# Petrogenesis of arc-related peridotite hosted chromitite deposits in Sikhoran-Soghan mantle section, South Iran: Evidence for proto-forearc spreading to boninitic stages

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## Abstract

The Soghan-Sikhoran ophiolite in southeast Iran (Outer Zagros Ophiolite Belt) is a remnant of series of Upper Triassic–Cretaceous supra-subduction zone (SSZ) ophiolites that formed along the Zagros suture zone, which is considered as the southern border of the Paleo-Tethyan sea. These SSZ ophiolites are older than the Late Cretaceous Zagros ophiolites. Layered ultramafic cumulates including dunites, wehrlites, and pyroxenites, and a tectonized mantle section including residual lherzolites, and depleted harzburgites with foliated/discordant dunite lenses form the main part of the Soghan-Sikhoran ophiolite. Podiform chromitites are common and are typically surrounded by thin dunitic haloes. Spinels in lherzolites-harzburgites are geochemically characterized by a low Cr# (42.0 to 52.6), and plot in an abyssal peridotite field on geochemical discrimination diagrams, whereas dunites and high-Cr chromitite spinel (Cr# = 52.4 to 76.4) show geochemical affinities to boninites. Lherzolites-harzburgites have low REE contents and experienced more than 17 vol. % partial melting. The Soghan-Sikhoran ophiolite contains both high Cr# and low Cr# podiform chromitite types. Trace and REE element patterns of Soghan-Sikhoran rocks are similar to those in SSZ-peridotites. The studied ophiolites show relatively moderate to high oxygen fugacities ($fO_2$), with log units ranging from $-0.4$ to $+0.4$ for harzburgites, $+0.2$ for lherzolites, from $-0.7$ to $+2.5$ for pyroxenites, from $+0.8$ to $+2.2$ for dunites, and from $+0.6$ to $+0.8$ for chromitites. The moderate to high oxidation state of the studied upper-mantle ophiolitic complexes also suggests a boninitic source in the mantle wedge of the arc setting. The two-pyroxene thermometer yields mean equilibrium temperatures of 879 °C, 895

°C, 912 °C and 912 °C for harzburgites, lherzolites, dunites and pyroxenites, respectively. We therefore interpret that the low-Cr# chromitites and spinels in the host peridotites crystallized from tholeiitic melt generated due to proto-forearc spreading and formation of the infant arc, whereas high-Cr# spinel in dunites and high-Cr# chromitite crystallized from boninitic melts during the mature arc stage, with an increasing contribution of slab-derived fluids at high $f$O$_2$.


## 1. Introduction

It is well known that juvenile oceanic lithosphere is formed from partial melting of mantle peridotite at a mid-ocean ridge (MOR) and/or in arc-related tectonic settings. Ophiolites represent fragments of the oceanic lithosphere that have been tectonically emplaced onto a continental margin, typically during plate tectonic convergence (e.g. Palin and Santosh, 2021). Ophiolites generated in MOR setting are characterized by fertile mantle rocks with Al-rich spinels (Cr# = atomic Cr/(Cr + Al) < 0.6), and a relatively low oxygen fugacity ($f$O$_2$) (Arai, 1997a; Dilek and Thy, 1998; Arai et al., 2011; Khedr et al., 2014). In contrast, arc-related ophiolites containing refractory mantle peridotites contain high-Cr# (>0.7) spinels and have a relatively high $f$O$_2$ (e.g. Zhou et al., 1998; Dare et al., 2009; Arai et al., 2011; Khedr and Arai, 2013, 2017; Moghadam et al., 2015). Most ophiolites originally form as dismembered types within subduction zone settings, and now occur as tectonic mélanges. In such cases, mineralogical and geochemical characteristics of their mantle sections (e.g. chromitites and peridotites) are useful to decipher the tectonic settings and genesis of ophiolitic complexes.

The most recent and widely accepted interpretation for the formation of ophiolitic chromitite is a model for melt-rock reaction in the upper mantle, where chromite crystallizes due to mingling of upwelling primitive melts and mantle peridotites (Kelemen, 1990; Arai and Yurimoto, 1994; Zhou et al., 1996, 1998; Arai, 1997a,b; Zhou and Robinson, 1997; Khedr and Arai, 2016, 2017). In addition, highly refractory mantle peridotites (e.g., harzburgites) are characterized by very high Cr# (>0.7) spinel, while fertile peridotites (e.g., lherzolites) show low-Cr# (<0.3) chromian spinel (Arai, 1997a). It is thus believed that podiform chromitites should be rare and form small volumes within both highly refractory harzburgites and fertile lherzolites, whereas the less refractory harzburgites with intermediate-Cr# (0.4–0.6) spinel are the best host rocks for large chromitite pods. Recent studies have shown that

podiform chromitites within highly refractory harzburgites can be related to hydrous melts saturated with chromian spinel and olivine (Matveev and Ballhaus, 2002). Therefore, the composition of chromitites is fundamentally controlled by the chemistry of both the wall-rock and the ascending melts. For example, the orthopyroxene ($Al_2O_3$ + $Cr_2O_3$ wt. %) content of the host peridotites may control the size and composition of podiform chromitites, due to its importance in the chromian spinel's crystallization (Arai and Abe, 1995; Arai, 1997b; Khedr and Arai, 2017).

The E–W trending Esfandagheh-Faryab ophiolitic belt, southern Iran, contains around 30 ultramafic complexes, including the Soghan (the study area), Sikhuran and Abdasht complexes (Fig. 1a) (Peighambari et al., 2011; Peighambari et al., 2016; Najafzadeh and Ahmadipour, 2014). Most of these complexes are highly charged with podiform chromitites of various sizes (Najafzadeh and Ahmadipour, 2014; Moghadam and Stern, 2014; Moghadam et al., 2015; Zaeimnia et al., 2017). All of these podiform chromitites occur within dunites, which themselves are surrounded by harzburgites, such as in the Soghan area (Fig. 1b). The Soghan complex is composed of dunites, harzburgites, lherzolites and chromitites.

Previous studies have established that the Soghan ophiolitic complex is likely a part of mantle diapir, although very little work has been performed on the constituent mineral chemistry or petrogenetic models for this complex. Here, we provide new bulk-rock and mineral chemistry information for mantle peridotites and associated chromitites from the Soghan ophiolitic complex, in order to determine the source and tectonic setting of chromitites and evaluate the nature of mantle melting. Therefore, it is useful to bring together data from various localities, especially in the southern Iranian ophiolite, in order to decipher the tectonic settings of the mantle section in the Late Cretaceous and to elucidate a geodynamic model for the southern Iranian ophiolitic belt. In the current study, the nature of the mantle section of this southern Iranian ophiolite complex is discussed in terms of mineral chemistry. The parental melt composition and tectonic settings of chromitites in the lithospheric upper mantle in the southern Iranian ophiolite have also been discussed. This study also reveals good evidence for multistage melting of peridotites and evolution or maturation of an arc system from a proto-forearc spreading stage (infant arc stage) including formation of fertile lherzolites-harzburgites and low-Cr chromitites from tholeiitic melts to mature-arc stage and formation of dunites and high-Cr chromitites from boninitic melts in the upper mantle of a Cretaceous ophiolite.

## 2. Geologic setting and locations

The Haji-Abad-Esfandagheh district in South Iran contains one of the largest exposed ophiolitic belts that developed at the southeastern edge of the main Zagros fold–thrust belt (Fig. 1a), which has been considered as slice of Tethyan oceanic lithosphere scraped off during NE-directed subduction underneath Iran (e.g. Babaie et al., 2001). Two distinct groups of ophiolitic/mélange units can be identified in the Haji-Abad-Esfandagheh district. The first group occurs towards the north of the district and has Upper Triassic–Cretaceous ages (Sikhoran–Soghan ophiolitic complex; Ghasemi et al., 2002; Ahmadipour et al., 2003), whereas another group is found further south, near the Main Zagros Thrust, and has Late Cretaceous ages (the Haji-Abad ophiolites). These ophiolitic/mélange units include total probable reserves of 8 million tons of chromite ores at a grade between 30 to 53% $Cr_2O_3$.

The Sikhoran-Soghan complex is separated by a mélange zone from glaucophane schist, serpentinites, flysch type sediments (the Sanandaj–Sirjan Zone metamorphic rocks), pillow lava and pelagic limestone to the north (Fig. 1b) and occasionally shows faulted contacts with younger Haji-Abad ophiolites. It contains, from bottom to top, depleted harzburgites, a transition zone comprising porphyroclastic dunites and podiform chromitites, and grading upward into layered ultramafic–mafic cumulates (Ghasemi, 2000) that include fine-grained dunites, wehrlite and pyroxenite cumulates (Fig. 2). These rocks are intruded by gabbroic–pyroxenitic dikes (Ghasemi et al., 2001), where ultramafic–mafic cumulates are overlain by crustal isotropic gabbros (Fig. 2). The Middle Jurassic–Cretaceous diabase dikes record the most recent igneous event (Ghasemi et al., 2002; Ahmadipour et al., 2003).

The Sikhoran-Soghan complex consists of three main lithological units intruded by dikes. Toward the south, foliated dunites and associated chromitite units show evidence of high-temperature ductile deformation (Fig. 3a), while toward south, these ultramafic units change from foliated to porphyroclastic dunites (Fig. 3b). In the transition zone between harzburgites and dunites, a specific dunitic facies occurs, containing giant and kinked olivine crystals up to 40 cm long with a concordant foliation. The thick piles (2–3 km) of ultramafic and gabbroic layered cumulates include a lower and upper portion. The lower ultramafic part contains dunites, wehrlites, olivine websterites, websterites and clinopyroxenites, and the dunites include concordant stratiform chromite horizons (Fig. 3c) and nodular chromitites. The upper gabbroic part (about 2 km thick), comprises of a series of pyroxenites and lherzolites in the basal part, overlaid by a series of layered gabbroic rocks. This layered series, from bottom to top, includes: troctolites, olivine gabbros, clinopyroxenites, noritic gabbros and hypersthene

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ferrogabbros. These ferrogabbros are composed of Ti-magnetite/ulvospinel, ± ilmenite, and disseminated sulphides that are mainly pyrite and chalcopyrite. The intrusive unit includes a two-pyroxene gabbroic pluton, showing variable grain sizes up to pegmatitic types. To the south, this pluton intrudes into the Sargaz–Abshur amphibolites that were transformed into mafic hornfels The well-developed volcanic complex consists of pillow basalts and massive mafic to felsic lavas and shows faulted contacts with other units, especially serpentinized peridotites. Lava consists of two types of pillow lavas including >2000 m thick boninites, ~1000 m thick E-MORB-type (Moghadam et al., 2012b) and one type of thick sequence of calc-alkaline lavas with unclear thickness due to faulting.

In the Sikhoran mine, chromitites show layered (Fig. 3c) and massive textures (Fig. 3d), whereas chromitites in the Soghan mine show a massive and nodular texture (Fig. 3e), occurring as discontinuous pods (up to 50 m in thickness) and disseminated (Fig. 3f) ores. The nodules with up to 2 cm in diameter show slight flattening and sit in a matrix composed of serpentine minerals and carbonates (Fig. 3g). Chromitites tend to act as rigid bodies during deformation, and individual chromite grains often show a series of sub-parallel tension fractures or pull-apart cracks, which have been formed perpendicular to the direction of stress (Fig. 3b).


**3. Petrographic investigation**

3.1. Chromitites

Chromitite pods in the Sikhoran-Soghan are mainly of massive type, but disseminated, nodular and layers (Fig. 3) are also common near the contact with the adjacent dunite envelopes. Primary olivine, interstitial to chromian spinels, is still preserved, but some olivine grains have been serpentinized in the dunitic-harzburgitic layers (Fig. 3g). Massive chromitites made up of about 90% chromian spinel are composed of coarse-grained, subhedral, and fractured chromian spinel crystals (Fig. 3h), which are found either as contiguous or form fairly compact masses with interstitial silicates (i.e., olivine). Semi-massive (>30–90% chromian spinel) and disseminated (≤30% chromian spinel) chromitites have smaller crystals than those in the massive chromitites, whereas chromian spinels are mainly euhedral to subhedral crystals that are found as a cumulus phase with olivine (Fig. 3i). Olivine is the main primary silicate mineral observed in massive nodular and layers chromitites, while serpentine, chlorite, talc and carbonates are the main secondary minerals especially in the disseminated chromitites. In many cases, the chromitite samples exhibit

conjugate fractures, and late shearing led to the formation of mylonitic and cataclastic textures. In chromitite layers, chromian spinel makes up to 95% of modal while and in dunite layers, olivine has been extremely serpentinized (Fig. 3g). Chromian spinels in all chromitite pods, especially the massive types, are almost completely fresh with very small, if any, ferritchromite rims associated with spinel fractures (Fig. 3h).

3.2. Dunites and harzburgites

Dunites in the study areas are found only as envelopes of various sizes around the chromitite pods up to 3 m thick around the thickest chromitite pod. Soghan dunites are composed of olivine (50–60 vol. %), orthopyroxene (<5 vol. %), chromian spinel and serpentine minerals. Some dunites are highly serpentinized, with the dominant replacement mineral being lizardite, which forms a mesh texture. Chrysotile is less abundant than lizardite and antigorite, and occurs mostly as long cross-fiber veinlets crossing the latter ones. The dunites contain more than 2% chromian spinels, which occur as subhedral to euhedral crystals (Fig. 3i). In highly serpentinized dunites, ferritchromite and Cr-rich magnetite are abundant in spinel rims and along the cracks.

Harzburgites formed the main part of the Soghan unit, and have medium to coarse grained textures and consist of olivine (up to 85 vol. %), orthopyroxene (up to 10 vol. %), clinopyroxene (<5 vol. %), and spinel (<1 vol. %). All harzburgites show diffuse alteration with a development of a serpentine-rich matrix. Secondary fibrous amphibole at the rim of pyroxenes is locally observed. The studied harzburgites mainly show a porphyroclastic texture, characterized by millimeter-sized (2.0-2.5 mm) porphyroclasts of olivine and orthopyroxene. Olivine also occurs as small neoblasts (<0.1 mm). Both olivine and orthopyroxene show internal deformation, where they display undulose extinction or kink bands, or sometimes show distorted lamellae. Orthopyroxene occurs as large prismatic crystals (2-2.5 mm) or as smaller interstitial crystals, which are replaced occasionally by amphiboles. The original cleavage of orthopyroxene is sometimes filled with secondary minerals, such as very fine magnetite trails. Orthopyroxene is rarely preserved in the highly serpentinized rocks, being pseudomorphically replaced by lizardite/chrysotile in the form of bastite texture. Chlorite and talc are sometimes observed as alteration products of olivine and orthopyroxene. The chromian spinel in the harzburgites (1-5 vol. %) is less abundant than in dunites (5–10 vol. %), and has a subhedral to vermicular shape. It sometimes contains few silicate inclusions, mostly olivine. Small chromian spinel crystals are severely altered and

show zonation, usually displaying Cr-rich part that irregularly distributed among the Cr-poor ones. Ferritchromite and/or Cr-rich magnetite occasionally rim the spinel crystals as well as along micro-scale fracture.

## 5. Results

### 5.1 Mineral chemistry

Spinels in the Soghan podiform chromitites display a wide range in Cr# (= atomic Cr/(Cr + Al) of 0.49 to 0.76 (av. 0.64) and Mg# (= atomic Mg/(Mg + $Fe^{2+}$) of 0.47 to 0.83 (av. 0.66). These compositional criteria can be used to classify podiform chromitites into high-Al types (i.e. low-Cr chromitites) with spinel Cr# < 0.6 and high-Cr types with spinel Cr# > 0.65 (Khedr and Arai, 2016). They are low in $TiO_2$ (0.1-0.7 wt. %), and $Fe_2O_3$ (13.6-18.3 wt. %), but high in $Cr_2O_3$ (53.2-64.3 wt. %) (Table 1), similar to forearc peridotites. Most spinels in the Soghan chromitites plot close to the Cr–Al line on the Cr–Al–Fe ternary discrimination diagram in the field of forearc peridotite spinels (Fig. 4a and Table 1). All spinels in the host harzburgites and lherzolites plot in the overlapping space of abyssal peridotites (Fig. 4a; e.g. Arai 1994; Arai et al. 2011; Khedr et al., 2014) and forearc peridotites, whereas spinels in dunites and pyroxenites are similar to chromitites, and plot in the field of forearc peridotites (Fig. 4a; e.g. Ishii et al. 1992; Khedr and Arai, 2017).

The chromian spinels in the studied dunites and harzburgites also display variable Cr# with ranges of 0.52–0.70 and 0.44–0.52, and average values of 0.64 and 0.48, respectively (Table 2). In lherzolites, spinel Cr# averages 0.49 (range of 0.42–0.56) and is lower than spinel Cr# in chromitites and host dunites. Their spinel Cr# thus resembles that for spinel Cr# of harzburgites (Fig. 4b). Chromian spinel composition (Cr# and Mg#) of lherzolites, harzburgites and dunites plot within the overlapping space of both abyssal and forearc peridotites (Fig. 4b).

Olivine occurs either in the matrix or as inclusions within spinels in the podiform chromitites. The olivine inclusions and interstitial matrix in chromitites have highly forsteritic composition with Fo (= Mg# × 100) content of Fo94 to Fo95, which is somewhat higher than forsterite of dunites (Fo91 to Fo94), harzburgites (Fo90 to Fo91) and lherzolites (Fo88 to Fo91) (Table 4). The NiO contents of olivine in chromitites are also high, ranging from 0.65 to 0.98 wt. %, and from 0.38 to 0.61 wt. % of olivine in peridotites from mantle section

(Table 1), which is consistent with primary mantle olivine (e.g. Arai, 1975, 1980; Takahashi et al., 1987).

The Soghan orthopyroxene in lherzolites, harzburgites and pyroxenites is mainly enstatite with high Mg# that ranges from 0.86 to 0.91 (Table 1). Its chemical composition averages 2.3 wt. % $Al_2O_3$, 0.54 wt. % CaO and 0.53 wt. % $Cr_2O_3$. The chemical composition of the investigated orthopyroxene is relatively similar to that of the residual depleted mantle harzburgites (Loney et al., 1971; Ernst, 1978; Arai, 1980; Khedr and Arai, 2013, 2017).

The clinopyroxene in lherzolites, pyroxenites and chromitites has diopsidic composition with average 0.94 Mg#, 2.44 wt. % $Al_2O_3$ and 0.77 wt. % $Cr_2O_3$ (Table 5). Its composition is also relatively similar to that of residual mantle harzburgites.

5.2. Whole-rock chemistry

Multiple samples of peridotite (lherzolites, harzburgites and pyroxenites) from the Soghan complex were examined for major and trace-element chemistry (Table 2). Analyzed Soghan peridotite samples (n = 18) show loss on ignition (LOI) contents in the range 0.8–8.8 wt. %, indicating low-to-high degrees of serpentinization. The peridotite samples show i) low $SiO_2$ (36.7–45.1 wt. %), $Al_2O_3$ (0.5–2.8 wt. %), CaO (0.1–2.4 wt. %) and $TiO_2$ (0.01-0.17 wt. %) contents; ii) $CaO/Al_2O_3$ values in the range 0.5–1.6, iii) MgO in the range 37.9–44.8 wt. % and iv) high Cr (20,537–372,233 ppm) and Ni (1298–4747 ppm) contents. These compositions reflect the depleted nature of the investigated peridotites. The Soghan peridotites show compositions close to those of primitive mantle (PM) composition (Mg# = 0.89, Cr =3240 ppm, Ni =2090 ppm; Hart and Zindler, 1986; McDonough and Sun, 1995; Workman and Hart, 2005). The higher values of $CaO/Al_2O_3$ (up to 1.5) could reflect carbonate alteration (e.g. Moghadam et al. 2013) or low olivine/clinopyroxene ratios (e.g. Whattam and Stern, 2011; Nouri et al., 2019). In an $Al_2O_3$ vs CaO diagram (Fig. 5a), Soghan peridotites plot in the fields of abyssal peridotites (Johnson et al. 1990; Ishii et al. 1992; Pearce et al. 1992; Snow et al. 1994) and south Sandwich forearc peridotites. Based on a CaO vs $Al_2O_3$ and $Al_2O_3/SiO_2$ vs $MgO/SiO_2$ diagram (Figs. 5a, b; Jagoutz et al., 1979; Hart and Zindler 1986), Soghan peridotites plot in the field of abyssal peridotites (Niu, 2004; Barnes et al., 2014) and south Sandwich forearc peridotite along the terrestrial array, or 'melt-depletion array' (Hart and Zindler, 1986). This diagram suggests a refractory nature of the investigated peridotites ($Al_2O_3/SiO_2 < 0.05$ and $MgO/SiO_2 = 0.75–1.20$). The distribution of these peridotites along the terrestrial array excludes MgO loss due to low-temperature (<100 °C)

sea-floor weathering (e.g. Niu, 2004) or $SiO_2$ metasomatism/enrichment due to serpentinization processes (e.g. Paulick et al. 2006; Malvoisin 2015; Bhat et al., 2019). Both Soghan and South-Sandwich arc-basin peridotites plot in the overlapping field of abyssal and forearc peridotites (Figs. 5a, b).

Chondrite (CI)-normalized rare earth element (REE) (Sun and McDonough, 1989) (Fig. 6a) of the Soghan peridotites show flat MREE and HREE patterns, with $(La/Yb)_N$ (0.9) and $(Dy/Yb)_N$ ( 0.83). The Eu anomaly $(Eu/Eu^* = [Eu_N/(Sm_N*Gd_N)^{1/2}])$ is low with average values of 3.5.

Primitive mantle (PM)-normalized multi-elements diagram (Sun and McDonough, 1989) (Fig. 6b) of all peridotites are enriched in fluid-mobile elements (Sr, Pb, Li, Ba and LREE; 0.1–100 times PM) and depleted in HFSE (Hf, Th, Zr and Nb; <0.7 times PM) + U (Table 3; Fig. 6b), similar to depleted peridotites from supra-subduction zone (SSZ) setting (e.g. McCulloch and Gamble 1991; Saccani et al. 2018a)

## 6. Discussion

### 6.1. Origin and petrogenesis of the Soghan peridotites

Partial melting is believed to be important for the genesis of peridotites (Dick and Bullen, 1984; Arai, 1994; Piccardo et al. 2007; Khedr et al., 2014). This process is usually evaluated via the whole rock and mineral composition of peridotites (Dick and Bullen, 1984; 195 Uysal et al. 2007; Wu et al. 2013; Khedr et al., 2014). Whole-rock compositions of the Soghan peridotites, along with the mineral chemistry of their primary assemblages (Spl+Ol+Cpx+Opx), are used to decipher partial melting, the origin, petrogenesis and the evolution of the Soghan complex (e.g. Moghadam and Stern, 2011, 2015; Warren 2016; Khedr and Arai, 2013, 2017; Arai et al., 2018; Lian et al. 2018).

The $Al_2O_3$ and CaO vs. MgO contents of the Soghan peridotites approximately follow the same fractionation trend suggested by Niu (1997) (Fig. 7a, b) for between 15 and 25 vol. % partial melting. In such diagrams, it can be shown that the Soghan mantle rocks are significantly different to primitive mantle, and so are likely residues after high degrees of partial melting (up to 25 vol. % melt loss). In addition, the melting model (Fig. 7c) applied to the primitive mantle spinel lherzolite mineralogy ($Ol_{0.53(-0.06)}$ + $Opx_{0.27(0.28)}$ + $Cpx_{0.17(0.67)}$ + $Sp_{0.03(0.11)}$) suggested by Palme and O'Neill (2004) indicates that the Soghan peridotites formed via melting of spinel lherzolites with 20–28 vol. % melt extraction (Fig. 7c ).The

degree of partial melting of the Soghan spinel lherzolites and harzburgites can be estimated based on chemistry of chromian spinel, which indicates a 20–25 vol. % melt fraction (Fig. 8b), whereas spinel dunites and chromitites tend to have a fractionation trend that has boninitic signatures (Fig. 8b). Furthermore, the calculated partial melting degree of the Soghan spinel lherzolites and harzburgites ranges from 17.3% to 20.2%, as determined by the equation of Batanova et al. (1998) ($F\% = 9.71 \times Ln(Cr\#Spl) + 25.8$) (Table 3), which is somewhat consistent with the results based on the Cr# and $TiO_2$ contents of the chromian spinels (Fig. 8b). The relationships between spinel Cr# and olivine Fo content (Fig. 8c) show that Soghan peridotites are refractory mantle residues, which match a ~20 vol. % melt extraction from the pristine fertile MORB mantle (FMM; e.g. Arai 1994; Pearce et al. 2000; Hellebrand et al. 2001; Choi et al. 2008; Moghadam et al. 2014; Khedr et al. 2014; Saccani and Tassinari 2015; Monsef et al. 2019). This result is confirmed by the application of the melting model equation of Batanova et al. (1998) based on spinel Cr#.

The Cr# and $TiO_2$ contents of spinel and the Fo content of coexisting olivine (Fig. 8) may be used to indicate tectonic settings for formation (Ishi et al., 1990; Arai, 1994, Khedr and Arai, 2013, 2017). The Cr-spinel in Soghan peridotites is characterized by high Cr# (0.42–0.76; 0.55 on average) and Mg# in the range 0.29–0.83 (0.65 on average). Based on an Al-Cr-$Fe^{3+}$ ternary diagram (Fig. 4a) and spinel Mg# vs. spinel Cr# diagram (Fig. 4b), spinels in harzburgites and lherzolites plot in the overlapping field of abyssal peridotites (e.g. Dick and Bullen, 1984; Arai, 1994; Arai et al., 2011; Khedr et al. 2014) and forearc peridotites. Where dunites and chromite spinels lay in the space of forearc peridotites (e.g. Ishii et al. 1992; Khedr and Arai, 2013, 2017).

Olivine in Soghan peridotites has the same Fo (88.9–92.5%) and NiO (0.38–0.61 wt. %) contents as typical residual mantle peridotites (e.g. Moghadam et al. 2013; 2015) and lie above the olivine-mantle array (e.g. Takahashi et al., 1987; Moghadam et al., 2013; Szilas et al., 2015). Indeed, it plots at the intersection field of olivine in both the Eastern and Western Makran forearc peridotites (Sepidbar et al., 2020) and Oman harzburgites (Khedr et al., 2013, 2014) (Fig. 8a).

In addition, Soghan peridotites plot within the olivine–spinel mantle array (OSMA; Arai 1994) and in the overlapping fields of abyssal peridotites (after Dick and Bullen 1984) and SSZ peridotites (e.g. Pearce et al. 2000; Moghadam et al. 2014, 2015a), similar to the south Sandwich arc (Pearce et al. 2000) and Oman SSZ peridotites (Tamura and Arai, 2006) (Fig.

8c). Moreover, Soghan pyroxene compositions are comparable to those from Eastern and Western Makran clinopyroxenites and peridotites (Monsef et al. 2019; Sepidbar et al., 2020), Fannuj peridotite (Sepidbar et al., 2020) and south Sandwich forearc peridotites (Pearce et al. 2000) (Fig. 9d).

Finally, the relationship between spinel Cr# and olivine Fo content shows that Soghan peridotites are refractory residues after high partial melting around 18–20 vol. % melt for lherzolite, 20 vol. % melt for harzburgites, and 30 vol. % melt for dunites. Therefore, Soghan peridotites are a mantle residuum after melt extraction from fertile MOR mantle (FMM; e.g. Arai 1994; Pearce et al. 2000; Khedr et al., 2014; Saccani and Tassinari 2015; Monsef et al. 2019).

The residual mantle signature is suggested by high Ni (1298–4172 ppm), Cr (20,532–372,233 ppm), MgO (37.9–44.8 wt. %) and low incompatible element ($Na_2O$ < 0.30 wt. %; $TiO_2$ < 0.04 wt. %.) contents (e.g. Ohara et al. 2002), along with depletion in $TiO_2$ for both the bulk rock (0.01–0.036 wt. %) and spinels ($TiO_2$ < 0.1 wt. %) (e.g. Kelemen et al. 1995; Pearce et al. 2000; Bhat et al., 2019) relative to those of the PM (McDonough and Sun, 1995; Workman and Hart, 2005) and of the DMM (Workman and Hart, 2005). CaO (0. 1 to 3.2 wt. %), and $Al_2O_3$ (0.29 to 2.45 wt. %) values of Soghan peridotites take the mantle depletion trend (Fig. 3a; Pearce et al. 1992), revealing the degree of partial melting (Lian et al., 2018). Moreover, higher $MgO/SiO_2$ (1.0 on average) and the lower $Al_2O_3/SiO_2$ (0.03 on average) ratio values than those of the PM (McDonough and Sun, 1995; Workman and Hart, 2005) and of the DMM (Workman and Hart, 2005), are the distinctive nature of mantle residue after low- to moderate-melting extraction (Fig. 4b) (Jagoutz et al. 1979; Barnes et al. 2014; Lian et al. 2018). The $Al_2O_3$ content (1.2 wt. % on average) and CI-normalized $Yb_N$ (1.0 on average, with $Yb_N$ representative of $HREE_N$) concentrations, with respect to ophiolitic harzburgites, also confirm a moderately refractory mantle source (e.g. Niu et al. 1997; Niu 2004; Bhat et al., 2019). The REE patterns well match with mantle residua that would remain after low to moderate degree of partial melting (Godard et al. 2008; Moghadam et al. 2014, 2015a).

The depletion of Nb and La (Fig. 6b) is consistent with a SSZ geochemical signature (Hawkins, 2003; Bhat et al., 2019). The complexity of the Soghan peridotite compositions is well resolved by a Th vs. Gd/Lu diagram (Figure 9; modified after Paulick et al., 2006; Deschamps et al., 2013), where all data points plot near the DMM region (Workman and

Hart, 2005), corresponding with the average abyssal peridotites (e.g. Niu, 2004; Deschamps et al., 2013; Lian et al., 2018) and South Sandwich forearc peridotites (Pearce et al., 2000).

The composition of chromian spinel gives valuable information on the geochemical signature of the chromite parental magmas, chromitite genesis and its tectonic setting. New studies of spinel-liquid equilibrium at 1 bar (Maurel and Maurel, 1982) show the $Al_2O_3$ value of the melt from which it crystallized can be back-calculated by the equation:

$$Al_2O_3(spinel) = 0.035 \times (Al_2O_3)^{2.42}$$

Kamenetsky et al. (2001) also suggested that there is a linear relationship between the $Al_2O_3$ and $TiO_2$ contents of spinels and the melt from which they crystallized. In addition, the FeO/MgO values in the melt can also be calculated from the FeO/MgO values of chromian spinel (Table 3), using the experiential equation suggested by Maurel and Maurel (1982):

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

where, FeO and MgO are in wt. %, $Al\# = $ atomic $Al/(Cr + Al + Fe^{3+})$ and $Fe^{3+}\# = $ atomic $Fe^{3+}/(Cr + Al + Fe^{3+})$. Using the Maurel and Maurel (1982) equation (Table 3), the melts that formed the spinel of Soghan lherzolites-harzburgites and dunites-chromitites contained 13–20 wt. % and 8.6–14.4 wt. % $Al_2O_3$, respectively. They also contained a relatively varied $TiO_2$ content of 0.03–0.1 wt. % and 0.1–0.9 wt. % for lherzolites-harzburgites and dunites-chromitites, respectively, with FeO/MgO ratios varying between 0.8 and 3.7 (Table 3). The compositions of the parental melts of chromian spinel in the Soghan chromitites are similar to boninitic melt composition, which have been suggested to be responsible for the formation of high-Cr chromitites from many localities, such as Kempirsai in Kazakhstan (Melcher et al., 1997). Parent melts of Soghan chromitite have similar $Al_2O_3$ (8.9–12.4 wt. %) and relatively higher FeO/MgO ratios (1.0 on average) than Oman and Thailand chromitites with $Al_2O_3$ ~11.4 wt.% and FeO/MgO average of 0.62 (Orberger et al., 1995; Rollinson, 2008). Calculated compositions of melts from which Soghan chromitites formed are similar to high-MgO boninites (Table 3). In addition, the high spinel Cr# in Soghan chromitites confirms a low-Ca boninite parental magma (Crawford et al., 1989). Chromitite samples from the Soghan complex have low $TiO_2$ (0.17–0.91 wt. %) and $Al_2O_3$ (8.9–12.3 wt. %) contents and have similar compositions to experimentally determined melts of depleted mantle (0.1–0.5 wt. %) and boninites (9–18 wt. %; Rollinson, 2008).

**6.2. Mantle oxidation state and thermometry**

It is believed that upper mantle oxygen fugacity ($fO_2$) is highly heterogeneous, ranging from reducing (low) to oxidizing (high) conditions according to tectonic setting. Mantle $fO_2$ is controlled by recycling processes (i.e., subduction) and addition of oxidized crustal component into a moderately reduced asthenosphere (Ballhaus et al., 1990). The mantle wedge above a subducting slab is thought to be notably more oxidized than mantle in other geodynamic settings (e.g. Ballhaus et al., 1990; Parkinson and Arculus, 1999).

The $fO_2$ of the magma from which the Soghan peridotites and associated chromitites were generated can be evaluated via the coexisting olivine, orthopyroxene (opx) and spinel thermometers in peridotites (O'Neill and Wall, 1987; Ballhaus et al., 1990, 1991; Sack and Ghiosro, 1991).

Estimated values of $fO_2$ were calculated using microprobe-derived analyses of olivine-chromian spinel pairs from the chromitites, dunites and harzburgites, following the proecdures documented by Ballhaus et al. (1990, 1991) and Sack and Ghiosro (1991). The $fO_2$ are quoted as log units relative to the FMQ (fayalite–magnetite–quartz) buffer at 1.0 GPa, where log $fO_2$ FMQ refers to the deviation from the FMQ conditions (Parkinson and Pearce (1998) (Fig. 10a and Table 3). The Soghan complex shows variation in oxidation state with log $fO_2$ units ranging from −0.4 to +0.4 for harzburgites, +0.2 for lherzolites, and from −0.7 to +2 for pyroxenites, +0.8 to +2.2 for dunites and +0.6 to +0.8 for chromitites. Most of samples plot in the field of oceanic arc peridotites and overlap with forearc peridotite, and so is consistent with tectonic settings obtained by mineral and whole rock geochemistry. Most of the studied chromitites, dunites and harzburgites are relatively oxidized (Fig. 10a), with oxidation states matching forearc peridotites, comparable to the mantle wedge above a relatively oxidized subduction zone (Parkinson and Arculus, 1999; Pearce et al., 2000). It should be mentioned that podiform chromitites and their dunite envelopes have relatively higher $fO_2$ and spinel Cr# than the related lherzolites and harzburgites. The increase of both Cr-ratio and $fO_2$ of chromian spinel in harzburgites to dunites and chromitites has been interpreted due to interaction process between residual mantle harzburgites and SSZ hydrous melts enriched with $H_2O$ (i.e. highly oxidizing condition and $fO_2$) to produce dunites and chromitites owing to the hydrous nature of the SSZ melts (e.g. Ishii et al., 1992; Parkinson and Pearce, 1998; Pearce et al., 2000; Dare et al., 2009).

The two-pyroxene thermometer of Wells (1977) yields mean equilibrium temperatures of 879 °C, 895 °C, 912 °C and 912 °C for harzburgites, lherzolites, dunites and

pyroxenites, respectively. These temperatures are relatively higher than those of the temperatures obtained by spinel chemistry (Fig. 10b) and consistent with low to moderate equilibrium temperatures (755 to 878 °C) of abyssal peridotites from the Coast Range in California (Hirauchi et al., 2008).

## 6.3. Evidence for proto-forearc spreading to boninitic stages

The tectonic setting of formation of Soghan-Sikhoran chromitites and their host peridotites is inferred to have evolved from a proto-forearc scenario to a boninitic stage (Fig. 11). The SSZ affinity of ophiolitic peridotites and associated chromitites in the south Iran has led to more debate as to whether they were formed in a forearc setting (Peighambari et al., 2011; 2016) or in a back-arc setting. Peighambari et al. (2011) stated that the harzburgites and dunites around Esfandagheh were possibly derived from a sub-arc mantle (mantle wedge) above a subduction zone. It is well known that spinel compositions can be used to distinguish the tectonic setting and geodynamic environments of ultramafic rocks. For instance Cr-poor spinels ($Cr\# < 0.6$) are generated in a MOR setting (e.g., Arai, 1994; Arai and Yurimoto, 1995; Arai et al., 2011; Kamenetsky et al., 2001; Khedr et al., 2014), whereas Cr-rich spinels ($Cr\# > 0.7$) are more common in arc settings (e.g., Arai et al., 2011; Dick and Bullen, 1984; Khedr and Arai, 2013, 2016, 2017; Zhou et al., 1998).

The $Cr\#$ of spinel (0.42–0.76) from the investigated rocks resembles the spinel $Cr\#$ for arc-related peridotites/or forearc peridotites (Fig. 4) (Kamenetsky et al., 2001; Arai and Ishimaru, 2008; Arai et al., 2011; Khedr and Arai, 2013, 2016, 2017). The lherzolites and harzburgites lie close to the field of MOR peridotites generated in an extensional environment or plot at the overlapping space of MOR peridotites and forearc peridotites (Figs. 4, 5, 6 and 8). This supports the probability of the extension environment origin of Soghan lherzolites and harzburgites during proto-forearc spreading at the subduction initiation stage (Fig. 11), whereas dunites and high-Cr chromitites were generated from boninitic magmas during a mature arc stage (e.g., Khedr and Arai, 2016). The chromian spinels in high-Cr dunite envelopes and chromitites are similar in compositions to both boninite spinels and spinels that form in a SSZ setting (Figs. 4, 5, 6, 9) (Kamenetsky et al., 2001; Parkinson and Pearce, 1998; Pearce et al., 2000; Stern et al., 2004; Khedr and Arai, 2013, 2016, 2017), reflecting a sub-arc origin for Soghan-Sikhoran chromitites (e.g., Arai and Yurimoto, 1995; Khedr and Arai, 2016, 2017). The spinel compositions of Soghan peridotites and associated chromitites plot closely near the Cr–Al line of the Cr–Al–Fe ternary diagram,

and show variable Cr and low $Fe^{3+}$ contents (Fig. 4a). Thus, such high-Al and high-Cr chromitites can form in an arc-related setting during two main stages; (i) proto-forearc spreading (initial arc stage) for high-Al chromitite and (ii) sub-arc (mature arc stage) setting for high-Cr chromitite (Fig. 4a). The magmas in equilibrium with Cr-rich chromitites have boninitic affinity and highly oxidizing conditions, with $fO_2$ values almost above the FMQ buffer (Fig. 10a). All of these criteria suggest that the studied ophiolitic upper mantle peridotites are highly depleted and most probably experienced high degrees of partial melting at a SSZ setting. In contrast, Al-rich spinel composition is close to spinel composition of mid-ocean ridge or back-arc basin basalts (Zhou et al., 1998) with more reduced oxygen fugacities (Fig. 10a). Melts with a boninitic affinity were formed due to second stage melting of previously depleted mantle in a SSZ setting during a mature arc stage, where highly amount of slab-derived water is abundant in the mantle wedge beneath the fore-arc region (Bonatti and Michael, 1989; Ulmer and Trommsdorff, 1995; Sisson and Bronto, 1998; Bedard, 1999). Therefore, we suggest a possible model (Fig. 12) to explain geodynamic evolution and petrogenesis of the Soghan ophiolitic complex.

A geodynamic evolution and petrogenesis model (Fig. 11) for the Soghan ophiolitic complex suggest that lherzolites and harzburgites were re-fertilized due to percolating MORB-like melts (e.g., tholeiitic melts) beneath a proto-forearc spreading center during subduction initiation and trench rollback (Moghadam et al., 2018, 2019), whereas dunites reflect more extensive mantle melting as a result of a more hydrous environment. We suggest that MORB-like melts interacted with Soghan mantle materials during subduction initiation and the resultant mixed melts precipitated low Cr# chromitites (melt-pocket shaped domains) (Fig. 11a). This is the mantle expression of the 'subduction initiation rule', articulated for well-studied ophiolite volcanic sections by Whattam and Stern (2011), whereby subduction initiation (SI) is manifested by outpourings of tholeiitic (FAB: forearc basalt-like melt) basalts followed by boninitic melts. Upwelling of asthenospheric mantle in the forearc region during early SI first causes decompression melting of upwelling asthenospheric mantle to form MORB-like melts (FAB of Reagan et al., 2010; Ishizuka et al., 2011) (Fig. 11a). Such melts crystallized spinels in lherzolites-harzburgites and early diabasic dikes. Production of FAB melts during early SI is commonly followed by generation of arc-like or boninitic melts as slab-derived fluids eventually reach the zone of melt generation in the mantle wedge as the sinking slab descends further (Fig. 11b). These later boninitic melts formed Soghan replacive dunites, high Cr# chromitites and depleted arc related intrusive dikes.

**Conclusions**

(i) The Soghan ophiolitic complex is a complete ophiolite sequence from upper mantle peridotites to extrusive basaltic lavas. The whole rock and the mineral chemistry of the FMO peridotites describe an SSZ-mantle residuum after ca. 20–25% partial melting and melt extraction.

(ii) Two main magma types occur: enriched lherzolite-harzburgite and dunite-high-Cr chromitites. The former crystallized from MORB-like tholeiitic melts at the incipient-arc stage, and the latter was generated from boninitic melts during the mature arc stage. Both magmas formed at different mantle depths (deep origin vs. shallow origin).

(iii) The spinels in the host lherzolite-peridotites crystallized from tholeiitic melt generated due to proto-forearc spreading and formation of the infant arc, whereas high-Cr# spinel in dunites and high-Cr chromitite crystallized from boninitic melts during the mature arc stage and record an increased contribution of slab-derived fluids at high $f\mathrm{O_2}$.

**Declaration of Competing Interest**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

**Appendix A. Supplementary data**

Supplementary data to this article can be found online at https://doi.org/10.1016/j.oregeorev.2021.104256.

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**Figure captions**

Fig. 1. A) Location of the study area within the Iranian ophiolitic belts; B) Geological map of the Haji–Abad-Esfandagheh area (modified after 1/250000 Geological map of Haji–Abad, GSI) showing the distribution of the high-P rocks, Haji–Abad-Soghan-Sikhoran ophiolites, SSNZ metamorphic rocks (Modified after Moghadam et al. (2017); ages shown as stars from Moghadam et al. (2017). C) Simplified geological map of the Soghan mafic ultramafic complex. Modified after Ahmadipour et al. (2003).

Fig. 2. Schematic stratigraphic column displaying idealized internal lithologic succession in the Soghan-Sikhoran ophiolite.

Fig. 3. a) marks of high temperature ductile deformations within the foliated dunites and associated chromititic units; b) porphyroclastic dunites; c) concordant stratiform chromite horizons; and d) massive; e) nodular; and f) disseminated chromitites; g) chromitite and dunitic layers, chromian spinel makes up to 95% of modal of chromitites layers while and in dunite layers, olivine has been extremely serpentinized; h) Massive chromitites with 90% coarse-grained, subhedral, and fractured chromian spinel crystals; i) subhedral to euhedral chromite grains in disseminated chromitites.

Fig. 4. a) Spinel compositions of Soghan ultramafic rocks. Al-Cr-Fe3+ ternary (a) and Cr# vs Mg# diagram (b) are after Melluso et al. (2014). Field of abyssal peridotites (after Arai, 1994; Arai et al., 2011; Khedr and Arai, 2017) and forearc (SSZ) peridotites (after Ishii et al., 1992; Khedr and Arai, 2017) are reported in Al–Cr–Fe3+ ternary diagram.

Fig. 5. (a) CaO (wt. %) vs Al2O3 (wt. %) diagram (after Ishii et al., 1992; Moghadam et al., 2013; Moghadam and Stern, 2014; Monsef et al., 2019) for peridotites. Mantle depletion trend, fields of abyssal peridotites, and forearc peridotites are from Pearce et al. (1992). (b) Al2O3/SiO2 vs MgO/SiO2 (after Jagoutz et al., 1979; Hart and Zindler, 1986; Barnes et al., 2014; Lian et al. 2019; Nouri et al. 2019) for peridotites. Depleted MORB Mantle (DMM: Workman and Hart 2005), primitive mantle (PM; McDonough and Sun 1995), abyssal peridotite (Niu 2004), forearc peridotite (Parkinson and Pearce 1998), metasomatism/ seafloor weathering trend (Paulick et al., 2006; Bhat et al., 2019; Nouri et al., 2019), interaction with olivine-rich melts (Bhat et al., 2019), and Terrestrial Array (Jagoutz et al., 1979) are shown.

Fig. 6. Chondrite-normalized rare earth elements (REEs) diagrams (Sun and McDonough 1989) (a) and primitive mantlenormalized multi-elements pattern diagrams (Sun and McDonough 1989) (b) for the Soghan peridotites. Data source for average depleted MOR mantle (DMM, Workman and Hart, 2005), and for OIB, N-MORB, and E-MORB (Sun and McDonough 1989), Eastern-Western Makran (E–W Makran) peridotites (after Monsef et al., 2019; Sepidbar et al., 2020), Abyssal peridotites (after Niu 2004; Lian et al. 2019), South Sandwich forearc peridotites (after Pearce et al. 2000).

Fig. 7. Major oxides versus MgO (a, b) and Ni/Yb × 1000 vs. Yb variation diagrams for the bulk-rock chemistry of the Soghan. Note systematic correlations, reflecting partial melting between 15 and 25% by either isobaric batch melting (broken line) or near-fractional polybaric melting (solid line) (e.g., Niu, 1997).

Fig. 8. Mineral phases compositional variations of mantle peridotites and chromitites from Soghan ophiolite. (a) XFo (in Ol) vs NiO (in Ol) diagram (after Pag´e et al. 2008). The olivine-mantle array is after Takahashi et al. (1987); the field of olivine compositions in forearc peridotites is after Pag´e et al. (2008, and references therein); (b) TiO2 (in Spl) vs Cr# (in Spl) diagram (after Tamura and Arai 2006). (c) XFo (in Ol) vs Cr# (in Spl) diagram (after Arai 1994). The olivine-spinel mantle array (OSMA), the melting trend (with indication of % of partial melting). (d) Mg# (in Cpx) vs Al2O3 (in Cpx) diagram (after Choi et al. 2008) for clinopyroxene compositions. The fields for clinopyroxene chemistry from abyssal and SSZ peridotites are after Choi et al. (2008). Fertile MORB Mantle (FMM) and the field of abyssal (ocean ridge) peridotites are after Arai (1994). The field of oceanic suprasubduction zone (SSZ) peridotites is after Pearce et al. (2000). The fields for boninites, Abyssal and SSZ peridotites are after Tamura and Arai (2006). Data for Oman peridotites and boninites (after Tamura and Arai 2006), Oman harzburgites (after Khedr and Arai 2017), South Sandwich forearc peridotites (Pearce et al. 2000), Eastern Makran (E-Makran) lherzolites, harzburgites, and clinopyroxenites (Monsef et al. 2019) and Western Makran (W-Makran) lherzolite and peridotites (Sepidbar et al., 2020) are shown for comparison.

Fig. 9. Discriminations diagram for ultramafic and mafic rocks from Soghan ophiolite complex and other forearc peridotites. Gd/Lu vs Th (ppm) diagram (modified after Paulick et al., 2006; Deschamps et al., 2013; Nouri et al., 2019). Petrological groups and references as in Figs. 6 and 8.

Fig. 10. Plot of oxygen fugacity as (a) ΔlogfO2 vs. Cr# of chromian spinels and (b) two pyroxene thermometry for the studied chromitites and associated ultramafic rocks of the Soghan ophiolitic complex. The tectonic discrimination fields of abyssal (MORB) peridotites, fore-arc peridotites, oceanic-arc peridotites and continental peridotites are adopted from Parkinson and Pearce (1998). The solid and dashed arrows denote the trends for residual peridotite compositions interacting with MORB and supra-

subduction zone (SSZ) melts, respectively. Note that all the data are above the FMQ buffer line.

Fig. 11. Schematic model for tectono-magmatic evolution and genesis of enriched peridotites, and high Cr# chromitites and dunitic rocks in the Soghan mantle section (modified after Moghadam et al., 2015).