A polygenetic origin for the Sikhoran ultramafic-mafic complex in southwest Iran

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Abstract

The nature and geodynamic setting of the Sikhoran ultramafic–mafic complex, southwest Iran, is controversial, with multiple competing theories having been proposed in recent years. The Sikhoran ultramafic–mafic complex is located in the southern part of the Sanandaj-Sirjan metamorphic-magmatic zone (SSMMZ) and is comprised of porphyroclastic, transitional and layered ultramafic-mafic sequences of Carboniferous age. These units are cut by isotropic gabbros, diabasic and pegmatoid gabbroic dikes with ages ranging from Permian to the Cretaceous. These bodies also intrude into Upper Paleozoic metamorphic rocks, where they induced high-temperature contact metamorphism during the Late Carboniferous. Here we present new textural descriptions, mineral and whole-rock geochemical analyses, and the results of zircon U–Pb dating of magmatic mafic rocks and metamorphic host rocks from this region in order to reconstruct the petrogenesis of the Sikhoran ultramafic–mafic complex. Zircon U–Pb ages from layered gabbros indicate Late Carboniferous (320.8 ± 6.4 Ma) crystallization, which is similar to the age of the Upper Paleozoic host gneisses (334.6 ± 4.9 Ma). By contrast, isotropic gabbros that cross-cut the complex have a crystallization age of 178.3 ± 2.3 Ma, and anatectic quartz diorite/plagiogranite produced from partial melting of hosting amphibolite in contact with the gabbros yielded Late Triassic-Early Jurassic ages

(187.2 ± 2.6 Ma). These data suggest a polygenetic origin and a range of tectonic settings for various parts of the ultramafic-mafic complex. Geochemical modeling shows that mantle plume-related melting (~5–40%) of metasomatized mantle could have formed the ultramafic-mafic layered cumulate complex during the Late Carboniferous. A geochemical transition from island arc tholeiite (IAT) isotropic gabbros to cross-cutting enriched mid-ocean ridge basalts (E-MORB)–diabasic dikes occurred between the Late Triassic-Early Jurassic to the Late Cretaceous. The mantle-derived ultramafic parts of the Sikhoran ultramafic-mafic complex have similarities ~~(e.g., a foliation parallel contact between amphibolite and ultramafic rock, internal foliation of the larger ultramafic bodies, location of the ultramafic rocks along strike of the regional foliation and their high TiO_2 contents along with the Late Carboniferous ages)~~ with other known ultramafic-mafic mantle plumes (e.g., Tinaqulllo, Lherz, and Ronda), and its layered ultramafic-mafic segment has similarity with layered intrusions (e.g., Pulur complex in Turkey). The highest observed crystallization temperature in porphyroclastic, transitional and layered ultramafic mantle sections of ~1140 °C is considered as the mantle potential temperature, since the highest crystallization temperatures are observed in the most forsteritic olivine crystals. Based on amphibole chemistry, the crystallization conditions of transitional to layered ultramafic rocks at 900–1005 °C occurred at high H_2O contents in a melt between nickel–nickel oxide (NNO) −8.5 and NNO −9.0, which lies along the NNO^{-2} buffer. These results are consistent with a model in which the opening of the Zagros Neotethys Oceanic basin in the Esfandagheh region initiated in the Late Carboniferous, alongside the formation of an extensional environment that allowed ascent of a subcontinental mantle plume. Subduction initiation of oceanic lithosphere began during the Late Triassic-Early Jurassic and ended in the Late Cretaceous.

Keywords: Ultramafic–mafic complexes, mantle plume-type, Sikhoran, Iran

1. Introduction

Ultramafic–mafic complexes can be divided into four types: (i) ophiolitic types (Gass, 1968, 1989; Moores et al., 1968; Coleman, 1977; Nicolas 1989; Dilek et al., 2000; Flower and Dilek, 2003), (ii) Alpine types (Challis, 1965; Dick,1977; Obata,1980; Quic, 1988) or subcontinental mantle plumes (Choi et al., 2007; Lenza et al., 2010; Le Roux et al., 2007; Uzel et al., 2020), (iii) layered types (Cawthorn, 1996; Charlier et al., 2015; Eyuboglu et al., 2010; Tupuz et al., 2023), and (iv) Alaskan-type intrusions (Himmelberg and Loney, 1995, Ye et al., 2015; Khedr and Arai 2016a; Abdallah et al., 2019; Khedr et al., 2020). The

Alaskan-type and layered intrusion complexes have relatively similar lithologies, but the former shows concentrically zoned intrusions, with ultramafic rocks in the center and mafic rocks at the edges. There are also numerous differences between these types in terms of their geochemistry and type of associated mineralization. The type example of Alaskan-type ultramafic rocks occurs along a 560-km-long belt situated west of the Coast Plutonic Complex in southeastern Alaska. They are concentrically zoned bodies with a dunite core surrounded by pyroxenite shells, and are generally associated with gabbro intrusions (Taylor, 1967; Himmelberg et al., 1985; Patton et al., 1994). Classically, they are interpreted as representing fractionated ultramafic intrusions (Taylor, 1967). In contrast, Alpine-type ultramafic rocks are generally fault-bounded, internally deformed, and serpentinized. They are classically interpreted as being segments of oceanic crust and/or mantle plume that were tectonically emplaced onto the continents during ocean closure (Hall, 1987). A common feature of Alpine-type ultramafic rocks is their occurrence along tectonic zones, including faults, shear zones and terrane boundaries (i.e. suture zones) (Mezger 2000). In the Alps, such ultramafic rocks may be remnants of an ocean basin that collapsed during subsequent accretion in the Late Mesozoic (Nokleberg et al., 1985, Patton et al., 1994). As such, Alpine-type ultramafic rocks are emplaced as hot solid masses; they show tectonite fabrics, have minor variations in modal mineralogy and mineral composition (except spinel), and refractory major and trace element compositions. Recently, these rocks have been interpreted as subcontinental mantle plumes that ascended and were emplaced into the continental crust (Choi et al., 2007; Le Roux et al., 2007; Pirajno, 2007, 2022; Puchkov, 2009; Lenaz et al., 2010; Dannberg, 2016; Ernst et al., 2019; Uzel et al., 2020; Srivastava et al., 2022; Topuz et al., 2023).

Some studies have suggested that Tethyan ophiolites resulted from diapiric upwelling and partial melting of mantle material during an early, possibly tensile, stage of the Alpine orogenic cycle. In this case, the ultramafic complex would have formed towards the end of the Jurassic, and would be described as an Alpine type (Maxwell, 1969). Field data and geophysical measurements suggest that most “Alpine-type” ultramafic masses are rootless and allochthonous (Maxwell, 1969). Wyllie (1969) noted that “Alpine” ultramafic rocks can be interpreted to be mantle-derived in four scenarios: (i) formation from an ultrabasic liquid, with or without suspended crystals; (ii) formation from a basic liquid magma (derived by partial fusion of the mantle), with or without suspended crystals; (iii) as a peridotite mush with an interstitial basic liquid, or (iv) as “solid intrusions” or via mechanical emplacement of rigid oceanic crust. However, other studies suggest that large ultramafic bodies commonly develop a foliation and shear sense that is similarly oriented to those in the adjacent rock, suggesting an 'Alpine-type' emplacement (Mezger 2000). In such a model, ultramafic rocks can occur as isolated outcrops, without needing to be associated with ophiolite sequences (see Khedr and Arai, 2009, 2010, 2012 and Khedr et al., 2010). As such, these ultramafic-mafic complexes have recently be re-considered to have formed via mantle plumes that are associated with extensional tectonic zones during the early stages of continental rifting (Choi et al., 2007; Le Roux et al., 2007; Pirajno, 2007, 2022; Puchkov, 2009; Lenaz et al., 2010; Dannberg, 2016; Ernst et al., 2019; Uzel et al., 2020; Srivastava et al., 2022; Topuz et al., 2023).

The Soghan-Sikhoran district, Iran, located within the southern parts of the Sanandaj-Sirjan metamorphic-magmatic zone (SSMMZ), contains the Abdasht, Soghan and Sikhoran ultramafic-mafic complexes (Sabzehei, 1974; Ghasemi, 2000; Ghasemi et al., 1998, 2001, 2002; 2004; Ahmadipour, 2000; Ahmadipour et al., 2003; Peighambari et al. 2011, 2016; Behzadi and Shahabpour, 2011; Moghadam et al., 2012, 2017; Sepidbar et al., 2021; Alesaadi et al., 2022; Asadi et al., 2022, 2023) (Fig. 1A, B). Previous studies have suggested that these complexes formed due to mantle diapirism (Sabzehei, 1974; Ghasemi et al., 2002; Ahmadipour et al., 2003), although few constraints exist on their ages and very little work has been performed on their constituent mineral chemistry, such that petrogenetic models remain uncertain.

Here, we discuss different mantle sections from the southern Iranian ultramafic–mafic complex in terms of their age, mineral and whole-rock chemistry, parental melt composition, tectonic setting, and the nature of the peridotites (i.e., ophiolite vs. subcontinental ascent and mantle plume-related melting). Consequently, we describe and interpret the possible origin of ultramafic-mafic rocks in the Sikhoran region, their mode of emplacement into the host metamorphic rocks. We also consider the implications of their origin for the tectonic evolution of the Soghan-Sikhoran district, south of Iran in the Zagros Neotethys realm, with particular reference to the initiation of rifting since the Late Carboniferous and subduction from Late Triassic-Early Jurassic in the Zagros basin. We show that the ultramafic-mafic rocks of the Sikhoran district are a polygenetic ultramafic-mafic complex that was tectonically interleaved within pre-existing metasedimentary rocks during final closure of Esfandagheh Ocean, which led to underplating and accretion onto the overriding Iranian plate during the Late Cretaceous.

2. Geological setting

2.1. General aspects

The Zagros Orogenic Belt (ZOB) in Iran is oriented parallel to the Arabian–Iranian plate boundary and can be divided into five domains (Alavi, 2004; 2007): (1) the Zagros fold-and-thrust belt, which extends ~2000 km from southeastern Turkey, through northern Syria and northeastern Iraq, to western and southern Iran (Alavi, 1994); (2) the Outer Zagros Ophiolitic Belt (OZOB) (Kermanshah–Neyriz–Hajiabad ophiolites) and associated HP/LT rocks; (3) the SSMMZ, which extends as a NW–SE-trending belt ~1500 km in length and ~150–200 km in width across south Iran; (4) the Inner Zagros Ophiolitic Belt (IZOB), which extends for 500–600 km and includes, from NW to SE, the Nain, Dehshir, Shahr-e-Babak and Baft ophiolites; and (5) the Urumieh–Dokhtar magmatic arc, which is a 50–80 km wide belt containing magmatic rocks that formed by NE-dipping subduction of the Neo-Tethyan oceanic lithosphere beneath the Central Iran block during the Cenozoic. Numerous partially dismembered ophiolite bodies crop out over several hundreds of km^2, forming extended complexes that follow the ZOB in a thin zone stretching from northeast Turkey in the northwest to Makran in the southeast (Fig. 1).

The SSMMZ exposes various metamorphic and magmatic rocks belonging to the Late Neoproterozoic basement (~550–600 Ma, i.e., Moghadam et al., 2017; Davoudian et al., 2008; Hassanzadeh et al., 2008; Hassanzadeh and Wernicke, 2016; Nutman et al., 2014; Azizi et al. 2011), but also Middle-Late Paleozoic (Ghasemi et al., 2002, 2004; Ahmadipour et al., 2003; Moghadam et al., 2015, 2017; Saccani et al., 2013; Azizi et al. 2017) and Mesozoic (Ghasemi et al., 2002, 2004; Ahmadipour et al., 2003; Moghadam et al., 2017, Asadi et al., 2022; Zaeimnia et al., 2017). The Mesozoic metamorphic rocks mainly include (i) Middle Jurassic amphibolites, migmatites, various schists, phyllites and slates, known as Hamedan schists (Baharifar et al. 2004; Sepahi et al., 2018), and (ii) Upper Triassic-Lower Jurassic schists, amphibolites and marbles and Upper Cretaceous blueschists (Ghasemi et al., 2002, 2004; Ahmadipour et al., 2003; Angiboust et al., 2016 Saccani et al., 2022).

2.2. Abdasht-Soghan-Sikhoran-Haji-Abad complex

The Upper Paleozoic Abdasht-Soghan-Sikhoran complex is the best-preserved and largest ultramafic–mafic complex in southeast Iran (ca. 1000 km^2 in area) and occurs in the southern part of the SSMMZ (Fig. 1B). Abdasht and Soghan units are mainly composed of residual mantle lherzolite and harzburgite along with small amounts of layered ultramafic-mafic rocks (Ahmadipour, 2000; Ahmadipour et al., 2003), while the Sikhoran complex is mainly composed of porphyroclastic harzburgite and dunite, contains chromitite pods, and has a magmatic crustal section consisting of layered cumulative ultramafic-mafic sequences (cumulative dunite, chromitites, harzburgite, lherzolite, wehrlite, websterite, pyroxenite, olivine gabbro, troctolite, gabbronorite, ferrogabbro, diorite/quartz diorite and plagiogranite). Based on K–Ar geochronology, the Sikhoran complex has been previously interpreted as an asthenospheric mantle plume that ascended through a Triassic–Jurassic back-arc basin within the SSMMZ (Ghasemi, 2000; Ghasemi et al., 2002, 2004; Asadi et al., 2022); however, we now present a polygenetic origin for this complex based on new findings.

The Sikhoran complex is cross cut by microgabbroic/diabasic and pegmatoid dikes and a large isotropic gabbro (the Abshour gabbro) that has a concordia U-Pb age of 178.3 ± 2.3 Ma (Early Jurassic). K–Ar dating performed on amphiboles of pegmatoid gabbroic dikes cutting the layered ultramafic-mafic sequence produced ages of 279.7 ± 17.1 Ma, 255.6 ± 7.5 Ma, 253 ± 13.8 Ma (Permian), and 184 ± 5.3 Ma (Early Jurassic) (Ghasemi, 2000, Ghasemi et al., 2002, 2004). The contact between the Sikhoran complex and host Upper Paleozoic metamorphic rocks is sharp and intrusive, and associated with a contact metamorphic aureole. The amphibolites have undergone partial melting, producing quartz diorite/ plagiogranite melts and pegmatoids that contain amphiboles and plagioclases with decimeter sizes. K–Ar dating performed on amphiboles in contact with the gabbro intrusion in the Abshour Valley produced ages of 222.6 ± 8 Ma and 201 ± 7.5 Ma (Ghasemi, 2000, Ghasemi et al., 2002, 2004; Table 1). Diabasic dikes with Late Jurassic (159.3 ± 12.4 Ma, 138.3 ± 12 Ma, 136.5 ± 11.7 Ma, 133.9 ± 10.9 Ma) and Late Cretaceous (81 ± 6.6 Ma) ages represent the youngest magmatic phases to intrude through the Sikhoran complex (Ghasemi, 2000; Ghasemi et al., 2002, 2004; Table 1). Sheeted dikes, massive and pillow basalts, hyaloclastites, cherts, radiolarites and pelagic limestones that are expected in a typical oceanic crustal section are absent in the complex. The ultramafic-mafic complex intruded in a succession of Sargaz-Abshour metamorphic rocks, which include garnet-bearing gneisses, amphibolites, schists, and marbles (Fig. 2).

Some previous studies have considered this complex as a mantle plume diapir (Ghasemi, 2000; Ghasemi et al., 1998, 2001, 2002; 2004; Ahmadipour et al., 2003; Behzadi and Shahabpour, 2011; Asadi et al., 2022), although others (Peighambari et al. 2011; Peighambari et al. 2016; Moghadam et al. 2017; Sepidbar et al., 2021; Alesaadi et al., 2022) considered it as an ophiolitic complex. Nevertheless, very little work has been performed on the chemistry of constituent mineral or petrogenetic models for different parts of the complex. Here, we provide new bulk-rock and mineral chemistry information for mantle peridotites including porphyroclastic harzburgite and dunites, chromitites, and layered ultramafic-mafic sections along with U-Pb dating on layered and isotropic gabbros from the Sikhoran complex and metamorphic host rocks (Table 1), in order to determine the source, tectonic setting, nature of mantle melting and age of complex emplacement. The geochemical results obtained in this work for the Sikhoran ultramafic-mafic rocks and associated amphibolite provide context to understand the petrogenesis of the complex, its magmatic and metamorphic history, and its implications for the Paleozoic-Mesozoic geodynamic evolution of SW Iran as an important part of the Iranian sector of Neo-tethys realm.

2.3 Sargaz-Abshur metamorphic rocks

The Sargaz-Abshur metamorphic rocks are the host rocks of the Sikhoran ultramafic-mafic complex, and are located in the southeast part of the SSMMZ. These metamorphic rocks mostly include garnet gneisses and amphibolites, which have Late Carboniferous ages, and interbedded amphibolite-marbles-calcschists and chlorite schists, which have Late Triassic-Early Jurassic ages (Fig. 2). These metamorphic rocks are unconformably overlain by Lower Jurassic conglomerates and sandstones, and Middle-Upper Jurassic limestones. Lower Jurassic conglomerates have many fragments of garnet gneisses and amphibolites of the Late Carboniferous. Some authors have interpreted these metamorphic complexes as Precambrian basement rocks (e.g., Stocklin, 1968); however, others have proposed that Early Cimmerian tectonics were responsible for their syn-tectonic regional metamorphism (Berberian, 1976; Hushmandzadeh et al., 1972; Majidi, 1972, 1974; Sabzehi and Berberian, 1972; Berberian and Nogol, 1974; Sabzehi, 1974). K–Ar dating performed on biotite, muscovite, and amphibole from these gneisses, alongside S-type granites derived from partial melting of the gneisses in contact with gabbro and amphibolite, produced ages of 329.6 ± 7.6 Ma, 324.6 ± 16.6 Ma and 304.9 ± 7 Ma, respectively (Ghasemi, 2000, Ghasemi et al., 2002, 2004). These ages agree with U-Pb ages of zircons extracted from gneisses (326.3 ± 3.3 Ma, 318.7 ± 4 Ma and 312.2 ± 4.1 Shafaii Moghadam et al., 2017; and 334.6 ± 4.9 Ma, this study) (Table 1). In contrast, K–Ar dating performed on amphibole from amphibolites produced ages of 202.1 ± 10.4 Ma and 199.2 ± 10.7 Ma (Ghasemi, 2000, Ghasemi et al., 2002, 2004). These are in agreement with U–Pb ages of zircons extracted from amphibolites (194 ± 1.8 Ma, 188.7 ± 1.4 Ma and 186.2 ± 1.4 Ma; Shafaii Moghadam et al., 2017).

2.4. Cretaceous Ashin-Seghin Blueschists

Despite the long-lived subduction history of the Zagros orogeny (~180 to ~35 Ma ago), very few exposures of high pressure-low temperature (HP/LT) metamorphic rocks (e.g., blueschists and eclogites) associated with its subduction phases occur along its suture zone (Sabzehei, 1974; Okay, 1989; Sabzehei et al., 1994; Davoudian et al., 2016). Fortunately, in the Esfandagheh region there are many well-preserved outcrops of blueschists (Sabzehei et al., 1994, Ghasemi, 2000, Ahmadipour, 2000, Ghasemi et al., 2002, 2004, Ahmadipour et al., 2003; Agard et al., 2006; Angiboust et al., 2016), including the Ashin and the Seghin complexes, which are associated with extensive colored mélange units (Agard et al., 2006; Angiboust et al., 2016). The contact between the Sanandaj–Sirjan Paleozoic-Mesozoic metamorphic rocks and these Cretaceous blueschists are bounded by the major Ashin thrust fault (Sabzehei et al., 1994; Ghasemi, 2000). This association is considered to represent remnants of a Late Cretaceous subduction zone system (e.g., Agard et al., 2006; Angiboust et al., 2016). The Ashin complex includes various rock types (e.g., paragneisses, garnet amphibolites, glaucophane schists, garnet mica schists, quartzites, and marbles) that have experienced an amphibolite-facies metamorphism event (500–550 °C, 0.9 GPa) followed by a blueschist-facies overprint (~500 °C and 1.3 GPa) (Angiboust et al., 2016). The Seghin complex is composed of HP/LT metabasic rocks that occur as meter-sized blocks or larger lenses (up to 200 m in length) in a serpentinite matrix, which were metamorphosed to 450–500 °C and 1.6–1.8 GPa (Angiboust et al., 2016). Whole-rock K-Ar and white mica ^{40}Ar-^{39}Ar and Rb-Sr dating yielded formation ages of ~115–65 Ma (Delaloye and Desmons, 1980; Ghasemi, 2000; Ghasemi et al., 2002, 2004; Agard et al., 2006; Monie and Agard, 2009; Angiboust et al., 2016). These petrological and geochronological data suggest that both tectonic units experienced subduction along different geothermal gradients (~17 °C/km versus ~7 °C/km, respectively) and record significant differences in the time of final deformation (~100–85 Ma versus ~65 Ma, respectively) (Angiboust et al., 2016).

3. Field observation and sample description

The Sikhoran complex comprises of a thick sequence of continuous ultramafic-mafic rocks including basal residual mantle peridotite (tectonite), transitional zone and layered ultramafic-mafic (layered gabbro) cumulates with a Late Carboniferous age (320 Ma; this study) which are cross cut by an Early Jurassic (178 Ma; this study) isotropic gabbro and scattered gabbroic/diabasic dikes, which range in age from Permian to Cretaceous. Primary contacts between mantle rocks, layered ultramafic-mafic cumulates and intrusive gabbros are well preserved. Mantle rocks and layered ultramafic-mafic cumulate fragments occasionally occur within the isotropic gabbro, indicating that they are relatively older than the gabbro itself.

The tectonite zone is the lowest part of the Sikhoran sequence (representing the upper mantle section) and is located in the northern part of the complex (Fig. 1B). Based on the modal content of mafic minerals (olivine, clinopyroxene, orthopyroxene) and following the recommended Ol-Opx-Cpx classification (Streckeisen 1973; in LeMaitre et al. 2002), this sequence is comprised of foliated porphyroclastic harzburgite that gradually transitions into porphyroclastic dunites and chromitite pods (Figs. 3A, B). These rocks are characterized by mantle metamorphism and deformation textures, such as foliations (Fig. 3B), kink bands, stretching, mechanical twinning, granulation, annealing, shadowy extinction and rotated porphyroclasts.

Porphyroclastic harzburgite contains coarse-grained olivine (less than 70 vol.% with Fo = 90%), orthopyroxene (less than 20 vol. % with En = 90%), chrome spinel (less than 1%) and clinopyroxene (~4 vol. %) as primary minerals, and serpentine (chrysotile-antigorite), tremolite-actinolite, magnetite, magnesite and brucite as secondary minerals. Olivines have kink bands, mechanical deformation, shadowy extinction (Fig. 3C) with irregular and interlocking margins.

Porphyroclastic dunites are comprised of olivine (up to 95 vol%), orthopyroxene and clinopyroxene (<5 vol. %), and spinel (~1 vol%) (Fig. 3D). All dunite samples show diffuse alteration with a development of a serpentine-rich matrix (up to 10 vol%). Secondary fibrous amphibole at the rim of pyroxenes is locally observed. All textural evidence for a high-temperature mantle origin and a lower temperature crustal metamorphic overprint found in porphyroclastic harzburgites are also seen in these rocks.

The tectonite zone grades upwards into a transitional mantle-crust zone consisting of interlayers of porphyroclastic dunite (residual upper mantle rocks), layered cumulative dunite and layered and massive stratiform chromitite beds. The dunites are characterized by textural characteristics similar to dunites from the tectonite zone, which includes various types of mantle-related, high-temperature plastic deformation; however, unlike with tectonite zone dunites, they contain both orthopyroxene and clinopyroxene, and sulfide phases (pyrrhotite, pentlandite, chalcopyrite, pyrite). The transitional zone lacks plastic deformation features, but instead shows the effects of viscous deformation, such as flow folding. Dunites contain concordant podiform chromite horizons (Fig. 3E). The invasion of basaltic melt into the dunites caused reactions with olivine to form pyroxenite veins (Khedr et al., 2023). These veins have pyroxenite walls and a gabbroic center (Fig. 3F). They are comprised of olivine (up to 95 vol. %), orthopyroxene, clinopyroxene (<5 vol. %), and spinel (~2 vol. %). Towards the top of the sequence, this zone gradually turns into an ultramafic-mafic cumulate zone with clear layering (Fig. 3G).

Layered ultramafic-mafic cumulates occur in the central parts of the Sikhoran complex, both north and south of Sikhoran village (Fig. 2) and mainly include layered cumulative dunite, harzburgite, lherzolite, wehrlite, websterite, olivine websterite, pyroxenite, chromitites, olivine gabbro, troctolite, gabbronorite, ferrogabbro, diorite/quartz-diorite and plagiogranite. They are characterized by undeformed cumulus textures. The appearance of plagioclase is an important mineralogical difference between this zone and the tectonite and transitional zones. The increase in the modal amount of plagioclase, decrease in the modal amounts of olivine and pyroxenes, and the presence of amphibole and Fe-Ti oxides above this zone are characterized by the appearance of gabbro, anorthosite (feldspar), ferrodiorite and plagiogranite layers. At least, four groups of gabbroic/diabasic dikes including Permian very coarse-grained gabbro-pegmatoid dikes, Triassic gabbros, Jurassic and Cretaceous diabasic dikes cut these rocks (Fig. 3H).

The Sikhoran crustal magmatic sequence contains layered and isotropic gabbros and differentiated rocks (leucogabbros, anorthosite, ferrodiorite/quartz-diorite and plagiogranite). Gabbros are the main rock types in the crustal sequence and can be categorized in two main groups: (i) Late Carboniferous layered gabbros, which are products of magmatic differentiation, and which belong to the crustal section of the Sikhoran complex, and (ii) Early Jurassic isotropic gabbro that intrudes the complex. These units are located structurally at the top of the complex, in its southern part, around Sargaz village and Abshour Valley (Fig. 2). The Early Jurassic isotropic gabbros intruded into the ultramafic-mafic sequence and Paleozoic-Triassic metamorphic rocks (Fig. 3I,J) caused partial melting of the host amphibolites. Metamorphic conditions reached the sanidinite facies, which led to emplacement of plagiogranitic pegmatoids (Fig. 3K).

Amphibolites occur intercalated within Paleozoic marbles, are weakly foliated and mainly contain coarse (0.5 mm) to fine-grained (0.2–0.1 mm) green hornblende and plagioclase. Accessory minerals include oxides, biotite, apatite, and titanite. The orientation of schistosity and shear sense indicators in the amphibolites and foliated ultramafic rocks are similar, suggesting a common deformation history (Fig. 3I). Plagioclase is interstitial between the amphibole grains, and is partially altered to clay minerals. Biotite, ilmenite, apatite and zircon occur in minor proportions or as accessory phases.

Felsic orthogneiss was additionally collected from southeast of Sargaz village. It has a weak foliation and contains variably sized garnet (>2 mm to <0.5 mm) in a matrix mainly composed of quartz, biotite, orthoclase, and plagioclase (Fig. 3J). Plagioclase preserves its magmatic zonation and shows slight sericitization. Apatite and white mica are minor phases.

5. Results

5.1. Zircon U–Pb geochronology

Four samples from the Sikhoran magmatic-metamorphic complex were dated using the U-Pb method on zircons via SHRIMP at the Beijing SHRIMP Center, China. These units comprised one layered gabbro (SF18-18), one isotropic gabbro (SF18-13), one granitic gneiss (SF18-8), and one anatectic plagiogranite dyke (SF18-12) that was derived from partial melting of the Upper Paleozoic amphibolites in contact of the Early Jurassic gabbro intrusion. All results are presented in the Supplementary Table 1 and discussed below.

Sample SF18-18 (Layered gabbro)

Zircon grains from sample SF18-18 are euhedral short prismatic zircons, and their CL images show broad oscillatory and sector zoning (Fig. 4A) typical of igneous zircons. Zircons are fine-grained (70 to <100 μm). Their U and Th concentrations are variable (U = 41–456 ppm, Th = 2–136 ppm), with very low to medium Th/U = 0.04–.81. Six spot analyses from six grains produced a near-concordant group on a concordia diagram, with a weighted-mean $^{206}Pb/^{238}U$ vs. $^{207}Pb/^{235}U$ age of 320.8 ± 6.4 Ma (MSWD = 1.6; Table 1; Fig. 4B,C).

Sample SF18-13 (*Isotropic gabbro*)

Zircon grains separated from sample SF18-13 are euhedral and their CL images (Fig. 4D) show sector and oscillatory zoning, typical of magmatic zircons. Their U and Th concentrations are variable (U = 83–343 ppm, Th = 19–130 ppm), with Th/U = 0.219–0.425. Fifteen spot analyses from 15 zircon grains produced a concordant group on a concordia diagram, with a weighted-mean $^{206}Pb/^{238}U$ vs. $^{207}Pb/^{235}U$ age of 178.3 ± 2.3 Ma, (MSWD = 0.62; Table; Fig. 4E,F).

Sample SF18-8 (*Gneiss*)

Zircon grains from sample SF18-8 exhibit euhedral to subhedral morphologies with CL oscillatory and sector zoning (Fig. 4G) (Corfu et al., 2003). They have variable U and Th concentrations (U = 81–158 ppm, Th = 30–96 ppm), with Th/U = 0.37–0.66. Ten spots from 10 grains produced a concordant group on a concordia diagram, with a weighted-mean $^{206}Pb/^{238}U$ vs. $^{207}Pb/^{235}U$ age 334.6 ± 4.9 Ma, (MSWD = 1.3; Table; Fig. 4H,I). Zircons from this sample also showed a restricted range in $^{206}Pb/^{238}U$ ages. The cores of long prismatic zircons yielded a weighted mean age of 349.7 ± 9.7 Ma (Supplementary Table 1), whereas the zoned rims indicate a younger age of 323.4 ± 7.4 Ma (Supplementary Table 1).

Sample SF18-12 (Anatectic plagiogranite dike)

Zircon grains separated from sample SF18-12 are small (<100 μm) and show euhedral to anhedral morphologies. Their CL images (Fig. 4J) show sector and oscillatory zoning, typical of magmatic zircons, Their U and Th concentrations are highly variable (U = 318–1817 ppm, Th = 87–881 ppm), with Th/U = 0.227–0.875. Ten spot analyses from 10 zircon grains produced a concordant group on a concordia diagram, with a weighted mean $^{206}Pb/^{238}U$ vs. $^{207}Pb/^{235}U$ age of 187.2 ± 2.6 Ma, (MSWD = 0.48; Table; Fig. 4K,L).

5.2. Whole-rock compositions

Whole-rock major, minor, and trace element analyses for the Sikhoran complex ultramafic-mafic sequence were determined for dunites of porphyroclastic zone (six samples), dunite, wehrlite and clinopyroxenite of the transitional zone (20 samples), dunite, lherzolite and harzburgites of layered ultramafic zone (13 samples), layered gabbros (7 samples), isotropic gabrros (22 samples) and diabasic dikes (7 samples). The results are presented in Supplementary Table 2 and discussed below.

The analyzed Sikhoran ultramafic samples show loss of ignition (LOI) values in the range 0.7–11.3 wt. %, suggesting highly variable degrees of serpentinization. Dunites from the porphyroclastic zone are characterized by low SiO_2 (35.8–38.7 wt. %), Al_2O_3 (0.04–0.32 wt. %), CaO (0.2–0.5 wt. %), and TiO_2 (0.01 wt. %) contents. Their CaO/Al_2O_3 (0.9–5.5) and Mg# (0.86–0.90) ratios, and Cr (1450–5000 ppm) and Ni (1750–2800 ppm) contents are close to those of primitive mantle (PM) values (CaO/Al_2O_3 = 0.8, Mg# = 0.88, Cr = 3240 ppm; Hart and Zindler, 1986; McDonough and Sun, 1995; Workman and Hart, 2005). The compositions of Sikhoran dunites (Al_2O_3, MgO, and SiO_2) from the porphyroclastic zone plot above the abyssal peridotite field (Barnes et al., 2014) along the terrestrial array on a MgO/SiO_2 vs Al_2O_3/SiO_2 bivariate plot (Fig. 5A), within the field denoting interaction with olivine-rich melt (Hart and Zindler, 1986). As such, they display characteristics of being refractory peridotites (Al_2O_3/SiO_2 < 0.04 and MgO/SiO_2 = 1.10–1.30) (Fig. 5A).

Dunites and wehrlite from the transitional zone also have low SiO_2 (36–41 wt. %), Al_2O_3 (0.1–0.8 wt. %), CaO (0.2–3.4 wt. %), and TiO_2 (0.01–0.05 wt. %) contents. Their CaO/Al_2O_3 (0.8–4.7) and Mg# (0.82–0.91) ratios, and Cr (1040–5200 ppm) and Ni (925–3300 ppm) contents are close to PM values of 0.8, 0.88, and 3240 ppm, respectively (Hart and Zindler, 1986; McDonough and Sun, 1995; and Hart, 2005). The Sikhoran dunites and wehrlites from the transitional zone plot above the abyssal peridotites field (Barnes et al., 2014) along the terrestrial array, or in the space of 'interaction with olivine-rich melt (Hart and Zindler, 1986; Fig. 5A) displaying a refractory peridotite character (Al_2O_3/SiO_2 < 0.04 and MgO/SiO_2 = 0.98–1.07) (Fig. 5A). However, clinopyroxenites from the transitional zone are characterized by higher SiO_2 (46–51 wt. %), Al_2O_3 (0.8–5.7 wt. %), CaO (12–19 wt. %) and TiO_2 (up to 0.2 wt. %) contents. They have CaO/Al_2O_3 (2.9–19) and Mg# (0.79–0.90) ratios, and Cr (1050–4400 ppm) and Ni (220–850 ppm) contents greater than those of PM (CaO/Al_2O_3 = 0.8, Mg# = 0.88, Cr = 3240 ppm; Hart and Zindler, 1986; McDonough and Sun, 1995; and Hart, 2005). The compositions of olivine-pyroxenite from the transitional zone plot along the terrestrial array, within the field of metasomatism/seafloor weathering (Hart and Zindler, 1986) (Fig. 5A), but display a refractory peridotite character (Al_2O_3/SiO_2 < 0.04 and MgO/SiO_2 = 0.40–0.65).

Dunite, wehrlite, lherzolite, and harzburgite from the layered ultramafic zone are characterized by low to moderate SiO_2 (dunite: 40 wt. %; lherzolite/harzburgite: 4l–45 wt. %; wehrlite: 49–52 wt. %), Al_2O_3 (0.7–2.8 wt. %), CaO (dunite: 0.45 wt. %; lherzolite/harzburgite: 0.7–5.5 wt. %; wehrlite: 16–19 wt. %), and TiO_2 (0.02–0.5 wt. %) contents. They have variable CaO/Al_2O_3 (dunite-lherzolite-harzburgite: 0.7–37; wehrlite: 6.4–8.4) and Mg# (0.80–0.89) ratios, and Cr (1450–5000 ppm) and Ni (1780–4300 ppm) contents that are close to those PM values (CaO/Al_2O_3 = 0.8, Mg# = 0.88, Cr = 3240 ppm; Hart and Zindler, 1986; McDonough and Sun, 1995; Workman and Hart, 2005). These compositions plot along the abyssal peridotite field (Barnes et al., 2014) along the terrestrial array on a MgO/SiO_2 vs Al_2O_3/SiO_2 bivariate plot, whereas, wehrlites plots below abyssal peridotites in the space defined by metasomatism/seafloor weathering (Hart and Zindler, 1986) (Fig. 5A).

All gabbro samples from the layered and isotropic zones of the Sikhoran complex have LOI ~ 0.6–3.7 wt. %, revealing generally low degrees of alteration. Isotropic gabbros have the following compositions: SiO_2: 40–49 wt. %; FeO_T = 4–9 wt. %; CaO: 8–19 wt. % and Al_2O_3 (11–28 wt. %). Layered gabbros have similar SiO_2 (42–48 wt. %), FeO_T (6–12 wt. %, except ferro-gabbro samples that have 16–20 wt. %), CaO = 11–18 wt. %, and Al_2O_3 (10–20 wt. %) contents compared to isotropic gabbros. Both types of gabbro have variable MgO (5.4–19.6 wt. %) and TiO_2 (0.04–0.20 wt. %) contents, with Mg# values ranging from 0.34 to 0.81. On a total alkali versus silica (TAS) diagram (LeMaitre et al., 2002), all plot within the gabbro field (Fig. 5B). In a Th vs Co diagram (Fig. 5C), the analyzed Sikhoran gabbros fall in the IAT (Hastie et al., 2007), forearc basalt (FAB) (Reagan et al. 2010; Ishizuka et al., 2011), and E-MORB fields. In a Nb/Yb vs Th/Yb diagram (Pearce 2008), both gabbro types fall within the subduction-related ophiolites field, which plots above the mantle array (Fig. 5D). The layered gabbros are characterized by low to moderate Cr (30–500 ppm), Ni (21–200 ppm), and V (118–330 ppm) contents. They show negligible variation in some LILEs (e.g. Sr = 80–190 ppm) and HFSEs (e.g. Nb < 1.5 ppm, Y = 0.7–6.4 ppm, and Zr = 2.5–10 ppm). Similarly, the isotropic gabbros are characterized by low to moderate Cr (80–470 ppm), Ni (20–235 ppm), and V (44–410 ppm) contents. They show variation in some LILEs (e.g. Sr = 62–302 ppm) and HFSEs (e.g. Nb = 0.1–6.4 ppm, Y = 0.9–6.0 ppm, and Zr = 2–33 ppm).

On a chondrite (CI)-normalized rare earth element (REEs) diagram (Sun and McDonough, 1989) (Fig. 6A), the Sikhoran cumulate layered gabbros show flat patterns, with mean $(La/Yb)_N$ and $(Dy/Yb)_N$ values of 1.2 and 0.8, without Eu anomalies (mean value of 1.6, comparable to that of N-MORB). The isotropic gabbros are also characterized by somewhat flat patterns to relatively enriched light REEs (LREEs) patterns, except two samples that are relatively enriched in middle REEs (MREEs). They typically have $(La/Yb)_N$ and $(Dy/Yb)_N$ mean values of 1.7 and 1.3, respectively, and lack an Eu anomaly (mean value of 1.02), similar to that of E-MORB (Sun and McDonough, 1989) (Fig. 6A). They display similar flat profiles characterized by minor depletion of Ti and Nb. In both CI- and primitive mantle (PM)-normalized diagrams, isotropic gabbros show relative enrichment compared to layered gabbros.

The studied diabasic dikes from the Sikhoran complex are characterized by low to moderate LOI (1.9–3.7 wt. %) contents, suggesting a low degree of post-crystallization alteration (Supplementary Table 2). They show SiO_2 contents of 44–55 wt. %, Al_2O_3 contents of 14–16 wt. %, and Fe_2O_3 contents of 9–13 wt. %, with variable CaO (8–18 wt. %), MgO (3.5–9.7 wt. %) and TiO_2 (0.4–1.1 wt. %) contents, and Mg# = 0.31–0.57. In a total alkali ($Na_2O + K_2O$) versus silica diagram (TAS; LeMaitre et al., 2002), the Sikhoran diabasic dikes fall in the gabbro and gabbro-diorite fields, which are intrusive equivalents of basalt and basaltic andesite (Fig. 5B). In a Th vs Co diagram (Fig. 5C), the dikes show an IAT and E-MORB affinity (Hastie et al., 2007). In a Nb/Yb vs Th/Yb diagram (Pearce 2008) (Fig. 5D), their compositions fall above the mantle array (Fig. 5D).

The Sikhoran diabases are characterized by moderate Cr (76–270 ppm), Ni (15–200 ppm), and V (234–500 ppm) contents, and REE diagrams reveal two distinct patterns: flat to slightly enriched LREE and slightly depleted LREE. The diabasic rocks with slightly enriched LREE profiles show relative enrichment in HFSE contents, including Nb (1–3 ppm), Y (1.1–2.7 ppm), and Zr (25–55 ppm), whereas slightly depleted LREE lavas are characterized by lower Nb (0.6 ppm), Y (0.5 ppm), and Zr (5.5 ppm) contents. On a chondrite (CI)-normalized REE diagram (Sun and McDonough, 1989) (Fig. 6C), diabase samples with flat patterns show $(La/Yb)_N$ and $(Dy/Yb)_N$ mean values of 1.7 and 1.1, respectively, but lack an Eu anomaly (mean value of 1.0). Their REE patterns are similar to those of E-MORB (Fig. 6C). The diabasic rocks with slightly depleted LREE contents have $(La/Yb)_N$ and $(Dy/Yb)_N$ values of 0.97 and 1.13, respectively, and also lack an Eu anomaly (mean value of 1.0).

In PM-normalized multi-element diagrams (Sun and McDonough, 1989) (Fig. 6D), Sikhoran diabasic dikes show enriched patterns, with negative Nb anomalies. They show affinities of normal- and enriched-type mid-ocean ridge basalts (N-MORB and E-MORB) respectively (Sun and McDonough, 1989), and are comparable with high-Ti N-MORB basalts from the Albanide–Hellenide Tethyan ophiolite (Saccani et al., 2018) (Fig. 6D).

5.3. Minerals of the Sikhoran ultramafic–mafic complex

Spinel

Compositional data were obtained for spinel from Sikhoran porphyroclastic, transitional zone and layered peridotites (Supplementary Table 3a). Spinels from porphyroclastic dunite and chromitite have low Al_2O_3 (5.4–13.6 wt. %), high Cr_2O_3 (49.3–60.3 wt. %), high FeO_{tot} (18.2–31.7 wt. %), and low TiO_2 0.06–0.27 wt. % contents. They show Mg# (= Mg/(Mg + Fe^{2+}) atomic ratio) values 0.30–0.68, with a mean of 0.49, and Cr# (= Cr/(Cr + Al) atomic ratio) values of 0.58–0.71, with a mean of 0.66. These correspond to the chemistry of chromite and magnesio-chrome spinel (Melluso et al., 2014).

Spinels in dunite from the transitional zones are characterized by higher Al_2O_3 (19.2–31.2 wt. %), Cr_2O_3 (29.7–44.8 wt. %), FeO_{tot} (24.2–31.6 wt. %), and TiO_2 0.15–0.54 wt. % contents. They have Mg# values of 0.34–0.58, with a mean of 0.46, and Cr# values of 0.39–

0.61, with a mean value of 0.52. These compositions correspond to the chemistry of chromite and magnesio-chromite, and hercynite–magnetite spinels (Melluso et al., 2014).

Spinels from layered lherzolite show wide variations in Al_2O_3 (23.2–49.5 wt. %), Cr_2O_3 (10.7–38.2 wt. %), FeO^{tot} (16.7–32.2 wt. %), and TiO_2 0.03–0.24 wt. %) contents. They display Mg# in the range 0.40–0.87 (mean value = 0.69) and Cr# in the range 0.12–0.50 (mean value = 0.30). The layered harzburgites are characterized by Al_2O_3 values of 26.6–44.1 wt. %, Cr_2O_3 (21.0–38.8 wt. %), FeO^{tot} (17.6–22.4 wt. %), TiO_2 0.07–0.31 wt. %), Mg# in the range 0.64–0.78 (mean value = 0.70) and Cr# in the range 0.24–0.50 (mean value = 0.39).

The range of compositional variation for spinel is shown in the Al–Cr–Fe^{3+} ternary diagram (Fig. 7A). Spinel from Sikhoran porphyroclastic dunites and related chromites and chromitites within the transitional zone has high Cr contents (mean value = 0.37) when compared to Al (mean value = 0.1) and Fe^{3+} (mean value = 0.23), and can thus be classified as Cr-spinel. Most spinel analyses plot in the field for supra-subduction zone (SSZ) forearc peridotites (Fig. 7A; e.g. Ishii et al., 1992; Khedr and Arai, 2017).

Spinel from Sikhoran transitional zone dunite and layered lherzolite-harzburgite show higher contents of Al (mean value = 0.3) with respect to Cr (mean value = 0.23) and Fe^{3+} (mean value = 0.13) and can be classified as Al- to Al–Cr–Fe spinel. Most analyses plot in the spinel field for abyssal to supra-subduction zone (SSZ) forearc peridotites (Fig. 7A) (e.g. Ishii et al., 1992; Khedr and Arai, 2017).

Olivine

Olivine grains in porphyroclastic mantle sections have mean forsterite contents (Fo = Mg/(Fe + Mg) atomic ratio) of 92% and 96% for dunite and chromitite respectively. Olivine NiO content varies from 0.22–0.46 wt. % in dunite and 0.31–0.48 wt. % in chromitite (Supplemental Table 3b) and lies below the olivine-mantle array, matching South Sandwich forearc peridotites (e.g. Takahashi et al., 1987; Moghadam et al., 2015) (Fig. 7B). Olivine in transitional zone peridotites is less magnesian, with a mean Fo value of 88% (~0.19 wt. % NiO) in dunitic rocks and ~86% Fo (~0.16 wt. % NiO) in wehrlite and olivine-clinopyroxenite, and plot below the olivine-mantle array (Fig. 7B). Olivine in layered ultramafic–mafic zones are characterized by low Mg contents compared to the above-described samples, with 84% Fo (~0.22 wt. % NiO), and 91% Fo (~0.38 wt. % NiO) in lherzolite and harzburgite rocks, respectively (Supplemental Table 3b). They plot within the field for abyssal peridotite in Fig. 7B.

Pyroxene

All lithologies within the Sikhoran–Soghan peridotites and mafic rocks contain both orthopyroxene and clinopyroxene (Supplementary Table 3c and d). Orthopyroxene in the transitional zone is Mg-rich, with Mg# of 82–90 (mean value = 86), 83-89 (mean value = 87) and 89-90 (mean value = 89.5) for wehrlite, olivine-clinopyroxenite and dunites, respectively. They have relatively the same Al_2O_3 up to 1.5–2.5 wt. %, TiO_2 up to 0.2 wt. %, CaO of 0.3–0.8 wt. %, and Na_2O < 0.04 wt. %. Cr_2O_3 contents lie in the range 0.46–0.56 wt. %, 0.05-0.29 wt. % and 0.09-0.28 wt. % for dunites, wehrlite, and olivine-clinopyroxenite, respectively. Similar to orthopyroxene, clinopyroxene shows Ca- and Mg-rich compositions, with Mg# values of 89.4–94.6 (mean value = 92.5), 86.1-90.0 (mean value = 88.6) and 90.8-94.8 (mean value = 92.9) for dunite, wehrlite and olivine-clinopyroxenite, respectively. Al_2O_3 contents of dunite, wehrlite and olivine-clinopyroxenite are in the range 0.5–2.0 wt. % (mean value = 1.0 wt. %), 1.2–3.2 wt. % (mean value = 2.8 wt. %), and 0.2–2.6 wt. % (mean value = 1.2 wt. %), respectively. Clinopyroxenes in dunites are characterized by higher Cr_2O_3 (0.4–0.9 wt. %; mean at 0.6 wt. %) contents than those of wehrlite (0.03–0.46 wt. %; mean at 0.3 wt. %) and olivine-clinopyroxenite (0.25–0.80 wt. %; mean at 0.48 wt. %). %, All grains have more TiO_2 (0.02–0.27 wt. %) and Na_2O (up to 0.28 wt. %) with respect to orthopyroxene (Supplementary Table 3). According to the Wo–En–Fs ternary system, orthopyroxene is classified as enstatite ($En_{83–89}Fs_{10–15}Wo_{0.7–5.5}$) and clinopyroxene as diopside ($En_{41–49}Fs_{3–6}Wo_{46–53}$).

Orthopyroxenes from wehrlite, lherzolite and harzburgites of the layered ultramafic zone have the same Mg-rich compositions, with Mg# values of 76.5–84.8 (mean value = 83) and 88.6-91.7 (mean value = 91), Al_2O_3 up to 1.1–3.3 wt. % (mean value = 2.3 wt. %) and 1.5-2.1 wt. % (mean value = 1.8 wt. %), TiO_2 of up to 0.11 wt. % and up to 0.16 wt. %, CaO of 0.42–0.82 wt. % (mean value = 0.57 wt. %) and 0.27-0.66 wt. % (mean value = 0.5 wt. %). Cr_2O_3 is in the range 0.0.08–0.33 wt. % (mean value = 0.21 wt. %) and 0.18-0.67 wt. % (mean value = 0.35 wt. %), respectively. Clinopyroxene also shows a Ca- and Mg-rich composition, with Mg# of 85–92 (mean value = 88), 79.4-81.4 (mean value = 80.5) and 90.8-93.7 (mean value = 92.4), Al_2O_3 in the range 1.4–3.4 wt. % (mean value = 2.3 wt. %), 1.4-1.9 wt. % (mean value = 1.8 wt. %) and 2.1-3.5 wt. % (mean value = 2.8 wt. %), Cr_2O_3 of 0.03–0.67 wt. % (mean value = 0.33 wt. %), up to 0.09 wt. % (mean value = 0.05 wt. %) and 0.56-0.83 (mean value = 0.76 wt. %) for lherzolite, pyroxenite and harzburgite, respectively. According to the Wo–En–Fs system, orthopyroxene can be classified as enstatite ($En_{82–91}Fs_{8–17}Wo_{0.5–1.3}$) and clinopyroxene as diopside ($En_{45–57}Fs_{3–8}Wo_{34–49}$).

Orthopyroxenes from layered and isotropic gabbro are characterized by lower Mg contents, with Mg# of 75–88, Al_2O_3 of 1.1–2.4 wt. %, TiO_2 of 0.06–0.18 wt. %, CaO of 0.4–1.1 wt. %, and Na_2O < 0.04 wt %. Cr_2O_3 is present up to 0.07 wt. %. Clinopyroxene has Mg# of 84–93, Al_2O_3 of 0.9–3.5 wt. %, Cr_2O_3 up to 0.44 wt. %, TiO_2 of 0.06–0.36 wt. %, and Na_2O of 0.05–0.35 wt. %. According to the Wo–En–Fs system, orthopyroxene can be classified as enstatite–hypersthene ($En_{66–82}Fs_{17–33}Wo_{0.8–2.3}$) and clinopyroxene as diopside ($En_{40–51}Fs_{6–14}Wo_{35–48}$). Clinopyroxene (Fig. 7C) and orthopyroxene (not shown) have SSZ (Parkinson et al., 1992; Parkinson and Pearce, 1998; Pearce et al., 2000; Pagé et al., 2008) and abyssal (Hellebrand et al., 2001; Choi et al., 2008 and references therein) peridotite signatures, although the extent of SSZ affinity decreases from porphyroclastic and transitional to layered ultramafic sequences.

Amphibole

Amphibole is present in all Sikhoran peridotites and mafic rocks. The amphibole in dunites of porphyroclastic sequences has high MgO (~23 wt. %) and low FeO (1.3–1.8 wt. %) contents (Supplementary Table 3e). Amphibole from the dunites of porphyroclastic sequences has a Mg# of 0.97. The analyzed amphibole crystals have Ca contents of 1.9–2.8 a.p.f.u, and so are calcic following the classification of Leake et al. (1997). In a Si vs. Mg/(Mg + Fe) binary diagram, amphibole from the dunites of porphyroclastic sequences plots in the tremolite field and was formed during serpentinization resulted from crustal metamorphism (Fig. 8A).

Amphibole from ultramafic layered sequences has low MgO (~16–18 wt. %) and high FeO (3.8–6.8 wt. %) contents (Table 3), which correspond to Mg# values of 0.82–0.89. The analyzed amphibole crystals are calcic, with Ca of 1.8–1.9 a.p.f.u, and plot in the tschermakite to magnesio-hornblende field on a Si vs. Mg/(Mg + Fe) binary diagram (Fig. 8A).

Amphibole in gabbro from the mafic layered and isotropic sequences has similar MgO (~17 wt. %) contents but higher FeO (5.3–13 wt. %) contents than amphibole in ultramafic sequences (Supplementary Table 3e). The Mg# values for amphibole from the mafic layered and isotropic sequences is 0.67–0.86. Grains from mafic layered and isotropic sequences are calcic, with Ca of 1.8–1.9 a.p.f.u, and plot in the tschermakite and magnesio-hornblende fields, respectively (Fig. 8A).

Amphibole in diabasic dikes from the crustal sequences have low MgO (~7.9–14.5 wt. %) and high FeO (21–23 wt. %) contents compared to ultramafic and mafic gabbro sequences (Supplementary Table 3e). These diabasic dike amphiboles have Mg# values of

0.48–0.99 and Ca contents of 0.4–3.9 a.p.f.u. They are calcic amphiboles, and plot in the tschermakite to magnesio-hornblende fields on classification diagrams (Fig. 8A).

5.4. Minerals of metamorphic rocks

Amphibole

Amphibole is the main constituent in the matrix of amphibolite samples (Supplementary Table 3e). Grains have an intense greenish pleochroism and are enriched in Fe^{+3} (0.6–1.0 p.f.u.). Contents of Si (6.3–6.7 p.f.u.), Ti (0.05–0.1 p.f.u.), Na (0.14–0.35 p.f.u.) and K (0.10–0.14 p.f.u.) are fairly uniform. Amphibole within the amphibolite has low MgO (~8.5–10.5 wt. %) and high FeO (20 wt. %) contents with respect to amphibole from ophiolitic sequences (Supplementary Table 3e). The Mg# values for amphibole from the amphibolite are 0.52–0.62 and they have Ca contents ranging from 1.9 to 2.1 a.p.f.u. These compositional criteria correspond to tschermakite to magnesio-hornblende according to Leake et al. (1997) (Fig. 8A). Amphibole in the in the layered ultramafic rocks, basaltic and pegmatoid dikes has an MOR affinity, whereas others plot in fields defining a forearc basin (Fig. 8B).

5.5. Mantle potential temperature and magma crystallization conditions

The chemical compositions of amphibole and spinel-olivine equilibrium pairs can be used to calculate the temperature, pressure, and oxygen fugacity of the mantle source and the magma from which they crystallized (Matthews et al., 2016; Henry et al., 2005; Ridolfi et al., 2008). Mantle potential temperature (*Tp*) estimates based on petrological observations work by inferring primary magma compositions, followed either by fitting this to predicted accumulated mantle melt compositions (Herzberg and Asimow, 2015; Hole and Millet, 2016) or estimating the temperature of olivine saturation and extrapolating back to the solidus (Putirka, 2008a, 2016). Aluminum exchange between olivine and spinel was estimated as a function of temperature by Coogan et al. (2014) via experimental statistics (Coogan et al., 2014; Wan et al., 2008), as shown by the following relationship: T (K) = {10000/[0.575 + 884Cr# + 0.897ln(kd)]}. Here, kd = $Al_2O_{3olivine}/Al_2O_{3spinel}$ and Cr# = Cr/(Cr + Al_{spinel}). Uncertainties in each parameter are reported by Coogan et al. (2014).

Estimates of mantle *Tp* for different parts of the Sikhoran mantle section are given in Supplementary Table 4. They show offsets and variable widths in crystallization temperature distributions between each part of mantle. This variation in crystallization temperature between each section appears to co-vary with the Fo content of olivine (Matthews et al., 2016). Olivine-spinel crystal pairs from the porphyroclastic dunites and transitional ultramafic rocks, which have Fo contents of 89–92, are estimated to have crystallized at temperatures of 1108–835 °C (average of 955 °C) and 1139–863 °C (average of 1007 °C), respectively (Supplementary Table 4a; using estimates of mantle *Tp* of Coogan et al., 2014). Olivine-spinel pairs from the layered ultramafic rocks have lower Fo contents of 82–91, and are estimated to have crystallized at temperatures of 949–724 °C (average of 861 °C) (Supplementary Table 4a; using estimates of mantle *Tp* method of Coogan et al., 2014).

The composition of olivine, spinel and amphibole from ultramafic–mafic complex has been also used to evaluate the temperature, pressure, and oxygen fugacity of magmas from which they crystallized (Balhouse et al., 1990). Most crystals from the porphyroclastic dunites are estimated to have crystallized at temperatures of 717–826 °C, which are significantly lower than mantle *Tp* values obtained by olivine–spinel Al-exchange thermometry. Nonetheless, this disparity is expected, as the highest crystallization temperatures of olivine–spinel Al-exchange thermometry reflect crystallization of melts derived directly from the mantle, whereas, the lower temperature reveal mineral crystallization .

Layered and transitional ultramafic rocks are estimated to have crystallized at temperatures of 752–957 °C and 747-918°C, respectively, which correspond to depths below the ocean floor of up to 10–11 km (Supplementary Table 4b). These temperatures are relatively lower than mantle *Tp* values (maximum mantle *Tp*) obtained by olivine–spinel Al-exchange thermometry.

Mineral compositions were used to constrain the fO_2 of magma following the method of Balhouse et al., (1990), with results shown in Supplementary Table 4. These data show that the $\Delta\log fO_2$ for olivine-spinel pair in the porphyroclastc, transitional and layered ultramafic rocks was between 2.8-4.4 (i.e. at 717–826 °C °C), 2.1-3.2 (i.e. at 747-918 °C) and 2.3-3.1 (i.e. at 752-957 °C). Amphibole compositions were also used to constrain the fO_2 of mafic magma following the method of Ridolfi et al. (2008), with results shown in Supplementary Table 4. Amphibole within the layered gabbros (320 Ma; this sudy) formed at relatively higher temperatures of (953–1003 °C and 291–470 MPa); however, amphibole in massive gabbros (178 Ma; this study) show lower temperature conditions of 722–827 ° (Supplementary Table 4). The calculated $\log fO_2$ values for amphibole in layered gabbro were between NNO −8.5 and NNO -9.0, along the stable range of the $(NNO)^{-2}$ buffer (Behrens and Gaillard, 2006), similar to those of the ultramafic sequences, whereas those of isotropic gabbro were between NNO −11 and NNO -13, significantly lower than those of ultramafic sequences.

6. Discussion

The new petrological and geochemical data obtained in this study allow four main aspects of the geological evolution of the Sikhoran complex to be refined: (1) petrogenesis of the Sikhoran ultramafic–mafic rocks, including (2) constraints on partial melting, melt–peridotite interaction and metasomatic processes; and the (3) origin and emplacement of the Sikhoran ultramafic units. These allow us to address a key question: (4) is the Sikhoran ultramafic–mafic complex a relic of an ascending and partially melted mantle plume?

6.1. Petrogenesis of the Sikhoran ultramafic–mafic rocks

The Sikhoran complex contains ultramafic and mafic lithologies, and associated metamorphic rocks, including (i) porphyroclastic, transitional, and layered ultramafic rocks, including lherzolite, harzburgite, dunite and minor wehrlite and pyroxenite; (ii) layered and isotropic gabbros; and (iii) metamorphic rocks including schist, amphibolite, gneiss and marble.

Whole-rock geochemical compositions obtained for Sikhoran ultramafic–mafic sequences, along with their mineral chemistry, have been used to decipher their origin and petrogenesis. The Cr-spinel in Sikhoran dunites from the porphyroclastic zone has high Cr# (0.58–0.71, with a mean of 0.66) and Mg# (0.30–0.68, with a mean of 0.49) values. This suggests its derivation from a more depleted mantle source region with respect to other parts of the Sikhoran ultramafic–mafic sequences, which typically have lower Cr# (0.23–0.70, mean value of 0.45) and Mg# (0. 0.29–0.63, mean value of 0.46) values. On an Al–Cr–Fe^{3+}ternary diagram (Fig. 7A) and Mg# vs. Cr# diagram (Fig. 9A), spinel analyses from dunites, chromitite within the porphyroclastic zone and chromitites within the transitional zone plot in the field of forearc peridotites (e.g., Arai et al., 2011; Khedr and Arai 2013, 2017; Khedr et al., 2014). By contrast, transitional ultramafic sequences show both forearc and abyssal peridotite affinities, close to and/or overlapping with a supra-subduction zone (SSZ) setting (e.g. Ishii et al., 1992; Khedr and Arai, 2017). A SSZ environment is also suggested by the affinity of spinels from forearc settings, such as the Soghan (Sepidbar et al., 2021) and from the South-Sandwich arc-basin system (SSABS) peridotites (Pearce et al., 2000). Layered ultramafic sequences show geochemical affinities to abyssal peridotites and/or those that form in MOR settings (e.g. Dick and Bullen, 1984; Khedr et al., 2014) (Fig. 7A, 9A).

Olivine in Sikhoran dunites from porphyroclastic sequence has a primary composition of Fo = 90–93% and NiO = 0.22–0.56 wt. %, and thus is considered residual mantle olivine (e.g. Arai, 1994; Moghadam et al., 2013; Khedr and Arai 2013, Khedr et al., 2023; Khedr et al., 2022) (Fig. 6B; Pagé et al., 2008; Moghadam et al., 2013), plotting below the olivine-mantle array, but parallel to it (e.g. Takahashi et al., 1987; Szilas et al., 2015). These compositions also plot at the intersection of abyssal and forearc peridotites (Pagé et al., 2008). The studied olivine is comparable in composition to mantle olivine from the Oman harzburgites (Khedr et al., 2013, 2014 and references therein), and abyssal and SSABS peridotites (after Pearce et al., 2000). When spinel Cr# is plotted against the Fo content of olivine in porphyroclastic dunites (Fig. 9B), results lie within the olivine–spinel mantle array (OSMA; Arai, 1994) and in the fields of average abyssal peridotites (Dick and Bullen, 1984) and SSZ peridotites (e.g. Moghadam et al., 2014). Such geochemical features of the studied peridotites have also been described for SSABS peridotites (Pearce et al., 2000), and the Soghan and Oman SSZ peridotites (Tamura and Arai 2006). The Cr-spinel in Sikhoran peridotites from transitional (Mg# values of 0.34–0.58, with a mean of 0.46 and Cr# values of 0.39–0.61, with a mean value of 0.52) and layered sequences (Mg# in the range 0.40–0.87 (mean value 0.69) and Cr# in the range 0.12–0.50 (mean value 0.30) for lherzolite and Mg# in the range 0.64–0.78 (mean value 0.70) and Cr# in the range 0.24–0.50 (mean value 0.39) for layered harzburgites, suggest derivation from a less depleted mantle source region with respect to the Sikhoran porphyroclastic sequence (with Mg# values 0.30–0.68, a mean of 0.49, and Cr# values of 0.58–0.71, with a mean of 0.66). Olivine in Sikhoran peridotites from the transitional and layered ultramafic sequences has residual composition of Fo = 88–93 % and NiO = 0.31–0.49 wt. %, (e.g. Moghadam et al., 2013; Khedr and Arai 2013) (Fig. 7B; Pagé et al., 2008; Moghadam et al., 2013), which is comparable to the compositions of mantle olivine from abyssal and SSABS peridotites (after Pearce et al., 2000)._These distinctions are emphasized in a TiO_2 vs. Cr# diagram, where spinels from layered ultramafic zone plot in the field of depleted peridotites (Fig. 9). Transitional dunite spinels plot in the overlapping field of fore-arc spinels, but porphyroclasic dunite–chromitite spinels have compositions similar to boninitic spinels (Fig. 9), suggesting a change from early proto-forearc spreading to generate tholeiites to late proto-forearc magmatism to generate boninites.

The Sikhoran mafic suite contains layered and isotropic gabbros, and diabase dikes. The dated layered gabbro associated with the ultramafic-mafic complex has a different age (sample SF18-18, 320.8 Ma) and condition crystallization (temperatures and fO_2; see section 5.5) from isotopic gabbro (sample SF18-13, 178.3 Ma), suggesting that the layered and isotropic gabbros formed in a different condition and tectonic regime. As the diabase dikes cross-cut the older layered gabbros, the layered gabbro and diabase dikes are likely unrelated to the ultramafic complex. However, both gabbro types (layered and massive) and diabasic dikes have a tholeiitic character, and thus are considered as subduction-related rocks, based on Th vs Co (Hastie et al., 2007) and Th/Yb vs Nb/Yb (Dilek and Furnes, 2011) systematics (Fig. 5C, D) (Moghadam and Stern, 2015; Omrani et al., 2017). The consistent Nb, Th, and Ti depletion in all Sikhoran gabbros and diabasic dikes again supports magma genesis in a SSZ environment (e.g. Sepidbar et al., 2019; Azizi et al., 2018a, 2018b) where a mantle source has been contaminated by fluids formed by dehydration of subducted oceanic crust and partial melting of subducted sediments (e.g. Pearce and Parkinson, 1993; Hawkesworth et al., 1997; Mac Donald et al., 2000; Pearce and Stern, 2006).

Gabbros show REE- and trace-patterns distributed between DMM and N-MORB (Fig. 6A, B), similar to low-Ti IAT from SSZ environments (Saccani et al., 2018a) and falling in the forearc basalt (FAB, Reagan et al., 2010; Ishizuka et al., 2011) field on a tectonic discrimination diagram. In contrast, diabasic dikes show (Fig. 6C, D) REE- and trace-patterns distributed between the N-MORB and E-MORB, similar to high-Ti IAT from SSZ (Saccani et al., 2018a) and falling in the forearc basalt (FAB, Reagan et al., 2010; Ishizuka et al., 2011) field on a tectonic discrimination diagram.

6.2. Constraints on partial melting, melt-peridotite interaction, and metasomatic processes

Mineral and whole-rock geochemical compositions of mantle peridotites can be used to decipher the extent of partial melting and melt/fluid phase enrichment and mantle-melt interaction (e.g. melt percolation) processes subsequent to melt extraction (e.g. Khedr etal., 2010, 2014, 2022, 2023). Based on the mineral major-element compositions, it is clear that the Sikhoran peridotites underwent differing degrees of partial melting (Jonnalagadda et al., 2019). Relationships among spinel Cr# vs. Mg# and olivine Fo contents (Fig. 9A, B) indicate that Sikhoran dunites belonging to the porphyroclastic sequence are mantle residua that formed after ~40 vol. % melt extraction from an original fertile MORB mantle (FMM; e.g. Khedr and Arai 2013, 2017, Khedr et al., 2014; Saccani and Tassinari, 2015), which corresponds to melting of orthopyroxene and clinopyroxene from primary mantle lherzolite. Such a high degree of partial melting can be also inferred from the whole-rock chemistry of peridotites, such as MgO versus Al_2O_3 (Fig. 10). Based on this relationship, the best fit for peridotite compositions suggests more than 35 vol. % melting for dunite, similar to Izu-Bonin–Mariana (IBM) forearc peridotites (Parkinson and Pearce, 1998). The residual mantle signature of the Sikhoran peridotites is also implied by their whole-rock chemistry, which is characterized by enrichment in compatible elements (Ni: 1750–2800 ppm; Cr: 2000–5000 ppm; MgO: 42–50 wt. %) and depletion of incompatible elements (Na_2O < 0.1 wt. %; TiO_2 < 0.1 wt. %) (e.g. Ohara et al., 2002).

The Sikhoran dunites belonging to the porphyroclastic sequence have higher MgO (mean 46 wt. %) but lower CaO (mean 0.31 wt. %) contents than PM (MgO = 36.8–37.8 wt. % and CaO = 3.2–3.7 wt. %; McDonough and Sun, 1995; Workman and Hart, 2005) and of DMM (MgO = 37.8 wt. % and CaO = 3.55 wt. %; McDonough and Sun, 1995; Workman and Hart, 2005). This indicates that they are a refractory residue that formed from a depleted mantle source after high degree of partial melting. Enriched MgO/SiO_2 ratios (mean = 1.23) and depleted Al_2O_3/SiO_2 ratios (mean = 0.003) (Supplementary Table 2) in Sikhoran dunites samples compared to PM (MgO/SiO_2 = 0.8; Al_2O_3/SiO_2 = 0.1; McDonough and Sun, 1995; Workman and Hart 2005) and DMM (MgO/SiO_2 = 0.84.; Al_2O_3/SiO_2 = 0.10; McDonough and Sun, 1995) suggest that the Sikhoran dunites are mantle residues that formed following high degrees of partial melting and melt extraction (Barnes et al., 2014; Lian et al., 2019).

Spinel Cr# data plotted against the Fo content of olivine in transitional peridotites (Fig. 9B) also plot within the olivine–spinel mantle array (OSMA; Arai, 1994), whereas, their compositions show that Sikhoran peridotites belonging to the transitional sequence are mantle residua that formed after ~15–28 vol. % partial melting and melt extraction (FMM; e.g. Khedr and Arai, 2013, 2017, Khedr et al., 2014; Saccani and Tassinari, 2015), in agreement with the whole-rock chemistry of transitional peridotites (e.g. MgO versus Al_2O_3; Fig. 10A). Based on this relationship, the best fit for peridotite compositions suggests 15–30 vol. % melting for peridotite, which is characterized by enrichment in compatible elements (Ni: 210–2300 ppm; Cr: 1040–5200 ppm; and MgO: 19–46 wt. %) and depletion of incompatible elements (Na_2O < 0.3 wt. % and TiO_2 < 0.2 wt. %) (e.g. Ohara et al., 2002).

Spinel Cr# plotted against the Fo content of olivine in layered ultramafic peridotites (Fig. 9B) also lies close to the OSMA (Arai 1994), whereas, their compositions suggest that the Sikhoran peridotites belonging to the layered ultramafic sequence are mantle residua that formed after just ~5–22 vol. % melt extraction (FMM; e.g. Khedr and Arai, 2013, 2017, Khedr et al., 2014; Saccani and Tassinari, 2015). Such a low degree of partial melting is also inferred from the whole-rock chemistry, which suggests 5–25 vol. % melting for peridotite (Fig. 10). The residual mantle signature of the Sikhoran peridotites also is demonstrated by its enrichment in compatible elements (Ni: 330–2450 ppm; Cr: 1780–4300 ppm; and MgO: 22–

44 wt. %) and depletion of incompatible elements (Na_2O up to 0.3 wt. % and TiO_2 < 0.2 wt. %) (e.g. Ohara et al., 2002).

Lherzolites and harzburgites within the layered ultramafic zone show more abyssal type peridotite signatures, and underwent depletion in a MOR system (Figs. 7C, 8B, 10), while the dunite and wehrlites in the transitional zone sequence show compositional overlap with abyssal and forearc peridotites (Figs. 7A-C, 8B, 10), and supra-subduction zone (SSZ) peridotites owing to the relatively high degree of melting recorded by their whole-rock and mineral chemistry. Dunite-chromitites of the porphyroclastic zone and chromitite within the transitional zone show geochemical signatures that better resemble fore-arc type peridotite (Figs. 7A-C, 8B, 10). Therefore, it can be suggested that the ultramafic-mafic parts of the Sikhoran complex were affected by boninitic magmas and magmatic fluids generated during early proto-forearc spreading. The mantle ultramafic part is very similar to the other known subcontinental ultramafic mantle plumes, such as Tinaqullo in Venezuela (Choi et al., 2007), Ronda in Spain (Lenza et al., 2010) and Lherz in France (Le Roux et al., 2007; Uzel et al., 2020) and its layered ultramafic-mafic igneous part is very similar to Carboniferous mafic-ultramafic Pulur intrusion in the Eastern Pontides, Turky (Eyuboglu et al., 2010; Tupuz et al., 2023).

Mantle peridotites containing spinels with Cr# > 0.70 are believed to form due to high degrees of partial melting, which generates pyroxene-free dunites (Kostopoulos, 1991). However, the Sikhoran porphyroclastic dunites contain spinel with high Cr# 0.71–0.84 and low TiO_2 contents (0.05–0.27 wt. %), along with minor interstitial clinopyroxene that may have formed from fertilization processes (i.e. magma percolation), as proposed by Khedr etal., 2014 and Khedr et al., 2023. Relationships between Cr# and TiO_2 in the studied spinel were investigated in order to determine if the peridotites were affected by metasomatic and/or partial melting processes (Fig. 9C). The Cr# from porphyroclastic dunites-chromitite and transitional chromitites show a positive correlation when plotted against the TiO_2 content of spinel after degrees of partial melting of greater than 20%. By contrast, the Cr# values from transitional dunite–wehrlite and layered lherzolite–harzburgite are also positively correlated, but only at low degrees of partial melting (~15 %), indicating that all host type rocks were likely affected by late-stage fluid/melt metasomatism (magma percolation and solid-melt reaction during Jurassic simultaneously with the formation of isotropic gabbro and diabase dikes) (Fig. 9C).

6.3. Is the Sikhoran ultramafic-mafic complex a polygenetic, partially melted mantle plume?

The Sikhoran ultramafic-mafic complex appears to be a polygenetic complex, completely different from most ophiolitic sequences. It has at least four different parts comprising of a mantle residual tectonite (part 1) transitionally changed to a thick sequence of ultramafic-mafic layered cumulates (part 2). Layered gabbros of this cumulative part have a Late Carboniferous age, whereas Late Triassic-Early Jurassic Abshour isotropic gabbros (part 3) intruded ultramafic-mafic layered cumulates (part 2) and induced very high grade contact metamorphism (sanidinite facies) at their contact with the host Upper Paleozoic amphibolites. Sporadic, pegmatoid gabbro and diabasic dikes belong to Permian, Triassic, Late Jurassic and Late Cretaceous (part 4) cut the Sikhoran complex. Parts 1 and 2 may represent an ascending partial melting mantle diapir (plume) intruded in the base of the crust in a rift setting during initial stage of opening of the Zagros Neotethys oceanic basin during the Upper Carboniferous. This intrusion event was associated with the HT/LP rift-type metamorphism in the Upper Carboniferous metamorphic rocks, which produced S-type granites that intrude gneisses. Carboniferous mafic-ultramafic intrusions of the Pulur Complex in the Eastern Pontides (Eyuboglu et al., 2010; Topuz et al., 2023) and the Carboniferous ultramafic-mafic complex of Misho, NW Iran (Saccani et al., 2013) show many similarities to the layered ultramafic-mafic part of the Sikhoran Complex. The Abshour gabbro (part 3) intruded into the ultramafic-mafic layered cumulates(part 2) during the Late Triassic-Early Jurassic, and the fourth part of the complex formed between the Permian and the Late Cretaceous.

Based on these geological relations, the rock association, ages and structures of the Sikhoran complex show key differences from typical Penrose-type ophiolites, and have key similarities with mantle plumes and ultramafic-mafic intrusions. Firstly, field observation and geochemical and geochronological data show that the gabbros in the Sikhoran complex appear to have formed as intrusions cutting a pre-existing ultramafic-mafic complex, suggesting that the gabbros are not related to the ultramafic-mafic complex. Secondly, juxtaposition of the ultramafic-mafic rocks against schist and amphibolites could have resulted from intrusion of the ultramafic-mafic complex into the metamorphic country rocks. A foliation-parallel contact between amphibolite and ultramafic rock, internal foliation of the larger ultramafic bodies, and location of the ultramafic rocks along strike of the regional foliation support mantle plume emplacement. These features also agree with new Jurassic (187.2 Ma) U–Pb ages for the amphibolite that is juxtaposed with this ultramafic complex. Finally, our study indicates that Carboniferous magmatism is also widespread in the Sikhoran region and is temporally related to occurrences of Hercynian magmatism elsewhere in Iran, which was probably related to the extensional tectonic regime responsible for the rifting of the Cadomian blocks from northern Gondwana to form the Zagros Neotethyan basin. The dated layered gabbro (sample SF18-18, 320.8 Ma) associated with the ultramafic-mafic complex has a different age from isotopic gabbro (sample SF18-13, 178.3 Ma), suggesting that the ultramafic-mafic complex formed in a different tectonic regime that was responsible for the rifting of the Cadomian fragments from northern Gondwana and opening of the Neotethys basin during the Middle-Late Paleozoic (Moghadam et al., 2015b; Jafari et al., 2023). This agrees with their geochemical affinities, including REE- and trace-patterns distributed between DMM and N-MORB, similar to low-Ti IAT from a SSZ geodynamic environment.

Our thermobarometric results indicate that the Sikhoran ultramafic-mafic rocks equilibrated under a wide range of conditions, and that some high-temperature ultramafic rocks have been re-intruded upwards through several kilometers of crust. In such conditions, mantle rocks are considered as parts below the Moho transition zone. The textures of rocks likely formed due to large volumes of high-temperature fluids generated at the main thrust zone (MTZ) infiltrating the dunites. These fluids would be highly reactive and so produced porphyroclastic textures. The MTZ is considered a weak surface due to the existence of shear-related deformation and fractures, and so allows channelization or conduit-assisted fluid flow and percolation; such fluids would have collected and accumulated olivine or orthopyroxene as clots or grain aggregation, forming porphyroclastic textures

Some models imply that a peridotite of any origin becomes an mantle plume if it is involved in regional metamorphism and associated deformational processes. O'Hara (1967c) suggested that the only feature that these such rocks have in common is that they occur in an Alpine setting, which allowed tectonic transport and re-intrusion. However, Sorensen (1967) reviewed evidence that peridotite and garnet peridotite, when enclosed in amphibolite, could be the products of metamorphic differentiation in zones of stress concentration. As mentioned above, the foliation orientations in the amphibolite and Sikhoran ultramafic complex match, which we use to argue that this complex has more similarities to mantle plumes than ophiolitic complexes.

6.4. ***Origin and emplacement of the Sikhoran ultramafic-mafic complex***

The mechanism by which ultramafic-mafic rocks can become interleaved within schist/amphibolite is poorly understood. One model suggests that during the underplating process, sedimentary rocks are ductilely deformed and metamorphosed to become schist/amphibolite. The direction of underplating or underthrusting can be deduced from the orientation of mineral lineation and sense of shear derived from rotated plagioclase porphyroclasts and shear bands in mylonitic schist/amphibolite. Tectonic mantle plume upwelling of the Sikhoran complex in an extensional rift setting occurred during the Late Carboniferous (Fig. 11A). This is constrained by the ages of layered gabbro samples (this study), Upper Paleozoic amphibolites, gneisses and S-type granites (Table 1), which formed at c. 334 Ma (Ghasemi, 2000; Ghasemi et al., 2002, 2004; Moghadam et al., 2017; and this study). Yet, later tectonic events that occurred in the Late Triassic-Early Jurassic to Late Cretaceous caused emplacement of the complex into the rock collages of the SSMMZ at the southern margin of the Central Iran block (Fig. 11B). The tectonic setting of the Sikhoran ultramafic complex is comparable to very similar subcontinental ultramafic mantle plumes, such as Tinaqullo in Venezuela (Choi et al., 2007), Ronda in Spain (Lenza et al., 2010) and Lherz in France (Le Roux et al., 2007; Uzel et al., 2020) and the layered ultramafic-mafic igneous part is very similar to Carboniferous mafic-ultramafic Pulur intrusion in the Eastern Pontides, Turky (Eyuboglu et al., 2010; Tupuz et al., 2023). Recently, many studies have emphasized the importance of replacement in the form of rising and melting mantle plumes in extensional rift setting (e.g., Choi et al., 2007; Le Roux et al., 2007; Pirajno, 2007, 2022; Puchkov, 2009; Lenaz et al., 2010; Dannberg, 2016; Ernst et al., 2019; Uzel et al., 2020; Srivastava et al., 2022).

The SSMMZ began to open as an intracontinental rift during the Late Carboniferous (Ghasemi, 2000, Ghasemi et al., 2004; Saccani et al., 2013; Azizi et al., 2017; Jafari et al., 2023) (Fig. 11A), producing mafic-ultramafic mantle rocks (Fig. 11A) which remained in crustal sections until the Late Triassic. These rocks were incorporated into a metamorphic-magmatic arc since the Late Triassic-Early Jurassic (Fig. (Berberian and King, 1981; Ghasemi, 2000; B) Ghasemi et al., 2004), with peak activity in the Middle-Late Jurassic (Shahbazi et al. 2010; Azizi et al. 2011). Less overprinting is documented in the Late Cretaceous, and final closure occurred in the Late Cretaceous (Ghasemi, 2000; Ghasemi et al., 2002, 2004). Our geochemical data for these mantle rocks suggest that lherzolites and harzburgites of layered ultramafic zone and dunites of transitional zone lie close to the fields of MOR peridotites generated in an extensional environment or plot in the overlapping space of MOR peridotites and forearc peridotites (Fig. 7, 9). This supports that the presence of an extensional environment for Late Triassic-Early Jurassic during early proto-forearc spreading accompanying subduction initiation (Figure 12a), which may have affected and/or metasomatized previous Carboniferous ultramafic rock. Our data also indicate that forearc peridotites have a stronger boninitic signature than back-arc peridotites, as expected for subduction-initiation peridotites. Chromian spinels in the dunite envelopes and chromitites of porphyroclastic zone are geochemically similar to spinels that formed in both boninite-forming late SI and in a mature SSZ setting (Figures 7, 9; Kamenetsky et al. 2001; Parkinson and Pearce 1998; Pearce et al. 2000; Khedr and Arai 2013, Khalatbari Jafari et al. 2016, Khedr and Arai 2017). This indicates an SSZ origin for the change in the geochemical features of these rocks (e.g. Khedr and Arai 2016, 2017). Spinel compositions of peridotites and associated chromitites plot near the Cr–Al line on the Cr–Al– Fe ternary diagram and show variable Cr and low Fe^{3+} contents (Fig. 7a). Such high-Al (for layered and transitional zone) and high-Cr spinels (for porphyroclastic zone) can metasomatize Carboniferous ultramafic-mafic rocks during two stages; (i) early subduction initiation for metasomatism of layered lherzolites-harzburgite and transitional dunites, which are most consistent with a FAB-like melt; (ii) late subduction initiation stage for formation of porphyroclastic dunite-chromitites, which is most consistent with a boninite-like melt.

All of these criteria suggest that lherzolites and harzburgites were re-fertilized due to percolating MOR-like tholeiitic melts beneath a proto-forearc spreading center during subduction initiation and trench rollback stage (Fig. 11B), producing isotropic gabbro, whereas the porphyroclastic dunites reflect more extensive metasomatism as a result of a more hydrous environment (Fig. 11C) with a boninite-like signature, which are responsible for making E-MORB diabasic dikes. This supports an evolutionary model in which the opening and closing of the Zagros Neotethyan Ocean in the Esfandagheh region during the Late Carboniferous (This study) and Late Cretaceous respectively, led to incorporation of the Late Neoproterozoic to Late Cretaceous rock units as metamorphic-magmatic-sedimentary collages with tectonic juxtaposition in the SSMMZ.

7. Conclusions

(1) Our new U–Pb zircon ages suggest that several generations of rocks are present in the Sikhoran ultramafic–mafic rocks. Two populations of ages identified: a Carboniferous (c. 330 Ma) event, likely documenting magmatism and metamorphism and the production of ultramafic–mafic complex (layered gabbros), amphibolites, gneisses and S-type granites, and an Early Jurassic (c. 187 Ma) event, characterized by metamorphism and magmatism (isotropic gabbro intrusion).

(2) Inter-foliation with schist/amphibolite, similar ductile fabrics in schist and ultramafic rocks, lack of internal zoning, and lack of intrusive relations with isotropic gabbros along with the Late Carboniferous ages of layered gabbros, suggest that the ultramafic-mafic

components of the Sikhoran complex are unrelated with isotropic gabbros, signifying that they are polygenetic ultramafic-mafic products associated with an ascending and partially melting mantle plume.

(3) We conclude that ultramafic-mafic (layered gabbro) rocks within the Sikhoran complex represent remnants of a subcontinental mantle plume associated with a rift basin onto which the sediments of the proto-Sikhoran were deposited. The basin began to develop during the Late Carboniferous, transformed to a sea floor spreading center during the Permian-Triassic, and collapsed during the Mesozoic in association with subduction that initiated during the Late Triassic.

(4) All fragments, including ultramafic-mafic components and metasedimentary rocks were accreted onto the overriding Iranian plate as the metamorphic-magmatic assemblages of the SSMMZ during the Mesozoic and the final closuring of the Esfandagheh Oceanic basin as a part of larger Zagros Neotethys oceanic system during the Late Cretaceous.

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10. Figure caption

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

Fig. 2. Simplified geological map of the Sikhoran mafic ultramafic complex. Modified after Asadi et al. (2022).

Fig. 3. (A) Tectonite zone including porphyroclastic dunite and harzburgite; (B) foliated porphyroclastic harzburgite in the tectonite zone; (C) porphyroclastic harzburgite comprised of olivine and orthopyroxene, and spinel with mechanical deformation and extinction; (D) porphyroclastic dunite comprised of olivine, and spinel with mesh texture of serpentine; (E) concordant foliation between dunite and chromitites in the transitional zone; (F) enclaves of pyroxenites in the intruding isotropic gabbro; (G) layered ultramafic cumulates (mostly cumulative dunite) zone; (H) Jurassic- Cretaceous diabasic dikes cutting the layered gabbros; (I) Late Carboniferous layered gabbros of layered cumulative ultramafic-mafic sequence of the Sikhoran complex; (J) isotropic gabbro intruded in the Sikhoran Complex and Upper Paleozoic metamorphics; (K) melting of Upper Paleozoic amphibolites in contact with Early Jurassic isotropic gabbro and formation of very large (meter-scale) amphibole and plagioclase in plagiogranite.

Fig. 4. CL images, concordia diagrams, and weighted mean $^{206}Pb/^{238}U$ age plots for investigated zircons from (A-C) layered gabbro, (D-F) isotropic gabbro, (G-I) gneiss and (J-L) plagiogranite, respectively.

Fig. 5. Peridotite compositions for the Sikhoran Complex, recalculated to 100% on a volatile-free basis. Data are shown on diagrams of (A) Al_2O_3/SiO_2 vs MgO/SiO_2 (after Jagoutz et al. 1979; Hart and Zindler, 1986; mantle depletion trend, fields of abyssal peridotites, and forearc peridotites are from Pearce et al., 1992). Depleted MORB Mantle (DMM: Workman and Hart 2005), primitive mantle (PM; McDonough and Sun 1995), abyssal peridotite (Niu

2004), forearc peridotite (Parkinson and Pearce 1998), metasomatism/seafloor weathering trend (Paulick et al. 2006; Bhat et al., 2019; Nouri et al. 2019), interaction with olivine-rich melts (Bhat et al., 2019), and Terrestrial Array (Jagoutz et al. 1979) are shown; (B) total alkali versus silica (TAS) diagram (LeMaitre et al. 2002); and (c) Th vs Co diagram (Hastie et al. 2007). (D) Th/Yb vs. Nb/Yb diagram. Rock type: B, basalt; BA/A, basaltic andesites/andesites; CA, calc-alkaline series; D/R, dacite/ rhyolite; H-K, high-K series; IAT, island arc tholeiites. Petrological group and references: BABB, back-arc basalts (Pearce et al. 2005; Buchs et al. 2013); FAB, forearc basalts (Reagan et al. 2010; Ishizuka et al. 2011); Tholeiite-OIB, ocean island basalts (Willbold and Stracke 2006; Buchs et al. 2013); E-MORB, enriched mid-ocean ridge basalts (after Jenner and O'Neill 2012; Azizi et al. 2018a); 85% probability contour of IAT composition (after Hastie et al. 2007).

Fig.6. Chondrite-normalized rare earth elements (REEs) diagrams (Sun and McDonough 1989) (a, c) and primitive mantle normalized multi-elements pattern diagrams (Sun and McDonough 1989) (b, d) for the Sikhoran gabbros, and diabases. Data source for average depleted MOR mantle (DMM, Workman and Hart 2005), and for OIB, N-MORB, and E-MORB (Sun and McDonough 1989), FAB forearc basalts (after Reagan et al. 2010; Ishizuka et al. 2011), BABB back-arc basalts (Pearce et al. 2005; Buchs et al. 2013).

Fig. 7. Mineral chemistry of Sikhoran ultramafic-mafic rocks. (A) Spinel compositions: in the Al-Cr-Fe^{3+} ternary diagram (Melluso et al., 2014). Field of abyssal peridotites (after Arai 1994; Arai et al. 2011; Khedr and Arai 2017) and forearc (SSZ) peridotites (after Ishii et al. 1992; Khedr and Arai 2017). (B) Olivine compositions: XFo vs NiO diagram (after Pagé et al. 2008). The olivine-mantle array is after Takahashi et al. (1987); the field of olivine compositions in forearc peridotites is after Pagé et al. (2008, and references therein). (c) clinopyroxene compositions: Mg# vs Al_2O_3 diagram (after Choi et al. 2008). The fields for clinopyroxene chemistry from abyssal and SSZ peridotites are after Choi et al. (2008).

Fig. 8. Amphibole chemistry of Sikhoran ultramafic-mafic rocks shown for the Mg/(Mg+Fe) vs. Si and TiO_2 vs. Na_2O binary diagrams.

Fig. 9. (A) Spinel Cr#-Mg# binary diagram (after Melluso et al., 2014); (B) XFo (in Ol) vs Cr# (in Spl) diagram (after Arai 1994). The olivine-spinel mantle array (OSMA), the melting trend (with indication of % of partial melting); (C) TiO_2 (in Spl) vs Cr# (in Spl) diagram (after Tamura and Arai 2006).

Fig. 10. Major oxides versus MgO vs. Yb variation diagrams for the bulk-rock chemistry of the Sikhoran. Note systematic correlations, reflecting partial melting between 15 and 25% by either isobaric batch melting (broken line) or near-fractional polybaric melting (solid line) (e.g., Niu, 1997).