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1 **Multi-scale development of a stratiform chromite ore body at the base of**  
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4  
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18  
19 **Abstract**

20 A stratiform chromite ore body crops out in the lower part of the dunitic mantle-crust  
21 transition zone (DTZ) that developed at the top of a mantle diapir in the Maqsad area in the  
22 Oman ophiolite. It is made of layers ranging in thickness from a few mm to a maximum of 3  
23 m, and in modal composition from massive to antinodular and disseminated ore. The ore body  
24 is about 50 m thick and its lateral extent does not exceed several hundred meters. The layering  
25 dips gently to the southeast, parallel to that of the overlying gabbroic cumulates.

26 The chromite composition is typical of a MORB kindred - moderate XCr ( $100 \times$   
27 Cr/(Cr + Al) atomic ratio), ranging from 48 to 60, and relatively high TiO<sub>2</sub> content, ranging  
28 from about 0.3 to 0.5 wt% -, a characteristic shared by most lithologies issued from the  
29 igneous activity of the Maqsad diapir. The silicate matrix is essentially made of slightly  
30 serpentinized olivine with minor clinopyroxene and rare pargasitic amphibole, orthopyroxene  
31 and garnet. This strongly contrasts with the nature of the mineral inclusions mostly made of  
32 the assemblage amphibole-orthopyroxene-mica, enclosed in the chromite grains and  
33 represented in abundance all along the ore body whatever the ore grade. The inclusions  
34 demonstrate the involvement of a silica- and water-rich melt and/or fluid, in addition to  
35 MORB, in the early stages of chromite crystallization.

36 The chemical composition of chromite, silicate matrix, together with the one of  
37 silicate inclusions display well-defined evolutions vertically along the stratiform chromitite.  
38 At the scale of the ore body, the compositional trends are independent of the ore concentration  
39 but the major kinks in these trends are well-correlated with levels of magmatic breccias. This  
40 shows that abrupt chemical changes can be attributed to sudden melt +/- fluids injection  
41 events followed mainly by melt-fluid-rock interaction and in a lesser extent by quieter  
42 evolution by fractional crystallization. At the thin section scale, second order chemical  
43 variations, essentially in the Mg# ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio) of chromite and Fo of  
44 olivine, are clearly attributable to re-equilibration between these two solid phases, possibly in  
45 the presence of an interstitial melt/fluid.

46

#### 47 **Keywords**

48 Oman ophiolite; chromite; stratiform chromitite; melt hybridization; dunitic mantle-crust  
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50

**51 Highlights**

- 52 - The Maqсад dunitic transition zone (Oman ophiolite) hosts a stratiform chromitite
- 53 - The matrix (ol, cpx) strongly contrasts with silicate inclusions (opx, amph, mica)
- 54 - Vertical chemical evolutions along the ore body correlate to magmatic breccias
- 55 - Melt injections (breccias) controlled the hybridization between contrasted melts

56

**57 1. Introduction**

58 The fate of chromium from its mantle source and its concentration by the  
59 crystallization and accumulation of chromite in cooling magma bodies is a long-standing  
60 poorly understood process of the magmatic history of mafic and ultramafic rocks (e.g.  
61 Bowen, 1928; Lombaard et al., 1956; Stoll, 1958), and is still the matter of intense  
62 discussions (cf. Arai and Miura, 2016 and references therein). Different mechanisms were  
63 proposed to account for the formation of massive chromitites in crustal magma chambers,  
64 following the crystallization from a chromite-oversaturated basaltic melt in which chromite  
65 was the only liquidus phase and by the chromite crystals settling and concentration in layers  
66 of variable thickness. Among others processes, the mixing between a basalt saturated in  
67 chromite and a more silicic melt, contamination of the basalt by the surrounding felsic crustal  
68 rocks, addition of H<sub>2</sub>O, increase in the total pressure or fO<sub>2</sub> of the parent magma, saturation in  
69 chromite alone in the parent magma following a decrease in the lithostatic pressure (e.g.  
70 Irvine, 1975, 1977a, 1977b; Latypov et al., 2018; Lipin, 1993; Matveev and Ballhaus, 2002;  
71 Mondal and Mathez, 2007; Naldrett et al., 2012; Spandler et al., 2005) were proposed. The  
72 most popular model consists in the massive chromite crystallization from a melt hybrid  
73 between a relatively primitive basalt and a silica-rich melt, issued from the country rock re-  
74 melting (Irvine, 1975) or from the same melt at a more evolved, differentiated stage (Irvine,  
75 1977a). Some authors also proposed the involvement of volatile- and sodium-rich melts or

76 fluids based on the nature of hydrous, Na-rich silicate inclusions in chromite grains (e.g.  
77 Borisova et al., 2012; Johan et al., 1983, 2017; McElduff and Stumpfl, 1991; Spandler et al.,  
78 2005; Talkington et al., 1984), while in extreme cases chromite may be entirely attributed to  
79 hydrothermal processes (Arai and Akizawa, 2014).

80 Chromitite ore bodies were mainly reported in continental layered intrusions and in  
81 ophiolites worldwide, whatever their tectonic setting (i.e. supra-subduction zone or mid-  
82 oceanic ridge). Recent sampling in present-day oceans documented cm-thick chromitites  
83 associated with abyssal peridotites (Abe, 2011; Arai and Matsukage, 1998; Payot et al.,  
84 2014). Silicate inclusions in their chromite grains (chromitites and chromite grains scattered  
85 in olivine-rich troctolites) are similar to the ones observed in layered intrusions and ophiolites  
86 (Matsukage and Arai, 1998; Tamura et al., 2014). This demonstrates that the formation of  
87 chromitite may occur in various magmatic environments involving a wide spectrum of parent  
88 melts and a silica-, alkali- and water-rich melt or fluid not necessarily issued from the  
89 subduction factory. Accordingly, the topic of the geochemical cycle of Cr has to be addressed  
90 by the study of the local processes involved in the formation of a given ore body (surrounding  
91 rocks, nature and proportion of the melts/fluids, physical settings, etc.), rather than through  
92 the tectonic setting in which the chromitite formed.

93 Stratiform chromitite ore bodies are among the most impressive expressions of Cr  
94 mobilization and fractionation, chromite-seams alternating with harzburgite, dunite, gabbro,  
95 orthopyroxenite, norite, anorthosite among other (e.g. Jackson, 1961; Mathez and Kinzler,  
96 2017; Thayer 1960). They were made famous by their occurrence in large (up to few tens to  
97 few hundreds km) and thick (up to few km), layered igneous complexes that intruded  
98 continental domains through massive magmatic events in old Archean to Proterozoic  
99 geological times (e.g. Muscox, Stillwater, Bushveld and Great Dyke among the most studied  
100 complexes). Stratiform chromitites of much smaller size were described in ophiolitic

101 sequences and share many features with chromite-seams in layered intrusions (e.g. texture,  
102 lithological associations, silicate inclusions) (e.g. Borisova et al., 2012; Gauthier et al., 1990;  
103 González-Jiménez et al., 2014; Lorand and Ceuleneer, 1989; Rahgoshay et al., 1981), such as  
104 in Oman where they are restricted to the dunitic mantle-crust transition zone (DTZ) (Augé,  
105 1987; Ceuleneer and Nicolas, 1985), which contrasts with the podiform chromite ore bodies  
106 more common in the underlying harzburgitic mantle section. They preserved their igneous  
107 structure when they escaped post-crystallization deformation resulting from off-ridge-axis  
108 transposition, associated to diverging mantle solid-state flow (Cassard et al., 1981; Ceuleneer  
109 and Nicolas, 1985).

110         The formation of chromitite in the upper mantle beneath oceanic ridges is thought to  
111 be closely related to the formation of dunite. As a matter of fact, a reactional dunitic envelope  
112 underlines, as a rule, the contact between chromitite and surrounding mantle harzburgites  
113 while abundant chromitites also occur within the DTZ (e.g. Arai and Yurimoto, 1994;  
114 Cassard et al., 1981; González-Jiménez et al. 2014; Lago et al., 1982). The chromite grains  
115 scattered in the DTZ share the same abundant mineral inclusions than the chromite grains  
116 from chromitites (e.g. Borisova et al., 2012; Rospabé et al., 2017). Consequently, both dunite  
117 and chromitite appear to be, in this context, two associated products formed at the expense of  
118 harzburgite following melt- or magma-peridotite reactions (Arai and Yurimoto, 1994;  
119 Borisova et al., 2012; Zhou et al., 1994).

120         In order to better understand the formation of stratiform chromitites beneath oceanic  
121 spreading centres, we performed a detailed sampling of an undeformed stratiform chromitite  
122 ore body cropping out within the Maqsad DTZ in Oman (Ceuleneer and Nicolas, 1985). Our  
123 results highlight that the processes that governed chromite composition are different  
124 depending on the scale: chemical re-equilibration at local (thin section) scale, and episodic  
125 magmatic influxes at a larger (ore body) scale.

126

## 127 **2. The Maqsad area of the Oman ophiolite**

128 The Oman ophiolite is a fragment of the Cretaceous oceanic lithosphere accreted in  
129 the Tethys ocean about 95-97 Ma ago (Tippit et al., 1981), detached during an intra-oceanic  
130 thrusting event that occurred contemporaneously, soon after or soon before the accretion  
131 (Montigny et al., 1988; Rioux et al., 2016, 2013; Warren et al., 2005) and obducted onto the  
132 Arabian margin during Maestrichtian times (circa 70 Ma) (Coleman, 1981; Glennie et al.,  
133 1973). The precise context of genesis and emplacement of the Oman ophiolite is still debated:  
134 a “pure” mid-ocean ridge setting according to some authors (e.g. Coleman, 1981), a  
135 subduction zone setting according to others (e.g. Pearce et al., 1981). The spatial distribution,  
136 along the ophiolite, of the nature of mafic dikes cross-cutting the harzburgitic mantle section  
137 and of lower crustal cumulates shows that both MORB and depleted calc-alkaline series  
138 coexisted at the time of the igneous accretion of the Oman ophiolite (Benoit et al., 1999;  
139 Clénet et al., 2010; Python and Ceuleneer, 2003).

140 The Maqsad area is located in the Sumail massif, in the southeastern part of the Oman  
141 ophiolite (Fig. 1a). There, a former axial mantle upwelling (“diapir”) has been frozen and has  
142 been mapped through the study of high-temperature plastic deformation structures in mantle  
143 harzburgites (Ceuleneer, 1991; Ceuleneer et al., 1988; Rabinowicz et al., 1987). The Maqsad  
144 diapir is less than 10 km large in cross-section but constitutes the central part of a much larger  
145 structure: a former spreading segment, about 80 km in length parallel to the paleo-spreading  
146 axis (NW-SE) (Python and Ceuleneer, 2003). The segment itself is characterized by mantle  
147 dykes (former channels used by partial melts to reach the crust) filled with cumulates of  
148 MORB affinity (Benoit et al., 1996; Ceuleneer et al., 1996; Python and Ceuleneer, 2003;  
149 Python et al., 2008). This contrasts markedly with the more common situation observed  
150 elsewhere in Oman where mantle dykes are essentially filled with crystallization products

151 from the depleted andesitic melts (Python and Ceuleneer, 2003). These andesites can be  
152 generated either by the hydrous re-melting of the oceanic lithosphere in a mid-ocean ridge  
153 setting (Benoit et al., 1999; Nonnotte et al., 2005), or in a subduction environment (see  
154 discussion in Python et al., 2008 and in Python and Ceuleneer, 2003).

155 The dunitic mantle-crust transition zone (DTZ) is particularly well-developed in the  
156 Maqsad area, reaching more than 300 m in thickness above the central part of the diapir  
157 (Abily and Ceuleneer, 2013; Boudier and Nicolas, 1995; Ceuleneer and Nicolas, 1985).  
158 Dunite from the Oman ophiolite DTZ is mainly interpreted as the product of melt-peridotite  
159 interaction (e.g. Boudier and Nicolas, 1995; Koga et al., 2001) while its upper part  
160 (shallowest 50 m) may be cumulative in origin after olivine fractionation from highly  
161 magnesian melts (Abily and Ceuleneer, 2013).

162 The DTZ was a very dynamic environment formed by several processes frozen at  
163 various stages of their development, where the current extensive petrological and geochemical  
164 variability is the footprints of dunitization (i.e. incongruent dissolution of the mantle  
165 harzburgites' orthopyroxene), melt percolation associated with chemical re-equilibration with  
166 the surrounding olivine matrix, refertilization by variably evolved melt batches, compaction,  
167 syn-magmatic faulting, later low temperature alteration, etc. (Rospabé et al., 2018, 2019).  
168 Among the crystallization products in the DTZ dunites, orthopyroxene and amphibole  
169 interstitial to olivine grains have been interpreted to indicate that a hydrous component was  
170 involved in the formation of dunites (Rospabé et al., 2017). We proposed that the mantle-crust  
171 transition zone beneath oceanic spreading centres is the place of hybridization between the  
172 uprising (relatively dry) MORB issued from the mantle partial melting (i.e. recorded by the  
173 troctolite and ol-gabbro cumulates in the mantle section - Benoit et al., 1996; Ceuleneer et al.,  
174 1996; Python et al., 2003), and a hydrated component coming from above, probably seawater-  
175 derived. The volatile-, Si-, Na- and trace elements-rich hybrid lithospheric melt formed during

176 or/and contributed to the harzburgite dunitization process (Rospabé et al., 2017, 2018, 2019).  
177 The presence of orthopyroxene and amphibole, together with mica, enclosed in chromite  
178 grains strongly suggests that such a hybrid melt was involved early in both the formation of  
179 dunite and the re-mobilization of chromium at the mantle-crust transition zone (Rospabé et  
180 al., 2017; see also Borisova et al., 2012).

181 Chromitite ore bodies are particularly abundant in the Maqsad area, probably resulting  
182 from the intense magmatic activity associated to the rise of a hot diapir up to the base of the  
183 crust (Ceuleneer and Nicolas, 1985; Ceuleneer et al., 1996). Those cropping out at the top of  
184 this structure, in the DTZ, were not deformed as they were not transposed by mantle solid-  
185 state flow. This offers a unique opportunity to study the details of magmatic processes related  
186 to chromite ore genesis (e.g. Borisova et al., 2012; Leblanc and Ceuleneer, 1991; Rollinson  
187 and Adetunji, 2013; Zagrtednov et al., 2018).

188

### 189 **3. Description of the studied stratiform chromitite**

#### 190 *3.1. Structuration of the stratiform chromitite*

191 The studied chromitite ore body is located a few tens of metres above the base of the  
192 DTZ near the centre of the Maqsad diapir (Fig. 1b; Ceuleneer and Nicolas, 1985). We  
193 collected 62 samples (labelled 06OM04-##) along a vertical section (Fig. 2), before the  
194 mining operations that started in 2008 (Fig. 3a-b). The stratiform ore body is made of  
195 alternating levels of chromite-bearing dunites and chromitite. It reaches about 50 m in  
196 exposed thickness (Fig. 2). The layering is near horizontal, slightly dipping to the SE and  
197 strictly parallel to the layering of overlying gabbroic cumulates. This supports the view that  
198 the stratiform chromitite formed during the same tectonic/igneous episode as the one that built  
199 the overlying crust, and that it preserved its primary structure. The lateral extent of the layers  
200 cannot be determined precisely due to erosion, but it does not exceed several hundred metres

201 as revealed by our general survey of this area (Rospabé, 2018; Rospabé et al., 2018). The  
202 studied samples preserve delicate igneous textures and display no evidence for post-  
203 crystallization deformation (Fig. 3), which is consistent with the fact that the DTZ in the  
204 Maqsad area was frozen at the axis of a spreading centre and not transposed by diverging  
205 mantle flow (Ceuleneer et al., 1988; Ceuleneer, 1991).

206 The vertical organization of the stratiform chromitite is characterized by 5 main  
207 chromite-rich layers (labelled I to V in Fig. 2), while levels in between are made of more  
208 scattered ore within dunite (Figs. 3c-e and 4a-d). We basically observed the global increase of  
209 the chromite content upsection, from rich-levels I to V. Massive ore occurs in layers ranging  
210 in thickness from a few cm up to 3 m (Fig. 3c) in the upper rich-layers III to V. In between  
211 massive layers (and along layers I and II), the texture of the ore is antinodular. The nodules,  
212 typically about 5 mm in size, are mostly made of subhedral single olivine crystals, more  
213 exceptionally of aggregates of smaller olivine grains embedded in interstitial chromites. The  
214 relative modal abundance of chromite and of olivine presents marked variations, gradual at  
215 some places defining graded bedding, sharper at other places (Figs. 3d-f and 4c-e). Thin  
216 sections observations reveal the absence of solid-state deformation (Fig. 4), and the  
217 antinodular texture of chromite is reminiscent of the poikilitic plagioclase and pyroxene  
218 encompassing olivine grains in DTZ dunites (e.g. Koga et al., 2001; Abily and Ceuleneer,  
219 2013; Rospabé et al., 2017), although the antinodular texture consists of aggregates of tiny  
220 chromite grains surrounding olivine crystals and never form large monocrystals as plagioclase  
221 and pyroxenes. No sintering texture was observed in chromite grains.

222 Magmatic breccias made of cm-sized dunite and wehrlite fragments embedded in  
223 chromite are rather common (Figs. 3g-i and 4f), pointing to a quite dynamic environment  
224 during the stratiform chromitite formation. Six main brecciated levels, a few cm to a few dm  
225 in thickness, were identified along the cross-section: one near its base (1020 m), two in

226 between altitudes 1030 and 1034 m (intermediate breccias system), then three in between  
227 1046 and 1052 m (higher breccias system) (Fig. 2). They show no systematic correlation with  
228 the abundance of chromite in surrounding rocks.

229 Olivine is the dominant silicate phase constituting the matrix between chromite grains  
230 (Figs. 2 and 4g). The serpentinization degree of olivine in this area, and in the surrounding  
231 DTZ, is unusually low compared to many other occurrences of chromite and dunite in the  
232 mantle-crust transition zone where the degree of serpentinization reaches a higher degree. Our  
233 observations of thin section revealed that it doesn't exceed 30%. Clinopyroxene is likewise  
234 regularly observed along the entire cross-section (Fig. 2), appearing as cm-sized poikilitic  
235 crystals both associated with and hosting chromite grains in disseminated, antinodular or  
236 massive ores (Fig. 4h), and impregnating the olivine matrix in surrounding dunitic levels (Fig.  
237 4f). Interstitial orthopyroxene occurs in only two samples (06OM04-12D2, 1041 m, and -  
238 19(2), 1049.5 m). Plagioclase is typically absent in the stratiform chromitite, although  
239 common as an interstitial phase between olivine grains in the surrounding DTZ dunites (e.g.  
240 Abily and Ceuleneer, 2013; Boudier and Nicolas, 1995; Koga et al., 2001; Rospabé et al.,  
241 2017, 2018). Among other minor phases, diopside (hybrid between magmatic and  
242 hydrothermal clinopyroxene, see Rospabé et al., 2017) is observed in the half lower part of  
243 the stratiform chromitite while amphibole and garnet, appearing together above the two  
244 highest magmatic breccias at the altitude 1050 m, are restricted to the upper part of the ore  
245 body (Fig. 2).

246

### 247 *3.2. Distribution of inclusions in chromite*

248 Chromites contain silicate phases in inclusion. Figure 3 represents their relative  
249 abundance in each sample, from 0 to 8 following thin sections observation, as a function of  
250 the altitude. Two levels show a particularly high abundance of inclusions, corresponding to

251 levels affected by magmatic breccias above 1030 m (intermediate breccias system) and  
252 around 1050 m (higher breccias system). In the second order, the abundance of inclusions is  
253 related to the modal content of chromite since at a given level, despite a quite lower amount  
254 of inclusion far away from breccias (for example at the altitude 1040 m), it appears that  
255 massive ore contains more inclusions in chromite than disseminated ore. It is worth  
256 mentioning that no sample got a “score” of 0, meaning that inclusions were observed in all the  
257 62 studied samples, including in dunites that contain only a few percent of disseminated  
258 chromite.

259         The inclusions have a size ranging from a few microns to more than 100  $\mu\text{m}$  and are  
260 generally round-shaped (Fig. 5), more rarely square-shaped (e.g. Fig. 5c and g). Their habitus  
261 and distribution within a chromite grain are a quite large inclusion at the centre of the grain  
262 and/or a corona of several inclusions that suggests they were trapped at the time the chromite  
263 grain growth (Fig. 5a, b and f).

264         The nature of silicate minerals in the inclusions have been determined using the  
265 electron probe. The resulting picture is in strong contrast with the silicate matrix.  
266 Orthopyroxene, amphibole and mica are by far the most common phases of the inclusions.  
267 They are observed all across the ore body independently from the chromite modal content of  
268 the host samples and from the location of magmatic breccias (Fig. 2). Orthopyroxene and  
269 mica occur as monomineralic inclusions as well as polymineralic assemblages associated,  
270 separately or together, with amphibole (Fig. 5c). When occurring as monomineralic  
271 inclusions, the host chromite also generally contains other inclusions filled with amphibole,  
272 making this latter phase the most abundant silicate inclusion type. Clinopyroxene (as well as  
273 diopside) appears only rarely in inclusion, identified in three samples only while it is an  
274 important major component of the matrix. They likewise occur as single inclusions or  
275 associated with amphibole. Garnet and nepheline were also identified as very minor

276 component inclusions. The two nepheline are parts of polymineralic inclusions, occurring in  
277 two different chromites from the same sample (06OM04-3A) and associated with amphibole  
278 as well as with mica in one case (Fig. 5d). Garnet occurs as monomineralic inclusions (Fig.  
279 5e). One inclusion (06OM04-17) displays a composition very rich in silica, aluminium and  
280 sodium (see paragraph 4.5.5.) and fits with a mixing between albite and quartz (Fig. 5g).  
281 High-Cr chromite inclusions or coronas are occasionally observed. Coronas are found at the  
282 contact between silicate single inclusions or assemblages made of amphibole and mica and  
283 the host chromite grain. These chromite inclusions and coronas are observed all along the  
284 stratiform chromitite and their presence is not related to the chromite modal content of the  
285 host sample, although never observed in massive ores, or to the location of magmatic  
286 breccias.

287

#### 288 **4. Mineral compositions**

##### 289 *4.1. Analytical note*

290 Major and minor element concentrations in chromite and silicate phases were  
291 determined by *in situ* electron probe microanalysis (EPMA) using a Cameca SX50  
292 microprobe with SAMx automation (Géosciences Environnement Toulouse, Université de  
293 Toulouse III, Toulouse, France) and a Cameca SXFive FE microprobe (Centre de  
294 MicroCaractérisation Raimond Castaing, Toulouse, France). The operating conditions using  
295 the Cameca SX50 were an accelerating voltage of 15 kV and beam current of 10 nA for  
296 silicate mineral phases and beam current of 20 nA for chromites with a spot size of 4  $\mu\text{m}$  in  
297 diameter. The operating conditions when using the Cameca SXFive FE were an accelerating  
298 voltage of 20 kV, a beam current of 20 nA and a spot size of 1  $\mu\text{m}$  in diameter for all  
299 minerals. The analysis counting time was of 10 s on peak for each element and 5 s on  
300 backgrounds on both sides of the peak for each type of minerals (chromite,

301 hydrous/anhydrous silicates). The two datasets obtained following the using of the two  
302 different microprobes were compared; no shift was observed. The following synthetic and  
303 natural minerals standards were used: albite (Na), periclase (Mg), corundum (Al), sanidine  
304 (K), wollastonite (Si, Ca), pyrophanite (Mn, Ti), hematite (Fe), chromium oxide (Cr), nickel  
305 oxide (Ni), sphalerite (Zn) and pure vanadium (V). Oxides concentrations were obtained by  
306 correcting nominal concentrations following the PAP data reduction method (Pouchou and  
307 Pichoir, 1985). The detection limits are 0.03 wt% for TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, Cr<sub>2</sub>O<sub>3</sub> and MgO, 0.04 wt%  
308 for SiO<sub>2</sub> and Na<sub>2</sub>O, 0.05 wt% for CaO and K<sub>2</sub>O, 0.06 wt% for V<sub>2</sub>O<sub>3</sub> and NiO, and 0.07% for  
309 FeO, MnO and ZnO, for all mineral phases analysed. The internal precision (%RSD) is better  
310 than 2% for major elements (e.g. Cr<sub>2</sub>O<sub>3</sub> in chromite, MgO in olivine, CaO in clinopyroxene),  
311 better than 6% for TiO<sub>2</sub> (chromite, clinopyroxene) and better than 30% for minor elements  
312 (e.g. NiO in olivine, Na<sub>2</sub>O in clinopyroxene). A few selected data are available in Table 1.  
313 The whole dataset is provided in Supplementary Material.

314

#### 315 4.2. Chromite

316 Chromites from the stratiform chromitite are Cr- and Al-rich relative to Fe<sup>3+</sup> (see  
317 ternary diagram in Figure 6). They display variable chemical compositions, with averaged per  
318 sample XCr (100 × Cr/(Cr + Al) atomic ratio) = 49.0-59.9 (Cr<sub>2</sub>O<sub>3</sub> = 39.4-47.1 wt%; Al<sub>2</sub>O<sub>3</sub> =  
319 20.7-27.6 wt%), XMg (100 × Mg/(Mg + Fe<sup>2+</sup>) atomic ratio) = 50.0-69.8, YFe<sup>3+</sup> (100 ×  
320 Fe<sup>3+</sup>/(Cr + Al + Fe<sup>3+</sup>) atomic ratio) = 1.1-7.8 (4.6 in average), and TiO<sub>2</sub> = 0.25-0.54 wt%  
321 (averaged values for each samples) (Fig. 6). Chromite inclusions and coronas have a higher  
322 XCr, ranging from 58.3 to 71.5 (Cr<sub>2</sub>O<sub>3</sub> = 43.7-52.7 wt%; Al<sub>2</sub>O<sub>3</sub> = 14.1-21.3 wt%), a similar to  
323 slightly lower XMg (50.5-63.0), and a less variable but globally higher in average YFe<sup>3+</sup> (4.3-  
324 6.3, 5.3 in average). The YFe<sup>3+</sup> is slightly higher in chromite inclusions (4.9-6.3) than in  
325 coronas (4.3-5.9). Chromite inclusions and coronas also display a clear negative correlation

326 between their TiO<sub>2</sub> content (0.12-0.51 wt%) and their XCr (Fig. 6).

327 At the scale of the whole ore body (~ 50 m-thick), chromite composition does not vary  
328 randomly. Progressive evolutions can be observed vertically (Fig. 7a): (1) the lowest XCr  
329 values are restricted around the lower breccia near the altitude 1020 m and do not exceed 55  
330 until the intermediate breccias system (1030-1034 m), then (2) XCr increases continuously to  
331 highest values (~60) toward the higher breccias before (3) a reverse decreasing tendency from  
332 the altitude 1050 m and within the uppermost part of the ore body. These continuous  
333 evolutions from a sample to another are independent of the amount of chromite in host rock at  
334 the scale of the whole stratiform chromitite. However, at a given altitude, XCr values higher  
335 than ~55 are generally observed in the more concentrated ores (scattered, antinodular or  
336 massive samples) rather than in the more disseminated ones (rare or very scattered chromite  
337 samples).

338 In detail, Cr and Al, which typically substitute for each other in spinel, show an  
339 unexpected decoupling at some levels. Although their concentrations are anti-correlated at the  
340 scale of the whole stratiform chromitite, they show a parallel decreasing trend within the  
341 intermediate breccias system (1030-1034 m) as well as above the altitude 1058 m (correlated  
342 to the TiO<sub>2</sub> content and negatively correlated to YFe<sup>3+</sup>). Otherwise, in between the three  
343 highest breccias (1049-1053 m), the Cr<sub>2</sub>O<sub>3</sub> content shows two successive increasing and  
344 decreasing trends within a few metres while Al<sub>2</sub>O<sub>3</sub> continuously increases on the same  
345 interval. XMg shares basically the same evolution as XCr and is particularly well-correlated  
346 to the Cr<sub>2</sub>O<sub>3</sub> content, while oppositely YFe<sup>3+</sup> is negatively correlated to XMg (Fig. 7a). The  
347 vertical evolution of the TiO<sub>2</sub> content is decoupled from the other chemical species/atomic  
348 ratios and shows, with few local exceptions, a global and relatively monotonous increase from  
349 about 0.30 wt% at the base of the stratiform chromitite to 0.50 wt% to its top (Fig. 7a). The  
350 few local exceptions occur generally at altitudes where magmatic breccias were observed.

351

352 *4.3. Silicate matrix*353 *4.3.1. Olivine*

354 The olivine matrix shows marked variations in composition with Fo ( $100 \times \text{Mg}/(\text{Mg} +$   
355  $\text{Fe}_{\text{total}})$  atomic ratio) = 90.4-95.2, NiO = 0.34 to 0.64 wt%, and CaO ranging from detection  
356 limit to 0.16 wt% (averaged per thin section). Vertical variations of both the Fo and the NiO  
357 content are closely correlated to the XCr, especially to the Cr<sub>2</sub>O<sub>3</sub> content, and to the XMg of  
358 chromite (Fig. 7a-c). We observe scattered values in Fo and NiO at the base of the stratiform  
359 chromitite, overlain by common and successive (1) decrease in between the two intermediate  
360 breccias, (2) increase to highest breccias (with a much higher variability at the altitude 1040  
361 m) and (3) decrease within the uppermost part of the ore body. MnO content is conversely  
362 anti-correlated to those chemical species/ratios (not shown). High CaO values of the olivine  
363 matrix (i.e. higher than 0.05 wt%) are restricted to the base of the stratiform chromitite,  
364 particularly around the lowest breccia at the altitude 1020 m (Fig. 7b). Beyond these global  
365 trends at the scale of the stratiform chromitite, alternations of increases and decreases in a  
366 given element or elemental ratio can be observed at a finest scale, especially from the highest  
367 breccias and above.

368

369 *4.3.2. Clinopyroxene*

370 Clinopyroxene, the second most abundant mineral phase after olivine in the matrix,  
371 shows XMg ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio, average per thin section) varying between  
372 91.1 and 94.9, and Al<sub>2</sub>O<sub>3</sub> = 1.8-4.4 wt%, Cr<sub>2</sub>O<sub>3</sub> = 1.1-1.5 wt, TiO<sub>2</sub> = 0.15-0.38 wt% and Na<sub>2</sub>O  
373 = 0.22-0.58 wt%. The XMg of clinopyroxene is correlated to Fo of olivine (and XMg of  
374 chromite) (Fig. 7c). While most difficult to decipher, the vertical evolution of both Al<sub>2</sub>O<sub>3</sub> and  
375 TiO<sub>2</sub> contents of clinopyroxene, as well as of Na<sub>2</sub>O (not shown), is negatively correlated to

376 the one of XMg. The few analysed hybrid diopsides (i.e. hybrid in composition between  
377 igneous clinopyroxene and pure hydrothermal diopsides; Rospabé et al., 2017) show higher  
378 XMg (94.3-97.0) and SiO<sub>2</sub> content (53.8-55.0 wt% instead of 51.8-53.6 wt%) coupled with  
379 lower Al<sub>2</sub>O<sub>3</sub> (0.47-1.4 wt%), TiO<sub>2</sub> (0.03-0.16 wt%) and Na<sub>2</sub>O (<0.23 wt%) contents than in  
380 clinopyroxene whose composition fits with an igneous origin.

381

#### 382 4.3.3. Orthopyroxene

383 The few orthopyroxene occurrences in the matrix are enstatites, with XMg (100 ×  
384 Mg/(Mg + Fe<sup>2+</sup>) atomic ratio) = 92.2-94.0, Al<sub>2</sub>O<sub>3</sub> = 1.33-1.38 wt%, CaO = 0.62-1.48 wt% and  
385 TiO<sub>2</sub> = 0.13-0.15 wt% (Table 1 and Supplementary Material).

386

#### 387 4.3.4. Amphibole

388 Excepting for one sample (06OM04-10, 1036 m), the matrix contains amphibole only  
389 above the higher breccias system in the uppermost part of the stratiform chromitite ore body.  
390 Amphibole is mostly pargasitic to edenitic hornblende. The variation ranges of XMg (100 ×  
391 Mg/(Mg + Fe<sup>2+</sup>) atomic ratio) and XNa (100 × Na/(Na + K) atomic ratio) are 89.6-92.4 and  
392 91.7-98.2 respectively (Na<sub>2</sub>O = 2.5-3.1 wt%; K<sub>2</sub>O = 0.08-0.40 wt%). The TiO<sub>2</sub> content of  
393 amphibole varies from 0.30 to 1.8 wt%. Similarly to what was described for the Al<sub>2</sub>O<sub>3</sub> content  
394 of chromite and for other chemical species in olivine and clinopyroxene (Fig. 7a-c),  
395 amphibole composition evolves differently below and above the altitude 1058 m. Below this  
396 level, XMg decreases from 92.4 to 90.1 and XNa and TiO<sub>2</sub> increase from 96.8 to 91.7 and  
397 from 1.5 to 1.8 wt% respectively. Above this level, the reverse trend is observed, XMg, XNa  
398 and TiO<sub>2</sub> evolving to 90.9, 98.2 and 0.30 wt% respectively.

399

#### 400 4.3.5. Garnet

401 Garnet observed at the same levels where the amphibole occurs is mostly grossular  
402 (sum of oxides ranging from 97.8 to 100.2 wt%), with  $\text{SiO}_2 = 36.0\text{-}39.6$  wt%,  $\text{CaO} = 36.9\text{-}$   
403  $38.2$  wt% and  $\text{Al}_2\text{O}_3 = 21.1\text{-}22.6$  wt%. The  $\text{Cr}_2\text{O}_3$  and  $\text{TiO}_2$  contents are comprised between  
404 0.09 and 1.36 wt%, and 0.03 and 0.15 wt% respectively (Table 1 and Supplementary  
405 Material).

406

#### 407 *4.4. Chemical profiles across samples: chromite-matrix composition co-variations*

408 In order to better describe the mm- to cm-scale composition co-variations in chromite  
409 and olivine (and clinopyroxene when present), we performed chemical profiles across four  
410 samples 06OM04-3B2 (alt. 1021.5 m), -5D (1030.5 m), -12E1 (1040.3 m) and 04OM31E  
411 (collected apart from the cross-section, very near the base of the stratiform chromitite) (Fig.  
412 8).

413 The samples selected display mm- to cm-scale extremely variable chromite and  
414 olivine modal contents, covering the entire variation range described previously (from rare  
415 chromite in dunite to massive chromitite). Despite differences in terms of absolute chemical  
416 compositions from one sample to another, these samples share common relationships between  
417 chemical and modal variations. Higher amounts of chromite are perfectly correlated to higher  
418  $\text{XMg}$  of chromite and  $\text{Fo}$  of olivine:  $\text{Fo}$  reaches values of about 90.5 in chromite-bearing  
419 dunites (rare disseminated chromite) to 94.5 in antinodular to massive chromitites. The  $\text{NiO}$   
420 content of olivine, despite a high scatter, is correlated to the  $\text{Fo}$ . The  $\text{TiO}_2$  content of chromite  
421 displays a small-scale high dispersion; however, it seems that highest values are generally  
422 observed in antinodular to massive ore portions.

423 The main difference with the chemical variations at the scale of the whole stratiform  
424 ore body resides in the distribution of  $\text{Cr}$ ,  $\text{IVAl}$  and  $\text{Fe}^{3+}$  in chromite. In samples - or parts of  
425 thin sections - with the highest amount of chromite (scattered, antinodular and massive),  $\text{Cr}$

426 and Al are negatively correlated (Fig. 8b-c-d), with the highest Cr values (up to 1.06 apfu)  
427 observed where the amount of chromite is high. Oppositely, in portions containing  
428 disseminated chromites (rare or very scattered), Cr and Al increase or decrease concomitantly  
429 while  $\text{Fe}^{3+}$  is negatively correlated to those two chemical species (Fig. 8a-b). In these cases,  
430 Cr and Al generally decrease together and  $\text{Fe}^{3+}$  increases as the sample becomes more dunitic.

431 The thin section of sample 06OM04-12E1 shows interstitial clinopyroxene, restricted  
432 to the decreasing peak of the amount of chromite and above, within the upper massive interval  
433 (Fig. 8d). Within this interval, we observed the increase of XMg (from 94.0 to up to 99)  
434 associated to the decrease of  $\text{Al}_2\text{O}_3$  (from 2.6 to 1.4 wt%),  $\text{TiO}_2$  (from 0.29 to 0.20 wt%) and  
435  $\text{Na}_2\text{O}$  (from 0.38 to 0.22 wt%) contents of clinopyroxene.

436 In a different view, Figure 9 shows that in the four samples (06OM04-3B2, -5D, -  
437 12E1 and 04OM31E), the evolution of Fo of olivine and of XMg, XCr,  $\text{YFe}^{3+}$  and  $\text{TiO}_2$  of  
438 chromite follows the chromite abundance. Systematically Fo and XMg increase and  $\text{YFe}^{3+}$   
439 decreases when variations in the chromite modal content ranges from 0 to 30-40% chromite.  
440 Above 40%, there is no more evolution of these chemical ratios which seem to be buffered to  
441 values of about 93.5, 65 and 4.5 respectively. XCr is constant around 53 in the two samples  
442 06OM04-5D and 04OM31E whatever the chromite modal abundance, and tends to this value  
443 in the two other samples when the chromite abundance becomes lower than 40% (i.e. steady  
444 values near 49 and 57 in 06OM04-3B2 and -12E1 respectively above 40% of chromite). The  
445  $\text{TiO}_2$  content does not show marked variation from below to above 40% of chromite, and  
446 appears not affected by the variation of the chromite modal content excepting a very slight  
447 and constant increase from scarce to massive ores.

448

#### 449 4.5. Silicate inclusions

##### 450 4.5.1. Orthopyroxene

451 Orthopyroxene in inclusion in chromite is enstatite, with a similar XMg (92.5-94.5),  
452 slightly higher Cr<sub>2</sub>O<sub>3</sub> (0.60-1.2 wt%) and lower Al<sub>2</sub>O<sub>3</sub> (0.41-1.3 wt%), CaO (0.21-0.83 wt%)  
453 and TiO<sub>2</sub> (0.05-0.11 wt%) contents than those of the few orthopyroxene crystals observed in  
454 interstitial position and associated to the other matrix minerals. Higher XMg values are  
455 observed along the interval 1040-1050 m (Fig. 7d), together with high XCr and XMg of  
456 chromite (Fig. 7a) and high Fo of olivine and XMg of clinopyroxene (Fig. 7b-c). In the same  
457 way, Al<sub>2</sub>O<sub>3</sub> and TiO<sub>2</sub> contents increase together from 1026 to 1036 m, decrease to about 1050  
458 m, then globally re-increase to the top of the stratiform chromitite with a clear disturbance  
459 around the altitude 1058 m (Fig. 7d). These variations largely mimic the ones observed for  
460 Al<sub>2</sub>O<sub>3</sub> of host chromites and for Al<sub>2</sub>O<sub>3</sub> and TiO<sub>2</sub> contents of matrix clinopyroxene.

461

#### 462 4.5.2. Clinopyroxene

463 The less common inclusions of clinopyroxene exhibit variable XMg (93.0-97.3) and  
464 Na<sub>2</sub>O content (0.54-1.3 wt%), both higher than those of the clinopyroxene in the matrix.  
465 These inclusions show more restricted Al<sub>2</sub>O<sub>3</sub> (2.4-2.6 wt%), Cr<sub>2</sub>O<sub>3</sub> (1.3-1.6 wt%) and TiO<sub>2</sub>  
466 (0.20-0.36 wt%) compositional ranges (averaged per thin section). Only one inclusion of  
467 diopside of likely hydrothermal origin (cf. Python et al., 2007; Rospabé et al., 2017) was  
468 found. Its XMg and Al<sub>2</sub>O<sub>3</sub>, Cr<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub> and Na<sub>2</sub>O contents values are 96.0, 1.0 wt%, 0.73  
469 wt%, 0.22 wt% and 0.20 wt% respectively.

470

#### 471 4.5.3. Amphibole

472 Amphiboles occurring as inclusions in chromites are pargasites/pargasitic  
473 hornblendes, for more than 60% of them, or magnesio-hastingsites/magnesio-hastingsitic  
474 hornblendes, for more than 30% of them. The XMg and TiO<sub>2</sub> content values, averaged for  
475 each sample, are much variable (88.8-93.3) and higher (1.8-2.8 wt%), respectively, compared

476 to those of amphiboles from the matrix. Within each sample, the variability from an  
477 amphibole inclusion to another is much higher than concerning amphibole in the matrix (Fig.  
478 7e). Their Na<sub>2</sub>O content ranges from 2.9 to 4.1 wt%, while their K<sub>2</sub>O content is lower than  
479 0.10 wt%, making their XNa to be restricted to very high values (98.2-99.3). The vertical  
480 evolution of XMg and Al<sub>2</sub>O<sub>3</sub> values basically follows the same tendency observed in  
481 orthopyroxene inclusions. However, the distribution of the TiO<sub>2</sub> content is much scattered,  
482 with a certain increase from lower to higher values from the base to the top of the stratiform  
483 chromitite (Fig. 7e).

484

#### 485 4.5.4. Mica

486 Micas, absent in the matrix and observed solely as inclusions, have compositions  
487 evolving from phlogopite to aspidolite (Na-rich phlogopite) end-members. Their averaged  
488 XMg and Cr<sub>2</sub>O<sub>3</sub> and (high) TiO<sub>2</sub> contents range from 93.7 to 96.2, from 1.9 to 3.0 wt% and  
489 from 2.6 to 4.3 wt% respectively. Micas are also characterized by high Na<sub>2</sub>O (3.1-5.6 wt%)  
490 and moderate K<sub>2</sub>O (0.61-3.8 wt%) contents (Na<sub>2</sub>O + K<sub>2</sub>O = 6.1-7.0 wt%), the XNa varying  
491 from 55.0 to 93.2 (averaged per thin section). Vertically along the stratiform chromitite, mica  
492 composition follows the same evolution observed in other minerals, with high XMg values  
493 along the interval of 1040-1050 m and a common evolution of their Al<sub>2</sub>O<sub>3</sub> and TiO<sub>2</sub> contents  
494 (not shown). The XNa values lower than 90 (enrichment in K<sub>2</sub>O relative to Na<sub>2</sub>O) are  
495 observed at the base of the ore body, in between the lowest and the intermediate breccias, as  
496 well as along its uppermost part within and above the higher breccias system (Fig. 7f). These  
497 lower XNa values result either, in a given sample, from compositions evolving continuously  
498 from Na-rich to K-rich end-members (e.g. sample 06OM04-5C2 (1030 m) with Na<sub>2</sub>O = 0.34-  
499 6.3 wt%, K<sub>2</sub>O = 0.32-6.9 wt% and XNa = 7.0-96.8), or from the average of values from two  
500 distinct populations within a single sample (e.g. sample 06OM04-4 (1026 m) with Na<sub>2</sub>O =

501 1.3-2.7 wt% and  $K_2O = 2.7-7.2$  wt% in the first population, and  $Na_2O = 5.9-6.9$  wt% and  $K_2O$   
502  $= 0.17-0.60$  wt% in the second population, giving  $X_{Na} = 27.9-42.7$  and  $93.7-98.3$   
503 respectively).

504

#### 505 *4.5.5. Other inclusions*

506 Similarly to what is observed in the matrix, the few garnet inclusions in chromite are  
507 grossular. We identified also two nepheline inclusions that are sodic (K-poor) and Si-rich,  
508 with  $Na_2O$ ,  $SiO_2$  and  $Al_2O_3$  averaged contents of 18.7 wt%, 49.3 wt% and 30.6 wt%,  
509 respectively ( $K_2O = 0.14$  wt%) (Supplementary Material).

510 A silica-rich inclusion identified in the sample 06OM04-17 has high  $SiO_2$  (76.0 wt),  
511  $Al_2O_3$  (17.7 wt%) and  $Na_2O$  (8.6 wt%) concentrations, and a minor FeO content (0.32 wt%).  
512 While the nature of this very minor phase is quite enigmatic, its composition fits with a  
513 submicronic-scale mixing between ~80% albite and ~20% quartz. It might be a micropertite.

514

## 515 **5. Discussion**

### 516 *5.1. Structure and petrology of the stratiform ore body*

517 The Maqsad area is characterized by a high abundance of podiform chromitites  
518 (Borisova et al., 2012; Ceuleneer and Nicolas, 1985; Leblanc and Ceuleneer, 1991; Rollinson  
519 and Adetunji, 2013; Rollinson et al., 2018; Zagrtednov et al., 2018), defining one of the most  
520 important chromitiferous districts in the Oman ophiolite. The ore deposits occurring within the  
521 dunitic mantle-crust transition zone at the top of the Maqsad diapir, including the studied  
522 stratiform chromitite ore body, developed during the last stage of the upwelling of the  
523 asthenospheric mantle diapir. Accordingly, they escaped solid-state deformation related to the  
524 spreading ridge activity and off-axis transposition. The layered structure of the stratiform  
525 chromitite, the preservation of delicate igneous texture, including graded bedding, the absence

526 of any sign of solid-state deformation (Figs. 3 and 4), and the entrapment of silicate inclusions  
527 (Fig. 5), evoke a pile of layered cumulates whose crystallization was synchronous with the  
528 one of the overlying gabbroic lower crust.

529 The stratiform ore body consists of interlayered levels of scattered chromite in dunites  
530 and of more concentrated levels, mostly with antinodular textures but hosting layers of  
531 massive chromitites a few cm- to 3 m-thick (Figs. 2c and 3). Another significant feature is the  
532 presence of levels of magmatic breccias, a few tens of cm in thickness at most, made of  
533 angular wehrlitic fragments (i.e. dunites displaying a higher clinopyroxene content than  
534 average in this part of the DTZ) in the chromitite, all together embedded in the olivine matrix  
535 (Fig. 3h-i). They most likely witness successive melt injection events, potentially with  
536 increased melt/mush ratios, during the formation of the stratiform ore body. Other related  
537 melt segregation and injection textures reflecting the former dynamic environment were  
538 frozen at some levels (Fig. 3f-g). Chromite-rich layers I, II and IV were found at the same  
539 levels where were described the lower, intermediate and higher brecciated systems  
540 respectively (Fig. 2). However, the fact that the two layers III and V seem to be unrelated to  
541 these features allows concluding that breccias and related melt injections do not solely control  
542 the local concentration of Cr.

543 The three-dimensional structure of the stratiform chromitite is difficult to assess due to  
544 erosion; however, its lateral extension apparently did not exceed a few hundred metres  
545 regarding the significant decrease of the chromite modal content in surrounding outcrops.  
546 Originally, as suggested by its shape before mining (Fig. 3a), it was probably formed as a  
547 circular lens bordered to the southwest by a high temperature N130-oriented normal fault.  
548 This fault presently makes the contact between, to the SW, layered troctolites that originally  
549 formed higher up in the section, and, to the NE, dunites from the DTZ that host the stratiform  
550 chromitite.

551 The mineral distribution confers to the stratiform chromitite a major and remarkable  
552 contrast between (1) minerals constituting the matrix between chromite grains, mostly  
553 anhydrous (olivine, clinopyroxene) except for its upper part (amphibole, garnet), and (2)  
554 minerals of the chromite-hosted inclusions which are, as a rule, hydrous or anhydrous silicates  
555 (mostly amphibole, orthopyroxene and mica) witnessing the high silica and water contents in  
556 their parental melt/fluid. The origin of the silicate inclusions enclosed in chromite grains from  
557 Oman ophiolite chromitites were interpreted either as *in situ* crystallization products from  
558 trapped melt inclusions (e.g. Schiano et al., 1997; Rollinson et al., 2018), or as solid phases  
559 already crystallized in the parent magma and that were subsequently trapped when chromite  
560 formed (e.g. Borisova et al., 2012). The fact that the surrounding Maqsad DTZ dunites  
561 contain numerous mineral impregnations that are actually fractional crystallization products  
562 from percolating melts (e.g. Boudier and Nicolas, 1995; Koga et al., 2001; Abily and  
563 Ceuleneer, 2013), including widespread orthopyroxene and amphibole in the upper level of  
564 the DTZ (Rospabé et al., 2017), suggests a common origin and supports the view that most  
565 silicate inclusions in chromite were already crystallized prior to their entrapment. Otherwise,  
566 it is worth noting that the silicate inclusions in the stratiform chromitite become more  
567 abundant approaching the intermediate and higher magmatic breccias systems (Fig. 2). This  
568 observation suggests that the same dynamic event leading to the breccias formation triggered  
569 the sudden crystallization of silicate crystals that would eventually be entrapped by chromite.

570

## 571 5.2. Chemistry of the stratiform ore body

572 Averaged XCr values of chromite from the stratiform ore body are relatively moderate  
573 (49.0-59.9), similar to those of the disseminated chromite from the surrounding DTZ and  
574 contrasting with those of other ore bodies and their host dunites elsewhere in Oman,  
575 especially in the northern massifs of the ophiolite where XCr can reach much higher values

576 (Ahmed and Arai, 2002; Augé, 1987; Miura et al., 2014; Rollinson, 2008; Rollinson and  
577 Adetunji, 2015). Evenly, the  $\text{TiO}_2$  content of the stratiform chromite is typically high (0.25-  
578 0.54 wt%), a characteristic shared by most chromitites and scattered chromites in dunites  
579 from the Maqsad area and contrasting with widespread lower values observed in almost all  
580 other massifs of the Oman ophiolite. This is a major evidence for the MORB kindred of the  
581 parent melt of the various igneous products issued from the Maqsad diapir (Clénet et al.,  
582 2010; Python et al., 2008), including chromitites (Borisova et al., 2012; Leblanc and  
583 Ceuleneer, 1991; Zagrtzenov et al., 2018).

584 Mineral chemical compositions show significant variations at different scales, both  
585 locally at a single altitude and at the scale of the whole stratiform chromitite (Figs. 6 and 7).  
586 At the local scale, variations mainly concern  $\text{XMg}$  and  $\text{YFe}^{3+}$  of chromite and Fo and the NiO  
587 content of olivine, and are clearly related to thin section scale variations in the modal  
588 proportion of chromite and olivine (Fig. 8). This is observed for intervals with a chromite  
589 modal content ranging from 0 to 30-40%. Above 40% of chromite, the chemical compositions  
590 do not evolve anymore, suggesting a buffering process around a composition possibly in  
591 equilibrium with the chromite parent melt (Fig. 9).

592 In contrast, chemical variations vertically along the stratiform ore body define  
593 complex trends of higher amplitude than locally, mostly concerning the XCr of chromite and  
594 in a lesser extent the Fo, NiO and CaO contents of olivine as well as the clinopyroxene  
595 composition (Fig. 7a-c). This is reminiscent of the trends observed globally all across the  
596 DTZ (Abily and Ceuleneer, 2013; Rospabé et al., 2018, 2019). Parallel trends are observed in  
597 the composition of silicate inclusions in chromite, as in the  $\text{TiO}_2$  content of amphibole and  
598 alkalis of mica (Fig. 7d-f). The high amplitude of these variations cannot have solely resulted  
599 from sub-solidus re-equilibration. *Interestingly, these large-scale trends do not mirror at all*  
600 *the variations in modal composition of the ore but correlate to the presence of the three*

601 *magmatic breccia systems (i.e. we observe continuous trends bounded by the three levels*  
602 *where brecciated levels are concentrated).*

603 Accordingly, this suggests that the current chemical characteristics along the  
604 stratiform chromitite result from multi-scale processes: the development of high temperature,  
605 magmatic breccias have extensively controlled the large scale chemical evolution of the  
606 whole chromitite ore body, while the chromite mode and subsolidus chemical re-equilibration  
607 affected the small, mm to cm-scale chemical variations.

608 More generally, the olivine matrix from the stratiform chromitite displays Fo and NiO  
609 content significantly higher than in nearby dunites from the DTZ and in mantle harzburgites  
610 from the Sumail massif, reaching values up to 95.3 and 0.63 wt% respectively (Fig. 10a).  
611 Even higher values were reported for mantle chromitites from the north of the Oman ophiolite  
612 (Augé, 1987; Miura et al., 2014; Rollinson, 2008, 2019) (Fig. 10b). In the Sumail massif, all  
613 podiform chromitites share the same feature, their olivine matrix falling in the same chemical  
614 variation range whatever they outcrop within the DTZ or the mantle section. Furthermore, a  
615 chromitite dike from the Maqsad DTZ, located few hundred metres to the northeast of and at  
616 a higher structural level than the present stratiform ore body, displays higher olivine Fo values  
617 in olivine-plagioclase nucleus around which chromite nodules grown than in the olivine  
618 matrix surrounding nodules (Zagrtdenov et al., 2018). Taken together, these two observations  
619 allow us suggesting that Mg-Fe<sup>2+</sup> post-magmatic subsolidus exchanges between chromite and  
620 olivine (e.g. Cameron, 1975; Fabriès, 1979; Grieco et al., 2018; Lehmann, 1983; O'Neill and  
621 Wall, 1987) cannot solely explain such high Fo values associated to high NiO content along  
622 the stratiform chromitite: (1) first, the contrast between DTZ and mantle chromitites from the  
623 Sumail massif on one hand and mantle chromitites from the northern massifs on the other  
624 hand matches with the two main magmatic series identified along the Oman ophiolite. These  
625 series were highlighted following the nature of cumulates frozen in the mantle section (i.e.

626 MORB mainly, but not only, in the SW of the Oman ophiolite, more widespread depleted  
627 calc-alkaline in the central and northern massifs; Python and Ceuleneer, 2003; Python et al.,  
628 2008); (2) then in the case of the other chromitite dike (Zagrtdenov et al., 2018), later re-  
629 equilibration should have affected equally olivine in nucleus and olivine in the matrix, which  
630 is not the case. However, it might strongly depend on the scale we are investigating these  
631 chemical variations.

632 The compositions of coexisting olivine and chromite (Figs. 7 and 8) suggest  
633 subsolidus re-equilibration at 800 to 900°C, as commonly observed in metamorphosed  
634 peridotites (Evans and Frost, 1975). In such rocks, the partitioning of Mg and Fe<sup>2+</sup> depend on  
635 the volume ratio of olivine to chromite as well as on temperature (e.g., Arai, 1980).  
636 Interestingly, other studies report on a negative (and not positive as in our case) correlation  
637 between Fo of olivine and XMg of chromite (e.g. Bussolesi et al., 2019). These authors  
638 interpret this correlation in terms of solid-state exchange of Mg and Fe between olivine  
639 (incorporation of Mg) and chromite (incorporation of Fe) grains. In our case this process  
640 alone doesn't fit simple mass balance consideration as the correlation between XMg and Fo is  
641 also associated to variations in the NiO content of olivine and in the XCr, YFe<sup>3+</sup>, and  
642 sometimes TiO<sub>2</sub> of chromite. An avenue of explanation might be to evoke a percolating  
643 interstitial melt (i.e. an open system), re-equilibration being likely triggered by such an  
644 interstitial melt that makes diffusional processes more efficient. The equilibration is limited in  
645 space, possibly in a few cm<sup>3</sup>, and the mass balance for Mg and Fe have been maintained in the  
646 restricted area.

647 At a much larger scale, the variations observed over up to 10 m vertically are likewise  
648 correlated to other chemical indexes such as XCr (Fig. 7), and are not correlated to the ore  
649 concentration as was stressed before (Fig. 2). The chemical composition thus better result  
650 from the primary processes that originated the stratiform chromitite, including the

651 development of the magmatic breccias, rather than from secondary processes.

652

### 653 *5.3. Tentative scenario for the formation of the stratiform ore body*

654 Since decades, several mechanisms were proposed to account for the formation of  
655 chromitite ore bodies: mixing between contrasted magmas, addition of SiO<sub>2</sub> and/or H<sub>2</sub>O,  
656 change in the pressure among others (e.g. Irvine, 1975, 1977a, 1977b; Latypov et al., 2018;  
657 Lipin, 1993; Matveev and Ballhaus, 2002; Mondal and Mathez, 2007; Naldrett et al., 2012;  
658 Spandler et al., 2005). One of the most popular models is the one of Irvine (1975, 1977a),  
659 which proposed that massive crystallization of chromite is triggered by the mixing between a  
660 basaltic melt saturated in chromite and a more silicic melt (the same melt at a more evolved  
661 stage, or felsic contamination by assimilation of crustal rocks surrounding the magma  
662 reservoir). More generally, it was also proposed that the nucleation of ‘crustal-like’ minerals  
663 associated to the massive chromite crystallization may result from the hybridization between  
664 the basaltic fraction and a siliceous volatile- and Na-rich fluid or melt fraction (e.g. Johan et  
665 al., 1983, 2017; McElduff and Stumpfl, 1991; Spandler et al., 2005; Talkington et al., 1984).  
666 In other studies, it was shown that this mixing process cannot account for the formation of  
667 some chromitites, especially in the Bushveld Complex (e.g. Mondal and Mathez, 2007;  
668 Naldrett et al., 2012). It is probable that a single model cannot consensually explain the  
669 formation of all chromitites worldwide, in old intracontinental intrusions or in ophiolites, but  
670 that different processes, working alone or together, may lead to the formation of a given ore  
671 body.

672 In the Maqsad area, we evidenced witnesses of the involvement of such hybrid melts  
673 or fluids. The partial melting within the Maqsad diapir produced ‘dry’ (i.e. water poor)  
674 MORB, evidenced by the petrology (nature, crystallization sequences) and geochemistry of  
675 troctolite and ol-gabbro that crop out in the harzburgitic mantle sequence (Benoit et al., 1996;

676 Ceuleneer et al., 1996; Python et al., 2003). However, the widespread occurrence of  
677 orthopyroxene, amphibole, diopside and garnet, interstitially between olivine grains in the  
678 DTZ dunites (i.e. thus at a higher structural level) attests to the genesis of hydrous melts at the  
679 mantle-crust transition zone (Rospabé et al., 2017; 2018). We proposed that it resulted from  
680 the hybridization between the uprising MORB and a hydrated component coming from above,  
681 probably seawater in origin and potentially trondhjemitic in nature following the hydrous  
682 melting of the country rocks (see also Amri et al., 1996; Benoit et al., 1999; Koepke et al.,  
683 2005, 2014; Nonnotte et al., 2005 concerning the hydrous re-melting of the shallow mantle  
684 and deep crust beneath oceanic spreading centres). The similar minerals, interstitially in  
685 dunites and in inclusions in both disseminated chromites in dunites and more concentrated  
686 chromitites, allowed concluding that the hybrid melt was strongly involved in the formation  
687 of both dunites and chromitites from the Maqsad DTZ following melt- or magma-fluids-  
688 harzburgite reactions (Rospabé et al., 2017, 2018). Returning to the stratiform chromitite, it  
689 allows us to favour the hypothesis of the mixing, or hybridization, between contrasted melts  
690 and/or fluids at its origin.

691 The relationship between (1) the location of brecciated levels, (2) the vertical chemical  
692 evolution of all mineral phases along the stratiform chromitite, and (3) the higher abundance  
693 of orthopyroxene, amphibole and micas included in chromite grains, supports that melt or  
694 magma injections strongly influenced the composition of the parent melt from which  
695 crystallized the ore body. The existence of such injections is witnessed by the several levels of  
696 breccia interlayered with the dunites and chromitites. Especially, the decoupling between the  
697 zigzag variation pattern in XCr, and the smooth increase of the TiO<sub>2</sub> content of chromite,  
698 from the base to the top of the ore body (Fig. 7a), cannot result solely from fractional  
699 crystallization from a single melt batch. It probably reflects the involvement of several melt  
700 injections and of melt fractions of contrasted composition. Additionally, it was shown in the

701 Maqsad area that fault zones developed early, during oceanic spreading and crystallization of  
702 the crust, and were probably the main vectors for the introduction of the hydrated component  
703 in the lower oceanic crust and at Moho level (Abily et al., 2011; Rospabé et al., 2019). In this  
704 way, magmatic breccias we observe within the stratiform chromitite may represent the  
705 nucleation state of such syn-magmatic faults, developed within a still unconsolidated mush  
706 and responsible of fluids and/or melt pulses that hybridized with magmatic mush containing  
707 interstitial MORB. The hybridization from brecciated levels may have thus been responsible  
708 for:

709 - The precipitation of Ti-rich chromites that entrapped mainly orthopyroxene-  
710 amphibole-mica and high XCr chromites, which were all microcrystals in suspension in the  
711 former hydrated Si- and Na-rich melt or fluid (e.g. Matveev and Ballhaus, 2002);

712 - The higher amount of orthopyroxene, amphibole and mica in inclusion in chromite at  
713 the same levels than the breccias (Fig. 2);

714 - The higher XCr of chromite and Fo and NiO content of olivine toward the higher  
715 breccias system (Fig. 7), buffered to high values due to a particularly strong involvement of  
716 the hydrated component at this specific level, probably with a higher supply of the hydrated  
717 melt (possibly saturated with a fluid);

718 - The local correlation between Cr and Al near breccias, negatively correlated to the  
719  $Fe^{3+}$  (Figs. 6 and 7), being the imprint of a disturbance of the Cr-Al substitution process by  
720 the involvement of such hydrated melt/fluids at these levels;

721 - The widespread occurrence of clinopyroxene and the total absence of plagioclase  
722 interstitially between olivine grains in the dunites hosting the stratiform chromitite (Fig. 2;  
723 wehrlitic dunites in Figs. 3h and 4h). It probably crystallized from the newly hybrid MORB,  
724 more evolved (continuous increase of the  $TiO_2$  content of chromite, Fig. 7) and hydrated  
725 (appearance of interstitial amphibole and garnet in the upper part of the stratiform chromitite;

726 Fig. 2) upsection, after the solely crystallization and concentration of chromite.

727         However, while the hybridization of the MORB with the hydrated Si- and Na-rich  
728 melt/fluid along the magmatic breccias is responsible for the distribution of the silicate  
729 inclusions and of the whole chemical structuration of the stratiform chromitite (e.g. richness  
730 in Cr), the hybridization influenced only partly the ore concentration (i.e. chromite modal  
731 content). It is explained here by the fact that the chromite-rich layers I, II and IV only are  
732 correlated to the presence the magmatic breccias, whereas the layers III and especially V (that  
733 is the most massive, 3 m in thickness; Figs. 2c and 3) show no evident influence of the  
734 magmatic breccias and calls for another explanation. Although as a working hypothesis, the  
735 decoupling between the chromite modal concentration and its Cr concentration may be  
736 accounted by the small-scale highly contrasted saturation in chromite of the former MORB  
737 before its hybridization. It was shown that highly variable fractions of melts percolated  
738 through the Maqsad DTZ (Rospabé et al., 2018), and that decoupling may thus also reflect the  
739 locally variable advancement of the hybridization process (or with which proportions of each  
740 of the two contrasted melt/fluid the hybridization operated). In this context, it is likely that  
741 podiform chromitites form early during the development of the mantle-crust transition zone as  
742 a companion product of the dunites' formation.

743

#### 744 **Conclusion**

745         This paper illustrates the usefulness of studies combining accurate field descriptions  
746 and dense sampling of chromite ore bodies. The high quantity of samples reveals the  
747 variability in composition of the mineral composition while the origin of this variability can  
748 be understood if and only if each sample is positioned in its geographical and structural  
749 context.

750         In the case of the Maqsad stratiform chromitite, our study has established that no

751 relationship exists between the distribution of the layers of massive ore (i.e. chromite modal  
752 content) and the variation patterns of the chemical composition of the ore (i.e. richness in Cr)  
753 and of associated silicate minerals. However, these patterns are consistent with the  
754 distribution of magmatic breccias, and hence of melt injection in crystallizing cumulates.  
755 Accordingly, different processes condition the variations in ore concentration on one hand  
756 and in ore composition on the other hand.

757         The major contrast between the nature of the silicate minerals making the ore matrix  
758 and of the silicates entrapped within the chromite grains confirms that hybridization between  
759 a primitive basaltic melt (of MORB-type in the case of this particular ore body) and a hydrous  
760 silica-rich melt of possible hydrothermal origin primarily conditions the formation of  
761 chromite ore bodies as well as of the dunitic matrix at the mantle-oceanic crust transition  
762 zone.

763

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773 de l'Univers).

774

#### 775 **Figure captions**

776 **Figure 1:** a) General map of the different massifs constituting the Oman ophiolite. b)  
777 Geological map of the studied Maqsad area in the Sumail massif, modified after Zagrtdenov  
778 et al. (2018) (enlargement of the square in Figure 1a). The red star indicates the stratiform  
779 chromitite investigated in this study, and located within the dunitic mantle-crust transition  
780 zone. White stars are other chromitite ore bodies occurring in the Maqsad area.

781  
782 **Figure 2:** Petrological log of the cross-section sampled along the stratiform chromitite ore  
783 body. Collected samples (labels are reported to the right of the petrological log) are classified  
784 in five main lithological groups depending on their chromite modal content, from rare  
785 (dunites containing few percent) to very scattered, scattered, antinodular and massive. The  
786 five layers that show a particularly high chromite modal content are annotated from I, the  
787 lowest chromite-rich level, to V, the highest one. Magmatic breccias are reported at the  
788 altitude they were described on the field (light blue lines). The six main brecciated levels are  
789 described as the lower breccia (1020 m), the intermediate breccias system (two breccias at  
790 altitudes 1030 and 1034 m), and the upper breccia systems (three breccias in between 1046  
791 and 1052 m). The table in the middle reports the presence/absence of each identified silicate  
792 phase in the matrix (mainly olivine and clinopyroxene) or occurring as inclusion in chromite  
793 (mainly orthopyroxene, amphibole and mica). The diagram to the right represents the relative  
794 abundance of inclusions, from 0 to 8, based on thin sections observations using optical  
795 microscope. The value 1 corresponds to samples containing only rare chromite grains that  
796 enclose one or two silicate inclusions at most; the value 8 was attributed to samples in which  
797 almost all chromite grains contain several (more than 10 in some cases) silicate inclusions.  
798 Inclusions were observed in all the samples from the studied set, explaining that no one was  
799 classified as 0 (absence of inclusion). Abbreviations as following: am: amphibole; cpx:  
800 clinopyroxene; di: hybrid diopside; gt: garnet; mi: mica; ne: nepheline; ol: olivine; opx:

801 orthopyroxene.

802

803 **Figure 3:** Field photographs along the stratiform chromitite. a and b) View of the stratiform  
804 chromitite before (a: 2006) and after (b: 2016) its exploitation by mining companies. c)  
805 Massive chromitite, about 3 m thick, near the top of the stratiform ore body. d) Layered  
806 gradual variations (“graded beddings”) of the relative modal abundance of chromite and of  
807 olivine. e) Sharp transition from layered scattered to antinodular texture to massive  
808 chromitite. f and g) Migration (f) and injection (g) textures within the olivine-rich matrix. h  
809 and i) Magmatic breccias composed of angular dunitic fragments included in surrounding  
810 chromitite.

811

812 **Figure 4:** Insights of increasing abundance of chromite in thin sections (natural light) from a)  
813 scarce (06OM04-1) to b) very scattered (06OM04-2A), c) scattered to antinodular (06OM04-  
814 3C1), and d) antinodular (to the left) to massive (to the right) (06OM04-6B2). e) Small-scale  
815 (mm to cm) variation of chromite mode within single samples (to the left 06OM04-5A, to the  
816 right 06OM04-5B). f) Brecciated sample 06OM04-6B1 composed of chromitite and  
817 clinopyroxene-bearing dunite fragments. g) Olivine matrix (two grains) in an antinodular to  
818 massive chromitite sample (06OM04-21C). h) Antinodular texture formed by chromite and  
819 oikocrystic clinopyroxene (06OM04-19(1)). Numerous inclusions are observed within  
820 chromite grains.

821

822 **Figure 5:** Photomicrographs of a few examples of silicate inclusions in samples from the  
823 stratiform chromitite. (a and b) The two first pictures show that silicate inclusions are  
824 sometimes organized as coronas within the chromite grain (circled by the white arrows),  
825 surrounding a central inclusion (06OM04-10 and -12B2C). (c) Enlargement of (b) showing

826 the central amph-opx-mica polymineralic inclusion. (d) Polymineralic inclusion made of the  
 827 amphibole-mica-nepheline assemblage (06OM04-03A). (e) Large ( $> 100 \mu\text{m}$ ) and well-  
 828 preserved inclusion of grossular garnet in transmitted and crossed-nicols lights to the left  
 829 (06OM04-26). The pictures in transmitted light to the right allow appreciate the isotropic  
 830 character of the garnet. (f and g) Another example of a corona of inclusions (f). The inclusion  
 831 to the top right (g), possibly a microperthite, has a silica-rich composition which fits with a  
 832 mixture made of albite and quartz (06OM04-17). Abbreviations are as follows: amph,  
 833 amphibole; gt, grossular garnet; ne, nepheline; opx, orthopyroxene.

834

835 **Figure 6:** Chromite chemical composition from the stratiform chromitite, illustrated in an  
 836 enlargement of the Cr-Al axis from the ternary Al-Cr-Fe(III) diagram (top) and in the  
 837 variation of XCr and of the  $\text{TiO}_2$  content as a function of XMg and XCr respectively  
 838 (bottom).  $\text{XCr} = 100 \times \text{Cr}/(\text{Cr} + \text{Al})$  atomic ratio;  $\text{XMg} = 100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio.  
 839 The colours of the dots represent the chromite modal content in each analysed sample, as rare  
 840 in light green, very scattered in green, scattered in orange, antinodular in red and massive in  
 841 black.

842

843 **Figure 7:** Vertical evolution of the mineral chemical composition along the studied stratiform  
 844 chromitite, plotted as a function of the absolute altitude. a) the XCr ( $100 \times \text{Cr}/(\text{Cr} + \text{Al})$   
 845 atomic ratio), the  $\text{Cr}_2\text{O}_3$  and  $\text{Al}_2\text{O}_3$  contents, the XMg ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio),  
 846 the  $\text{YFe}^{3+}$  ( $100 \times \text{Fe}^{3+}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$  atomic ratio) and the  $\text{TiO}_2$  content of chromite, in b)  
 847 the Fo ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}_{\text{total}})$  atomic ratio) and the NiO and CaO contents of olivine matrix,  
 848 in c) the XMg ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio) and the  $\text{Al}_2\text{O}_3$  and  $\text{TiO}_2$  contents of  
 849 clinopyroxene matrix, in d) the XMg ( $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$  atomic ratio) and the  $\text{Al}_2\text{O}_3$  and  
 850  $\text{TiO}_2$  contents of orthopyroxene in inclusion in chromite grains, in e) the XMg ( $100 \times$

851 Mg/(Mg + Fe<sup>2+</sup>) atomic ratio) and the TiO<sub>2</sub> content of amphibole in inclusion, and in f) the  
852 XNa (100 × Na/(Na + K) atomic ratio) of mica in inclusion. Coloured dots are averaged  
853 values per sample, depending on the chromite abundance in the sample, while small grey dots  
854 show the variability of un-averaged, analysed values. Blue lines are magmatic breccias.

855  
856 **Figure 8:** Chemical profiles along the four samples 06OM04-3B2 (a), 06OM04-5D (b) and  
857 06OM04-12E1 (d), all from the cross-section collected along the stratiform chromitite, and  
858 04OM31E (c), collected near the base of the cross-section. For each profile along the thin  
859 section (about 3.5 cm in height) the variation of the chromite modal content is represented as  
860 well as of the XCr, Cr, Al and Fe<sup>3+</sup> expressed in *apfu* (atom per formula unit, recalculated on  
861 the basis of O = 4), XMg and TiO<sub>2</sub> content of chromite, and of the Fo and NiO content of the  
862 surrounding matrix olivine.

863  
864 **Figure 9:** Evolution of the Fo of olivine and XMg, XCr, YFe<sup>3+</sup> and TiO<sub>2</sub> content of chromite  
865 in the samples 06OM04-3B2, 06OM04-5D, 06OM04-12E1 and 04OM31E depending on the  
866 modal content of chromite.

867  
868 **Figure 10:** Comparison of the Fo and NiO content of olivine, between the matrix from  
869 chromitites from different areas in Oman, and in dunites from the Maqsad mantle-crust  
870 transition zone (Rospabé, 2018) and mantle harzburgites from the Sumail massif (Gerbert-  
871 Gaillard, 2002; Monnier et al., 2006; Quatrevaux, 1995). a) The Fo and NiO content are much  
872 higher in olivine matrix from the stratiform chromitite than in olivine from the surrounding  
873 DTZ or from mantle harzburgites. b) In the Sumail massif the high Fo and NiO values in  
874 matrix are shared both by DTZ chromitites (Rollinson, 2019; Zagrtdenov et al., 2018) and by  
875 mantle chromitites (Ceuleneer, unpublished data). In the northern massifs even more extreme

876 values were reported in the matrix from mantle chromitites (Miura et al., 2014; Rollinson,  
877 2008, 2019, see also Augé, 1987) (no data about the olivine matrix composition in chromitites  
878 outcropping within the mantle-crust transition zone in this area). Values were averaged for  
879 each sample.

880

881 **Table 1.** Selected analyses of minerals from the matrix in chromitites and of silicate  
882 inclusions in chromites (full dataset in Supplementary Material).

883

#### 884 **References**

885 Abe, N., 2011. Petrology of podiform chromitite from the ocean floor at the 15° 20'N FZ in  
886 the MAR, Site 1271, ODP Leg 209. *J. Mineral. Petrol. Sci.* 106, 97-102.

887 doi:10.2465/jmps.101022

888 Abily, B., Ceuleneer, G., 2013. The dunitic mantle-crust transition zone in the Oman  
889 ophiolite: Residue of melt-rock interaction, cumulates from high-MgO melts, or both?

890 *Geology* 41, 67-70. doi:10.1130/G33351.1

891 Abily, B., Ceuleneer, G., Launeau, P., 2011. Synmagmatic normal faulting in the lower  
892 oceanic crust: Evidence from the Oman ophiolite. *Geology* 39, 391-394.

893 doi:10.1130/G31652.1

894 Ahmed, A.H., Arai, S., 2002. Unexpectedly high-PGE chromitite from the deeper mantle  
895 section of the northern Oman ophiolite and its tectonic implications. *Contrib. to Mineral.*

896 *Petrol.* 143, 263-278. doi:10.1007/s00410-002-0347-8

897 Amri, I., Benoit, M., Ceuleneer, G., 1996. Tectonic setting for the genesis of oceanic  
898 plagiogranites: Evidence from a paleo-spreading structure in the Oman ophiolite. *Earth*

899 *Planet. Sci. Lett.* 139, 177-194. doi: 10.1016/0012-821X(95)00233-3

900 Arai, S., 1980. Dunite-harzburgite-chromitite complexes as refractory residue in the Sangun-

- 901 Yamaguchi zone, western Japan. *J. Petrol.* 21, 141-165. doi: 10.1093/petrology/21.1.141
- 902 Arai, S., Akizawa, N., 2014. Precipitation and dissolution of chromite by hydrothermal  
903 solutions in the Oman ophiolite: New behavior of Cr and chromite. *Am. Mineral.* 99, 28-  
904 34. doi:10.2138/am.2014.4473
- 905 Arai, S., Matsukage, K., 1998. Petrology of a chromitite micropod from Hess Deep,  
906 equatorial Pacific: a comparison between abyssal and alpine-type podiform chromitites.  
907 *Lithos* 43, 1-14. doi:10.1016/S0024-4937(98)00003-6
- 908 Arai, S., Yurimoto, H., 1994. Podiform chromitites of the Tari-Misaka ultramafic complex,  
909 southwestern Japan, as mantle-melt interaction products. *Econ. Geol.* 89, 1279-1288.  
910 doi: 10.2113/gsecongeo.89.6.1279
- 911 Benoit, M., Ceuleneer, G., Polvé, M., 1999. The remelting of hydrothermally altered  
912 peridotite at mid-ocean ridges by intruding mantle diapirs. *Nature* 402, 514-518.  
913 doi:10.1038/990073
- 914 Benoit, M., Polvé, M., Ceuleneer, G., 1996. Trace element and isotopic characterization of  
915 mafic cumulates in a fossil mantle diapir (Oman ophiolite). *Chem. Geol.* 134, 199-214.  
916 doi:10.1016/S0009-2541(96)00087-3
- 917 Borisova, A.Y., Ceuleneer, G., Kamenetsky, V.S., Arai, S., Béjina, F., Abily, B., Bindeman,  
918 I.N., Polvé, M., De Parseval, P., Aigouy, T., Pokrovski, G.S., 2012. A new view on the  
919 petrogenesis of the Oman ophiolite chromitites from microanalyses of chromite-hosted  
920 inclusions. *J. Petrol.* 53, 2411-2440. doi:10.1093/petrology/egs054
- 921 Boudier, F., Nicolas, A., 1995. Nature of the moho transition zone in the Oman ophiolite. *J.*  
922 *Petrol.* 36, 777-796. doi:10.1093/petrology/36.3.777
- 923 Bowen, N.L., 1928, *The evolution of the igneous rocks.* Princeton, N. J., Princeton Univ.  
924 Press, 332 p. (reprinted 1956 by Dover Pubs., Inc., N. Y.)
- 925 Bussolesi M., Grieco G., Tzamos E., 2019. Olivine-spinel diffusivity patterns in chromitites

- 926 and dunites from the Finero phlogopite-peridotite (Ivrea-Verbano Zone, Southern Alps):  
927 implications for the thermal history of the massif. *Minerals* 9, 75.  
928 doi:10.3390/min9020075
- 929 Cameron, E.N., 1975. Postcumulus and subsolidus equilibration of chromite and coexisting  
930 silicates in the Eastern Bushveld Complex. *Geochim. Cosmochim. Acta* 39, 1021-1033.  
931 doi:10.1016/0016-7037(75)90044-7
- 932 Cassard, D., Nicolas, A., Rabinovitch, M., Moutte, J., Leblanc, M., Prinzhofer, A., 1981.  
933 Structural classification of chromite pods in southern New Caledonia. *Econ. Geol.* 76,  
934 805-831. doi:10.2113/gsecongeo.76.4.805
- 935 Ceuleneer, G., 1991. Evidence for a Paleo-Spreading Center in the Oman Ophiolite: Mantle  
936 Structures in the Maqsad Area. Springer Netherlands, pp. 147-173. doi:10.1007/978-94-  
937 011-3358-6\_9
- 938 Ceuleneer, G., Monnereau, M., Amri, I., 1996. Thermal structure of a fossil mantle diapir  
939 inferred from the distribution of mafic cumulates. *Nature* 379, 149-153.  
940 doi:10.1038/379149a0
- 941 Ceuleneer, G., Nicolas, A., 1985. Structures in podiform chromite from the Maqsad district  
942 (Sumail ophiolite, Oman). *Miner. Depos.* 20, 177-184. doi:10.1007/BF00204562
- 943 Ceuleneer, G., Nicolas, A., Boudier, F., 1988. Mantle flow pattern at an oceanic spreading  
944 centre: the Oman peridotite record. *Tectonophysics* 151, 1-26. doi:10.1016/0040-  
945 1951(88)90238-7
- 946 Clénet H., Ceuleneer G., Pinet P., Abily B., Daydou Y., Harris E., Amri I. and Dantas C.,  
947 2010. Thick sections of layered ultramafic cumulates in the Oman ophiolite revealed by  
948 an airborne hyperspectral survey: petrogenesis and relationship to mantle diapirism.  
949 *Lithos*, 114, 265-281. doi:10.1016/j.lithos.2009.09.002
- 950 Coleman, R.G., 1981. Tectonic setting for ophiolite obduction in Oman. *J. Geophys. Res.*

- 951 Solid Earth 86, 2497-2508. doi:10.1029/JB086iB04p02497
- 952 Evans, B.W., Frost, B.R., 1975. Chrome-spinel in progressive metamorphism - a preliminary  
953 analysis. *Geochim. Cosmochim. Acta* 39, 959-972. doi: 10.1016/B978-0-08-019954-  
954 2.50020-2
- 955 Fabriès, J., 1979. Spinel-olivine geothermometry in peridotites from ultramafic complexes.  
956 *Contrib. to Mineral. Petrol.* 69, 329-336. doi:10.1007/BF00372258
- 957 Gauthier, M., Corrivaux, L., Trottier, L.J., Cabri, J., Laflamme, J.G., Bergeron, M., 1990.  
958 Chromitites platinifères des complexes ophiolitiques de l'Estrie-Beauce, Appalaches du  
959 Sud du Québec. *Miner. Depos.* 25, 169-178. doi:10.1007/BF00190378
- 960 Gerbert-Gaillard, L., 2002. Caractérisation géochimique des péridotites de l'ophiolite  
961 d'Oman: processus magmatiques aux limites lithosphère/asthénosphère. PhD thesis,  
962 Université Montpellier II - Sciences et Techniques du Languedoc.
- 963 Glennie, K.W., Boeuf, M.G.A., Clarke, M.H., Moody-Stuart, M., Pilaar, W.F.H., Reinhardt,  
964 B.M., 1973. Late Cretaceous nappes in Oman Mountains and their geologic evolution.  
965 *Am. Assoc. Pet. Geol. Bull.* 57, 5-27.
- 966 González-Jiménez, J.M., Griffin, W.L., Proenza, J.A., Gervilla, F., O'Reilly, S.Y., Akbulut,  
967 M., Pearson, N.J., Arai, S., 2014. Chromitites in ophiolites: how, where, when, why?  
968 Part II. The crystallization of chromitites. *Lithos*, 189, 140-158.
- 969 Grieco, G., Bussolesi, M., Tzamos, E., Rassios, A.E., Kapsiotis, A., 2018. Processes of  
970 primary and re-equilibration mineralization affecting chromitite ore geochemistry within  
971 the Vourinos ultramafic sequence, Vourinos ophiolite (West Macedonia, Greece). *Ore*  
972 *Geol. Rev.* 95, 537-551. doi:10.1016/j.oregeorev.2018.03.009
- 973 Irvine, T.N., 1977a. Origin of chromitite layers in the Muskox intrusion and other stratiform  
974 intrusions: A new interpretation. *Geology* 5, 273-277. doi:10.1130/0091-  
975 7613(1977)5<273:OOCLIT>2.0.CO;2

- 976 Irvine, T.N., 1977b. Chromite crystallization in the join  $Mg_2SiO_4$ - $CaMgSi_2O_6$ - $CaAl_2Si_2O_8$ -  
977  $MgCr_2O_4$ - $SiO_2$ . Carnegie Institute. Washington Yearbook 76, 465-472.
- 978 Irvine, T.N., 1975. Crystallization sequences in the MuskoX intrusion and other layered  
979 intrusions-II. Origin of chromitite layers and similar deposits of other magmatic ores.  
980 *Geochim. Cosmochim. Acta* 39, 991-1020. doi:10.1016/0016-7037(75)90043-5
- 981 Jackson E.D., 1961. Primary textures and mineral associations in the ultramafic zone of the  
982 Stillwater complex, Montana. U.S. Geol. Surv. Prof. Paper 358, 106 pp.
- 983 Johan, Z., Dunlop, H., Le Bel, L., Robert, J.L., Volfinger, M., 1983. Origin of chromite  
984 deposits in ophiolitic complexes: evidence for a volatile- and sodium-rich reducing fluid  
985 phase. *Fortschr. Mineral.* 61, 105-107.
- 986 Johan, Z., Martin, R.F., Ettler, V., 2017. Fluids are bound to be involved in the formation of  
987 ophiolitic chromite deposits. *Eur. J. Mineral.* 29, 543-555. doi:10.1127/ejm/2017/0029-  
988 2648
- 989 Koepke, J., Berndt, J., Horn, I., Fahle, J., Wolff, P.E., 2014. Partial melting of oceanic gabbro  
990 triggered by migrating water-rich fluids: A prime example from the Oman Ophiolite.  
991 *Geol. Soc. Lond.* 392, 195-212 (Spec. Publ.). doi:10.1144/SP392.10
- 992 Koepke, J., Feig, S.T., Snow, J., 2005. Hydrous partial melting within the lower oceanic crust.  
993 *Terra Nova* 17, 286-291. doi:10.1111/j.1365-3121.2005.00613.x
- 994 Koga, K.T., Kelemen, P.B., Shimizu, N., 2001. Petrogenesis of the crust-mantle transition  
995 zone and the origin of lower crustal wehrlite in the Oman ophiolite. *Geochemistry,*  
996 *Geophys. Geosystems* 2. doi:10.1029/2000GC000132
- 997 Lago B., Rabinowicz M., Nicolas A., 1982. Podiform chromite ore bodies - a genetic model.  
998 *J. Petrol.* 23, 103-125. doi:10.1093/petrology/23.1.103
- 999 Latypov, R., Costin, G., Chistyakova, S., Hunt, E. J., Mukherjee, R., Naldrett, T., 2018.  
1000 Platinum-bearing chromite layers are caused by pressure reduction during magma ascent.

- 1001 Nat. com. 9, 462. doi:10.1038/s41467-017-02773-w
- 1002 Leblanc, M., Ceuleneer, G., 1991. Chromite crystallization in a multicellular magma flow:  
1003 Evidence from a chromitite dike in the Oman ophiolite. *Lithos* 27, 231-257.  
1004 doi:10.1016/0024-4937(91)90002-3
- 1005 Lehmann, J., 1983. Diffusion between olivine and spinel: application to geothermometry.  
1006 *Earth Planet. Sci. Lett.* 64, 123-138. doi:10.1016/0012-821X(83)90057-2
- 1007 Lipin, B. R., 1993. Pressure increases, the formation of chromite seams, and the development  
1008 of the ultramafic series in the Stillwater Complex, Montana. *J. Petrol.* 34, 955-976.  
1009 doi:10.1093/petrology/34.5.955.
- 1010 Lombaard, B.V., 1956. Chromite and dunite of the Bushveld Complex. *South Afr. J. Geol.*  
1011 59, 59-74.
- 1012 Lorand J.-P. and Ceuleneer G., 1989. Silicate and base-metal sulfide inclusions in chromites  
1013 from the Maqsad area (Oman ophiolite, Gulf of Oman): a model for entrapment. *Lithos*,  
1014 22, 173-190.
- 1015 Matsukage, K., Arai, S., 1998. Jadeite, albite and nepheline as inclusions in spinel of  
1016 chromitite from Hess Deep, equatorial Pacific: their genesis and implications for  
1017 serpentinite diapir formation. *Contrib. to Mineral. Petrol.* 131, 111-122.  
1018 doi:10.1007/s004100050382
- 1019 Mathez, E.A., Kinzler, R.J., 2017. Metasomatic chromitite seams in the Bushveld and Rum  
1020 Layered intrusions. *Elements*, 13, 397-402.
- 1021 Matveev, S., Ballhaus, C., 2002. Role of water in the origin of podiform chromitite deposits.  
1022 *Earth Planet. Sci. Lett.* 203, 235-243. doi:10.1016/S0012-821X(02)00860-9
- 1023 McElduff, B., Stumpfl, E.F., 1991. The chromite deposits of the Troodos complex, Cyprus-  
1024 evidence for the role of a fluid phase accompanying chromite formation. *Miner. Depos.*  
1025 26, 307-318. doi:10.1007/BF00191079

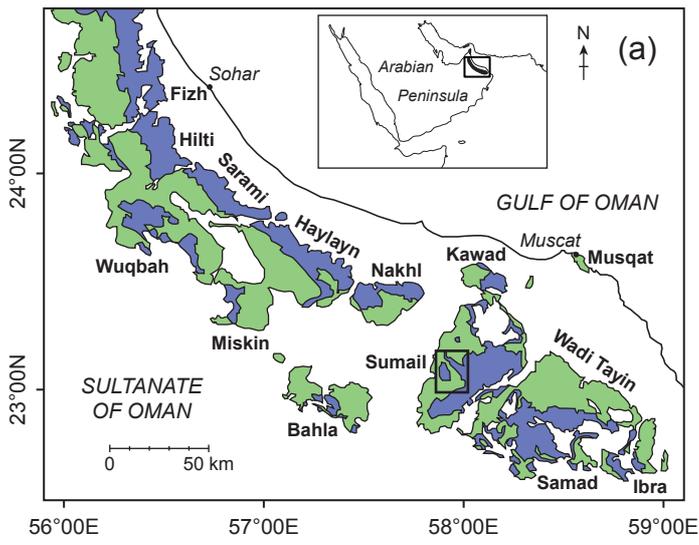
- 1026 Miura, M., Arai, S., Tamura, A., 2014. Formation of discordant chromitite at the initiation of  
1027 sub-arc mantle processes: Observations from the northern Oman ophiolite. *J. Mineral.*  
1028 *Petrol. Sci.* 109, 38-43. doi:10.2465/jmps.131006
- 1029 Mondal, S. K., Mathez, E. A., 2006. Origin of the UG2 chromitite layer, Bushveld Complex.  
1030 *J. Petrol.* 48, 495-510. doi:10.1093/petrology/egl069
- 1031 Monnier, C., Girardeau, J., Le Mée, L., Polvé, M., 2006. Along-ridge petrological  
1032 segmentation of the mantle in the Oman ophiolite. *Geochemistry, Geophys. Geosystems.*  
1033 doi:10.1029/2006GC001320
- 1034 Montigny, R., Le Mer, O., Thuizat, R., Whitechurch, H., 1988. K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  study of  
1035 metamorphic rocks associated with the Oman ophiolite: Tectonic implications.  
1036 *Tectonophysics* 151, 345-362. doi:10.1016/0040-1951(88)90252-1
- 1037 Naldrett, A. J., Wilson, A., Kinnaird, J., Yudovskaya, M., Chunnnett, G., 2012. The origin of  
1038 chromitites and related PGE mineralization in the Bushveld Complex: new mineralogical  
1039 and petrological constraints. *Mineralium Deposita* 47, 209-232. doi:10.1007/s00126-  
1040 011-0366-3
- 1041 Nonnotte, P., Ceuleneer, G., Benoit, M., 2005. Genesis of andesitic-boninitic magmas at mid-  
1042 ocean ridges by melting of hydrated peridotites: Geochemical evidence from DSDP Site  
1043 334 gabbro-norites. *Earth Planet. Sci. Lett.* 236, 632-653. doi:10.1016/j.epsl.2005.05.026
- 1044 O'Neill, H.S.C., Wall, V.J., 1987. The olivine - orthopyroxene - spinel oxygen geobarometer,  
1045 the nickel precipitation curve, and the oxygen fugacity of the earth's upper mantle. *J.*  
1046 *Petrol.* 28, 1169-1191. doi:10.1093/petrology/28.6.1169
- 1047 Payot, B.D., Arai, S., Dick, H.J., Abe, N., Ichiyama, Y., 2014. Podiform chromitite formation  
1048 in a low-Cr/high-Al system: An example from the Southwest Indian Ridge (SWIR).  
1049 *Mineral. Petrol.* 108, 533-549. doi:10.1007/s00710-0134-0317-z
- 1050 Pearce, J.A., Alabaster, T., Shelton, A.W., Searle, M.P., 1981. The Oman ophiolite as a

- 1051 Cretaceous arc-basin complex: evidence and implications. *Philos. Trans. R. Soc. London*  
1052 300, 299-317. doi:10.1098/rsta.1892.0001
- 1053 Pouchou, J.-L., Pichoir, F., 1985. "PAP" phi-rho-Z procedure for improved quantitative  
1054 microanalysis., in: Armstrong, J. (Ed.), *Microbeam Analysis*. San Francisco, CA, pp.  
1055 104-106.
- 1056 Python, M., Ceuleneer, G., 2003. Nature and distribution of dykes and related melt migration  
1057 structures in the mantle section of the Oman ophiolite. *Geochemistry, Geophys.*  
1058 *Geosystems* 4. doi:10.1029/2003GC000354
- 1059 Python, M., Ceuleneer, G., Arai, S., 2008. Chromian spinels in mafic-ultramafic mantle  
1060 dykes: Evidence for a two-stage melt production during the evolution of the Oman  
1061 ophiolite. *Lithos* 106, 137-154. doi:10.1016/j.lithos.2008.07.001
- 1062 Python, M., Ceuleneer, G., Ishida, Y., Barrat, J.-A., Arai, S., 2007. Oman diopsidites: a new  
1063 lithology diagnostic of very high temperature hydrothermal circulation in mantle  
1064 peridotite below oceanic spreading centres. *Earth Planet. Sci. Lett.* 255, 289-305.  
1065 doi:10.1016/j.epsl.2006.12.030
- 1066 Quatrevaux, F., 1995. Etude pétrologique des péridotites des massifs de Maqсад et Wuqbah,  
1067 ophiolite d'Oman. PhD thesis, Université Paris 7.
- 1068 Rabinowicz, M., Ceuleneer, G., Nicolas, A., 1987. Melt segregation and flow in mantle  
1069 diapirs below spreading centres: evidence from the Oman ophiolite. *J. Geophys. Res.*  
1070 *Solid Earth* 92, 3475-3486. doi:10.1029/JB092iB05p03475
- 1071 Rahgoshay, M., Juteau, T., Whitechurch, H., 1981. Kizilyuksek Tepe: un gisement  
1072 exceptionnel de chromite stratiforme dans un complexe ophiolitique (massif de Pozanti-  
1073 Karsanti, Taurus, Turquie). *C.R. Acad. Sc. Paris* 293, 765-770
- 1074 Rioux, M., Bowring, S., Kelemen, P., Gordon, S., Miller, R., Dudás, F., 2013. Tectonic  
1075 development of the Samail ophiolite: High-precision U-Pb zircon geochronology and

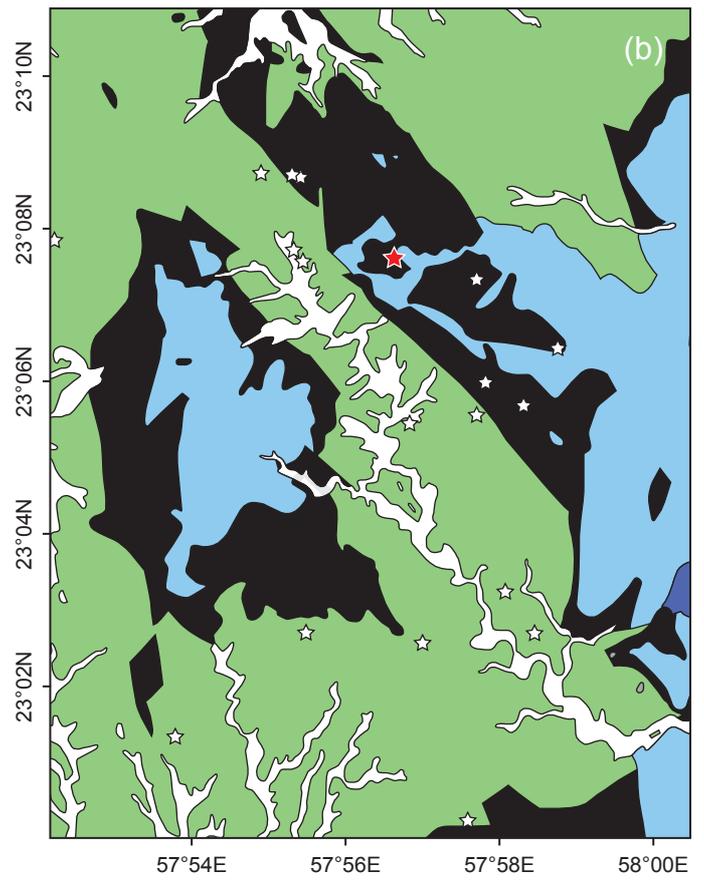
- 1076 Sm-Nd isotopic constraints on crustal growth and emplacement. *J. Geophys. Res. Solid*  
1077 *Earth* 118, 2085-2101. doi:10.1002/jgrb.50139
- 1078 Rioux, M., Garber, J., Bauer, A., Bowring, S., Searle, M., Kelemen, P., Hacker, B., 2016.  
1079 Synchronous formation of the metamorphic sole and igneous crust of the Semail  
1080 ophiolite: New constraints on the tectonic evolution during ophiolite formation from  
1081 high-precision U-Pb zircon geochronology. *Earth Planet. Sci. Lett.* 451, 185-195.  
1082 doi:10.1016/j.epsl.2016.06.051
- 1083 Rollinson, H., 2019. Dunites in the mantle section of the Oman ophiolite-the boninite  
1084 connection. *Lithos*. doi:10.1016/j.lithos.2019.03.008
- 1085 Rollinson, H., 2008. The geochemistry of mantle chromitites from the northern part of the  
1086 Oman ophiolite: Inferred parental melt compositions. *Contrib. to Mineral. Petrol.* 156,  
1087 273-288. doi:10.1007/s00410-008-0284-2
- 1088 Rollinson, H., Adetunji, J., 2015. Chromite in the Mantle Section of the Oman Ophiolite:  
1089 Implications for the Tectonic Evolution of the Oman Ophiolite. *Acta Geol. Sin.*  
1090 doi:10.1111/1755-6724.12308\_44
- 1091 Rollinson, H., Adetunji, J., 2013. Mantle podiform chromitites do not form beneath mid-  
1092 ocean ridges: A case study from the Moho transition zone of the Oman ophiolite. *Lithos*  
1093 177, 314-327. doi:10.1016/j.lithos.2013.07.004
- 1094 Rollinson, H., Mameri, L., Barry, T., 2018. Polymineralic inclusions in mantle chromitites  
1095 from the Oman ophiolite indicate a highly magnesian parental melt. *Lithos* 310, 381-391.  
1096 doi:10.1016/j.lithos.2018.04.024
- 1097 Rospabé, M., 2018. Etude pétrologique, géochimique et structurale de la zone de transition  
1098 dunitique dans l'ophiolite d'Oman: Identification des processus pétrogénétiques à  
1099 l'interface manteau/croûte. PhD thesis, Université Paul Sabatier, Toulouse III.
- 1100 Rospabé, M., Benoit, M., Ceuleneer, G., Hodel, F., Kaczmarek, M.-A., 2018. Extreme

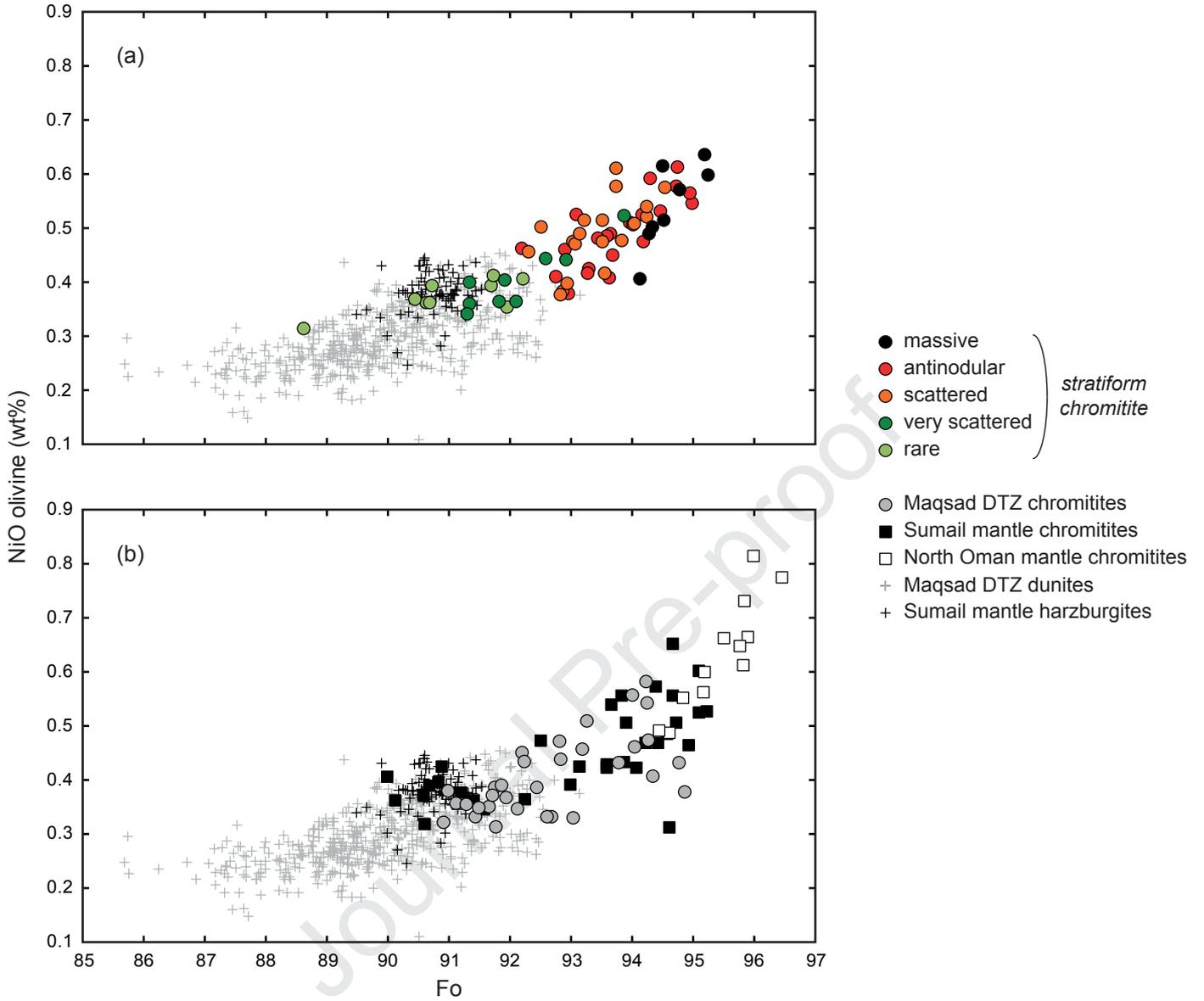
- 1101 geochemical variability through the dunitic transition zone of the Oman ophiolite:  
1102 Implications for melt/fluid-rock reactions at Moho level beneath oceanic spreading  
1103 centers. *Geochim. Cosmochim. Acta* 234, 1-23. doi:10.1016/j.gca.2018.05.012
- 1104 Rospabé, M., Benoit, M., Ceuleneer, G., Kaczmarek, M.-A., Hodel, F., 2019. Melt  
1105 hybridization and metasomatism triggered by syn-magmatic faults within the Oman  
1106 ophiolite: A clue to understand the genesis of the dunitic mantle-crust transition zone.  
1107 *Earth Planet. Sci. Lett.* 516, 108-121. doi:10.1016/j.epsl.2019.04.004
- 1108 Rospabé, M., Ceuleneer, G., Benoit, M., Abily, B., Pinet, P., 2017. Origin of the dunitic  
1109 mantle-crust transition zone in the Oman ophiolite: The interplay between percolating  
1110 magmas and hightemperature hydrous fluids. *Geology* 45, 471-474.  
1111 doi:10.1130/G38778.1
- 1112 Schiano, P., Clocchiatti, R., Lorand, J. P., Massare, D., Deloule, E., Chaussidon, M., 1997.  
1113 Primitive basaltic melts included in podiform chromites from the Oman Ophiolite. *Earth*  
1114 *Planet. Sci. Lett.* 146, 489-497. doi:10.1016/S0012-821X(96)00254-3
- 1115 Spandler, C., Mavrogenes, J., Arculus, R., 2005. Origin of chromitites in layered intrusions:  
1116 Evidence from chromite-hosted melt inclusions from the Stillwater Complex. *Geology*  
1117 33, 893-896. doi:10.1130/G21912.1
- 1118 Stoll, W.C., 1958. Geology and Petrology of the Masinloc chromite deposit Zambales, Luzon,  
1119 Philippine Islands. *Geol. Soc. Am. Bull.* 69, 419-448. doi:10.1130/0016-  
1120 7606(1958)69[419:GAPOTM]2.0.CO;2
- 1121 Talkington, R.W., Watkinson, D.H., Whittaker, P.J., Jones, P.C., 1984. Platinum-group  
1122 minerals and other solid inclusions in chromite of ophiolitic complexes: occurrence and  
1123 petrological significance. *Tschermaks Mineral. Petr. Mitt.* 32, 285-301.  
1124 doi:10.1007/BF01081619
- 1125 Tamura, A., Morishita, T., Ishimaru, S., Arai, S., 2014. Geochemistry of spinel-hosted

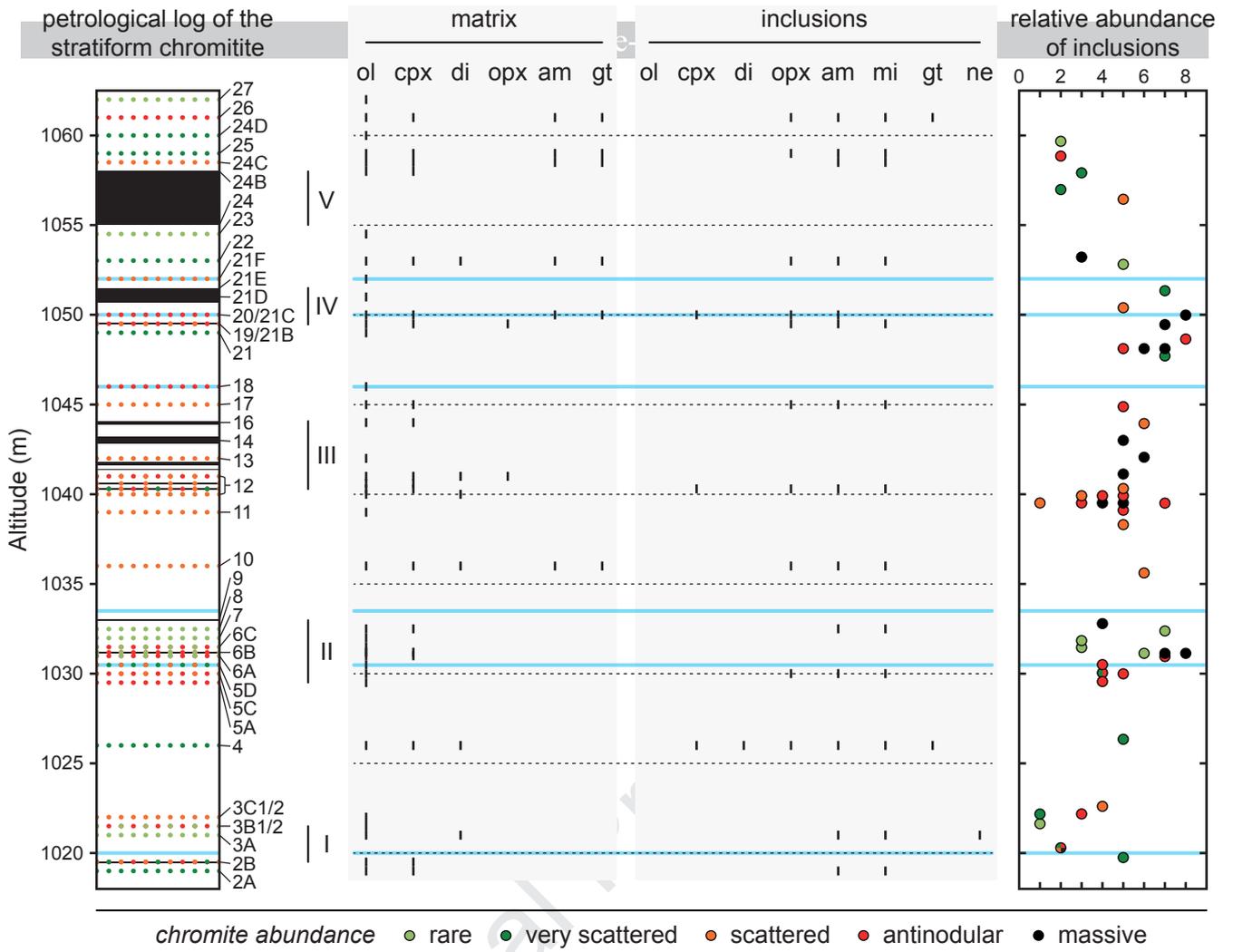
- 1126 amphibole inclusions in abyssal peridotite: insight into secondary melt formation in  
1127 melt–peridotite reaction. *Contrib. to Mineral. Petrol.* 167, 974. doi:10.1007/s00410-014-  
1128 0974-x
- 1129 Thayer, T.P., 1960. Some critical differences between alpine type and stratiform peridotite-  
1130 gabbro complexes. *Int. Geol. Congr.* 21th 13, 247-259.
- 1131 Tippit, P.R., Pessagno, E.A., Smewing, J.D., 1981. The Biostratigraphy of Sediments in the  
1132 Volcanic Unit of the Samail Ophiolite. *J. Geophys. Res. Solid Earth* 86, 2756-2762.  
1133 doi:10.1029/JB086iB04p02756
- 1134 Warren, C.J., Parrish, R.R., Waters, D.J., Searle, M.P., 2005. Dating the geologic history of  
1135 Oman’s Semail ophiolite: Insights from U-Pb geochronology. *Contrib. to Mineral.*  
1136 *Petrol.* 150, 403-422. doi:10.1007/s00410-005-0028-5
- 1137 Zagrtdenov, N.R., Ceuleneer, G., Rospabé, M., Borisova, A.Y., Toplis, M.J., Benoit, M.,  
1138 Abily, B., 2018. Anatomy of a chromitite dyke in the mantle/crust transition zone of the  
1139 Oman ophiolite. *Lithos* 312-313. doi:10.1016/j.lithos.2018.05.012
- 1140 Zhou, M.F., Robinson, P.T., Bai, W.J., 1994. Formation of podiform chromitites by  
1141 melt/rock interaction in the upper mantle. *Miner. Depos.* 29, 98-101.  
1142 doi:10.1007/BF03326400

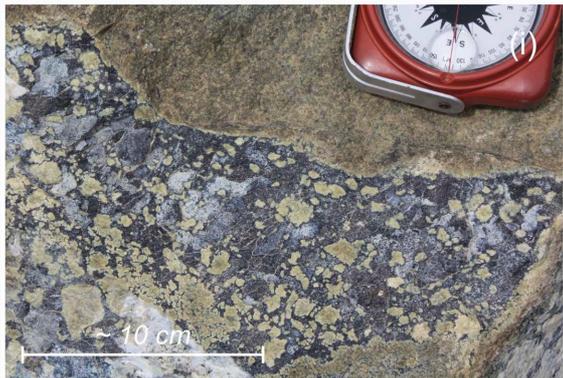
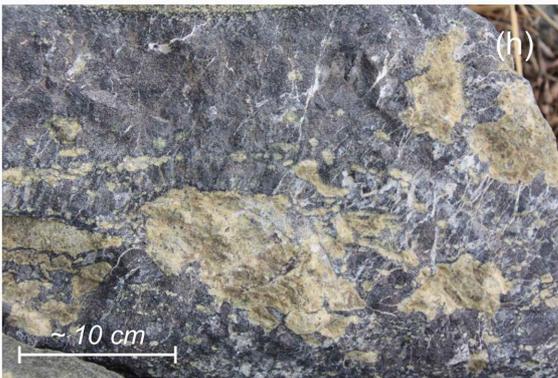
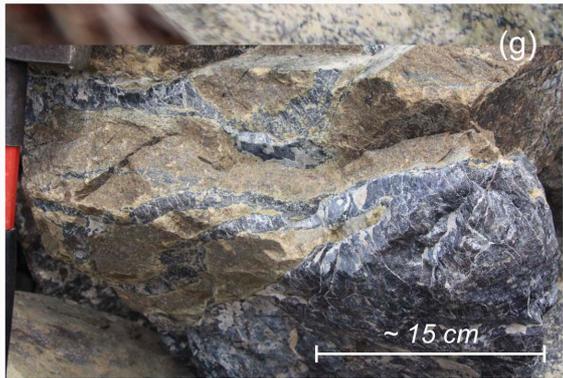


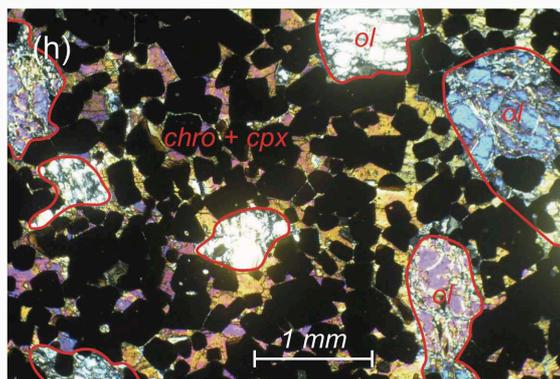
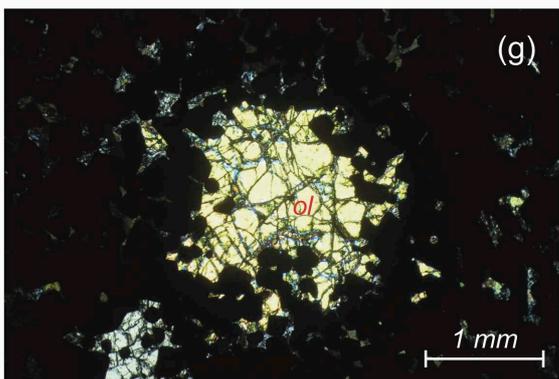
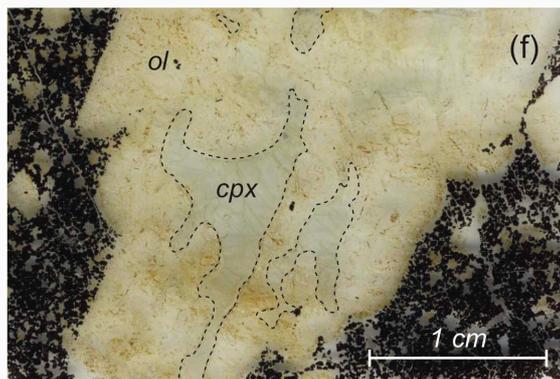
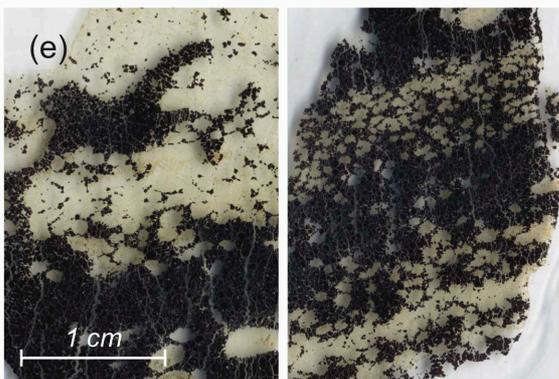
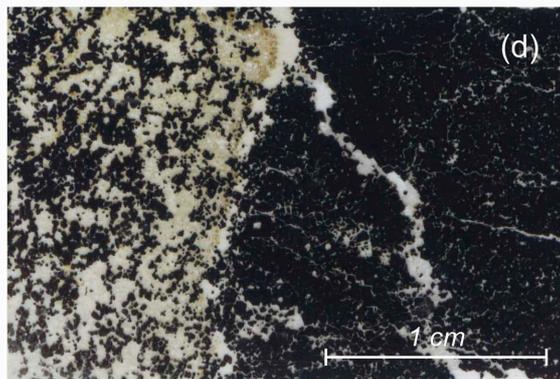
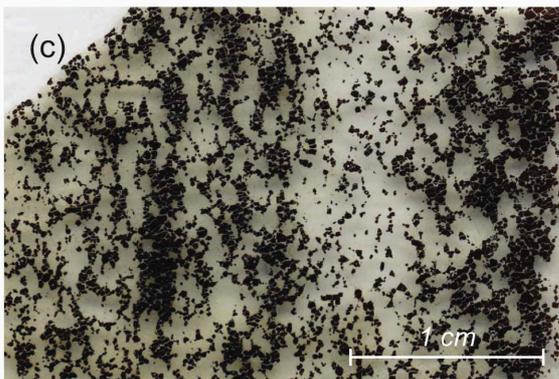
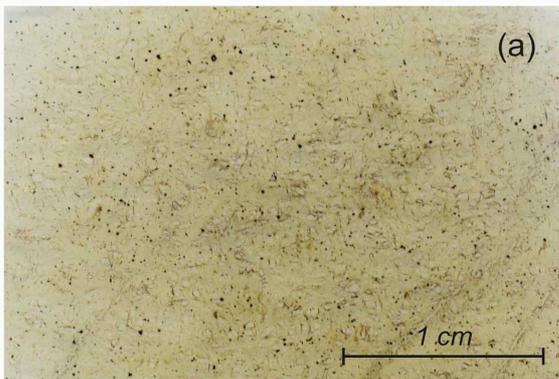
- Crustal units
- Lower crustal layered cumulates
- Dunitic transition zone (DTZ)
- Mantle harzburgites
- ★ Stratiform chromitite (this study)
- ☆ Other chromitite occurrences

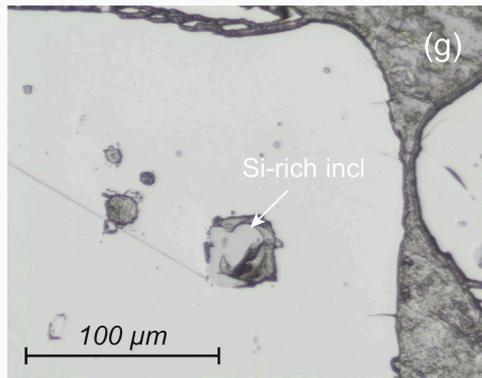
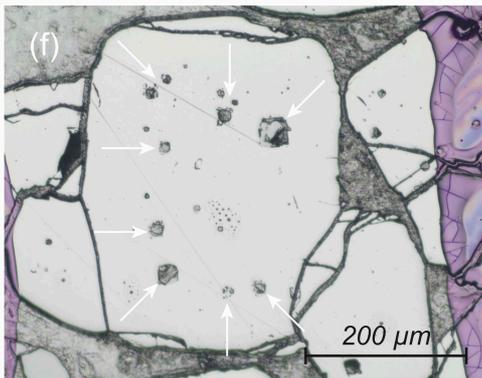
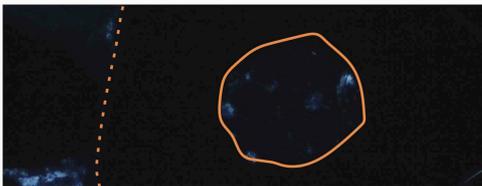
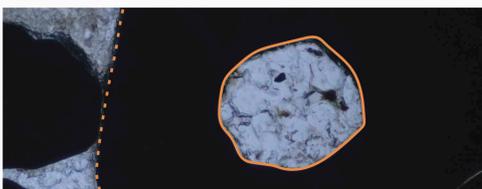
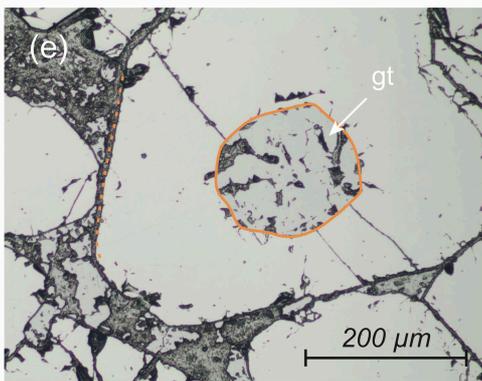
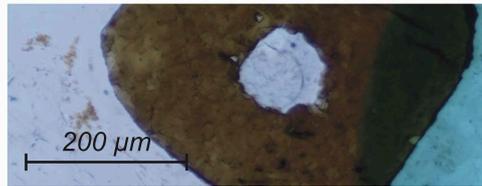
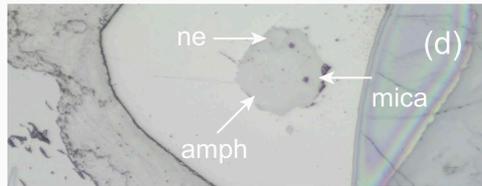
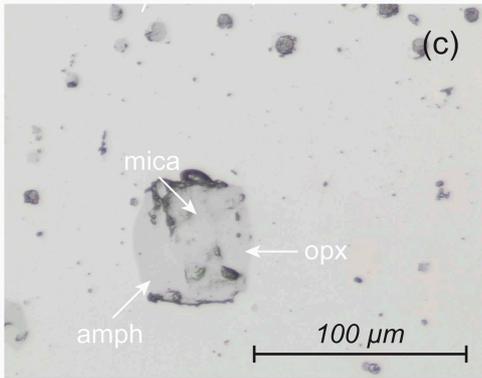
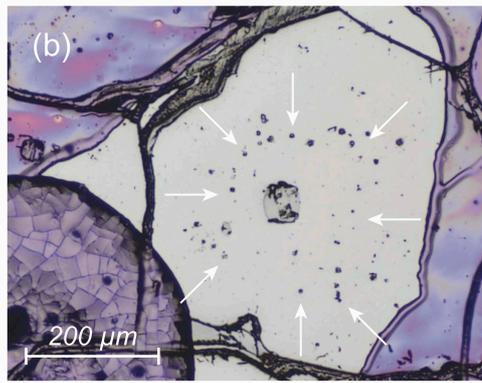
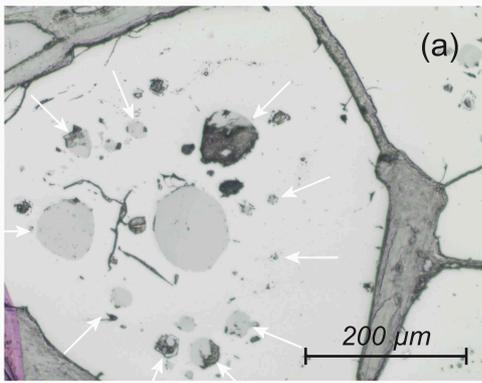


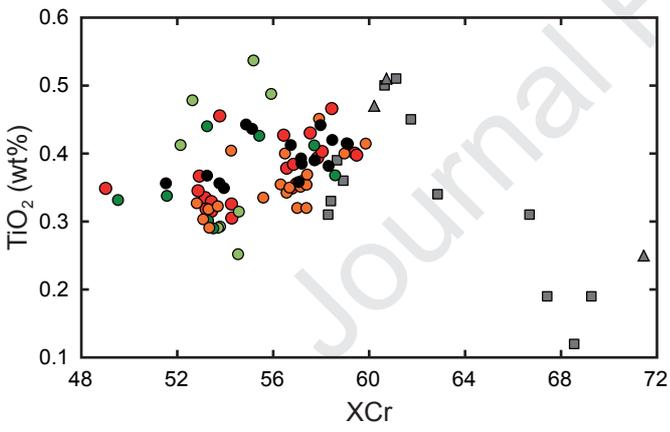
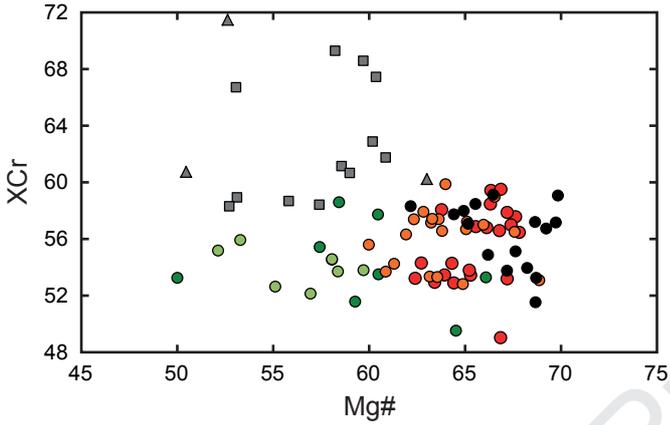
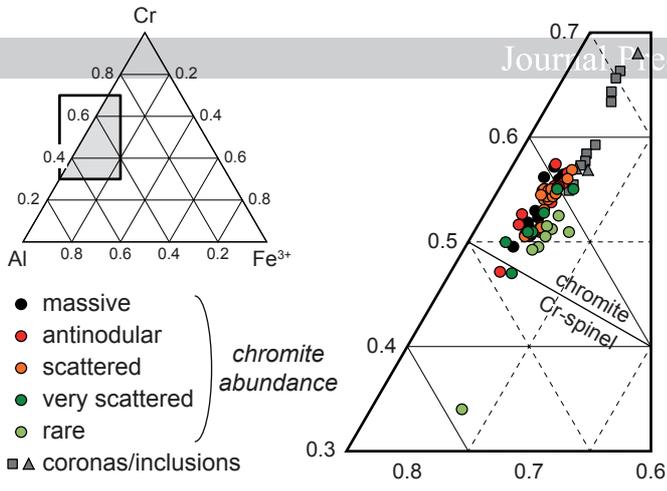


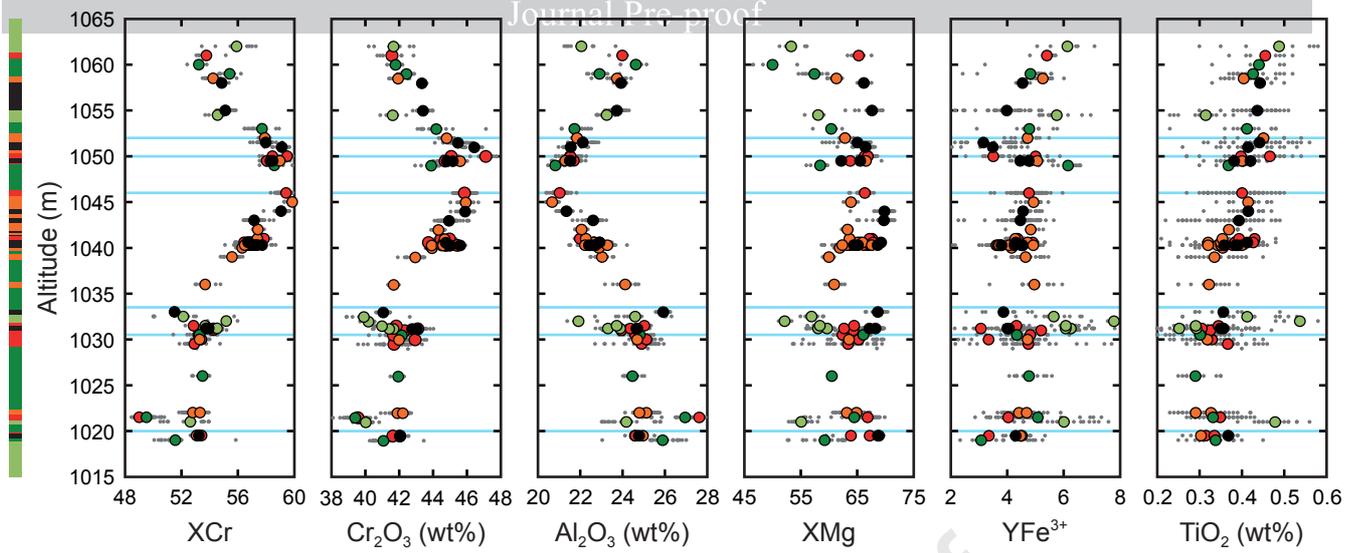




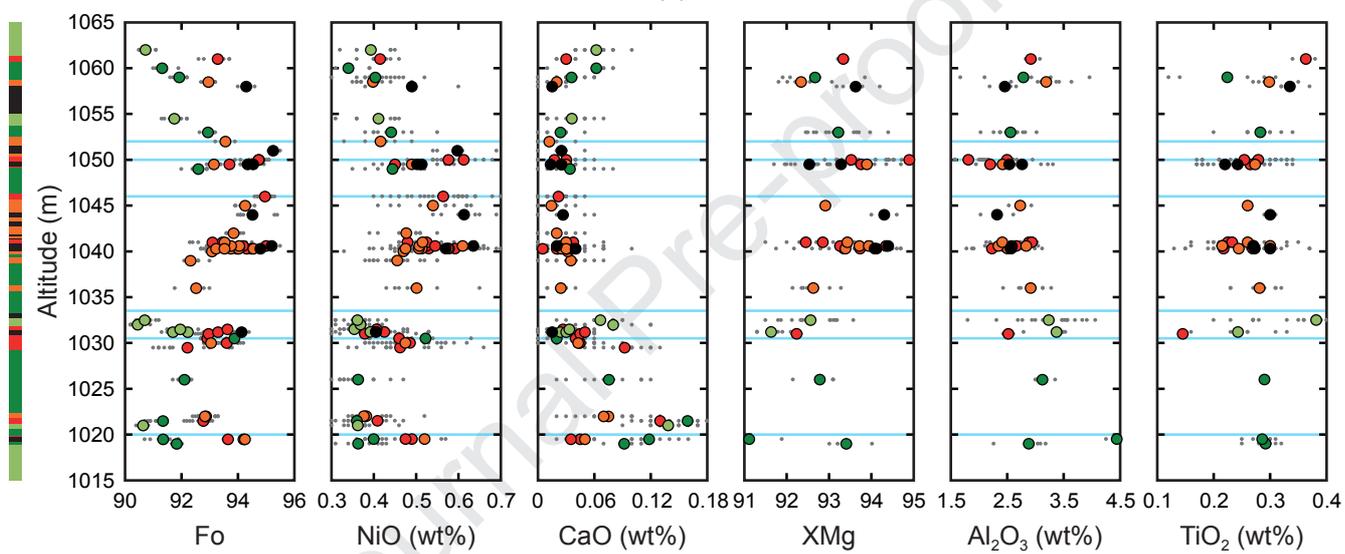






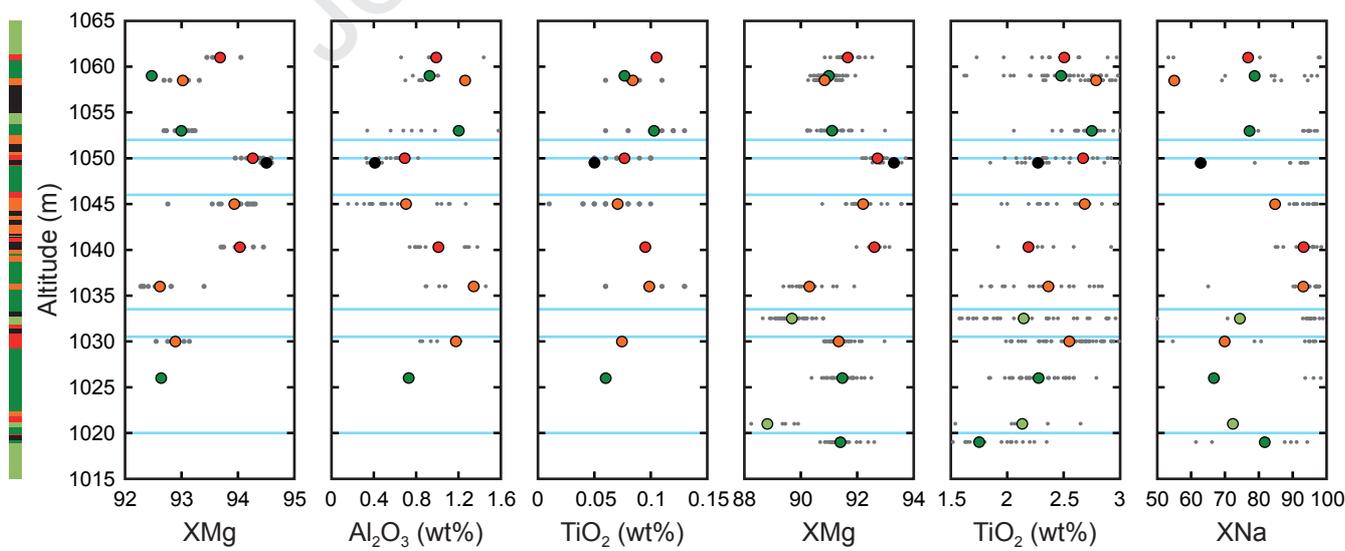


(a) chromite



(b) olivine matrix

(c) clinopyroxene matrix



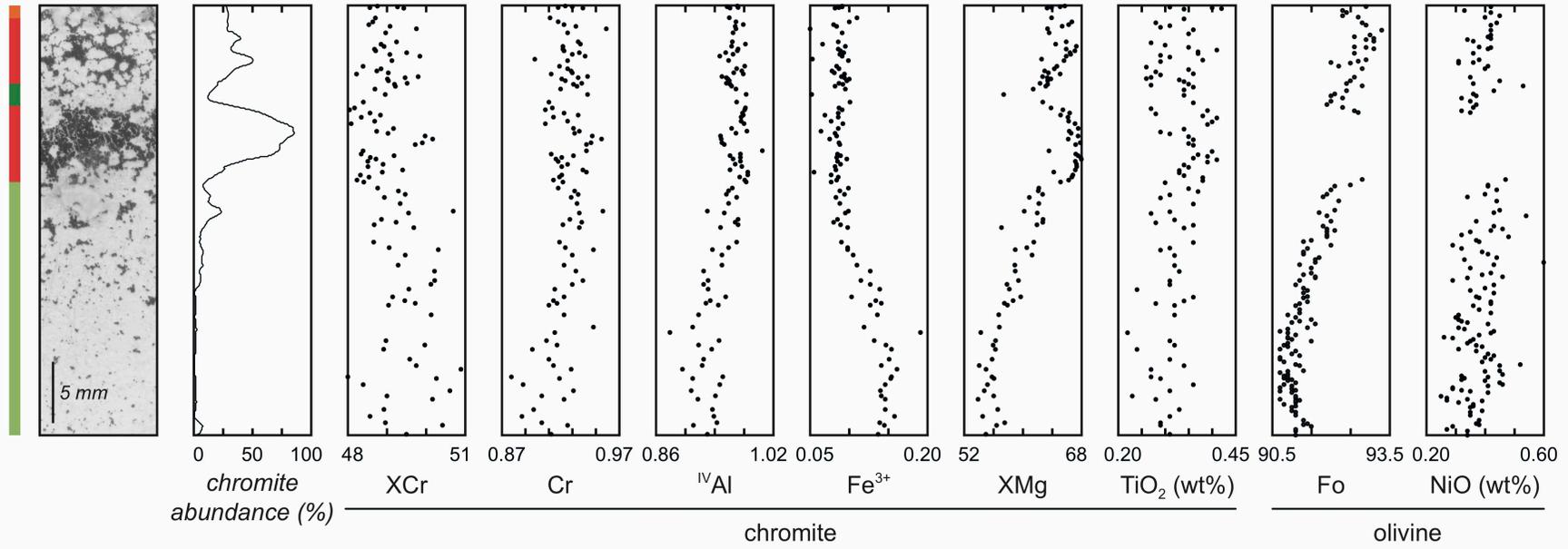
(d) orthopyroxene inclusions

(e) amphibole inclusions

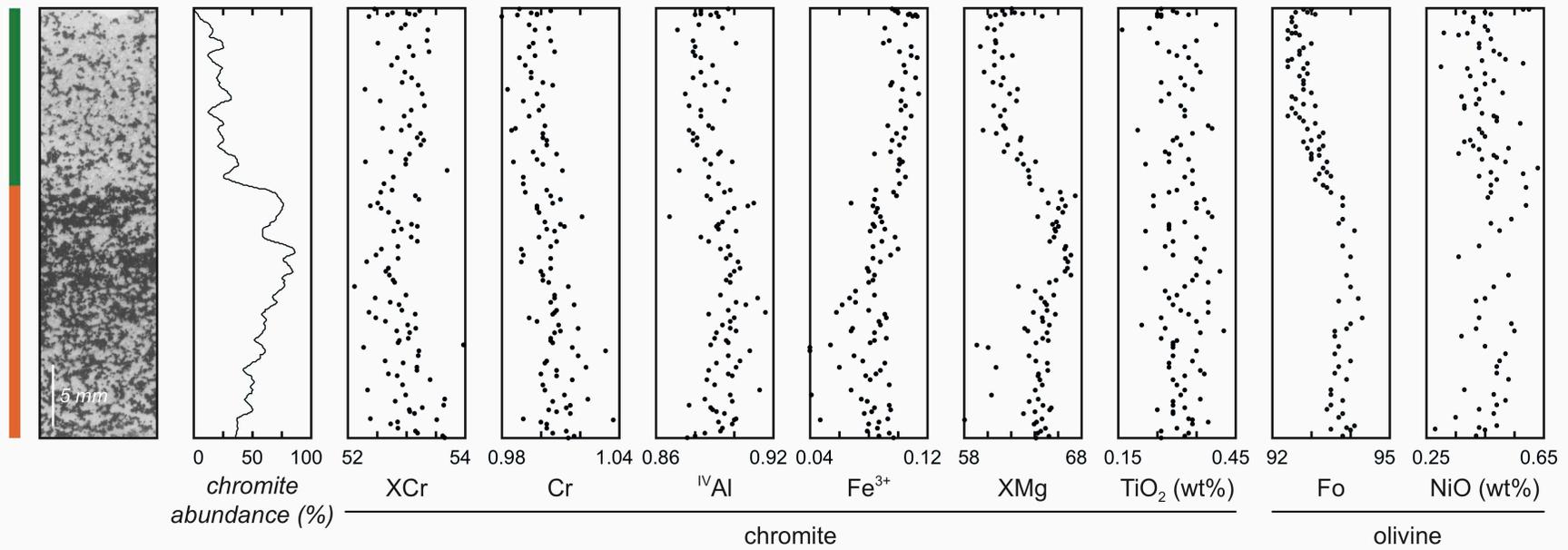
(f) mica inclusions

*chromite abundance* ● rare ● very scattered ● scattered ● antinodular ● massive

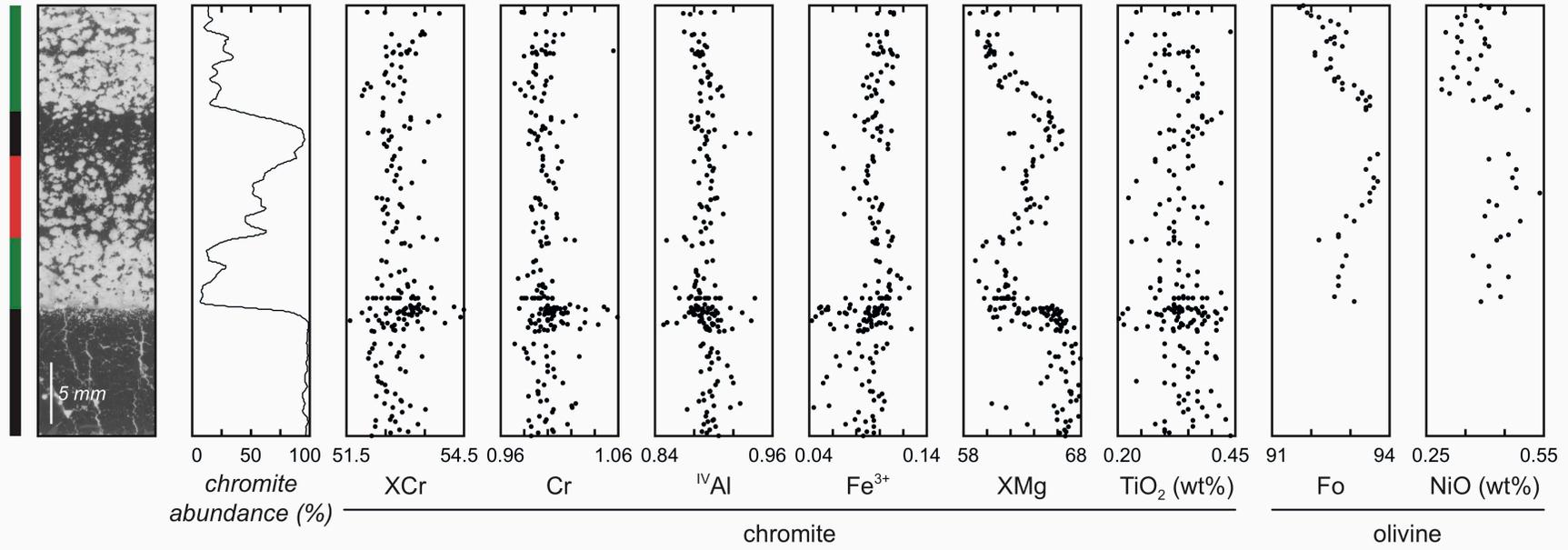
(a) 06OM04-3B2



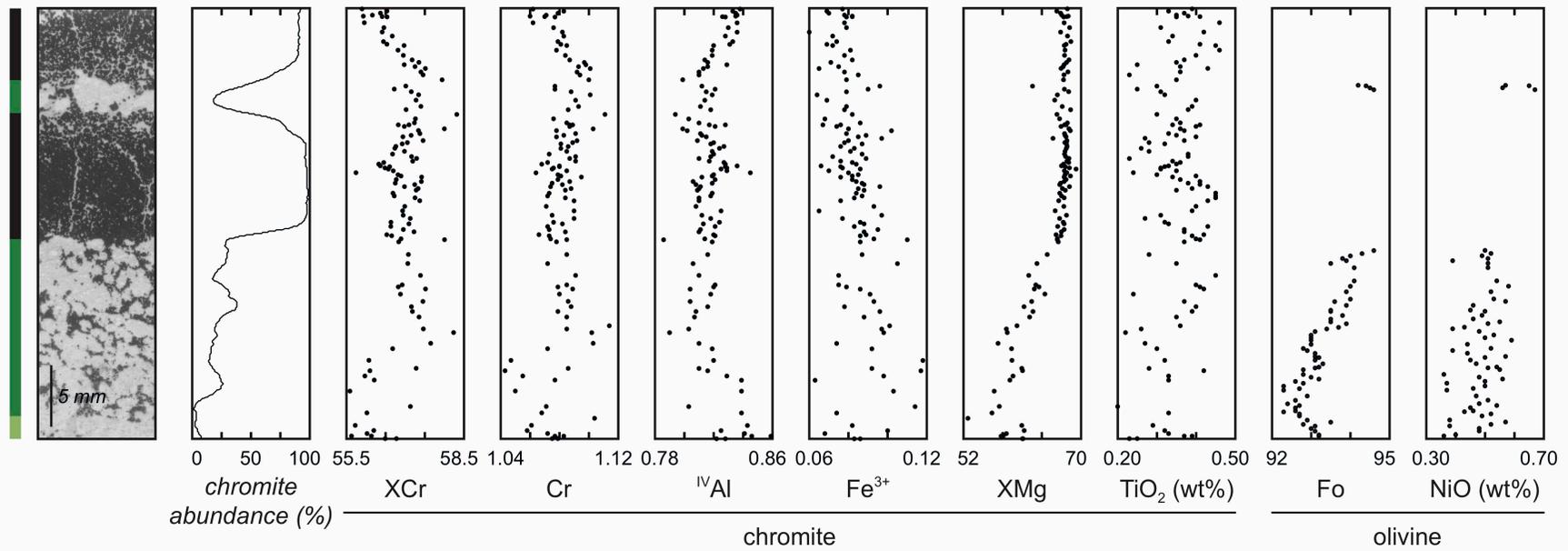
(b) 06OM04-5D

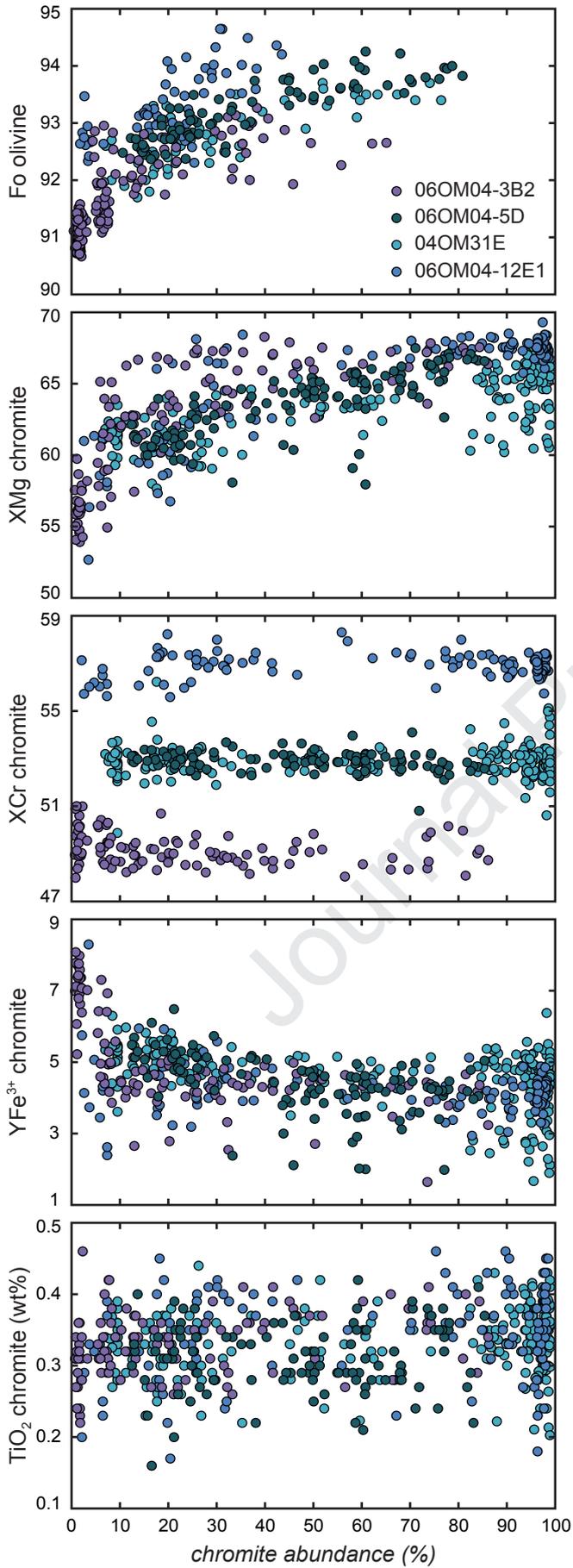


(c) 04OM31E



(d) 06OM04-12E1





1 **Multi-scale development of a stratiform chromite ore body at the base of**  
2 **the dunitic mantle-crust transition zone (Maqsad diapir, Oman ophiolite):**  
3 **the role of repeated melt and fluid influxes.**

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18  
19 **Highlights**

- 20 - The Maqsad dunitic transition zone (Oman ophiolite) hosts a stratiform chromitite  
21 - The matrix (ol, cpx) strongly contrasts with silicate inclusions (opx, amph, mica)  
22 - Vertical chemical evolutions along the ore body correlate to magmatic breccias  
23 - Melt injections (breccias) controlled the hybridization between contrasted melts

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19 **Declaration of interest**

20 Declarations of interest: none

21 This work has never been submitted to another journal than Lithos on 27/07/19 (ms#  
22 LITHOS-S-19-00730).

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25 (Institut National des Sciences de l'Univers).

Table 1. Selected analyses of minerals from the matrix in chromitites and of silicate inclu

Sample	06OM04-2A	06OM04-1	06OM04-4	06OM04-12B2C	
Chromite abundanc	very scattered	rare	very scattered	antinodular	
Mineral mode of occurrenc	Chro matrix	OI matrix	Opx inclusion	Cpx matrix	Cpx inclusion
SiO <sub>2</sub>	0.05	40.90	57.42	52.35	52.69
TiO <sub>2</sub>	0.34		0.06	0.27	0.28
Al <sub>2</sub> O <sub>3</sub>	25.89		0.73	2.53	2.38
Cr <sub>2</sub> O <sub>3</sub>	41.09	0.02	0.97	1.34	1.46
FeO <sub>total</sub>	18.34	11.02	5.01	2.13	2.01
MnO	0.25	0.20	0.17	0.08	0.03
MgO	12.99	48.23	35.37	17.74	16.94
ZnO	0.04				
CaO		0.26	0.29	22.23	23.26
Na <sub>2</sub> O			0.03	0.47	0.54
K <sub>2</sub> O			0.01		
NiO	0.15	0.31	0.06	0.05	0.02
Sum Ox%	99.12	100.93	100.12	99.19	99.61
Fo / XMg	59.3	88.6	92.6	93.7	93.8
XCr	51.6				
YFe <sup>3+</sup>	3.11				
XNa					

## isions in chromites (full dataset in Supplementary Material).

06OM04-12E1	06OM04-20		06OM04-5C2	06OM04-10	06OM04-3A
massive / antinodular	antinodular		scattered	scattered	rare
Diop	Amph		Mica (asp.)	Garnet (gr.)	Neph
matrix	matrix	inclusion	inclusion	matrix	inclusion
54.89	45.86	44.81	40.97	39.41	46.51
0.16	1.47	3.42	4.37	0.04	0.04
0.79	11.37	10.78	16.45	21.14	32.63
0.37	1.68	2.69	2.29	1.36	0.62
0.84	2.73	2.53	2.19	0.78	0.39
0.15	0.06		0.01	0.10	
18.05	18.57	18.08	23.34	0.52	0.77
25.31	12.31	12.17	0.08	36.89	0.95
0.04	2.86	2.97	5.52	0.01	18.60
	0.14	0.07	0.57		0.17
	0.14	0.14	0.24		0.03
100.60	97.18	97.66	96.03	100.25	100.71
97.0	92.4	92.7	95.0		
	96.8	98.5	93.6		