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Petrochemistry of High-Cr and High-Al chromitites occurrences of Dargai Complex along Indus Suture Zone, Northern Pakistan

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The present study has been carried out to determine the chromian spinel and platinum group elements (PGEs) geochemistry of podiform chromitites of Dargai Complex which provides significant information regarding the mantle-crust transition zone (MTZ) to mantle section during the emplacement and genesis of ophiolitic complexes. The mineral chemistry and platinum group elements (PGEs) were analysed using the Electron Probe Micro Analyzer (EPMA) and inductively-coupled plasma mass spectrometer (ICP-MS). Chromian spinels in the high-Cr chromitites (Cr# >60), harzburgite and mantle dunites have much higher Cr# values in the range of 78.4 to 81.9, 72.2 to 77.5 and 79.8 to 82.6, respectively. However, the chromian spinels in the high-Al chromitites (Cr# <60) and MTZ dunites have much lower Cr# values in the range of 47.5 to 52.9 and 46.7 to 49.4, respectively. Based on behaviours of PGEs, the chemical differences are reflected in these two types of chromitites. The Ir-group (IPGE) concentration is relatively higher than the Pt-group (PPGE) in the high-Cr chromitites, while high-Al chromitites have relatively greater PPGE/IPGE ratios. The melts responsible for the formation of High-Cr chromitites exhibit boninitic affinities, while that of the high-Al chromitites show mid ocean ridges basalt (MORB) signature with hydrous, oxidized and TiO₂ poor characteristics. Chromian spinel mineral chemistry and PGE geochemistry data confirm that the coexistence of both types of chromitites in the Dargai Complex may indicate a change in the composition of the parent magma from MORB to boninitic type in a proto fore-arc environment during subduction initiations. This coexistence of compositionally different mantle suites with episodic melting histories in a single ophiolitic complex can be attributed to the mechanically juxtaposition of mantle convection during recycling process within the mantle.

Introduction

In various ophiolitic complexes, chromitite bodies are randomly distributed within the mantle section and mantle-crust transition zone (MTZ). The composition of chromitites and residual mantle peridotites of ophiolites is sensitive to the nature of reactant magmas and hence can be used to track tectonic setting and melt-peridotites interaction (Dick and Bullen, 1984; Arai, 1992; Zhou and Robinson, 1997; Barnes and Roeder, 2001; Kamenetsky et al., 2001; Arai et al., 2006; Rollinson, 2008; Gonzalez-Jimenez et al., 2011, 2014). Ophiolite display a variety of tectonic settings, from low degree of partial melting at mid ocean ridge (MOR) environments to higher degrees in arc-related settings (Llanes et al., 2015; Fareeduddin and Dilek, 2015). However, the coexistence of both high-Cr (Cr# >60) and high-Al (Cr# <60) chromitites with specific composition in the same body of ophiolitic mantle has made their crystallization processes and associated tectonic changes quite debatable (Zhou and Robinson, 1997; Ballhaus, 1998; Ahmed and Arai, 2002; Morishita et al., 2010; Miura et al., 2012; Rollinson and Adetunji, 2013; Xiong et al., 2013; Arai and Miura, 2016).

It is the general consensus that high-Cr chromitites are formed from boninitic magma in a supra-subduction zone (SSZ) settings while high-Al chromitites are formed from less refractory tholeiitic magma in a back arc setting or mid ocean ridges (MOR) environments (Zhou et al., 1994; Robinson et al., 1997; Ballhaus, 1998; Ahmed and Arai, 2002; Miura et al., 2012). In some situation, the dual-mode distribution and vertical partitioning have been observed in a given complex due to different degrees of partial melting, with high-Cr chromitites occurring in deeper parts of the mantle section and high-Al chromitites occurring in the MTZ (Roberts and Neary, 1993; Ahmed and Arai, 2002; Gonzalez-Jimenez et al., 2011, 2014). The podiform chromitites have been proposed to be generated at a shallow depth of <30 km (Zhou et al., 1994; Zhou and Robinson, 1997; Bao et al., 1999; Rollinson and Adetunji, 2015). The worldwide existence of many unusual minerals including highly reduced and crustal minerals and ultra-high pressure (UHP) minerals in both high-Cr and high-Al chromitites in ophiolites indicate a range of various processes for the formation of chromitites

Table 1. Representative composition of high-Cr and high-Al chromitites of Dargai Complex through EPMA (wt%)

Sample No	Sakhakot-Qila chromitites (high-Cr chromitites)								Heru-Shah chromitites (high-Al chromitites)							
	SQ1-1	SQ1-2	SQ1-3	SQ1-4	SQ1-5	SQ1-6	SQ1-7	SQ1-8	HS2-1	HS2-2	HS2-3	HS2-4	HS2-5	HS2-6	HS2-7	HS2-8
SiO ₂	0.08	0.05	1.07	0.09	1.16	0.04	0.02	0.08	0.03	0	0.01	0	0.06	0.02	0.03	0.01
TiO ₂	0.16	0.18	0.11	0.14	0.06	0.12	0.09	0.14	0.13	0.19	0.14	0.24	0.14	0.16	0.18	0.21
Al ₂ O ₃	9.48	9.89	8.65	10.4	8.99	9.99	10.6	9.39	27.8	26.6	27.8	24.9	23.1	26.9	25.5	25.2
Cr ₂ O ₃	59.5	57.9	58.4	58.7	58.5	59.3	57.5	57.8	37.5	40.8	38.9	40.5	38.8	37.9	38.3	38.4
FeO	15.6	15.9	13.8	14.9	14.8	15.4	14.9	15.9	19.3	15.2	15.4	18.5	22.2	19.1	20.2	20.7
MnO	0.24	0.32	0.34	0.35	0.29	0.34	0.36	0.32	0.25	0.28	0.22	0.27	0.23	0.22	0.25	0.28
MgO	14.4	14.2	15.8	14.3	15.3	14.1	14.9	15.2	14.1	15.9	16.3	14.4	15.1	14.9	15.2	14.8
CaO	0	0	0.02	0.01	0	0.02	0.01	0.02	0	0	0.01	0.01	0	0	0.01	0
Na ₂ O	0	0.02	0.02	0	0.01	0	0	0.01	0.08	0	0.06	0.04	0	0	0.04	0
K ₂ O	0	0.03	0	0.03	0.01	0	0.02	0	0.04	0.01	0.04	0	0	0.02	0.02	0
NiO	0.11	0.13	0.14	0.12	0.14	0.11	0.12	0.13	0.24	0.22	0.14	0.18	0.13	0.21	0.22	0.19
Total	99.6	98.6	98.4	99	99.3	99.4	98.5	98.9	99.5	99.2	99	99	99.8	99.4	99.9	99.8
O = 32																
Si	0.021	0.013	0.276	0.023	0.297	0.01	0.005	0.021	0.007	0	0.002	0	0.014	0.005	0.007	0.002
Ti	0.031	0.035	0.021	0.027	0.012	0.023	0.017	0.027	0.023	0.034	0.025	0.044	0.025	0.029	0.032	0.038
Al	2.876	3.023	2.629	3.16	2.716	3.036	3.219	2.846	7.839	7.48	7.763	7.118	6.571	7.58	7.173	7.128
Cr	12.111	11.872	11.906	11.966	11.856	12.088	11.713	11.752	7.093	7.696	7.287	7.767	7.404	7.165	7.227	7.287
Fe	3.359	3.448	2.976	3.213	3.172	3.32	3.21	3.419	3.861	3.033	3.051	3.752	4.48	3.819	4.032	4.155
Mn	0.052	0.07	0.074	0.076	0.063	0.074	0.079	0.07	0.051	0.057	0.044	0.055	0.047	0.045	0.051	0.057
Mg	5.527	5.491	6.074	5.497	5.847	5.42	5.723	5.828	5.03	5.656	5.758	5.207	5.433	5.312	5.409	5.296
Ca	0	0	0.006	0.003	0	0.006	0.003	0.006	0	0	0.003	0.003	0	0	0.003	0
Na	0	0.01	0.01	0	0.005	0	0	0.005	0.037	0	0.028	0.019	0	0	0.019	0
K	0	0.01	0	0.01	0.003	0	0.007	0	0.012	0.003	0.012	0	0	0.006	0.006	0
Ni	0.023	0.027	0.029	0.025	0.029	0.023	0.025	0.027	0.046	0.042	0.027	0.035	0.025	0.04	0.042	0.037
Fe ²⁺	2.449	2.42	2.094	2.43	2.354	2.511	2.18	2.108	2.806	2.273	2.117	2.706	2.534	2.625	2.486	2.651
Fe ³⁺	0.909	1.028	0.881	0.783	0.819	0.809	1.03	1.312	1.056	0.759	0.935	1.047	1.946	1.194	1.545	1.504
*Cr#	80.8	79.7	81.9	79.1	81.4	79.9	78.4	80.5	47.5	50.7	48.4	52.2	53	48.6	50.2	50.5
**Mg#	69.3	69.4	74.4	69.3	71.3	68.3	72.4	73.4	64.2	71.3	73.1	65.8	68.2	66.9	68.5	66.6
Parental melt compositions																
TiO ₂	0.26	0.28	0.19	0.23	0.12	0.21	0.17	0.23	0.44	0.56	0.46	0.65	0.46	0.5	0.54	0.59
Al ₂ O ₃	10.7	10.9	10.2	11.2	10.4	10.9	11.3	10.6	15.8	15.7	15.8	15.4	15.1	15.7	15.5	15.5

Oxide values in weight percent, chemical formulae calculated on the basis of 32 oxygens, *Cr# = $100 \times \text{Cr}/(\text{Cr} + \text{Al})$, **Mg# = $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$, EPMA- electron probe micro analyzer.

(Robinson et al., 2004; Yang et al., 2007, 2011, 2014, 2105; Yamamoto et al., 2009; Yazhou et al., 2015; Xiong et al., 2016, 2017; Lian et al., 2017; Wu et al., 2017; Liu et al., 2018). Therefore, genetic relationship of both high-Cr and high-Al chromitites and their formation in a specific tectonic setting is still ambiguous.

Moreover, abundant high-Cr chromitites in Dargai Complex of ophiolitic sequence along Indus Suture Zone also suggest that these ophiolites are formed in a SSZ setting (Ahmed, 1982, 1984; Arif and Jan, 2006). In contrast, detailed field investigations of MTZ of the Dargai Complex suggest that economically significant high-Al chromitites, such as the Heru-Shah chromitites (Table 1), could be considered for the best for exploring the issues stated above. So far a limited data on these chromitites are available, which are restricted to the description of the chromian spinel (Ahmed, 1982, 1984; Arif and Jan, 2006). Based on field observations, petrographic analysis, mineral

chemistry and platinum-group elements (PGEs) geochemistry, this study outlines (1) the mineralogy and geochemistry of high-Cr and high-Al chromitites and peridotites, (2) the occurrence and distribution of PGEs within the chromitites and peridotites (harzburgite, mantle dunite and MTZ dunite), and (3) the genesis of the chromitites and possible tectonic setting of their ophiolitic Dargai Complex.

Geological Setting of Dargai Complex

The Indus Suture Zone (ISZ) marks the northern boundary of Indian-Plate with the Asian landmass. But in Pakistan the two plates (i.e., Indian and Karakoram) are separated from each other by exhumed sandwich section of Kohistan Island Arc (KIA) which is formed due to the intra-oceanic subduction of the Neo-Tethys and later on collided

Figure 1. (a) Geological map of the Dargai Complex (Ahmed, 1982, 1984) with some economical grade podiform chromite ore deposits. The small m-set map showing the location of Dargai Complex in north Pakistan (modified after Asrarullah et al., 1979); (b) Compiled geological profile showing rock units of the Dargai Complex modified after Jan et al. (1993), and existences of the Sakhakot-Qila high-Cr chromitites and the Heru-Shah high-Al Chromitites.

with the Karakoram block (Bard, 1983; Burg et al., 1998; Schaltegger et al., 2002). The Indian plate has northwestern tectonic contact with the southeastern part of Afghan block (Fig. 1a). Both the Indus Suture Zone and western collisional zone with Afghan block contains numerous mafic-ultramafic (ophiolitic) complex (Jan, 1979; Bard, 1983; Miller and Christensen, 1994; Burg et al., 1998; DiPietro et al., 2000; Arif and Jan, 2006). The mafic-ultramafic mélange that occurs along the Indus Suture is known as Dargai Complex (Fig. 1a) which is considered a south-directed thrust fault that was active in the Late Cretaceous age (Kazmi et al., 1984, 1986).

The Dargai Complex is also called as Harichand Ophiolitic Complex (Ahmed, 1982, 1984; Arif and Jan, 2006), which is mainly consisting of southern arcuate mass of ultramafic complex that is extended east to west for 30 km, with an average width of 3 to 6 km. The complex was obducted on metamorphosed pelitics, carbonaceous and calcareous rocks that ranges in age from Precambrian to Triassic (Ahmad et al., 1998). The compositional and textural variations of chromitites and associated phases have been studied in details by Ahmad (1982, 1984). The chromitites in Dargai Complex occur in two horizon based on their occurrences in a southern (lower or Sakhakot-Qila) zone with $Cr\# > 60$, and a northern (upper or Heru-Shah) zone with $Cr\# < 60$ (Fig. 1a). The chemistry of chromite are used to indicate that the high-Cr chromitites are formed in an environment similar to that of supra-subduction zone (SSZ) setting, whereas the high-Al chromitites are crystallized under mid-ocean ridges (MOR) or back arc environment (Robinson et al., 1997; Ballhaus, 1998; Zhou et al., 1998; Ahmed and Arai, 2002; Miura et al., 2012). The ultramafic rocks of Dargai Complex

(Fig. 1a) are comprised of 65% harzburgites, 23% dunites, 8% minor bodies of wehrlite, 4% chromitites and dikes of pyroxenites and serpentinites (Fig. 1a). The dikes of pyroxenite contain mostly clinopyroxene, with less commonly websterite (Asrarullah et al., 1979; Ahmed, 1988).

From the bottom to top (also from east to west), the complex is composed of clinopyroxenite, harzburgites, dunites and widespread bodies of chromitites with a 400 to 600 m thick zone of extensive dunites (Fig. 1b), which correspond to the mantle-crust transition zone (MTZ) (Asrarullah et al., 1979; Ahmed, 1982, 1984; Jan et al., 1993). There is a sharp contact between the MTZ dunites and harzburgites. The crustal section of the Dargai Complex generally occurs as 1 to 2 km thick discontinuous outcrops and is mainly represented by gabbro and pyroxenite along its western margins. The lower most part of mafic crustal unit is the top section of the interlayered wehrlites-dunites gabbro with mantle-crust transition zone (Fig. 1b).

In Dargai Complex, the Sakhakot-Qila chromitites hosted in the lenticular bodies of harzburgites and dunite are of mantle origin (Fig. 1b). The large ore bodies of these chromitites are formed as lenses, pods and seams morphologies, with massive, semi-massive and disseminated textures and concordant and folded structures due to the shearing and deformation caused by Main Mantle Thrust (MMT) of Indus Suture Zone (Fig. 2a) (Ahmed, 1982, 1984; Arif and Jan, 2006). The chromitite ore bodies range in thickness from 1 to 10 m, and are usually enclosed by the dunites envelope (Fig. 2b), which are several cm to 2.5 m thick. Locally, the host harzburgites have sharp contact with some of the chromitites. The Heru-Shah chromitites are small concordant bodies, which are hosted in the dunites of mantle-crust transition

Figure 2. Field photographs showing the relationship between the host rocks and ores in Dargai Complex. (a) Distribution of podiform chromite in the Sakhakot-Qila high-Cr chromitites which are strongly faulted and deformed due to thrusting along the Indus Suture Zone; (b) Massive chromitite body with dunites wall-rock; (c) Gradual transition between Heru-Shah high-Al chromitites and dunites; (d) Various tectonic structures present at the chromite hand specimen of MTZ dunites.

zone (MTZ) (Fig. 1b). They are usually fractured and generally from lenticular to rod like bodies (Fig. 2c-d) with nodular and massive textures (Ahmed, 1982, 1984; Jan et al., 1993; Arif and Jan, 2006).

Petrographic Features of Dargai Peridotites and Chromitites

Harzburgite

The harzburgite of Dargai Complex mainly consists of olivine (74 to 84 vol%), orthopyroxene (14 to 24 vol%) with minor clinopyroxene

and chromian spinel (Fig. 3a). The textures are mostly porphyroblastic and are characterized by porphyroblasts of olivine and orthopyroxene with plastic deformation features such as, recrystallization, rotation and kink bands. Chromian spinel occurs as inclusions in orthopyroxene as well as in small, euhedral crystals of olivine or along the contacts between these two minerals (Fig. 3a). The orthopyroxene grains show crystal flexing, twin gliding and undulatory extension. Some of the orthopyroxene grains host clinopyroxene exsolution lamellae and also exhibit deformation features along their boundaries (Fig. 3b). There are irregular interstitial grains of clinopyroxene which occur between orthopyroxene and olivine. In

Figure 3. Photomicrographs of representative harzburgites, dunites and chromitites of the Dargai Complex. (a) the fine-grained olivine have contacts with coarse-grained olivines and orthopyroxene (Opx) in harzburgite, (b) orthopyroxene grains having exsolution lamellae of clinopyroxene (Cpx) in the harzburgite, (c) mantle dunites, (d) mantle-crust transition zone (MTZ) dunite, (e) mono or polyminerallc inclusions in chromian spinels of Sakhakot-Qila chromitites, (f) chromian spinels of the Heru-Shah chromitites showing alteration to ferrit-chromitites at their boundaries and cracks, (g) polyminerallc inclusions occur in the chromian spinels of Heru-Shah chromitites which include amphibole (Amp), Opx, Cpx, and Ni-Fe-Pt alloy. Olivine (Ol), clinopyroxene (Cpx), orthopyroxene (Opx), chromian spinel (Cr-spl), serpentine (Srp), chlorite (Chl), ferrit chromite (Ferrit-Chr), amphibole (Amp), chalcopyrite (Ccp).

harzburgite, the residual chromian-spinel grains are brown to reddish-brown in color and smaller in size which mostly occurs as interstitial grains along the triple junctions of olivine grains. The harzburgites are mostly fresh with only sparse cracks occupied by chlorite and serpentine. Exsolution lamellae and kink deformation characteristics in harzburgites show that these rocks were originally shaped at mantle pressures and temperatures.

Mantle Dunite and MTZ Dunite

Mechanical properties such as crystal growth, deformation and qualitative description of texture provide significant information about the geotectonic nature of rocks (Esmailzadeh et al., 2017; Ahmadi et al., 2018). Mantle dunites in the harzburgites occur as lenses which extensively consist of granular olivine (93 to 97 vol%) with minor clinopyroxene and chromian spinel (Fig. 3c). The olivine crystals are 0.2 to 1 mm in diameter and show deformational twinning and wavy extinction. Some of the large crystals have been intensively fractured (Fig. 3c). Clinopyroxene occurs as small (0.2 to 0.4 mm) euhedral crystals. The chromium spinel grains (0.1 to 0.4 mm) are generally distributed in the olivine, but their local concentration is high enough to form disseminated-chromitites. Chromian is an important element in helping to determine the conditions that rocks, soil and ophiolite form (Kyere et al., 2018). Chromian spinel in the studied dunites hosting high-Cr chromitites are largely anhedral in shape (Fig. 3c), while chromian spinel in dunites, enclosing high-Al chromitites, are predominantly euhedral or subhedral in shape (Fig. 3d). This is in consistency with the observations of Matsumoto and Arai (2001).

The MTZ dunites have a granular texture (Fig. 3d) similar to that of mantle dunites but it is characterized by weak bands or layers which have been obliterated by massive hydrothermal alteration. Chlorite and serpentine are formed along the cracks and grain borders by alteration, eventually leading to small “island” of olivine relicts (Fig. 3d). The chromian spinel grains are weakly directed parallel to bandings in the MTZ dunites (Fig. 2d).

Chromitites

Chromitites is chromite deposit, which is composed of chromite and silicate minerals, such as olivine, pyroxene. Chromium element of chromite and other heavy metals have great importance in industrial manufacturing and also in our daily life (Arasaretnam and Kirudchayini, 2019; Nkansah et al., 2018). The textures of chromitites of Sakhakot-Qila Dargai Complex are primarily massive, but merge into disseminated types when reaches to the mantle dunite contacts. Massive chromitites usually contain >94 vol% chromian spinel grains, which are coarse grained (1 to 5 mm) and are typically anhedral to subhedral in shape. The olivine, chlorite and serpentine are the inter-granular minerals with small amount of base metal sulfides. The disseminated chromitites with ≤ 40 vol% chromian spinel grains occur at the edges of massive chromitites which form irregular patches of few centimeters to a few meters across. Disseminated chromitites consist of variable amounts of chromian spinel and olivine. The size of spinel in these rocks is smaller (0.5-1 mm in diameter) than those of massive chromitites. Chromian spinel grains in both chromitites consist of inclusions of silicates, platinum-group minerals (PGM) and base metal

sulfide (BMS). Clinopyroxene and olivine are the most common silicates inclusions which are present in the chromian spinel as monomineralic grains with rounded-subhedral shape, while they can also be associated as poly-mineralic and with base metal sulfide (Fig. 3e). The base metal sulfides are mainly chalcopyrite, millerite and pentlandite. The chromian spinel contains anhedral inclusions of Os-Ir alloys along with the BMS (Fig. 3e).

The chromitites of Heru-Shah Dargai Complex are generally characterized by semi-massive textures, with predominance of disseminated to nodular textures near the contact with the host mantle-crust transition zone (MTZ) dunite. As compared to Sakhakot-Qila chromitites these varieties are considerably more altered and occur at deeper levels in the mantle section. The chromian spinel grains are altered to ferrite chromite along the margins and cracks (Fig. 3f), whereas the olivine present along the boundaries of chromian spinel grains are replaced by the serpentine with chlorite bands along the rims. The chromian spinel crystals have abundant solid poly-mineralic inclusions ranging in size from 5 to 100 μm in size. These inclusions mainly consist of amphibole, clinopyroxene and orthopyroxene with minor olivine, base metal sulfide (BMS) and platinum-group minerals (PGM). The amphibole and orthopyroxene inclusions are associated with some nickel-iron-platinum (Ni-Fe-Pt) alloys (Fig. 3f-g).

Analytical Methods

A total number of 30 fresh samples were collected which include 8 chromitites from Sakhakot-Qila, 8 peridotites from Heru-Shah, 5 harzburgites, 5 mantle dunites and 4 MTZ dunites from the Dargai Complex. These samples were collected for mineral geochemistry and whole rock platinum group elements (PGE) composition. The chromian spinel and associated silicate minerals were analyzed for major elements, after the preparation of polished thin sections, by JEOL JXA-8100 Electron Probe Micro Analyzer (EPMA) equipped with four wavelength-dispersive spectrometers (WDS) in the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Wuhan). The analysing parameters of EPMA were accelerating voltage of 15 kV, beam current of 20 nA and a beam diameter of 1 μm size. Natural and synthetic standards were used to check calibration precision of the instrument. By assuming the ideal chromian spinel stoichiometry of $A^{2+}B^{3+}_2O_4$ the amount of Fe^{3+} in chromite were calculated and the results of chromian spinel and silicates representative analysis are given in Tables 1 to 5.

The PGE concentration of the chromitites and peridotites were determined at the Institute of Geochemistry, Chinese Academy of Sciences by adopting nickel-sulfide fire assay and Te-coprecipitation method followed by inductively-coupled plasma mass spectrometry (ICP-MS) measurements (Jackson et al., 1990; Sun et al., 1993). A standard sample (GBW079290) was used as a PGE reference to clarify the consistency of analytical results. The accuracies of the analysis were enhanced than 5% for Ir, Rh, and Pd, and 10% for the Os, Ru, Pt elements. The detection limits for PGEs and results of representative analyses are given in Table 6. For detail procedures see the technique of Liang et al. (2000) and Sun and Sun (2005) and summary of the research methodology displayed in Figure 4.

Results

Mineral Chemistry

The probe analyses were carried out on chromian spinels and silicates after preparation of polished thin sections of chromitites and enclosed dunite and host harzburgite. The chemical composition of Sakhakot-Qila chromian spinel is particularly different as compared to that of Heru-Shah. The Cr_2O_3 contents of Sakhakot-Qila chromian spinels are in the range of 57.5–59.5 wt% and Al_2O_3 from 8.65–10.6 wt%, whereas in the Heru-Shah chromitites Cr_2O_3 contents vary from 37.5 to 40.8 wt% and Al_2O_3 from 23.1 to 27.8 wt%. The Cr# values are ranging from 78.4 to 81.9 and 47.5 to 52.9 in chromian spinel and chromitites ore of Sakhakot-Qila and Heru-Shah, respectively (Fig. 5a; Table 1). Thus these chromitites are classified as High-Cr metallurgical and high-Al refractory chromitites. The high-Cr types spinels have 69.3 to 74.4 Mg# and 0.06 to 0.18

Figure 4. Flow chart for the research methodology.

Figure 5. (a) Chemical composition of chromian spinel from chromitites and peridotites plotted on Al_2O_3 versus Cr_2O_3 diagram, (b) TiO_2 versus Cr_2O_3 diagram, (c) Cr# versus Mg# diagram, (d) Cr# versus TiO_2 diagram. Fields for the stratiform and podiform chromite are taken from Uysal et al. (2009b), while boninitic and mid ocean ridges basalt (MORB) fields are from Dick and Bullen (1984) and Arai (1992).

wt% TiO₂ contents, whereas the Mg# values (64.2-73.1) are slightly lower in high-Al chromitites with TiO₂ contents 0.13 to 0.24 wt% (Fig. 5b-c; Table 1). The Cr# vs. Mg# values in chromian spinel and Cr# vs. TiO₂ values in the chromitites (high-Cr chromitites) of the Sakhakot-Qila plot in the boninitic field, while the Heru-Shah chromitites (high-Al chromitites) partially fall in the field of MORB (Fig. 5c-d).

The mantle dunite chromian spinels have similar or somewhat higher Cr# (79.8-82.6) and TiO₂ (0.12 to 0.17 wt%) as compared to that of the high-Cr chromitites (Fig. 5c-d; Table 2). The MTZ dunites have much lower Cr# values in the chromian spinels (46.7 to 49.4) but higher Mg# values (59.4 to 67.3) than those of mantle dunite chromian spinels (Fig. 5c). The Cr# and Mg# contents of chromian spinels of MTZ dunites are comparable or slightly lower than those in the high-Al chromitites, but are having slightly higher contents of TiO₂ (0.17 to 0.22 wt%) than those of mostly occurring in the high-Al chromitites and mantle dunites (Fig. 5d).

The Cr# values are ranging from 72.2 to 77.5 in the chromian spinels of harzburgites which are lower than those of high-Cr chromitites and mantle dunites. The increasing trend of Cr# values is also accom-

panied by an increase in the TiO₂ concentration from the host harzburgite to mantle dunite and then to high-Cr chromitites; which is <0.08 wt% in chromian spinel of harzburgite, 0.12 to 0.17 wt% in chromian spinel of mantle dunite and 0.06 to 0.18 wt% in those of high-Cr chromitites (Fig. 5d).

Forsterite (Fo) contents of olivines in the harzburgites are ranging from 88.6 to 89.6 with NiO concentration between 0.24 to 0.35 wt%, whereas the Fo contents in the mantle dunites are slightly higher with a range of 90.0 to 91.9 and NiO contents are ranging from 0.37 to 0.44 wt%. The Fo values of MTZ dunite olivines are between 88.8 to 89.8 with a range of 0.36 to 0.44 wt% NiO contents. High-Cr chromitites and high-Al chromitites olivine grains have high Fo contents ranging from 94.2 to 95.1 and 88.4 to 91.8, respectively (Table 3).

All of the studied clinopyroxenes are diopside (Fig. 6a), but they have relatively different contents of Cr₂O₃, TiO₂ and MgO (Fig. 6b-c). In harzburgite, the clinopyroxenes have greater amount of Cr₂O₃ which is varying from 0.39 to 0.58 wt% and Al₂O₃ from 0.74 to 1.16 wt% while the mantle dunites clinopyroxenes have Cr₂O₃ and Al₂O₃ contents in the range of 0.16 to 0.18 wt% and 0.16 to 0.17 wt%, respectively. The high-Cr chromitites interstitial clinopyroxenes contain lowest Al₂O₃

Table 2. Representative composition of chromian spinel in Dargai Complex peridotites through EPMA (wt%)

	Harzburgite					Mantle Dunite					MTZ Dunite				
Sample No	SQ4-1	SQ4-2	SQ4-3	SQ4-4	SQ4-5	SQ5-1	SQ5-2	SQ5-3	SQ5-4	SQ5-5	HS4-1	HS4-2	HS4-3	HS44	HS4-5
SiO ₂	1.19	1.27	1.25	1.08	1.17	0.03	0.04	0.01	0	0	0.01	0	0.03	0.02	0
TiO ₂	0.08	0.05	0.07	0.05	0.06	0.12	0.15	0.14	0.17	0.16	0.17	0.19	0.22	0.18	0.19
Al ₂ O ₃	12.8	13.2	10.6	13.9	11.7	9.56	8.22	9.06	8.87	8.96	27.5	25.6	24.1	25.9	26.1
Cr ₂ O ₃	54.8	53.7	54.5	53.7	53.9	56.2	58.1	54.4	58.5	56.9	35.9	36.5	35.1	36.9	36.3
FeO	21.9	22.5	21.7	20.8	22.3	25.6	24.9	24.9	22.1	24.5	20.5	23.9	25.2	23.4	23.8
MnO	0.41	0.36	0.39	0.41	0.38	0.35	0.37	0.42	0.39	0.41	0.26	0.27	0.25	0.27	0.24
MgO	8.21	8.78	9.99	9.58	9.91	8.01	8.33	9.13	9.05	8.37	14.8	13.1	14.9	13.2	13.1
CaO	0	0	0	0	0	0.01	0	0.02	0.02	0.02	0.01	0	0.01	0	0
Na ₂ O	0.01	0.04	0.02	0	0.01	0.04	0	0.06	0.04	0.03	0.07	0	0.05	0	0
K ₂ O	0	0	0	0.01	0.01	0.01	0	0.01	0.01	0	0.01	0	0.01	0	0.01
NiO	0.09	0.02	0.08	0.11	0.09	0.05	0.05	0.08	0.07	0.06	0.19	0.16	0.17	0.19	0.21
Total	99.5	99.9	98.6	99.6	99.5	99.9	100.2	98.2	99.2	99.4	99.42	99.7	100	100.1	99.9
O = 32															
Si	0.315	0.333	0.332	0.282	0.307	0.008	0.011	0.003	0	0	0.002	0	0.007	0.005	0
Ti	0.016	0.01	0.014	0.01	0.012	0.024	0.03	0.028	0.034	0.032	0.03	0.035	0.04	0.033	0.034
Al	3.994	4.075	3.318	4.27	3.618	3.018	2.602	2.886	2.809	2.844	7.727	7.311	6.81	7.364	7.426
Cr	11.47	11.121	11.445	11.066	11.181	11.901	12.336	11.623	12.43	12.116	6.767	6.993	6.653	7.038	6.928
Fe	4.848	4.928	4.82	4.533	4.893	5.734	5.592	5.627	4.967	5.518	4.087	4.843	5.052	4.721	4.804
Mn	0.092	0.08	0.088	0.091	0.084	0.079	0.084	0.096	0.089	0.094	0.052	0.055	0.051	0.055	0.049
Mg	3.241	3.429	3.956	3.723	3.877	3.199	3.335	3.679	3.626	3.361	5.26	4.733	5.326	4.748	4.715
Ca	0	0	0	0	0	0.003	0	0.006	0.006	0.006	0.003	0	0.003	0	0
Na	0.005	0.02	0.01	0	0.005	0.021	0	0.031	0.021	0.016	0.032	0	0.023	0	0
K	0	0	0	0.003	0.003	0.003	0	0.003	0.003	0	0.003	0	0.003	0	0.003
Ni	0.019	0.004	0.017	0.023	0.019	0.011	0.011	0.017	0.015	0.013	0.036	0.031	0.033	0.037	0.041
Fe ²⁺	4.969	4.789	4.265	4.448	4.322	4.692	4.611	4.163	4.25	4.528	2.61	3.215	2.582	3.198	3.224
Fe ³⁺	0.121	0.139	0.555	0.085	0.571	1.041	0.981	1.464	0.716	0.99	1.477	1.627	2.47	1.523	1.581
*Cr#	74.2	73.2	77.5	72.2	75.6	79.8	82.6	80.1	81.6	80.9	46.7	48.9	49.4	48.9	48.3
**Mg#	39.5	41.7	48.1	45.6	47.3	40.5	41.9	46.9	46	42.6	66.8	59.5	67.4	59.8	59.4

Oxide values in weight percent, chemical formulae calculated on the basis of 32 oxygens, *Cr# = $100 \times \text{Cr}/(\text{Cr} + \text{Al})$, **Mg# = $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$.

Table 3. Representative composition of olivine in Dargai Complex peridotites and chromitites through EPMA (wt%)

Lithology	S. No	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	NiO	Total	Fo
Harzburgite	SQ4-6	41.4	0	0.01	0.02	8.47	0.1	49.2	0	0	0	0.24	99.4	89.6
	SQ4-7	41.5	0.02	0	0.04	8.53	0.16	48.9	0.02	0	0.01	0.35	99.5	89.2
	SQ4-8	41.8	0.03	0.02	0.05	8.33	0.13	48.8	0	0	0	0.32	99.5	88.6
	SQ4-9	41.6	0.01	0.02	0.05	8.92	0.11	48.8	0.01	0.01	0	0.24	99.8	88.7
	SQ4-10	41.8	0.04	0	0.01	8.59	0.1	49.1	0.03	0	0	0.3	100	88.9
Mantle Dunite	SQ5-6	40.6	0.01	0	0.03	8.41	0.13	49.9	0.08	0	0.01	0.44	99.6	91.9
	SQ5-7	40.9	0	0	0.19	8.39	0.08	50.1	0.05	0.02	0	0.39	100.1	91.7
	SQ5-8	41.1	0	0.01	0.11	8.41	0.11	49.8	0	0.01	0.01	0.37	99.9	91
	SQ5-9	41.1	0.02	0.01	0.09	8.43	0.08	49.9	0.02	0	0	0.42	100.1	91
	SQ5-10	40.8	0	0.01	0.04	8.48	0.09	50.2	0	0.01	0	0.43	100.1	91.9
MTZ Dunite	HS4-6	40.6	0.01	0.01	0.01	10.3	0.18	48.4	0.05	0	0.02	0.43	100	89.5
	HS4-7	40.4	0	0.02	0	10.4	0.11	47.7	0.13	0	0	0.36	99.1	88.7
	HS4-8	40.5	0.03	0	0.03	10.4	0.13	48.5	0.11	0.03	0.01	0.44	100.2	89.8
	HS4-9	40.7	0.01	0.01	0.02	10.5	0.13	48.4	0.06	0.01	0	0.39	100.2	89.2
	HS4-10	40.6	0.02	0	0.02	10.3	0.14	48.6	0.1	0.01	0	0.44	100.2	89.7
MTZ Chromitites	HS2-9	41.9	0.01	0.02	0.02	7.31	0.14	49.1	0.05	0.02	0	0.55	99.1	89.4
	HS2-10	41.6	0.01	0.02	0.05	9.91	0.08	48.7	0.02	0	0	0.45	100.8	88.4
	HS2-11	40.8	0	0.01	0.03	8.54	0.11	49.9	0.04	0.01	0.02	0.52	100	91.7
	HS2-12	40.3	0.02	0	0.05	9.22	0.14	49.5	0.05	0.04	0.02	0.55	99.9	91.8
	HS2-13	41.7	0	0.01	0.04	9.66	0.09	48.8	0.02	0	0	0.47	100.8	88.5
Mantle Chromitites	SQ1-9	42.2	0	0.01	0.13	4.39	0.05	52.7	0.1	0	0	0.4	100	94.2
	SQ1-10	41.7	0	0	0.28	4.33	0.11	52.4	0.12	0.01	0.02	0.47	99.4	94.8
	SQ1-11	41.9	0.01	0.01	0.07	4.82	0.07	52.5	0.1	0	0.01	0.42	99.9	94.3
	SQ1-12	41.7	0.01	0	0.04	4.51	0.09	52.7	0.12	0.02	0.02	0.45	99.7	95.1
	SQ1-13	41.8	0.01	0.01	0.18	4.89	0.06	52.3	0.11	0	0	0.48	99.8	94.2

Oxide values in weight percent, Fo (forsterite) or $Mg\# = 100 \times Mg/(Mg + Fe^{2+})$.

contents (0.25 to 0.64 wt%) and Cr₂O₃ (0.11 to 0.49 wt%) with highest Mg# values (97.6 to 99.3). Chromian spinel clinopyroxenes inclusions in high-Cr chromitites have the maximum Cr₂O₃ contents ranging from 1.13 to 1.22 wt%. The clinopyroxene inclusions in spinels of high-Al chromitites are compared with those of high-Cr chromitites. These are having low Mg# (96.5-97.0) and Cr₂O₃ (0.41-0.44 wt%) but have high TiO₂ contents (0.08 to 0.09 wt%). The Al₂O₃ contents have positive correlation with Cr₂O₃ in all clinopyroxenes except for inclusions in the high-Cr chromitites chromian spinel (Fig. 6c).

Based on the composition of major elements all of orthopyroxenes fall in enstatite range (Fig. 6a). Those in the harzburgite and mantle dunite have almost comparable compositional characteristics of Al₂O₃ (0.16 to 0.59 wt%), Cr₂O₃ (0.25 to 0.33 wt%), and Mg#s (90.4 to 91.3) as shown in Table 4, but those occurring as inclusions in high-Al spinel have higher Cr₂O₃ ranging from 0.56 to 0.58 wt% and TiO₂ contents are about 0.05 wt% as compared to harzburgite and mantle dunite.

The amphiboles are mainly edenite or pargasite with high Mg#s ranging from 91.7 to 93.3 (Fig. 6d). The TiO₂ contents (0.29 to 0.64 wt%) in amphibole exhibit a negative correlation with SiO₂ contents (44.5 to 47.2 wt%) (Table 5; Fig. 6e).

Geochemistry of Platinum Group Elements (PGEs)

The total PGE concentrations in the harzburgite samples are rang-

ing from 17.3 to 23.9 ppb (Table 6), which are PGE depleted as compared to chondrite and there is a relatively flat pattern in about 0.01×chondrite (Fig. 7a). Mantle dunites samples have total PGE contents between 7.83 to 13.3 ppb which are lower as compared to that harzburgite samples having Pt and Pd negative anomalies in the normalized PGE mood of chondrite values. The total PGE concentrations of the samples of MTZ dunite are lower (6.74 to 10.2 ppb) as compared to the total PGE concentrations of the mantle dunite which show flat PGE moods relative to the negative anomalies of Ir and Pt portion (Fig. 7a-b).

The concentrations of PGEs (112-258 ppb) of the high-Cr chromitites occurring in harzburgite and those of high-Al chromitites (39.1 to 200 ppb) found in the MTZ dunites are much higher than their host rocks, and also that of the primitive mantle (Table 6; Fig. 7a-b). The chondrite-normalized PGE patterns are steep in high-Cr chromitites which are enriched in the more refractory IPGE (iridium-group platinum group elements i.e., Os, Ir and Ru), comparative to the PPGE (palladium-group platinum group elements i.e., Rh, Pt and Pd) with negative anomalies of Ir and Pt. The high-Cr chromitites PPGE/IPGE proportions are 0.04 to 0.40 as shown in Table 6. These high-Cr chromitites have lower Pd/Os and Pd/Ir values as compared to primitive mantle (Fig. 7c). Chondrite-normalized PGE pattern displaying positively sloping characteristics in high-Al chromitites (Fig. 7b) with relatively high ratios of PPGE/IPGE range from 0.60 to 6.24. These high-Al

Figure 6. (a) Variation in Ca, Mg# and Fe in the high-Cr and high-Al chromitites of Dargai Complex compared with host peridotites. All the clinopyroxenes are diopside and orthopyroxenes fall in enstatite range, (b) high-Cr and high-Al chromitites and peridotites TiO₂ versus Mg#, (c) Cr₂O₃ (wt%) versus Al₂O₃ (wt%), (d) Mg# versus Si (atoms per formula unit) of high-Al chromitites amphiboles, (e) TiO₂ (wt%) versus SiO₂ (wt%).

chromitites also show increase in Pd/Os values with increasing in the Pd/Ir values, which plots on a different trend than the chromitites of high-Cr varieties (Fig. 7c).

Discussion

Partial Melting Processes and Products

Experimental work has demonstrated that the chemistry of chromite and olivine is strongly dependent on the composition of their parental

magma (Dick and Bullen, 1984; Arai, 1992; Kamenetsky et al., 2001; Irvine, 2009; Page and Barnes, 2009; Zhou et al., 2014). It is commonly acknowledged that the chromitites parental melts are the product of melt-melt interactions or melt-rock reactions previously depleted in mantle sources in supra subduction zone environment (SSZ) (Zhou et al., 1994, 1998; Zhou and Robinson, 1997; Melcher et al., 1999; Kamenetsky et al., 2001; Uysal et al., 2005, 2009b). However, there is still a discussion on the genesis of different geochemical characteristics of chromitites with different melts in similar tectonic environments (e.g., high-Cr and high-Ti, high-Cr and low-Ti, and high-Al and high-Ti) (Rollinson, 2008; Arai et al., 2011; González-Jiménez et al., 2011; Pal,

Table 4. Representative composition of pyroxenes in peridotites and chromitites of Dargai Complex through EPMA (wt%)

Lithology	Harzburgite				Mantle Dunite				High-Al Chromitites				High-Cr Chromitites			High-Cr Chromitites		
	matrix minerals				matrix minerals				inclusions in chromian spinel				matrix minerals			inclusions in chromian spinel		
Sample	*Cpx	Cpx	**Opx	Opx	Cpx	Cpx	Opx	Opx	Opx	Opx	Cpx	Cpx	Cpx	Cpx	Cpx	Cpx	Cpx	Cpx
SiO ₂	54.3	54.1	56.8	57.2	53.9	54.6	56.2	56.1	57.2	56.8	54.1	54.5	55.2	54.9	54.5	54.8	54.7	54.6
TiO ₂	0.01	0.01	0.02	0.01	0	0.01	0	0.01	0.02	0.05	0.09	0.08	0.05	0.05	0.04	0.01	0.02	0.04
Al ₂ O ₃	0.74	1.16	0.59	0.55	0.16	0.17	0.49	0.16	0.69	0.71	0.27	0.31	0.64	0.25	0.33	0.31	0.33	0.34
Cr ₂ O ₃	0.39	0.58	0.33	0.27	0.16	0.18	0.26	0.25	0.56	0.58	0.44	0.41	0.21	0.49	0.11	1.16	1.13	1.22
FeO	1.58	1.65	6.13	6.16	1.97	1.91	6.66	6.62	6.69	6.73	1.42	1.13	0.73	0.77	0.99	0.96	1.04	0.97
MnO	0.09	0.05	0.14	0.15	0.08	0.07	0.13	0.12	0.13	0.12	0.06	0.06	0.04	0.02	0.04	0.07	0.06	0.07
MgO	17.2	17.4	35.2	34.9	17.3	17.4	34.3	34.5	34.3	34.4	17.5	17.4	17.7	17.8	18.1	18.6	18.7	18.5
CaO	24.7	24.9	0.68	0.63	24.5	24.3	0.33	0.35	0.41	0.47	25.2	24.9	24.6	25.2	25.3	23.3	23.4	23.4
Na ₂ O	0.08	0.07	0.02	0.01	0.08	0.09	0.02	0.02	0.02	0.01	0.05	0.01	0.33	0.19	0.05	0.15	0.19	0.15
K ₂ O	0.01	0.01	0	0	0.02	0.02	0.01	0.01	0.01	0	0.01	0.01	0.01	0	0.01	0.01	0	0.01
NiO	0.06	0.05	0.08	0.06	0.06	0.05	0.14	0.13	0.08	0.11	0.05	0.04	0.07	0.08	0.08	0.09	0.11	0.07
Total	99.2	100	100	99.9	98.2	98.8	98.5	98.3	100	100	99.2	98.9	99.6	99.8	99.6	99.5	99.7	99.4
O = 6																		
Si	1.990	1.965	1.954	1.971	1.994	2.008	1.968	1.969	1.976	1.964	1.980	2.002	2.004	1.993	1.981	1.994	1.985	1.989
Ti	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.001	0.002	0.002	0.001	0.001	0.001	0.000	0.001	0.001
Al	0.032	0.050	0.024	0.022	0.007	0.007	0.020	0.007	0.028	0.029	0.012	0.013	0.027	0.011	0.014	0.013	0.014	0.015
Cr	0.011	0.017	0.009	0.007	0.005	0.005	0.007	0.007	0.015	0.016	0.013	0.012	0.006	0.014	0.003	0.033	0.032	0.035
Fe ³⁺	0.000	0.008	0.059	0.028	0.008	0.000	0.038	0.051	0.006	0.025	0.014	0.000	0.000	0.000	0.023	0.000	0.000	0.000
Mn	0.003	0.002	0.004	0.004	0.003	0.002	0.004	0.004	0.004	0.004	0.002	0.002	0.001	0.001	0.001	0.002	0.002	0.002
Mg	0.940	0.942	1.806	1.793	0.954	0.954	1.791	1.805	1.766	1.773	0.955	0.953	0.958	0.963	0.981	1.009	1.012	1.005
Ca	0.970	0.969	0.025	0.023	0.971	0.957	0.012	0.013	0.015	0.017	0.988	0.980	0.957	0.980	0.985	0.908	0.910	0.913
Fe ²⁺	0.048	0.042	0.117	0.149	0.053	0.059	0.158	0.144	0.187	0.170	0.030	0.035	0.022	0.023	0.007	0.029	0.032	0.030
Na	0.006	0.005	0.001	0.001	0.006	0.006	0.001	0.001	0.001	0.001	0.004	0.001	0.023	0.013	0.004	0.011	0.013	0.011
K	0.000	0.000	0.000	0.000	0.001	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ni	0.002	0.001	0.002	0.002	0.002	0.001	0.004	0.004	0.002	0.003	0.001	0.001	0.002	0.002	0.002	0.003	0.003	0.002
Mg#	95.1	95.7	93.9	92.3	94.7	94.2	91.9	92.6	90.4	91.3	97.0	96.5	97.7	97.6	99.3	97.2	97.0	97.1

Oxide values in weight percent, chemical formulae calculated on the basis of 6 oxygens, *Cpx = Clinopyroxene; **Opx = Orthopyroxene; Mg# = 100 × Mg/(Mg + Fe²⁺).

2011; Zaccarini et al., 2011). In various types of melts, the total concentration of Cr₂O₃ and Al₂O₃ is controlling the Cr/Al ratio of chromian spinel, although these two elements have negative correlation to each other. The concentrations of Al and Ti in chromian spinel are directly transferred from their parent magma and can be used to restrain the primary melt composition, thus providing a basis for further understanding of tectonic background of podiform chromitites and the host ophiolites (Barnes and Roeder, 2001; Kamenetsky et al., 2001; Page and Barnes, 2009; Uysal et al., 2009a; Gonzalez-Jimenez et al., 2011; Zaccarini et al., 2011; Rollinson and Adetunji, 2013).

Kamenetsky et al. (2001) presented that there is a linear relationship between the TiO₂ and Al₂O₃ contents of chromian spinel and the melt from which it crystallized by using parental melt in chromian spinel from volcanic rocks. Following the method of Rollinson (2008) (for arc sequences acquired from Kamenetsky et al., 2001), the Al₂O₃ and TiO₂ contents of parental melts are in equilibrium with the chromitites samples. The Al₂O₃ and TiO₂ contents of the magmas from which the Sakhakot-Qila high-Cr chromitites have been crystallized are determined by using the equations proposed by Rollinson (2008) and Rollinson and Adetunji (2015);

$$\text{Al}_2\text{O}_3 - \text{melt} = 7.1518 \times \text{Al}_2\text{O}_3 - \text{spinel}^{0.2387} \quad (1)$$

$$\text{TiO}_2 - \text{melt} = 1.5907 \times \text{TiO}_2 - \text{spinel}^{0.6322} \quad (2)$$

To conclude Al₂O₃ and TiO₂ contents of magmas due to which the Heru-Shah high-Al chromitites precipitated the following equations are used for MOR settings;

$$\text{Al}_2\text{O}_3 - \text{melt} = 5.2181 \times \ln(\text{Al}_2\text{O}_3 - \text{spinel}) - 1.0505 \quad (3)$$

$$\text{TiO}_2 - \text{melt} = 1.0963 \times \text{TiO}_2 - \text{spinel}^{0.7863} \quad (4)$$

Based on calculations, the high-Cr chromitites precipitated from magmas with Al₂O_{3(melt)} ranging from 10.2 to 11.2 wt% and TiO_{2(melt)} contents ranging from 0.12 to 0.30 wt%, correspondingly (Fig. 8a-b). The values that are similar to those of high-Cr chromitites parental melts in the Bonin Island, Japan (Hickey and Frey, 1982), Semail Ophiolite, Oman (Rollinson, 2008), the Therford Mines Ophiolite, Canada (Page and Barnes, 2009), the Sagua de Tanamo Ophiolite, Eastern Cuba (Gonzalez-Jimenez et al., 2011) and the Harmancik and Kizilyuksek Ophiolite, Turkey (Uysal et al., 2015; Avcı et al., 2017). In these regions the high-Cr chromitites are supposed to have formed in the subduction zone environments from boninitic melts similar to that of

Table 5. Representative composition of amphibole inclusions in chromian spinel of the high-Al chromitites in the mantle-crust transition zone (MTZ) dunites of Dargai Complex through EPMA (wt%)

Sample No	HS2-a3	HS2-a5	HS2-a6	HS2-a7	HS2-a9	HS2-a11	HS2-a14	HS2-a15
SiO ₂	45.7	47.2	46.4	45.8	45.2	44.6	44.9	44.5
TiO ₂	0.52	0.29	0.64	0.59	0.63	0.61	0.43	0.52
Al ₂ O ₃	9.65	8.91	9.39	9.02	8.99	9.01	9.01	8.41
Cr ₂ O ₃	2.57	1.93	2.13	2.51	2.09	2.16	2.05	1.99
FeO	3.14	2.96	3.13	3.11	2.51	2.77	2.46	2.78
MnO	0.08	0.06	0.04	0.07	0.05	0.04	0.07	0.05
MgO	19.4	18.9	19.3	19.5	19.2	19.4	19.1	18.8
CaO	13.2	12.4	12.4	12.3	12.7	12.5	12.2	12.3
Na ₂ O	4.36	4.11	3.09	2.51	2.77	2.49	2.18	2.15
K ₂ O	0.21	0.44	0.25	0.19	0.37	0.22	0.34	0.25
NiO	0.11	0.14	0.11	0.15	0.12	0.11	0.18	0.14
Total	99.9	99.3	99.9	99.8	99.6	99.9	99.9	99.9
O = 23								
Si	6.469	6.733	6.618	6.564	6.617	6.533	6.618	6.658
Ti	0.055	0.031	0.069	0.064	0.069	0.067	0.048	0.059
Al	1.61	1.498	1.578	1.523	1.551	1.555	1.565	1.483
Cr	0.288	0.218	0.24	0.284	0.242	0.25	0.239	0.235
Mn	0.01	0.007	0.005	0.008	0.006	0.005	0.009	0.006
Mg	4.094	4.019	4.104	4.166	4.19	4.237	4.197	4.194
Ca	2.002	1.895	1.895	1.888	1.992	1.962	1.926	1.972
Na	1.196	1.137	0.854	0.697	0.786	0.707	0.623	0.624
K	0.038	0.08	0.045	0.035	0.069	0.041	0.064	0.048
Ni	0.013	0.016	0.013	0.017	0.014	0.013	0.021	0.017
Fe ⁽ⁱⁱⁱ⁾	0	0	0	0	0	0	0	0
Fe ⁽ⁱⁱ⁾	0.372	0.353	0.373	0.373	0.307	0.339	0.303	0.348
Al ^(IV)	1.531	1.267	1.382	1.436	1.383	1.467	1.382	1.342
Al ^(VI)	0.078	0.231	0.196	0.087	0.168	0.089	0.183	0.141
Ca ^(C)	0.104	0.141	0.013	0.017	0.017	0.013	0.021	0.017
Ca ^(B)	1.898	1.754	1.882	1.871	1.975	1.949	1.905	1.955
Na ^(B)	0.102	0.246	0.118	0.129	0.025	0.051	0.095	0.045
Na ^(A)	1.094	0.891	0.737	0.569	0.761	0.656	0.528	0.579
K ^(A)	0.038	0.08	0.045	0.035	0.069	0.041	0.064	0.048
Mg#	91.7	91.9	91.7	91.8	93.2	92.6	93.3	92.3

Oxide values in weight percent, chemical formulae calculated on the basis of 23 oxygens, Mg# = $100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$.

the Sakhakot-Qila chromitites (Ahmed, 1982, 1984).

The Al₂O₃ and TiO₂ contents of the parental magma of the high-Al chromitites are normally ranging from 15.1 to 15.8 wt% and 0.44 to 0.71 wt%, respectively. The Al₂O₃ contents (16 wt%, Fig. 8a) of the studied high-Al chromitites samples are in good match with those of MORB environment (Wilson, 1989). This indicates their precipitation in a MORB type environment. However, the parental magmas of chromitites of high-Al are not of typical mid oceanic ridges basaltic magmas because they have lower TiO₂ than the normal concentrations (Fig. 8b). This suggests that, the magma is likely to have undergone low degree of melt extraction through a partial melting of residual mantle.

Oxygen Fugacity (fO₂) of Chromitite and Peridotite

Oxygen fugacity (fO₂) of the earth's mantle is one of the intensive

variables restraining the progression of mantle derived magmatic rocks. The fO₂ of podiform chromitites and associated ultramafic rocks contributes the direct evidence of the deep physical and chemical phenomenon occurred in the earth's mantle (Ballhaus et al., 1991). In the other tectonic settings, it is found that the mantle wedge above the subduction zone setting is more oxidized than the mantle (Ballhaus et al., 1991; Parkinson and Arculus, 1999). The oxygen fugacity of high-Cr chromitites, mantle dunites and harzburgites has been calculated on the basis of microprobe analysis of the studied olivine and chromian spinel pairs according to Ballhaus et al. (1991) and Sack and Ghiorso (1991). The fO₂ contents of high-Al chromitites and MTZ dunites have also been considered for comparisons. On the basis of average mineral chemistry, the fO₂ contents have been calculated and the results are shown in Table 7 as deviation ($\Delta \log f\text{O}_2$ FMQ) from the fayalite-magnetite-quartz (FMQ) buffer at 1.0 GPa. The fO₂ values

Table 6. Platinum group element compositions of whole rock samples of the chromitites and peridotites in the Dargai Complex

Lithology	S.No	Os	Ir	Ru	Rh	Pt	Pd	ΣPGE	Pd/Ir	Os/Ir	Pd/Os	Ir/(Pt+Pd)	PPGE/IPGE
High-Cr Chromitites	SQ3	33.5	14.8	43.6	3.39	17.2	12.8	125.3	0.86	2.38	0.36	0.49	0.37
	SQ9	49.2	39.3	38.3	7.06	3.19	0.48	137.5	0.02	1.60	0.01	10.7	0.04
	SQ10	45.3	12.7	98.1	2.77	13.6	4.99	177.5	0.50	4.56	0.11	0.68	0.06
	SQ12	72.4	47.6	92.7	1.83	2.36	41.6	258.5	1.11	1.94	0.57	1.08	0.12
	SQ14	87.8	16.5	22.8	5.45	0.95	19.4	152.9	1.49	6.80	0.22	0.81	0.14
	SQ16	38.7	11.1	71.3	10.5	8.38	2.07	142.1	0.24	4.45	0.05	1.06	0.07
	SQ19	35.9	32.2	29.2	6.91	6.13	1.92	112.3	0.08	1.42	0.05	4.00	0.08
	SQ20	42.4	44.7	36.9	5.71	1.47	9.18	140.4	0.26	1.21	0.21	4.20	0.08
High-Al Chromitites	HS1	7.98	2.47	22.6	1.58	14.9	53.9	103.4	27.6	4.13	6.69	0.04	0.97
	HS7	9.15	1.84	16.8	9.97	52.2	23.1	113.1	15.9	6.35	2.50	0.02	1.43
	HS9	3.01	5.26	7.12	2.94	39.4	125	182.7	30.1	0.73	41.1	0.03	6.24
	HS10	9.91	4.39	5.55	1.79	5.52	11.9	39.1	3.43	2.88	1.19	0.25	0.64
	HS11	2.96	7.09	11.7	2.13	18.8	91.2	133.9	16.3	0.53	30.5	0.06	2.75
	HS12	7.11	2.17	28.9	1.43	89.2	71.4	200.2	41.7	4.19	9.95	0.01	1.84
	HS13	6.35	7.98	14.2	7.36	4.44	25.8	66.1	4.09	1.02	4.03	0.26	0.66
	HS14	4.51	1.88	25.4	2.04	8.16	64.1	106.1	43.2	3.07	14.1	0.03	0.97
Harzburgite	SQ6	4.19	3.08	5.98	1.21	4.79	4.53	23.8	1.86	1.74	1.07	0.33	0.43
	SQ8	5.31	2.54	6.58	0.84	5.67	2.99	23.9	1.49	2.67	0.56	0.29	0.36
	SQ11	4.87	2.93	3.87	0.92	4.63	0.41	17.6	0.18	2.12	0.08	0.58	0.31
	SQ15	3.75	3.38	6.12	0.75	5.85	1.46	21.3	0.55	1.42	0.39	0.46	0.33
Mantle Dunites	SQ7	3.22	1.21	4.11	0.31	1.12	2.97	12.9	3.11	3.40	0.91	0.30	0.28
	SQ13	0.78	2.89	2.75	0.55	0.55	0.31	7.83	0.14	0.34	0.39	3.36	0.11
	SQ17	3.49	1.09	1.01	0.44	2.19	0.59	8.81	0.69	4.09	0.17	0.39	0.42
	SQ17	2.83	0.84	5.81	0.38	1.85	1.63	13.3	2.46	4.30	0.57	0.24	0.19
MTZ Dunites	HS4	0.74	0.37	2.03	0.21	2.34	1.05	6.74	3.59	2.56	1.41	0.11	0.53
	HS5	0.99	1.22	1.11	0.91	0.53	2.53	7.29	2.63	1.04	2.53	0.40	0.71
	HS6	2.33	0.78	4.52	0.18	1.67	0.75	10.2	1.22	3.82	0.32	0.32	0.16
	HS8	1.65	0.93	1.58	0.15	3.79	1.93	10.0	2.63	2.27	1.16	0.16	0.87
*PM		514	540	690	200	1020	545						
**CC1		3.40	3.20	2.00	0.90	7.10	3.90						

Contents are in ppm, IPGE = Ir, Os, Ru, PPGE = Pd, Pt, Rh, *PM (primitive mantle) values used from McDonough and sun (1995); **CC1 (chondrite of average type C1) values used from Naldrett and Duke (1980).

obtained from the Sakhakot-Qila high-Cr chromitites (log units ranging from 0.83 to 1.85), mantle dunite (-0.04 to 1.26) and harzburgite (-1.76 to 0.21) are higher or less than the FMQ buffer (Fig. 9). The high-Cr chromitites and mantle dunites chromian spinel have higher Cr# values and oxygen fugacity with respect to associated harzburgite, which is thought to be due to interface between residual mantle harzburgite and boninitic melt in a supra-subduction zone (SSZ) setting (Ishii et al., 1992; Parkinson and Pearce, 1998).

The oxygen fugacity values of the Heru-Shah high-Al chromitites and mantle-crust transition zone (MTZ) dunites are higher than the fayalite-magnetite-quartz buffer, having log units for high-Al chromitites ranging from 0.43 to 1.53 and for MTZ dunites ranging from 1.09 to 1.98 (Table 7; Fig. 9). For MORB, these values have high oxygen fugacity than expected which is supposed to approximate the curve of FMQ (Cottrell and Kelley, 2011), but equivalent to those for high-Cr chromitites, mantle dunites and harzburgites which originated under supra-subduction zone setting. In addition, the occurrence of amphibole inclusions in the high-Al chromitites infers that they are precipi-

tated by parental melts with hydrous nature. Therefore, the parent melts of high-Al chromitites are considered to be oxidized, hydrous in nature and poor in TiO₂ but Al₂O₃ rich in characteristics.

PGE Occurrences and Potential in Dargai Complex Chromitites and Peridotites

In igneous rocks, the three main factors that usually control the PGE concentrations are the magmatic fractionation, partial melting and hydrothermal alteration processes (Barnes et al., 1985). The first two processes seem to be more effective while the last one have minor effect on the redistribution and remobilization to a slighter extent Pt and Pd. On the basis of compatibility of their melting temperature, PGEs are divided into two groups: the Ir-group, more refractory elements (IPGEs = Os, Ir and Ru) that have melting points > 2000 °C, and the Pd-group, less refractory elements (PPGEs = Rh, Pt, Pd) with melting points < 2000 °C (Barnes et al., 1985; Bockrath et al., 2004). The combined processes of partial melting and melt/rock interaction produced PGEs

Figure 7. (a) Chondrite-normalized platinum group elements (PGE) patterns for the representative mantle dunites, harzburgite and high-Cr chromitites; (b) Chondrite-normalized platinum group elements patterns for the MTZ dunites and high-Al chromitites, (c) Chondrite-normalized Pd_N/Ir_N versus Pd_N/Os_N for the peridotites and high-Cr and high-Al chromitites. Platinum group elements (PGE) values of the PM (primitive mantle) are from McDonough and Sun (1995), and chondrite values are from Naldrett and Duke (1980).

Figure 8. The parental melts (a) Al_2O_3 and (b) TiO_2 contents are in equilibrium with the studied chromitites from Dargai Complex. The Arc and MORB lines are reported by Kamenetsky et al. (2001) and Rollinson (2008). The calculated melt compositions and the chromian spinel ranges from Dargai complex are shown as light blue green and ocean green fields.

and Cr contents in the mantle rocks (Zhou et al., 1998).

In Dargai Complex, the total PGE contents of both the high-Al and high-Cr chromitites are much higher than the concentrations of PGE in MTZ dunites, mantle dunites and harzburgites (Fig. 7a-b). It is gen-

erally believed that platinum group elements are added in basaltic melts during partial melting of residual sulfide, and to a smaller extent alloys, that reside in the peridotites of mantle (Luguet et al., 2001; Alard et al., 2002; Bockrath et al., 2004; Lorand et al., 2010). Hence,

the melts PGE compositions depend to a large extent on partial melting degrees. Due to the low degree of partial melting (< 20%), sulfides are usually retained in the mantle, so the PGE abundances of the melts is low, and any chromitites precipitated from such melts will also have very low PGE concentrations (some of ppb). On the other hand, melts produced by the high degrees of partial melting ($\geq 20\%$), are PGE enriched, leads to higher PGE abundances in the related chromitites (Economou-Eliopoulos, 1996; Zhou et al., 1998; Ahmed and Arai, 2002; O'Hara et al., 2002; Gervilla et al., 2005; Prichard et al., 2008; Proenza et al., 2008; Uysal et al., 2009b; Gonzalez-Jimenez et al., 2011). In Dargai Complex, the high-Al chromitites have similar, but slightly lower, total concentrations of PGEs from 39.1 to 200 ppb than the high-Cr chromitites ranging from 112 to 258 ppb (Fig. 7a; Table 6). These outcomes illustrate that their magmas were both originated by relatively higher degree of partial melting, but probably the melts of high-Al chromitites were formed by somewhat low degrees of melting than the high-Cr types.

Under the conditions of sulfur under-saturation, PGEs distribution of podiform chromitites is basically consistent with initial chromite accumulation from mafic characteristics magmas. Early research on chromitites suggest that the stratiform chromitites are rich in PPGEs and have normalized PGE patterns with positive sloping chondrite, whereas podiform chromitites in ophiolites are IPGE-rich and have negative sloping pattern characteristics (Page et al., 1982; Page and

Talkington, 1984). Therefore, PPGE enriched podiform chromitites have since been studied from Philippines Zambales Ophiolite (Bacuta et al., 1988), Greece Pindos Ophiolite (Konstantopoulou and Economou-Eliopoulos, 1991), Scotland Pt- and Pd-rich Ophiolite include Unst in the Shetland Island (Prichard and Lord, 1993), from Norway Leka Ophiolite (Pedersen et al., 1993) and from Saudi Arabia Al'Ays Ophiolite (Prichard et al., 2008). According to Prichard et al. (2008), chromium in the mantle can be liberated over a much higher partial melting range than the narrow critical melting interval at around 20 to 25% proposed for PPGE, which is considered for the largely variable PPGE concentrations of podiform chromitite. The high-Cr chromitites of Dargai Complex, are poor in PPGE concentrations comparative to IPGE with lower Pt/Ir ratio which is ranging from 0.02 to 1.49 (Table 6; Fig. 7c). This indicates that the degree of partial melting exceeded the critical interval, thus preventing sulfide saturation and diluting volatile PGE (i.e., Pt and Pd) in melting and is resulted in enrichment of PGEs chromitites, while Pt and Pd are particularly barren (Keays, 1995; Najafzadeh et al., 2008; Prichard et al., 2008; Gonzalez-Jimenez et al., 2011). Such explanations are consistent with the crystallization of high-Cr chromitites due to boninitic type magmas, which are sulfur under-saturated in composition (Zhou et al., 1998; Melcher et al., 1999; Ahmed et al., 2009).

In contrast, high-Al chromitites of the Dargai Complex are low in IPGE contents relative to PPGE with higher Pd/Ir ratio ranging from

Table 7. Calculated oxygen fugacities (fO_2) of Dargai Complex peridotites and chromitites

Lithology	S. No	Spinel Cr#	Spinel Mg#	Spinel $Fe^{3+}/\Sigma Fe$	Olivine Mg#	Temp (K)	$\Delta \log(fO_2)$ FMQ	Spinel $Fe^{3+}\#$	Spinel TiO_2 (wt%)
Harzburgite	SQ4-1	74.2	39.5	0.053	89.6	924.5	-1.759	0.016	0.08
	SQ4-2	73.2	41.7	0.104	89.2	961.8	-0.650	0.033	0.05
	SQ4-3	77.5	48.1	0.194	88.6	1111.1	0.157	0.059	0.07
	SQ4-4	72.2	45.6	0.093	88.7	1010.1	-1.120	0.027	0.05
	SQ4-5	75.6	47.3	0.188	88.9	1072.2	0.214	0.059	0.06
Mantle Dunite	SQ5-1	79.8	40.5	0.168	91.9	910.1	0.962	0.060	0.12
	SQ5-2	82.6	42.0	0.169	91.7	949.5	0.839	0.060	0.15
	SQ5-3	80.1	46.9	0.238	91.0	1015.0	1.263	0.085	0.14
	SQ5-4	81.6	46.0	0.127	91.0	994.7	-0.043	0.040	0.17
	SQ5-5	81.0	42.6	0.165	91.9	937.4	0.841	0.057	0.16
MTZ Dunite	HS4-1	46.7	66.8	0.335	89.5	1142.8	1.093	0.086	0.17
	HS4-2	48.9	59.5	0.323	88.7	1065.2	1.303	0.099	0.19
	HS4-3	49.4	67.3	0.460	89.8	1207.4	1.976	0.147	0.22
	HS4-4	48.9	59.8	0.312	89.2	1050.6	1.278	0.093	0.18
	HS4-5	48.3	59.4	0.316	89.7	1024.0	1.453	0.096	0.19
High-Cr Chromitite	SQ1-1	80.8	69.3	0.264	94.2	1195.6	1.122	0.056	0.16
	SQ1-2	79.7	69.4	0.279	94.8	1145.2	1.487	0.061	0.18
	SQ1-3	81.9	74.4	0.393	94.3	1407.7	1.491	0.075	0.11
	SQ1-4	79.1	69.3	0.234	94.2	1178.2	0.834	0.047	0.14
	SQ1-5	81.4	71.3	0.361	95.1	1239.4	1.852	0.073	0.06
High-Al Chromitite	HS2-1	47.5	64.2	0.245	89.4	1071.6	1.533	0.060	0.13
	HS2-2	50.7	71.3	0.238	88.4	1337.9	0.427	0.045	0.19
	HS2-3	48.4	73.1	0.274	91.7	1166.8	0.562	0.053	0.14
	HS2-4	52.2	65.8	0.260	91.8	1051.0	0.979	0.061	0.24
	HS2-5	53.0	68.2	0.420	88.5	1316.0	1.269	0.119	0.14

Formula units are based on O = 32 and Fe^{2+}/Fe^{3+} assuming full site occupancy. Cr# = $100 \times Cr/(Cr + Al)$; Mg# = $100 \times Mg/(Mg + Fe^{2+})$; $Fe^{3+}\# = 100 \times Fe^{3+}/(Fe^{3+} + Cr + Al)$; Spinel $Fe^{3+}/Fe = Fe^{3+}/(Fe^{3+} + Fe^{2+})$. S.No = number of samples; Tem p = temperature; FMQ = fayalite-magnetite-quartz buffer.

3.43 to 43.2 (Table 6; Fig. 7c). This shows that degree of partial melting of the mantle sources were just within the critical range of 20 to 25%, which could have effectively extracted PGEs from mantle sulfides, but were not high enough to dilute the parental melts in the refractory elements (PPGEs) (Prichard et al., 2008; Gonzalez-Jimenez et al., 2011).

Tectonic Setting of Chromitites and Peridotites of Dargai Complex

The coexistence of both high-Cr and high-Al chromitites in the same ophiolitic block has been commonly interpreted as reflecting different melt activities that originated from different mantle sources in different tectonic environments (Zhou et al., 1994, 1996, 1998; Economou-Eliopoulos, 1996; Proenza et al., 1999; Ahmed and Arai, 2002; Rollinson, 2008; Tsoupas and Economou-Eliopoulos, 2008). Although, the progressive evolution of parental melts during melt-rock reaction or via melt-migration in supra-subduction zone (SSZ) environment can also provide interpretation for the coexistence of both high-Cr and high-Al chromitites in a single ophiolitic belt (Leblanc, 1995; Graham et al., 1996; Melcher et al., 1997; Ahmed and Arai, 2002; Rollinson, 2005; Uysal et al., 2009a).

In the Dargai Complex, the mantle section comprises high-Cr chromitites, whereas mantle-crust transition zone (MTZ) encloses the high-Al chromitites at higher level. Chemical characteristics of chromian spinel such as high Cr# values (78.4 to 81.9) and low TiO₂ (0.06 to 0.18 wt%) and Al₂O₃ (8.65 to 10.6 wt%) contents for the parental melts of high-Cr

chromitites suggest that in Dargai Complex the high-Cr chromitites were formed from parental melt with boninitic composition related to supra-subduction zone (SSZ) tectonic environment. However, chromian spinels in the high-Al chromitites fall in the mid-ocean ridge basalt (MORB) field based on Cr# ranging from 47.5 to 52.9, Mg# from 64.2 to 73.1 and TiO₂ from 0.13 to 0.24 wt%. These concentrations of chromian spinel in the high-Al chromitites are distinguished by a continuous and large spectrum, with some values plotted at the border of MORB field (Fig. 5d).

The high-Al chromitites display a continuous spectrum of calculated TiO₂ concentrations of the parental melts lower than typical MORB melts (Fig. 8b), which suggests that the MORB magmas have a less role in formation of high-Al chromitites. Other than that, the presence of hydrous silicate inclusions (Fig. 3f-g) and their high oxygen fugacities (Fig. 9) in the chromian spinel of the high-Al chromitites shows that they are formed by typical supra-subduction zone (SSZ) from the hydrous oxidized and low TiO₂ magma. It is consistent with the degree of high partial melting assumed for the creation of high-Al chromitites parental melts (20 to 25%), which is greater than the average degrees of partial melting (10 to 20%) for mid-ocean ridge basalt (MORB) magma (Klein and Langmuir 1987; Hofmann, 1988), but within the scope of supra-subduction zone (SSZ) tectonic environment melt (Prichard et al., 2008). Thus it is proposed that the high-Al chromitites of Heru-Shah have been generated at a spreading center in a supra-subduction zone (SSZ) setting.

Recent research on the Izu-Bonin-Mariana (IBM) arc system, estab-

Figure 9. Plot of $\Delta\log f_{O_2}$ (FMQ) vs. Cr# of chromian spinel from the high-Cr and high-Al chromitites, harzburgites, mantle dunites and mantle-crust transition zone dunites of the Dargai Complex. The $\Delta\log f_{O_2}$ (Ballhaus et al., 1991) values have been calculated from geothermometric data derived from olivine-spinel equilibria following the approach of Wan et al. (2008); the data are provide in Table 7. MOR-SSZ discrimination boundaries for dunites (solid line) and harzburgite (dashed line) are shown. MOR (mid ocean ridge), SSZ (supra-subduction zone), BON (boninite), IAT (island arc tholeiite) (Parkinson and Pearce, 1998; Dare et al., 2009).

lished that there is existence of chemical transition of lavas between the forearc-basalt and boninite (Reagan et al., 2010; Li et al., 2013; Brounce et al., 2015). The changing mid-ocean ridge basalt (MORB) magmas composition to boninitic type could explain the higher range of Cr#s in chromian spinel in the Izu-Bonin-Mariana (IBM) forearc mantle (Morishita et al., 2011a; Tamura et al., 2016). In the other hand, the existence of MORB like and boninitic type magmas in the Dargai Complex indicates a two-phase process of magmatism, which usually occurs in a proto-forearc spreading during the initiation of subduction (Ahmed, 1982, 1984; Arif and Jan, 2006). The first phase is characterized by upwelling of asthenosphere over the sinking oceanic slab generating MORB like melts, which are Ti rich. Both the chromian spinels in the MTZ dunites and pyroxenes and amphibole inclusion in high-Al chromitites spinels having comparatively higher TiO₂ contents are well-matched to such a model (Fig. 5d, 6b, 6e). In the second phase, the parental melts altered from MORB type to boninitic type in chemical characteristics as the proto-forearc environment evolves into a forearc setting, causing the crystallization of high-Al chromitites and then high-Cr chromitites. These processes can describe the relationships observed in Dargai Complex Ophiolite which are well-matched with the models proposed by some authors (Morishita et al., 2011b; Qiu et al., 2018), who presented distinct peridotite (harzburgites and dunites) in the mantle section and concluded that clinopyroxene bearing harzburgite rocks in the mantle section are a residues associated to MORB type forearc basalt. This model shows that, the highly refractory harzburgites, mantle dunites (and chromitites) may have genetically associated to boninitic type magma in the mantle-wedge, and by interaction with transitional melt between MORB and boninitic types magmas, some spinel bearing dunites may have usually formed. Hence, our preliminary conclusions is that the lithological variation and the two kinds of chromitites in the Dargai Complex reflect evolutions in magma composition from MORB type to boninitic type during subduction initiation process.

Conclusions

The petrography, mineralogy and geochemistry based on the chromian spinel and platinum group elements of the chromitites and peridotites of the Dargai Complex suggest that the high-Cr chromitites were existed in peridotites (harzburgites and dunites) of the mantle section, while high-Al chromitites were hosted in the mantle-crust transition zone (MTZ) massive dunites.

The total PGE concentrations in high-Cr chromitites indicate that these are higher in IPGE than that of the high-Al chromitites. For the parental magma of the high-Cr chromitites, these differences are related to the higher degree (>25%) of partial melting, which prevented the saturation of sulfides and caused the dilution of Pt and Pd in the parental magma. The high-Al chromitites parental magmas were produced at lower degree (20-25%) of partial melting, which could have effectively extracted the PGE from the mantle sulfide and as a result caused PPGE enrichment in high-Al chromitites type.

In the Dargai Complex, the coexistence of high-Cr chromitites and high-Al chromitites suggests that the compositions of magmas evolved from MORB type to boninitic type in a proto-forearc setting during subduction initiation. The transitional lavas, which are generally MORB

type with affinities of low TiO₂, oxidized and hydrous nature, could have been the parental magma of high-Al chromitites, whereas boninitic type melts could have been responsible for the formation of high-Cr chromitites.

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