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Mathieu Rospabé, Georges Ceuleneer, Nicolas Granier, Shoji Arai, Anastassia Yu. Borisova. Multi-scale development of a stratiform chromite ore body at the base of the dunitic mantle-crust transition zone (Maqsad diapir, Oman ophiolite): The role of repeated melt and fluid influxes. *Lithos*, 2019, 350-351, pp.105235. 10.1016/j.lithos.2019.105235 . hal-02383759

HAL Id: hal-02383759

<https://hal.science/hal-02383759>

Submitted on 2 Dec 2019

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1 **Multi-scale development of a stratiform chromite ore body at the base of**
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5 Mathieu Rospabé ^{a,b,*}, Georges Ceuleneer ^a, Nicolas Granier ^a, Shoji Arai ^c, Anastassia Y.
6 Borisova ^{a,d}

7
8 ^a Géosciences Environnement Toulouse (GET), Observatoire Midi-Pyrénées, Université de
9 Toulouse, CNRS, IRD, 14 avenue E. Belin, F-31400 Toulouse, France

10 ^b Research Institute for Marine Geodynamics (IMG), Japan Agency for Marine-Earth Science
11 and Technology (JAMSTEC), 2-15 Natsushima, Yokosuka, Kanagawa 237-0061, Japan

12 ^c Department of Sciences, College of Science and Engineering, Kanazawa University,
13 Kanazawa, Ishikawa 920-1192, Japan

14 ^d Geological Department, Lomonosov Moscow State University, Vorobievsky Gory, 119899,
15 Moscow, Russia

16
17 * Corresponding author. E-mail address: mrospabe@jamstec.go.jp

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 TiO₂ 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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9 Toulouse, CNRS, IRD, 14 avenue E. Belin, F-31400 Toulouse, France

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12 ^c Department of Sciences, College of Science and Engineering, Kanazawa University,
13 Kanazawa, Ishikawa 920-1192, Japan

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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 H₂O, increase in the total pressure or fO₂ 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 μm 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 µm 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 µm 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 TiO₂, Al₂O₃, Cr₂O₃ and MgO, 0.04 wt%
308 for SiO₂ and Na₂O, 0.05 wt% for CaO and K₂O, 0.06 wt% for V₂O₃ 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. Cr₂O₃ in chromite, MgO in olivine, CaO in clinopyroxene),
311 better than 6% for TiO₂ (chromite, clinopyroxene) and better than 30% for minor elements
312 (e.g. NiO in olivine, Na₂O 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 Fe³⁺ (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 (Cr₂O₃ = 39.4-47.1 wt%; Al₂O₃ =
319 20.7-27.6 wt%), XMg (100 × Mg/(Mg + Fe²⁺) atomic ratio) = 50.0-69.8, YFe³⁺ (100 ×
320 Fe³⁺/(Cr + Al + Fe³⁺) atomic ratio) = 1.1-7.8 (4.6 in average), and TiO₂ = 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 (Cr₂O₃ = 43.7-52.7 wt%; Al₂O₃ = 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 YFe³⁺ (4.3-
324 6.3, 5.3 in average). The YFe³⁺ 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 TiO₂ 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 TiO₂ content and negatively correlated to YFe³⁺). Otherwise, in between the three
343 highest breccias (1049-1053 m), the Cr₂O₃ content shows two successive increasing and
344 decreasing trends within a few metres while Al₂O₃ continuously increases on the same
345 interval. XMg shares basically the same evolution as XCr and is particularly well-correlated
346 to the Cr₂O₃ content, while oppositely YFe³⁺ is negatively correlated to XMg (Fig. 7a). The
347 vertical evolution of the TiO₂ 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 Cr₂O₃ 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 Al₂O₃ = 1.8-4.4 wt%, Cr₂O₃ = 1.1-1.5 wt, TiO₂ = 0.15-0.38 wt% and Na₂O
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 Al₂O₃ and
375 TiO₂ contents of clinopyroxene, as well as of Na₂O (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 SiO₂ content (53.8-55.0 wt% instead of 51.8-53.6 wt%) coupled with
379 lower Al₂O₃ (0.47-1.4 wt%), TiO₂ (0.03-0.16 wt%) and Na₂O (<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 + Fe²⁺) atomic ratio) = 92.2-94.0, Al₂O₃ = 1.33-1.38 wt%, CaO = 0.62-1.48 wt% and
385 TiO₂ = 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 + Fe²⁺) atomic ratio) and XNa (100 × Na/(Na + K) atomic ratio) are 89.6-92.4 and
392 91.7-98.2 respectively (Na₂O = 2.5-3.1 wt%; K₂O = 0.08-0.40 wt%). The TiO₂ content of
393 amphibole varies from 0.30 to 1.8 wt%. Similarly to what was described for the Al₂O₃ 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 TiO₂ 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 TiO₂ 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 Cr_2O_3 and 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 XMg of chromite and Fo of olivine: Fo reaches values of about 90.5 in chromite-bearing
419 dunites (rare disseminated chromite) to 94.5 in antinodular to massive chromitites. The NiO
420 content of olivine, despite a high scatter, is correlated to the Fo . The 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 Cr , IVAl and Fe^{3+} in chromite. In samples - or parts of
425 thin sections - with the highest amount of chromite (scattered, antinodular and massive), 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 Fe^{3+} is negatively correlated to those two chemical species (Fig. 8a-b). In these cases,
430 Cr and Al generally decrease together and 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 Al_2O_3 (from 2.6 to 1.4 wt%), TiO_2 (from 0.29 to 0.20 wt%) and
435 Na_2O (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, YFe^{3+} and TiO_2 of
438 chromite follows the chromite abundance. Systematically Fo and XMg increase and 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 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 Cr₂O₃ (0.60-1.2 wt%) and lower Al₂O₃ (0.41-1.3 wt%), CaO (0.21-0.83 wt%)
453 and TiO₂ (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, Al₂O₃ and TiO₂ 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 Al₂O₃ of host chromites and for Al₂O₃ and TiO₂ 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 Na₂O content (0.54-1.3 wt%), both higher than those of the clinopyroxene in the matrix.
465 These inclusions show more restricted Al₂O₃ (2.4-2.6 wt%), Cr₂O₃ (1.3-1.6 wt%) and TiO₂
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 Al₂O₃, Cr₂O₃, TiO₂ and Na₂O 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 TiO₂ 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_2O content ranges from 2.9 to 4.1 wt%, while their K_2O 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_2O_3 values basically follows the same tendency observed in
481 orthopyroxene inclusions. However, the distribution of the TiO_2 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_2O_3 and (high) TiO_2 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_2O (3.1-5.6 wt%)
490 and moderate K_2O (0.61-3.8 wt%) contents ($\text{Na}_2\text{O} + \text{K}_2\text{O} = 6.1-7.0$ 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_2O_3 and TiO_2 contents
494 (not shown). The XNa values lower than 90 (enrichment in K_2O relative to Na_2O) 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 $\text{Na}_2\text{O} = 0.34-$
499 6.3 wt%, $\text{K}_2\text{O} = 0.32-6.9$ wt% and $\text{XNa} = 7.0-96.8$), or from the average of values from two
500 distinct populations within a single sample (e.g. sample 06OM04-4 (1026 m) with $\text{Na}_2\text{O} =$

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; Zagrtidenov 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 TiO₂ 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; Zagrtednov 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 XMg and YFe³⁺ 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 TiO₂ 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-Fe²⁺ 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^{2+} 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^{3+} , and
642 sometimes TiO_2 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^3 , 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 SiO₂ and/or H₂O,
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 TiO₂ 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

764 **Acknowledgments**

765 M.R. and G.C. are very indebted to F. de Parseval for thin sections preparation and to Ph. de
766 Parseval for assistance during microprobe data acquisition. We are extremely grateful to our
767 colleagues from the Public Authority of Mining, M. Al Araiimi, M. Almusharrafi, M. Al
768 Batashi, A. Al-Rajhi and H. Al Azri, for their constant support and hospitality during field
769 work in Oman. Constructive comments and suggestions from A. Kapsiotis, E. Mathez and the
770 editor A. Kerr greatly helped to improve the manuscript. This work has benefited from a
771 financial support provided by the French CNRS (Centre National de la Recherche
772 Scientifique) through the TELLUS/SYSTER program of INSU (Institut National des Sciences
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 Zagrtednov
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 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 Cr_2O_3 and Al_2O_3 contents, the XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio),
 846 the YFe^{3+} ($100 \times \text{Fe}^{3+}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$ atomic ratio) and the 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 Al_2O_3 and TiO_2 contents of
 849 clinopyroxene matrix, in d) the XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio) and the Al_2O_3 and
 850 TiO_2 contents of orthopyroxene in inclusion in chromite grains, in e) the XMg ($100 \times$

851 Mg/(Mg + Fe²⁺) atomic ratio) and the TiO₂ 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 Fe³⁺ expressed in *apfu* (atom per formula unit, recalculated on
861 the basis of O = 4), XMg and TiO₂ 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, YFe³⁺ and TiO₂ 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

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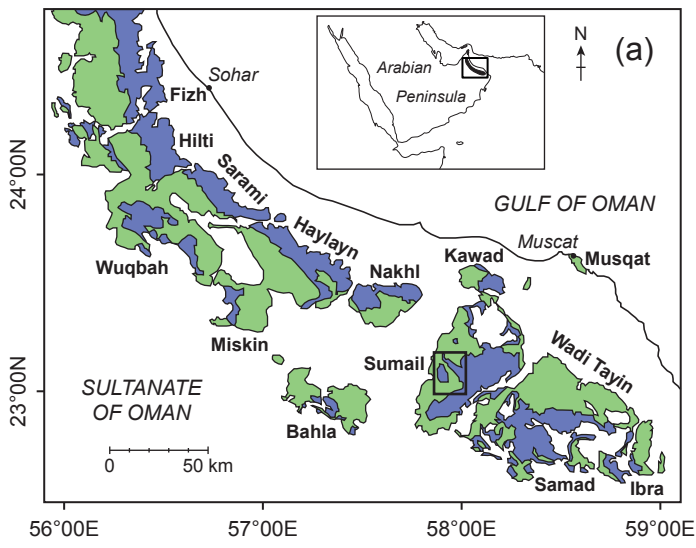
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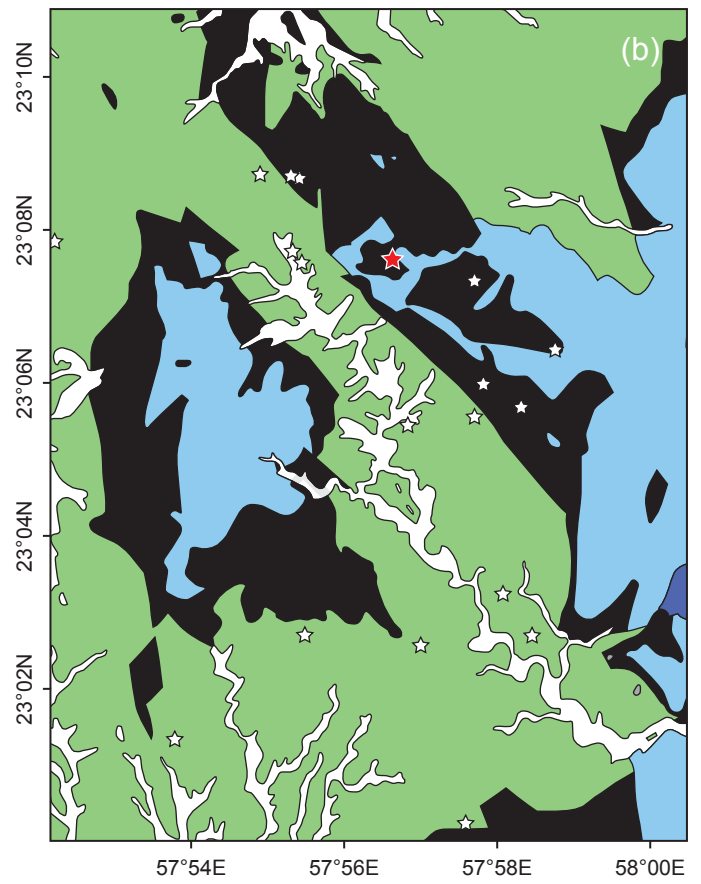
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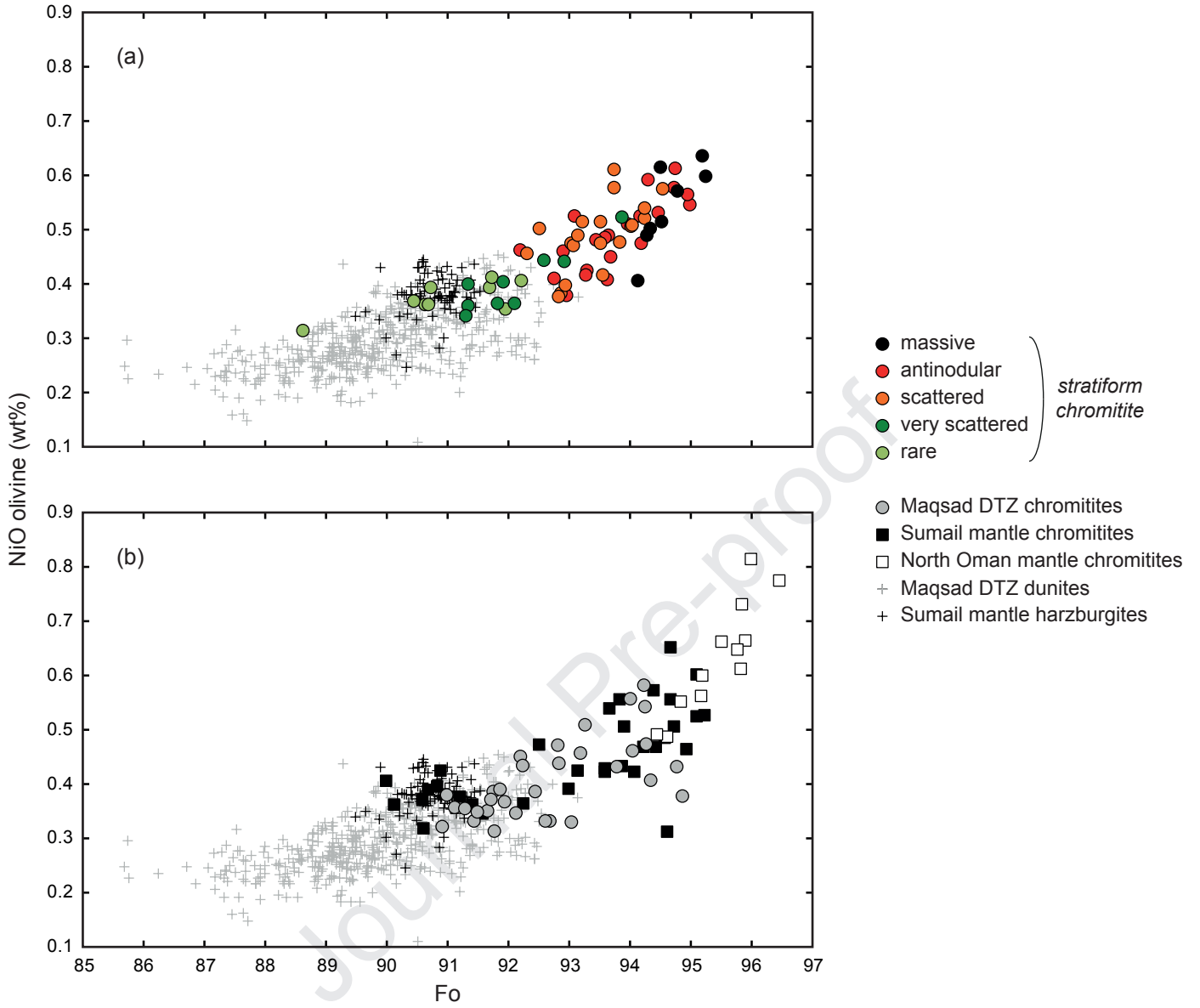
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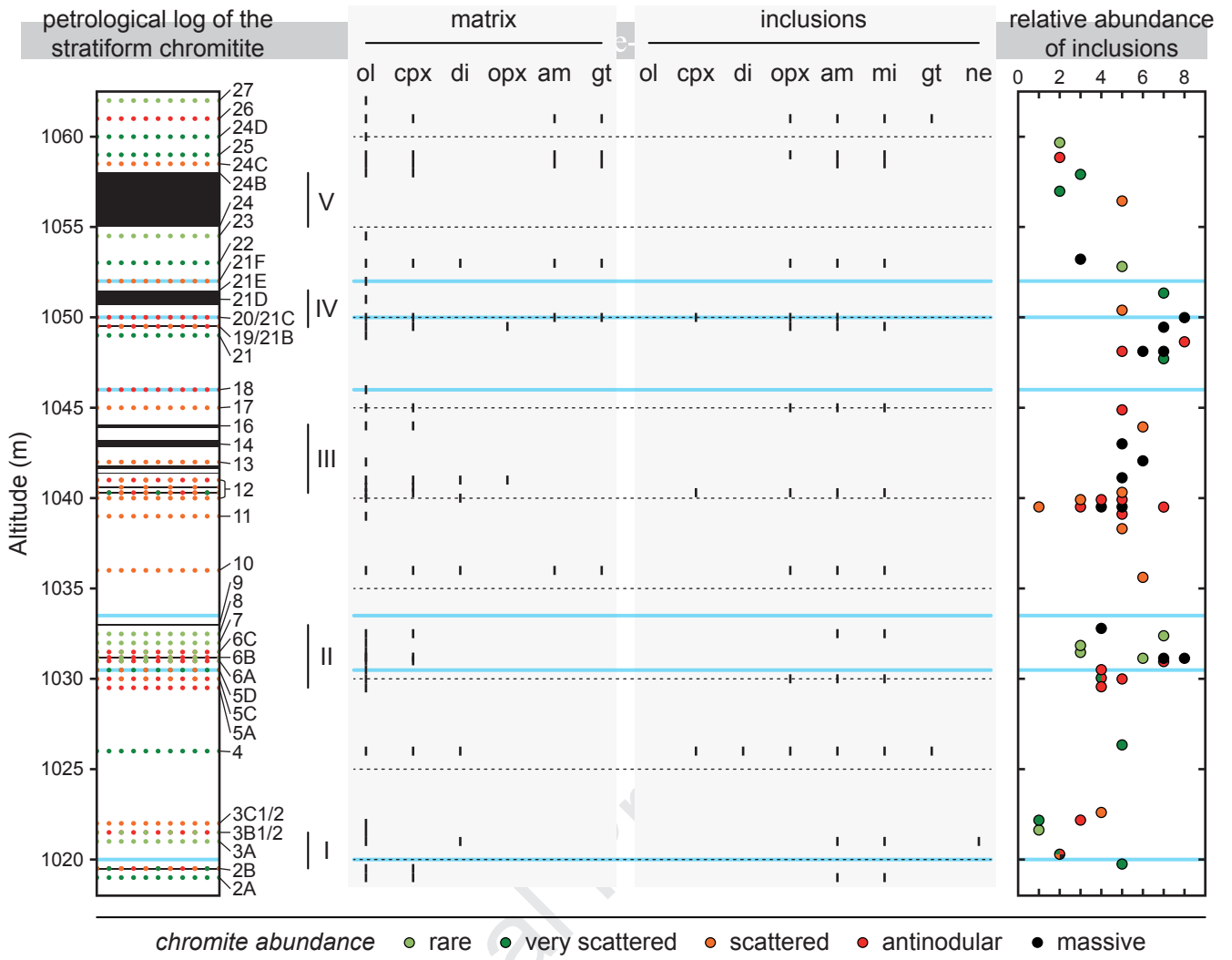
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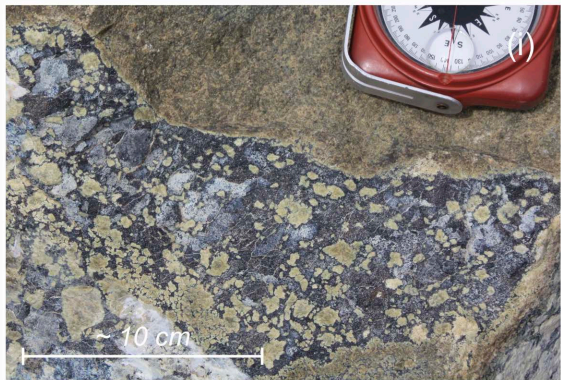
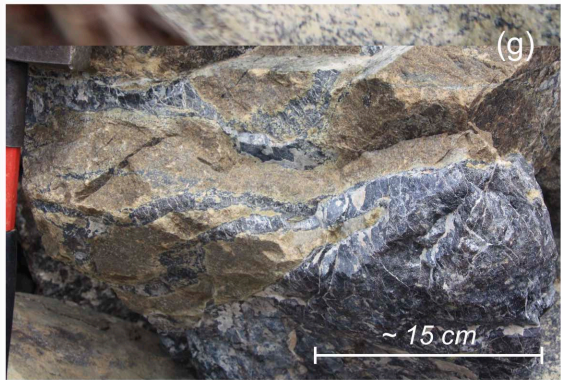
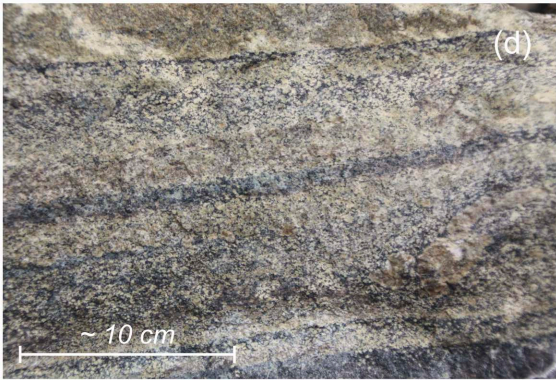
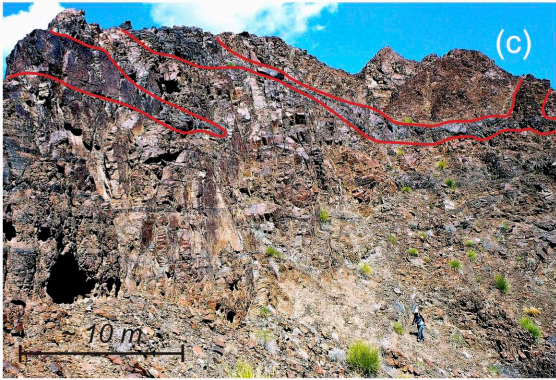


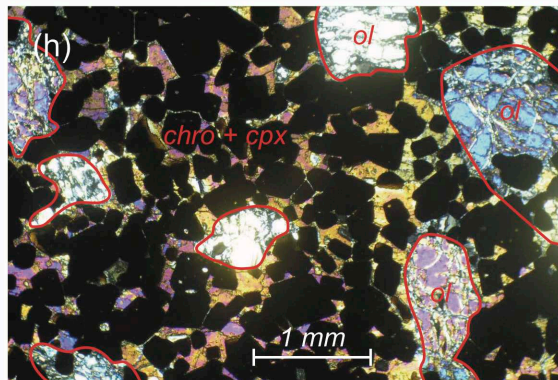
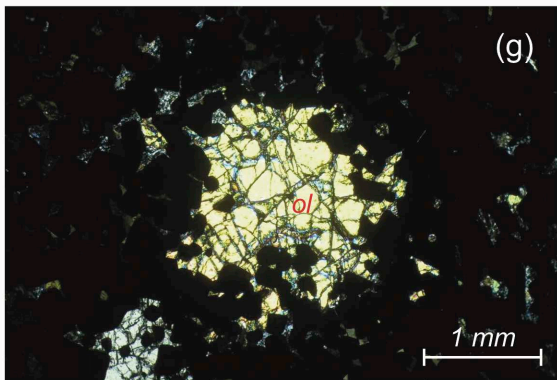
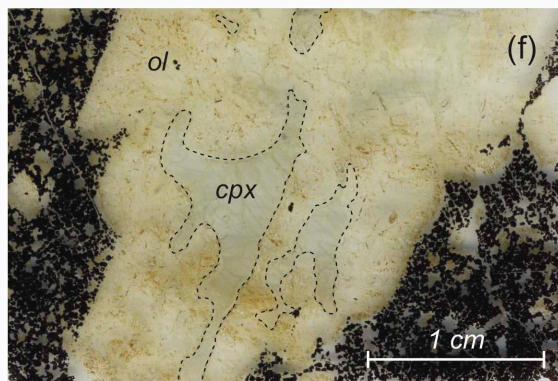
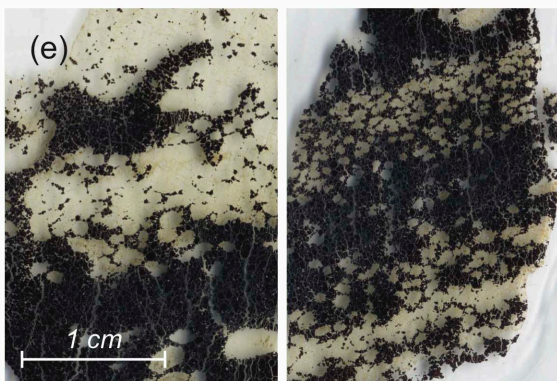
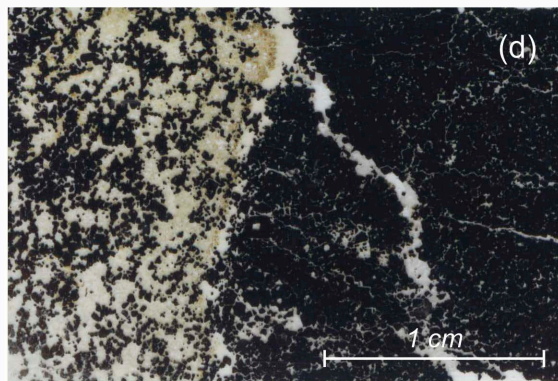
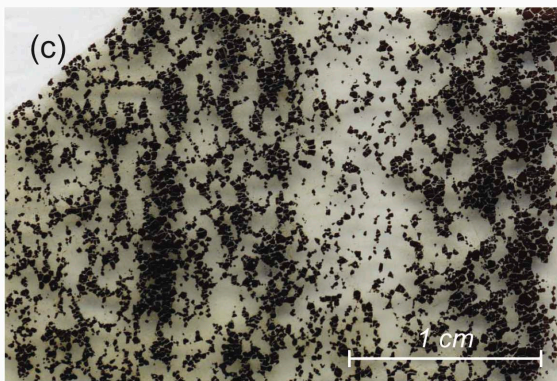
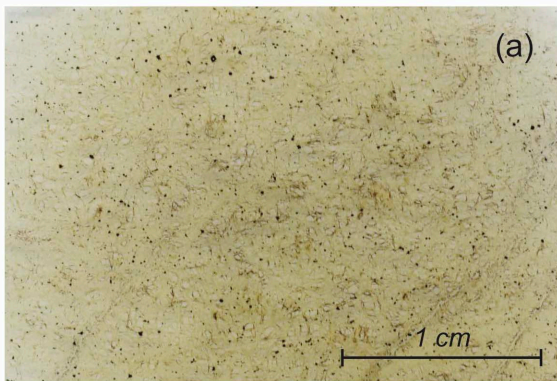
- Crustal units
- Lower crustal layered cumulates
- Dunitic transition zone (DTZ)
- Mantle harzburgites
- ★ Stratiform chromitite (this study)
- ☆ Other chromitite occurrences

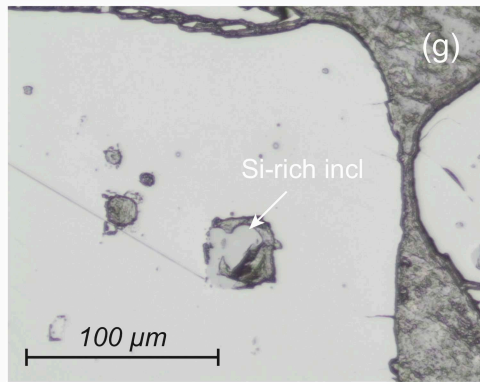
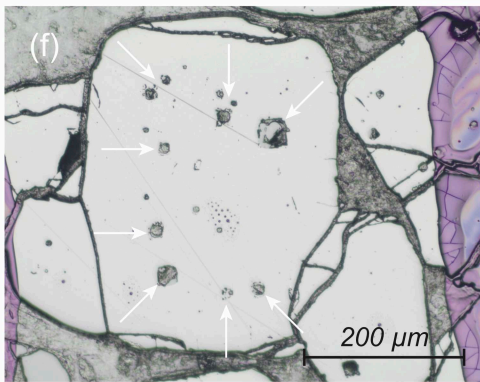
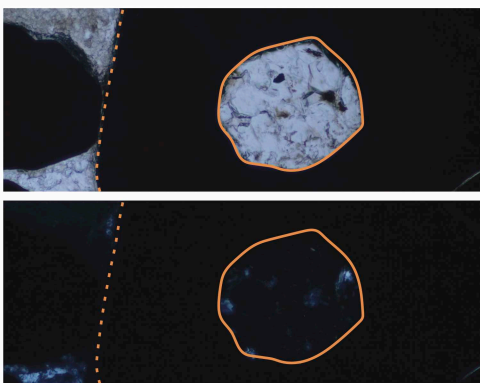
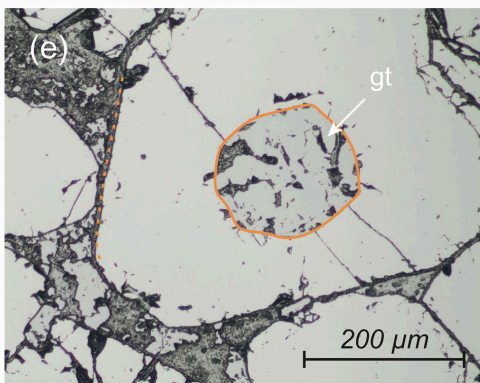
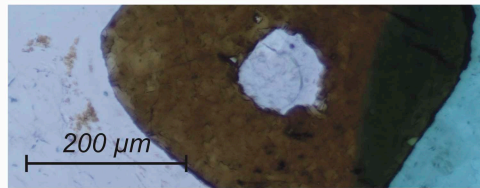
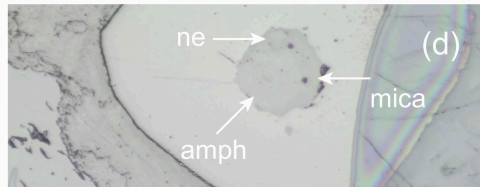
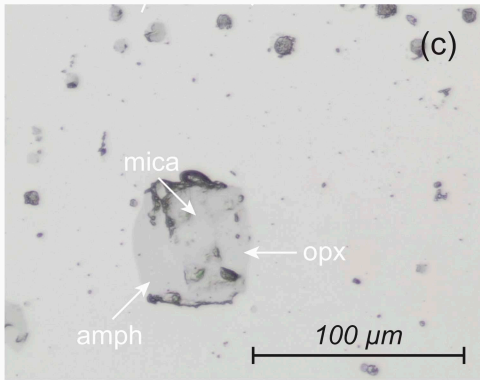
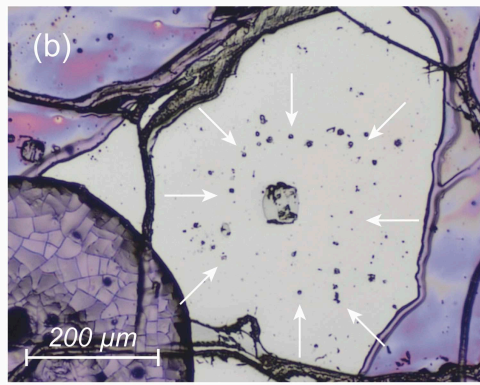
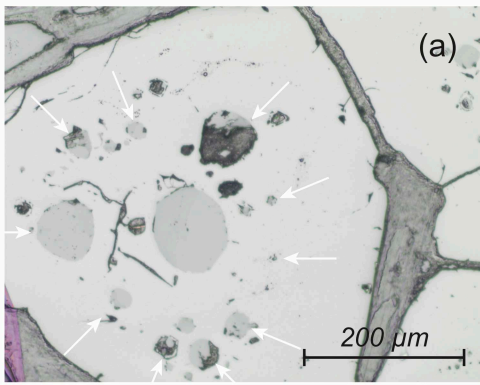


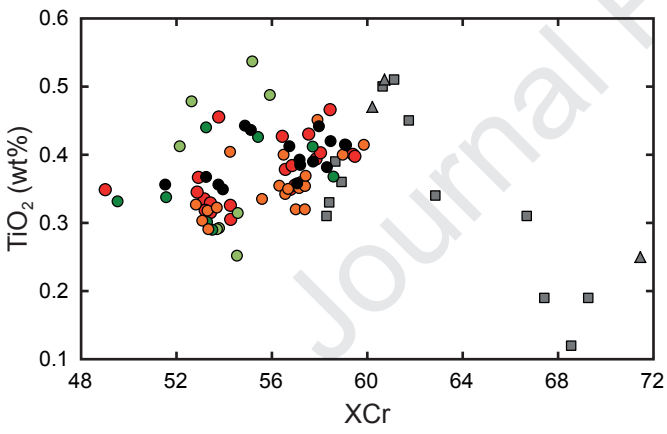
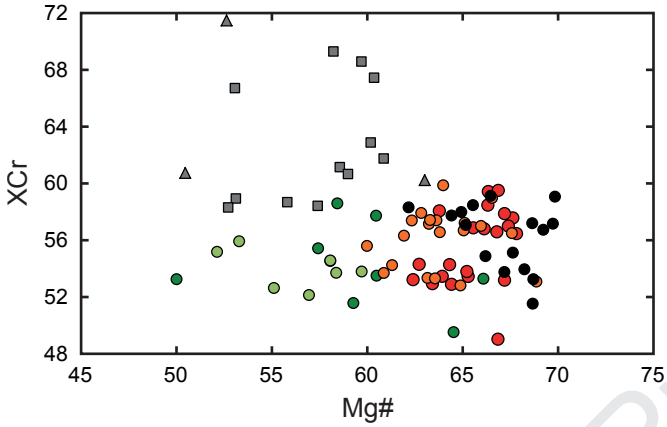
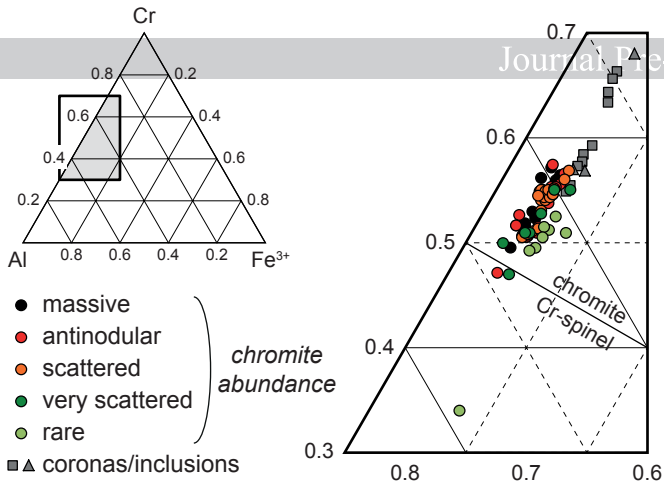


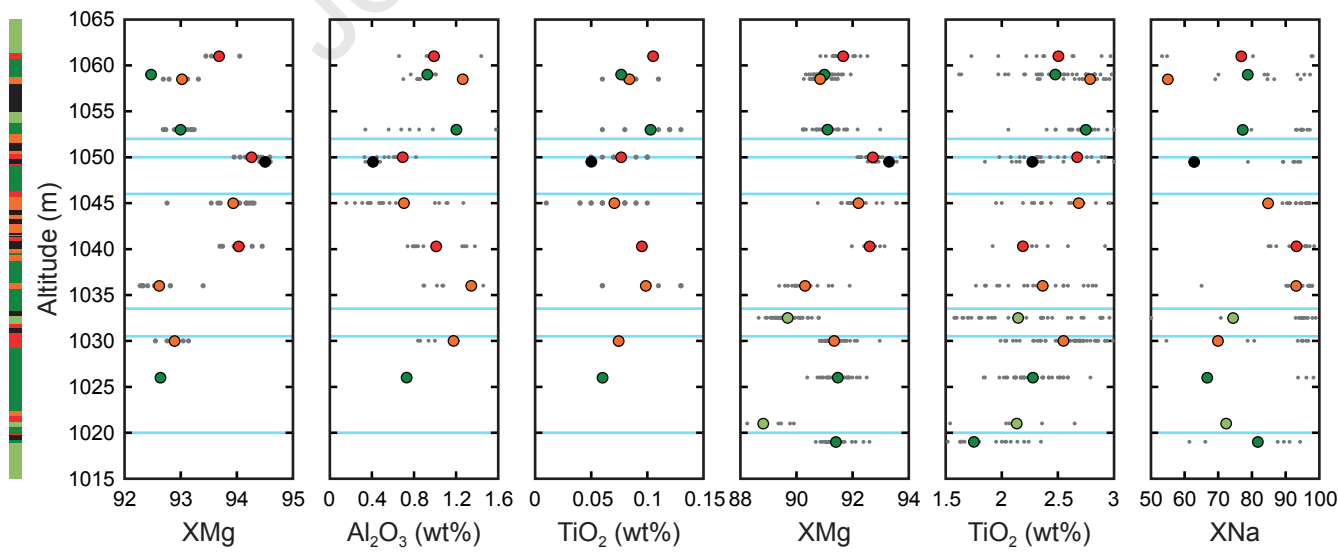
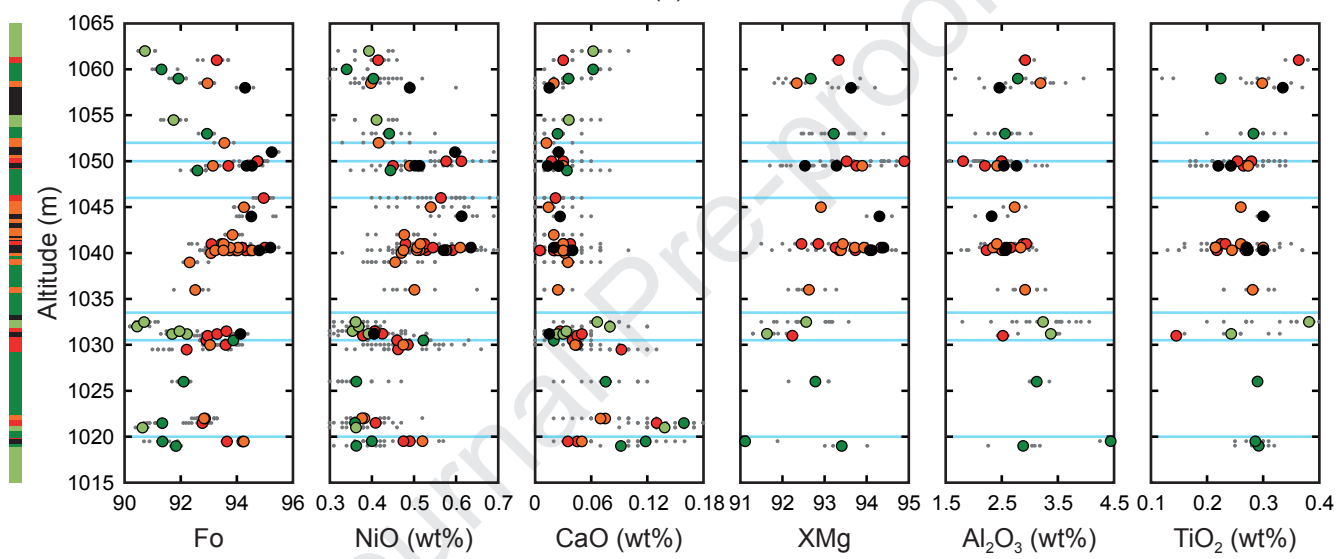
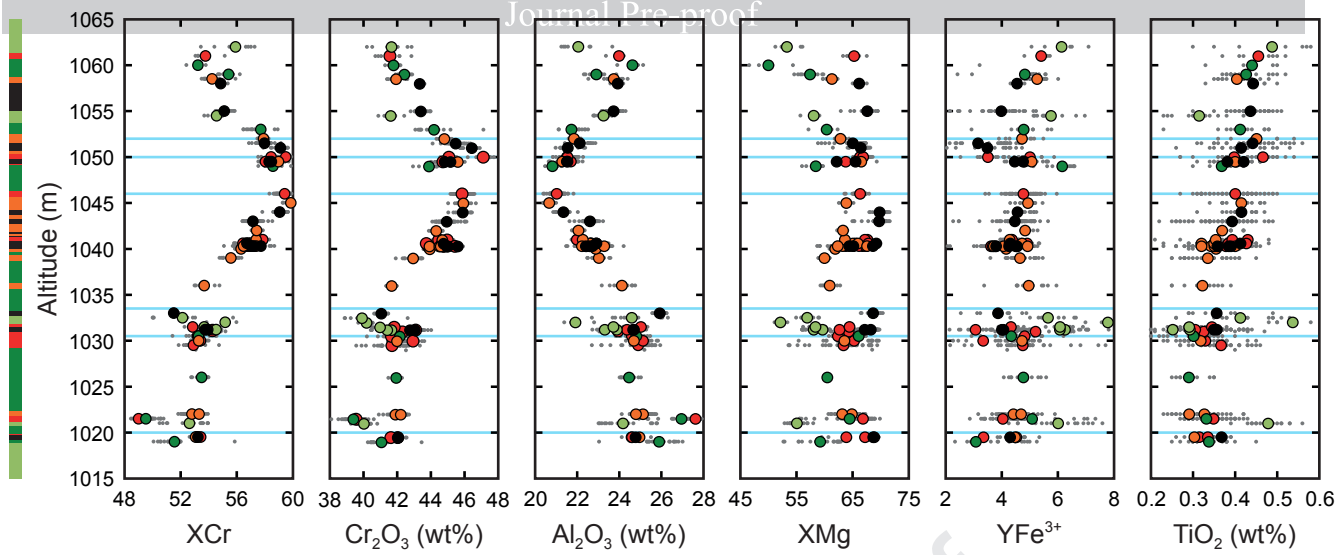






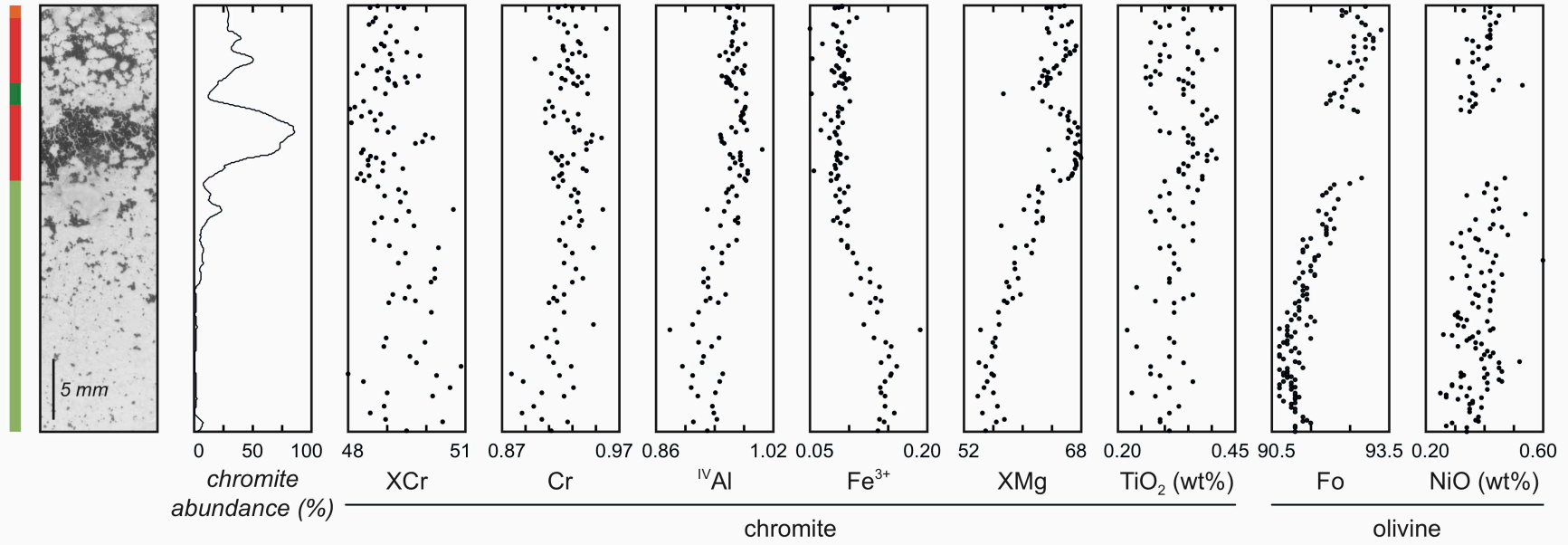




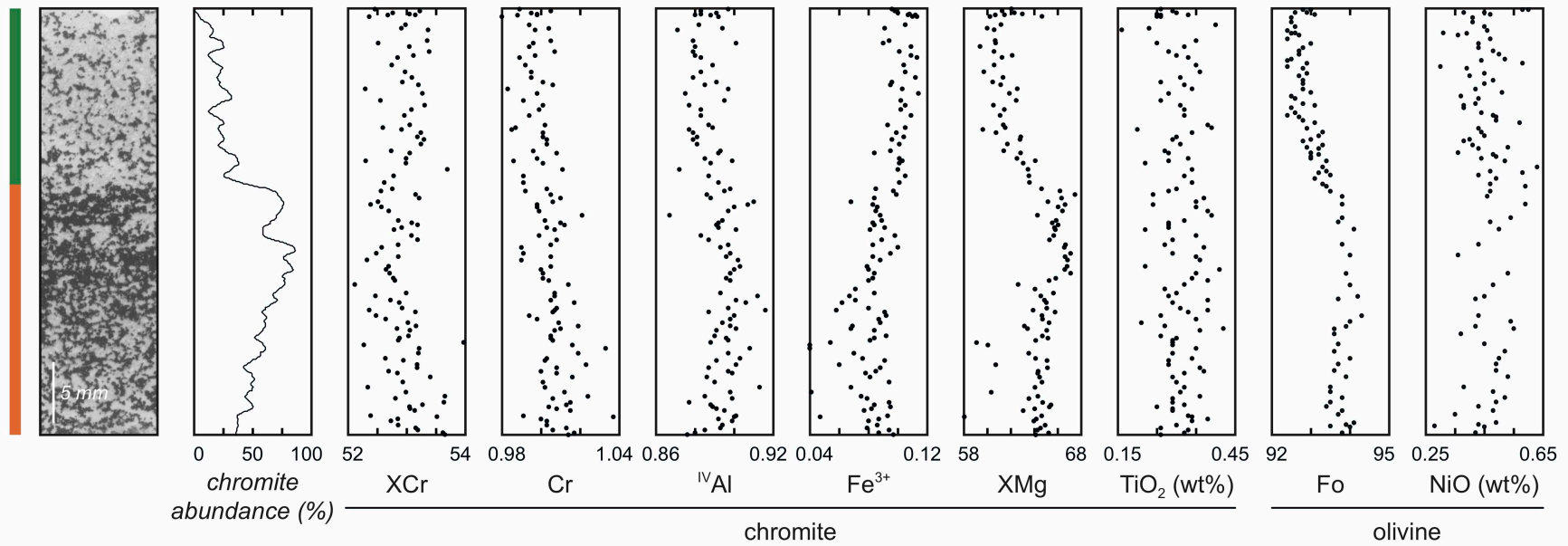


chromite abundance ● rare ● very scattered ● scattered ● antinodular ● massive

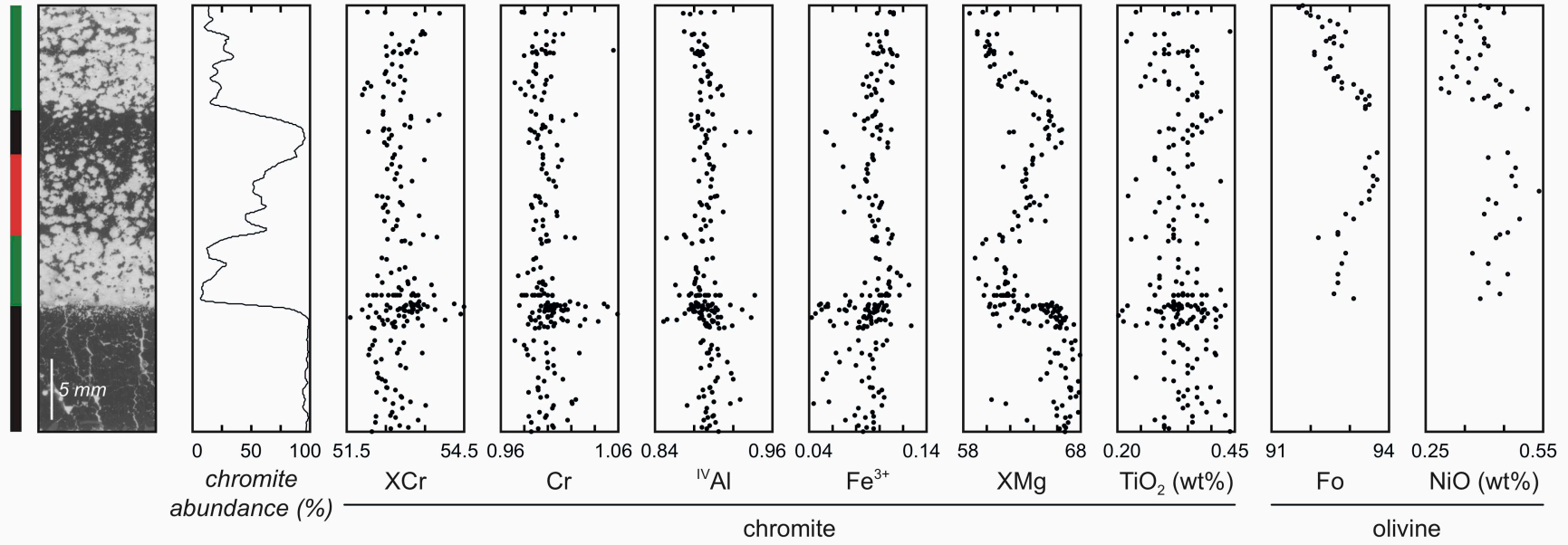
(a) 06OM04-3B2



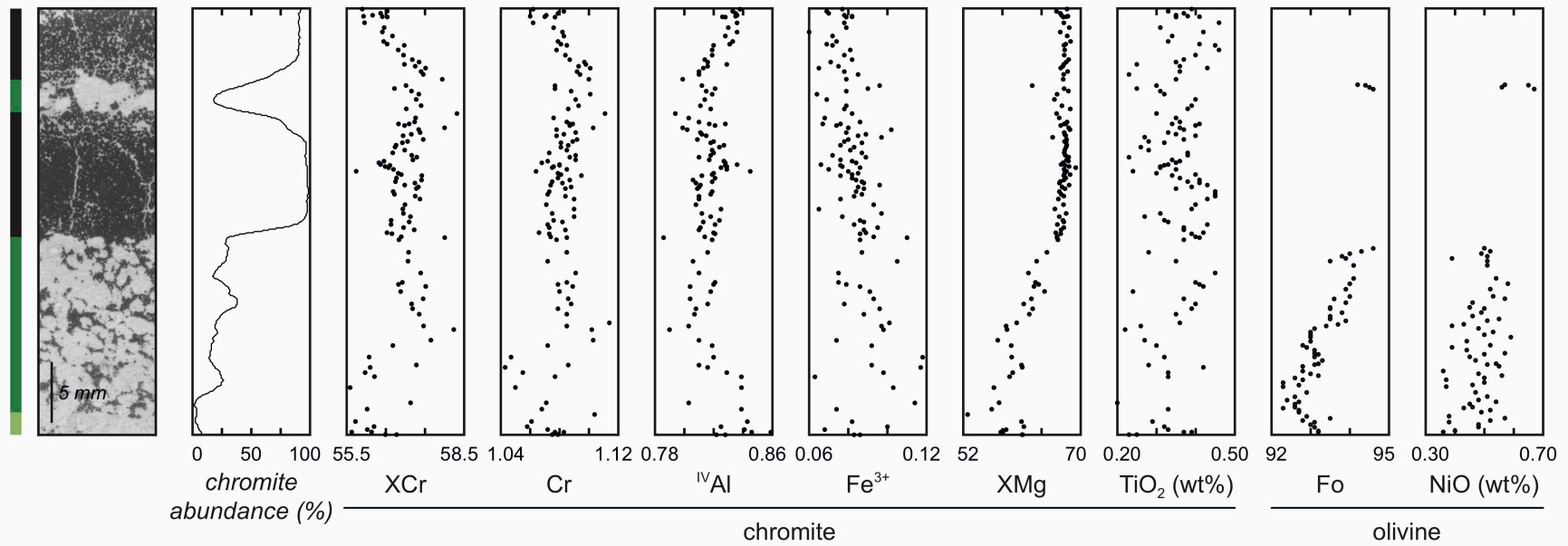
(b) 06OM04-5D

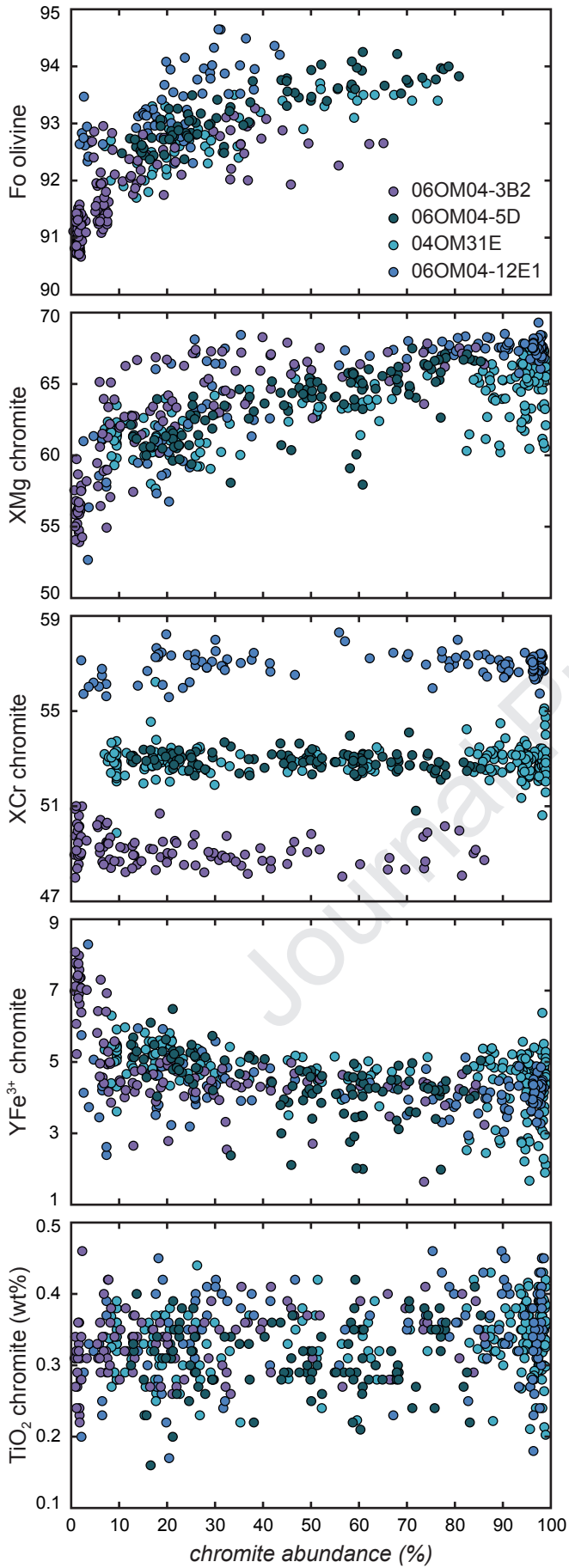


(c) 04OM31E



(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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5 Mathieu Rospabé ^{a,b,*}, Georges Ceuleneer ^a, Nicolas Granier ^a, Shoji Arai ^c, Anastassia Y.
6 Borisova ^{a,d}

7
8 ^a Géosciences Environnement Toulouse (GET), Observatoire Midi-Pyrénées, Université de
9 Toulouse, CNRS, IRD, 14 avenue E. Belin, F-31400 Toulouse, France

10 ^b Research Institute for Marine Geodynamics (IMG), Japan Agency for Marine-Earth Science
11 and Technology (JAMSTEC), 2-15 Natsushima, Yokosuka, Kanagawa 237-0061, Japan

12 ^c Department of Sciences, College of Science and Engineering, Kanazawa University,
13 Kanazawa, Ishikawa 920-1192, Japan

14 ^d Geological Department, Lomonosov Moscow State University, Vorobievsky Gory, 119899,
15 Moscow, Russia

16
17 * Corresponding author. E-mail address: mrospabe@jamstec.go.jp

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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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17 * Corresponding author. E-mail address: mrospabe@jamstec.go.jp

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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).

23 This work has benefited from a financial support provided by the French CNRS (Centre
24 National de la Recherche Scientifique) through the TELLUS/SYSTER program of INSU
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 ₂	0.05	40.90	57.42	52.35	52.69
TiO ₂	0.34		0.06	0.27	0.28
Al ₂ O ₃	25.89		0.73	2.53	2.38
Cr ₂ O ₃	41.09	0.02	0.97	1.34	1.46
FeO _{total}	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 ₂ O			0.03	0.47	0.54
K ₂ 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 ³⁺	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		