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Polymineralic inclusions in mantle chromitites from the Oman ophiolite indicate a highly magnesian parental melt

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Abstract

Polymineralic inclusions interpreted as melt inclusions in chromite from the dunitic Moho Transition Zone in the Maqsad area of the Oman ophiolite have been analysed and compositions integrated using a rastering technique on the scanning electron microscope. The inclusions now comprise a range of inter-grown hydrous phases including pargasite, aspidolite, phlogopite and chlorite, indicating that the parental melts were hydrous. Average inclusion compositions for seven samples contain between 23.1-26.8 wt% MgO and 1.7-3.6 wt % FeO. Compositions were corrected to allow for the low FeO concentrations using coexisting olivine compositions. These suggest that the primary melt has between 20-22 wt% MgO and 7-9.7 wt % FeO and has an affinity with boninitic melts, although the melts have a higher Ti content than most boninites. Average rare earth element concentrations suggest that the melts were derived from a REE depleted mantle source although fluid-mobile trace elements indicate a more enriched source. Given the hydrous nature of the inclusions this enrichment could be fluid driven.

An estimate of the melt temperature can be made from the results of homogenisation experiments on these inclusions and suggests ~ 1300 °C, which implies for a harzburgite solidus, relatively shallow melting at depths of less than 50 km and is consistent with a boninitic origin.

The current 'basaltic' nature of the chromite host to highly magnesian melt inclusions suggests that the dunitic Moho Transition Zone operated as a reaction filter in which magnesian melts were transformed into basalts by the removal of high magnesian olivines, particularly in areas where the Moho Transition Zone is unusually thick. We propose therefore that podiform mantle chromitites, even those with an apparent MORB-like chemical signature, have crystallised from a highly magnesian parental melt. The data presented here strongly support the view that this took place in a subduction initiation setting.

1. Introduction

Chromitites in the dunitic Moho Transition Zone of the Oman ophiolite have compositions typical of those formed from a basaltic melt, i.e. $cr\# = 49$ to 58 (Rollinson and Adetunji, 2013), and the standard interpretation of their origin has been their formation in melt channels in the mantle created by the migration of basaltic melts. Here we report chromite grains from mantle chromitites which contain abundant polymineralic inclusions which we interpret as melt inclusions and which have highly magnesian, hydrous melt compositions. This leads us to challenge the prevailing view of a basaltic origin for mantle chromitites and suggest instead that they were formed from an ultramafic melt.

The underlying petrological logic used in this study is based upon the observation that chromite is normally the first phase to crystallise from a melt in mafic and ultramafic magmas. Thus it has the potential to record information about the parental melt, in particular its primary composition and redox state (Irvine, 1965; 1967; Rollinson 2008; Rollinson et al. 2017). This is important in situations where the melt composition has subsequently evolved through fractional crystallisation. In this study, polymineralic inclusions trapped in early crystallising chromite are interpreted as melt inclusions and thus have the potential to provide direct information on the composition of the magma from which the chromite crystallised.

Our observation that melt inclusions in chromite described here are both highly magnesian and hydrous raises two important questions. The first question relates to the immediate host of the chromitites – the dunitic Moho Transition Zone. Typically this is thought to have formed through melt-rock reaction between mantle harzburgite and a basaltic melt (Koga et al., 2001). With the recognition of more magnesian melts in this region of the mantle the conventional basalt-mantle harzburgite melt-rock reaction relationship needs to be revisited. Secondly, it is instructive to consider whether the presence of highly magnesian melts in the mantle are consistent with a mid-ocean ridge model for the Oman ophiolite.

2. Geological background

2.1 The Moho Transition Zone

The samples examined in this study were collected from the Moho Transition Zone of the Oman ophiolite and are from the Maqсад area (Figure 1). Maqсад is in the southern part of the Oman ophiolite and the geology of this region has been well described (Ceuleneer, et al.,

1988; Ceuleneer et al. 1996; Jouselin et al. 1998; Abily and Ceuleneer, 2012; Rollinson and Adetunji, 2013).

The Moho Transition Zone lies at the interface between the crust and mantle sections of the ophiolite and comprises a sequence of dunites (Figure 2). The Moho Transition Zone dunites vary in thickness but may be up to 400m thick in the southern part of the ophiolite (Boudier and Nicolas, 1995; Nicolas and Boudier, 2000). In the north they are even thicker (Goodenough et al., 2010). Abundant chromitite bodies are also found in this horizon and there are some gabbro melt lenses (Jousselin et al., 2012). The Moho Transition Zone has been interpreted as the product of melt-rock interaction between mantle harzburgite and upwelling MORB (Kelemen, et al., 1997; Koga et al. 2001), although the uppermost dunites have been interpreted as an olivine cumulate, part of the lowermost oceanic crust (Abily and Ceuleneer, 2013).

The geology of the Maqсад area has been interpreted as a ridge axis (Ceuleneer, 1991; Ceuleneer et al, 1996) on which has been superimposed an area of mantle upwelling (Figure 1b). This is indicated by steeply dipping lineations (up to 70°) mapped in the mantle harzburgites and is one of a number of zones of mantle upwelling or 'mantle diapirs' which have been mapped in the mantle section of the Oman ophiolite (Nicolas and Boudier, 2000). There is a strong correlation between zones of mantle upwelling and the thickness of the Moho Transition Zone, such that thick mantle dunites (>100m and up to 400m), often containing chromitites, are located structurally above these zones of upwelling. In addition the crustal section of the ophiolite is thinner in these zones of upwelling (Nicolas and Boudier, 2000). Boudier and Nicolas (1995) also described an association between the presence of olivine-clinopyroxene (wehrlite) intrusions in the crustal section of the ophiolite above zones of upwelling and the thick dunite layers.

Chromitites in the Moho Transition Zone at Maqсад form stratiform and dyke-like bodies a few meters thick (Figure 3). They were described in detail by Rollinson and Adetunji (2013) who showed that, whilst superficially they have the geochemical character of MORB-like chromitites ($cr\# = 0.486-0.577$ and high Al_2O_3 , Figure 4), they are more oxidised than typical MORB melts. The chromites in this study also have relatively high TiO_2 concentrations (average values up to 0.5 wt%, Leblanc and Ceuleneer, (1991); Borissova et al. (2012)), similar to the values found for shallow mantle chromitites at Wadi Rajmi in the north

(Rollinson, 2008). Rollinson and Adetunji (2013) also noted that these chromitites contain hydrous melt inclusions, the subject of this present study, and contain small inclusions of more chrome-rich chromites ($cr\# = 0.598-0.697$). On the basis of these observations it was argued that these features are not consistent with the origin of these chromitites at a mid-ocean ridge, but rather they formed in an arc environment, perhaps during subduction initiation.

2.2 Sample selection

The samples used in this study are from two localities in the Maqsad area (Figure 1b). They were previously described by Rollinson and Adetunji (2013) and GPS locations are given in this paper. Most samples are from near to Tuf village near Maqsad, a locality first described by Ceuleneer and Nicolas (1985). Samples from the Tuf area are from a layered chromite-dunite body in which the layering is on the cm to dm scale (Figure 3 a,b). Sample 03-06 is a dunite containing about 20% disseminated chromite. Samples 03-07 and 03-08 are massive chromitites collected about 50 cm above sample 03-06 (Rollinson and Adetunji, 2013, Figure 2c). Sample 04-02 is a disseminated chromitite with equal proportions of chromite and olivine together with some large patches of olivine. Samples 13-02 and 13-03 were collected adjacent to each other from a layered chromitite-dunite sequence made up of alternating, thin (1-2 cm thick) dunite-chromitite bands (Rollinson and Adetunji, 2013, Figure 2b). Sample 13-03 is from an antinodular chromitite layer with equal amounts of chromite and olivine and containing some large olivine patches.

Sample 03-22 is a nodular chromitite from the Maqsad dyke (Figure 3 c,d,e) located 2 km east of Tuf (Rollinson and Adetunji, 2013). The Maqsad dyke is made up of massive chromitite as nodules up to about 1 cm across. Between the nodules are patches of olivine. The nodules have a silicate-rich core and are an aggregate of individual chromite grains 100-200 μm across. This sample is described in detail by Leblanc and Ceuleneer (1991) who showed that elsewhere in this outcrop the nodules are up to 3cm across.

2.3 Previous work

Lorand and Ceuleneer (1989) were the first authors to report in detail inclusions in chromites from a Maqsad dyke. Schiano et al., (1997) interpreted them as hydrous melt inclusions and carried out step heating and homogenisation experiments between 1050-1300 $^{\circ}\text{C}$ and showed

that the homogenised inclusions had compositions similar to those of primitive basaltic melts with 47.2 wt % SiO₂, 11.6 wt % MgO, mg#=64-67 and depleted MORB-like REE patterns. These findings were criticised by Arai (1998) who suggested by analogy with melt inclusions in chromite from the Hess Deep that the mineralogical variability between the inclusions indicated the presence of multiple melts created through the mixing of two different initial melts. However, Schiano et al., (1998) argue that the variability between inclusions is an artefact of the sectioning process and that the inclusions homogenise to a single melt composition.

More recently Borisova et al. (2012) conducted homogenisation experiments at 1300 °C and found a wide variety of compositions but with an average basaltic composition, similar to that of Schiano et al. (1997) with 47.3 wt % SiO₂, 10.5 wt % MgO, although enriched in FeO (11.1 wt %) and depleted in CaO (7.6 wt %) relative to N-MORB. However, they conclude that these inclusions represent the entrapment of a variety of silicate phases. They do not interpret their data in detail but suggest that these polyminerale inclusions may not be fully equilibrated and that the high Fe content may imply diffusive input from chromite indicating some equilibration with the host. Thus the measured compositions might be, in part, an artefact of the homogenisation experiment.

We follow Schiano et al., (1997; 1998) in interpreting these polyminerale inclusions as former, fully molten melt inclusions, now crystallized to solid mineral phases. This is based upon

- the morphology of the inclusions which show rounded and negative crystal shapes;
- the observation that they appear as clusters in the core of some grains and in the rims of large grains as is found with melt inclusions (Figure 5);
- the internal morphology of the inclusions shows that the phases have crystallised together (see Supplementary Figure 2). This textural evidence strongly indicates that the inclusions were originally fully molten.
- some inclusions contain small rounded sulphide grains 2-3 µm across composed of Ni-Fe sulphide with a small amount of Cu. These were also noted by Borisova et al. (2012) and are thought to represent immiscible sulphide droplets in the melt inclusion. The existence of these sulphide droplets also supports the view that the inclusions were initially fully molten.

- the presence of highly magnesian (up to 23 wt % MgO) glassy melt inclusions in chromite from Japan (Umino et al., 2015) shows that inclusions of the type described here can exist as melt inclusions in chromite.

However there are also solid mineral inclusions within our inclusion suite and these include the anhydrous phases olivine, orthopyroxene and clinopyroxene. These phases are rarely observed within the polymineralic inclusions in our study, although they are reported by Borisova et al (2012). Thus it is possible that in a few inclusions analysed here, there are included solid phases within the melt. If this is the case then some reported inclusion compositions will be slightly more magnesian than they should be. However, this effect is minimised in the averaging process described below.

The presence of hydrous phases in the mantle section of the Oman ophiolite is also well known (Lorand and Ceuleneer, 1989; Ahmed and Arai, 2002; Rollinson, 2008). There are two main views on the origin of the water. Many authors have suggested that the water was derived from below, i.e. from the dehydration of a subducting slab (Schiano et al., 1997, Ahmed et al., 2006) or from serpentinised oceanic lithosphere (Borisova et al., 2012). In contrast other authors have suggested that the water was sea-water introduced from above and penetrated through deep hydrothermal circulation (Augé and Johan, 1988). The most recent advocates of this view are Rospabé et al. (2017) and Dygert et al. (2017). Rospabé et al. (2017) argue that the presence of hydrous mineral phases both as inclusions in chromite and interstitial in the dunites is the product of the mixing between upwelling basaltic melts and the downward movement of supercritical water. Dygert et al. (2017) argue from the very rapid cooling rates in the mantle section of the ophiolite that cooling must have been facilitated by deep hydrothermal circulation.

3. Methodology

3.1 Melt inclusions

Melt inclusions in the chromite grains examined in this study are 10-300 µm diameter and vary in shape from rounded to polygonal. They are now universally made up of an aggregate of mineral phases (Figure 5). As noted by Borisova et al. (2012) these include the hydrous phases pargasite, phlogopite and Na-phlogopite (aspidolite) together with olivine,

orthopyroxene and clinopyroxene. In addition we note the presence of chlorite and grossular (see also Rosapbé et al., 2017) with minor occurrences of magnesite, titanite and apatite.

Some inclusions have a rim of chromite (Figure 5) which is more Cr-rich ($cr\# = 0.6-0.7$) than the enclosing chromite grain ($cr\# = 0.49-0.58$). There is no obvious pattern to either the size or distribution of these particular inclusions and they are not fundamentally different in composition from those which are not associated with high-Cr chromite grains apart from slightly lower CaO and FeO concentrations relative to the inclusion population as a whole. These Cr-rich chromite grains tend to form euhedral grains within the host chromite. It is possible that these grains represent early-formed chromite crystals containing trapped melt which are included in later chromite growth (see Figure 5 and Rollinson and Adetunji (2013)).

Melt inclusions are susceptible to post-entrapment modification due to the diffusive re-equilibration of the trapped melt with the host mineral. This problem is of particular concern for melt inclusions in olivine in mafic and ultramafic rocks where there may be significant loss of Fe to the host. However, Hanski and Kamanetsky (2013) and Umino et al. (2015) argue that this is much less of a problem in melt inclusions in chromite. We have already described previous attempts to obtain the bulk composition of the melt inclusions from Maqsad by Schiano et al. (1997) and Borisova et al. (2012) using high temperature homogenisation experiments. These show that the primary melt compositions were basaltic, although it is possible that during the homogenisation experiment the inclusions equilibrated with their host and the composition of the inclusion was modified.

In this study therefore, a different approach was adopted to obtain a more accurate estimate of the original composition of the melt inclusions. The compositions of the inclusions were measured using a rastering technique to obtain an 'average area' on the scanning electron microscope (SEM). For each inclusion we defined the area of the inclusion as close to the rim as possible and measured the composition. We used a counting time of 100 seconds deadtime. In this way we avoided any high temperature modification of the inclusion composition and loss of volatiles from the electron beam. Clearly, the surface expression of individual inclusions in a polished thin section is not representative, for the inclusions now represent a composite aggregate of mineral phases which has been cut through at random. Hence the principal logic of this method is statistical. We have assumed that the melt

composition found as inclusions in any one sample is constant and therefore have measured a large number of inclusions in each rock in order to obtain a representative estimate of the composition of the original melt in that sample. We have allowed for the fact that melt compositions might vary between samples.

Images were made on a Zeiss Leo 1450VP scanning electron microscope. Quantitative results were obtained on this instrument using an energy dispersive Oxford Instruments X-Max 20 silicon detector together with Inca software. The operating conditions were 20 kV, 50 μ A beam current with a nominal beam size of 1 μ m.

We find a wide disparity between the results presented here and previous high temperature homogenisation studies and we propose, as indicated above, that there was high temperature re-equilibration of the homogenised melt with the host chromite grain during the heating experiments, in particular the diffusive Mg-Fe exchange from melt to the chromite. Our new results suggest that the melt inclusions are more magnesian but Fe-poor relative to the compositions determined in the homogenisation experiments. Further support for this finding is the highly magnesian nature of all the constituent silicate and oxide phases which make up the inclusions. These compositions when integrated predict a highly magnesian melt composition as is found here.

3.2 Data quality

We measured the olivine standard MGK3 as an unknown, using the same rastering technique that we used for our melt inclusions and found that our method over-reads SiO₂ by 2.8% relative, under-reads MgO by 0.7% relative and under-reads FeO by 4.4% relative. Individual point analyses of chromite measured on sample 03-08 also under-read FeO by 4.7% relative and in this case over-reads MgO by 5.4% relative.

It will be noted that some of the melt compositions reported in Table 1 have low totals. These have been retained rather than artificially recasting the result to 100% because it can be seen from the composition of the mineral phases which now make up these inclusion that many are hydrous. Thus at least some of the additional material is water. However, some inclusions may have low totals in part due to surface defects created during polishing. The data were screened as follows.

- Samples with very low totals were removed from the dataset on the basis that they represent analytical error. Often low totals correlate with low SiO₂, indicating a poor analysis. Thus individual raster analyses had totals between 84.3-100.1 wt % oxide, although the averages for each sample had totals in the range 92.0-96.5 wt % oxide (Table 1).
- Individual analyses with high Cr₂O₃ were removed from the data set on that basis that the analysis had accidentally included some Cr₂O₃ from the chromite host. Indeed average Cr₂O₃ values are high, between 2.38-3.64 wt % indicating some post entrapment redistribution of Cr between the host mineral and inclusion (Hanski and Kamanetsky, 2013)
- Inclusions where the composition can be reduced to a single phase such as clinopyroxene or olivine were also removed on the assumption that these are probably solid mineral inclusions rather than melt inclusions.

All the melt inclusions measured have a low FeO content for a mafic or ultramafic liquid, with average compositions in the range 1.72-3.59 wt %. Given the possibility for Fe to diffuse in a chromite host we examined the potential for Fe-loss through diffusion carefully. In Supplementary Figure 1 we show back-scatter images for four inclusions from samples 03-07 and 03-22 and element maps for each inclusion for FeK α 1 in order to investigate whether or not the chromite host shows a zone of Fe-enrichment adjacent to the inclusion indicative of Fe-diffusion out of the inclusion. No such rims are visible on the ten elemental maps that we made. In addition we measured diffusion profiles across grains making analyses at between 1-3 μ m intervals. Of six profiles made on inclusions in sample 03-22 only one inclusion shows a possible Fe-enrichment zone of about 5 μ m width with a 0.7 wt% increase in elemental Fe although this enrichment is not visible on the elemental FeK α 1 map.

On this basis we conclude that the low Fe-content of the inclusions is not primarily due to Fe diffusion into the chromite host, but rather is primary-magmatic. Also, as noted above relative to the olivine standard FeO is under-read in our samples by 4.4% relative, which means that our measurements might be up to 0.2 wt % lower than they should be for FeO. This small difference indicates that the 'measured' low FeO of the melt inclusions is real.

4. Results

4.1 Melt inclusion major element compositions

Here we present results for seven samples of chromitite collected from the Maqсад area of the Oman ophiolite. Six of the seven samples come from the Tuf locality, a stratiform chromitite-dunite body within the dunitic Moho Transition Zone. Sample 03-22 comes from the Maqсад dyke locality a 5m-wide dyke containing chromitite nodules ca 1 cm in diameter in a dunitic matrix. The petrology of these samples was described in detail by Rollinson and Adetunji (2013), although it should be noted that in the original description of this locality by Leblanc and Ceuleneer (1991) they also found nodules up to 3 cm across in a troctolite matrix. In total we have measured 296 melt inclusions from which we have derived seven average values, one for each rock.

We present a summary of our average data compositions in Table 1 and in Figure 6. The average compositions vary between 23.05-26.84 wt % MgO (average 25.59 wt%), 39.44-44.78 wt % SiO₂ (average 43.21 wt%), but have very low FeO – 1.72-3.59 wt % (average 2.58 wt %). When compared with the melt homogenisation data Borisova et al. (2012) our data show that the melt inclusions are much more magnesian and Fe-poor. As noted Cr₂O₃ values are probably too high indicating some post entrapment redistribution of Cr between the host mineral and inclusion (Hanski and Kamanetsky, 2013)

4.2 Melt inclusion trace element compositions

We analysed 28 melt inclusions in sample 03-08 for the following trace elements – Cs, Rb, Ba, Th, Sr, La, Ce, Gd, Yb using laser ablation ICP-MS. The details are given in the supplementary information. Our results are shown in Figure 7 as primitive mantle normalised plots for all nine elements and as chondrite normalised plots for the four rare earth elements (La, Ce, Gd, Yb). The results show a range of measured compositions reflecting the varied mineralogy of the ablated area in the near-surface area of the inclusions and for this reason, as with our major element data, we have taken the mean value of all our measurements as indicative of the former melt composition. Our results are similar to those previously measured by ion microprobe by Schiano et al. (1997), although the middle and heavy REE concentrations are lower (Figure 7(a)). However, when our results are compared with melt compositions recovered from inverted REE in clinopyroxene, also obtained by ion microprobe (Koga et al., 2001), there are a number of differences. Melt compositions calculated by Koga et al. (2001) have relatively high total REE concentrations compared to the 2-3 x chondrite found here. The Koga et al. (2001) data also show great variability in light

REE concentrations from depleted to enriched relative to the middle and heavy REE whereas our average data have a relatively flat REE pattern with slight depletion in the middle REE (Figure 7(b)). Koga et al. (2001) suggested that the REE in their calculated melt compositions was MORB-like, although a variety of MORB-types are preserved. In contrast, the melt inclusion data reported here are not MORB-like in their REE systematics.

4.3 Former melt compositions

In plots of the average major element compositions there are weak negative correlations between 'MgO and CaO' and 'MgO and FeO' and a weak positive correlation between MgO and Al₂O₃ (Figure 6). These correlations, together with the very low FeO contents indicate that the average melt compositions may not be primary compositions but modified through crystal fractionation. On a MgO-FeO plot the average melt compositions lie on a common trend with the early-formed chromite compositions (data from Rollinson and Adetunji, 2013), i.e. those which occur as inclusions within the host chromite (Figure 8). This relationship strongly suggests that the FeO content of the average melt compositions as determined in this study has been reduced by the removal of chromite from the primary melt.

We have combined the data from this trendline with the composition of the coexisting unmodified olivines in dunites associated with the chromitites in order to obtain an estimate of the MgO and FeO content of the parental melt. The dunites are interlayered with the chromitites, but are sufficiently distant from them for the olivine compositions to have not equilibrated with the chromite. Since they are dunites containing no other phases the olivine compositions have retained their primary igneous compositions. Coexisting olivines have compositions in the range Fo=91.5-94 (data from Rollinson and Adetunji, 2013) and using the recently re-determined olivine-melt Fe-Mg partition coefficient of 0.34 (Matzen et al., 2011) it is possible to calculate the FeO and MgO content of their equilibrium liquids. On Figure 8 we show the intersection of the olivine-melt equilibria and the chromite-melt inclusion trendline indicating that the parental melts had compositions in the range MgO = 20-22 wt % and 7-9.7 wt% FeO. The reduction in FeO content of the melts from the primary melt composition to the observed suggests that some chromite was removed.

Constraining other elemental concentrations in the parental melts is more difficult, however, we have adjusted the melt compositions to MgO₂₀ and MgO₂₂ and in Figure 9 plot our average compositions together with a range of high-MgO melts from other sources. These

include high MgO basalts and picrites from western Greenland (Larsen and Pedersen, 2009), komatiites from Gorgona Island (Kerr et al., 1996), boninites and high magnesian andesites from Tonga (Falloon et al., 1989; Sobolev and Danyushevsky, 1994) and boninites from the Bonin archipelago (Umino et al. 2015). We also plot the glass compositions for 1.5 and 2.0 GPa from the harzburgite melting experiments of Falloon and Danyushevsky (2000), which for the most part plot in the boninite field. Our estimated melt compositions for MgO_{20} and MgO_{22} are given in Table 1. For the most part these estimated compositions plot in the boninite field on Figure 9 (although TiO_2 may be too high (Georges Ceuleneer pers com)) indicating that the melt compositions found as inclusions in the chromitites of the Moho Transition Zone of the Oman ophiolite are of boninitic affinity.

The trace elements show low REE concentrations (ca 2-3 x chondrite) with a slight depletion in the middle REE implying their derivation from a depleted mantle source and as noted above a source different from that of MORB. On a primitive mantle normalised plot REE concentrations are low (ca 1x PM concentrations) whereas the fluid-mobile elements Cs, Rb and Ba have higher concentrations (up to 10 x PM) which could imply the derivation of the fluid-mobile elements from a different, more enriched source. It is possible therefore that the high magnesian melts reported here were derived by the melting of a depleted mantle source which was re-enriched by fluid-borne LILE. These features imply an origin different from that of a MORB-source.

5. Discussion

Our data presented here indicate that the melts associated with chromite genesis in the Moho Transition Zone were highly magnesian with a MgO content between 20-22 wt % and a FeO content between 7-9.7 wt %. The primary melt compositions that we have been able to estimate are very similar to those of boninitic affinity. Previous authors have also suggested that melts associated with the Moho transition zone were highly magnesian. For example, Abbou-Kebir et al. (2011) suggested that late highly magnesian dunitic veins (Fo_{91-94}) devoid of chromite and cross-cutting chromitites from Wadi Rajmi in northern Oman were derived from a komatiitic parental melt. At Maqсад, Abily and Ceuleneer (2013) suggested Mg-rich melts as parental to olivine cumulates found at the top of the Moho Transition Zone, although they proposed a melt MgO content of ~ 15 wt %.

Our REE data indicate that the melts were derived from a depleted source (Figure 7b), which in the Oman context could indicate a harzburgitic source. The higher concentrations of fluid mobile trace elements would suggest a source which has been enriched or re-fertilised with the elements Cs, Rb, Ba and Sr (Figure 7a). This enrichment process could be through the presence of subduction zone fluids. Thus the trace element data are consistent with a boninitic melt.

5.1 Conditions of melting

Using the semiquantitative geothermometer of Matzen et al. (2011) for Hawaiian magma 'compositions' the melt compositions calculated for the melt inclusions discussed here of MgO = 20-22 wt% have melt temperatures in the range 1430-1470 °C. However, Falloon and Danyushevsky (2000) have shown that the solidus temperature of harzburgite is lowered during hydrous melting and that, for example in a melt with 24% MgO and 2 wt % water, this can be by as much as 100 °C. This would bring the melt temperatures closer to 1300 °C, the homogenisation temperatures reported by Schiano et al. (1997) and Borisova et al. (2012). Falloon and Danyushevsky (2000) have recorded a harzburgite solidus at ca 1500 °C and 1.5 GPa and so at the lower temperatures indicated by these melts indicate relatively shallow mantle melting at depths of less than 50 km.

5.2 Tectonic setting

Given the boninitic affinity of these melts it is pertinent to note that Falloon and Danyushevsky (2000) and Parman and Grove (2004) show that high-Ca boninites are most likely the product of the water-undersaturated melting of mantle harzburgite contaminated with a 'subduction zone component'. They show that experimental melts in equilibrium with depleted harzburgite overlap the compositions of high-Ca boninites from Tonga with 19-24 wt % MgO, although their results suggest higher temperatures (1430-1470 °C) than estimated in this study.

Whattam and Stern (2011) have argued that boninites are produced in association with subduction initiation and fore-arc spreading and are indicative of supra-subduction processes and the tectonic constraints outlined here lend further support to the idea that the Oman ophiolite formed in an environment akin to subduction initiation.

5.3 The origin of the Moho Transition Zone

A particular paradox in this study is the observation that the host chromite grains have compositions which are consistent with their equilibration with a basaltic, MORB-like parental melt and yet these same grains contain melt inclusions which are high MgO in character. Resolving this paradox requires a revision in the way we think about the Moho Transition Zone in the Oman ophiolite. The thick dunitic sequence of the Moho Transition Zone has long been regarded as the product of melt percolation and melt reaction (Koga et al., 2001). The standard model has been that of the percolation of a basaltic melt through depleted mantle harzburgite creating a dunitic residue. Here it is proposed on the basis of the high magnesian melt inclusions and the presence of high magnesian olivines that, at least in some cases, the initial percolating melt was ultramafic, but that the process of melt reaction transformed the melt composition into a basaltic composition. Thus a high MgO-melt was supplied at the base of the Moho transition zone but a basaltic melt was produced at the top. A possible reaction would be

in which a boninitic melt reacts with mantle harzburgite, dissolves the orthopyroxene and precipitates chromite (as discussed below) and a more magnesian olivine. In this way the MTZ operates as a reaction zone filter converting magnesian melts into melts of basaltic composition. This process may be more common above areas of mantle upwelling (mantle diapirs), such as at Maqсад, where the greater supply of melt would lead to a thicker dunitic Moho transition zone containing areas of highly magnesian olivine.

5.4 The origin of mantle-hosted chromitites

Irvine (1977) in a seminal study showed that chromite crystallisation in a mafic or ultramafic melt may be triggered by the addition of a siliceous component to the melt. This has the effect of moving the melt composition from the olivine-chromite cotectic boundary into the field of chromite crystallisation. A number of different models of siliceous contamination have been proposed - from the assimilation of a felsic melt into a mafic magma (Irvine, 1977), to melt-rock reaction between a mafic melt and orthopyroxene-rich mantle harzburgite (Arai and Miura, 2015). These competing approaches have led to two of different models for the origin of melt inclusions in the Maqсад chromitites, in particular to explain the origin of the water. For example Rosabé et al. (2017) argue that a hydrous mafic melt was created by

the mixing of a mafic melt with ‘supercritical hydrothermal fluids rich in silica, ie trondhjemitic melt issued from the hydrous melting of the country rocks’. This model combines both the need for a siliceous component to trigger the precipitation of chromite and water to produce a hydrous melt from which hydrous phases have crystallised. Whilst this model is elegant in its simplicity our field observations in the Oman ophiolite suggest that where trondhjemitic and mafic melts coexist they do not mix but rather form a ‘vinaigrette’ structure – a term coined by French workers in Oman to describe the lack of mixing between oil and vinegar in a salad dressing.

In contrast, in this study we solve the problem of chromite crystallisation in two steps. First, we have shown that our data require a hydrated melt with a high MgO-content. Secondly, we argue that chromite crystallisation is triggered by reaction between that hydrous melt and orthopyroxene in mantle harzburgite, in part following the model of Arai and Miura, (2015). We have argued that this approach is consistent with the chemical data obtained on the inclusions themselves. A further advantage of this model is that it simplifies the problem of from where the large volumes of chromium might be derived. Current models for the origin of mantle chromitites have to presume enormous volumes of basalt fluxing through the mantle, each melt batch crystallising a small volume of chromite. However, if the parental melt were more magnesian, the scale of this problem is reduced. An average Cr concentration for MORB is ~ 250 ppm (Gale et al., 2013), whereas a highly magnesian boninite is ~ 950-1150 ppm (Falloo et al, 2008; Li et al., 2013). Thus the flux of a more magnesian melt might reduce the required melt volume by 4 or 5 times.

6. Conclusions

The principal findings of this study are that

- Mantle podiform chromitites at Maqsad formed from a highly magnesian parental melt with ~ 20-22 wt % MgO;
- Trace element concentrations indicate a source depleted in the REE, but more enriched in fluid mobile trace elements.
- The presence of hydrous mineral phases present in these now recrystallised melts indicates that they were originally hydrous;

- The major and trace element chemistry and the observation that the melt was hydrous all support the view that these melts are boninitic in origin and probably of high-Ca boninite parentage and formed by melting of mantle harzburgite at ~ 1300 °C;
- The current ‘basaltic’ nature of the chromite host (Rollinson and Adetunji, 2013) and the highly magnesian nature of the melt inclusions suggests that the dunitic Moho Transition Zone operated as a reaction filter in which magnesian melts were transformed into basalts, particularly in areas where the Moho Transition Zone is unusually thick.

We propose therefore that podiform mantle chromitites, even those with an apparent MORB-like chemical signature, have crystallised from a highly magnesian parental melt. The data presented here strongly support the view that this took place in a subduction initiation setting.

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Figure captions

1. Geological map of the study area showing (a) the location of Maqсад in the context of the Oman ophiolite; (b) the geology of the Maqсад area showing the samples sites (after Rollinson and Adetunji, 2013).
2. Schematic stratigraphic section of the Oman ophiolite showing the location of the Moho Transition Zone. This transition zone is dunitic in composition, ranges in thickness from a few metres to several hundred metres and is the host to podiform chromitites.
3. Field photographs of chromitites from the Moho Transition Zone at Maqсад. (a-b) layered chromitite-dunite units at Tuf; (c) chromitite dyke at Maqсад, dunite on the right; (d) detail of photograph in (c) showing nodules of chromitite; (e) chromitite nodules from the Maqсад dyke seen in thin section; typically nodules are about 1 cm in diameter.
4. $\text{TiO}_2\text{-Al}_2\text{O}_3$ plot of Maqсад chromites showing that most grains in the chromitite host have compositions consistent with a MORB origin (after Kamenetsky et al., 2001). Chromite compositions are from Rollinson and Adetunji (2013)
5. Scanning electron microscope images of inclusions. (a, b) zone of inclusions at the margin of chromitite nodules in Maqсад dyke sample 03-22; (c) inclusion cluster in chromite in dunite 03-06; (d) isolated inclusions in chromitite 03-08, some with Cr-rich chromite rims; (e, f) inclusions with individual phases identified in chromitite 03-07.
6. MgO-plots for individual inclusions (black circles) and averages for each of seven rock samples (yellow circles)
7. (a) Primitive mantle normalised plots for 28 melt inclusions in sample 03-08 (shaded yellow field) and the average composition shown as the black line. There is some similarity with these compositions and those obtained by Schiano et al. (1997) on melt inclusions from Maqсад (shaded green field), although in this study the heavy REE concentrations are lower.
(b) Chondrite normalised rare earth element plots for 28 melt inclusions in sample 03-08 (shaded yellow field) and the average composition shown as the red line. These compositions are different from melt compositions calculated from the REE content of

- interstitial clinopyroxenes in Oman mantle harzburgites (shaded green field and black and grey lines, Koga et al., 2001).
8. FeO and MgO compositions for melt inclusion averages (yellow circles), the early formed chromites found as inclusions in the host chromite (open circles), calculated compositions for melts in equilibrium with olivines Fo₉₃ and Fo₉₄ (black circles) and the estimated primary melt compositions (red squares).
  9. MgO plots for average melt compositions (small red circles) and the calculated primary melt composition (larger red circles) compared to the composition of boninites (black circles), Gorgona komatiites (yellow circles), basalt-picrite (grey circles) and melts of harzburgite (green circles) (for data sources see text). Note: in most plots the basalt-picrite field and the komatiite field overlap.

Tables.

Table 1. Average melt compositions and calculated primary melt compositions.

Table 1. Average melt compositions and calculated primary melt compositions

Averages primary melt									Calculated	
	03-06	03-07	03-08	03-22	04-02	13-02	13-03	average		
n	47	25	35	23	48	71	47	296	MgO20	
	MgO22									
MgO	23.05	26.24	26.57	25.30	25.18	26.84	25.95	25.59	20.00	22.00
SiO2	44.78	43.93	47.73	40.85	43.88	39.44	41.87	43.21	51.00	50.00
TiO2	2.26	2.20	2.78	1.13	1.98	1.49	1.63	1.93	1.00	1.00
Al2O3	10.06	12.88	11.27	12.59	8.96	14.47	14.07	12.04	7.00	8.70
FeO	3.59	1.72	2.19	3.44	3.01	2.24	1.91	2.58	9.70	7.00
MgO	23.05	26.24	26.57	25.30	25.18	26.84	25.95	25.59	20.00	22.00
CaO	6.74	2.79	2.63	4.65	2.85	4.28	3.65	3.94	8.40	7.20
Na2O	2.81	2.83	3.64	1.97	2.67	1.84	2.84	2.66	1.00	1.00
K2O	0.36	0.38	0.59	0.10	0.63	0.70	0.70	0.49	0.20	0.20
Cr2O3	2.38	2.57	2.53	2.97	2.71	3.51	3.64	2.90	1.00	1.00
NiO	0.10	0.44	0.00	0.05	0.13	0.47	0.28	0.21	0.05	0.05
tot	96.13	95.99	99.93	93.06	92.00	95.27	96.52	95.56	99.35	98.15

Note: the oxide compositions for the calculated primary melts shown in italics have a higher degree of uncertainty.

Highlights

- Chromitites at Maqsad in the Oman ophiolite appear to have formed from basaltic melts
- Melt inclusions in chromite are highly magnesian, 20-22 wt % MgO
- Compositionally the melt inclusions are of boninitic character
- the moho transition zone formed by melt reaction with an input of high MgO melts and a basaltic output

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9