spatial variations in the minerals composition within a sample was estimated by considering all possible combinations between cores, intermediate position, rim and matrix analyses for the different minerals in each sample; it is represented by plotting the median with errorbars representing the dispersion of the estimated values for each microstructural class.

181x134mm (300 x 300 DPI)

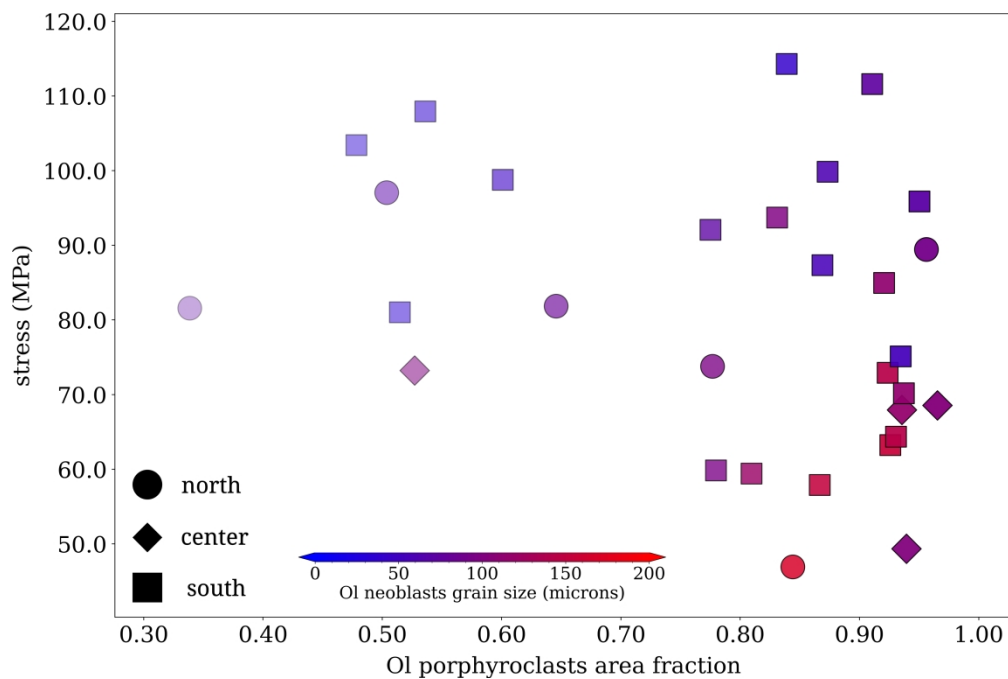


Figure 15. Paleostresses determined based on the mean olivine neoblast equivalent grain size using the piezometer of Van de Wal et al. (1993) for all studied peridotites.

378x251mm (300 x 300 DPI)

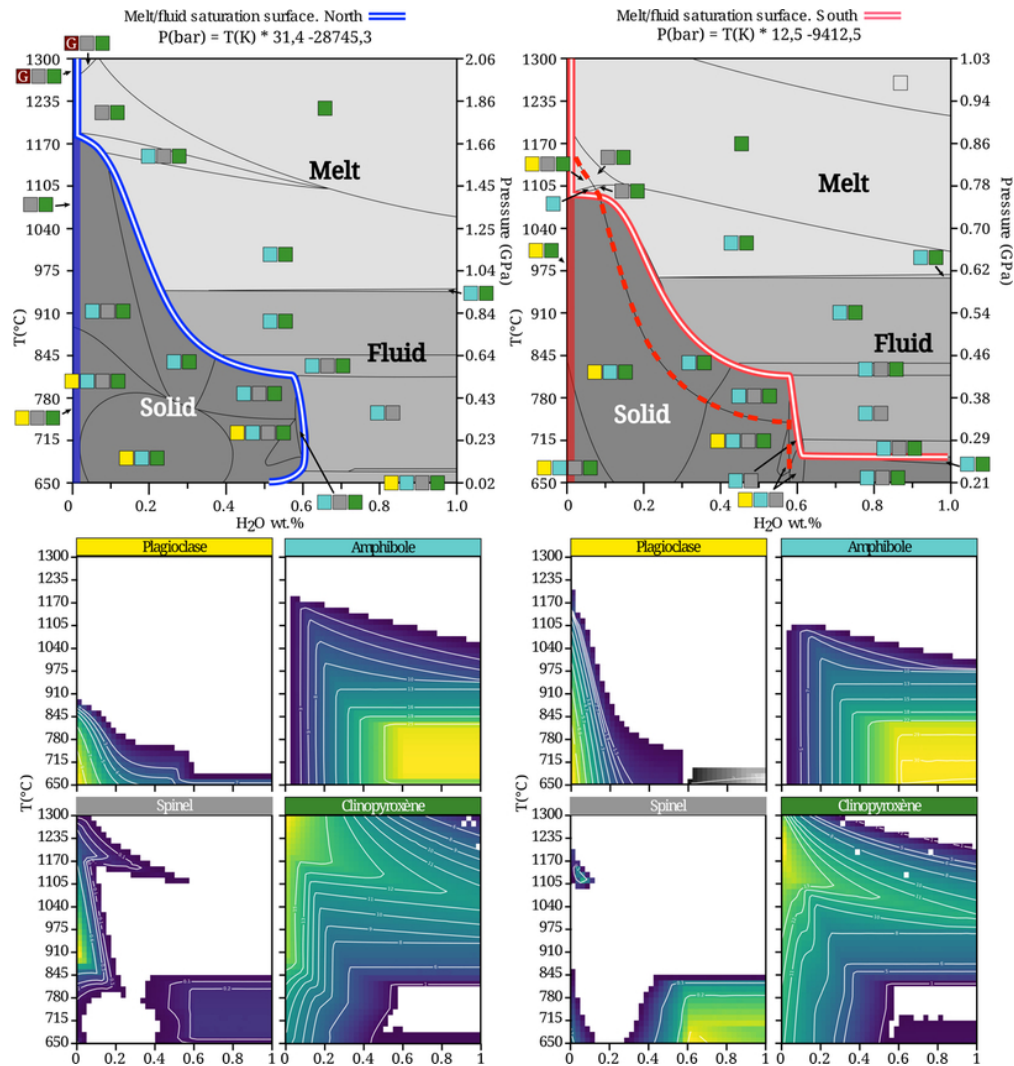


Figure 16. Evolution of the system composition as a function of the water content (XH<sub>2</sub>O wt. %) for the pT paths inferred for the northern/central (left, blue) and southern (right, red) massifs for the bulk rock composition of sample 85ZA36, chosen as representative of the average composition of the massif. The full lines on the pseudosections indicate the paths used for calculating the evolution of mineral compositions during exhumation: the simple line marks an almost anhydrous path (100 ppm H<sub>2</sub>O), the composite line, a path following the melt or fluid saturation surface, and the dashed line, a melt/aqueous fluid undersaturated path along the near-zero mode of plagioclase. Coloured squares represent the stable modal compositions of the solid matrix: Ol and Opx are not represented because they are stable over the entire field, for the other phases colors as in the phase maps: green = Cpx, yellow = Plg, cyan = Amph, gray = Sp. Garnet, which is not observed in the studied peridotites is indicated in brown. The panels below each pseudosection represent the stability field and modal contents of the four Al-bearing phases. Grey contours in the bottom left of the plagioclase content panel for the southern massif document the stability field and modal content of chlorite.

34x36mm (600 x 600 DPI)

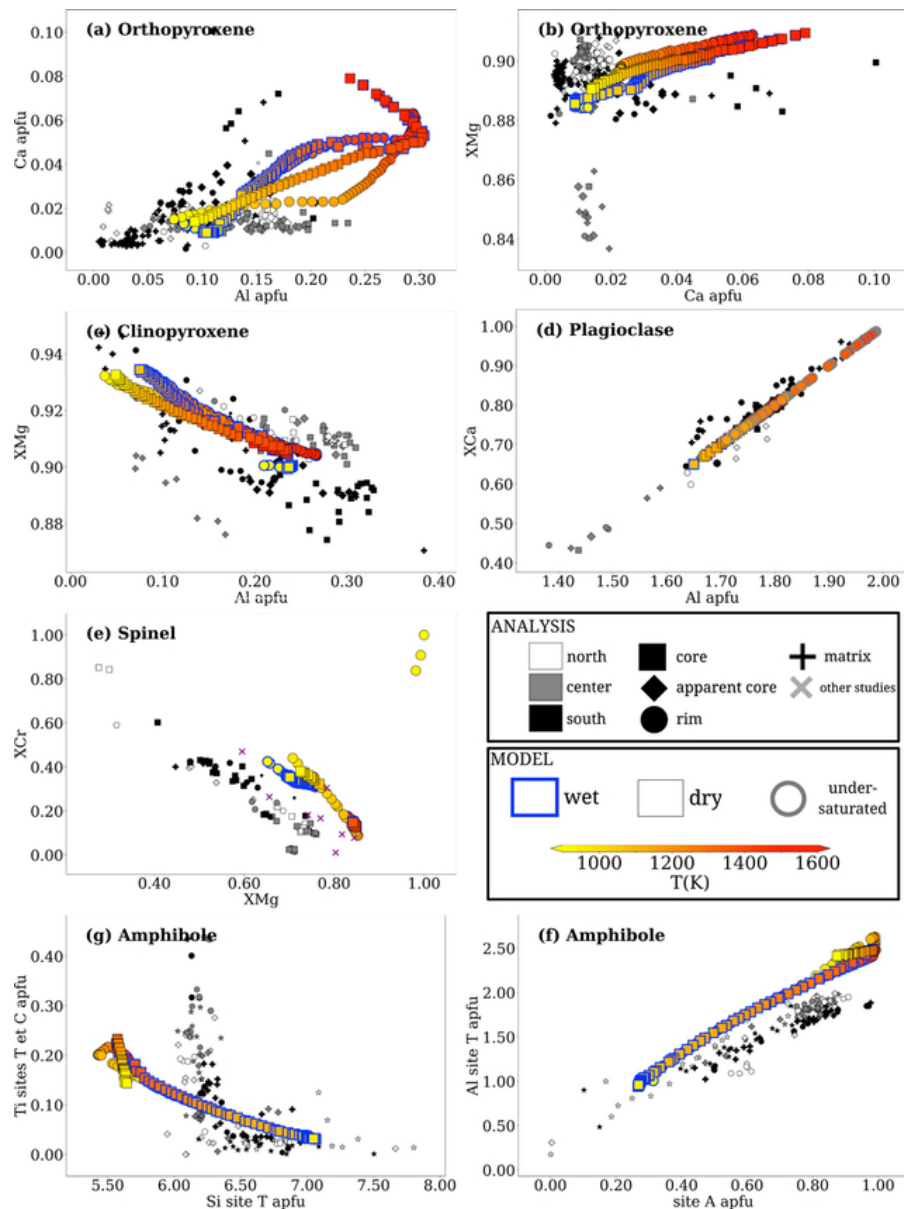


Figure 17. Comparison between the mineral compositions predicted by the thermodynamical calculations for the four pT-hydration paths and those measured in the Zabargad peridotites. For the calculated compositions, the filling of the symbols indicates the temperature and their contours the nature of the path: black= almost anhydrous, blue=melt or fluid-saturated, with squares for the northern/central pT path and circles for the southern pT path. Gray circles in Figure 17d indicate the evolution of XCa along a melt/aqueous fluid undersaturated path along the near-zero mode of plagioclase. For the measured data, symbols as in Figures 11 and 12 with data from the southern massif in black, central massif in gray, and northern massif in white.

27x36mm (600 x 600 DPI)