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.

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Abstract

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

The chromite composition is typical of a MORB kindred - moderate XCr ($100 \times \text{Cr}/(\text{Cr} + \text{Al})$ atomic ratio), ranging from 48 to 60, and relatively high TiO_2 content, ranging from about 0.3 to 0.5 wt% -, a characteristic shared by most lithologies issued from the igneous activity of the Maqsad diapir. The silicate matrix is essentially made of slightly serpentinized olivine with minor clinopyroxene and rare pargasitic amphibole, orthopyroxene and garnet. This strongly contrasts with the nature of the mineral inclusions mostly made of the assemblage amphibole-orthopyroxene-mica, enclosed in the chromite grains and represented in abundance all along the ore body whatever the ore grade. The inclusions demonstrate the involvement of a silica- and water-rich melt and/or fluid, in addition to MORB, in the early stages of chromite crystallization.

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

Keywords

Oman ophiolite; chromite; stratiform chromitite; melt hybridization; dunitic mantle-crust transition zone

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Highlights

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

1. Introduction

The fate of chromium from its mantle source and its concentration by the crystallization and accumulation of chromite in cooling magma bodies is a long-standing poorly understood process of the magmatic history of mafic and ultramafic rocks (e.g. Bowen, 1928; Lombaard et al., 1956; Stoll, 1958), and is still the matter of intense discussions (cf. Arai and Miura, 2016 and references therein). Different mechanisms were proposed to account for the formation of massive chromitites in crustal magma chambers, following the crystallization from a chromite-oversaturated basaltic melt in which chromite was the only liquidus phase and by the chromite crystals settling and concentration in layers of variable thickness. Among others processes, the mixing between a basalt saturated in chromite and a more silicic melt, contamination of the basalt by the surrounding felsic crustal rocks, addition of H₂O, increase in the total pressure or fO₂ of the parent magma, saturation in chromite alone in the parent magma following a decrease in the lithostatic pressure (e.g. Irvine, 1975, 1977a, 1977b; Latypov et al., 2018; Lipin, 1993; Matveev and Ballhaus, 2002; Mondal and Mathez, 2007; Naldrett et al., 2012; Spandler et al., 2005) were proposed. The most popular model consists in the massive chromite crystallization from a melt hybrid between a relatively primitive basalt and a silica-rich melt, issued from the country rock re-melting (Irvine, 1975) or from the same melt at a more evolved, differentiated stage (Irvine, 1977a). Some authors also proposed the involvement of volatile- and sodium-rich melts or

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

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

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

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

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

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

2. The Maqsad area of the Oman ophiolite

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

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

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

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

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

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

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

3. Description of the studied stratiform chromitite

3.1. Structuration of the stratiform chromitite

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

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

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

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

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

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

3.2. Distribution of inclusions in chromite

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

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

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

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

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

4. Mineral compositions

4.1. Analytical note

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

hydrous/anhydrous silicates). The two datasets obtained following the using of the two different microprobes were compared; no shift was observed. The following synthetic and natural minerals standards were used: albite (Na), periclase (Mg), corundum (Al), sanidine (K), wollastonite (Si, Ca), pyrophanite (Mn, Ti), hematite (Fe), chromium oxide (Cr), nickel oxide (Ni), sphalerite (Zn) and pure vanadium (V). Oxides concentrations were obtained by correcting nominal concentrations following the PAP data reduction method (Pouchou and Pichoir, 1985). The detection limits are 0.03 wt% for TiO_2 , Al_2O_3 , Cr_2O_3 and MgO , 0.04 wt% for SiO_2 and Na_2O , 0.05 wt% for CaO and K_2O , 0.06 wt% for V_2O_3 and NiO , and 0.07% for FeO , MnO and ZnO , for all mineral phases analysed. The internal precision (%RSD) is better than 2% for major elements (e.g. Cr_2O_3 in chromite, MgO in olivine, CaO in clinopyroxene), better than 6% for TiO_2 (chromite, clinopyroxene) and better than 30% for minor elements (e.g. NiO in olivine, Na_2O in clinopyroxene). A few selected data are available in Table 1. The whole dataset is provided in Supplementary Material.

4.2. Chromite

Chromites from the stratiform chromitite are Cr- and Al-rich relative to Fe^{3+} (see ternary diagram in Figure 6). They display variable chemical compositions, with averaged per sample XCr ($100 \times \text{Cr}/(\text{Cr} + \text{Al})$ atomic ratio) = 49.0-59.9 (Cr_2O_3 = 39.4-47.1 wt%; Al_2O_3 = 20.7-27.6 wt%), XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio) = 50.0-69.8, YFe^{3+} ($100 \times \text{Fe}^{3+}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$ atomic ratio) = 1.1-7.8 (4.6 in average), and TiO_2 = 0.25-0.54 wt% (averaged values for each samples) (Fig. 6). Chromite inclusions and coronas have a higher XCr , ranging from 58.3 to 71.5 (Cr_2O_3 = 43.7-52.7 wt%; Al_2O_3 = 14.1-21.3 wt%), a similar to slightly lower XMg (50.5-63.0), and a less variable but globally higher in average YFe^{3+} (4.3-6.3, 5.3 in average). The YFe^{3+} is slightly higher in chromite inclusions (4.9-6.3) than in coronas (4.3-5.9). Chromite inclusions and coronas also display a clear negative correlation

between their TiO_2 content (0.12-0.51 wt%) and their XCr (Fig. 6).

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

In detail, Cr and Al, which typically substitute for each other in spinel, show an unexpected decoupling at some levels. Although their concentrations are anti-correlated at the scale of the whole stratiform chromitite, they show a parallel decreasing trend within the intermediate breccias system (1030-1034 m) as well as above the altitude 1058 m (correlated to the TiO_2 content and negatively correlated to YFe^{3+}). Otherwise, in between the three highest breccias (1049-1053 m), the Cr_2O_3 content shows two successive increasing and decreasing trends within a few metres while Al_2O_3 continuously increases on the same interval. XMg shares basically the same evolution as XCr and is particularly well-correlated to the Cr_2O_3 content, while oppositely YFe^{3+} is negatively correlated to XMg (Fig. 7a). The vertical evolution of the TiO_2 content is decoupled from the other chemical species/atomic ratios and shows, with few local exceptions, a global and relatively monotonous increase from about 0.30 wt% at the base of the stratiform chromitite to 0.50 wt% to its top (Fig. 7a). The few local exceptions occur generally at altitudes where magmatic breccias were observed.

4.3. Silicate matrix

4.3.1. Olivine

The olivine matrix shows marked variations in composition with Fo ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}_{\text{total}})$ atomic ratio) = 90.4-95.2, NiO = 0.34 to 0.64 wt%, and CaO ranging from detection limit to 0.16 wt% (averaged per thin section). Vertical variations of both the Fo and the NiO content are closely correlated to the XCr, especially to the Cr_2O_3 content, and to the XMg of chromite (Fig. 7a-c). We observe scattered values in Fo and NiO at the base of the stratiform chromitite, overlain by common and successive (1) decrease between the two intermediate breccias, (2) increase to highest breccias (with a much higher variability at the altitude 1040 m) and (3) decrease within the uppermost part of the ore body. MnO content is conversely anti-correlated to those chemical species/ratios (not shown). High CaO values of the olivine matrix (i.e. higher than 0.05 wt%) are restricted to the base of the stratiform chromitite, particularly around the lowest breccia at the altitude 1020 m (Fig. 7b). Beyond these global trends at the scale of the stratiform chromitite, alternations of increases and decreases in a given element or elemental ratio can be observed at a finest scale, especially from the highest breccias and above.

4.3.2. Clinopyroxene

Clinopyroxene, the second most abundant mineral phase after olivine in the matrix, shows XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio, average per thin section) varying between 91.1 and 94.9, and Al_2O_3 = 1.8-4.4 wt%, Cr_2O_3 = 1.1-1.5 wt, TiO_2 = 0.15-0.38 wt% and Na_2O = 0.22-0.58 wt%. The XMg of clinopyroxene is correlated to Fo of olivine (and XMg of chromite) (Fig. 7c). While most difficult to decipher, the vertical evolution of both Al_2O_3 and TiO_2 contents of clinopyroxene, as well as of Na_2O (not shown), is negatively correlated to

the one of XMg. The few analysed hybrid diopsides (i.e. hybrid in composition between igneous clinopyroxene and pure hydrothermal diopsides; Rospabé et al., 2017) show higher XMg (94.3-97.0) and SiO₂ content (53.8-55.0 wt% instead of 51.8-53.6 wt%) coupled with lower Al₂O₃ (0.47-1.4 wt%), TiO₂ (0.03-0.16 wt%) and Na₂O (<0.23 wt%) contents than in clinopyroxene whose composition fits with an igneous origin.

4.3.3. Orthopyroxene

The few orthopyroxene occurrences in the matrix are enstatites, with XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio) = 92.2-94.0, Al₂O₃ = 1.33-1.38 wt%, CaO = 0.62-1.48 wt% and TiO₂ = 0.13-0.15 wt% (Table 1 and Supplementary Material).

4.3.4. Amphibole

Excepting for one sample (06OM04-10, 1036 m), the matrix contains amphibole only above the higher breccias system in the uppermost part of the stratiform chromitite ore body. Amphibole is mostly pargasitic to edenitic hornblende. The variation ranges of XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio) and XNa ($100 \times \text{Na}/(\text{Na} + \text{K})$ atomic ratio) are 89.6-92.4 and 91.7-98.2 respectively (Na₂O = 2.5-3.1 wt%; K₂O = 0.08-0.40 wt%). The TiO₂ content of amphibole varies from 0.30 to 1.8 wt%. Similarly to what was described for the Al₂O₃ content of chromite and for other chemical species in olivine and clinopyroxene (Fig. 7a-c), amphibole composition evolves differently below and above the altitude 1058 m. Below this level, XMg decreases from 92.4 to 90.1 and XNa and TiO₂ increase from 96.8 to 91.7 and from 1.5 to 1.8 wt% respectively. Above this level, the reverse trend is observed, XMg, XNa and TiO₂ evolving to 90.9, 98.2 and 0.30 wt% respectively.

4.3.5. Garnet

Garnet observed at the same levels where the amphibole occurs is mostly grossular (sum of oxides ranging from 97.8 to 100.2 wt%), with $\text{SiO}_2 = 36.0\text{-}39.6$ wt%, $\text{CaO} = 36.9\text{-}38.2$ 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 0.09 and 1.36 wt%, and 0.03 and 0.15 wt% respectively (Table 1 and Supplementary Material).

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

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

The samples selected display mm- to cm-scale extremely variable chromite and olivine modal contents, covering the entire variation range described previously (from rare chromite in dunite to massive chromitite). Despite differences in terms of absolute chemical compositions from one sample to another, these samples share common relationships between chemical and modal variations. Higher amounts of chromite are perfectly correlated to higher XMg of chromite and Fo of olivine: Fo reaches values of about 90.5 in chromite-bearing dunites (rare disseminated chromite) to 94.5 in antinodular to massive chromitites. The NiO content of olivine, despite a high scatter, is correlated to the Fo. The TiO_2 content of chromite displays a small-scale high dispersion; however, it seems that highest values are generally observed in antinodular to massive ore portions.

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

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

The thin section of sample 06OM04-12E1 shows interstitial clinopyroxene, restricted to the decreasing peak of the amount of chromite and above, within the upper massive interval (Fig. 8d). Within this interval, we observed the increase of XMg (from 94.0 to up to 99) 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 Na_2O (from 0.38 to 0.22 wt%) contents of clinopyroxene.

In a different view, Figure 9 shows that in the four samples (06OM04-3B2, -5D, -12E1 and 04OM31E), the evolution of Fo of olivine and of XMg, XCr, YFe^{3+} and TiO_2 of chromite follows the chromite abundance. Systematically Fo and XMg increase and YFe^{3+} decreases when variations in the chromite modal content ranges from 0 to 30-40% chromite. Above 40%, there is no more evolution of these chemical ratios which seem to be buffered to values of about 93.5, 65 and 4.5 respectively. XCr is constant around 53 in the two samples 06OM04-5D and 04OM31E whatever the chromite modal abundance, and tends to this value in the two other samples when the chromite abundance becomes lower than 40% (i.e. steady values near 49 and 57 in 06OM04-3B2 and -12E1 respectively above 40% of chromite). The TiO_2 content does not show marked variation from below to above 40% of chromite, and appears not affected by the variation of the chromite modal content excepting a very slight and constant increase from scarce to massive ores.

4.5. Silicate inclusions

4.5.1. Orthopyroxene

Orthopyroxene in inclusion in chromite is enstatite, with a similar XMg (92.5-94.5), slightly higher Cr₂O₃ (0.60-1.2 wt%) and lower Al₂O₃ (0.41-1.3 wt%), CaO (0.21-0.83 wt%) and TiO₂ (0.05-0.11 wt%) contents than those of the few orthopyroxene crystals observed in interstitial position and associated to the other matrix minerals. Higher XMg values are observed along the interval 1040-1050 m (Fig. 7d), together with high XCr and XMg of chromite (Fig. 7a) and high Fo of olivine and XMg of clinopyroxene (Fig. 7b-c). In the same way, Al₂O₃ and TiO₂ contents increase together from 1026 to 1036 m, decrease to about 1050 m, then globally re-increase to the top of the stratiform chromitite with a clear disturbance around the altitude 1058 m (Fig. 7d). These variations largely mimic the ones observed for Al₂O₃ of host chromites and for Al₂O₃ and TiO₂ contents of matrix clinopyroxene.

4.5.2. Clinopyroxene

The less common inclusions of clinopyroxene exhibit variable XMg (93.0-97.3) and Na₂O content (0.54-1.3 wt%), both higher than those of the clinopyroxene in the matrix. These inclusions show more restricted Al₂O₃ (2.4-2.6 wt%), Cr₂O₃ (1.3-1.6 wt%) and TiO₂ (0.20-0.36 wt%) compositional ranges (averaged per thin section). Only one inclusion of diopside of likely hydrothermal origin (cf. Python et al., 2007; Rospabé et al., 2017) was found. Its XMg and Al₂O₃, Cr₂O₃, TiO₂ and Na₂O contents values are 96.0, 1.0 wt%, 0.73 wt%, 0.22 wt% and 0.20 wt% respectively.

4.5.3. Amphibole

Amphiboles occurring as inclusions in chromites are pargasites/pargasitic hornblendes, for more than 60% of them, or magnesio-hastingsites/magnesio-hastingsitic hornblendes, for more than 30% of them. The XMg and TiO₂ content values, averaged for each sample, are much variable (88.8-93.3) and higher (1.8-2.8 wt%), respectively, compared

to those of amphiboles from the matrix. Within each sample, the variability from an amphibole inclusion to another is much higher than concerning amphibole in the matrix (Fig. 7e). Their Na_2O content ranges from 2.9 to 4.1 wt%, while their K_2O content is lower than 0.10 wt%, making their XNa to be restricted to very high values (98.2-99.3). The vertical evolution of XMg and Al_2O_3 values basically follows the same tendency observed in orthopyroxene inclusions. However, the distribution of the TiO_2 content is much scattered, with a certain increase from lower to higher values from the base to the top of the stratiform chromitite (Fig. 7e).

4.5.4. Mica

Micas, absent in the matrix and observed solely as inclusions, have compositions evolving from phlogopite to aspidolite (Na-rich phlogopite) end-members. Their averaged 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 from 2.6 to 4.3 wt% respectively. Micas are also characterized by high Na_2O (3.1-5.6 wt%) 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 from 55.0 to 93.2 (averaged per thin section). Vertically along the stratiform chromitite, mica composition follows the same evolution observed in other minerals, with high XMg values along the interval of 1040-1050 m and a common evolution of their Al_2O_3 and TiO_2 contents (not shown). The XNa values lower than 90 (enrichment in K_2O relative to Na_2O) are observed at the base of the ore body, in between the lowest and the intermediate breccias, as well as along its uppermost part within and above the higher breccias system (Fig. 7f). These lower XNa values result either, in a given sample, from compositions evolving continuously from Na-rich to K-rich end-members (e.g. sample 06OM04-5C2 (1030 m) with $\text{Na}_2\text{O} = 0.34-6.3$ 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 distinct populations within a single sample (e.g. sample 06OM04-4 (1026 m) with $\text{Na}_2\text{O} =$

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 $K_2O = 0.17-0.60$ wt% in the second population, giving $X_{Na} = 27.9-42.7$ and $93.7-98.3$ respectively).

4.5.5. Other inclusions

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

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

5. Discussion

5.1. Structure and petrology of the stratiform ore body

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

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

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

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

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

5.2. Chemistry of the stratiform ore body

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

(Ahmed and Arai, 2002; Augé, 1987; Miura et al., 2014; Rollinson, 2008; Rollinson and Adetunji, 2015). Evenly, the TiO_2 content of the stratiform chromite is typically high (0.25-0.54 wt%), a characteristic shared by most chromitites and scattered chromites in dunites from the Maqsad area and contrasting with widespread lower values observed in almost all other massifs of the Oman ophiolite. This is a major evidence for the MORB kindred of the parent melt of the various igneous products issued from the Maqsad diapir (Clénet et al., 2010; Python et al., 2008), including chromitites (Borisova et al., 2012; Leblanc and Ceuleneer, 1991; Zagrtedenov et al., 2018).

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

In contrast, chemical variations vertically along the stratiform ore body define complex trends of higher amplitude than locally, mostly concerning the XCr of chromite and in a lesser extent the Fo , NiO and CaO contents of olivine as well as the clinopyroxene composition (Fig. 7a-c). This is reminiscent of the trends observed globally all across the DTZ (Abily and Ceuleneer, 2013; Rospabé et al., 2018, 2019). Parallel trends are observed in the composition of silicate inclusions in chromite, as in the TiO_2 content of amphibole and alkalis of mica (Fig. 7d-f). The high amplitude of these variations cannot have solely resulted from sub-solidus re-equilibration. *Interestingly, these large-scale trends do not mirror at all the variations in modal composition of the ore but correlate to the presence of the three*

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

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

More generally, the olivine matrix from the stratiform chromitite displays Fo and NiO content significantly higher than in nearby dunites from the DTZ and in mantle harzburgites from the Sumail massif, reaching values up to 95.3 and 0.63 wt% respectively (Fig. 10a). Even higher values were reported for mantle chromitites from the north of the Oman ophiolite (Augé, 1987; Miura et al., 2014; Rollinson, 2008, 2019) (Fig. 10b). In the Sumail massif, all podiform chromitites share the same feature, their olivine matrix falling in the same chemical variation range whatever they outcrop within the DTZ or the mantle section. Furthermore, a chromitite dike from the Maqsad DTZ, located few hundred metres to the northeast of and at a higher structural level than the present stratiform ore body, displays higher olivine Fo values in olivine-plagioclase nucleus around which chromite nodules grown than in the olivine matrix surrounding nodules (Zagrtdenov et al., 2018). Taken together, these two observations allow us suggesting that Mg-Fe²⁺ post-magmatic subsolidus exchanges between chromite and olivine (e.g. Cameron, 1975; Fabriès, 1979; Grieco et al., 2018; Lehmann, 1983; O'Neill and Wall, 1987) cannot solely explain such high Fo values associated to high NiO content along the stratiform chromitite: (1) first, the contrast between DTZ and mantle chromitites from the Sumail massif on one hand and mantle chromitites from the northern massifs on the other hand matches with the two main magmatic series identified along the Oman ophiolite. These series were highlighted following the nature of cumulates frozen in the mantle section (*i.e.*

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

The compositions of coexisting olivine and chromite (Figs. 7 and 8) suggest subsolidus re-equilibration at 800 to 900°C, as commonly observed in metamorphosed peridotites (Evans and Frost, 1975). In such rocks, the partitioning of Mg and Fe^{2+} depend on the volume ratio of olivine to chromite as well as on temperature (e.g., Arai, 1980). Interestingly, other studies report on a negative (and not positive as in our case) correlation between Fo of olivine and XMg of chromite (e.g. Bussolesi et al., 2019). These authors interpret this correlation in terms of solid-state exchange of Mg and Fe between olivine (incorporation of Mg) and chromite (incorporation of Fe) grains. In our case this process alone doesn't fit simple mass balance consideration as the correlation between XMg and Fo is also associated to variations in the NiO content of olivine and in the XCr, YFe^{3+} , and sometimes TiO_2 of chromite. An avenue of explanation might be to evoke a percolating interstitial melt (i.e. an open system), re-equilibration being likely triggered by such an interstitial melt that makes diffusional processes more efficient. The equilibration is limited in space, possibly in a few cm^3 , and the mass balance for Mg and Fe have been maintained in the restricted area.

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

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

5.3. Tentative scenario for the formation of the stratiform ore body

Since decades, several mechanisms were proposed to account for the formation of chromitite ore bodies: mixing between contrasted magmas, addition of SiO₂ and/or H₂O, change in the pressure among others (e.g. Irvine, 1975, 1977a, 1977b; Latypov et al., 2018; Lipin, 1993; Matveev and Ballhaus, 2002; Mondal and Mathez, 2007; Naldrett et al., 2012; Spandler et al., 2005). One of the most popular models is the one of Irvine (1975, 1977a), which proposed that massive crystallization of chromite is triggered by the mixing between a basaltic melt saturated in chromite and a more silicic melt (the same melt at a more evolved stage, or felsic contamination by assimilation of crustal rocks surrounding the magma reservoir). More generally, it was also proposed that the nucleation of ‘crustal-like’ minerals associated to the massive chromite crystallization may result from the hybridization between the basaltic fraction and a siliceous volatile- and Na-rich fluid or melt fraction (e.g. Johan et al., 1983, 2017; McElduff and Stumpfl, 1991; Spandler et al., 2005; Talkington et al., 1984). In other studies, it was shown that this mixing process cannot account for the formation of some chromitites, especially in the Bushveld Complex (e.g. Mondal and Mathez, 2007; Naldrett et al., 2012). It is probable that a single model cannot consensually explain the formation of all chromitites worldwide, in old intracontinental intrusions or in ophiolites, but that different processes, working alone or together, may lead to the formation of a given ore body.

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

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

The relationship between (1) the location of brecciated levels, (2) the vertical chemical evolution of all mineral phases along the stratiform chromitite, and (3) the higher abundance of orthopyroxene, amphibole and micas included in chromite grains, supports that melt or magma injections strongly influenced the composition of the parent melt from which crystallized the ore body. The existence of such injections is witnessed by the several levels of breccia interlayered with the dunites and chromitites. Especially, the decoupling between the zigzag variation pattern in XCr, and the smooth increase of the TiO₂ content of chromite, from the base to the top of the ore body (Fig. 7a), cannot result solely from fractional crystallization from a single melt batch. It probably reflects the involvement of several melt injections and of melt fractions of contrasted composition. Additionally, it was shown in the

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

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

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

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

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

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

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

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

Conclusion

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

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

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

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

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Figure captions

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

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

orthopyroxene.

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

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

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

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

Figure 6: Chromite chemical composition from the stratiform chromitite, illustrated in an enlargement of the Cr-Al axis from the ternary Al-Cr-Fe(III) diagram (top) and in the variation of XCr and of the TiO_2 content as a function of XMg and XCr respectively (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. The colours of the dots represent the chromite modal content in each analysed sample, as rare in light green, very scattered in green, scattered in orange, antinodular in red and massive in black.

Figure 7: Vertical evolution of the mineral chemical composition along the studied stratiform chromitite, plotted as a function of the absolute altitude. a) the XCr ($100 \times \text{Cr}/(\text{Cr} + \text{Al})$ atomic ratio), the Cr_2O_3 and Al_2O_3 contents, the XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio), 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) the Fo ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}_{\text{total}})$ atomic ratio) and the NiO and CaO contents of olivine matrix, 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 clinopyroxene matrix, in d) the XMg ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio) and the Al_2O_3 and TiO_2 contents of orthopyroxene in inclusion in chromite grains, in e) the XMg ($100 \times$

Mg/(Mg + Fe²⁺) atomic ratio) and the TiO₂ content of amphibole in inclusion, and in f) the XNa (100 × Na/(Na + K) atomic ratio) of mica in inclusion. Coloured dots are averaged values per sample, depending on the chromite abundance in the sample, while small grey dots show the variability of un-averaged, analysed values. Blue lines are magmatic breccias.

Figure 8: Chemical profiles along the four samples 06OM04-3B2 (a), 06OM04-5D (b) and 06OM04-12E1 (d), all from the cross-section collected along the stratiform chromitite, and 04OM31E (c), collected near the base of the cross-section. For each profile along the thin section (about 3.5 cm in height) the variation of the chromite modal content is represented as well as of the XCr, Cr, Al and Fe³⁺ expressed in *apfu* (atom per formula unit, recalculated on the basis of O = 4), XMg and TiO₂ content of chromite, and of the Fo and NiO content of the surrounding matrix olivine.

Figure 9: Evolution of the Fo of olivine and XMg, XCr, YFe³⁺ and TiO₂ content of chromite in the samples 06OM04-3B2, 06OM04-5D, 06OM04-12E1 and 04OM31E depending on the modal content of chromite.

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

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

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

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- 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 ● antinodal ● massive

(a) 06OM04-3B2

(b) 06OM04-5D

(c) 04OM31E

(d) 06OM04-12E1

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.

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Highlights

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

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Declaration of interest

Declarations of interest: none

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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	Chro	Ol	Opx	Cpx	
mode of					
occurrence	matrix	matrix	inclusion	matrix	inclusion
SiO₂	0.05	40.90	57.42	52.35	52.69
TiO₂	0.34		0.06	0.27	0.28
Al₂O₃	25.89		0.73	2.53	2.38
Cr₂O₃	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
Na₂O			0.03	0.47	0.54
K₂O			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				
YFe³⁺	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		