High-pressure experiments provide insights into the Mantle Transition Zone history of
chromitite in Tibetan ophiolites

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16 Abstract

17 The chromitites in the Luobusha ophiolite (Tibet) have been proposed as
18 messengers from the deep mantle. The exsolution of diopside, MgSiO_3 and coesite in
19 chromite from these bodies has been used to argue that the chromite previously
20 existed as a phase with the CaFe_2O_4 structure (idealized composition $(\text{Mg}, \text{Fe})(\text{Cr},$
21 $\text{Al})_2\text{O}_4$), an indicator of Mantle Transition Zone (MTZ) conditions. However, evidence
22 about the stability of the CaFe_2O_4 -structured phase and its formation mechanism are
23 limited. Here we present experimental evidence for the depth of metamorphism and
24 subsequent petrogenesis of the podiform chromitite. The CaFe_2O_4 -structured phase
25 was observed at $\sim 14\text{--}18$ GPa, and can contain several weight percent of CaO and SiO_2 .
26 Partial-melting experiments show that chromite cannot be formed through direct
27 crystallization in the MTZ. Therefore, we suggest that the Tibetan chromitites formed
28 under shallow conditions, were then subducted, and were metamorphosed near the top
29 of MTZ. During the tectonics/buoyancy-driven ascent of the enclosing peridotites, the
30 CaFe_2O_4 -phase transformed to chromite at depths of ~ 400 km, accompanied by the
31 simultaneous exsolution of diopside and other phases, which were then preserved
32 during transportation to shallow depths. These observations provide a new window
33 into the processes of mantle geodynamics, and constraints on mantle convection in
34 major collision zones.

35

36 Introduction

37 Podiform chromitites in the Luobusha ophiolite (southeastern Tibet) are typically
38 lens-shaped and surrounded by dunite envelopes, enclosed in well-preserved depleted
39 harzburgites. The chromitites consist mainly of chromite, associated with forsterite $\pm$
40 diopside. Most chromite grains are chemically homogenous and have high Cr# (molar
41 $Cr/(Cr+Al)$ and Mg# (molar $Mg/(Mg+Fe)$) of ~ 0.7 - 0.8 and ~ 0.70 - 0.76 , respectively
42 (e.g., [Zhou et al., 2014](#); [Zhang et al., 2016](#)). Previous studies have shown that these
43 chromitites can be explained by interaction between mantle peridotite (especially
44 harzburgite) and melt in the upper mantle (less than ~ 30 km depth) ([Arai and](#)
45 [Yurimoto, 1994](#); [Zhou et al., 1996](#); [Arai, 1997](#)), or partial melting of mantle peridotite
46 and crystal fractionation at shallow depths ([Jin et al., 1996](#); [Edwards et al., 2000](#)), or
47 mixing of mafic magmas ([González-Jiménez et al., 2014](#)). However, discoveries of
48 ultrahigh-pressure (UHP) mineral inclusions (e.g., coesite, diamond) within the
49 chromitites have suggested a much deeper origin for both the chromitites (“UHP
50 chromitites”), and their host rocks (no less than ~ 120 km) ([Yang et al., 2007](#)). Recent
51 studies have also revealed numerous exsolution lamellae of diopside and an $MgSiO_3$
52 phase within chromites in the podiform chromitites from Tibet and Oman ([Yamamoto](#)
53 [et al., 2009](#); [Miura et al., 2012](#); [Satsukawa et al., 2015](#)). Previous experimental studies
54 have shown that only trace amounts of SiO_2 (less than ~ 1 wt%) can be incorporated
55 into chromite at pressures less than 6 GPa and temperature ranges of 1000-1500 °C
56 ([Brey et al., 1999](#); [Klemme and O'Neill, 2000](#); [Girnis et al., 2003](#)). Recent
57 experimental studies also show no obvious increase of SiO_2 in chromite with pressure

up to 12 GPa at 1400-1600 °C (Xu, 2011; Jin et al., 2014; Wu et al., 2016). Thus, the exsolution of diopside ($\text{CaMgSi}_2\text{O}_6$), MgSiO_3 and coesite/quartz in chromite may suggest the previous existence of a CaFe_2O_4 -phase (CF phase, idealized composition $(\text{Mg}, \text{Fe})(\text{Cr}, \text{Al})_2\text{O}_4$), stable at pressures of >12.5 GPa (Chen et al., 2003), which was then replaced by chromite with exsolution of diopside and other phases during mantle upwelling (Yamamoto et al., 2009; Arai, 2013; McGowan et al., 2015). However, little is known about the stability field of the CF phase at realistic mantle temperatures, and how much CaO and SiO_2 can be incorporated into chromite and its high-pressure polymorphs.

Recent experimental studies by Ishii et al. (2014, 2015) have shown that in the FeCr_2O_4 system the CF phase can only be observed at pressures above ~17-18 GPa and temperatures below 1300 °C. However, the temperature at the base of the upper mantle is thought to be ~1400 °C (Ito and Katsura, 1989), and the temperature of a mantle plume rising adiabatically from these depths is thus ~200 °C higher than normal upper mantle (Herzberg et al., 2007). Thus, whether the observed diopside, MgSiO_3 and coesite exsolution within chromite can be thought of an indicator of a Mantle Transition Zone (MTZ) origin is debatable.

In this study, we use new experimental data in the magnesiochromite + diopside system to constrain the depth of origin of the Tibetan chromitites. Here, magnesiochromite, instead of natural chromite (with small amounts of iron), was used as the starting material; iron was omitted to avoid complexity. This system is relevant because mineral separates of these chromitites contain abundant large (50-250 μm)

grains of diopside with crystal faces controlled by chromite, implying equilibration in a diopside-rich system (Griffin et al., 2016; Zhang et al., 2016). We have observed the formation of the CF phase at pressures corresponding to the top of the MTZ. The solubility of CaO and SiO₂ in magnesiochromite and its high-pressure polymorphs provides direct experimental constraints on the depth of metamorphism of the UHP chromite. Furthermore, partial-melting experiments in the pyrolite+chromite system have been used to test petrogenetic models for the UHP chromite.

Methods

2.1 Starting materials

Two types of magnesiochromite [$Mg(Cr_{1.5}Al_{0.5})O_4$ and $MgCr_2O_4$] were prepared from stoichiometric mixtures of reagent grade MgO, Cr₂O₃ and Al₂O₃. Cr# of the starting magnesiochromite [$Mg(Cr_{1.5}Al_{0.5})O_4$] was set at 0.75, consistent with that of the natural chromite from Luobusha (Zhou et al., 2014; Zhang et al., 2016). The oxide mixtures were first crushed and mixed to form fine powders, and then heated at 1000 °C for 24 h. After heating, the products were confirmed to consist of single phases by observations in a field-emission scanning electron microscope (FE-SEM), fitted with an energy-dispersive spectrometer (EDS). CaMgSi₂O₆ glass was synthesized from a mixture of CaCO₃, MgO and SiO₂ with a 1:1:2 molar ratios by heating at 1350 °C for 0.5 h. Three starting materials, *Mixes A*, *B* and *C*, were synthesized and used in the present study. *Mixes A* and *B* were synthesized by mixing 90 mol.% $Mg(Cr_{1.5}Al_{0.5})O_4$ + 10 mol.% CaMgSi₂O₆, and 95 mol.% $Mg(Cr_{1.5}Al_{0.5})O_4$ + 5 mol.% CaMgSi₂O₆,

respectively. *Mix C* was synthesized by mixing 90 mol.% MgCr_2O_4 + 10 mol.% $\text{CaMgSi}_2\text{O}_6$. The two types of magnesiochromite [$\text{Mg}(\text{Cr}_{1.5}\text{Al}_{0.5})\text{O}_4$ and MgCr_2O_4] and $\text{CaMgSi}_2\text{O}_6$ glass were first crushed and ground into very fine powder (grain size $\sim < 5$ μm) separately, and then ground together in alcohol with an agate mortar to produce homogeneous mixtures.

The pyrolite+chromite starting materials were synthesized by mixing 95 wt% pyrolite + 5 wt% chromite (*PC-A*) and 90 wt% pyrolite + 10 wt% chromite (*PC-B*), respectively. The pyrolite was synthesized by mixing approximately 53 wt% olivine + 31 wt% enstatite + 14 wt% clinopyroxene + 2 wt% spinel. The average chemical composition of this synthetic pyrolite is (in wt %): $\text{SiO}_2 = 46.29$, $\text{TiO}_2 = 0.09$, $\text{Al}_2\text{O}_3 = 3.87$, $\text{Cr}_2\text{O}_3 = 0.51$, $\text{MgO} = 38.67$, $\text{CaO} = 3.09$, $\text{MnO} = 0.12$, $\text{FeO} = 7.17$, $\text{Na}_2\text{O} = 0.18$ and $\text{K}_2\text{O} = 0.01$. The chromite was hand-picked from a natural chromitite sample from Luobusha, Tibet. Its average chemical composition is (in wt%): $\text{SiO}_2 = 0.07$, $\text{TiO}_2 = 0.24$, $\text{Al}_2\text{O}_3 = 10.33$, $\text{Cr}_2\text{O}_3 = 59.96$, $\text{FeO} = 11.98$, $\text{MnO} = 0.22$, $\text{MgO} = 16.74$, $\text{ZnO} = 0.14$ and $\text{NiO} = 0.25$, with Cr# of 0.76 and Mg# of 0.79. Detailed microscopic study did not show any pyroxene or coesite exsolution lamellae in the chromite grains (Xu, 2011; Wu et al., 2016).

2.2 Experimental procedures

High-pressure, high-temperature (HPHT) experiments were carried out in a 1000-ton Walker type multi-anvil press at the Laboratory for the Study of the Earth's Deep Interior (SEDI-Lab), China University of Geosciences. Cell assemblages with

10 and 8 mm octahedral edge lengths and 5 and 3 mm truncation edge lengths, respectively, were used (10/5 and 8/3 cell assemblies). The detailed descriptions and pressure calibrations of the cell assemblies were reported by [Zhang et al. \(2013\)](#). A type C thermocouple was used to measure the temperature at the center of the furnace. The estimated temperature gradient across the sample length was no more than ~25 °C.

2.3 Data analysis

After recovery, Pt capsules containing the samples were embedded in epoxy resin and polished to expose the center of the charges. A field-emission scanning electron microscope (Quanta 2000-type) (FE-SEM), fitted with a LINK EDS detector, was used to study the microstructure of the recovered samples. Micro-Raman spectrum and electron-backscattered diffraction (EBSD) were used to identify phases in the recovered charges. Quantitative compositional analyses were conducted with an electron probe micro-analyzer (EPMA, JXA-8100) at the Key Laboratory of Submarine Geosciences, State Oceanic Administration, China, using an accelerating voltage of 15 kV, a beam size of 1.0-2.0 μm, and a beam current of 20 nA.

Results

Details of the experimental conditions and run products are summarized in [Tables 1 and 2](#). Quantitative compositional data for minerals are listed in [Tables S1 and S2](#) (see supplementary material). Raman spectra of magnesiochromite and its

high-pressure polymorphs are shown in [Figure 1](#). We observed one characteristic excitation peak of magnesiochromite [idealized composition $Mg(Cr, Al)_2O_4$] at 697 cm^{-1} in the Raman spectrum, in contrast to the two strong peaks of the CF phase [idealized composition $Mg(Cr, Al)_2O_4$] at 628 and 658 cm^{-1} . Two strong bands at 574 and 643 cm^{-1} can be assigned to the mLd phase [modified ludwigite-type phase, idealized composition $Mg_2(Cr, Al)_2O_5$], in comparison to the one strong peak of the CT phase [CaTi₂O₄-structured phase, idealized composition $Mg(Cr, Al)_2O_4$] at 626 cm^{-1} . These results were further confirmed by EBSD analyses, which can be indexed in the space groups *Fd3m*, *Pnma* and *Pbam* for magnesiochromite, the CF phase and the CT phase respectively.

3.1 Phase relations

$MgCr_{1.5}Al_{0.5}O_4 + CaMgSi_2O_6$ (*Mixes A and B*)

We conducted a series of experiments in the system $MgCr_{1.5}Al_{0.5}O_4 + CaMgSi_2O_6$ over pressure ranges of 12-21 GPa, at 1400 and 1600 °C ([Table 1](#)). Microstructures of representative run products are shown in [Figures 2a-b](#). The obtained phase relations are presented in [Figure 2c](#). At a pressure of 12 GPa and temperatures of 1400 and 1600 °C, the starting materials were transformed into assemblages consisting of magnesiochromite + eskolaite (Cr₂O₃) + garnet. When pressure was increased to 14 GPa, the CF phase was observed in contact with garnet ([Figures 2a, b](#)), and is stable up to at least 18 GPa. These observations indicate that garnet (or another silicate phase?) can catalyze the formation of the CF phase. Magnesiochromite can be

preserved up to ~16 GPa, coexisting with the CF or the mLd phase. The assemblage mLd phase + eskolaite, which formed by the decomposition of magnesiochromite, can be observed in experimental charges at 15-18 GPa (Figure 2b). In experiments at 21 GPa, the CT phase was observed to coexist with eskolaite + garnet (Figure 2c).

Two experiments were conducted using $MgCr_2O_4 + CaMgSi_2O_6$ (Mix C) as the starting material: one at 12 GPa and 1400 °C, and another at 15 GPa and 1400 °C. At 12 GPa, the run products consisted of magnesiochromite + eskolaite + garnet. At 15 GPa, the run products consisted of the CF phase + mLd phase + eskolaite + Cr-garnet (Table 1), which is inconsistent with the results from Mixes A and B under similar P - T conditions. We attribute this discrepancy to the presence of Al in Mixes A and B. On the other hand, experimental results in the $MgCr_{1.5}Al_{0.5}O_4$ and $MgCr_2O_4$ systems (15 GPa, 1600 °C) show that the decomposition pressure of $MgCr_{1.5}Al_{0.5}O_4$ magnesiochromite is higher than in the $MgCr_2O_4$ system. Therefore, it can be concluded that some amount of Al has the potential to expand the stability field of magnesiochromite.

Pyrolite + chromite

We conducted a series of experiments in the system *pyrolite + chromite* (PC-A and PC-B) at pressures of 10 and 15 GPa, respectively (Table 2). In experiments conducted at 10 GPa, the run products consisted of Ol + Cpx + Grt at 1400 °C; small amounts of melt can be observed in the recovered samples when temperatures were increased to 1600-1700 °C. At 15 GPa, the starting material was transformed into

assemblages consisting of wadsleyite + garnet at 1500 °C, while small amounts of melt were observed in the recovered charges when temperatures were increased to 1700-1800 °C (Figures 3a, b).

3.2 SiO₂ and CaO contents of chromite and its high-pressure polymorphs

SiO₂ contents of chromite, CF phase and CT phase

Figure 2d shows the SiO₂ contents of chromite, the CF phase and the CT phase in the experimental charges. The SiO₂ contents of chromite taken from literature sources are shown for comparison. For pressure ranges of 4-8 GPa and temperature ranges of 1200-1600 °C, the SiO₂ content of chromite is less than 1 wt%. With pressure increased to 12-15 GPa, the SiO₂ content of chromite ranges from ~1wt% to ~2 wt%. Our experiments also show that the SiO₂ content of chromite at 1600 °C is higher than that at 1400 °C, indicating a moderate increase of SiO₂ solubility with temperature. This tendency has also been observed in experiments of Bery et al. (1999) at pressures of 3-5 GPa and temperatures of 1200-1500 °C.

In the CF phase, the SiO₂ content is higher than that in chromite, and reaches ~3-5 wt% between 14-18 GPa; it shows a positive correlation with pressure, but no correlation with temperature. The result obtained here is lower than that reported by Kojitani et al. (2007) in an MgAl₂O₄-Mg₂SiO₄ system, where the SiO₂ content can reach up to ~11-14 wt% in the CF phase. This discrepancy may reflect the different starting materials and pressures adopted in these two studies, as Kojitani et al. (2007) conducted their experiments at 23-27 GPa. In the CT phase, the SiO₂ content ranges

from ~2 wt% to ~9 wt% in experiments conducted at 21 GPa. These low values could reflect the slow diffusion rate of Si^{4+} between garnet and other mantle minerals (van Mierlo et al., 2013). Very recently, the SiO_2 content in the CT phase was reported to be ~3.8 wt% at 20 GPa, but zero SiO_2 at 21 and 23 GPa in the $\text{Mg}_4\text{Si}_4\text{O}_{12}$ - $\text{Mg}_3\text{Cr}_2\text{Si}_3\text{O}_{12}$ system (Sirotkina et al., 2015). The differences between these two studies could be caused by the difference in starting materials.

CaO content in chromite, CF phase and CT phase

Figure 2e shows the CaO content of magnesiochromite, the CF phase and the CT phase in the present study. The results demonstrate that only traces of CaO (<0.5 wt%) can be incorporated into magnesiochromite at pressures of 12-15 GPa, this is consistent with previous experimental results at lower pressures (~1-2 GPa) (Gasparik, 1987; Liu and O'Neill, 2004a, 2004b). A definite increase of CaO content in the CF phase can be observed over pressure ranges of 14-18 GPa; CaO reaches up to ~7-8 wt%, while the CaO content of the CT phase ranges from ~0.5 wt% to ~8 wt%. These low values could reflect the slow inter-diffusion of Si^{4+} - M^{3+} + M^{2+} (Ca^{2+}) between garnet and other minerals (van Mierlo et al., 2013; Nishi et al., 2013).

Discussion

4.1 Stability of the CF phase

The stability of the CF phase plays a key role in the interpretation of the ultra-deep origin of the UHP chromites. In this study, we have used $\text{MgCr}_{1.5}\text{Al}_{0.5}\text{O}_4$ +

MaCaSi₂O₆ as the starting material. The CF phase was observed coexisting with magnesiochromite at 14-15 GPa, and coexisting with the mLd phase at 15-18 GPa. However, [Ishii et al. \(2014, 2015\)](#) have examined the phase relations in the FeCr₂O₄ and MgCr₂O₄ systems. They found that the CF phase can only be observed in the FeCr₂O₄ system at pressures ~>17-18 GPa and temperatures of no more than 1300 °C; no FeCr₂O₄-chromite or Fe₂Cr₂O₅-mLd phase was observed coexisting with CF phase. In order to confirm the stability of the CF phase that coexists with magnesiochromite or mLd phase in our study, three experiments with longer durations (30 hours at 1400 °C, 12 hours at 1600 °C) were conducted (R676, R679 and R455, for details see [Table 1](#)). The run products in the recovered samples are consistent with experiments over shorter durations (e.g., R350, R374 and R377). EPMA analyses show that compositions of the coexisting phases are similar to those in experiments with short durations ([Table S1](#)). This provides evidence for the stability of the CF phase coexisting with magnesiochromite or the mLd phase, and indicates that chemical equilibrium was at least approached during the experiments. In the following section, we discuss three possible factors potentially accounting the stability of CF phase in our experiments and previous studies.

Effect of FeO

The temperature in the MTZ is about 1400-1600 °C under normal mantle thermal conditions ([Ito and Katsura, 1989](#)). In the MgCr₂O₄ system, the decomposition pressure of magnesiochromite to mLd phase + eskolaite is ~13.5 GPa and ~12.5 GPa at 1400 °C and 1600 °C, respectively; the CT phase was observed at ~17.6 GPa and

1400 °C, and at ~16.6 GPa and 1600 °C (Ishii et al., 2015). When Mg was substituted by iron, the decomposition pressure of Fe-chromite to the mLd phase + eskolaite is also located at ~13.5 GPa and ~12.5 GPa, at 1400 °C and 1600 °C, respectively, while the CT phase was observed at ~18.6 GPa and 1400 °C, and at ~16.3 GPa and 1600 °C (Ishii et al., 2014). The CF phase can only be observed in the FeCr₂O₄ system at temperatures of no more than 1300 °C and pressures above ~17-18 GPa (Ishii et al., 2014). Therefore, we suggest that iron has the potential to stabilize CF phase at relatively low temperature conditions (~< 1300 °C), but has only minor effects on the phase relations of chromite in the MTZ under normal mantle thermal conditions (~1400-1600 °C).

Effect of SiO₂ (or silicate phase)

Previous experiments have shown that the transition between spinel and the CF phase occurs at ~27 GPa at 1600 °C in the pure MgAl₂O₄ system (Akaogi et al., 1999), while the transition pressure decreases to ~23 GPa in the MgAl₂O₄-Mg₂SiO₄ system (Kojitani et al., 2007). This tendency is consistent with the experimental results in Ishii et al. (2015) and our reports. Thus, it can be concluded that the presence of SiO₂ (or a silicate phase) may have the potential to decrease the formation pressure (stability field) of the CF phase.

Effect of Al₂O₃

The CF polymorph has been observed as a stable phase at pressures of ~>26-27

GPa and temperatures of 1400-1600 °C in the pure MgAl_2O_4 system (Irifune et al., 2002; Akaogi et al., 1999), but no CF phase has been observed in the pure MgCr_2O_4 system (Ishii et al., 2015). In another experimental study in an Al-bearing and FeCr_2O_4 -rich system, Chen et al. (2003) reported the coexistence of chromite and the CF phase at pressures above 12.5 GPa and temperatures of 2000 °C; the CF phase then transforms to the CT phase at 20 GPa. Here, it is worth noting that the stable region of the CF phase in their experiments is roughly consistent with that in the present study, given the differences in temperature. Therefore, it can be concluded that Al in the starting material has the potential to expand the stability fields of both chromite and the CF phase. Based on these considerations, we suggest that the lack of Al and Si in the experiments of Ishii et al. (2014, 2015) may be responsible for the difference in the stability region of the CF phase between their reports and the present study.

4.2 Depth of metamorphism of the Tibetan chromitites

A key issue regarding the formation of the UHP chromite is their depth of metamorphism. The discovery of coesite and diamond inclusions in chromitite implies depths of not less than ~80 km and ~120-150 km, respectively. Exsolution of diopside, MgSiO_3 and coesite in chromite has been suggested as an indicator of ultra-deep origin (~>300 km), but experimental evidence has been limited previously. In the present study, the CF phase was observed at 14-18 GPa in the magnesiochromite+diopside system, and it incorporates ~3-5 wt% SiO_2 and ~7-8 wt%

CaO. In contrast, magnesiochromite can incorporate almost no CaO and only minor amounts of SiO₂ (<1-2 wt%) at 12-16 GPa. The trace levels of CaO in chromite cannot account for the abundant diopside exsolution (Yamamoto et al., 2009; Griffin et al., 2016), although even the limited solubility of SiO₂ in chromite could allow the exsolution of rare coesite lamellae through either decompression or cooling. Therefore, the presence of a former CF phase is required to explain the exsolution of diopside in chromite grains. If the “chromite” were (re)crystallized in the stability field of the CT phase, it will break down to the mLd phase + eskolaite during ascent to the surface. This process may release CaO and SiO₂ to the surrounding mantle, as the mLd phase can only take up traces of CaO and SiO₂ in its structure (Table S1). Therefore, it is unlikely that the Luobusha chromitites experienced mantle conditions deeper than the stability field of the CF phase. During their ascent, the CaO- and SiO₂-bearing CF phase transformed to chromite at depths of ~400 km, which resulted in the expulsion of CaO and SiO₂ from its structure and the simultaneous exsolution of diopside. On the other hand, no evidence for the presence of an mLd phase has been reported in natural chromitites. Thus, the Tibetan UHP chromite may have experienced pressures corresponding to depths around the top of the MTZ.

4.3 Petrogenesis of the UHP chromite

UHP minerals and highly-reduced phases have been reported in both podiform chromites and the host peridotites of the Tibetan ophiolites (Yang et al., 2007; Dobrzhinetskaya et al., 2009; Xu et al., 2015). In view of these discoveries, two

explanations have been proposed to reconcile the presence of UHP minerals in chromites with abundant evidence for the shallow formation of podiform chromitites. (1) “Deep crystallization” model: a Cr-rich melt was transported from the MTZ to the upper mantle; chromite crystals crystallized directly from the Cr-rich melt at the bottom of upper mantle, then were transported upward and incorporated into the rising asthenosphere beneath a spreading center and eventually into suprasubduction zone mantle wedges (Robinson et al., 2004; Ruskov et al., 2010; Yang et al., 2014; Xiong et al., 2015; Wu et al., 2016). (2) “Subduction-recycling” model: low-pressure chromitite was subducted to the MTZ as part of a slab system; the chromite and its host dunite/harzburgite then ascended to appear at a spreading center as a mantle diapir, driven by mantle convection or slab rollback (Arai, 2013; McGowan et al., 2015; Griffin et al., 2016). It seems that each of these hypotheses, when combined with later modification in suprasubduction zones, could be used to explain the presence of both UHP minerals and low-pressure hydrous phases (e.g., phlogopite, pargasite) (Lorand and Geuleneer, 1989; Arai, 2013; Zhou et al., 2014). However, the geological processes involved and their geodynamic implications are quite different.

The “deep crystallization” model includes two hypotheses: (a) Cr is preferentially enriched in melts at the top of MTZ; (b) chromite, as a stable phase, can coexist with the mineral assemblages and melt at the bottom of upper mantle and the top of MTZ. Thus, a series of experiments was conducted in the systems *pyrolite* + *chromite* (PC-A and PC-B) at pressures around the 410 km depth range to test the above two hypotheses. The average Cr₂O₃ contents in the systems (~3.5 wt% and 6.5

wt% in *PC-A* and *PC-B*, respectively) are high enough, when compared with normal pyrolitic mantle ($\sim < 0.5$ wt %, [McDonough and Sun, 1995](#)), to constrain the behavior of Cr in the deep mantle. The results in [Table 2](#) and [Figure 3](#) show that chromite (or its high pressure polymorphs) was absent in all experiments; thus it cannot coexist with mantle mineral assemblages at depths around the 410 km seismic discontinuity. The compositional diagrams of Al_2O_3 vs Cr_2O_3 and $\text{FeO}+\text{MgO}$ vs $\text{Cr}_2\text{O}_3+\text{Al}_2\text{O}_3$ ([Figures 3c, d](#)) for the partial melting experiments show that garnet has incorporated the bulk of the Cr (or Cr+Al), while the Cr content in the corresponding melt is very low. For an experiment (R495) conducted at 15 GPa and 1800 °C, the partition coefficient of Cr between garnet and melt is ~ 30 . These results indicate that Cr is preferentially enriched in garnet, rather than melt, at the top of the MTZ. Previous studies have shown that garnet is the liquidus phase in both dry and hydrous pyrolite systems at depths below the 410 km seismic discontinuity ([Zhang and Herzberg, 1994](#); [Litasov and Ohtani, 2002](#)). Thus melts formed by the deep partial melting of pyrolite should be Cr-poor under both dry and hydrous conditions. Based on these observations, we suggest that neither chromite, nor its high-pressure polymorphs can be formed through direct crystallization from melts at the top of the MTZ.

[Arai \(2010, 2013\)](#) has first presented the “subduction-recycling” model; it has been further developed using dynamic modeling ([McGowan et al., 2015](#); [Griffin et al., 2016](#)). Low-pressure chromite, which contains abundant diopside and hydrous phases inclusions, is one of the members of the depleted oceanic or subcontinental lithosphere ([Lorand and Geuleneer, 1989](#); [Yamamoto et al., 2009](#); [Arai, 2010, 2013](#);

366 [Satsukawa et al., 2015; Griffin et al., 2016](#)). This model considers that a segment of
367 this lithospheric mantle was subducted into the deep mantle, and stagnated around the
368 Transition Zone, where the diopside-associated chromite transformed to the
369 CF-structured phase. [Satsukawa et al. \(2015\)](#) have presented microstructural and
370 mineralogical evidence that the chromitites at this stage consisted of wadsleyite and a
371 non-cubic polymorph of chromite, presumably the CF phase, with a strongly
372 deformed and lineated microstructure. After a long residence ($\sim \geq 150$ Ma) and heating
373 to the ambient mantle T, the depleted harzburgite became buoyant with respect to the
374 surrounding mantle, then was recycled back to shallow depth due to the effects of slab
375 rollback and mantle convection ([McGowan et al., 2015; Griffin et al., 2016](#)).
376 Thermo-mechanical modeling has shown that the uplift of highly depleted (and hence
377 buoyant) harzburgite volumes in the fore-arc side of a subduction system is a
378 predictable result of the initial steepening and subsequent rollback of the slab ([Afonso](#)
379 [and Zlotnik, 2011; Moresi et al., 2014; McGowan et al., 2015](#)), which produces
380 large-scale passive upwelling and then develops into a self-maintained low-viscosity
381 channel. During the uplift process, the CF-structured phase transformed back to
382 chromite at depths around 400 km, with the simultaneous exsolution of diopside and
383 other phases, accompanied by grain growth in the chromite ([Satsukawa et al., 2015](#)).
384 Finally, the recycled mantle rocks, which contained UHP chromite, reappeared
385 beneath a spreading center as part of a mantle diapir; UHP chromitites appear in the
386 harzburgite diapir and low-pressure chromite occur within associated lherzolites (for
387 details see [Griffin et al., 2016](#)). During the deep recycling processes, the initial

igneous textures of the podiform chromitites can be preserved because of the absence of reaction between olivine (and its high-pressure polymorphs) and chromite (and its high-pressure polymorphs) (Arai, 2013). Therefore, the “subduction-recycling” model may provide a new window into the processes of mantle convection.

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References

- Afonso, J.C., Zlotnik, S., 2011. The subductibility of continental lithosphere, in: Brown, D., Ryan, P.D. (Eds.), Arc-Continent Collision: Berlin, Heidelberg, Springer-Verlag, Frontiers in Earth Sciences, pp. 53–86.
- Akaogi, M., Hamada, Y., Suzuki, T., Kobayashi, M., Okada, M., 1999. High pressure transitions in the system $\text{MgAl}_2\text{O}_4\text{-CaAl}_2\text{O}_4$: A new hexagonal aluminous phase with implication for the

410 lower mantle. *Phys. Earth Planet. Inter.* 115, 67-77.

411 Arai, S., 1997. Origin of podiform chromitites. *J. Asian Earth Sci.* 15, 303-310.

412 Arai, S., 2010. Possible recycled origin for ultrahigh-pressure chromitites in ophiolites. *J. Mineral.*

413 *Petrol. Sci.* 105 (5), 280–285.

414 Arai, S., 2013. Conversion of low-pressure chromitites to ultrahigh-pressure chromitites by deep

415 recycling: A good inference. *Earth Planet. Sci. Lett.* 379, 81-87.

416 Arai, S., Yurimoto, H., 1994. Podiform chromitites of the Tari–Misaka ultramafic complex,

417 southwestern Japan, as mantle–melt interaction products. *Economic Geology* 89(6), 1279–

418 1288.

419 Brey, G.P., Doroshev, A.M., Gurnis, A.V., Turkin, A.I., 1999. Garnet-spinel-orthopyroxene

420 equilibria in the FeO-MgO-Al₂O₃-SiO₂-Cr₂O₃ system: I. Composition and molar volumes of

421 minerals. *Eur. J. Mineral.* 11(4), 599-617.

422 Chen, M., Shu, J., Mao, H.K., Xie, X., Hemley, R.J., 2003. Natural occurrence and synthesis of

423 two new postspinel polymorphs of chromite. *Proc. Natl. Acad. Sci.* 100, 14651-14654.

424 Dobrzhinetskaya, L.F., Wirth, R., Yang, J., Hutcheon, I.D., Weber, P. K., Green, H.W., 2009. High

425 pressure highly reduced nitrides and oxides from chromite of a Tibetan ophiolite: *Proc. Natl.*

426 *Acad. Sci.* 106, 19233-19238.

427 Edwards, S.J., Pearce, J.A., Freeman, J., 2000. New insights concerning the influence of water

428 during the formation of podiform chromitite. *Geological Society of America Special Papers.*

429 *Geological Society of America Inc., Boulder*, pp. 139–147.

430 Gasparik, T., 1987. Orthopyroxene thermobarometry in simple and complex systems. *Contrib.*

431 *Mineral. Petrol.* 96, 357-370.

432 Girnis, A. V., Brey, G. P., Doroshev, A. M., Turkin, A. I., Simon, N., 2003. The system
 433 MgO-Al₂O₃-Cr₂O₃ revisited: reanalysis of Doroshev et al.'s (1997) experiments and new
 434 experiments. *Euro. J. Mineral.* 15, 953-964.
 435 González-Jiménez, J. M., Griffin, W. L., Proenza, J. A., Gervilla, F., O'Reilly, S. Y., Akbulut, M.,
 436 Pearson, N. J., Arai, S., 2014. Chromitites in ophiolites: How, where, when, why? Part II. The
 437 crystallization of chromitites. *Lithos* 189, 140-158.
 438 Griffin, W.L., Afonso, J.C., Belousova, E.A., Gain, S.E., Gong, X.H., González-Jiménez, J.M.,
 439 Howell, D., Huang, J.X., McGowan, N., Pearson, N.J., Satsukawa, T., Shi, R., Williams, P.,
 440 Xiong, Q., Yang, J.S., Zhang, M., O'Reilly, S.Y., 2016. Mantle Recycling: Transition Zone
 441 Metamorphism of Tibetan Ophiolitic Peridotites and its Tectonic Implications. *J. Petrol.* doi:
 442 10.1093/petrology/egw011.
 443 Herzberg, C., Asimow, P.D., Arndt, N., Niu, Y., Leshner, C.M., Fitton, J.G., Cheadle, M.J.,
 444 Saunders, A.D., 2007. Temperatures in ambient mantle and plumes: Constraints from basalts,
 445 picrites, and komatiites. *Geochem. Geophys. Geosy.* 8, Q02006.
 446 Irifune, T., Naka, H., Sanehira, T., Inoue, T., Funakoshi, K., 2002. In situ X-ray observations of
 447 phase transitions in MgAl₂O₄ spinel to 40 GPa using multianvil apparatus with sintered
 448 diamond anvils. *Phys. Chem. Miner.* 29, 645-654.
 449 Ishii, T., Kojitani, H., Tsukamoto, S., Fujino, K., Mori, D., Inaguma, Y., Tsujino, N., Yoshino, T.,
 450 Yamazaki, D., Higo, Y., Funakoshi, K., Akaogi, M., 2014. High-pressure phase transitions in
 451 FeCr₂O₄ and structure analysis of new post-spinel FeCr₂O₄ and Fe₂Cr₂O₅ phases with
 452 meteoritical and petrological implications. *Am. Mineral.* 99, 1788-1797.
 453 Ishii, T., Kojitani, H., Fujino, K., Yusa, H., Mori, D., Inaguma, Y., Matsushita, Y., Yamaura, K.,

454 Akaogi, M., 2015. High-pressure high-temperature transitions in MgCr_2O_4 and crystal
 455 structures of new $\text{Mg}_2\text{Cr}_2\text{O}_5$ and post-spinel MgCr_2O_4 phases with implications for
 456 ultrahigh-pressure chromitites in ophiolites. *Am. Mineral.* 100, 59-65.

457 Ito, E., Katsura, T. 1989. A temperature profile the mantle transition zone. *Geophys. Res. Lett.*
 458 16(5), 425-428.

459 Jin, Z.M., Bai, Q., Kohlstedt, D.L., Jin, S.Y., 1996. Experimental study on the relation ship of
 460 partial melting in the upper mantle to chromite pre-concentration. *Geological Review* 42,
 461 424-429 (in Chinese with English abstract).

462 Jin, Z.M., Wu, Y., Xu, M.J., Fei, Y.W., Robinson, P.T., 2014. How deep is the Tibetan chromitite:
 463 experimental study: International Workshop on Ophiolites, Mantle Process and Related Ore
 464 Deposits, Beijing, 14-15 April.

465 Klemme, S., O'Neill, H.S.C., 2000. The effect of Cr on the solubility of Al in orthopyroxene:
 466 experiments and thermodynamic modeling. *Contrib. Mineral. Petrol.* 140, 84-98.

467 Kojitani, H., Hisatomi, R., Akaogi, M., 2007. High-pressure phase relations and crystal chemistry
 468 of calcium ferrite-type solid solutions in the system $\text{MgAl}_2\text{O}_4\text{-Mg}_2\text{SiO}_4$. *Am. Mineral.* 92,
 469 1112-1118.

470 Litasov, K., Ohtani, E., 2002. Phase relations and melt compositions in CMAS–pyrolite– H_2O
 471 system up to 25 GPa. *Phys. Earth Planet. Inter.* 134, 105-127.

472 Liu, X., O'Neill, H.S.C., 2004a. The Effect of Cr_2O_3 on the Partial Melting of Spinel Lherzolite
 473 in the System $\text{CaO-MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-Cr}_2\text{O}_3$ at 1.1 GPa. *J. Petrol.* 45(11), 2261-2286.

474 Liu, X., O'Neill, H.S.C., 2004b. Partial Melting of Spinel Lherzolite in the System
 475 $\text{CaO-MgO-Al}_2\text{O}_3\text{-SiO}_2\pm\text{K}_2\text{O}$ at 1.1 GPa. *J. Petrol.* 45(7), 1339-1368.

476 Lorand, J.P., Ceuleneer, G., 1989. Silicate and base-metal sulfide inclusions in chromites from the
477 Maqsad area (Oman ophiolite, Gulf of Oman): a model for entrapment. *Lithos* 22 (3),
478 173-190.

479 McDonough, W.F., Sun, S.S. 1995. The composition of the Earth, *Chem. Geol.*, 120, 223-253.

480 McGowan, N.M., Griffin, W.L., González-Jiménez, J.M., Belousova, E., Afonso, J.C., Shi, R.,
481 McCammon, C.A., Pearson, N.J., O'Reilly, S.Y., 2015. Tibetan chromitites: Excavating the
482 slab graveyard. *Geology* 43, 179-182.

483 Miura, M., Arai, S., Ahmed, A.H., Mizukami, T., Okuno, M., Yamamoto, S., 2012. Podiform
484 chromitite classification revisited: A comparison of discordant and concordant chromitite
485 pods from Wadi Hilti, northern Oman ophiolite. *J. Asian Earth Sci.* 59, 52-61.

486 Moresi, L., Betts, P.G., Miller, M.S., Cayley, R.A., 2014. Dynamics of continental accretion.
487 *Nature* 508, 245-248.

488 Nishi, M., Kubo, T., Ohfuji, H., Kato, T., Nishihara, Y., Irifune, T., 2013. Slow Si-Al interdiffusion
489 in garnet and stagnation of subducting slabs. *Earth Planet. Sci. Lett.* 361, 44-49.

490 Robinson, P.T., Bai, W.-J., Malpas, J., Yang, J.S., Zhou, M.F., Fang, Q.-S., Hu, X.F., Cameron, S.,
491 Staudigel, H., 2004. Ultra-high pressure minerals in the Luobusa Ophiolite, Tibet, and their
492 tectonic implications. Geological Society, London, Special Publications, 226, pp. 247-271.

493 Ruskov, T., Spirov, I., Georgieva, M., Yamamoto, S., Green, H.W., McCammon, C.A.,
494 Dobrzhinetskaya, L. F., 2010. Mössbauer spectroscopy studies of the valence state of iron in
495 chromite from the Luobusa massif of Tibet: implications for a highly reduced deep mantle. *J.*
496 *Metamorph. Geol.* 28, 551-560.

497 Satsukawa, T., Griffin, W.L., Piazzolo, S., O'Reilly, S.Y., 2015. Messengers from the deep: Fossil

498 wadsleyite-chromite microstructures from the Mantle Transition Zone. *Scientific Reports* 5,
499 16484.

500 Sirotkina, E.A., Bobrov, A.V., Bindi, L., Irifune, T., 2015. Phase relations and formation of
501 chromium-rich phases in the system $\text{Mg}_4\text{Si}_4\text{O}_{12}$ – $\text{Mg}_3\text{Cr}_2\text{Si}_3\text{O}_{12}$ at 10–24 GPa and 1,600 °C.
502 *Contrib. Mineral. Petrol.* 169, 1-14.

503 van Mierlo, W. L., Langenhorst, F., Frost, D.J., Rubie, D.C., 2013. Stagnation of subducting slabs
504 in the transition zone due to slow diffusion in majoritic garnet. *Nature Geosci.* 6, 400-403.

505 Wu, Y., Xu, M., Jin, Z., Fei, Y., Robinson, P.T., 2016. Experimental constraints on the formation of
506 the Tibetan podiform chromitites. *Lithos* 245, 109-117.

507 Xiong, F.H., Yang, J.S., Robinson, P.T., Xu, X.Z., Liu, Z., Li, Y., Li, J.Y., Chen, S.Y., 2015. Origin
508 of podiform chromitite, a new model based on the Luobusa ophiolite, Tibet. *Gondwana Res.*
509 27, 525–542.

510 Xu, M., 2011. Phase relations of chromite+ SiO_2 system at high temperature and high pressure and
511 implications for the origin of coesite exsolution in chromite. Master thesis, China University
512 of Geosciences.

513 Xu, X., Yang, J., Robinson, P.T., Xiong, F., Ba, D., Guo, G., 2015. Origin of ultrahigh pressure and
514 highly reduced minerals in podiform chromitites and associated mantle peridotites of the
515 Luobusa ophiolite, Tibet. *Gondwana Res.* 27, 686-700.

516 Yamamoto, S., Komiya, T., Hirose, K., Maruyama, S., 2009. Coesite and clinopyroxene exsolution
517 lamellae in chromites: In-situ ultrahigh-pressure evidence from podiform chromitites in the
518 Luobusa ophiolite, southern Tibet. *Lithos* 109, 314-322.

519 Yang, J.S., Dobrzhinetskaya, L., Bai, W.J., Fang, Q.S., Robinson, P.T., Zhang, J., Green, H.W.,

2007. Diamond- and coesite-bearing chromitites from the Luobusa ophiolite, Tibet. *Geology* 35, 875-878.
- Yang, J.S., Robinson, P.T., Dilek, Y., 2014, Diamond in ophiolites. *Elements* 10, 127-130.
- Zhang, J.Z., Herzberg, C., 1994. Melting experiments on anhydrous peridotite KLB-1 from 5.0 to 22.5 GPa. *J. Geophys. Res.* 99(B9), 17729–17742.
- Zhang, R.Y., Yang, J.S., Ernst, W.G., Jahn, B.M., Lizuka, Y., Guo, G.L., 2016. Discovery of in situ super-reducing, ultrahigh-pressure phases in the Luobusa ophiolitic chromitites, Tibet: New insights into the deep upper mantle and mantle transition zone. *Am. Mineral.* 101, 1285-1294.
- Zhang, Y., Wang, Y., Wu, Y., Bina, C.R., Jin, Z., Dong, S., 2013. Phase transitions of harzburgite and buckled slab under eastern China. *Geochem. Geophys. Geosy.* 14, 1182-1199.
- Zhou, M.F., Robinson, P.T., Malpas, J., Li, Z.J., 1996. Podiform chromitites in the Luobusa ophiolite (southern Tibet): Implications for melt-rock interaction and chromite segregation in the upper mantle. *J. Petrol.* 37, 3-21.
- Zhou, M.F., Robinson, P.T., Su, B.X., Gao, J.F., Li, J.W., Yang, J.S., Malpas, J., 2014. Compositions of chromite, associated minerals, and parental magmas of podiform chromite deposits: The role of slab contamination of asthenospheric melts in suprasubduction zone environments. *Gondwana Res.* 26, 262-283.

Figure captions

Figure 1. Raman spectra of chromite and its high-pressure polymorphs. (a), (b), (c) and (d) show the Raman spectra of magnesiochromite, the CF phase, the mLd phase and the CT phase, respectively.

542

543 **Figure 2.** Experimental results in magnesiochromite+diopside systems (*mixes A and*
544 *B*). (a) and (b) microstructures of run R676 (15 GPa, 1400 °C) and run R679 (18 GPa,
545 1600 °C), respectively; (c) phase relations of magnesiochromite+diopside based on
546 this study; (d) and (e) the SiO₂ (d) and CaO (e) content (wt %) in chromite and its
547 high pressure polynorphs. The SiO₂ content in chromite from previous studies is also
548 shown for comparison. KO2000: Klemme and O'Neill, 2000; G2003: Girnis et al.,
549 2003; W2015: Wu et al., 2016. Mineral abbreviations: Chr, chromite; CF,
550 CaFe₂O₄-type chromite; mLd, modified ludwigite-type phase; Es, eskolaite; CT,
551 CaTi₂O₄-type chromite; Grt, garnet.

552

553 **Figure 3.** Experimental results in the pyrolite+chromite system. (a) and (b)
554 microstructures of run R495 (15 GPa, 1800 °C); (c) and (d) compositional diagrams
555 for the partial melting experiments in the pyrolite + chromite system; (c) Al₂O₃ vs
556 Cr₂O₃, (d) FeO+MgO vs Cr₂O₃+ Al₂O₃. Mineral abbreviations: Q-melt, quenched melt;
557 Grt, garnet.

Figure 1

Intensity

Figure 2

Figure 3

Table 1. Summary of experimental conditions and run products in chromite+diopside system.

Run no.	P (GPa)	T (°C)	D (h)	Starting materials	Run products
R355	12	1400	6	<i>Mixes A, B</i>	Chr+Es+Grt
R412-1	14	1400	6	<i>Mix A</i>	Chr+CF+Es+Grt
R350	15	1400	6	<i>Mixes A, B</i>	Chr+CF+Es+Grt
R676	15	1400	30	<i>Mix A</i>	Chr+CF+Es+Grt
R448-1	16	1400	6	<i>Mix A</i>	Chr+CF+mLd+Es+Grt
R371	18	1400	6	<i>Mixes A, B</i>	CF+mLd+Es +Grt
R375	21	1400	6	<i>Mixes A, B</i>	CT+Es+Grt
R373	12	1600	1	<i>Mixes A, B</i>	Chr+Es+Grt
R406-1	14	1600	3	<i>Mix A</i>	Chr+CF+Es+Grt
R367	15	1600	3	<i>Mixes A, B</i>	Chr+CF+mLd+Es+Grt
R449-1	16	1600	3	<i>Mix A</i>	CF+mLd+Es+Grt
R374	18	1600	3	<i>Mixes A, B</i>	CF+mLd+Es+Grt
R679	18	1600	12	<i>Mix A</i>	CF+mLd+Es+Grt
R377	21	1600	3	<i>Mixes A, B</i>	CT+Es+Grt
R455	21	1600	12	<i>Mix A</i>	CT+Es+Grt
R342	12	1400	4	<i>Mix C</i>	Chr+Es+Grt
R347	15	1400	6	<i>Mix C</i>	CF+mLd+Es+Grt
R412-2	14	1400	6	MgCr _{1.5} Al _{0.5} O ₄	Chr
R448-2	16	1400	6	MgCr _{1.5} Al _{0.5} O ₄	Chr+mLd+Es
R406-2	14	1600	3	MgCr _{1.5} Al _{0.5} O ₄	Chr
R415-2	15	1600	3	MgCr _{1.5} Al _{0.5} O ₄	Chr
R449-2	16	1600	3	MgCr _{1.5} Al _{0.5} O ₄	mLd+Es
R415-1	15	1600	3	MgCr ₂ O ₄	mLd+Es

Mineral abbreviations: Chr, magnesiochromite, idealized composition Mg(Cr,Al)₂O₄; Grt, garnet; CF, CaFe₂O₄-type phase, idealized composition Mg(Cr,Al)₂O₄; mLd, modified ludwigite-type phase, idealized composition Mg₂(Cr,Al)₂O₅; Es, eskolaite, idealized composition Cr₂O₃; CT, CaTi₂O₄-type phase, idealized composition Mg(Cr,Al)₂O₄. **Mix A**: 90 mol.% Mg(Cr_{1.5}Al_{0.5})O₄ + 10 mol.% CaMgSi₂O₆. **Mix B**: 95 mol.% Mg(Cr_{1.5}Al_{0.5})O₄ + 5 mol.% CaMgSi₂O₆. **Mix C**: 90 mol.% MgCr₂O₄ + 10 mol.% CaMgSi₂O₆.

Table 2. Summary of experimental conditions and run products in pyrolite+chromite system.

Run no.	P (GPa)	T (°C)	D (h)	Starting materials	Run products
R490	10	1400	4	<i>PC-A</i>	Ol+Cpx+Grt
R474	10	1600	3	<i>PC-A</i>	Ol+Cpx+Grt+ trace melt
R494	10	1700	2	<i>PC-A</i>	Ol+Grt+ melt
R681	10	1650	1	<i>PC-B</i>	Ol+Grt+ trace melt
R488	15	1500	3	<i>PC-A</i>	Wd+Grt
R472	15	1700	2	<i>PC-A</i>	Wd+Grt+ trace melt
R495	15	1800	1	<i>PC-A</i>	Wd+Grt+ melt
R680	15	1800	1	<i>PC-B</i>	Wd+Grt+ melt

Note: the starting material of *PC-A* was composed of 95 wt% pyrolite and 5 wt% chromite, *PC-B* was composed of 90 wt% pyrolite and 10 wt% of chromite. Mineral abbreviations: Ol, olivine; Cpx, clinopyroxene; Grt, garnet; Wd, wadsleyite.