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Genesis of the Sartohay Podiform Chromitite Based on Microinclusions in Chromite

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Abstract: Here, we present a petrographic and microanalytical study of microinclusions in chromite from podiform chromitites hosted by the Sartohay ophiolitic mélange in west Junggar, northwestern China, to investigate the parental magma evolution and chromitite genesis. These silicate inclusions comprise olivine, enstatite, diopside, amphibole, and Na-phlogopite. Their morphological characteristics suggest that most inclusions crystallized directly from the captured melt, with a few anhydrous inclusions (olivines and pyroxenes) as solid silicates trapped during the chromite crystallization. Equilibrium pressure–temperature conditions of coexisting enstatite–diopside inclusions are 8.0–21.6 kbar, and 874–1048 °C. The high Na₂O and TiO₂ contents of hydrous minerals indicate that the parental magma of chromitites was hydrous and enriched in Mg, Na, Ca, and Ti. The calculated Al₂O₃ content and FeO/MgO ratio of the parental melts in equilibrium with chromite showed MORB affinity. However, the TiO₂ values of parental melts, TiO₂ contents of chromite, and estimated *f*O₂ values for chromitites (1.3–2.0 log units above the FMQ buffer) evoked parental MORB-like tholeiitic melts. The composition of olivine inclusion was determined, and it was revealed that the primary melts of the Sartohay podiform chromitites had MgO contents of ~22.7 wt %. This aligns with the observed high magnesian signature in mineral inclusions (Fo = 96–98 in olivine, Mg[#] = 0.91–0.97 in diopside, and Mg[#] = 0.92–0.97 in enstatite). We propose that Sartohay podiform chromitites initially formed through the mixing/mingling of primary hydrous Mg-rich melt and the evolved MORB-like melt derived from the melt–peridotite reaction in the upper mantle. In this process, the continuous crystallization of chromite captured micro-silicate mineral inclusions, finally leading to the formation of the Sartohay podiform chromitites.



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Keywords: podiform chromitites; silicate mineral inclusion; chromite; olivine; enstatite; diopside; amphibole; Sartohay

1. Introduction

The mechanism for the origin of podiform chromitites remains debatable, with various models proposed, including fractional crystallization [1], partial melting [2], fluid-related models [3], deep mantle mineralization models related with UHP (ultra-high pressure) minerals [4–6], and rheomorphic model associated with the localization of plastic flow in the dunite [7–9]. The most prevalent petrogenetic model is melt–peridotite interaction and subsequent melt mixing and/or mingling [10–12]. Arai and Miura [13] noted that mixing melts with different silica components is an essential process to precipitate chromite during the early stage of magmatic crystallization. Matveev and Ballhaus [14] suggested that the melts responsible for chromitite formation are hydrous and Cr-rich.

Mineral inclusions hosted by chromite have been identified in podiform chromitites from various orogenic belts of the world [15–17]. These silicate inclusions are commonly composed of olivine, enstatite, diopside, pargasite, and Na-phlogopite. Hydrous phases such as amphibole and phlogopite may represent hydrous SiO₂-rich melt droplets trapped by the chromite host [18]. Olivine and pyroxene inclusions occurring as independent

grains might be genetically linked to chromite crystallization and re-equilibrium, reflecting the oxygen fugacity and physical–chemical nature of chromitite parental melts [19–21]. Differences in mineral assemblages of inclusions reflect the chemical heterogeneity of primary magma [22].

The chemical composition of chromite serves as a valuable tool for elucidating the characteristics of the parental melt in which the chromite crystallized [10,23]. The trapped micromelt inclusions are also valid petrogenetic indicators for the parental melt of chromitites and peridotites because host chromite is resistant to modification during post-magmatic processes [24]. Homogenization experiments on mineral inclusions in chromitites from the Maqсад in southern Oman suggest a hydrous hybrid MORB nature of parental magma [25]. SEM analysis conducted by Rollinson et al. [26] on melt inclusions in chromitite indicates hydrous, highly magnesian (20–22 wt %) primary melts. Therefore, integrated studies of the microstructure and composition of microinclusions and host chromite have the potential to provide insights into the petrological and dynamic processes that contribute to chromitite formation [18].

The genesis of podiform chromitites in the Sartohay ophiolitic mélange remains a matter of contention. Previous studies suggested that chromitite is a product of reactions between MORB-like melts and depleted harzburgites [17,27]. We provide new data to illustrate the genesis of microinclusions in chromite, as well as the physical–chemical characteristics of the parental melts and the formation process of Sartohay chromitites.

2. Regional Geology

West Junggar in north Xinjiang of China is characterized by the occurrence of complex emplacements of ophiolitic mélanges, volcanic–sedimentary rocks, and intermediate-to-granitic intrusions (Figure 1a), which is considered to be a Palaeozoic orogenic belt resulting from the convergence of the Siberian and Kazakhstan–Junggar plates [28,29]. The ages of the ophiolitic mélanges occurring in the west Junggar range from the Cambrian to Devonian periods [30–33]. These ophiolitic mélanges are widely regarded as relics of the Ordovician to Silurian oceanic crust, which subducted to the south under the Junggar plate [29]. As the major component of the ophiolitic mélanges in west Junggar, the Sartohay ophiolitic mélanges are covered by flysch formation, which mainly consists of siltstone, mudstone, tuff, and chert. These mélanges and flysch underwent intense deformation and are unconformably covered by undeformed Devonian to early Carboniferous volcanic–sedimentary rocks [34–36]. All these units were intruded by late Carboniferous to early Permian monzogranite (with a zircon U–Pb age of 308 ± 3 Ma) [37], pyroxene diorite and granite porphyry (310.7 ± 3.7 Ma and 312 ± 3 Ma, respectively) [33], hornblende gabbro and granite (334 ± 1 Ma and 319 ± 4 Ma, respectively) [38], pyroxene-bearing granitoid and charnockite (306.5 ± 2.6 Ma) [39].

The Sartohay ophiolitic mélanges exhibit a zonal distribution along the NE–NEE fault belt (Figure 1a). The ophiolite consists mainly of ultramafic rocks, podiform chromitites, cumulates, pillow lavas, and radiolarian cherts, and it is cut by numerous veins of gabbro and diabase. The ultramafic rocks are mostly harzburgite with minor amounts of lherzolite, dunite, and lenses of metagabbro (Figure 1b). Most of the harzburgites have been altered to serpentinite with listwaenite [40], whereas dunite occurs as dykes or envelopes surrounding the chromitite ore bodies. Metagabbro lenses distributed in serpentinite matrix are strongly deformed. The cumulate gabbro and pillow lavas with interlayers of thin red chert or Jasper-siltstone, tuffaceous sandstone, and conglomerate tectonically overlie ultramafic rocks.

The discovered chromitites are mainly distributed in the northeastern part of Sartohay ophiolitic mélanges [42,43]. Most ore bodies hosted in the harzburgite are typically surrounded by dunite envelopes of varying thicknesses (0.5–30 cm), whereas some bodies have direct contact with serpentinitized harzburgites (Figure 1c) [27,41]. The ore types are mainly massive, disseminated, and banded (Figure 2). The size and shape of orebodies vary greatly and mostly occur as pods, lenses, and veins [17,44]. The orebodies stretch in the shear direction, and the smaller ones are less than half a meter, while the larger ones

are tens of meters along the long axis. The Sartohay chromitite deposit has several clusters of ore bodies, including the major mining areas named as No. 22 Group, 24 Group, and 25 Group. Chromitite ore bodies and serpentinite with chromitite lenses occasionally occur at the stratigraphic bottom of the Sartohay ultramafic massif (Figure 1c). The samples studied here were collected from the bottom of the 24 Group chromitite bodies.

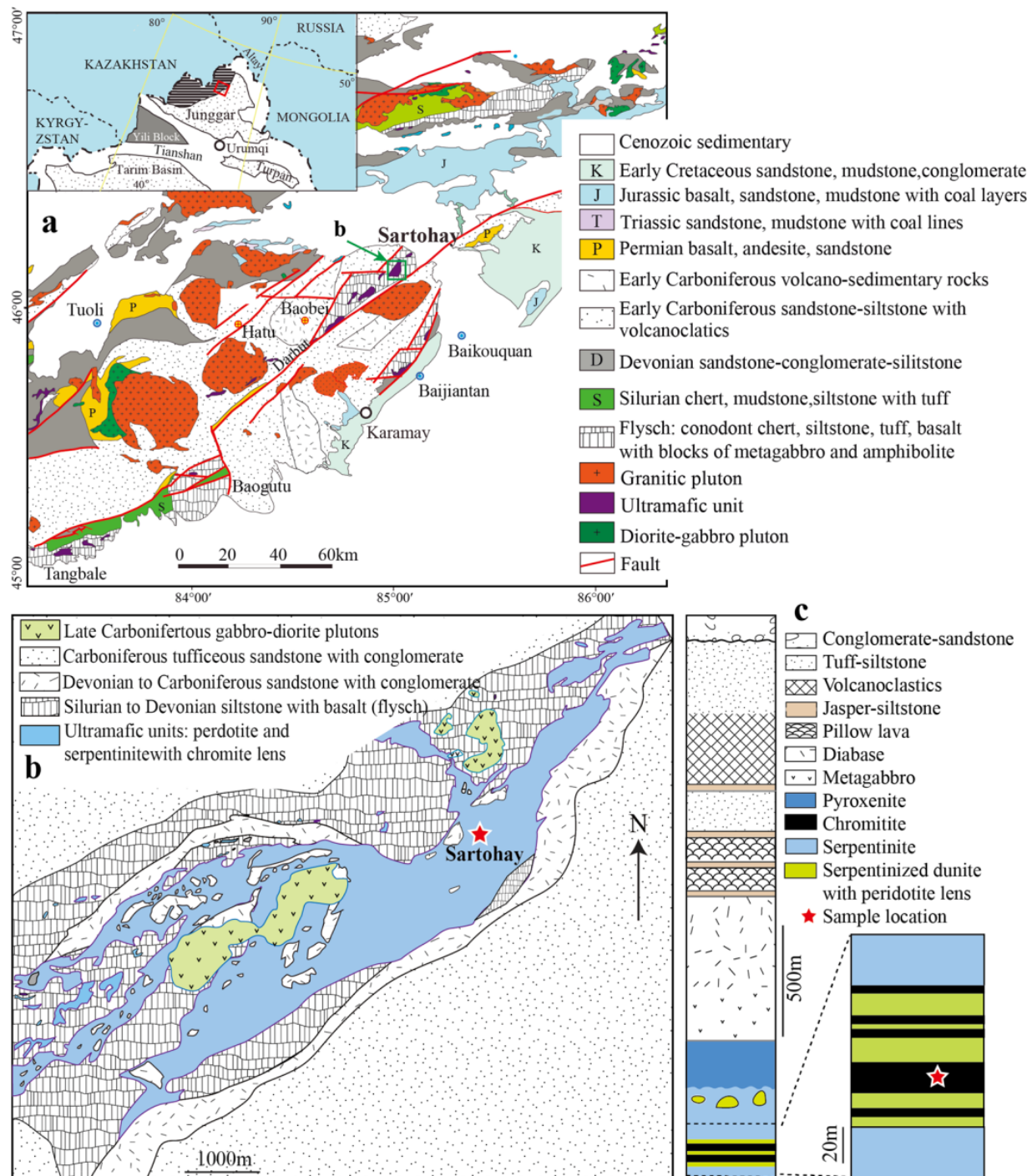


Figure 1. (a) Simplified geological map of west Junggar showing the distributions of the ophiolitic mélanges and plutons. (b) Geological map showing major geological units consisting of the Sartohay ophiolitic mélanges with the location of the Sartohay chromitite deposit. (c) Compiled geological profile showing rock units of the Sartohay ophiolite mélangé with an enlarged section showing the lower part containing chromitite lenses with sample location (from Zhu et al. [41]).

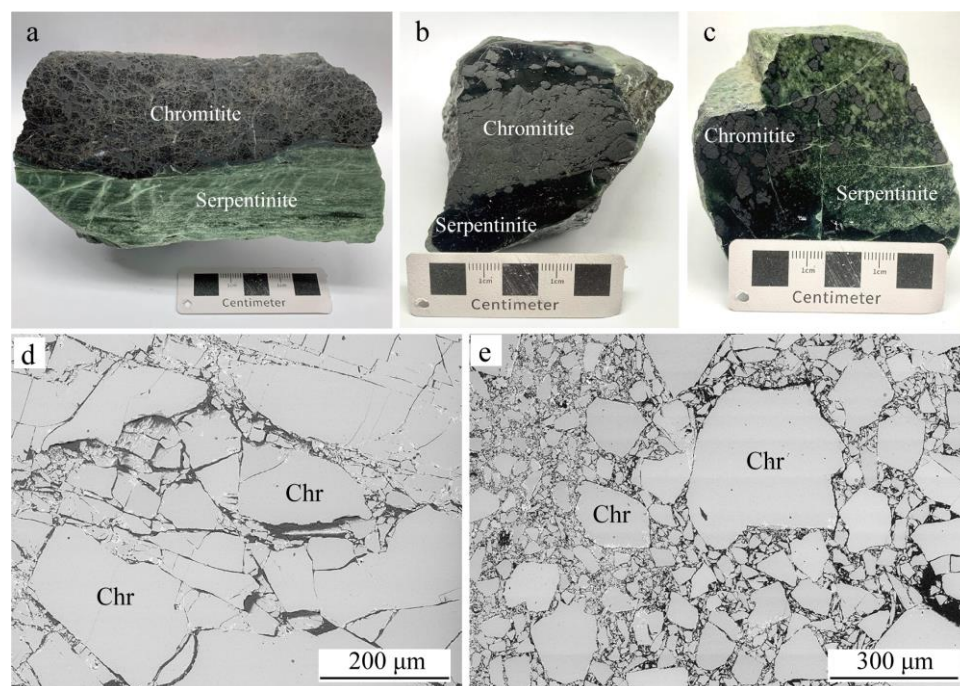


Figure 2. Photos showing the banded chromitite (a,b), disseminated chromitite (c), and photomicrographs showing the massive chromitite (d) and massive chromitite with cataclastic textures (e) in the Sartohay ophiolitic mélange. The black area on the right of (d) is a marker handwriting. Chr = chromite.

3. Analytical Methods

Samples 2022Cr-2a, 2022Cr-3, and 2022Cr-6 were collected from the 24 Group chromitite bodies mined in Sartohay. Mineral phases were first studied in polished thin sections under both transmitted and reflected light and then examined by electron microscope (SEM) using the back-scattered electron (BSE) mode. The inclusions were tentatively identified by semi-qualitative energy-dispersive spectroscopy (EDS). The acquisition time for each spot was 60 s with an accelerating voltage of 15 kv and spot size 4–5 μm . SEM-EDS analyses were performed using the ThermoFisher Quattro ESEM at the Electron Microscopy Laboratory of Peking University, Beijing, China.

The major element compositions of chromite, olivine, enstatite, diopside, amphibole, and Na-phlogopite in chromitite samples were determined using the JEOL JXA-8100 electron probe microanalyzer (EMPA, at the Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China). A beam current of 10 nA, an acceleration voltage of 15 kV, and a 1 μm focused beam were used for all analyses. Characteristic peaks and background positions of all elements were measured with acquisition times of 10 s and 5 s, respectively. The SPI 53 mineral standards (US) were utilized for the quantitative oxide–silicate analysis: diopside (Ca, Mg), rutile (Ti), jadeite (Na, Al, Si), sanidine (K), chromium oxide (Cr), nickel silicide (Ni), rhodonite (Mn), and hematite (Fe). All the standards (natural and synthetic) were tested for homogeneity before their utilization for quantitative analysis, and the PRZ correction was adopted for the final results of oxide–silicate. Detection limits of the oxides were 0.06 wt % for SiO_2 ; 0.04 wt % for Cr_2O_3 , MnO, FeO, and NiO; 0.03 wt % for MgO and CaO; 0.02 wt % for Al_2O_3 , Na_2O , and TiO_2 ; and 0.01 wt % for K_2O .

4. Results

4.1. Silicate Mineral Inclusions in Chromite

Silicate mineral inclusions in the host chromite from the Sartohay podiform chromitites mainly consist of olivine, diopside, enstatite, amphibole, and Na-phlogopite. These inclusions mostly vary from 10 μm to 35 μm and occasionally may reach 50 μm in size

(Figure 3a). Base-metal sulfide (BMS), rutile, and zircon grains generally occur along the rim of these silicate minerals (1–3 μm ; e.g., Figure 3c). The distribution of inclusions in chromite grains is random. All silicate inclusions found in Sartohay chromitite are primary inclusions, which are located away from cracks and fractures.

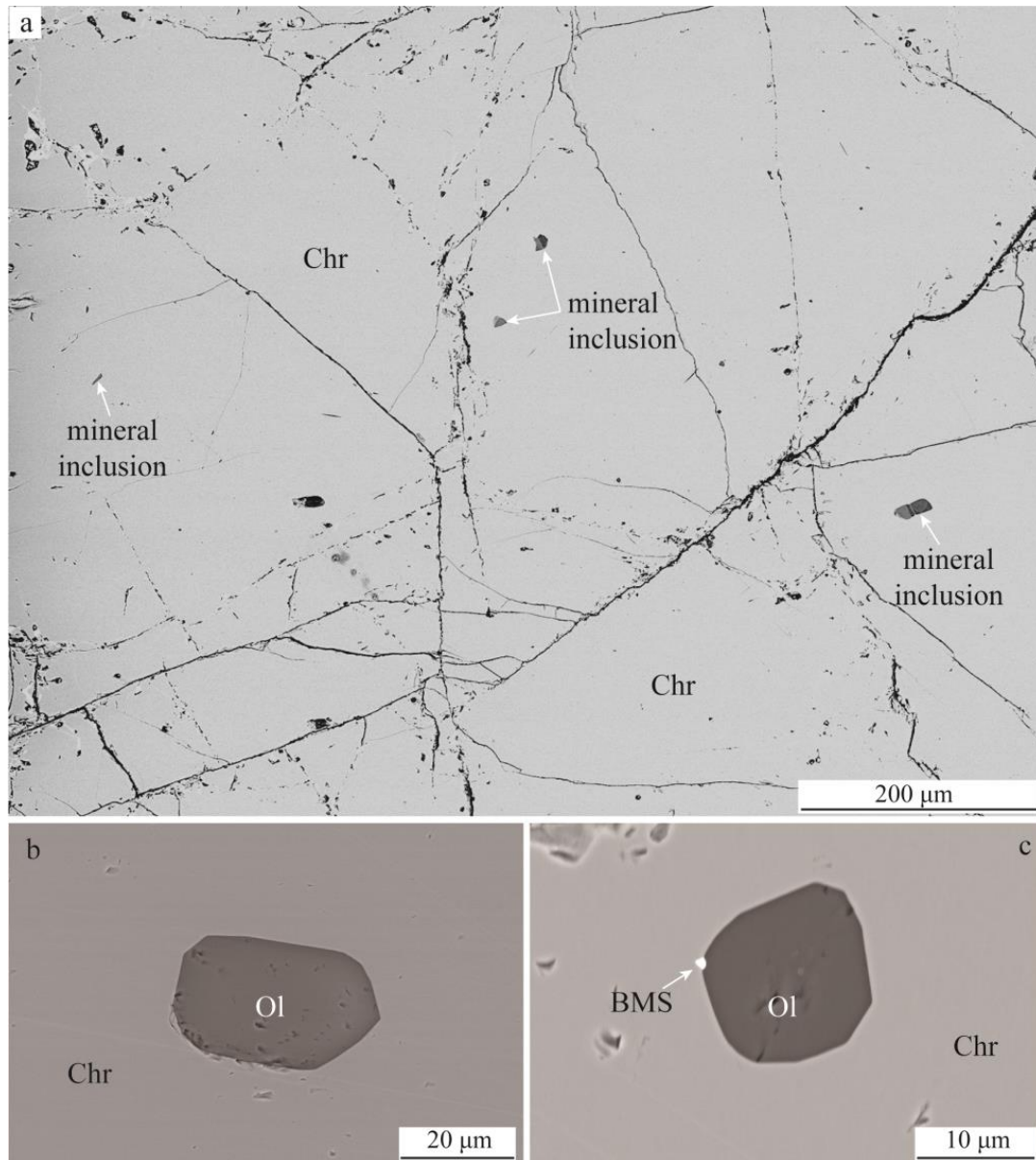


Figure 3. (a) Back-scattered electron (BSE) images showing silicate mineral inclusions within the chromite grain of Sartohay chromitites. (b,c) Isolated olivine inclusions trapped in the chromite. Abbreviation: BMS = base-metal sulfide, Chr = chromite, Ol = olivine.

Olivine inclusions are euhedral to slightly subhedral in shape with sizes ranging from 15 up to 40 μm (Figure 3b,c). Enstatite inclusions display two types with different mineral assemblages: Type 1 are isolated enstatite grains with sizes ranging from 5 μm to 35 μm (Figure 4a,b). Most of them have globular/spheroidal to irregular shapes against cubic chromite, with some preserving subhedral crystal forms. Type 2 inclusions are characterized by the enstatite coexisting with amphibole and/or Na-phlogopite (Figure 4c–h). These inclusions exhibit irregular or subroundish shapes with grain sizes of 5–30 μm , and with zircon or rutile occasionally observed at the margin of the grain.

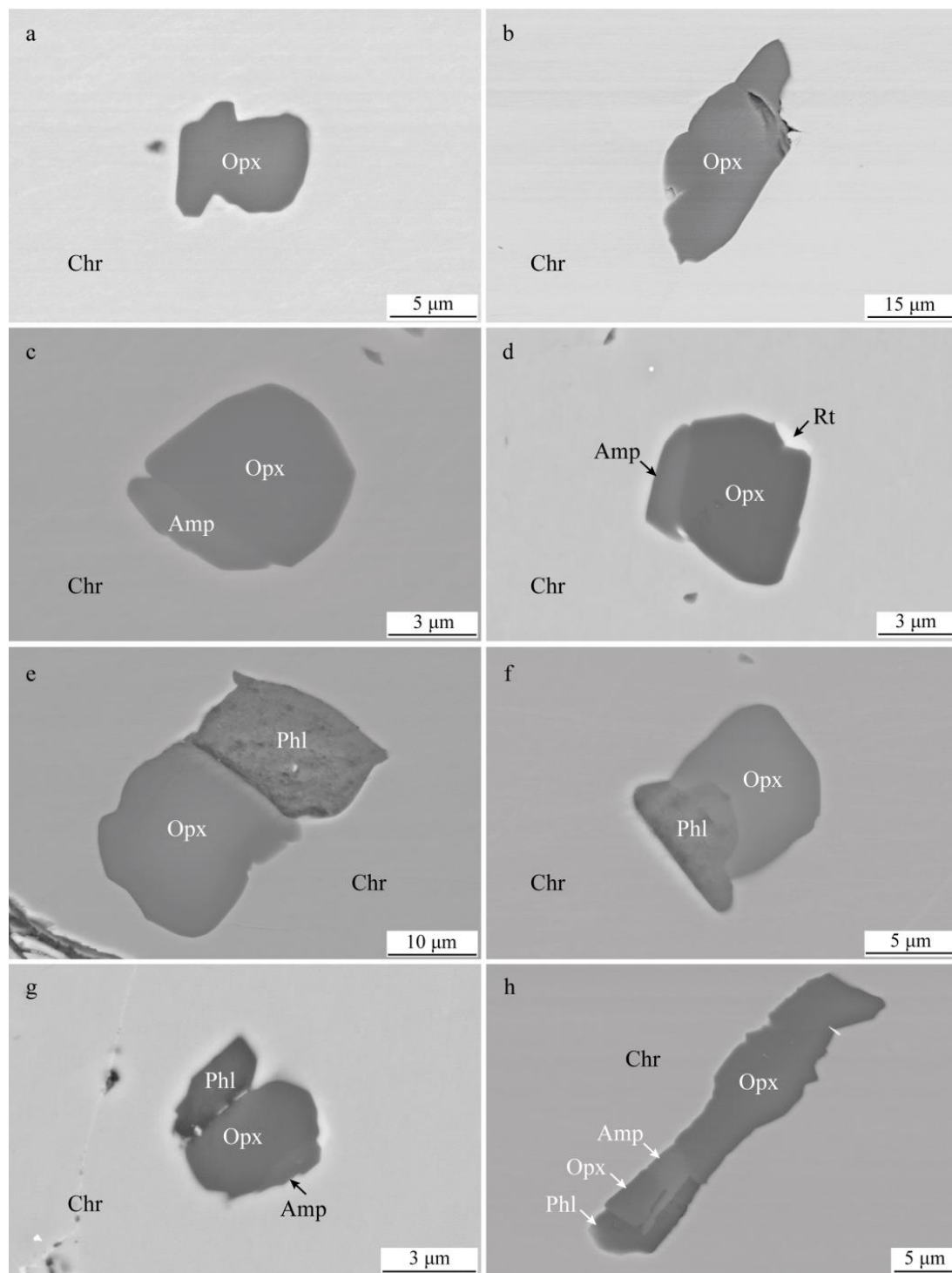


Figure 4. Back-scattered electron (BSE) images of clinopyroxene inclusions trapped in the chromite from Sartohay chromitites. (a,b) Single orthopyroxene crystal. (c–h) Orthopyroxene coexisting with amphibole and/or phlogopite. Abbreviation: Amp = amphibole, Chr = chromite, Opx = orthopyroxene, Phl = phlogopite, Rt = rutile.

There are three types of diopside inclusions (~20 μm) based on mineral assemblages. Type 1 are single granular or subhedral diopside grains with sizes around 5 μm (Figure 5a,b). Type 2 diopside is replaced by amphibole and occurs as a relict texture (Figure 5c). Type 3 inclusions are characterized by diopside with subrounded to irregular forms containing lensoid enstatite lamellae, displaying tiny holes or microfractures inside (Figure 5d–f).

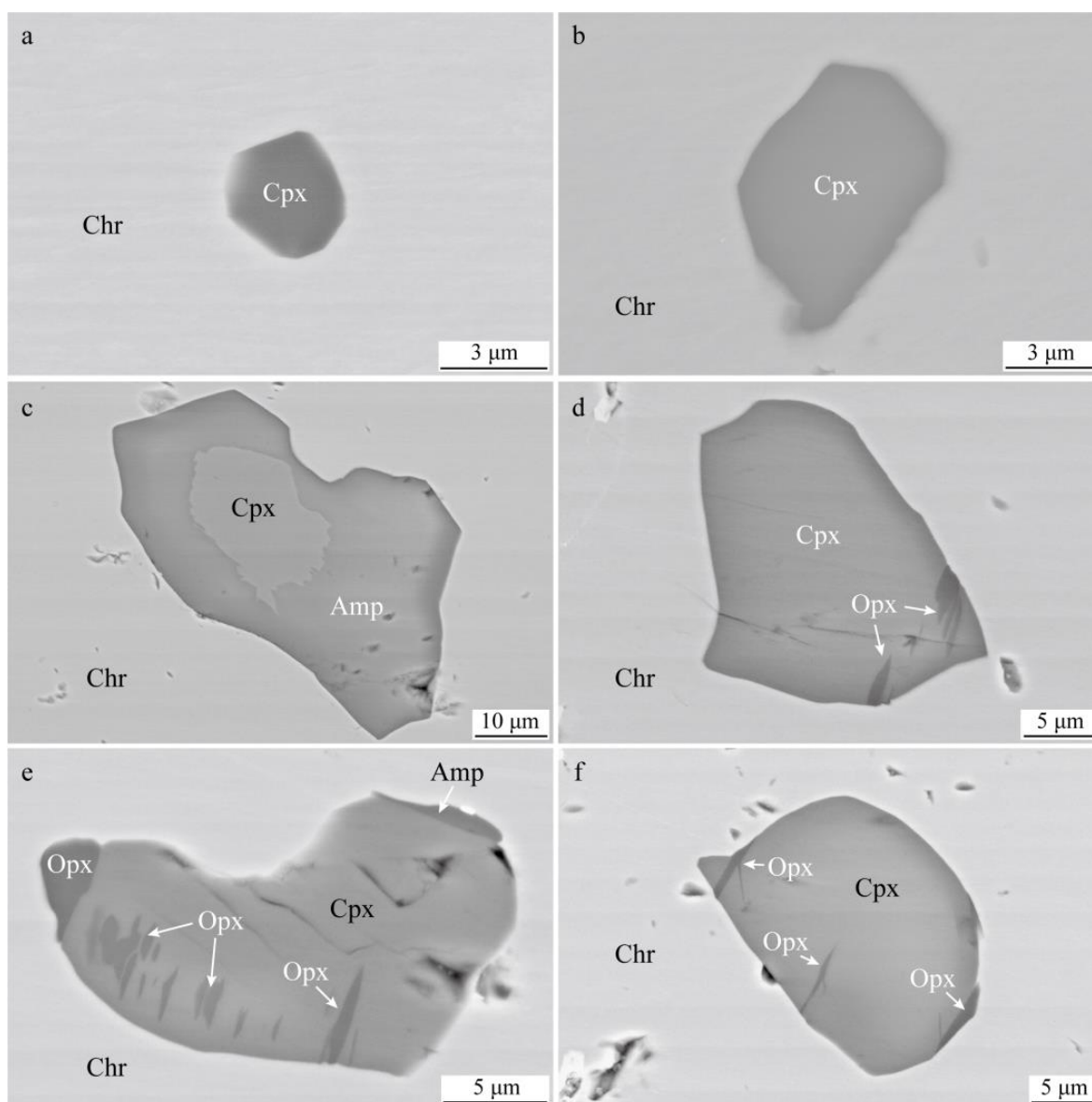


Figure 5. Back-scattered electron (BSE) images displaying clinopyroxene inclusions in Sartohay chromitites. (a,b) Solely clinopyroxene inclusion. (c) Clinopyroxene was replaced by amphibole. (d–f) Coexistence of clinopyroxene and orthopyroxene. Orthopyroxene occurs as exsolution lamellae in clinopyroxene. Abbreviation: Amp = amphibole, Chr = chromite, Cpx = clinopyroxene, Opx = orthopyroxene.

Amphibole inclusions are the most prevalent in the studied chromitites, and they can be categorized into two types according to their mineral assemblages. Type 1 inclusions are isolated amphibole (15–30 µm), showing tabular to irregular morphologies (Figure 6a,b). Type 2 inclusions comprising amphibole, enstatite, and/or Na-phlogopite (5–30 µm) with occasional BMS, zircon, or rutile grains (Figure 6c–h). Occasionally, amphibole replaces enstatite (Figure 6c).

Two types of Na-phlogopite inclusions can be distinguished. Type 1 inclusions occur as tiny solitary Na-phlogopite (<5 µm) with thin sheet shapes (Figure 6f). Type 2 inclusions consist of Na-phlogopite, enstatite, and/or amphibole (5–20 µm), and Na-phlogopite often replaces enstatite and/or amphibole (Figure 7). Some of the Na-phlogopite inclusions exhibit sharp protuberances into the host chromite (Figure 7e,f).

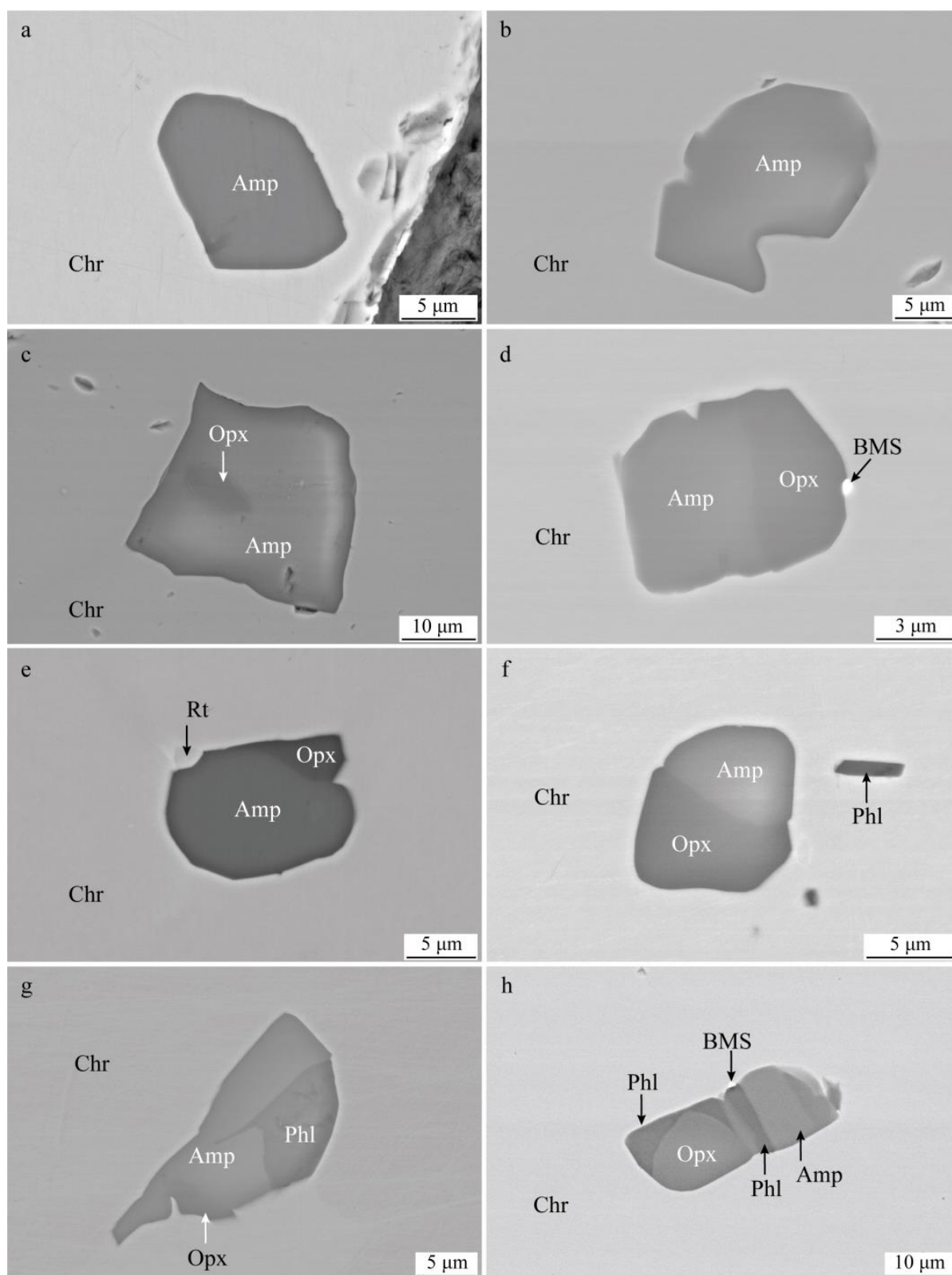


Figure 6. Back-scattered electron (BSE) images of amphibole inclusions enclosed in Sartohay chromitites. (a,b) Solitary amphibole crystal. (c) Amphibole replaces orthopyroxene. (d–h) Amphibole coexisting with orthopyroxene and/or phlogopite. Abbreviation: Amp = amphibole, BMS = base-metal sulfide, Chr = chromite, Opx = orthopyroxene, Phl = phlogopite, Rt = rutile.

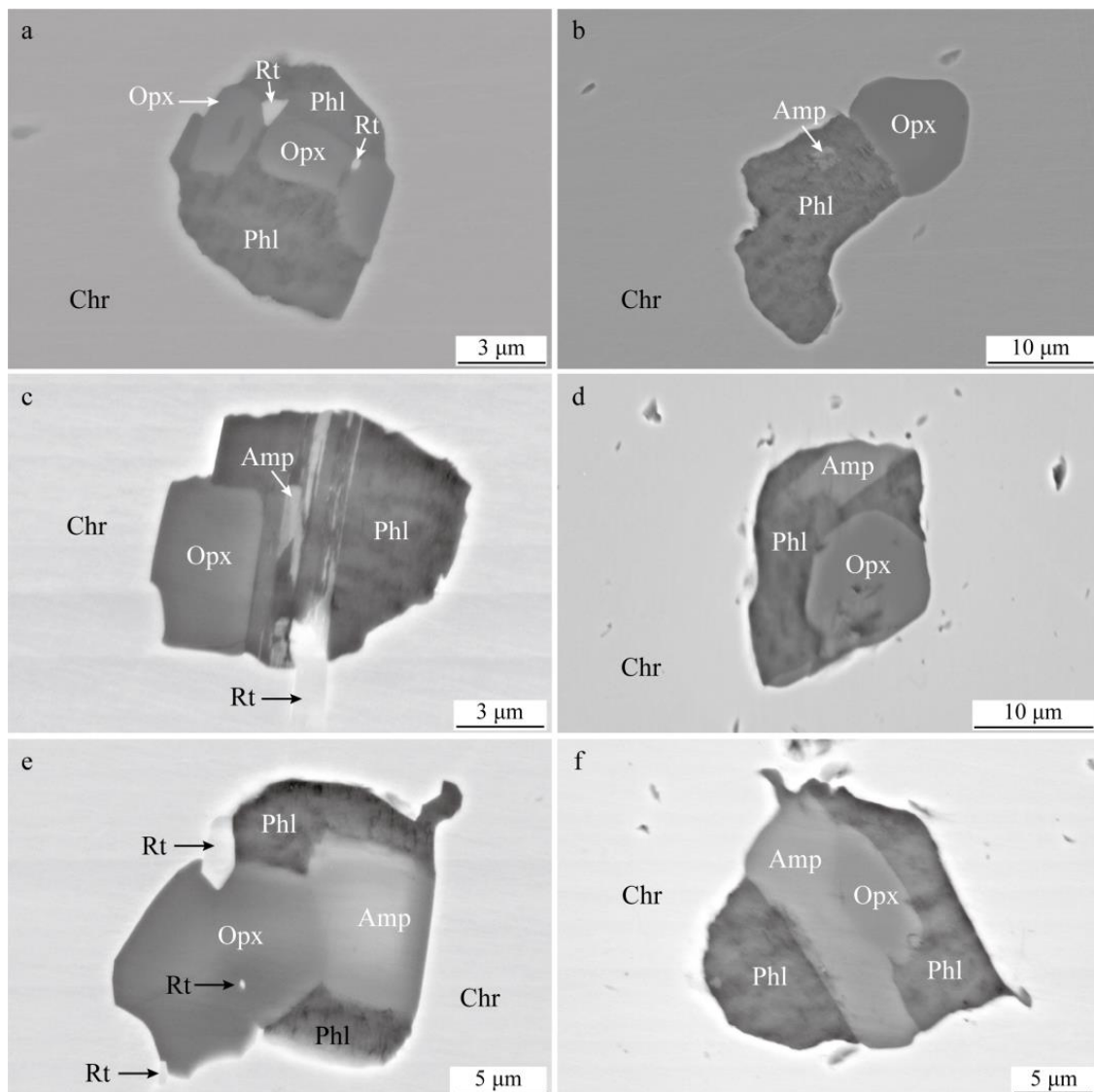


Figure 7. Back-scattered electron (BSE) images showing phlogopite inclusions in Sartohay chromitites. (a) Phlogopite coexisting with orthopyroxene. (b–f) Phlogopite coexisting with orthopyroxene and amphibole. Abbreviation: Amp = amphibole, Chr = chromite, Opx = orthopyroxene, Phl = phlogopite, Rt = rutile.

4.2. Mineral Composition

The $\text{Cr}^\#$ [$100 \times \text{Cr}/(\text{Cr} + \text{Al})$] values of chromites in Sartohay chromitites range from 52.2 to 55.3 with $\text{Mg}^\#$ [$100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$] values varying between 68.2 and 72.7, chemically akin to those from MORB (Figure 8a,b). They show low variability in Al_2O_3 (23.93–26.28 wt %), MgO (15.11–16.23 wt %), Cr_2O_3 (42.75–45.25 wt %), and FeO (14.11–15.99 wt %, Table S1). Chromites develop a trend with constant $\text{Cr}^\#$ and increasing TiO_2 wt % compositions, which coincides with the MORB–peridotite interaction trend [45] in Figure 8c. Their TiO_2 contents (0.22–0.50 wt %) are lower than those of chromite in MORB (usually >0.6 wt %) and overlap both those of MORB and the low-Ti tholeiitic lavas in the TiO_2 versus Al_2O_3 discrimination diagram (Figure 8d). According to the $\text{Cr}^\#$ values of chromite, the studied chromitites are classified as high-Al type [46,47], chemically similar to typical high-Al type chromitites [46] and high-Al chromitites previously reported in Sartohay [17]. However, they exhibit quite different features from high-Cr chromitites in Oman [48], Luobusa [49], and Sartohay [17].

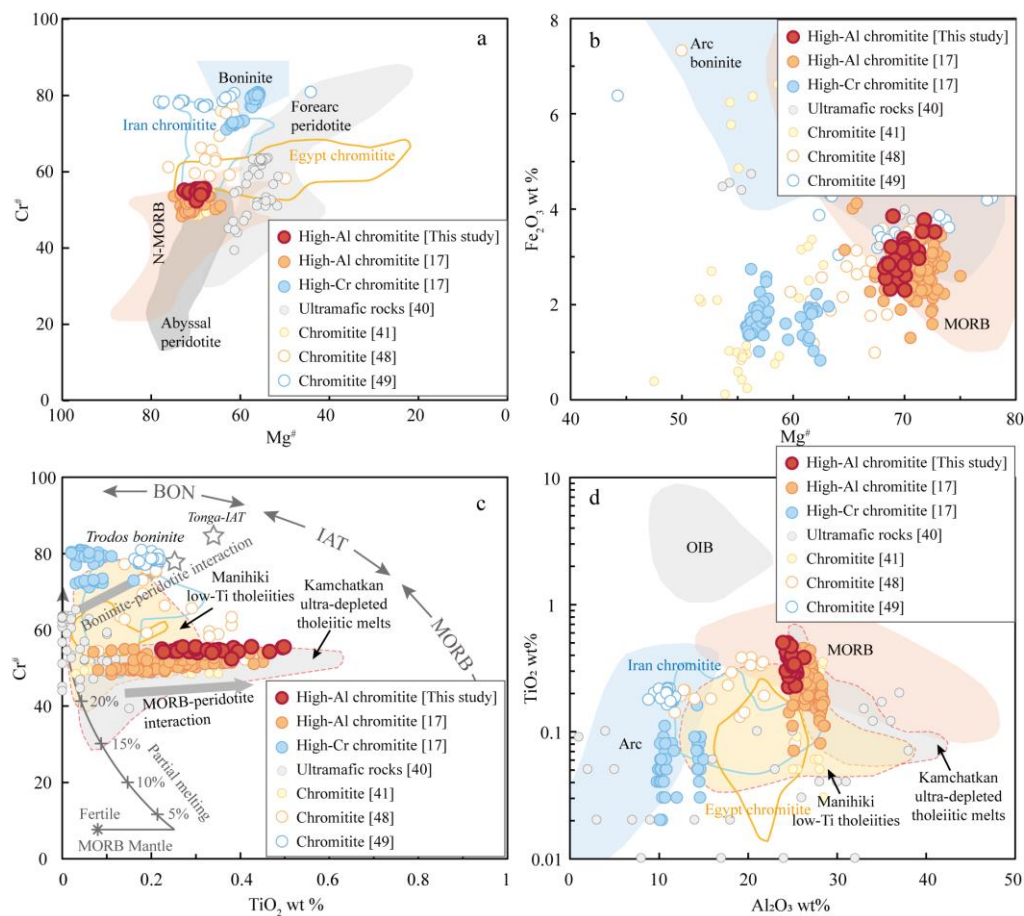


Figure 8. (a) Compositional variations in $\text{Cr}^\#$ versus $\text{Mg}^\#$ in chromite of Sartohay high-Al chromitites. Boninite, forearc-peridotite, abyssal peridotite, and N-MORB fields are from Dubois-Côté et al. [50]. (b) Fe_2O_3 contents (wt %) versus $\text{Mg}^\#$ values of high-Al chromitites from the Sartohay ophiolitic mélange. The field of MORB and arc boninites represents chromite compositions from Kamenetsky [51]. (c) Compositional variations in $\text{Cr}^\#$ versus TiO_2 (wt %) of chromite. The diagram discriminates between partial melting trends and melt-rock reaction fields, which are modified from Pearce and Parkinson [45]. (d) Compositional variations in TiO_2 (wt %) and Al_2O_3 (wt %) of chromite. ARC, MORB, and OIB compositional fields are adapted from Kamenetsky [51]. The data and ranges of chromite from the low-Ti tholeiitic lavas/melts of the Kamchatkan ophiolite (Far East) and Manihiki Plateau (SW Pacific) are from Portnyagin et al. [52] and Golowin et al. [53], respectively. Data of chromite in chromitites [17,41] and ultramafic rocks [40] from Sartohay ophiolitic mélange, and in chromitites from Luobusa [49], Oman [48], Iran [10], and Egypt [24] are shown for comparison. Abbreviation: BON = boninite, IAT = island-arc tholeiite, MORB = mid-ocean ridge basalt, OIB = ocean island basalt.

Olivine inclusions have high Fo [$100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$] values (96.4–97.6) and NiO contents (0.75–0.93 wt %) as a result of subsolidus re-equilibration with the host chromite. These compositions are comparable with those of Luobusa and Ray-Iz chromitites (Fo = 96–98) (Figure 9a) [13]. The MnO contents of olivine vary from 0.02 to 0.11 wt %, Cr_2O_3 contents from 0.63 to 0.96 wt %, and CaO contents < 0.30 wt % (Figure 9b,c; Table S2). The high concentrations of Cr_2O_3 in olivine are possibly obtained from the chromite micro (nano)-inclusions within olivine grains that are too tiny for observation (Figure 3b,c).

Enstatite in the studied host chromite of chromitites is dominant by ($\text{En}_{94-96}\text{Fs}_{4-6}\text{Wo}_{0-1}$), with $\text{Mg}^\#$ values ranging from 94.6 to 96.5 (Figure 10a). These inclusions contain 0.74–1.37 wt % Cr_2O_3 , 0.03–0.39 wt % TiO_2 , 0.10–0.30 wt % CaO, 0.48–1.11 wt % Al_2O_3 , and 2.76–3.72 wt % FeO. Na_2O content is <0.07 wt % (Figure 10b–d; Table S3). Diopside from the

chromite-hosted inclusions is dominant by (En_{48–50} Fs₃ Wo_{44–46}) with a wide range of Mg[#] values (93.6–96.4) (Figure 10a). These inclusions are low in Na₂O (0.50–0.79 wt %), TiO₂ (0.20–0.36 wt %), and FeO (1.74–2.10 wt %), and slightly higher in Cr₂O₃ (1.54–2.65 wt %) (Figure 10b–d; Table S4).

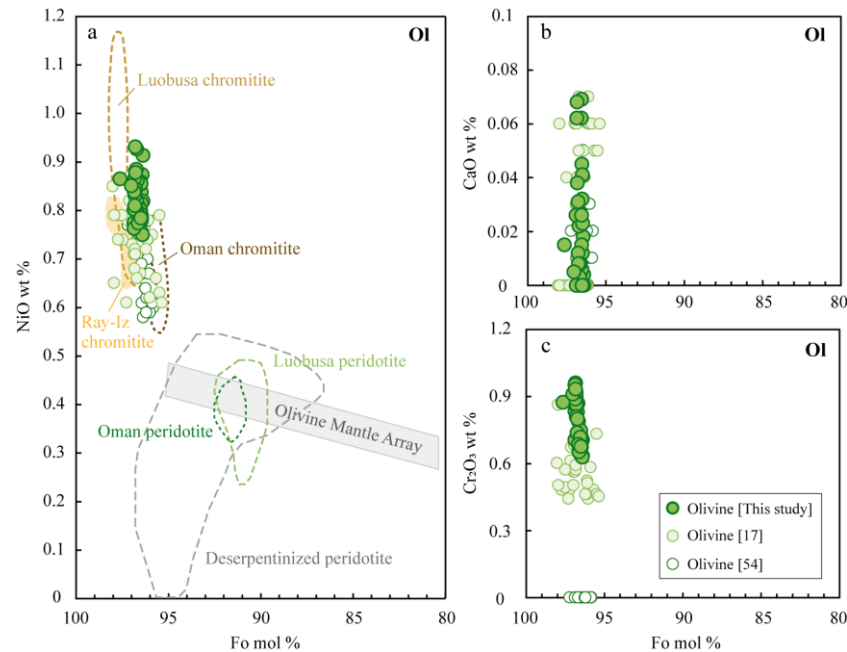


Figure 9. (a) NiO, (b) CaO, and (c) Cr₂O₃ (wt %) contents versus forsterite (Fo) of olivine inclusions in the Sartohay chromite. The fields of olivine in chromitites and mantle peridotites from Luobusa, Ray-Iz, and Oman are in accordance with Arai and Miura [13]. Data on olivine from Sartohay chromitites in previous studies [17,54] are plotted.

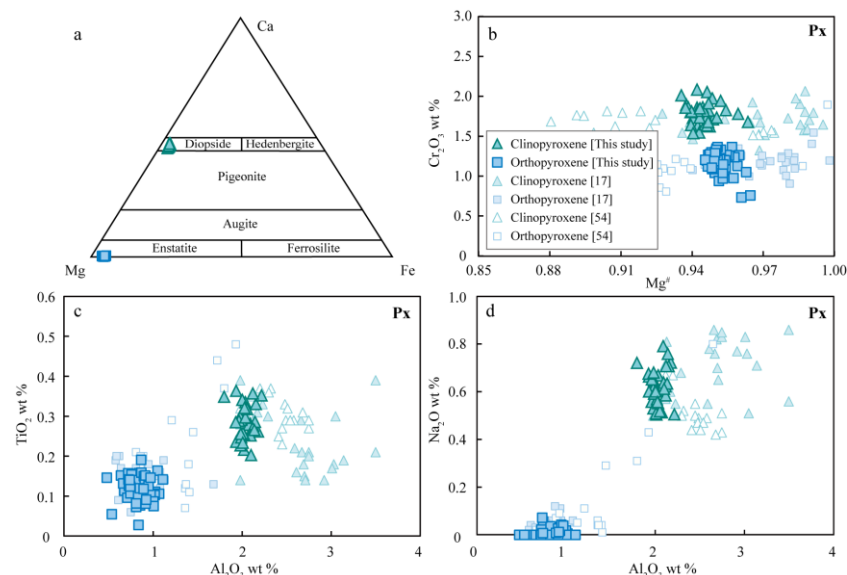


Figure 10. (a) Classification diagram of pyroxene inclusions within the chromite of Sartohay chromitites. (b) Cr₂O₃ (wt %) versus Mg[#], (c) TiO₂ versus Al₂O₃ (wt %), and (d) Na₂O versus Al₂O₃ (wt %) diagram for pyroxene inclusions. Data on pyroxene inclusions in chromitites from the Sartohay ophiolitic mélange of previous studies [17,54] are shown for comparison.

Amphibole inclusions are classified as pargasite with Mg[#] values above 93.4 and Si atoms vary from 6.2 to 6.4 apfu (atoms per formula unit) (Figure 11a). These inclusions

contain high Na_2O (2.82–3.48 wt %) and TiO_2 (2.14–4.17 wt %) contents but low K_2O (<0.06 wt %) contents (Figure 11b; Table S5).

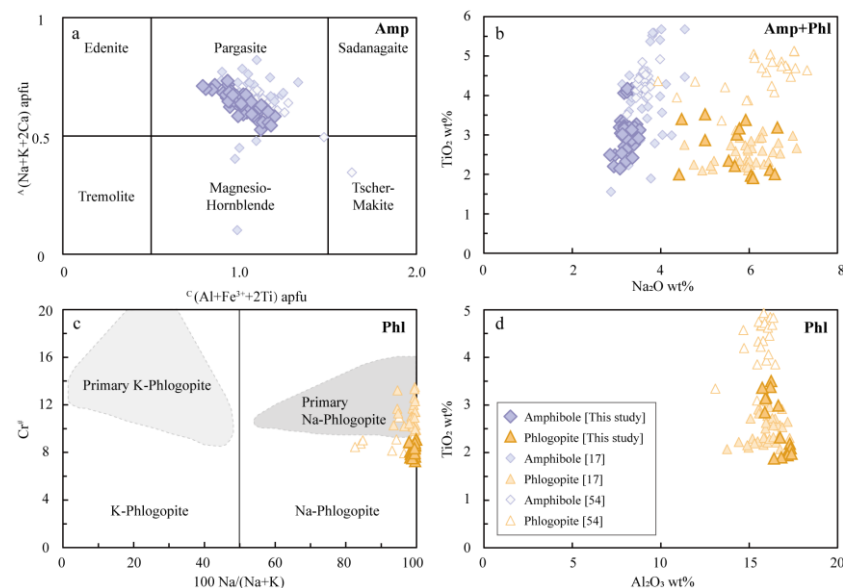


Figure 11. (a) Classification diagram of amphibole inclusions in chromite from Sartohay chromitites according to Hawthorne et al. [55]. (b) TiO_2 versus Na_2O (wt %) of amphibole and Na-phlogopite inclusions; (c) $100 \times \text{Na}/(\text{Na} + \text{K})$ versus $\text{Cr}^\#$, and (d) TiO_2 versus Al_2O_3 (wt %) of Na-phlogopite inclusions in chromite from Sartohay chromitites. Compositions are compared with data from the literature on amphibole and Na-phlogopite inclusions in Sartohay chromite [17,54].

Phlogopite inclusions are classified as Na-phlogopite (Figure 11c) with high $\text{Mg}^\#$ values (97.6–98.0) and Na_2O contents (4.38–6.58 wt %) (Figure 11b). TiO_2 , Al_2O_3 , and Cr_2O_3 contents in Na-phlogopite are 1.88–3.51 wt %, 15.76–17.42 wt %, and 2.06–2.45 wt %, respectively (Figure 11d; Table S6).

5. Discussion

5.1. The Genesis of Silicate Inclusions

Primary silicate mineral inclusions are frequently encountered in chromite, with documented occurrences reported in podiform chromitites from Oman [56]; Sabzevar in Iran [10]; Eastern Desert of Egypt [24]; and Luobusa [57], Zedang [58], and Zunhua [16] in China. Various hypotheses on the origin of silicate mineral inclusions embedded in chromite have been proposed based on the differences in inclusion sizes, morphologies, distribution, and coexistence relationships.

The solid-phase mechanism [6,59–61] provides a rational genesis model for silicate inclusions and host chromite grains. This hypothesis suggests that the formation of chromites is caused by plastic deformation of rock-forming silicates, as evidenced by typical petrographic facts like chromite, pyroxene, and pargasite lamellae or neoblasts in silicate matrix of peridotites. With this process, the decomposition of pyroxenes and olivines forms chromites associated with tiny grains of amphibole and diopside, which may be captured by growing chromite grains during progressive deformation and recrystallization. However, no petrographic evidence of plastically deformed olivine and pyroxene crystals has been identified in Sartohay chromitites and host peridotites [17,40,41,54]. Moreover, the common existence of Na-phlogopite, especially as a single inclusion in the chromite of Sartohay chromitites (Figure 6f), cannot be adequately explained. Notably, the implementation of the rheomorphic mechanism requires no external sources, making it hard to rationalize the rutile as a crustal-derived mineral in inclusions (Figure 7a,e). Therefore, the studied inclusions in Sartohay chromitites may be more applicable to genesis models of magmatic origin.

Three models considering magmatic genesis for the inclusions are mainly presented: (1) trapped droplets of influx liquids/melts, enriched with volatile components and incompatible elements resulting from melt–peridotite interactions and melt mixing, rapidly crystallized upon closed-system cooling within liquidus chromite [62,63]; (2) the entrapment of exotic fluids [64] or volatile-rich self-differentiated melts [65] reacting with anhydrous silicates; (3) the entrapment of previously crystallized minerals as xenoliths or xenocrysts at the early magmatic stage, in presence of hydrothermal fluids during subsolidus annealing of chromite [16,46,57]. However, neither of these interpretations can explain the formation of all inclusion types [66]. Inclusions in chromite are characterized by varied features and therefore may have different geneses.

Olivine, orthopyroxene, and clinopyroxene are the dominant liquidus line minerals of the mantle peridotites related to podiform chromitites. Isolated inclusions of these minerals in the chromite from Sartohay chromitites are relatively small and preserve euhedral–subhedral crystal shapes (Figures 3b and 5a). The morphology of the inclusions is related to the crystallization sequence with its host chromite [25], implying that these inclusions crystallized earlier than chromite, and were captured as individual crystals. Additionally, entrapped melts and post-entrapment interactions with solid phases are unlikely to precipitate single nominally anhydrous phases [24], supporting them as previously crystallized minerals. The dunite envelope surrounding podiform chromitites in Sartohay is significant evidence for melt–peridotite interaction [62]. These inclusions possibly crystallized from the melt during the interaction, whereas some inclusions with a corroded or subrounded form might experience various degrees of resorption due to changing melt physical–chemical conditions while ascending with host chromite.

Except as solitary inclusions, orthopyroxene and clinopyroxene coexist as multiple-phase mineral assemblages within a single inclusion (Figure 5d–f). They display subrounded boundaries with tiny cracks inside, indicating their formation as primary melt inclusions crystallized into solid phases. Foliated enstatite lamellae with crystallographic orientation exsolved from the host diopside (Figure 5e,f). Similar exsolution textures of orthopyroxene in clinopyroxene were reported from igneous rocks in Tianshan [67] and Bohemian Massif [68], and they were considered to have been formed by initial homogeneous clinopyroxene separation into equilibrium phases during post-magmatic cooling and/or decompression. Since the melt is rapidly enclosed by the chromite, the evolution of the melt occurs individually within each inclusion. The homogeneous diopside solid solution may have initially crystallized from entrapped melt and then became metastable due to changes in physicochemical conditions [67]. The exsolution process may be facilitated by a decrease in pressure and temperature as the inclusion ascends with host chromite in the mantle.

Amphibole, the most common hydrous silicate, occurs as isolated inclusion or co-occurs with enstatite–diopside and/or Na-phlogopite in an individual inclusion within chromite from Sartohay chromitites (Figure 6). The absence of primary amphibole or Na-phlogopite as interstitial minerals among chromite grains indicates that they do not represent solid phases trapped by chromite during grain boundary migration owing to sub-solidus annealing [69]. These composite inclusions generally exhibit anhedral granular shapes, and the minerals in them follow a crystallization sequence (Figures 5c and 6c). These observations suggest that the inclusions represent the gradual crystallization product of trapped melt. Alternatively, these inclusions could be formed by a reaction between solid minerals captured by chromite with melt.

Amphibole could form in three mechanisms: (1) the reaction between enstatite and Na-, and Al-rich melt (<1000 °C, ~3.8 wt % H₂O) [70]; (2) diopside reacts with encapsulated magma [71]; and (3) amphibole and Na-phlogopite crystallized straight from a Na-riched silicate melt (700–1000 °C, ≥0.5 kbar in the presence of H₂O or H₂O–CO₂ phase) [25]. Regarding single-amphibole and Na-phlogopite inclusions with irregular shapes, they may directly precipitate from melt globules trapped within chromite (Figure 6b,f), whereas composite inclusions containing amphibole can be classified into two types. Type 1 are inclusions in which amphibole metasomatized diopside (Figure 5c) and enstatite (Figure 6c),

leaving residual pyroxene with corroded margins at the core of amphibole. Their genesis may be attributed to the enstatite–diopside crystallization in the trapped melt at the early stage, followed by the reaction with the residual Na, and Al-rich melts to form amphibole. Type 2 exhibits a clear boundary between amphibole and enstatite (Figure 4c,d) diopside (Figure 5e), or Na-phlogopite (Figure 6f), with these minerals in close proximity to each other. Such inclusions may have crystallized directly from primary melt inclusions into equilibrated solid mineral phases, as evidenced by the petrography below: (1) the interlocking morphology of the inclusions suggests that the phases crystallized together; thus, the captured inclusion was originally fully molten (Figure 7c); (2) small cracks or holes embedded in these inclusions may be the product of the final evolution stage of the parental melt (Figure 7d); (3) some inclusions display angular shapes or project into the host chromite, indicating a reaction between melt and chromite (Figure 7e,f); and (4) some inclusions contain small sulfide grains representing immiscible sulfide melt droplets (Figure 6d). The chemical compositions of these inclusions display high Na₂O (up to 3.48 wt % in amphibole and 6.58 wt % in Na-phlogopite) and TiO₂ contents (up to 4.17 wt % in amphibole and 3.51 wt % in Na-phlogopite) (Figure 11b). The trapped MORB-like melts produced after melt–peridotite interaction are generally enriched in SiO₂, Na₂O, TiO₂, and H₂O [18,62,66], which is capable to precipitate Na- and Ti-rich amphibole and Na-phlogopite. Hence, most amphibole inclusions crystallized directly from the captured melt, with a few originating from melt interactions with pre-existing minerals. The diversity of mineral assemblages found in the studied inclusions likely reflects the heterogeneous nature of entrapped melt.

5.2. Estimates of the P-T-fO₂ Conditions

Estimation of P-T-fO₂ formation conditions can be carried out using primary mineral assemblages in chromitite (Figure 12). Equilibrated clinopyroxene–orthopyroxene assemblages in chromite inclusions serve as well-calibrated thermobarometers for magmatic systems [15,56,72]. The estimation model’s precision improved in mafic systems where the Mg[#] of clinopyroxene was >0.75 [21]. The enstatite–diopside inclusions in Sartohay chromitites show interlocking texture, rounded form, and small grain size (Figure 5d–f), indicating that they are equilibrated primary inclusions. Considering the best calibration, twenty-five enstatite–diopside pairs with K_D (Fe-Mg)_{Cpx-Opx} of 1.01–1.55 (within the range of 3σ) were selected for estimation based on Fe-Mg exchange coefficients [73]. The two-pyroxene thermometer by Putirka [21] yielded results of 874–1048 °C (Figure 12a). Composition profiles across pyroxene inclusions show that the Mg[#] of pyroxene grain gradually increases from core to rim (Figure 13a,b), indicating a subsolidus re-equilibration process between the host chromite and pyroxene. Thus, the calculated equilibrium temperature condition can only represent the closure temperature of the Mg-Fe exchange of the two pyroxenes and is not equal to the primary magmatic condition of chromitites. The temperature-independent two-pyroxene barometer by Putirka [21] yielded a pressure of 8.0–21.6 kbar (±4.0 kbar) for pyroxene crystallization in Sartohay chromitites (Figure 12a). The wide pressure variation can be interpreted as prints of different depth levels preserved in the chromitites as they continuously crystallized during the melt ascending in the uppermost mantle [46]. This is in accordance with the exsolution texture found in pyroxene (Figure 5e,f). The calculated P-T ranges are consistent with those reported from Zedang chromitites (996–1097 °C, 8.5–21.5 kbar) [58] and Hegenshan chromitites (940–1048 °C, 4.9–12.9 kbar) [74]. Moreover, the common presence of amphibole and Na-phlogopite coexisting in a single inclusion (Figure 6g,h) constrains a rapid cooling rate in chromite at a temperature of 1000–1050 °C and pressure ~10–20 kbar [75].

Estimation of fO₂ values of parental melts of podiform chromitite and peridotites is mainly based on equilibrium mineral assemblages of coexisting olivine and chromite [19]. However, the Sartohay podiform chromitites and host peridotites have been significantly altered by later metamorphic deformation and hydrothermal alteration [17], transforming the primary interstitial olivine grains in chromitite into secondary hydrous minerals. Olivine

inclusions are considered early crystallized solid silicates trapped by chromite; hence, they can be used herein to indicate primitive magmatic conditions (Figure 3b). Compositions of chromite and olivine, temperature, and pressure govern the calculations. The temperature was assumed to be 874–1048 °C, and the temperature was 8.0–21.6 kbar, as yielded by the two-pyroxene geothermobarometer. The oxygen barometer yielded 1.3–2.0 log units above the fayalite–magnetite–quartz (FMQ) buffer for Sartohay chromitites (Figure 12b).

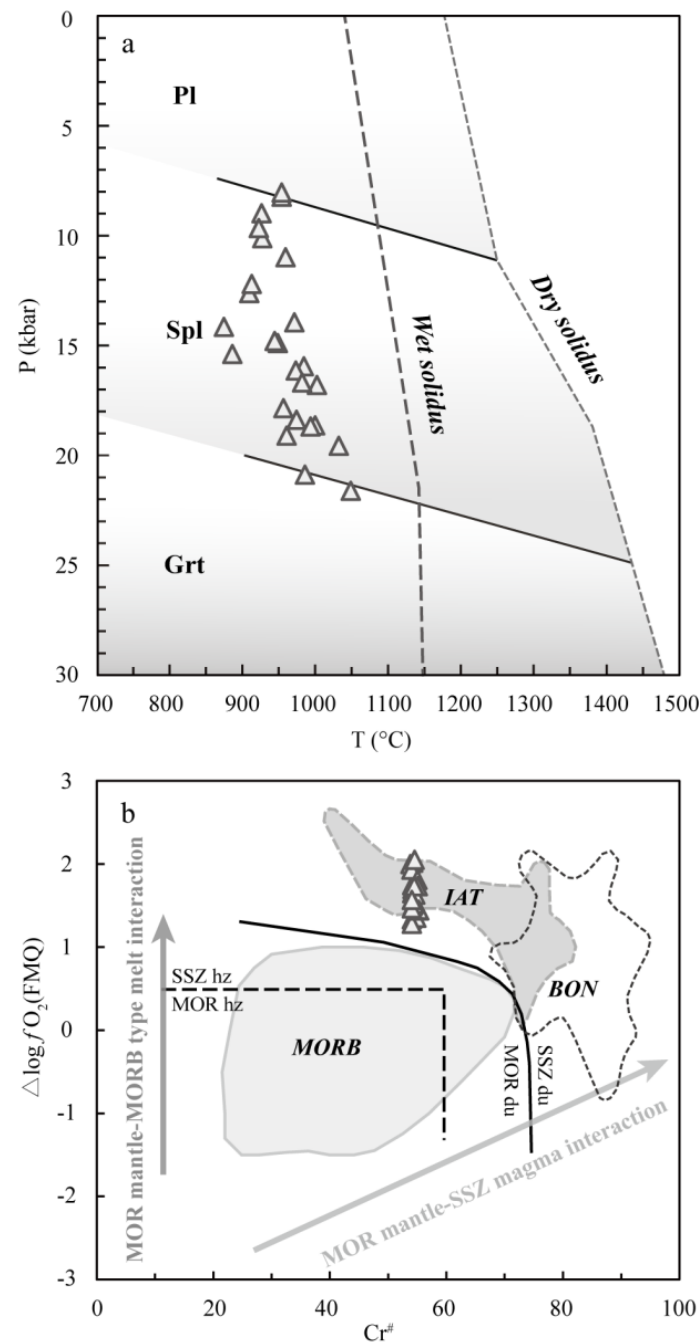


Figure 12. (a) Diagram of P-T conditions based on pyroxene inclusions in Sartohay chromitites using geothermometers and geobarometers [21]; the solidus curves are for mantle peridotites. (b) $\Delta \log f_{O_2}(\text{FMQ})$ versus $\text{Cr}^\#$ of chromite in Sartohay chromitites. The regions of various magmas and the MOR-SSZ discrimination boundaries for dunites (solid line) and harzburgites (dashed line) are from Dare et al. [20]. Abbreviation: BON = boninite, IAT = island-arc tholeiite, MORB = mid-ocean ridge basalt, SSZ = supra-subduction zone.

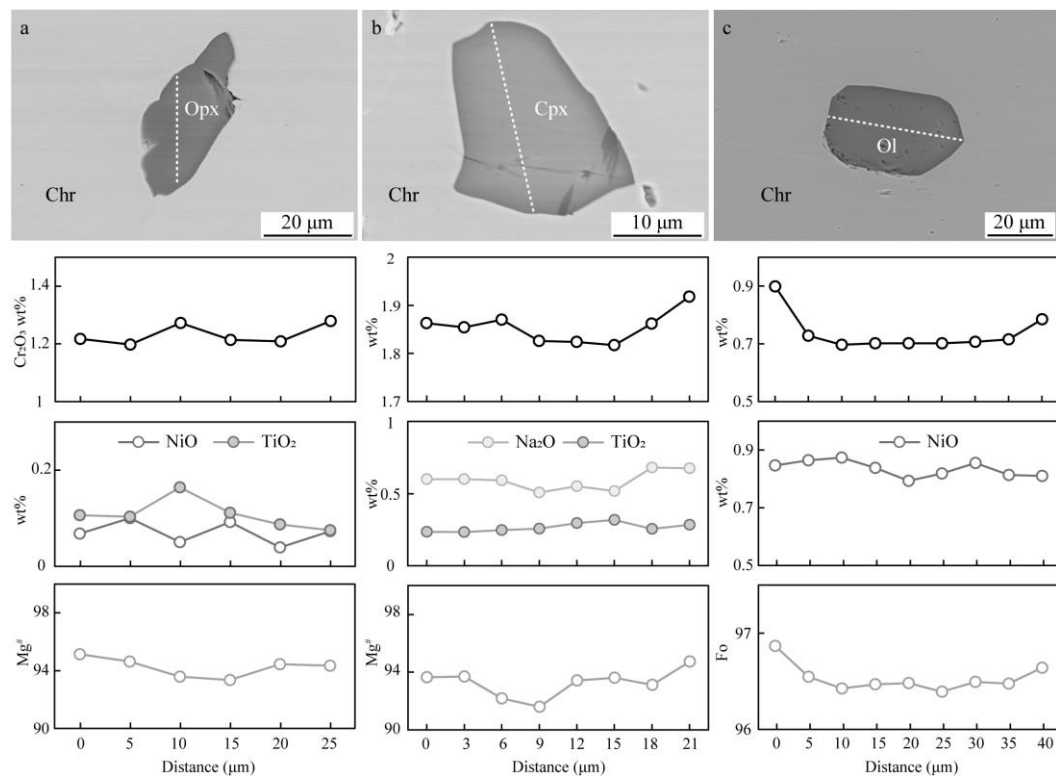


Figure 13. Compositional correlation diagrams, characteristics, and oxide (wt %) concentration profiles of mineral inclusions in Sartohay podiform chromitites. (a) BSE image and composition profile of orthopyroxene. (b) BSE image and composition profile of clinopyroxene. (c) BSE image and composition profile of olivine. Abbreviation: Chr = chromite, Cpx = clinopyroxene, Ol = olivine, Opx = orthopyroxene.

5.3. Constraints for Parental Melts of Chromitite

Chromite can be used to constrain the nature of parental melts formed in different tectonic environments [10,23]. The Al_2O_3 and TiO_2 concentrations in chromite are directly correlated with the composition of the parental melt, as Al and Ti experience very little change during subsolidus re-equilibration between chromite and inclusions due to their low diffusivity [51,76]. The equation proposed by Rollinson [47] was used to calculate the Al_2O_3 and TiO_2 contents in a modal parental melt of the Sartohay high-Al podiform chromitite.

$$\text{Al}_2\text{O}_3 (\text{melt}) = 7.1518 \text{ Al}_2\text{O}_3 (\text{chromite})^{0.2387} (\text{Al}_2\text{O}_3 \text{ in wt \%})$$

$$\text{TiO}_2 (\text{melt}) = 1.5907 \text{ TiO}_2 (\text{chromite})^{0.6322} (\text{TiO}_2 \text{ in wt \%})$$

Subsolidus Fe-Mg exchange between chromite and silicate mineral is limited with ≥ 80 vol% chromite in the massive texture Sartohay chromitites. Thus, chromite can retain its original components. The FeO/MgO ratios of the melt in equilibrium with chromite were determined using the empirical formulation of Maurel and Maurel [77] as follows:

$$\ln(\text{FeO/MgO})_{(\text{chromite})} = 0.47 - 1.07 \text{ Al}^{\#}_{(\text{chromite})} + 0.64 \text{ Fe}^{3+\#}_{(\text{chromite})} + \ln(\text{FeO/MgO})_{(\text{melt})}$$

where the FeO and MgO are in wt %, $\text{Al}^{\#} = \text{Al}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$ and $\text{Fe}^{3+\#} = \text{Fe}^{3+}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$.

Implementing these approaches, the melt in equilibrium with studied chromitites has 15.26–15.61 wt % Al_2O_3 (Figure 14a), 0.62–1.02 wt % TiO_2 (Figure 14b), and 0.86–1.03 FeO/MgO ratio (Figure 14c). The calculated Al_2O_3 content and FeO/MgO ratio of the parental melts are comparable to those of other high-Al podiform chromitites, including the Oman ophiolite

($\text{Al}_2\text{O}_3_{(\text{melt})} = 14.5\text{--}15.4 \text{ wt } \%$) [47], Sagua de Tanamo in Cuba ($\text{FeO}/\text{MgO}_{(\text{melt})} = 0.90\text{--}1.10$, $\text{Al}_2\text{O}_3_{(\text{melt})} = 15.0\text{--}16.0 \text{ wt } \%$) [76], and Othris massif in Greece ($\text{FeO}/\text{MgO}_{(\text{melt})} = 0.40\text{--}1.23$, $\text{Al}_2\text{O}_3_{(\text{melt})} = 14.40\text{--}16.13 \text{ wt } \%$) [18], showing MORB affinity. However, the TiO_2 values of parental melts are overall lower than the TiO_2 ranges of MORB lavas ($\sim 0.6\text{--}1.7 \text{ wt } \%$) [23,51]. TiO_2 contents of chromite are lower than that of chromite in MORB (Figure 8c), and the $f\text{O}_2$ values of the investigated chromitites plotted in the IAT field indicate a more oxidized parental magma than that of MORB (Figure 12b). This implies that the high-Al chromitites are not products of typical MORB-like magmas but those with more depleted compositions, for example, with low-Ti tholeiitic affinities (Figure 8c,d). Such tholeiitic lavas/melts associated with chromite were reported in the Kamchatkan ophiolite (Far East) and Manihiki Plateau (SW Pacific) and are comparable in composition to the chromite in our chromitites (Figure 14b) [52,53]. The inferred parental melts of Sartohay chromitites fall within the low-Ti tholeiitic lavas field of Lasail–Alley from Semail ophiolite ($\text{MgO} = 6.3 \text{ to } 13.7 \text{ wt } \%$) [78], with a trend to migrate toward the MORB field (Figure 15a). Data from previous studies of the same region from Sartohay chromitites [17] are partially plotted in the field of experimental melts of the depleted mantle ($\text{MgO} = 11 \text{ to } 17 \text{ wt } \%$) [79], suggesting that the parental melts of Sartohay chromitites may initially have high Mg contents. A strong positive correlation between MgO and FeO for chromite (Figure 15b) suggests that the composition of the parental melt is continuously evolving, and chromite may undergo various degrees of modification. Therefore, the above-calculated values are not representative of the primary melt for the Sartohay chromitite.

The chromite effectively preserves the original composition of inclusions due to its stable structure. These inclusions trapped in early crystallizing chromite offer direct insights into the composition of the magma from which the chromite crystallized. To determine the composition of the primary melt, the method developed by Rollinson et al. [26] that combines the trendline of chromite on the MgO–FeO diagram and the composition of the olivine in the dunite envelope was utilized. Olivine inclusions, likely crystallized from melt interacting with peridotite to form the dunite envelope, sharing genesis with dunite envelope olivine. The Fo values of olivine as inclusions in chromite were higher than that of olivine in most igneous and metamorphic rocks. This is because the $\text{Mg}^\#$ s of chromite and olivine in chromitites change from their initial magmatic values via Mg–Fe²⁺ redistribution during subsolidus cooling and the low modal abundance of olivine [13]. The well-preserved olivine inclusion ($\sim 40 \mu\text{m}$) used for the calculation exhibits a constant Fo value at the core (Fo = 96.4) (Figure 13c), suggesting minimal Fe–Mg exchange between the olivine core and the host chromite. Using this core Fo value as the initial Fo value, and with the redetermined olivine–melt Fe–Mg partition coefficient of 0.34 [80], the FeO and MgO contents of the original melt were determined. The intersection of the olivine–melt equilibria and the chromite trendline (Figure 15b) revealed that the parental melts of the Sartohay podiform chromitites had MgO contents of $\sim 22.7 \text{ wt } \%$ and FeO contents of $\sim 4.4 \text{ wt } \%$, signifying a Mg-rich nature. This is consistent with observed high $\text{Mg}^\#$ values in mineral inclusions that record the composition of the primary magma, such as olivine (Fo = 96.4–97.6) (Figure 9), pyroxene (0.94–0.96) (Figure 10b), amphibole (>0.93) (Figure 11a), and Na-phlogopite (~ 0.98). The notable difference between primary melt (MgO = 22.7 wt %) and calculated tholeiitic melt (Figure 15a) suggests an evolutionary process from high Mg to low Mg for primary magma in Sartohay podiform chromitites.

Melt inclusions within chromite could represent the composition of parental magma [24,47]. The composition of primary hydrous inclusions indicates that the trapped melts are enriched in Na_2O (up to 3.48 wt % and 6.58 wt % in amphibole and Na-phlogopite, respectively) (Figure 11b). The presence of high-CaO silicate minerals in Sartohay chromitites, such as diopside (Wo = 44.4–46.4) and amphibole (CaO = 11.1–12.2 wt %), indicates the Ca-rich nature of the chromitite parental magmas. CaO plays a crucial role in the formation of chromitite [22]. An experimental study exploring the effect of melt composition on $\text{Cr}^{2+}/\text{Cr}^{3+}$ in silicate melts has revealed a negative correlation between $\log \gamma^\# \text{Cr}^{3+}\text{O}_{1.5}$ and X_{CaO} [81]. Hence, the involvement of a Ca-rich component in parental melts of chromitite facilitates the transportation

of Cr^{3+} in magma conduits, ultimately leading to the deposition of chromite under suitable physical conditions.

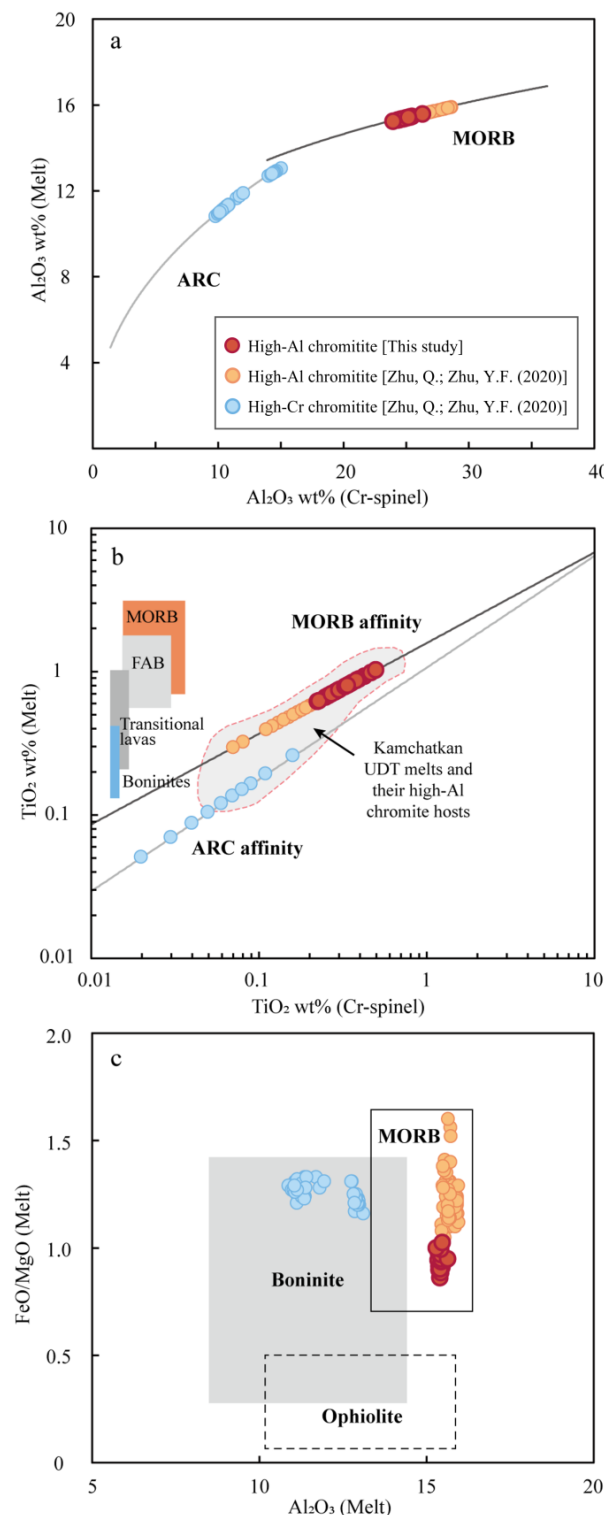


Figure 14. Calculated (a) Al_2O_3 (wt %) and (b) TiO_2 (wt %) contents of the melts in equilibrium with Sartohay podiform chromitites [17]. The empirical formulas and data for Oman chromitites are from Rollinson [47]. The range of ultra-depleted tholeiitic melts and their hosted high-Al chromite grains from the Kamchatkan ophiolite from Portnyagin et al. [52] are shown. (c) FeO/MgO versus Al_2O_3 (wt %) of the parental melt was estimated based on the chemical composition of chromite from Sartohay chromitites. The tectonic discrimination fields are from Barnes and Roeder [82].

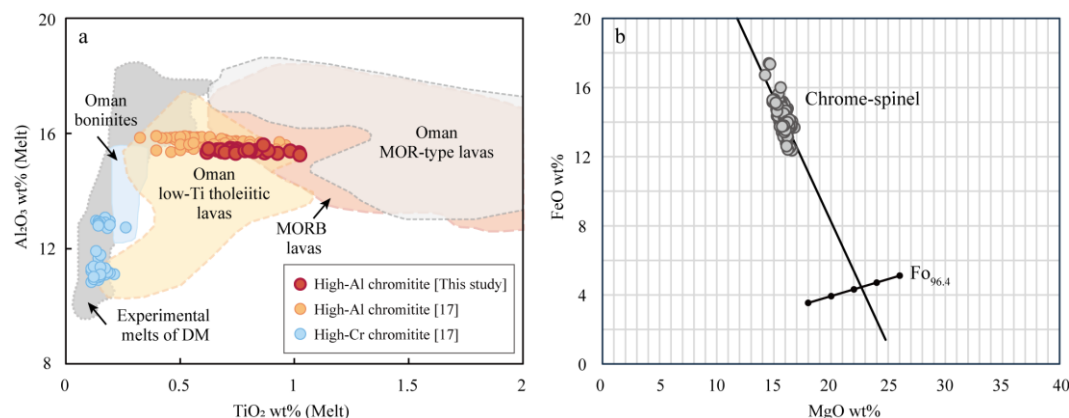


Figure 15. (a) TiO_2 vs. Al_2O_3 (wt %) diagram of parental melts in equilibrium with chromite from Sartohay chromitites; the compositional field for experimental melts of depleted mantle and boninites and lavas from the Oman ophiolite are shown according to Rollinson [47]. High-Al and -Cr Sartohay chromitite data are from Zhu and Zhu [17]. (b) MgO versus FeO (wt %) diagram, which is used for calculating the composition of primitive melt by the combination of chromite and olivine inclusion in Sartohay chromitites.

5.4. Genesis of Chromitites in Sartohay

Two mechanisms are proposed to facilitate the continuous accumulation of chromite in chromite-saturated melts: (1) in immiscible basalt–water systems saturated with olivine and chromite, olivine tends to reside in the melt, while chromite consistently enriches in the fluid phase [11,14], and (2) when droplets of low-silica (low-viscosity) melt mingle with siliceous melt, chromite crystallizes only in the low-silica melt owing to the lowest interfacial energy [83]. The prevalence of hydrous silicate inclusions (amphibole and Nephlogopite) in the Sartohay podiform chromitite suggests the involvement of fluids in the chromitite parental magma (Figure 7). Olivine inclusions with high Fo values and relatively low Al_2O_3 and CaO contents likely indicate their crystallization from hydrous melts (Figure 9b) [84]. The elevated Cr_2O_3 content in olivine may be attributed to a component exchange with Cr-rich fluids released from the melt (Figure 9c) [85]. Given that the water content of the primary mid-ocean ridge basalts typically remains below 0.4 wt %, and the low-Ti tholeiite magmas exhibit even lower H_2O contents (<0.2 wt %) [53], the presence of fluid incorporation has consequently elevated the water content in the MORB-like parental magma of podiform chromitites. This elevation facilitates the crystallization of hydrous silicates from the captured melt. Since Cr is a compatible element in all phases of mantle peridotite (with distribution coefficients mostly between 1 and 10) [86], the addition of hydrous melts or fluids into the source enhances the degree of partial melting of mantle peridotites, resulting in Cr accumulation in the parental melt and the formation of podiform chromitites.

Dunite envelopes surrounding podiform chromite ore are observed in the Sartohay and are common features in almost all ophiolites [i.e., Oman [26], Troodos in Cyprus [66], Kizildag in Turkey [3], and Luobusa [12] in China]. As proposed in the rheomorphic model, the localization of plastic flow in these dunites results in the formation of chromite grains by redistribution of mineral phases [7,8], with evidence including plastic deformation texture, “squeezed” interstitial silicates, or chromite folding in chromitite. However, such textural and structural features were not observed in Sartohay chromitites (Figure 2). Thus, alternative interpretations need to be addressed for the dunites. The presence of these dunite envelopes is considered significant evidence for the melt–peridotite interaction model [62], supported by the petrographic transition relationships, the PGE partitioning signature, and isotopic characteristics [12,87]. It should be noted that the system of podiform chromitite production should be much larger than the size of chromitite mining areas to ensure that the melt in the interaction can provide sufficient Cr for chromite precipitation [13,87]. The

disparity in scale between melt and peridotite suggests that the majority of the melt is likely to be located at depth. Moreover, the area for Cr collection is different from that for chromite precipitation [13]. The host rocks involved in the melt–peridotite reaction are not the only mantle peridotites corresponding to the range of dunite, which accounts for the weak correlation in the size of chromitite with the dunite envelope and the locally direct contact between chromitite and harzburgite (Figure 16a–c). Despite the possible differences in melt chemistry, the melt–peridotite interaction will result in the dissolving of pyroxene and precipitation of olivine in the mantle peridotite [62]. This reaction forms dunite envelopes, driving the primary magma into the field of chromite precipitation and forming a modified, silica-enriched melt [88].

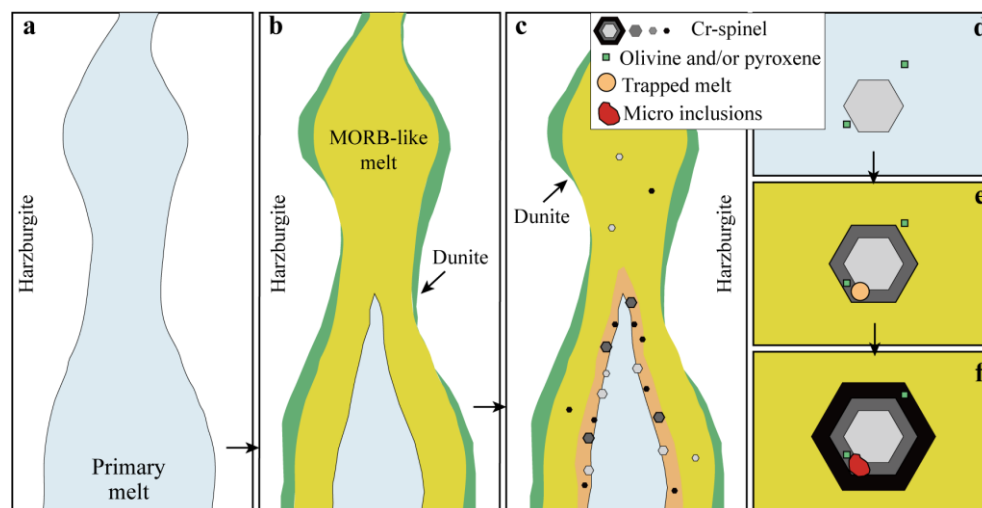
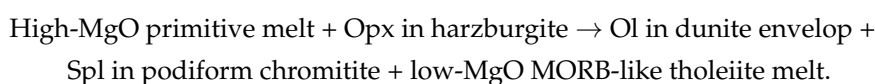


Figure 16. Formation model of microinclusions and host podiform chromitites in Sartohay ophiolitic mélanges (not to scale). The highly magnesian primary melt (a) reacts with mantle harzburgites and (b) forms an evolved MORB-like melt. (c) Mixing/mingling of the evolved melt with subsequently injected primary melt triggered chromite precipitation, leading to the formation of the Sartohay podiform chromitites. (d) Crystallization of chromite and solid silicate minerals like olivine and/or pyroxene in the primary melt. (e) Chromite grows continually to capture solid minerals and melt. (f) The trapped melt crystallizes into micro-silicate inclusions in chromite.

The composition of chromite (Figure 8) and calculated parental melts (Figure 14) are similar to those previously reported for Sartohay high-Al chromitites [17]. The formation of studied chromitites might resemble the genetic relationship proposed by Zhu and Zhu [17] via interaction between upwelling MORB-like melts and the harzburgites. The petrographic observations of pyroxenes replaced by amphibole (Figures 5c and 6c) and Na-phlogopite (Figure 7a) provide strong evidence for the hydrous melt–peridotite reaction. However, separate melt–peridotite interaction cannot adequately explain the massive aggregation of chromite [12]. Previous studies have concluded that a combination of melt–peridotite interaction with the mixing of primary melt and subsequent secondary melts can elucidate the genesis of podiform chromitites [46,57,76,88,89]. The hypothesis of melt mingling aligns with the abundant occurrence of silicate mineral inclusions in Sartohay chromitites (Figure 3a). Melt–peridotite interactions and subsequent mixing/mingling of melts could contribute to the evolution of Mg-rich, low-silica primary melts to tholeiite melts in Sartohay chromitites. A possible melt–peridotite reaction would be as follows:



In conjunction with the preceding discussion, a mechanism integrating the melt mixing process and the hydrous nature of the parental magma can elucidate the genesis of

chromitites. According to the equilibrium P-T conditions of clinopyroxene–orthopyroxene (874–1048 °C, 8.0–21.6 kbar), the Sartohay chromitite formed in the uppermost mantle. The primary melt of Sartohay podiform chromitites, being hydrous and enriched in Mg, undergoes melt–peridotite reactions with the surrounding mantle peridotite as it ascends through the upper mantle (Figure 16a). This reaction process elevated Si, Al, Ca, Na, and Cr in the evolved MORB-like tholeiite melt, leaving dunite envelopes at locations with a high degree of reaction (Figure 16b). As the evolved siliceous melt unevenly mixes with the primary low-silica melt, chromite continuously enriches the low-silica melt. Sustained melt–peridotite reactions ensure the ongoing mixing in the magma conduit, effectively preserving the composition of the mixed melt within the stable domain of chromite for an extended period and facilitating the continuous crystallization of chromite (Figure 16c). The high water content in the parental magma may lead to the exsolution of hydrous fluid phases, resulting in the enrichment of chromite in the hydrous fluid phase, thereby promoting the formation of chromitite ore.

The liquidus minerals crystallizing no later than chromite from the parental magma include olivine, enstatite, and diopside (Figure 16d). Some of these minerals are entrapped during the crystallization of chromite, forming solitary mineral inclusions. As the chromite grows, the heterogeneity of the melt leads to disparities in the captured melt composition and the crystal/melt ratio (Figure 16e). Upon the rapid enclosure of the melt by chromite, the evolution of the melt within each inclusion proceeds independently. As the melt ascends in the upper mantle, exsolution texture develops in pyroxene, and amphibole and Na-phlogopite form through reactions between hydrous melt and liquidus minerals, ultimately resulting in microinclusions with varied mineral assemblages (Figure 16f). In these inclusions, S and other chalcophile elements from the parental magma are captured as base-metal sulfides. The dynamic process described above facilitates the participation of hydrous silicate inclusions and the massive crystallization of chromite, culminating in the eventual formation of the Sartohay podiform chromitites.

Furthermore, the mechanism of chromite concentration as ore bodies is substantiated in the melt–peridotite reaction model insufficiently [90]. This has been interpreted by the rheomorphic model [7,8,61], which provides thermodynamic reasons for the compaction or sintering mechanism of chromitites. The lack of petrographical evidence for plastic deformation in Sartohay chromitites is possibly due to the unattained high pressure required for pressure sintering [8], which may have enabled the retention of primary mineral inclusions in the chromite. However, the high mantle temperature needed for chromitite compaction as proposed by the rheomorphic model [90] is recorded in Sartohay chromitites. Hence, progress similar to sintering under a tectonic setting at high temperatures is a possible hypothesis for the coalescence of ore grains to form chromitite aggregates. As discussed above, the geochemical characteristics of inclusions and host chromite record melt mixing of different silica contents and the hydrous, Ca-rich nature of the primary melts. These features favor the crystallization and concentration of chromitite in a dynamic environment. A combination of the sintering mechanism with the melt–peridotite interaction and melt mixing model will improve the thermodynamic interpretation of the chromitite aggregates.

The tholeiitic character of the parental melts in equilibrium with Sartohay high-Al chromitites suggests that they did not originate in a typical mid-ocean ridge setting. The petrographic evidence of rutile as a crustal-derived mineral in microinclusions (Figure 7a,e) indicates that they are crustal fragments separated from the subducting slab and included in the circulating mantle [16]. Chemical compositions of primary hydrous inclusions within the high-Al chromitites (Figures 6 and 7) indicate a Na, Ti, and H₂O enrichment in the trapped melts (Figure 11b). These hydrous melts possibly may have been generated from the interaction of MORB-like melts with harzburgites in the subduction-related environment [17,62]. Moreover, the studied chromitites reveal high fO_2 values, indicating that small amounts of slab-derived high- fO_2 materials may get involved in the mantle source of the parental magma [49]. Collectively, we favor the hypothesis that the Sartohay chromitites

reflect the interaction between the MORB-like melt and harzburgite at a spreading center in a subduction-related environment.

6. Conclusions

Silicate mineral inclusions within chromite from the Sartohay podiform chromitites primarily consist of olivine, enstatite, diopside, Na-phlogopite, and amphibole. The most common cases are amphibole coexisting with enstatite and/or diopside in a single inclusion, with microtextural features suggesting their crystallization directly from the captured melt during the precipitation of chromite. Geothermobarometric estimations, based on the equilibrium of coexisting mineral phases within inclusions, indicate P-T conditions of 874–1048 °C and 8.0–21.6 kbar. The calculated fO_2 values for the chromitites ranged from 1.3 to 2.0 log units above the FMQ buffer, suggesting a more oxidized state compared to typical MORB magmas. The presence of hydrous mineral phases indicates that the parental melt of chromitites was hydrous and typically enriched in Mg, Na, Ca, and Ti. Chemical compositions of primary silicate inclusions revealed a highly magnesian parental magma with ~22.7 wt % MgO for Sartohay chromitites, while the recrystallized melt constrained by chromite exhibited a MORB-like tholeiitic affinity. The MORB-like signature of the chromite and the highly magnesian nature of the microinclusions in the Sartohay podiform chromitites suggest a reaction between Mg-rich primary melt and harzburgite, which then evolved into MORB-like melt. The mixing/mingling of the evolved melt with subsequently injected primary melt triggered massive chromite precipitation and entrapped solid minerals and melts in the upper mantle.

Supplementary Materials: The following supporting information can be downloaded at: <https://www.mdpi.com/article/10.3390/min14060530/s1>, Table S1: Representative composition (wt %) of chromite in the Sartohay podiform chromitites. Table S2: Representative composition (wt %) of chromite-host olivine inclusions from the Sartohay podiform chromitites. Table S3: Representative composition (wt %) of chromite-host enstatite inclusions from the Sartohay podiform chromitites. Table S4: Representative composition (wt %) of chromite-host diopside inclusions from the Sartohay podiform chromitites. Table S5: Representative composition (wt %) of chromite-host amphibole inclusions from the Sartohay podiform chromitites. Table S6: Representative composition (wt %) of chromite-host Na-phlogopite inclusions from the Sartohay podiform chromitites.

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