

Mineral records of the pyroxenites formed within harzburgites (Ulaş, Sivas, Turkey): implications on petrogenesis and tectonic setting*

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Abstract: In this article, the petrological processes of the clinopyroxenite dykes and host peridotites occurring in Ulaş District (Sivas, mid-Anatolia, Turkey) are discussed. The new geochemical data from major minerals in the clinopyroxenite dykes and host harzburgites revealed differences to the supra-subduction zone (SSZ)-type pyroxenites and peridotites. In particular, the NiO content of olivines in the host harzburgites showed the signature of the mantle, whereas rare olivines of the clinopyroxenites had a NiO content and Fo number that were inferior to those of mantle peridotites. The high Mg-number of clinopyroxenes in the clinopyroxenites was likely associated with the partial remelting of the host harzburgites. Additionally, the rare earth element pattern of clinopyroxenes from the studied clinopyroxenites exhibited a similar pattern to those of the other SSZ or fore-arc clinopyroxenes. Additionally, the low Mg-number and relatively high Cr-number of spinels in the clinopyroxenites showed similarity to the subduction-related origin. Based on textural and geochemical evidence, the harzburgites were interpreted as depleted mantle rock, which was modified by melt-peridotite interactions. Consequently, the pyroxenites likely occurred as a crystallizing or cumulative zone of the SSZ-type melt and the minerals were gained from partial melting of the harzburgites through the interaction with such magma.

Key words: Pyroxenite dyke, peridotite, subduction zone, partial melting, Sivas, Turkey

1. Introduction

Pyroxenites generally occur as veins or dykes in peridotites related to ophiolites and represent up to 5% of all ultramafic bodies (Pearson and Nowell, 2004; Downes, 2007; Van Acken et al., 2010). However, these subject pyroxenites have been found in relationship with the abyssal peridotites, which are extremely rare (Dantas et al., 2007; Van Acken et al., 2010). Despite the fact that the pyroxenites represent a very small portion of the upper mantle, essential information is provided within on the petrological and dynamic processes of the Earth. This includes crustal recycling and melt-rock interaction in order to interpret the different tectonic environments (Downes, 2007; Gonzaga et al., 2010; France et al., 2015). The origin of the pyroxenite is still debated, supporting an enormous scope of theories, which are often reciprocal. Previous studies proposed the following to clarify the origin of the mantle pyroxenites: (1) metamorphic resolution of the peridotites (Dick and Sinton, 1979), (2) crystal precipitation of silicate magmas derived from the asthenosphere (Loubet and Allegre, 1982), (3) remnants of the subducted oceanic lithosphere (Allegre and Turcotte,

1986), and (4) high pressure crystal segregates of melting subducted oceanic crust (Pearson et al., 1993). Numerous studies examined the pyroxenite developments in the mantle, including the collaboration between peridotite and penetrating melt or the response amongst the peridotite and liquefied subducted lithosphere (Yaxley and Green, 1998; Garrido and Bodinier, 1999; Santos et al., 2002; Bodinier and Godard, 2003; Brooker et al., 2004; Berly et al., 2006; Downes, 2007; Van Acken et al., 2010; France et al., 2015).

The pyroxenite veins/dykes in the mantle peridotites were interpreted by crystallization products of the mantle melts, indicating partial melting of the host peridotites (Menzies and Allen, 1974; Buchl et al., 2004). However, the variety of origins suggest that each can be generated in different tectonic environments through various petrologic processes.

In this paper, the mineral geochemical characteristics of the Ulaş (Sivas, mid-Turkey) clinopyroxenite dykes and the host rocks are demonstrated. This provides the first operation of petrogenesis, which is compared with the typical supra-subduction zone (SSZ) and abyssal

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pyroxenites, related to host peridotites. The objective of this study is to identify whether the pyroxenite occurred in an SSZ setting and to determine any differences to typical pyroxenites. Detailed mineral major and trace element data were obtained and interpreted on a series of numerical diagrams of melt-rock interaction processes in order to better understand the petrological mechanisms of origin.

2. Geological setting

In the Anatolian sector of the Alpine–Himalayan Orogenic Belt included in the study area, the ophiolitic rocks of Turkey are the ruins of the Tethyan ophiolites (MTA, 2002). These ophiolitic rocks are found in tectonic units of Turkey, from south to north, composed of Taurides, Anatolides, Pontides, and the Border Folds (Ketin, 1983; Okay and Tüysüz, 1999), as shown in Figure 1a.

The Sivas basin is a residual basin between the Tauride and Pontide platforms (Cater et al., 1991; Poisson et al., 1996) and is 60 km long and 30 km wide, formed over the course of the closure of the northern branch of the Neo-Tethys within the early Tertiary (Yılmaz and Yılmaz, 2006). The ophiolitic units, previously called the Divriği ophiolite (Yılmaz et al., 2001; Yılmaz and Yılmaz, 2004), bear ultramafics mostly of Cretaceous age transported in soils that are ubiquitous and stretch over hundreds of kilometers in Turkey, especially in the Sivas region (MTA, 2002), as shown in Figure 1b. This region is composed of autochthonous platform carbonates, ophiolitic mélange, metamorphic sole, ophiolitic rocks, volcano-sedimentary units, granitoid rocks, and sediments (Yılmaz et al., 2001; Yılmaz and Yılmaz, 2004; Parlak, 2016). The mantle tectonites of this ophiolite are composed of harzburgite containing dunitic lenses with chromite pods (Parlak, 2016). The widespread harzburgites were cut by pyroxenite and diabasic (partly gabbroic) dykes at different levels (Bilici, 2015; Parlak, 2016).

In the study area (southeastern Ulaş, Sivas), the dark red, partly altered dunites were preserved in narrow fields within the harzburgitic peridotites. The harzburgites are more widespread than other ultramafics, which are cut by thin, coarse-grained, and grayish green-black pyroxenitic dykes as well as less dioritic, diabasic dykes several meters thick. Observed pyroxenite dykes have been mapped locally, in the study area around the Yaycı Mountain, as shown in Figure 2 (Bilici, 2015).

3. Analytical methods

Representative minerals were selected from the harzburgite and clinopyroxenite samples. A series of thin sections of these lithologies were examined under a polarizing microscope to determine the common

textural and mineralogical properties of the rock varieties. The major oxide analyses of olivine, clinopyroxene, orthopyroxene, and chromian spinels were performed on polished thin sections at the Electron Microprobe Laboratory of the University of Maine (Orono, ME, USA). All element analyses were applied by energy-dispersive X-ray spectroscopy (EDX) using EDS and WDS detectors attached to a Cameca SX 100 scanning electron microscope. The accelerating voltage was 15 kV and the beam current was 3.3 nA with a 0.6- μ m beam diameter. The detection limits were ~0.01% and accuracy was better than 5%. Cationic ratios of elements were calculated on the basis of 6 oxygen atoms and 32 oxygen atoms assuming pyroxene and spinel stoichiometry, respectively.

On the other hand, the trace and rare earth element concentrations of clinopyroxenes were detected in situ for single mineral phases using the LA-ICP-MS method at the University of Houston (Houston, TX, USA). The system combines a 193-nm Ar-F Excimer laser ablation system and a quadrupole ICP-MS with collision and reaction cell in pulse counting mode. Laser ablation was applied at a constant point on the mineral surface at a fluence of 20 J cm⁻², a stroke repetition rate of 10 Hz, and a typical point diameter of 60–120 μ m.

4. Results

4.1. Field observations and petrography

The pyroxenites generally occur as dykes with varying sizes having relatively sharp contacts with the host harzburgitic peridotites (Figures 3a–3d). No lherzolites have been observed in the study area. Although the degree of serpentinization of the host peridotites is quite intensive, the pyroxenite dykes remain fresh and visually distinguishable, pyroxene-rich in composition (Figures 3a–3c). The variable thickness of these pyroxenite dykes ranges from about 50 cm–1 m to about 5–10 m in length. (Figure 3c). Dunite and chromite bodies are locally observed near these pyroxenite dykes (Figure 3d).

The mineral paragenesis of the clinopyroxenites comprises coarse euhedral crystals of clinopyroxene and rare orthopyroxene, which define it as a cumulate texture. The intercumulus space is occupied by fine-grained mosaic olivine and rarely by spinel. Based on the modal analysis with point counting on the thin sections, the pyroxenite samples from various levels of the harzburgitic peridotite suite can be defined as clinopyroxenites or olivine clinopyroxenites.

The harzburgites show typical porphyroclastic texture with >80 vol. % olivine and orthopyroxene, and also <3 vol. % cpx and ~1–2 vol. % chromian spinels and serpentine minerals (Figures 4a and 4b). The major mineral assemblages of the clinopyroxenites are dominated by

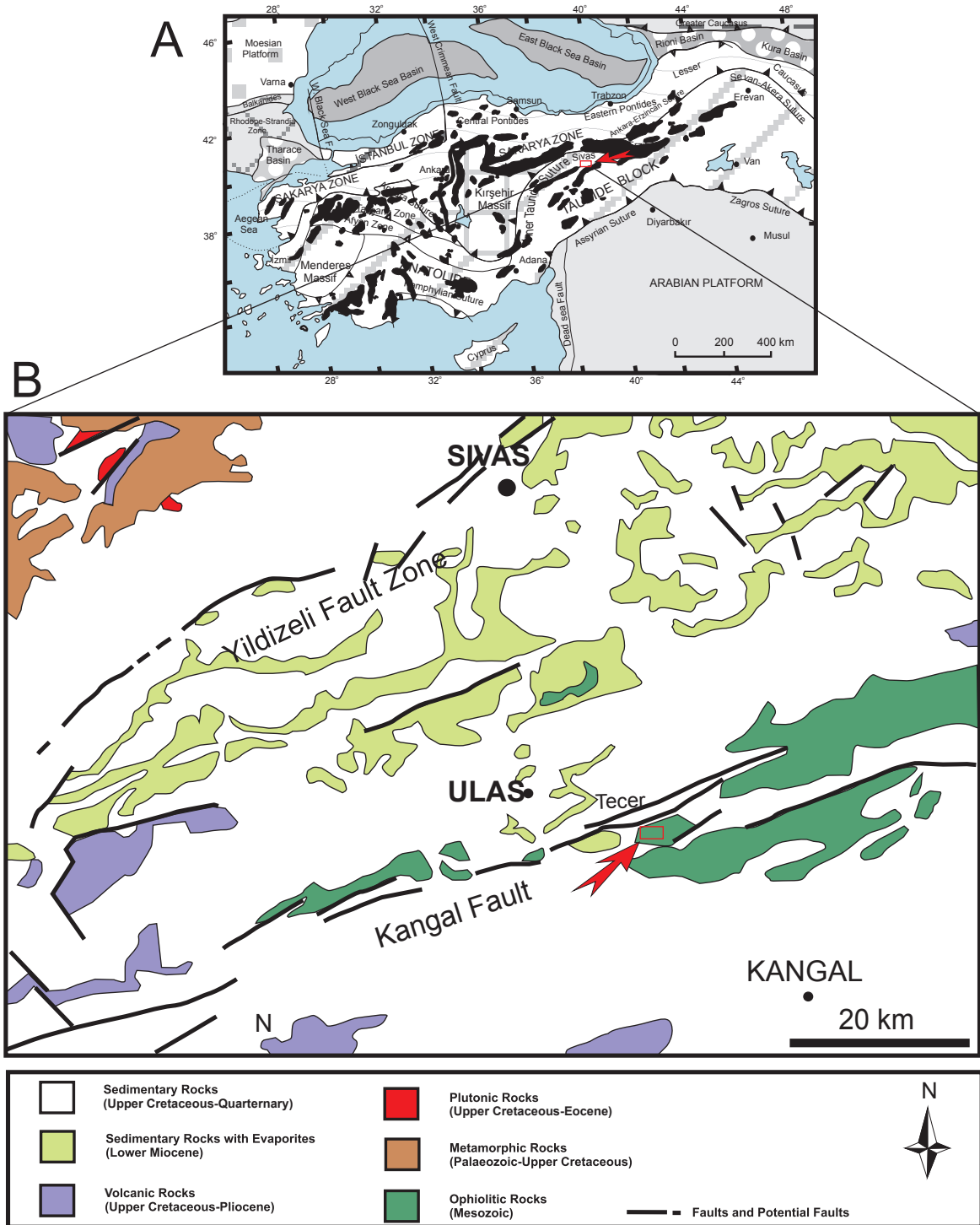


Figure 1. (a) Distribution of ophiolite belts in Turkey and tectonic map of the northeastern Mediterranean region showing the major sutures and continental blocks (modified from MTA (2002) and Okay and Tüysüz (1999)), (b) Geological map of the Ulaş (Sivas) area (modified from Yılmaz et al. (1989)).

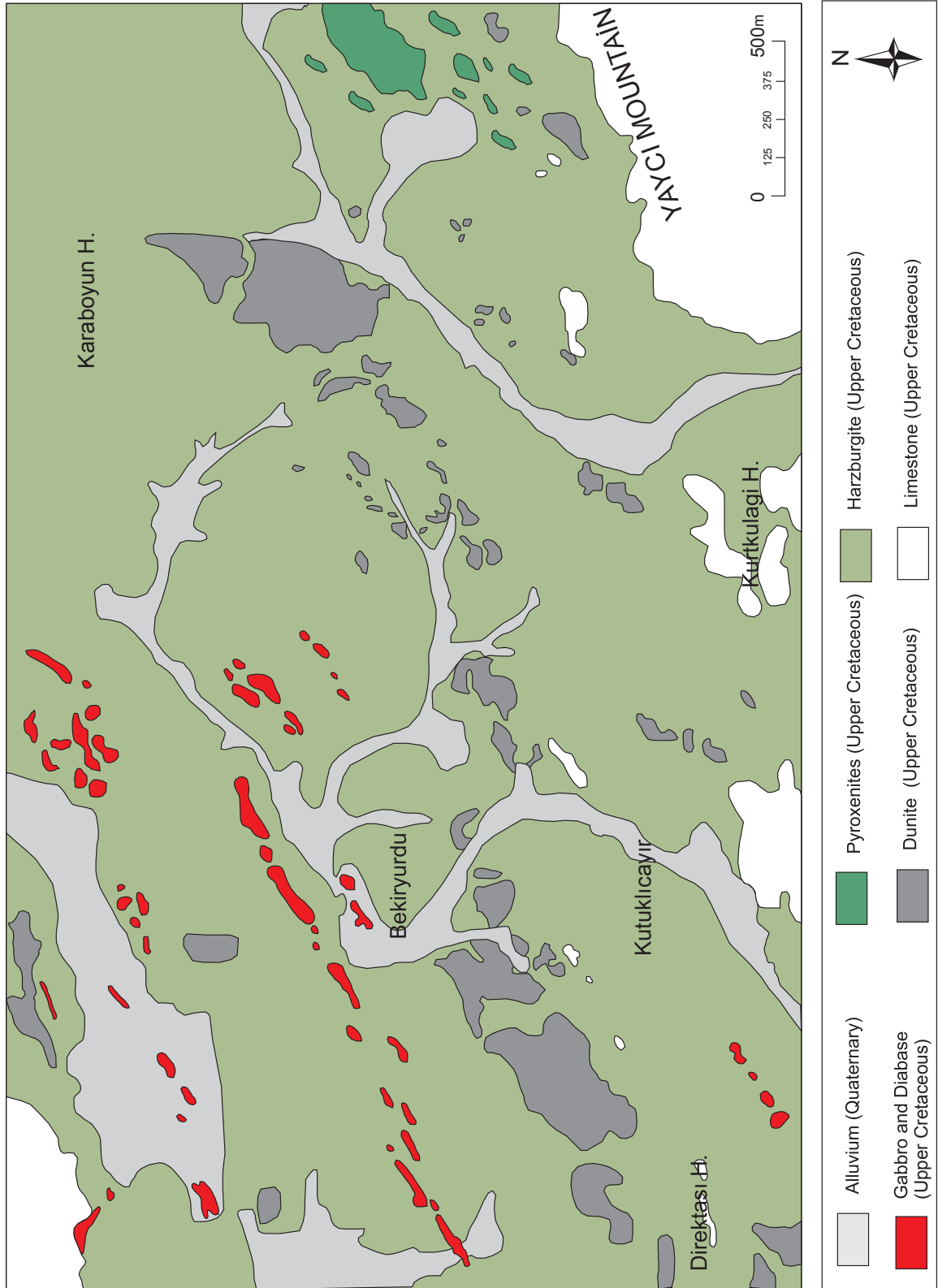


Figure 2. Detailed geological map of the study area (southeastern Ulaş, Sivas) (Bilici, 2015).

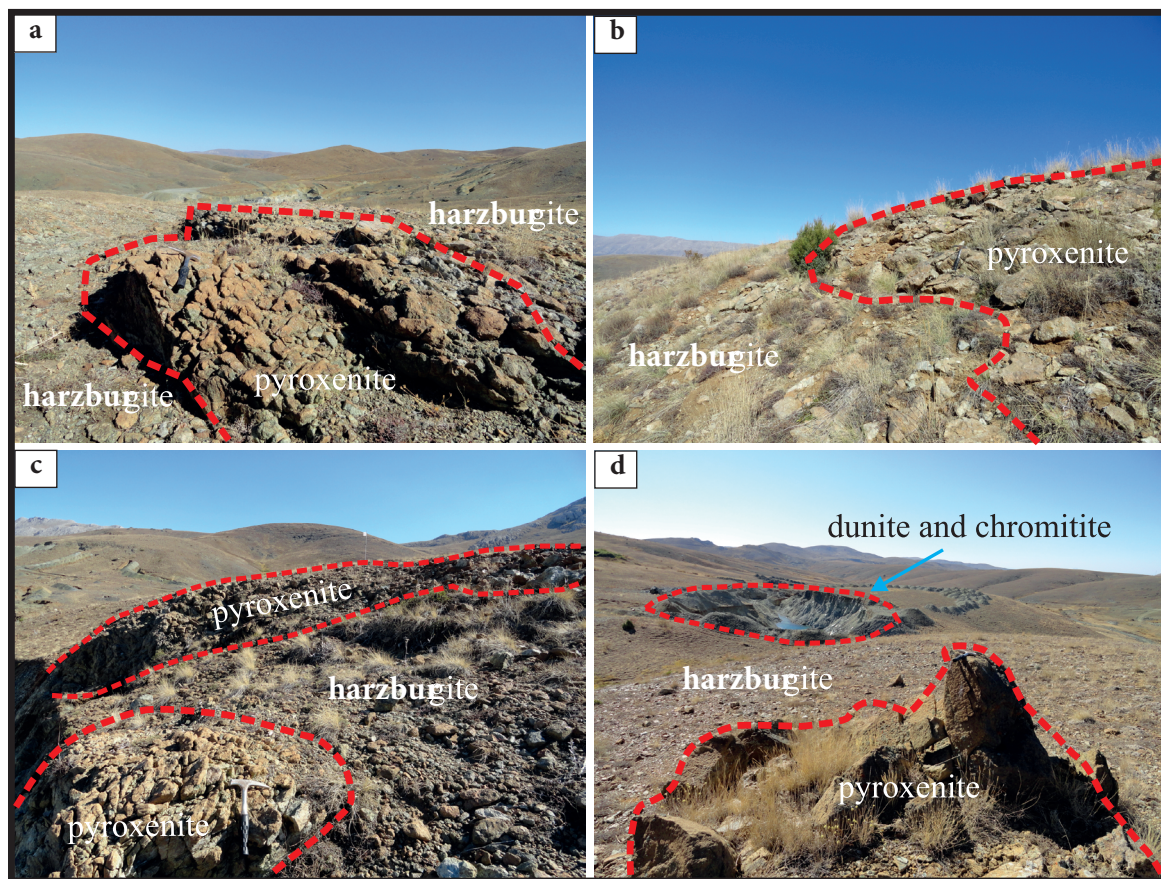


Figure 3. Field photographs showing the relationship between pyroxenite dykes and host peridotites (a–c). Pyroxenite dykes occur in variable thickness; (d) field view of the pyroxenite, harzburgite, and dunite with chromitite bodies.

clinopyroxene (>80 vol. %) with small amounts olivine–orthopyroxene (1–5 vol. %) and chromian spinel (~1 vol. %) and alteration minerals, such as serpentine minerals (Figures 4c–4f). Microscopic studies show that olivines in the studied clinopyroxenites are serpentinized and appear interstitially between clinopyroxenes (Figures 4c and 4d). The texture of these dykes is granular with large subhedral clinopyroxene crystals (Figures 4c–4f).

4.2. Mineral chemistry

4.2.1. Olivine

Results of the olivine analyses representing the clinopyroxenites and host harzburgite are given in Table 1. Olivines in the harzburgite samples show higher Fo-number than those in the clinopyroxenite samples and range from 90.34 to 91.82 and 88.20 to 88.83, respectively. Similarly, NiO (wt. %) contents of olivines are very low and show a decrease towards the clinopyroxenites from the harzburgites. The Fo-number and NiO contents of all olivine grains exhibit a positive correlation from the

clinopyroxenites to the harzburgites (Figure 5). The Fo-number vs. NiO diagram shows that olivines in the harzburgite plot mostly within the mantle olivine array and are comparable with those from the upper mantle peridotites. However, in the same diagram, the olivines in the clinopyroxenites have lower NiO contents (<0.2 wt. %) when compared with the other SSZ clinopyroxenites (Figure 5).

4.2.2. Orthopyroxene

Results of the major element analyses of orthopyroxenes are given in Table 2. The clinopyroxenite dykes contain very low modal orthopyroxene mineral abundance relative to the harzburgites. All of these orthopyroxenes are enstatite in composition in both rock types (Figure 6a). Although the Mg-number of orthopyroxenes in the clinopyroxenite ranges from 0.91 to 0.95, these values show a wider range in harzburgites (i.e. 0.89–0.94). The Al_2O_3 contents of orthopyroxene are low in both the harzburgites and the clinopyroxenites, ranging from 0.79 to 2.26 (wt. %) and from 1.24 to 2.44 (wt. %), respectively.

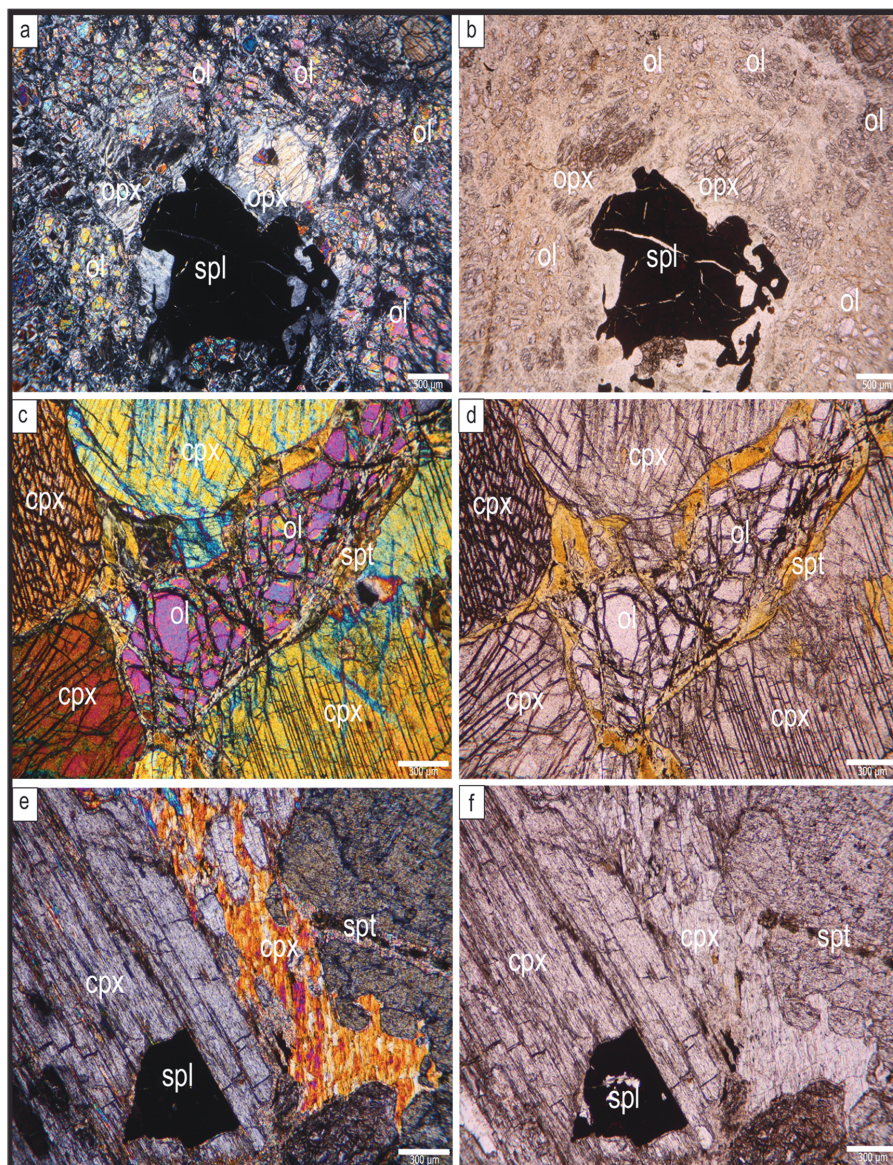


Figure 4. Photomicrographs from the harzburgite (a and b) and pyroxenite (c-f) of the study area (a, c, and e: crossed-nicol in polarized light; b, d, and f: plane polarized light; Abbreviations: spl: chromian spinel; ol: olivine; opx: orthopyroxene; cpx: clinopyroxene; spt: serpentine).

Overall, the harzburgites and the clinopyroxenites display similar distribution in the Mg-number vs. Al_2O_3 diagram and are comparable with other SSZ peridotites and clinopyroxenites (Figure 6b).

4.2.3. Clinopyroxene

All clinopyroxenes in both rock types have limited compositional variations (Table 2). Most of the analyzed clinopyroxenes are diopside in composition except for two samples, which are in augite composition in the clinopyroxenites. Generally, it can be said that the

Mg-numbers of clinopyroxenes decrease towards the harzburgites from the clinopyroxenites. The Al_2O_3 contents of all clinopyroxenes are low concentrations and range between 0.61 and 2.45 (wt. %) in the harzburgites and between 0.01 and 5.88 (wt. %) in the clinopyroxenites. Similarly, the Cr_2O_3 contents of clinopyroxenes range from 0.23 to 1.21 (wt. %) and from 0.00 to 1.12 (wt. %) in the harzburgites and the clinopyroxenites, respectively. Based on the increasing Mg-number, overall, the Al_2O_3 and Cr_2O_3 contents of clinopyroxenes decrease from the harzburgites to the clinopyroxenites (Figures 6c and 6d). In this regard,

Table 1. Representative major element analyses (in wt. %) of olivines from the Ulaş ultramafics (Hrz: harzburgite; Pyrx: pyroxenite; Ol: olivine).

Sample	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.
SiO ₂	41.16	41.82	41.16	41.44	41.03	41.27	41.09	41.31	41.15	41.71	41.23	41.54	41.67	41.14
TiO ₂	0.00	0.00	0.02	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Al ₂ O ₃	0.00	0.00	0.02	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00
Cr ₂ O ₃	0.01	0.07	0.00	0.07	0.01	0.30	0.01	0.03	0.05	0.02	0.07	0.00	0.01	0.07
FeO	8.93	8.87	8.97	9.05	8.69	8.03	8.47	8.88	8.81	9.04	8.78	7.74	9.19	9.29
MnO	0.14	0.16	0.15	0.13	0.16	0.09	0.05	0.14	0.16	0.18	0.18	0.11	0.16	0.12
MgO	49.61	48.64	49.37	48.73	49.41	50.08	49.69	48.67	49.25	48.84	48.89	48.76	48.50	48.72
CaO	0.01	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
NiO	0.41	0.39	0.40	0.42	0.38	0.42	0.46	0.44	0.48	0.41	0.40	0.48	0.36	0.42
Total	100.27	99.95	100.07	99.86	99.68	100.19	99.78	99.47	99.90	100.22	99.55	98.64	99.89	99.76
Fo	90.83	90.72	90.75	90.56	91.02	91.75	91.27	90.72	90.88	90.59	90.85	91.82	90.39	90.34
Sample	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.	Ol.
Rock type	Hrz.	Hrz.	Hrz.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.
SiO ₂	41.19	41.39	41.58	39.64	39.32	39.39	40.08	39.28	38.10	40.49	38.09	39.18	38.71	39.95
TiO ₂	0.00	0.00	0.00	0.05	0.01	0.01	0.01	0.02	0.01	0.01	0.02	0.01	0.00	0.02
Al ₂ O ₃	0.02	0.01	0.01	0.02	0.05	0.04	0.01	0.03	0.03	0.03	0.05	0.02	0.03	0.01
Cr ₂ O ₃	0.06	0.02	0.01	0.02	0.02	0.03	0.12	0.01	0.05	0.12	0.08	0.01	0.09	0.04
FeO	8.81	8.92	8.52	10.52	10.84	10.53	10.99	10.44	10.78	10.20	10.76	10.45	10.71	10.54
MnO	0.15	0.15	0.11	0.18	0.05	0.14	0.26	0.20	0.21	0.09	0.07	0.09	0.16	0.15
MgO	49.57	48.69	48.82	46.24	47.11	46.76	46.05	46.35	46.36	45.49	46.05	45.27	45.16	45.24
CaO	0.00	0.00	0.02	0.03	0.03	0.04	0.01	0.01	0.02	0.01	0.05	0.03	0.01	0.01
Na ₂ O	0.00	0.00	0.00	0.01	0.02	0.03	0.01	0.01	0.01	0.04	0.02	0.02	0.01	0.02
NiO	0.42	0.44	0.48	0.00	0.01	0.01	0.00	0.02	0.01	0.00	0.01	0.01	0.02	0.00
Total	100.22	99.62	99.55	96.66	97.44	96.97	97.53	96.34	95.57	96.46	95.17	95.08	94.91	95.94
Fo	90.93	90.68	91.08	88.68	88.57	88.78	88.20	88.79	88.46	88.83	88.41	88.54	88.26	88.44

all clinopyroxenes differ from residual abyssal peridotite clinopyroxenes (Seyler et al., 2003) except for two samples in augite composition (Figure 6e). Moreover, the Al₂O₃ and Cr₂O₃ contents of most clinopyroxenes in the clinopyroxenite show close distribution with secondary clinopyroxene (compared with Nozaka (2005), Seyler et al. (2007), and Wojtulek et al. (2016)). However, the harzburgitic clinopyroxenes exhibit modified characters (Figure 6e).

The rare earth element (REE) contents of clinopyroxenes in the clinopyroxenite samples have low concentrations (Table 3). The chondrite-normalized REE patterns of clinopyroxenes in the clinopyroxenite exhibit a flat-shape pattern from MREE to HREE and depletion in LREE (Figure 7).

4.2.4. Spinel

Analyzed chromian spinel compositions from the clinopyroxenites and host harzburgites are presented in Table 4. The variations in cationic ratios and oxides such as Mg-number and Cr-number along with the Fe³⁺-number and TiO₂ contents of the chromian spinels from the Ulaş ultramafic rocks are illustrated in Figures 8–11. In general, spinels show slightly higher Cr-number in the clinopyroxenites than that of the harzburgites, ranging between 0.57 and 0.65 and 0.46 and 0.65, respectively. In contrast, the Mg-number of chromian spinels decreases towards the clinopyroxenite (0.30–0.34) from the harzburgites (0.52–0.70). In the Mg-number vs. Cr-number diagram, chromian spinels in the harzburgites and the clinopyroxenites are clustered in different fields and

Table 2. Representative major element analyses (in wt. %) of pyroxenes from the Ulaş ultramafics (Opx: orthopyroxene; Cpx: clinopyroxene).

Sample	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.
SiO ₂	57.56	55.84	57.23	56.36	56.66	56.72	57.58	56.89	56.46	56.96	56.37	56.57	56.66	56.84
TiO ₂	0.04	0.05	0.03	0.01	0.00	0.03	0.01	0.07	0.01	0.06	0.03	0.00	0.05	0.06
Al ₂ O ₃	1.92	1.52	1.79	1.78	1.94	1.81	1.76	1.58	1.81	1.56	1.84	1.94	2.01	0.79
Cr ₂ O ₃	0.57	0.64	0.44	0.99	0.36	0.34	0.55	0.40	0.54	0.71	0.63	0.63	0.72	0.68
FeO	6.36	6.42	6.11	6.31	6.36	6.41	6.46	6.45	6.84	6.85	7.12	7.26	7.02	7.13
MnO	0.13	0.15	0.17	0.21	0.18	0.16	0.17	0.17	0.15	0.16	0.14	0.19	0.20	0.15
MgO	32.96	34.64	33.54	33.59	33.68	33.56	32.48	33.62	33.44	32.89	32.41	33.01	32.83	32.78
CaO	0.48	0.77	0.64	0.76	0.64	0.71	0.88	0.76	0.66	1.10	0.73	0.65	0.59	0.78
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.02	100.03	99.95	100.01	99.82	99.74	99.89	99.94	99.91	100.29	99.27	100.25	100.08	99.21
Mg #	0.90	0.94	0.91	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.89	0.89	0.89	0.89
Sample	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.	Opx.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.
SiO ₂	57.13	56.88	57.23	56.81	56.37	56.34	56.84	56.21	56.53	56.62	57.74	56.32	56.64	55.93
TiO ₂	0.01	0.00	0.00	0.01	0.02	0.01	0.06	0.05	0.03	0.07	0.00	0.01	0.03	0.06
Al ₂ O ₃	1.82	2.03	1.95	2.26	1.22	2.04	1.85	1.96	1.55	1.51	1.35	1.44	1.52	1.19
Cr ₂ O ₃	0.79	0.74	0.68	0.98	0.74	0.66	0.57	0.69	0.32	0.44	0.53	0.31	0.57	0.85
FeO	6.65	6.47	6.02	6.27	6.43	5.43	5.81	6.03	5.92	5.77	5.87	6.58	5.69	5.99
MnO	0.14	0.14	0.17	0.12	0.14	0.16	0.16	0.12	0.15	0.10	0.19	0.14	0.16	0.18
MgO	32.73	32.59	32.86	32.68	33.62	34.00	33.28	33.74	33.24	34.56	33.15	34.52	33.08	34.55
CaO	0.79	0.88	0.64	0.85	0.83	0.67	1.16	0.89	0.52	0.87	0.69	0.61	0.78	0.84
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.06	99.73	99.55	99.98	99.37	99.31	99.73	99.69	98.26	99.94	99.52	99.93	98.47	99.59
Mg #	0.90	0.90	0.91	0.90	0.91	0.92	0.91	0.91	0.91	0.93	0.91	0.93	0.91	0.94

Table 2. (Continued).

Sample	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.
SiO ₂	53.83	54.26	53.81	54.31	53.48	53.25	55.44	55.63	55.58	55.61	54.80	55.68	53.79	53.24
TiO ₂	0.02	0.06	0.07	0.03	0.06	0.09	0.00	0.01	0.04	0.04	0.05	0.01	0.15	0.12
Al ₂ O ₃	1.77	1.41	1.96	1.68	2.38	2.45	0.02	0.03	0.04	0.85	1.30	0.77	5.88	5.88
Cr ₂ O ₃	0.94	0.99	1.21	0.94	1.12	0.76	0.00	0.00	0.01	0.29	0.59	0.18	1.08	1.12
FeO	2.16	2.15	2.22	2.26	1.90	1.86	0.07	0.06	0.10	2.46	2.59	2.49	3.48	3.68
MnO	0.06	0.13	0.11	0.14	0.11	0.10	0.00	0.03	0.00	0.10	0.06	0.07	0.06	0.09
MgO	17.65	17.15	16.22	17.47	16.77	17.48	18.43	18.60	18.63	18.32	18.25	18.50	21.45	21.18
CaO	23.61	23.89	24.37	23.19	23.41	23.98	25.87	25.70	25.73	24.69	23.87	24.12	12.97	13.07
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.04	0.02	0.02	0.05	0.07	0.08	0.73	0.75
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.04	100.04	99.97	100.02	99.23	99.97	99.88	100.08	100.16	102.44	101.61	101.92	99.63	99.20
Mg #	0.94	0.93	0.93	0.93	0.94	0.95	1.00	1.00	1.00	0.93	0.93	0.93	0.92	0.91
Sample	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.
Rock type	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.
SiO ₂	54.45	54.51	54.75	54.60	54.59	55.14	57.97	54.69	54.71	55.06	54.89	54.52	54.64	55.02
Al ₂ O ₃	0.03	0.04	0.10	0.03	0.07	0.06	0.03	0.05	0.01	0.10	0.07	0.03	0.04	0.03
TiO ₂	1.29	1.30	1.32	1.30	1.26	1.15	1.24	1.33	1.30	1.27	1.19	0.81	1.32	1.28
Cr ₂ O ₃	0.58	0.57	0.58	0.55	0.56	0.57	0.40	0.55	0.61	0.49	0.41	0.21	0.60	0.60
FeO	2.52	2.65	2.52	2.54	2.61	2.33	7.30	2.65	2.66	2.51	2.52	2.39	2.62	2.61
MnO	0.15	0.08	0.10	0.08	0.09	0.10	0.15	0.10	0.11	0.09	0.10	0.10	0.11	0.09
MgO	18.58	18.31	18.25	18.10	18.23	18.39	34.83	18.39	18.12	18.30	18.20	18.14	18.26	18.31
CaO	22.82	23.99	23.84	23.95	23.83	24.50	0.75	23.65	23.85	24.26	24.23	24.31	23.87	23.69
Na ₂ O	0.08	0.09	0.06	0.08	0.09	0.09	0.02	0.07	0.09	0.07	0.09	0.06	0.11	0.09
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.51	101.55	101.59	101.27	101.33	102.34	102.74	101.50	101.48	102.17	101.75	100.57	101.62	101.77
Mg #	0.93	0.93	0.93	0.93	0.93	0.93	0.90	0.93	0.92	0.93	0.93	0.93	0.93	0.93

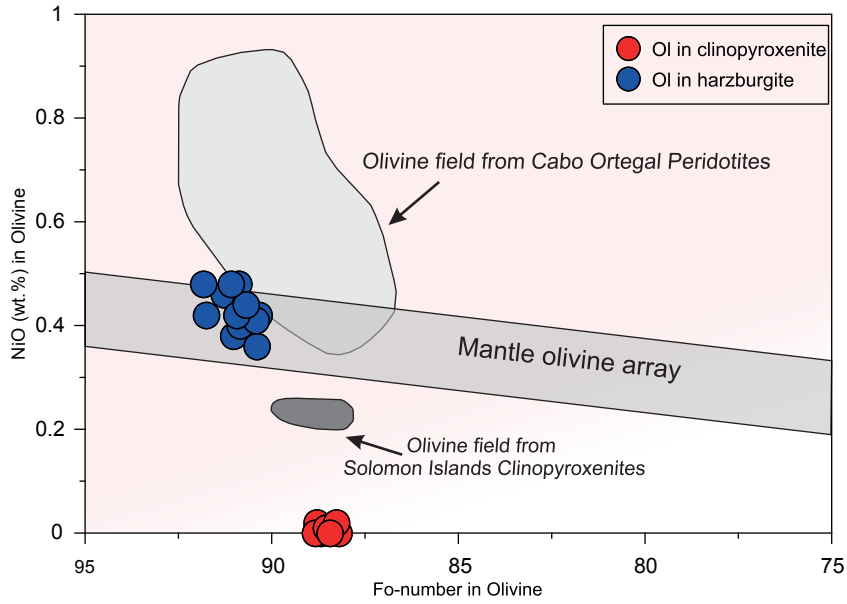


Figure 5. NiO vs. Fo-number diagram for olivine from the Ulaş (Sivas) ultramafic rocks (olivine field of the clinopyroxenite from Solomon Islands is taken from Berly et al. (2006), peridotitic olivine field of the Cabo Ortegal is taken from Girardeau and Ibarguchi (1991), and mantle olivine array from Takahashi et al. (1987)).

Table 3. Representative rare earth element analysis (in ppm) of clinopyroxenes in the pyroxenites from the Ulaş ultramafics.

Sample	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.	Cpx.
Rock type	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.
La	0.012	0.006	0.018	0.007	0.009	0.011	0.007	0.007	0.007
Ce	0.029	0.030	0.033	0.029	0.034	0.029	0.024	0.027	0.028
Pr	0.008	0.004	0.008	0.008	0.009	0.007	0.006	0.006	0.006
Nd	0.074	0.068	0.072	0.074	0.059	0.079	0.075	0.065	0.042
Sm	0.070	0.055	0.050	0.067	0.079	0.058	0.038	0.057	0.061
Eu	0.024	0.029	0.034	0.028	0.025	0.033	0.026	0.028	0.027
Gd	0.109	0.142	0.114	0.120	0.127	0.138	0.115	0.134	0.133
Tb	0.027	0.030	0.029	0.030	0.027	0.025	0.025	0.030	0.033
Dy	0.211	0.201	0.235	0.195	0.259	0.182	0.202	0.224	0.196
Ho	0.050	0.048	0.046	0.055	0.052	0.050	0.054	0.054	0.050
Er	0.146	0.142	0.145	0.165	0.154	0.150	0.149	0.120	0.138
Tm	0.019	0.022	0.019	0.022	0.024	0.022	0.022	0.018	0.024
Yb	0.166	0.135	0.129	0.113	0.174	0.122	0.141	0.154	0.150
Lu	0.020	0.019	0.019	0.020	0.023	0.021	0.017	0.021	0.019
Sr	4.080	3.530	4.010	3.480	4.040	3.830	3.430	3.310	3.440
Zr	0.207	1.117	0.099	0.113	0.116	0.110	0.113	0.110	0.102
Zr/Sm	2.957	20.309	1.980	1.687	1.468	1.897	2.974	1.930	1.672
Sr/Nd	55.135	51.912	55.694	6.486	68.475	48.481	45.733	50.923	81.905

Table 4. Representative major element analysis (in wt. %) of spinels from the Ulaş ultramafics (Spl: spinel).

Sample	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.
SiO ₂	0.01	0.02	0.00	0.00	0.04	0.00	0.00	0.03	0.02	0.00	0.00	0.00	0.02	0.00
TiO ₂	0.01	0.01	0.03	0.04	0.03	0.07	0.09	0.00	0.00	0.03	0.06	0.02	0.04	0.03
Al ₂ O ₃	23.69	26.71	26.00	27.38	23.45	26.79	28.21	27.05	28.89	31.06	27.23	29.34	22.64	18.89
Cr ₂ O ₃	41.35	41.48	43.08	41.93	41.01	41.12	41.18	42.92	40.87	40.52	40.57	39.74	44.57	46.00
FeO	18.91	16.32	17.38	17.04	17.02	16.67	16.50	17.05	18.64	17.19	18.31	19.47	20.59	20.54
MnO	0.33	0.27	0.31	0.31	0.22	0.28	0.29	0.30	0.33	0.27	0.27	0.29	0.24	0.18
MgO	12.74	14.47	13.24	12.55	13.21	15.84	13.69	13.35	11.96	11.64	14.00	14.43	11.96	15.25
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	97.04	99.28	100.04	99.25	94.98	100.77	99.96	100.70	100.71	100.71	100.44	103.29	100.06	100.89
Fe ⁺³	0.79	0.52	0.34	0.07	0.67	0.89	0.23	0.23	0.08	-0.38	0.67	0.77	0.69	1.73
Cr #	0.54	0.51	0.53	0.51	0.54	0.51	0.49	0.52	0.49	0.47	0.50	0.48	0.57	0.62
Mg #	0.60	0.65	0.60	0.57	0.63	0.70	0.61	0.60	0.54	0.52	0.63	0.62	0.55	0.69
Sample	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.	Spl.
Rock type	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Hrz.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.	Pyrx.
SiO ₂	0.06	0.02	0.03	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TiO ₂	0.04	0.07	0.05	0.06	0.03	0.04	0.05	0.21	0.12	0.16	0.14	0.09	0.19	0.22
Al ₂ O ₃	22.40	22.18	22.06	20.37	21.41	27.00	26.32	26.55	19.86	18.29	17.90	20.01	16.04	19.31
Cr ₂ O ₃	44.59	45.15	44.97	37.96	45.48	40.81	40.81	33.91	39.39	43.50	41.01	40.27	42.89	39.52
FeO	20.04	20.65	20.14	21.91	20.36	21.42	20.71	22.76	34.76	33.16	33.76	31.49	33.39	34.63
MnO	0.23	0.23	0.31	0.35	0.29	0.24	0.18	0.37	0.50	0.63	0.45	0.68	0.55	0.61
MgO	13.38	12.01	15.36	11.93	12.63	11.47	11.74	15.31	6.74	6.27	6.32	7.01	6.98	5.89
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.14	0.07	0.17	0.08	0.18	0.11
Total	100.74	100.31	102.92	92.58	100.20	100.98	99.83	99.11	101.51	102.08	99.75	99.63	100.22	100.29
Fe ⁺³	0.96	0.71	1.43	1.65	0.90	0.50	0.52	2.04	2.08	1.62	1.96	1.65	2.14	1.91
Cr #	0.57	0.58	0.58	0.56	0.59	0.50	0.51	0.46	0.57	0.61	0.61	0.57	0.64	0.58
Mg #	0.61	0.55	0.68	0.59	0.58	0.52	0.54	0.68	0.32	0.30	0.31	0.34	0.34	0.29

compared with the known SSZ peridotites and clinopyroxenites (Figure 9). The TiO₂ contents of spinels are in a narrow range between 0.00 and 0.21 (wt. %) in the harzburgites and 0.09 and 0.22 (wt. %) in the clinopyroxenites. Additionally, there is a slight positive correlation between TiO₂ contents and Cr-number of spinels from the harzburgites towards the clinopyroxenites (Figure 10). In general, the Fe³⁺-number of spinels in the clinopyroxenite is higher than that from the host harzburgites (Figure 11).

5. Discussion

5.1. Origin of the clinopyroxenite dykes

Mantle pyroxenites are generally composed of all pyroxenite types, such as clinopyroxenite, orthopyroxenite, and websterite.

However, only the clinopyroxenites were found in the harzburgitic host rock of this investigation. The current chemical data from the main minerals (ol, opx, cpx, and Cr-spinel) in the studied clinopyroxenite dykes and host harzburgites revealed differences to the subduction-related origin. Furthermore, the presence of spinel, together with absence of plagioclase and garnet, in the pyroxenites and host rocks confined the depth of the pyroxenite formation to be within the spinel–peridotite stability field (Berly et al., 2006). Field studies showed that the clinopyroxenites occur mainly as dykes within the host harzburgites (Figure 3). This revealed that each was formed in an SSZ mantle wedge, supported by evidence of infrequent orthopyroxene crystals of mantle affinity in the studied clinopyroxenites.

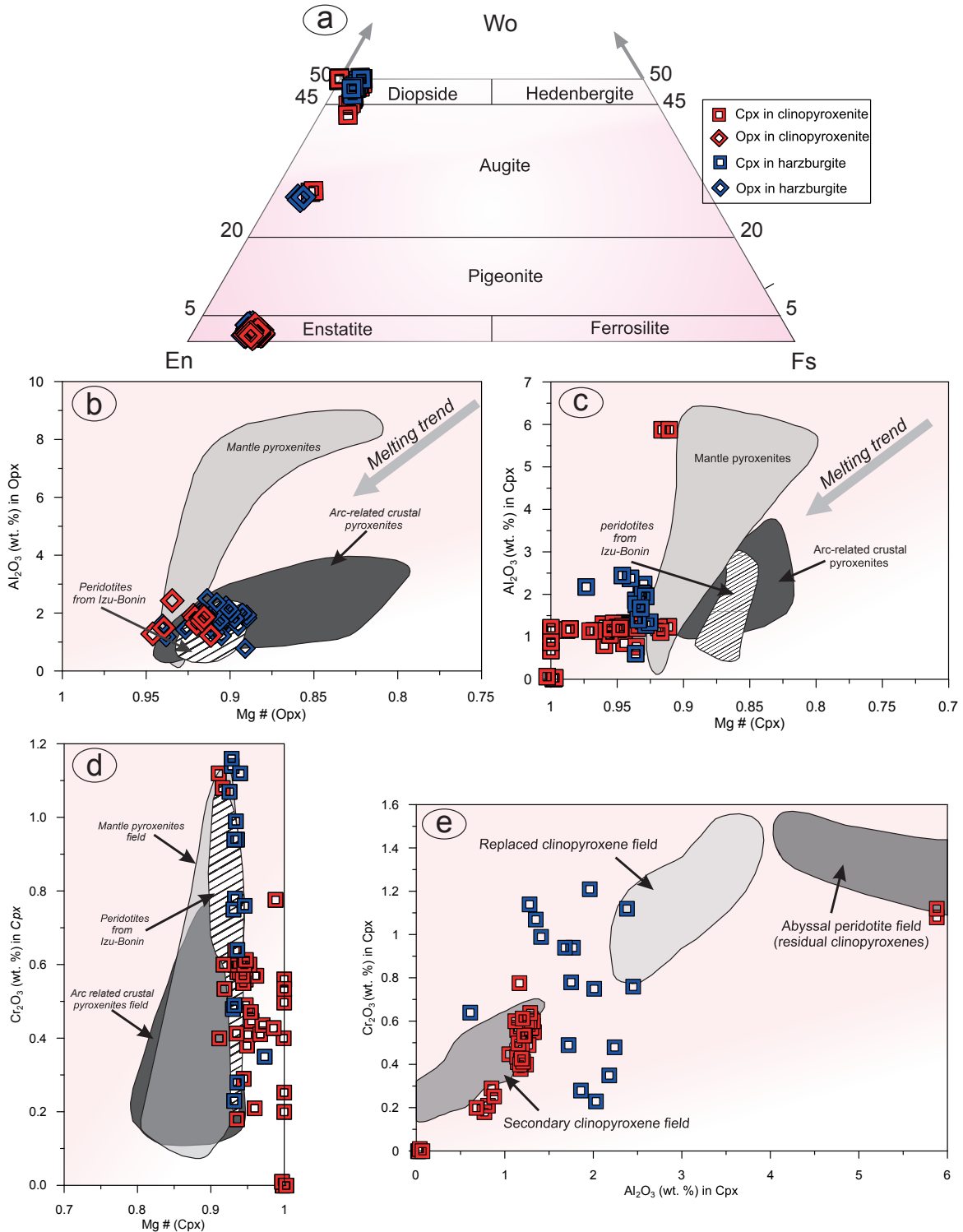


Figure 6. (a) Pyroxene nomenclature diagram for the pyroxenes from the studied ultramafic rocks (after Morimoto et al. (1988)); Al_2O_3 vs. Mg-number diagram for orthopyroxene (b) and for clinopyroxene (c); (d) Cr_2O_3 vs. Mg-number diagram for clinopyroxene (references of the fields after Berly et al. (2006)); (e) Cr_2O_3 vs. Al_2O_3 (wt.%) diagram for clinopyroxenes (fields of residual clinopyroxenes after Seyler et al. (2003), field of replaced clinopyroxenes after Seyler et al. (2007), field of secondary clinopyroxenes after Nozaka (2005)).

In a plot of Fo-number vs. NiO, olivines from the harzburgites showed higher NiO content than those of the clinopyroxenites and a plot within the “mantle olivine array” indicated a residual source for the harzburgites, comparable with peridotitic olivines from the Cabo Ortegal peridotites (Girardeau and Ibarguchi, 1991). In contrast, olivines from the clinopyroxenites had very low NiO content and also showed a different distribution to the other SSZ clinopyroxenites, especially when compared to the Solomon Islands clinopyroxenites (Berly et al., 2006), as in Figure 5.

The major element contents of ortho-clinopyroxenes from the clinopyroxenite and host harzburgite were evaluated together. In a plot of Mg-number vs. Al_2O_3 (Figure 6b), orthopyroxenes showed Al_2O_3 depletion towards the clinopyroxenites from the harzburgites, indicating Al_2O_3 extraction during partial melting compatible with the melting trend (Page et al., 2008) and orthopyroxene fractionation trend (Varfalvy et al., 1996, 1997). Moreover, the host harzburgites resembled the fore-arc (SSZ) peridotites and the clinopyroxenites that show an affinity to the mantle pyroxenites rather than the arc-related crustal pyroxenites by orthopyroxene chemistry (Figure 6b). The high Mg-number and low Al_2O_3 contents of clinopyroxenes exhibit a trend as a result of partial melting (Figure 6c). Clinopyroxenes from the host harzburgites plot within the fore-arc peridotite field suggested by Page et al. (2008) prominently differ from the Izo-Bonin fore-arc peridotites proposed by Parkinson and Pearce (1998). Clinopyroxenes from the clinopyroxenites have higher Mg-numbers than the host harzburgites, a few of which were similar to the mantle pyroxenites (Figure 6c). However, in the Cr_2O_3 content vs. Mg-number diagram, most clinopyroxenes from the clinopyroxenites plot were within the mantle pyroxenite field. In this same diagram, clinopyroxenes from the host harzburgites showed a very similar distribution to the Izo-Bonin fore-arc peridotites (Figure 6d). Generally, the high Mg-number of clinopyroxenes in the clinopyroxenites is associated with the high degree of partial melting of the host harzburgites. Furthermore, the Cr_2O_3 and Al_2O_3 contents of clinopyroxenes in the clinopyroxenite and the host harzburgites saw major elements decrease and exhibit a positive correlation towards the harzburgites from the clinopyroxenites (Figure 6e). Conversely, two clinopyroxene samples with high Al_2O_3 contents showed abyssal character, indicative of residual clinopyroxene in the clinopyroxenites. Overall, clinopyroxenes from the clinopyroxenites with lower Cr_2O_3 and Al_2O_3 content than modified clinopyroxenes in the host harzburgites displayed secondary clinopyroxene characteristics, as suggested by

Nozaka (2005) and Wojtulek et al. (2016); see Figure 6e.

Trace element contents of clinopyroxenes are crucial in defining the origin of mantle pyroxenites. Many studies show that clinopyroxene is the main mineral for REE, and its pattern reflects the whole rock in the absence of garnet (Garrido et al., 2000). In the presence of garnet, the REE would be distributed between garnet and pyroxene. The REE pattern of clinopyroxenes from the studied clinopyroxenites showed similar patterns to that of clinopyroxene from the SSZ and fore-arc ophiolite (Figure 7). These were depleted in LREE to MREE and HREE, as compared to those of abyssal clinopyroxenes (Johnson et al., 1990), and instead exhibited a pattern closer that of SSZ clinopyroxenes (Bizimis et al., 2000) rather than fore-arc clinopyroxenes (Parkinson et al., 1992). Slight La enrichments of clinopyroxene in the clinopyroxenite support this similarity, as in Figure 7. The slight upward inflection of La could also be clarified by a melt percolation event.

The high Cr-number of spinels indicated that the peridotites formed within a supra-subduction setting (Bonatti and Michael, 1989; Wang et al., 2001). The Fo content of olivines and the Cr-number of spinels are essential indicators of the partial melting degree for mantle peridotite and are used to differentiate tectonic environments (Dick and Bullen, 1984; Arai, 1994; Pearce et al., 2000; Tamura and Arai, 2006). Under this criteria, spinels from the clinopyroxenites and the host harzburgites show similar formation to the subduction-related mantle rocks (Figure 8). Additionally, mid-ocean ridge (MOR) or similar zone (fracture) peridotites have a Cr-number lower than 0.6, whereas SSZ peridotites have Cr-numbers up to 0.8 (Dick and Bullen, 1984; Arai, 1994; Kelemen et al., 1995; Gaetani and Grove, 1998; Choi et al., 2008; Kaczmarek et al., 2015). A typical subduction-related harzburgite, considered to have formed by 20%–25% partial melting, contains olivine with higher forsterite content (Parkinson and Pearce, 1998). The high Fo-number of olivines and relatively high Cr-number of spinels support the depletion trend of the studied harzburgites as a product of high-degree partial melting (25%) and denote similarity with the fore-arc (SSZ) and less so with the abyssal peridotites (Figure 8). However, relatively high Cr-numbers of spinels and low Fo-numbers of olivines from the clinopyroxenites plot outside of the olivine spinel mantle array (OSMA) indicated an eventual fractional crystallization from a magma with MORB-like affinity or initiation of boninitic affinity in the SSZ (Figure 8). The low Mg-number of spinels in the clinopyroxenite shows closer distribution to the SSZ clinopyroxenite field (Figure 9). An important feature to note is that the Mg-number of MORB-spinels

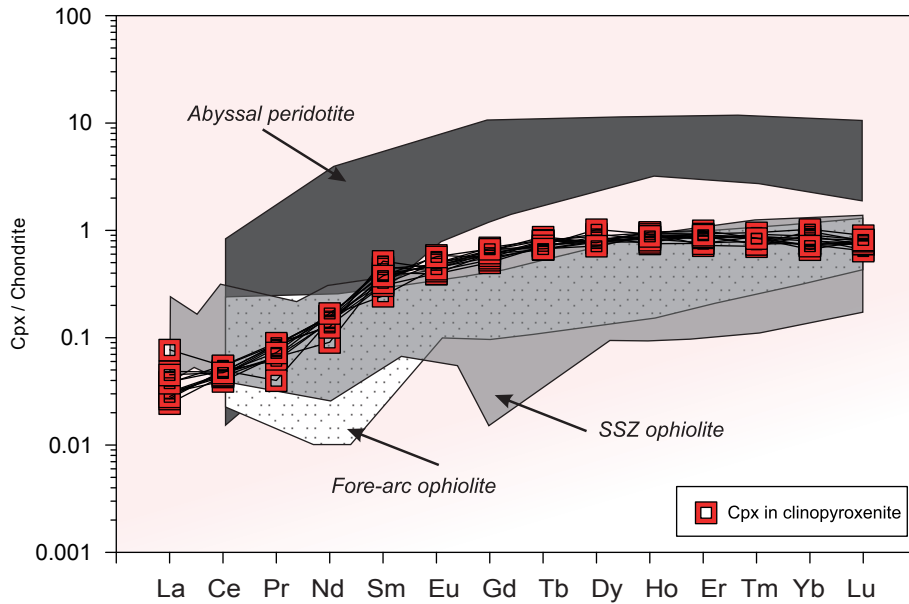


Figure 7. Chondrite-normalized REE patterns for clinopyroxenes from the clinopyroxenites of Ulaş (Sivas) ultramafic rocks. For comparison, a range of clinopyroxene REE pattern from abyssal peridotite (Johnson et al., 1990), from suprasubduction peridotite (Bizimis et al., 2000) and from fore-arc peridotite (Parkinson et al., 1992) is shown. Normalizing values of REE for chondrite are taken from Sun and McDonough (1989).

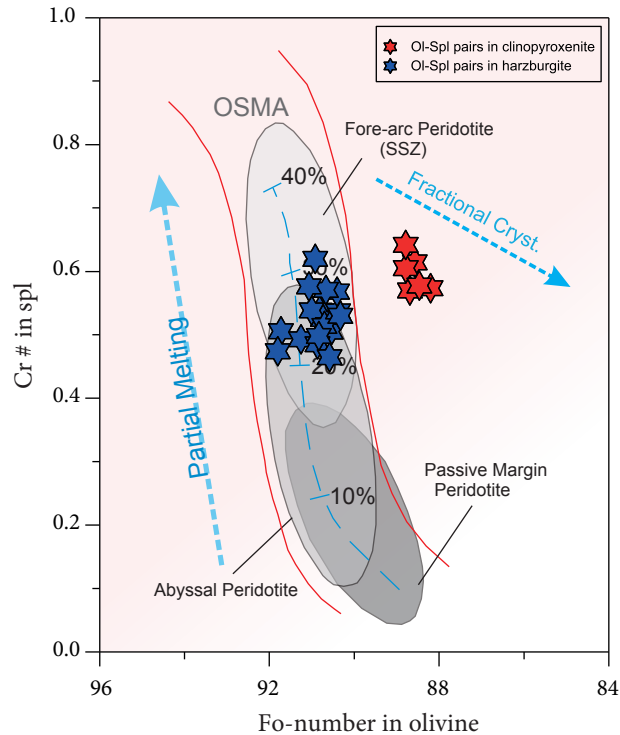


Figure 8. Compositional relationship between forsterite content of olivine and Cr-number of chromian spinels of the studied Ulaş ultramafic rocks on the olivine-spinel mantle array (OSMA) diagram of Arai (1994) (abyssal peridotite field from Dick and Bullen (1984), forearc peridotite field from Ishii et al. (1992), Parkinson and Pearce (1998), and Pearce et al. (2000).

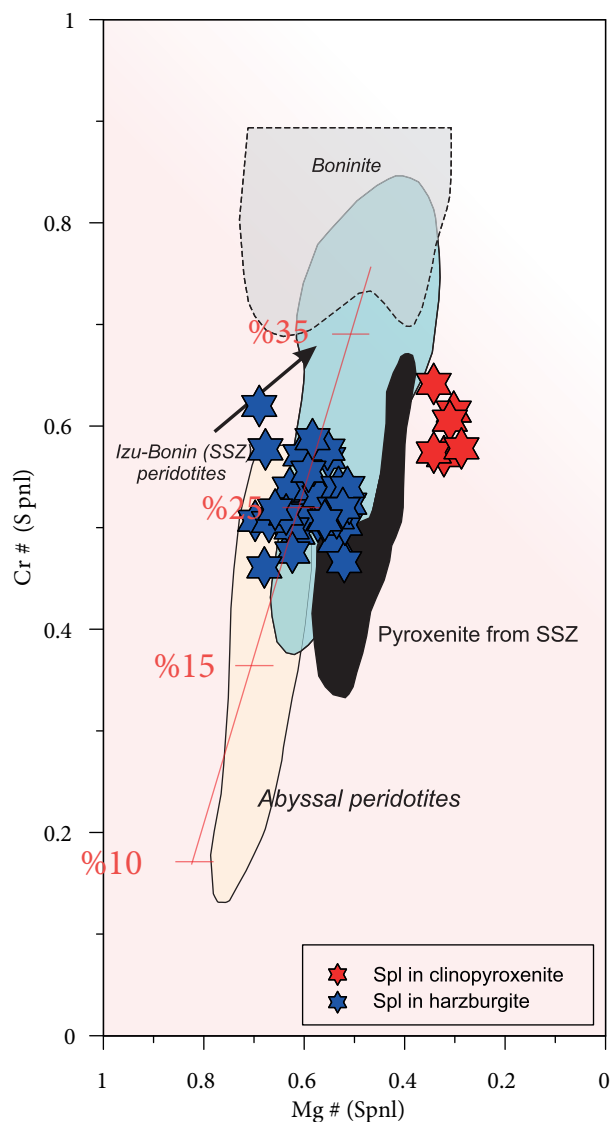


Figure 9. Relationship between Cr-number and Mg-number in chromian spinel from the studied Ulaş ultramafic rocks (fields for abyssal peridotites, fore-arc peridotites, and boninites are taken from Tamura and Arai (2006). SSZ pyroxenite field is taken from Berly et al. (2006). Partial melting trend is taken from Arai (1992)).

is higher when compared to the cumulates of the Alpine-type ophiolites and the mantle peridotites (Lee, 1999). Hence, the high Cr-number and low Mg-number of the spinels in the studied pyroxenites indicate a subduction-related origin, and an extensive re-equilibration between the pyroxenites and the mantle peridotites.

5.2. Effects of the melt-peridotite interactions on the pyroxenite formation

The melt-rock interaction is a key factor in the compositional evolution of both melt and host rock in the

mantle (Kelemen, 1990; Edwards, 1995; Varfalvy et al., 1996; Kelemen et al., 1997; Garrido and Bodinier, 1999; Bodinier et al., 2004; Batanova et al., 2011). The Al and Ti contents of chromian spinels are also very useful metrics in order to understand the characteristics of the parental melts of the spinels and the tectonic environment in which these melts were generated (Zhou et al., 1996; Melcher et al., 1997; Uysal et al., 2007; Rollinson, 2008; Page and Barnes, 2009; Gonzalez-Jimenez et al., 2011; Zaccarini et al., 2011; Uysal et al., 2012).

In the present study, spinel mineral chemistry was used to investigate the reactions between the melts forming the clinopyroxenite dykes. The significant reduction of olivine compositions (Fo and NiO) in the clinopyroxenite can be explained by a compositional change during fractional crystallization of magma or melt interaction with the host rocks over the course of dyke growth. Laukert et al. (2014) proposed that the presence of olivine in the pyroxenite would demonstrate a relatively low melt/rock ratio, whereas the absence of olivine indicates the final reaction components at high melt/rock ratios.

The TiO_2 content and other minor elements of spinels are sensitive monitors of melt-rock interaction and denote larger variations than the major elements during the partial melting of peridotites and fractional crystallization processes (Arai, 1992; Dare et al., 2009). Additionally, the TiO_2 content of spinels shows initial records of the impregnating melts as opposed to the primary structure of the host rocks (Cannat et al., 1990), because the impregnating melts are generally richer in concentrations than the nonreacted peridotites (Pearce et al., 2000; Whattam et al., 2011). In general, low TiO_2 content (<0.2 wt.%) spinels are considered to be plagioclase-free, depleted peridotites that do not react with MORB-like melts at low-P (Dick, 1989; Seyler and Bonatti, 1997). Spinels in the host harzburgites surrounding the dykes also have TiO_2 content (>0.1 wt. %) that is higher than the background harzburgite. This result could be explained by the reaction between the intruding melts and the host rocks. Plots of spinel Cr-number vs. spinel TiO_2 (Figure 10) demonstrate that spinels from the host harzburgite underwent intensive melt-rock interactions with a Ti-rich melt, subsequent to having undergone 25% partial melting. Spinels from the clinopyroxenite were similar those of the host harzburgites and reflect a melt-rock reaction trend towards an SSZ reaction field more than a MORB reaction field (Figure 10).

The high Fe^{3+} -number of spinels indicated subduction-related melts, because of the presence of water associated with an SSZ setting, which promotes oxidizing conditions relative to MORB melts (Arai, 1992). The melt-rock

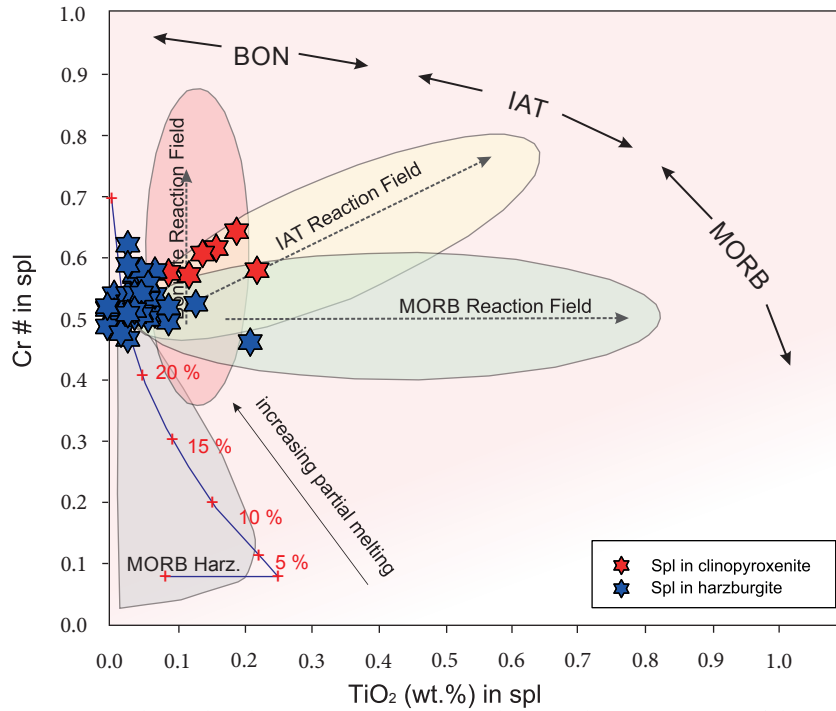


Figure 10. Relationships between TiO_2 contents and Cr-number atomic ratios of chromian spinels in the studied Ulaş ultramafic rocks (after Pearce et al. (2000)).

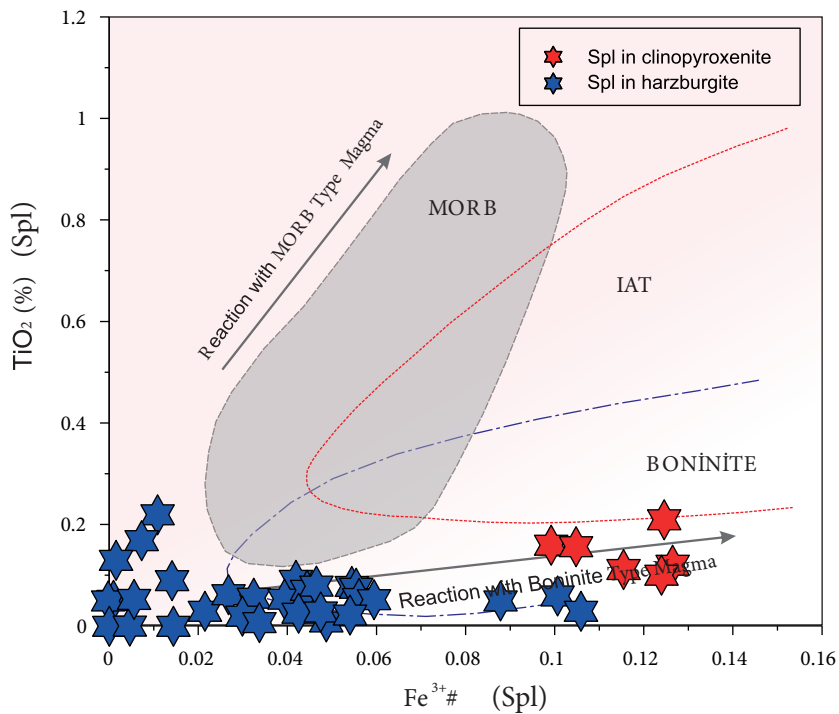


Figure 11. Relationships between TiO_2 contents and Fe^{3+} -number atomic ratios of chromian spinels in the studied Ulaş ultramafic rocks (after Pearce et al. (2000)).

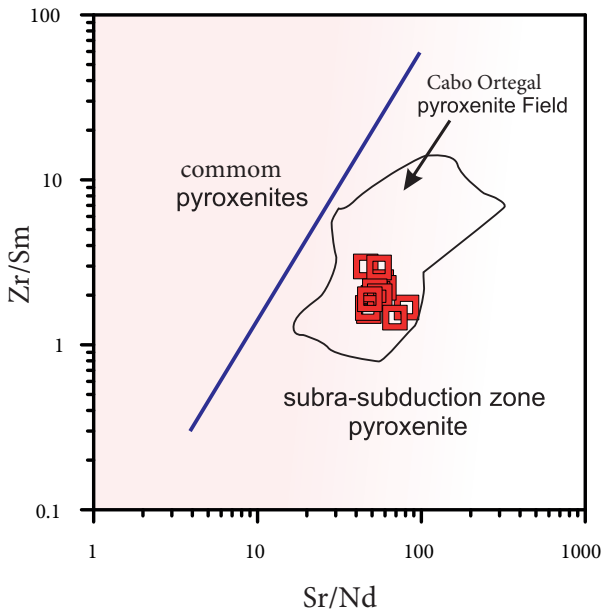


Figure 12. Sr/Nd vs. Zr/Sm ratios in clinopyroxenes from pyroxenite in the studied Ulaş ultramafic rocks (after Downes (2007)).

reaction effect in MOR and SSZ settings is shown through the Fe^{3+} -number and TiO_2 (wt.%) contents of spinels from the clinopyroxenites and host harzburgites in Figure 8. Spinel from the SSZ harzburgites have higher Fe^{3+} -numbers (0.06) than abyssal harzburgites (0.03) (Dare et al., 2009). Although the obtained data of spinels in the harzburgites showed a wide range of Fe^{3+} -numbers and indicated both SSZ and abyssal characteristics, the low Fe^{3+} -numbers of these spinels show that reacting melts were more SSZ-like than MORB-like (Figure 11). Spinel in the clinopyroxenite that had higher Fe^{3+} -numbers and low TiO_2 (wt. %) content in a narrow range indicate formation from SSZ-like melt interactions. This is compatible with the isolated dyke geochemistry of Divriği ophiolite (Parlak et al., 2006), where the diabasic isolated dykes cut the mantle tectonites, and were ascribed to subsequent magmatism fed by melts that occurred within an asthenospheric window caused by slab break-off, shortly before the settlement of the ophiolite onto the Tauride Platform. Hence, the clinopyroxenite dykes within the host harzburgite are thought to have crystallized from this SSZ-type magma with similar conditions.

5.3. Comparison with the arc-related (SSZ) and abyssal pyroxenites

Occurrences of the pyroxenites remain controversial in different tectonic environments. The pyroxenite dykes can be formed through various mechanisms, depending on the tectonic setting or formation and a detailed study with

field evidence, petrography, and geochemistry is required in every individual case in order to understand the origins thoroughly.

Within this context, the pyroxenites and associated peridotites were discussed in many studies on the arc-related settings, including the Beni Bousera massif, Morocco (Pearson et al., 1993), the Alaskan complexes (DeBari and Coleman, 1989), the North America–Canada ultramafics (Snook et al., 1981), the Europe and northwestern Africa ultramafics (Downes, 2007), the pyroxenites from Cabo Ortegal, Spain (Girardeau and Ibarra, 1991; Santos et al., 2002), the Andong Ultramafic Complex in Korea (Whattam et al., 2011), the Ronda peridotite massif in Spain (Marchesi et al., 2012), the intraoceanic arc-related cumulates in southern New Zealand (Spandler et al., 2003), the Solomon Islands pyroxenites (Berly et al., 2006), pyroxenite dykes in the North Qaidam Peridotites, China (Xiong et al., 2014), and the Marum ophiolite complex, Papua New Guinea (Kaczmarek et al., 2015).

In contrast, the investigations on abyssal pyroxenites are more limited (Dick et al., 1984, 2010; Fujii, 1990; Kempton and Stephens, 1997; Hellebrand et al., 2005; Arai and Takemoto, 2007; Dantas et al., 2007; Warren et al., 2009; Seyler et al., 2011; Laukert et al., 2014).

In general, when all of the studies are examined in terms of mineral chemistry, it can be seen that the arc-related and abyssal pyroxenites differ in terms of formation. The low Cr-number (20–30) of spinels and relatively low Mg-number (90) of clinopyroxenes are characteristic properties of abyssal pyroxenites and the presence of websterites at lower melt/peridotite ratios, signifying the products of melt–rock interaction. However, the high Cr-number of spinels and high Mg-number of clinopyroxene are typical of arc-related (SSZ) pyroxenites. The clinopyroxenites can be considered melt-dominated systems with high melt/peridotite ratios (Van Aken, 2008).

In the present study, the mineral chemistry of the clinopyroxenites within the host harzburgites indicates that these pyroxenites have a similar affinity to the SSZ pyroxenites with relatively high Cr-numbers of spinels and high Mg-numbers of clinopyroxenes (Figure 6). Furthermore, the REE concentrations of clinopyroxenes from the clinopyroxenites were consistent with the SSZ pyroxenites (Figure 7). The high Sr/Nd and low Zr/Sm ratios in these clinopyroxenes suggest crystallization from LREE-enriched melts of a subduction zone (Figure 12), and these ratios show a similar distribution to the Cabo Ortegal pyroxenites reported by Downes (2007).

6. Concluding remarks

Field observations and petrographic properties show that clinopyroxenite dykes occur within the harzburgite from Ulaş District, as part of the Divriği ophiolite. A detailed mineral geochemical investigation of the clinopyroxenites

and the host peridotites was evaluated using petrological interpretation of different diagrams presenting the obtained analytical data. The sharp margins with the host harzburgites together with the mineral geochemical evidence suggested that the pyroxenite dykes were generated through a series of petrological processes, including fractional crystallization and melt–rock reactions. The dykes may have been formed by a focused flow of melts during the migration towards the crust in the mantle section of this ophiolite. The results have shown that these clinopyroxenites formed as a product of fractional crystallization from magma of relatively high-grade partial melting of the host harzburgite, which reacted with the SSZ-type melts in the mantle wedge of a subduction zone. Consequently, the compositions of spinels and other

silicate minerals suggest that the clinopyroxenites were derived from a nonboninitic SSZ type-magma that resulted from slab breakoff. This situation could be explained by the close proximity of the Cr-numbers of spinels in both lithologies.

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