Subduction initiation and back-arc opening north of Neo-Tethys: Evidence from the Late Cretaceous Torbat-e-Heydarieh ophiolite of NE Iran

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ABSTRACT

How new subduction zones form is an ongoing scientific question with key implications for our understanding of how this process influences the behavior of the overriding plate. Here we focus on the effects of a Late Cretaceous subduction-initiation (SI) event in Iran and show how SI caused enough extension to open a back-arc basin in NE Iran. The Late Cretaceous Torbat-e-Heydarieh ophiolite (THO) is well exposed as part of the Sabzevar-Torbat-e-Heydarieh ophiolite belt. It is dominated by mantle peridotite, with a thin crustal sequence. The THO mantle sequence consists of harzburgite, clinopyroxene-harzburgite, plagioclase lherzolite, impregnated lherzolite and dunite. Spinel in THO mantle peridotites show variable Cr# (10-63), similar to both abyssal and fore-arc peridotites. The igneous rocks (gabbros and dikes intruding mantle peridotite, pillowed and massive lavas, amphibole gabbros, plagiogranites and associated diorites and diabase dikes) display rare earth element patterns similar to MORB, arc tholeiite and back-arc basin basalt. Zircons from six samples, including plagiogranites and dikes within mantle peridotite, yield U-Pb ages of *ca* 99-92 Ma, indicating that the THO formed during the Late Cretaceous and was magmatically active for ~7 Myr. THO igneous rocks have variable εNd(t) of +5.7 to +8.2 and εHf(t) ranging from +14.9 to +21.5; zircons have εHf(t) of +8.1 to +18.5. These isotopic compositions indicate that the THO rocks were derived from an isotopically depleted mantle source similar to that of the Indian Ocean, which was slightly affected by the recycling of subducted sediments. We conclude that the THO and other Sabzevar-Torbat-e-Heydarieh ophiolites formed in a back-arc basin well to the north of the Late Cretaceous fore-arc, now represented by the Zagros ophiolites, testifying that a broad region of Iran was affected by upper-plate

extension accompanying Late Cretaceous subduction initiation.

KEYWORDS: Hf-Nd isotopes, U-Pb geochronology, back-arc basin, ophiolite, Iran.

INTRODUCTION

Ophiolites are fragments of upper mantle and oceanic crust (Coleman, 1977; Nicolas, 1989) and are often tectonically incorporated into continental margins during continent-continent and arc-continent collisions (Dilek and Flower, 2003), ridge-trench interactions (Lagabrielle et al., 2000), and/or subduction-accretion events (Cawood, 1989). Ophiolites and their dismembered equivalents are especially common in suture zones in both collisional and accretionary orogens of Neoproterozoic and Phanerozoic age (e.g., (Furnes et al., 2014; Hébert et al., 2012; Saccani et al., 2013). (Dilek and Furnes, 2011) classified ophiolites into continental margin-, mid-ocean-ridge (MOR)-, plume-, supra-subduction zone-, volcanic arc- and accretionary- types according to their tectonic settings. Evidence is growing that many ophiolites form during subduction initiation in a fore-arc setting (Whattam and Stern, 2011), and this concept is useful for understanding the Late Cretaceous ophiolites of the Zagros mountains in Iran (Moghadam and Stern, 2011; Moghadam et al., 2010).

The emplacement of an ophiolite on continental crust can occur via various mechanisms, including obduction, subduction and then exhumation of oceanic lithosphere or even through trapping and compression of oceanic basins between two neighboring continental blocks, as is the case for NE Iran ophiolites. Despite the fact that it is geodynamically difficult to emplace ophiolites that form anywhere other than the fore-arc (Stern, 2002), ophiolites are known to form in other extensional settings such as marginal basins or back-arc basins. Back-arc basins need strong extension to open and this can be achieved as a result of subduction initiation (SI). Back arc extension is not as easily accomplished if regional compression causes induced SI but is expected to accompany spontaneous SI, due to the large lateral density contrasts across the lithospheric weakness separating the pertinent plates (Stern, 2004; Stern and Gerya, 2017). Spontaneous SI, (i.e., when oceanic lithosphere begins to sink vertically before down-dip motion is established), will cause extension in the overlying plate, and this can open a back-arc basin. How subduction initiation causes extension in the overlying plate also depends on plate configuration and relative motions, thickness of the subducting oceanic slab, and whether SI is spontaneous or induced.

Marginal basin is a general term for basins of unknown origin associated with continental margins and island arcs. Back-arc basins are a type of marginal basin that open at a convergent plate margin behind an active magmatic arc. Most back-arc basins are associated with oceanic arcs, although rifting of continental lithosphere also produces back-arc basins (i.e. ensialic back-arc basins) (Gerya, 2011; Stern, 2002). Back-arc ophiolites are represented by MORB-like, BABB-like and calc-alkaline type intrusions and lavas, generated by partial melting of a MORB-type mantle source (through mantle wedge flow; e.g., (Long and Wirth, 2013)) into the sub-arc mantle wedge (Saccani et al., 2008). In many cases, the tectonic setting of ophiolites is obscured by alteration and deformation, so that workers rely heavily on relict mineralogy and immobile trace element geochemistry to resolve the original tectonic setting. This approach is problematic because BAB and fore-arc ophiolites share many geochemical and petrographic features and can be difficult to distinguish solely on a basis of petrology and geochemistry; other considerations are required to distinguish back-arc and fore-arc ophiolites. This report addresses this problem by focusing on a well-preserved ophiolite belt in NE Iran that clearly formed in a back-arc basin distal to coeval ophiolites that formed in a fore-arc, subduction-initiation (SI) setting. Our research supports results from recent IODP drilling in the Philippine Sea indicating that extensional magmatism related to subduction initiation can affect much larger regions than just the immediate fore-arc environment (Hickey-Vargas et al., 2018).

The Sabzevar-Torbat-e-Heydarieh ophiolites (= STHO) are excellent examples of Late Cretaceous ophiolites of Iran. The STHO is exposed over a large area of NE Iran, approximately 130 x 400 km (Fig. 1). From a regional-tectonic perspective, the STHO is situated in a back-arc setting, separated by ~600km from the Zagros fore-arc ophiolites (Moghadam and Stern, 2011; Monsef et al., 2018), the Urumieh-Dokhtar Magmatic Belt (UDMB) and continental crust of the Lut block (Fig. 2). The faulted contact between the southernmost STHO ophiolites and Ediacaran-Cambrian (Cadomian) basement to the south is exposed on the southern margin of the STHO (Fig. 3). Cadomian basement comprises granitoids, gabbros, rhyolites and pyroclastic rocks with ages of 531-553 Ma (Moghadam et al., 2017b). There is no suture to which STHO ophiolites are clearly related, which probably were exhumed during latest Cretaceous-Paleogene formation of the Oryan Basin (Fig. 3). This is also confirmed by low-temperature thermo-chronometric and structural evidence, which show that uplift of the ophiolite domain was accompanied by dextral transpression beginning in early Paleocene (~60 Ma) and ended in the Miocene-Pliocene (5 Ma) (Tadayon et al., 2018). This conclusion is supported by the presence of an early Paleocene unconformity and deposition of middle Paleocene conglomerates which seal the STHO ophiolites. These conglomerates were produced by the erosion of the exhuming ophiolites during dextral transpression (Tadayon et al., 2018).

The modern episode of subduction of the Zagros orogen subducted Neotethys and began at 104-98 Ma, as determined by zircon U-Pb dating of Zagros fore-arc ophiolites (Moghadam et al., 2013; Moghadam and Stern, 2015). Late Cretaceous subduction initiation was accompanied by a major change in the direction and velocity of Arabia-Eurasia convergence (Agard et al., 2007). We suggest that Late Cretaceous spontaneous SI caused the observed change in plate motion; that subsidence of Mesozoic Neotethys lithosphere commenced along a transform margin adjacent to buoyant Eurasian lithosphere, causing proto-forearc seafloor spreading and producing the Zagros ophiolites (Moghadam and Stern, 2011; Moghadam et al., 2010). Foundering Neotethys oceanic lithosphere induced extension on the southern margin of Eurasia, producing back-arc basins, tilting of crustal blocks, uplift and intense erosion of the Iranian plateau, as well as core complex exhumation in some parts of the Iranian plateau (Malekpour-Alamdari et al., 2017; Moritz et al., 2006; Verdel et al., 2007). For testing the Late Cretaceous SI model for southern Eurasia, it is essential to understand the significance of the Late Cretaceous back-arc basins in NE Iran, of which the Sabzevar-Oryan-Cheshmehshir-Torbat-e-Heydarieh ophiolite is the largest representative (Fig. 3). Besides the suspected back-arc basins in NE Iran, there are several other Late Cretaceous oceanic basins (ophiolites) whose opening seems to be related to the Late Cretaceous extension within the Iranian plateau caused by SI along the Zagros suture zone. These ophiolites include the Khoy ophiolite in NW Iran (e.g, (Khalatbari-Jafari et al., 2004)) and scattered ophiolitic slices along the Talysh-Qare Dagh mountains in NNW Iran (e.g., (Burtman, 1994; Omidvar et al., 2018)).

We present here the results of our study of Late Cretaceous mantle peridotites, gabbros, lavas and plagiogranites from the Torbat-e-Heydarieh ophiolite. We provide zircon U-Pb ages for six samples of geochemically-distinct plagiogranites, gabbros and dikes, and compare these with geochronological data from the Sabzevar ophiolite to the NW (Fig. 3). Peridotite mineral compositions as well as major, trace and rare-earth element data and bulk rock Nd-Hf and zircon Hf isotopes for selected samples are used to constrain the tectonic setting of the different magmatic suites. We show that the mantle rocks and magmatic rocks were generated during back-arc basin extension and use these results and their magmatic-stratigraphic associations suggest that magmatism occurred in response to SW-directed rollback of a NE-dipping subduction zone during SI (Fig. 2).

GEOLOGICAL SETTING

The STHO is bordered to the north by the major Sangbast-Shandiz strike-slip fault delimiting the Binalud Mountains (Fig. 1). To the south, the STHO is bounded by the major Great Kavir-Dorouneh sinistral strike-slip fault (Figs. 1 & 3). The STHO comprises four main regions where ophiolites crop out (Fig. 3), separated by the Paleocene-Eocene Oryan sedimentary basin: (1) NNW of Sabzevar (the Sabzevar ophiolite, studied by (Moghadam et al., 2014b; Moghadam et al., 2015; Nasrabady et al., 2011; Omrani et al., 2013; Rossetti et al., 2014; Rossetti et al., 2010; Shojaat et al., 2003)); (2) SSW of Sabzevar (Oryan ophiolite, no published data); (3) Cheshmehshir in the far south (Maghfouri et al., 2016), and (4) north of Torbat-e-Heydarieh, the focus of this study.

The STHO lies between the Lut block in the south and the Kopeh-Dagh (Turan platform (Thomas et al., 1999) in the north. The STHO formed during the Late Cretaceous (Moghadam and Stern, 2015) and contains a well-preserved mantle sequence, but some parts are fragmented and sheared. There is consensus that the STHO formed above a subduction zone (-SSZ) setting, as most Sabzevar ophiolite lavas show appropriate geochemical signatures (Ghazi et al., 1997; Moghadam et al., 2014a; Shojaat et al., 2003). Sabzevar mantle peridotites have spinels with both MORB-like (Cr#<50%) and SSZ-type (Cr#>50%) signatures but most peridotite spinels have high Cr# (>50) with low TiO_2 and resemble those of SSZ peridotites (Moghadam et al., 2014a). Some Sabzevar ophiolite pillow lavas have OIB-like whole-rock compositions, suggesting a plume or subcontinental lithospheric mantle source. The Sabzevar ophiolite is covered by Late Campanian to Early Maastrichtian (~75-68 Ma) pelagic sediments. Sabzevar plagiogranites yield zircon U-Pb ages of 99.9-77.8 Ma (Moghadam et al., 2014a). Magmatic rocks of the Sabzevar ophiolite have positive εNd (t) values (+5.4 to +8.3) and most have high $^{207}Pb/^{204}Pb$, indicating a significant contribution of subducted sediments to their mantle source (Moghadam et al., 2014a).

The Torbat-e-Heydarieh ophiolite (THO) constitutes the southeasternmost of the four STHO outcrops, covering an area 60 km long and 50 km wide (Fig. 4). The THO shows no contact with older rocks and is unconformably overlain by Paleocene-Eocene conglomerates and sandstones. The THO is mostly composed of mantle peridotites including lherzolites, impregnated lherzolites, minor harzburgites and discordant dunites and chromitite lenses (Figs. 5, 6A). Harzburgites with 2-3 modal% clinopyroxene and plagioclase-lherzolites are also common. Dunites also occur as small veins and pods/lenses within the harzburgite. Diabasic-gabbroic-pyroxenitic dikes crosscut the mantle sequence and in most cases are boudinaged (Fig. 5). Diabasic-gabbroic dikes are locally converted into rodingite (Fig. 6B). Some diabasic dikes are metamorphosed to greenschist or lower amphibolite facies. Small intrusions of cumulate and fine-grained isotropic gabbro and pyroxenite (Figs. 6C, D) are common in harzburgites and lherzolites; pyroxene-rich gabbroic dikes are also common within these gabbros. Cumulate gabbroic lenses occasionally are layered, characterized by plagioclase-rich (anorthosite) and Opx- and Cpx-rich (olivine-beearing) bands (olivine-bearing melanocratic gabbros) (Fig. 6D). The gabbroic sequence is crosscut by diabasic and microgabbroic to more-evolved plagiogranitic dikes or small pockets (Fig. 6E). Some dikes within the gabbros also intrude the underlying peridotites. In some cases, angular gabbroic xenoliths are found within the plagiogranitic pockets (agmatite, Fig. 6F). Late plagiogranitic dikes were injected into both gabbros and underlying peridotites.

The THO crustal sequence is more poorly exposed than the mantle section but includes massive and pillow lavas (Fig. 6G), isotropic and coarse-grained massive gabbros and plagiogranite lenses/dikes within the gabbros (Fig. 5). The contact between the mantle sequence and crustal rocks in most cases is tectonized, but in some cases the mantle rocks are overlain by cumulate gabbros or by crustal lavas and pelagic limestones. The presence of pelagic limestone on top of the mantle section suggests that Late Cretaceous extension on the seafloor was locally amagmatic. Crustal gabbros differ from mantle gabbros by containing more diabasic dikes and large plagiogranitic lenses, but plagiogranitic dikes are also common. Lava flows are slightly metamorphosed to greenschist or lower amphibolite facies. Other rocks from both crustal and mantle sequences show no trace of metamorphism, but instead they show slight to moderate alteration. High-temperature alteration of mantle dikes into rodingite is also common. Pillow lavas show pyrite-chalcopyrite mineralization. Massive and pillow lavas are occasionally tectonically interlayered with peridotites. There is no obvious geochemical difference between most pillow and massive lavas (see next section). Late Cretaceous (Cenomanian to Turonian, ~ 99-90 Ma) pelagic sediments and pyroclastic rocks are interlayered with and conformably cover the lavas (Figs. 6H, I). Cold breccia including basaltic fragments set in a pelagic limestone matrix is abundant. THO melanges include serpentinites, lavas and pelagic sediments. These melanges are especially abundant in the Sabzevar ophiolites, where various ophiolitic units, Paleocene-Eocene magmatic rocks and shallow-water limestones are dispersed with serpentinites, showing that these melanges are related to Eocene or younger deformation (Moghadam et al., 2014a).

PETROGRAPHY

Lherzolites are the predominant THO mantle rocks; they contain serpentinized olivine, orthopyroxene and large clinopyroxenes (Fig. 7A). Nearly all lherzolite samples are moderately (20-30%) serpentinized. Vermicular brown spinel is abundant and generally intergrown with clinopyroxene. Harzburgites are minor and contain serpentinized olivine and orthopyroxene porphyroclasts with fine-grained clinopyroxene (2-3 %, Fig. 7C) and deep brown spinel. Serpentinization of harzburgites (50-60%) is greater than in lherzolites (20-30 %). The content of clinopyroxene is higher in Cpx-harzburgites (3-5 %). Here, fine clinopyroxenes occur in embayments of orthopyroxene porphyroclasts (Fig. 7B), associated with vermicular spinels (Fig. 8A). Part of the THO mantle section was impregnated with a percolating basaltic melt which filtered through peridotites, crystallizing plagioclase and clinopyroxene. Impregnated lherzolites have more coarse-grained clinopyroxene aggregates and lighter-brown spinel compared to normal lherzolites (Fig. 7C). Plagioclase lherzolites show traces of altered plagioclase (5-7 modal %) along with melt-impregnated clinopyroxenes (up to 20-30 %) and light brown vermicular spinel. Large clinopyroxenes enclose plagioclase and brown spinels (Figs. 6D and 7B). Plagioclase also occurs at the contact of orthopyroxene with impregnated clinopyroxenes (Fig. 8C). Blebs of sulfide minerals are also observed. Clinopyroxene accumulation in plagioclase lherzolites makes these rocks similar to olivine websterites (Fig. 9A). Plagioclase lherzolites show melt-assisted crystallization of new olivine grains in plagioclase and clinopyroxene embayments (Fig. 8C). Dunites are altered (50-70 % serpentinization) with mesh-textured olivines (>90%) and coarse-grained black spinel. Spinels contain abundant inclusions of serpentinized olivine, clinopyroxene and amphibole (Fig. 8D). Amphibole inclusions are altered to chlorite.

Gabbroic lenses within peridotites vary from gabbro (~1-2 % Opx) to gabbronorite (10-30 % Opx, Fig. 9B). Cumulate gabbros are heterogeneous, with dark bands rich in clinopyroxene, orthopyroxene and minor plagioclase and whitish bands rich in plagioclase with minor clinopyroxene. Olivine is rare, whereas coarse-grained, subhedral orthopyroxene and anhedral clinopyroxene are the main components of cumulate gabbros (Fig. 7E, F). Plagioclase is interstitial between pyroxene crystals (Fig. 8E, F). Fine-grained, isotropic gabbros containing plagioclase, orthopyroxene and clinopyroxene are also common in the mantle sequence, closely associated with cumulate gabbros. Occasional primary brown amphiboles also occur as interstitial laths between other minerals. Rodingitized dikes contain fine-grained hydrogrossular, rare diopside and wollastonite as well as pectolite.

Coarse-grained crustal plutonic rocks contain plagioclase and amphibole in diorites and clinopyroxene and plagioclase in gabbros. Some gabbros contain interstitial amphibole. Plagiogranites have granular texture and are dominated by plagioclase, quartz and amphibole with secondary epidote, chlorite, sericite, iron oxide and titanite. Apatite and zircon are minor components. Plagiogranites can be petrographically divided into tonalites and diorites with more amphibole. Granophyric intergrowths of plagioclase and quartz are also common in these rocks.

Diabasic dikes are fine-grained and contain magnesio-hornblende and sodic plagioclase. Pillow lavas contain clinopyroxene microphenocrysts, plagioclase microlites and palagonitized glass. Fine-grained clinopyroxene is also common in the groundmass. Calcite, chalcedony, titanite, sericite, epidote, prehnite and chlorite are secondary phases.

Pillow lavas display hyaloporphyritic to intersertal texture with clinopyroxene and plagioclase micro-phenocrysts. Massive lavas are similar to pillow lavas but are less altered. The glassy groundmass of some massive lavas is altered into chlorite and clay, but most are holocrystalline without glass.

ANALYTICAL METHODS

We used six main analytical procedures to study THO rocks: 1) JEOL wavelength dispersive electron probe X-ray micro-analyzer (JXA 8800R) to determine the composition of minerals; 2) Inductively Coupled Plasma-Atomic Emission (ICP-AES) and Inductively Coupled Plasma-Mass Spectrometry (ICP-MS) to determine whole-rock contents of major- and trace elements; 3) Cathodoluminescence (CL) imaging of zircons; 4) Sensitive High Resolution Ion Microprobe (SHRIMP) analyses to determine zircon U-Pb ages; 5) Laser Ablation Inductively Coupled Plasma-Mass Spectrometry (LA-ICPMS) analyses to determine zircon U-Pb ages and to analyze the trace-element compositions of clinopyroxene and orthopyroxene from mantle peridotites and gabbros; 6) Multi Collector-Inductively Coupled Plasma-Mass Spectrometry (MC-ICPMS) equipped with LA for *in situ* analysis of zircon Hf isotopic compositions. The same MC-ICPMS was used to analyze the isotopic compositions of Nd and Hf in whole-rock samples. We studied ~100 samples for petrography, 13 polished thin sections for electron microprobe analysis, 39 for whole-rock major- and trace-element compositions, 6 for U-Pb zircon ages, 8 for whole-rock Nd and Hf analyses, 6 for *in situ* mineral trace elements and 6 for *in situ* Lu-Hf isotope analyses of zircons. The Nd and Hf isotopic compositions discussed below are corrected for 100 Ma of radiogenic ingrowth. The details of each technique are given in Electronic Appendix A.

WHOLE ROCK GEOCHEMISTRY

Three mantle peridotites, fourteen mantle gabbros, ten pillowed and massive lavas, four amphibole gabbros, two mantle-intruding dikes, and six plagiogranites and associated diorites and diabase dikes samples were selected for whole rock analysis.

Sample locations are shown on Fig. 4. The analyzed samples are characterized by variable loss on ignition; 0.6-4.2 wt % for crustal rocks, 5.9-11.7 wt % for peridotites and 0.6-5.8 wt % for mantle gabbros. Because of the mobility of some major- and trace elements during alteration, emphasis is placed on immobile trace elements such as the REE and high field strength elements (HFSE) to evaluate the original composition and tectono-magmatic setting of Torbat-e-Heydarieh magmatic rocks. Fluid-mobile elements such as Cs, Rb, U, Pb, and Sr may be discussed but are generally de-emphasized in the following sections. Similarly, to avoid the effects of alteration and seawater exchange on isotopic composition of the rocks, alteration-resistant Nd and Hf isotopes are reported.

Mantle Units

Harzburgite has low contents of CaO and Al_2O_3 (~1.6 wt %), whereas Cpx-harzburgite and plagioclase lherzolite contain more CaO (~2.3 and ~8.9 wt % respectively) and Al_2O_3 (~2.3 and ~5.1 wt % respectively) (Supplementary Table 1). Harzburgites show highly depleted REE patterns, with steep slopes from LREEs to MREEs (Fig. 10A). REE abundances are higher in Cpx-harzburgite and plagioclase lherzolite with nearly flat patterns from MREEs to HREEs, but with steep slope from LREEs to MREEs. These patterns are similar to those of the cumulate gabbros (Fig. 10A). Enrichment in MREEs to HREEs for Cpx-harzburgite and plagioclase lherzolite is consistent with the formation of Cpx by metasomatism. Enrichment in fluid-mobile elements such as U, Rb, Ba, Sr, and Pb relative to LREEs is characteristic of THO mantle peridotites.

Mantle gabbroic rocks have variable contents of SiO_2 (~45-54 wt %), Al_2O_3 (~10-22 wt %), MgO (~4-16 wt %), and CaO (~10-20 wt %), reflecting different extents of fractionation and/or modal contents of olivine, orthopyroxene, clinopyroxene, and plagioclase. Gabbroic rocks are Ti-poor, with ~0.06-0.9 wt % TiO_2; a microgabbro dike within the gabbro cumulate contains more TiO_2 (0.9 wt %). Isotropic gabbros show an IAT signature in the Ti *vs* V diagram (Shervais, 1982) (Fig. 11B). Gabbroic rocks show three distinct patterns of rare earths and other trace elements (Fig. 10A and B). Cumulate gabbros have low REE concentrations compared to fine-grained isotropic gabbros and the microgabbro dike. Cumulate gabbros are highly depleted in LREE ($La_{(n)}/Yb_{(n)}$ = 0.04-0.23) but with nearly flat REE patterns from HREE to MREE ($Gd_{(n)}/Yb_{(n)}$ = 0.7-1.2). Eu enrichment is also noticeable. Extreme depletion in Nb, Ta and Zr (e.g., $Nb_{(n)}/La_{(n)}$ = 0.05-0.2) and enrichment in U, Pb, Ba and Sr (e.g., $Sr_{(n)}/La_{(n)}$ = 93-338) relative to LREE is characteristic of these rocks (Fig. 10B).

Isotropic gabbros are enriched in bulk REE compared to the cumulate gabbros and show LREE-depletion to almost flat patterns ($La_{(n)}/Yb_{(n)}$ = 0.3-0.7). These rocks are slightly depleted in Nb and Zr (e.g., $Nb_{(n)}/La_{(n)}$ = 0.3-0.9) and enriched in Sr, Th, and U (Fig. 10B). The microgabbro dike shows a flat REE pattern ($La_{(n)}/Yb_{(n)}$ = 1.2), weak negative anomalies in Nb-Ta and enrichment in Sr, Th, U, Ba and Rb.

Crustal rocks

The Torbat-e-Heydarieh massive lavas have similar contents of SiO_2 (~47-49 wt%) and TiO_2 (1.1-1.6 wt %) and Mg# (100Mg/Mg+Fe^{+2}) (54-63) (Supplementary Table 4). Pillow lavas show greater variability, ranging from 49-55 wt % SiO_2, Mg# values of 33-60, and TiO_2 contents of 0.4-2.4 wt %. These volcanic rocks exhibit both tholeiitic and calc-alkaline tendencies in the FeO^t/MgO *vs* SiO_2 diagram (Miyashiro, 1974) (Fig. 11A). In a Ti *vs* V diagram (Shervais, 1982), THO volcanic rocks tend to plot in both the island-arc tholeiitic (IAT) and MORB fields. Sample TH11-39 shows affinity to E-MORB-type lavas (Fig. 11B).

Massive lavas are depleted in LREE relative to HREE ($La_{(n)}/Yb_{(n)}$=0.5-0.8) and have a slight to moderate depletion in Nb ($Nb_{(n)}/La_{(n)}$=0.4-0.7). They are not enriched in Ba, Th and U ($Th_{(n)}/La_{(n)}$=0.7-1) relative to N-MORB (Figs. 10 C and D). These characteristics are similar to back-arc basin basalts or early arc tholeiites (Peate and Pearce, 1998). These lavas differ from Sabzevar calc-alkaline lavas, which are enriched in LREE, Th, U and extremely depleted in Nb (Moghadam et al., 2014a).

Pillow lavas can show both enrichment and depletion in LREE relative to HREE ($La_{(n)}/Yb_{(n)}$~0.6-2.7) as well as variable Nb-Ta depletion and Th enrichment (e.g., $Nb_{(n)}/La_{(n)}$=0.3-0.8 and $Th_{(n)}/La_{(n)}$=1.2-4.3) relative to N-MORB (Fig. 10 C and D). They are geochemically similar to E-MORB and depleted tholeiitic lavas. There is no clear relationship between these two types of pillow lavas in the field.

Plagiogranites and their host rocks

Plagiogranites and their host rocks (diorites and diabasic dikes) have variable SiO_2 (71-79 wt % for plagiogranites and 56-58 wt % SiO_2 for host rocks) and TiO_2 (0.2-0.5 wt %) contents (Supplementary Table 4). Plagiogranites and their host rocks have LREE-enriched and flat REE patterns respectively ($La_{(n)}/Yb_{(n)}$=0.8-3.7) with depleted Nb and Ta contents ($Nb_{(n)}/La_{(n)}$=0.3-0.9) accompanied by Ba, U, K, and Th enrichment ($Th_{(n)}/La_{(n)}$~2-9) relative to N-MORB (Fig. 10F). Plagiogranites also show strong positive anomalies in Sr and sometimes in Eu. These geochemical features are similar to Sabzevar plagiogranites (Moghadam et al., 2014a) and felsic igneous rocks of convergent plate margins (Pearce et al., 1984). These geochemical signatures are also similar to the geochemical characteristics of plagiogranites from other Tethyan ophiolites of the eastern Mediterranean realm (e.g., (Anenburg et al., 2015; Bonev and Stampfli, 2009; Dilek and Thy, 2006; Osozawa et al., 2012; Uner et al., 2014).

Mantle dikes

Diabasic and dioritic dikes injected into mantle peridotites have SiO_2~50-57 wt %, with Mg# ranging between 43 and 82 (Supplementary Table 4). Their TiO_2 contents are low, ranging from 0.1-1.2 wt %. In the FeO^t/MgO *vs* SiO_2 diagram, the dikes are broadly calc-alkaline, excluding sample TH11-19 (Fig. 11A). In a Ti *vs* V diagram (Shervais, 1982), these rocks plot in the IAT field (Fig. 11B). Diabasic dikes are characterized by flat REE patterns ($La_{(n)}/Yb_{(n)}$=1.1-1.8), depletions in Nb and Ta ($Nb_{(n)}/La_{(n)}$=0.3-0.5) and Ba, U, Sr, and Th enrichment ($Th_{(n)}/La_{(n)}$=2.6-3.2), resembling island arc tholeiites (IAT). Amphibole gabbros are characterized by an IAT signature in the Ti *vs* V diagram (Fig. 11B). They show LREE-enriched concave-upward patterns ($La_{(n)}/Yb_{(n)}$~2-10) and depletion in Nb and Ta ($Nb_{(n)}/La_{(n)}$~0.4-0.6).

MINERAL COMPOSITION

Major element compositions of the main rock-forming minerals were determined in mantle peridotites and cumulate gabbroic rocks: olivine, spinel, orthopyroxene, clinopyroxene, plagioclase and amphibole. Trace element contents of clinopyroxenes are also presented. These results are discussed below.

Olivine

Olivine in dunite has forsterite contents ranging from 92.8% to 93.1% ($Fo_{92.8-93.1}$) with high NiO content (~0.3-0.4 wt %) (Supplementary Table 1), typical for mantle olivine. Olivine in the harzburgites and Cpx-harzburgites is slightly less magnesian (but still mantle-like) with Mg# and NiO between 91.4-91.9 ($Fo_{91.4-91.9}$) and 0.34-0.43 wt % respectively. Lherzolites and impregnated lherzolites are characterized by olivine Mg# and NiO content of 89.5-91.3 ($Fo_{89.5-91.3}$) and 0.34-0.52 wt% respectively, also mantle-like. Olivine in plagioclase lherzolites has lower Fo and NiO contents ($Fo_{84.8-85.6}$ and 0.18-0.24 wt %) than commonly are found in mantle peridotite. These olivines have a magmatic composition (compared to high Mg# mantle olivines) and are interpreted to have crystallized from impregnating mafic melts.

Spinel

Peridotite spinels show variable Cr# (100Cr/(Cr+Al); 10-63), similar to those in abyssal, back-arc basin, and fore-arc peridotites (Dick and Bullen, 1984) (Fig. 12). Spinel from dunites has higher Cr# (59-63) and TiO_2 contents (0.18-0.25 wt %) and follows the peridotite-boninitic melt interaction trend of (Pearce et al., 2000a) (Figs. 12A and B). Spinels from harzburgites and Cpx-harzburgites have variable contents of TiO_2 (0.03-0.07 wt % in harzburgites and 0.1 to ~0.3 in Cpx-harzburgites) and FeO (~12.2-12.8 wt % in harzburgites and 14.2-15.5 wt % in Cpx-harzburgites). Harzburgite spinels have lower Cr# (22-24) than Cpx-harzburgite spinels (24-36).

Spinels of lherzolites and impregnated lherzolites have variable contents of TiO_2 (0.02-0.12 wt %) and Cr# (10-20). Spinel compositions of lherzolites and impregnated lherzolites overlap at low Cr#. This is because these two rocks have same mineral assemblages; impregnated lherzolites have more coarse-grained clinopyroxene aggregates with the lightest brown spinels. The Cr# of their spinels is similar to that of abyssal peridotites (Dick and Bullen, 1984). Spinels from plagioclase lherzolites have high TiO_2 (0.15-0.25 wt%) and FeO (~31-39 wt%) content and high Cr# (41-48).

Orthopyroxene

Orthopyroxene in harzburgite and Cpx-harzburgite has Mg# (91-92) and elevated contents of Al_2O_3 (2.7-5 wt %) and Cr_2O_3 (0.56-1 wt %) (Supplementary Table 1). Orthopyroxene in lherzolites and impregnated lherzolites has constant Mg# (89-91) but variable Al_2O_3 (~3-6.1 wt %) content. Plagioclase-lherzolite orthopyroxenes have low Mg# (86) and Al_2O_3 (2.1-2.3 wt %) compared to orthopyroxene from other THO peridotites. Orthopyroxene in gabbros has lower Mg# (79-84) and Al_2O_3 contents (1.5-2.1 wt %). The Al_2O_3 content of the THO peridotite orthopyroxenes are relatively high, ranging from 1.5-6.1 wt %; the higher values are encountered in impregnated lherzolites. However, the Al_2O_3 content of orthopyroxene from fertile spinel peridotites from ophiolites in other worldwide localities (e.g., the Vardar ophiolite, (Bazylev et al., 2009) and fertile peridotites from the Yarlung Zangbo suture zone, (Bedard et al., 2009)) and oceanic back-arcs (e.g., Mariana Trough, (Ohara et al., 2002) and South Sandwich arc-basin, (Pearce et al., 2000b)) are also high, resembling the THO impregnated lherzolites, and show the fertile nature of the host mantle rocks.

Clinopyroxene

Clinopyroxene is abundant in dunite as inclusions in large chromites. Its composition is $Wo_{49-51}En_{47-49}Fs_{2-3}$. It has high Mg# (94-97) and low TiO_2 (0.06-0.19 wt. %) and Al_2O_3

(0.6-~2 wt. %). Harzburgites and Cpx-harzburgites contain clinopyroxene with $Wo_{44-52}En_{48-49}Fs_{3-4}$. Clinopyroxene in Cpx-harzburgites have higher TiO_2 (0.26-0.35 wt. %), but lower Al_2O_3 (3.9-4.6 wt. %) compared to those in harzburgites (TiO_2=0.12-0.17 wt. %; Al_2O_3=3-5.2 wt. %) (Fig. 12C) with identical Mg# (92-94). Clinopyroxene in mantle lherzolites and impregnated lherzolites has variable composition ($Wo_{36-51}En_{45-59}Fs_{4-6}$) with high Mg# (91-93), Al_2O_3 (4.9-7.1 wt %) and TiO_2 (0.3-0.5 wt %) (Supplementary Table 1). Plagioclase lherzolites have clinopyroxene with more constant composition ($Wo_{41-48}En_{46-52}Fs_{6-7}$) but with low TiO_2 (0.06-0.16 wt %) and Al_2O_3 (2.2-3.8 wt %) and lower Mg# (87-89) compared to clinopyroxene from other THO lherzolites. On a TiO_2 *vs* Mg# diagram (Fig. 12C) clinopyroxenes from THO peridotites are similar to those found in both abyssal and supra-subduction zone peridotites. Clinopyroxene and spinel from nearly all mantle units of Tethyan ophiolites show both abyssal and supra-subduction-zone geochemical signatures (e.g., (Aldanmaz, 2012; Ghazi et al., 2010; Monsef et al., 2010)). This geochemical duality can be also observed in the crustal lavas of Neotethyan ophiolites (e.g., (Dilek and Furnes, 2009; Dilek et al., 2007; Sarifakioglu et al., 2009)).

Clinopyroxene in gabbro has variable composition ($Wo_{37-47}En_{45-51}Fs_{7-13}$). The more evolved compositions (with low Ca but high Fe) are found in smaller crystals between large clinopyroxenes. Gabbroic clinopyroxenes have TiO_2 and Al_2O_3 contents from 0.07-0.21 and 1.7-2.8 wt % respectively; these are lower than in Sabzevar gabbroic clinopyroxenes.

Plagioclase

Plagioclase in plagioclase lherzolites is highly altered; unaltered portions show very high An contents (96.2-98.3%). Gabbro plagioclase is also anorthite-rich (~An 91.1-98.2)

(Supplementary Table 1); amphibole-bearing gabbros with less orthopyroxene contain An_{91-92} plagioclase. Comparison of the clinopyroxene Mg# against anorthite content in plagioclase (Sanfilippo et al., 2013), shows a pattern similar to SSZ-type gabbros (Fig. 12D).

Amphibole

Analyses of amphibole inclusions within dunite chromites, plotted in the Mg# *vs* Si diagram (*not shown*), show that these are pargasite to edenite according to (Leake et al., 1997). The Cr_2O_3 contents of these amphiboles are high (2.4-3.1 wt. %).

Clinopyroxene: Trace Element Geochemistry

Trace-element abundances (including REEs) in THO lherzolites, impregnated lherzolites, plagioclase lherzolites and cumulate gabbros are listed in Supplementary Table 2.

All mantle clinopyroxenes from the Torbat-e-Heydarieh ophiolite are enriched in REEs with LREE-depleted patterns. They are similar to REE abundances in Cpx from mid-ocean ridge (MOR) cumulates and abyssal peridotites and are much more enriched that those in SSZ peridotites (Fig. 13A, C).

Chondrite-normalized REE patterns of clinopyroxenes from THO lherzolites are strongly depleted in light REEs (LREE) but are flat in middle to heavy REEs (MREE-HREE). These patterns are similar to those of Cpx in oceanic abyssal peridotites (Bizimis et al., 2000; Johnson et al., 1990), although total REE concentrations are greater (Fig. 11A). The REE content of THO lherzolite Cpx is even higher than Cpx from Sabzevar plagioclase lherzolites (Fig. 13B) (Shafaii Moghadam et al., 2015). Clinopyroxene trace-element patterns are depleted in Ti and Zr, but not in Sr and Eu (Fig. 13B). Clinopyroxene in THO impregnated lherzolites exhibits REE patterns similar to Cpx from THO lherzolites, characterized by higher total REE contents compared to Cpx from abyssal peridotites and Sabzevar lherzolites (Fig. 13C). Negative anomalies in Ti and Zr are also obvious (Fig. 13D).

Cpx in THO plagioclase lherzolites have REE patterns similar to those of Cpx from MOR cumulates (Ross and Elthon, 1993), but with more depleted LREE (La-Sm, Fig. 13C). These patterns are also similar to those for Cpx from Sabzevar Cpx-harzburgites. MREE-HREE abundances in Cpx of plagioclase lherzolite ar depleted compared to Cpx from THO lherzolites and impregnated lherzolites (Fig. 13C). Slight depletion in Ti and strong Zr negative anomalies are conspicuous in plagioclase lherzolites (Fig. 13D).

The REE abundances and patterns of clinopyroxenes in THO gabbro cumulates are in the range of those exhibited by MOR cumulates (Ross and Elthon, 1993) (Fig. 13A), but with more convex-upward MREEs with Sm-Ho peaks. They are similar to clinopyroxene patterns from Sabzevar cumulate gabbros but with lower REE abundances (Fig. 13A). The Cpx from cumulate gabbros are depleted in Zr and Ti but enriched in Sr (Fig. 13B). Trace element patterns of Cpx in cumulate gabbros are between those from Cpx of plagioclase and impregnated lherzolites.

Nd-Hf Isotopes

Eight samples were selected including depleted tholeiitic pillow lavas, back-arc basin-type massive lavas, supra-subduction zone-type plagiogranites, IAT-like diabasic dike and back-arc to fore-arc-type gabbros for analysis of whole-rock Nd and Hf isotopes. Hf-Nd isotope compositions are useful for inferring mantle source characteristics and the role of sediments and older continental crust in generating subduction-related magmas (Vervoort and Blichert-Toft, 1999b; Vervoort et al., 1999; Woodhead et al., 2012; Yogodzinski et al., 2010).

The εNd(t) and εHf(t) values for THO rocks vary between +5.7 to +8.2 and +14.9 to +21.5 respectively; gabbro shows especially high εHf(t) (+21.5) (Fig. 14A). All the samples plot near the field for modern Indian Ocean MORBs, suggesting a similar mantle source (Chauvel and Blichert-Toft, 2001b). Figure 13B shows the results of bulk-mixing calculations between mantle-derived MORB melts and various types of sediments (Chauvel et al., 2009). Involvement of subducted Fe-Mn rich sediments (as well as calcareous clays) in the mantle source affects Nd- more than Hf isotopic compositions (Fig. 14B). In contrast, addition of continental sands will affect Hf as well as Nd isotopic compositions, depending upon whether bulk assimilation of sediments or sediment melts are involved. Also, the different Hf/Nd ratios of sediments are expected to have a large influence, especially with the contribution of zircons in the subducted sediments.

In contrast to the Oman plagiogranites, which show strong involvement of sediment melts in their magmatic reservoir (Haase et al., 2015), the mantle sources of the Torbat-e-Heydarieh plagiogranites were not significantly affected by sediment melt addition or crustal assimilation. However, the mantle source of the Torbat-e-Heydarieh magmatic rocks could have been slightly affected by subducted pelagic sediments (high Nd/Hf ratio), because they vary considerably in εNd(t). They have less radiogenic Nd- isotope compositions than the igneous rocks of the Dehshir Late Cretaceous ophiolite (inner belt Zagros ophiolites, Fig. 1 (Moghadam et al., 2010)). The Nd-isotope compositions of the

THO magmatic rocks are similar to those of the Sabzevar ophiolite from NE Iran, which have high but variable εNd (t) values (+5.4 to +8.3) and moderately high $^{207}Pb/^{204}Pb$ ratios (15.50-15-65) (Moghadam et al., 2014a). The Nd-Hf-Pb isotopic compositions of the Sabzevar-Torbat-e-Heydarieh rocks suggest a minor contribution of subducted sediments to the mantle source of these ophiolites from NE Iran.

ZIRCON GEOCHRONOLOGY

Four plagiogranites and two diabasic and gabbroic mantle-intruding dikes from the Torbat-e-Heydarieh ophiolite have been dated. We also analyzed five of these zircon separates for Hf-isotope compositions.

Sample TH11-23

This sample is taken from a plagiogranitic dike injected into the mantle diabasic dikes. Zircons are euhedral and prismatic and show magmatic concentric zoning (Fig. 15). These zircons contain 215-762 ppm U and have high Th/U ratios of 0.6-1.7 (Supplementary Table 6), consistent with a magmatic origin. Twenty-one analyses define a mean $^{206}Pb/^{238}U$ age of 99.32 ± 0.72 Ma (MSWD=1.4) (Fig. 15). This is interpreted as the time of plagiogranite crystallization. Zircons from this sample show a wide range of εHf (t), between ~ +8 and +16.1 (Supplementary Table 8).

Sample TH11-12

This sample is taken from a rodingitized gabbroic dike injected into mantle peridotites. The Th/U ratio of analyzed zircons varies between 0.47-0.69, consistent with a magmatic origin. Six analyzed grains yield a mean $^{206}Pb/^{238}U$ age of 96.7 ± 2.1 Ma (MSWD=0.32) (Fig. 15). This is the age of gabbroic dike intrusion and provides a minimum age for the Torbat-e-Heydarieh ophiolite formation. Zircons from this sample show εHf (t) values that vary from ~ +9 to +16.9.

Sample TH11-82B

This plagiogranite intrudes mantle cumulate gabbros. Twenty-five zircons from this sample were dated by LA-ICPMS. Cathodoluminescence (CL) images show that the grains are medium-grained (<100 μm) and euhedral to subhedral (Fig. 15). Zircons have low to medium U (25-420 ppm) and Th (10-208 ppm) concentrations, and Th/U ratios vary between 0.4 and 0.7 (Supplementary Table 7), consistent with a magmatic origin. The $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ data define a concordia age of 91.9 ± 0.33 Ma (MSWD = 1.7). This is interpreted as the age of plagiogranite crystallization. Zircons from this sample show εHf (t) values of ~ +9.8 to +15.1 (Supplementary Table 8).

Sample TH11-74

This plagiogranite intrudes mantle cumulate gabbros. We analyzed fourteen zircons from this sample. CL images indicate magmatic zonation. Zircons have low to medium U (53-613 ppm) and Th (26-364 ppm) concentrations, and Th/U ratios vary between 0.5 and 1.3 (Supplementary Table 7). In the concordia diagram, fourteen analyses yield an age of 97.0 ± 1.3 (MSWD = 0.2) (Fig. 15). Zircons from this sample show εHf (t) values that vary from ~ +11.3 to +17.6.

Sample TH11-29

This sample is a diabasic dike within the mantle peridotites and contained only eight zircons. Zircons from this sample are prismatic (~<50 μm to ~100 μm). CL images show oscillatory zoning in some grains and some are less luminescent with weak zonation. Eight analyzed spots have high Th/U values (0.3 to 1.6). Five of the analyzed zircons show concordant ages with weighted mean of $^{206}Pb/^{238}U$ age of 97.1 ± 1.2 (MSWD=0.6) (Fig. 15). The striking feature of this sample is the presence of three xenocrystic zircons with $^{207}Pb/^{235}U$ ages of ~1.1-1.5 Ga. Late Cretaceous zircons from this sample show εHf (t) values that vary from ~ +8.1 to +12.7, whereas the old zircons have εHf (t) values between +1.7 and +6.2 (Supplementary Table 8).

Sample TH11-20

Zircon grains from this sample, a plagiogranitic dike within the crustal gabbros, are long-prismatic and euhedral. Most crystals have magmatic zonation. Twenty-five analyses show low to moderate Th/U (~0.2-0.7) (Supplementary Table 7). Analyses are concordant, yielding a weighted mean age of 97.9 ± 0.6 Ma (MSWD= 1.4). This is interpreted as the crystallization age for this dike. Zircons from this sample show εHf (t) values that vary from ~ +13 to +18.5.

DISCUSSION

This new geochemical and isotopic dataset from the Torbat-e-Heydarieh ophiolites allows a detailed investigation of the mantle characteristics and the source of crustal rocks; the timing of back-arc basin opening and the relationship of this ophiolite to other Iranian ophiolites; the tectonic setting in which the Torbat-e-Heydarieh ophiolite formed; and the implications for understanding Late Cretaceous tectonic evolution in the region.

Sources of mantle peridotites and crustal melts

Mineral- and whole-rock compositions of various mantle peridotites, crustal gabbros, and effusive lavas classify the rocks into four different tectonic settings: abyssal peridotite, supra-subduction zone fore-arc (FA), island-arc (IA), and back-arc basin (BAB) settings.

Mantle peridotites of MOR origin

Lherzolites and impregnated lherzolites show affinities with abyssal (MOR) peridotites. Compositions of Cr-spinel and co-existing olivine all plot in a confined field between fertile MORB-source mantle (FMM) and its partial melts (Figs. 9A-B). The clinopyroxene and spinel compositions of impregnated lherzolites are similar to the mineral compositions of lherzolites, suggesting these phases precipitated from the same melts during the impregnation. The overall compositional trends show mixing between a FMM source and its partial melting residue (F = ~ 9%). The trend may also represent various degrees of extraction (lherzolite) and addition (impregnated lherzolite) of basaltic melts (Fig. 10A). Clinopyroxene compositions in these rocks also plot within the enriched part (TiO_2>0.2 wt. %) of the abyssal-peridotite field (Fig. 10C). Such variations can form in the fractional to reactive melting regime in the spinel-stability field beneath a spreading ridge (Kimura and Sano, 2012), as envisaged by the clinopyroxene REE patterns with flat MREE to HREE and depleted LREE (Figs. 10A-C).

Dunite of SSZ fore-arc origin

Spinels in dunite samples are rich in Cr and Ti and their compositions plot close to the boninite and IAT basalt field (Figs. 9A-B). They plot away from the abyssal peridotite field, into the SSZ field. Because of the proximity of the compositions of spinel and clinopyroxene to those of boninite or IAT, these are most like SSZ fore-arc peridotites.

Plagioclase lherzolite and cumulate gabbro of SSZ-IA origin

Cumulate gabbros contain low-Mg# clinopyroxene and An-rich (~90) plagioclase. These are features of assemblages formed from water-rich arc basalt magmas (e.g., (Hamada and Fujii, 2008)). Therefore, the gabbros are most similar to arc magmas (Ishizuka et al., 2014).

Spinels in plagioclase lherzolites show intermediate abundances of Cr and Ti (Figs. 9A-B). Clinopyroxenes have low Mg# and are poor in Ti, similar to cumulate gabbros. The low Ti content of clinopyroxenes from plagioclase lherzolites suggest crystallization from a SSZ-derived melt during magmatic impregnation. The Cr and Ti compositions of plagioclase lherzolite spinels lie between the fields of abyssal peridotite and basalt (Fig. 10A); however, Mg# in the spinel plots away from abyssal peridotite but close to SSZ-FA peridotite. Considering the Mg#-rich spinel and Mg-poor clinopyroxene together (Figs. 9B-C), the plagioclase lherzolites suggest affinities with island-arc tholeiites.

The REE patterns of clinopyroxenes from cumulate gabbros show depleted MREE to HREE relative to MOR gabbros with elevated Sr (Fig. 11B). Plagioclase lherzolites show the same feature, with more depleted MREE to HREE and elevated Sr (Fig. 11D). Low REE abundance is the signature of a depleted mantle source and elevated Sr may reflect addition of metasomatic slab-derived fluid. Sr enrichment in plagioclase-bearing peridotites is also suggested to postdate melt impregnation and could be the result of high-temperature (370-850 °C) breakdown of plagioclase, which liberated Sr to enrich adjacent pyroxenes (Pirnia et al., 2014).

The lower abundances of Zr in plagioclase lherzolites relative to lherzolites and depletion in HFSE in the mantle and gabbro sources are consistent with hydrous metasomatism (Arai et al., 2006; Khedr and Arai, 2009). This feature is most consistent with a SSZ-FA origin of the mantle-crust lithologies.

Harzburgite and Cpx-harzburgite of BAB origin

Spinels in harzburgite have intermediate Mg# (66-76) and partially overlap lherzolite fields (Figs. 9A-B). In contrast, spinels with lower Ti and higher Cr than those of lherzolite are more depleted (Figs. 9A-B). Ti contents are low in both clinopyroxenes and dunite, suggesting a SSZ signature (Fig. 10C). Any further constraints on harzburgite tectonic affinities are difficult to identify.

Cpx-harzburgite forms common geochemical trends with harzburgite. They plot between abyssal basalt, abyssal peridotite, and SSZ dunite (Figs. 9A-C). The depleted nature of these rocks implies that harzburgite was impregnated by MORB or IAT melt (Figs. 9A-C). It is difficult to identify the impregnated melts as either MOR or IA. However, the harzburgites are distinct from abyssal peridotites suggesting involvement of strongly depleted mantle (F = ~12%: Fig. 9A). Such highly depleted sources can occur in fore-arc or BAB settings (Hirahara et al., 2015). We thus conclude that harzburgite and cpx-harzburgite probably formed in a fore-arc or BAB.

No Cpx-REE data are available for the Cpx-harzburgites. Some insight may be gained by considering harzburgite clinopyroxenes from the nearby Sabzevar ophiolite. These are extremely depleted (Fig. 11D) consistent with a SSZ mantle origin of the NE Iran ophiolite harzburgites and impregnated harzburgites.

Isotropic and cumulate gabbros: back-arc basin or fore-arc origin?

Isotropic gabbros show flat to slightly LREE-depleted patterns with lower total abundances than N-MORB, along with Nb-Ta troughs and elevated La and Th (Fig.

12B). These geochemical features are close to those of FAB from the IBM fore-arc (Reagan et al., 2010), although Nb-Ta depletions are greater in THO rocks. One microgabbroic dike is similar, with more elevated trace element abundances and no Eu anomaly (Figs. 11A-B). This may represent a melt-rich part of the gabbros. Although some depleted IA basalts generated during BAB opening show similar trace element patterns, these are generally richer in Pb, Th, and LREEs (e.g., (Shinjo et al., 1999)). Similarly, enriched E-type MORB erupted in the Sea of Japan back-arc basin during its opening at ~17 Ma (Hirahara et al., 2015). Depleted D-type MORB is also associated with the opening of the Sea of Japan (Hirahara et al., 2015). Sea of Japan D-type basalts are geochemically

similar to THO isotropic gabbros, apart from depleted HREE due to residual garnet in the source of the former basalts (Hirahara et al., 2015). According to these considerations, we find that the isotropic gabbros are most similar to FA or BAB.

In contrast to the isotropic gabbros, cumulate gabbros show elevated Sr, Pb, Ba, and Eu anomalies due to plagioclase accumulation. LREE-depleted patterns reflect cumulate clinopyroxene, but the patterns still show Nb-Ta depletion relative to La and Th. This is similar to gabbros of FA or BAB origin. The loss of melt from cumulate gabbros and the concentration of some cumulate minerals could explain the different chemistries of isotropic and cumulate gabbros (Figs. 11A-B). Plagioclase is anorthite-rich (Fig. 10D), suggesting that these gabbros most likely originated from water-rich basaltic melts (Hamada and Fujii, 2008). All of the above observations indicate that the THO gabbros are most similar to FA or BAB basalts.

It is hard to specify the tectonic setting of the SSZ gabbros from geochemistry alone; this is consistent with fore-arc, arc, or back-arc basin settings.

Pillow and massive basaltic lavas, dikes within mantle of IA-BAB origin

Most lavas show slightly LREE-depleted flat patterns with slight Nb depletion and slight Th enrichment. The pattern is akin to N-MORB, FAB, or BABB. These features are further examined below using Th/Yb and Th/Ce ratios.

A useful observation is the lack of boninite in the THO. Fore-arc volcanic sequences are frequently but not always associated with boninite and high-Mg andesite ((Ishizuka et al., 2014; Reagan et al., 2010) for IBM; (Ishikawa et al., 2002) for Oman ophiolite). In the fore-arc setting, boninites are suggested to be generated 2-4 m.yr. later than fore-arc basalts, when the residual, highly depleted mantle melted at shallow levels after interaction with a water-rich fluid derived from the subducting slab (Ishizuka et al., 2014; Reagan et al., 2010). However, boninite does not occur in the THO, suggesting that it did not form in a fore-arc subduction-initiation setting.

One pillow-lava sample is enriched in LREE, like E-MORB (Fig. 12D). This sample is similar to OIB-like pillow and massive lavas from the Sabzevar ophiolite (Moghadam et al., 2014a). E-MORB occurs either in MOR or in BAB as an along-ridge geochemical variation (e.g., (Kelley et al., 2013)). E-MORB also occurred in the SW Japan FA when subduction re-initiated at 15 Ma (Kimura et al., 2005). E-MORB occurs all over MOR-FA-BAB settings, so its presence does not constrain the setting of THO. Melting of enriched subcontinental lithosphere might be responsible for generating THO E-MORB-like lavas.

Two of the most depleted samples have strong Nb depletions and strong Th enrichments (Fig. 12D). REEs are more depleted than MORBs, but with elevated LILEs such as Sr, U, and Ba. These are IAT-like basalts; such basalts occur in BAB and in fore-arcs.

Plagiogranites and amphibole gabbros of IA-BAB origin

Plagiogranites and amphibole gabbros all have arc-magma signatures including Nb-Ta troughs and strong Th enrichment (Fig. 12F). HREEs are strongly depleted, similar to IAT-type pillow lavas. LREEs are highly enriched in both lithologies with greater LILE enrichment in amphibole gabbros. Sample TH11-9B geochemically resembles E-MORBs with enriched LREEs and less Nb depletion. Elevated Sr and Eu contents in the plagiogranites are probably due to plagioclase accumulation. Positive Zr-Hf anomalies in the amphibole gabbros and host rocks of plagiogranites are likely to reflect preferential partitioning of these elements into amphibole during fractional crystallization (Tiepolo et al., 2007) (Fig. 12F).

Nb/Yb - Th/Yb discrimination

(Pearce, 2008) proposed that Th/Yb *vs* Nb/Yb relationships could distinguish MORB-OIB from SSZ igneous rocks. Most THO lavas and crustal gabbros show elevated Th/Yb at moderate to low Nb/Yb, consistent with subduction contributions (sediment melt) to a moderately depleted mantle source (Fig. 16A). Exceptions are TH11-70, TH11-33, TH11-32, TH11-64, TH11-67, TH11-66, which plot along the MORB-OIB array within the N-MORB field, and TH11-39 pillow lava and TH11-9B amphibole gabbro, which plot along the MORB-OIB array within the E-MORB field. Other lavas and gabbros have Nb/Yb similar to N- and E-MORB, but have elevated

Th/Yb (Fig. 16A). Pillow and massive lavas, mantle dikes and cumulate gabbros have low Nb/Yb, whereas amphibole gabbros and plagiogranites show high Nb/Yb indicating derivation of these magmas from more- and less-depleted mantle sources, respectively (Pearce, 2008).

Th/Nb - Ce/Nb discrimination

Similar conclusions are reached based on the Th/Nb - Ce/Nb plot (Fig. 16B) (Sandeman et al., 2006). Most lava flows (TH11-70, TH11-33, TH11-32, TH11-64, TH11-67, TH11-66) and the TH11-9A amphibole gabbro have MORB signatures whereas TH11-39 pillow lava and TH11-9B amphibole gabbro are similar to Sabzevar OIB-type lavas. These are in turn similar to Lau Basin BAB and Okinawa Trough intra-arc rift lavas. In contrast, other IA-BAB-type lavas, isotropic gabbros, amphibole gabbros, and mantle dikes have lower Ce/Th, plotting close to Okinawa Trough BABs. Overall, THO magmatic rocks including lavas, gabbros, and plagiogranites plot between MORB and Mariana arc lavas, and therefore indicate generation from a heterogeneous DMM source mantle that was variably fluxed by slab fluids.

The above considerations support a BAB origin for the THO magmatic rocks. Only plagiogranites and amphibole gabbros show strong IAT signatures (Fig. 12D). These geochemical signatures resemble Sabzevar ophiolite magmatic rocks (Moghadam et al., 2014a).

Zr/Nb - Hf/Nb and Th/Nd - $^{143}Nd/^{144}Nd$ discrimination diagrams

The composition of BAB basalts varies significantly, indicating that BABB melts are derived from heterogeneous mantle sources, which range from depleted to enriched and from MORB-like to arc-like (Stern, 2002; Taylor and Martinez, 2003). The THO igneous rocks are LREE-depleted or slightly enriched relative to N-MORB, similar to BABB (Fig. 12) like those of the Lau Basin (Tian et al., 2011) and Okinawa Trough (Shinjo, 1999). The THO igneous rocks involve both depleted (FAB and D-type BABB-like igneous rocks) and enriched (E-BABB and E-MORB-like rocks) mantle components, a conclusion which is supported by the plot of Hf/Nb *vs* Zr/Nb (Sorbadere et al., 2013) (Fig. 16C).

To better understand the contribution of slab-derived components to the mantle source of THO igneous rocks, we used a plot of Th/Nb *vs* $^{143}Nd/^{144}Nd$ (Fig. 16D). THO igneous rocks reflect variable enrichment by sediment melts, as shown on Fig. 16D (Sorbadere et al., 2013). THO melts are like those of the Sabzevar ophiolite, which sampled multiple mantle and slab components, again consistent with a BAB interpretation.

Whole rock Nd-Hf and zircon Hf isotopes

Whole rock (εHf=+14.9 - +21.5) and zircon (εHf= +8.1 - +18.5) Hf- isotope data from the Torbat-e-Heydarieh ophiolite are comparable. These data indicate that the mantle source beneath the THO back-arc was broadly MORB-like (Todd et al., 2011; Woodhead et al., 2012; Woodhead and Devey, 1993). However, this mantle source also had slightly enriched components (Stracke et al., 2005), as shown by Hf and Nd isotopic data (Fig. 14).

All THO lavas seem to be derived from an Indian MORB source mantle (I-DMM), similar to the source of the Oman ophiolitic gabbros. The addition of HFSE such as Hf occurs if subducted sediments are melted or if supercritical fluids are involved (Kessel et al., 2005). The THO igneous rocks vary modestly in εHf and εNd, indicating restricted contributions of subducted sediment melts. We conclude that the THO pillow lavas and massive lavas reflect derivation from an Indian Ocean DMM mantle that may have been slightly affected by melts of subducted sediments, although some isotropic gabbros and E-MORB lavas show a modestly-enriched mantle signature.

It is worth noting that THO Hf-isotope compositions are similar to those of the Zagros fore-arc ophiolites, both inner belt (Nain-Baft; εHf (t) = +8 - +21) and outer belt (Neyriz-Kermanshah; εHf (t) = +10 - +23) ophiolites, confirming a broadly similar mantle source for the generation of Late Cretaceous Iranian ophiolitic magmas. However, the conspicuous variations in the zircon εHf values (+8.1 - +18.5) (which are similar to the whole rock Hf isotope variations; εHf=+14.9 - +21.5), can reflect mixing with new pulses of similar magmas but with slight isotopic differences entering the magma chamber duringcooling of the plagiogranitic melts (Shaw and Flood, 2009).

Heterogeneous mantle sources for THO magmas: further thoughts

Table 1 summarizes possible sources and tectonic affinities of the studied THO rocks. As discussed above, lherzolites and impregnated lherzolites are similar to MOR abyssal peridotite. Dunites are most like SSZ-FA mantle whereas plagioclase lherzolites are most like sub-arc mantle. Plagiogranites and amphibole gabbros also have strong arc affinities. Lavas, dikes in peridotite, other gabbros, and harzburgites have the strongest affinities with BAB (Table 1).

Cumulate gabbros are melt-deficient equivalents of isotropic gabbros; both are probably derived from Nb-Ta depleted N-MORB-like magmas like THO pillow and sheet lavas. Harzburgite is probably melt-depleted residual BAB mantle as indicated by Cr# and TiO_2 contents of its spinels (Fig. 10). The Cpx-harzburgite is similar to melt-impregnated harzburgite but with greater addition of abyssal basalt melt (Fig. 10A). "Abyssal basalt" in this case is not N-MORB but BAB basalt with signatures of slab-derived fluids. Cpx-harzburgites may represent reactive melt flow channels (Kimura and Sano, 2012). Magmas so generated could have formed dikes within the mantle near the Moho mantle and could have fractionated to form isotropic and cumulate gabbros. Some of this magma may have erupted as pillow lavas and sheet lavas to form BAB oceanic crust Layer 2 (Hirahara et al., 2015).

The presence of heterogeneous mantle rocks (with affinities to FA, IA, and MOR) associated with THO BAB is consistent with the inferred tectonic setting of the THO.

Tectonic significance of the Torbat-e-Heydarieh ophiolite

Timing of back-arc basin opening and its relation to Zagros subduction initiation

Strong upper plate extension accompanies the initiation of some subduction zones such the Izu-Bonin-Mariana convergent margin, south of Japan. Strong extension on the upper plate of a nascent subduction zone, above the sinking lithosphere, allows asthenospheric upwelling and leads to seafloor spreading, forming infant arc crust of the proto-forearc (Stern, 2004; Stern and Gerya, 2017). (Gurnis et al., 2004; Hall et al., 2003) suggest that rapid trench retreat and extension in the overriding plate also occur after subduction becomes self-sustaining, others believe that trench roll-back (and back-arc extension) can start after the slab pull becomes greater than the far-field push (Baes et al., 2011).

Late Cretaceous Zagros ophiolites and their ~3000 km long equivalents in Oman and from the Mediterranean area (Cyprus, Turkey, Syria to Iraq) show a site of Late

Cretaceous subduction initiation along the southern margin of Eurasia (Moghadam and Stern, 2011; Monsef et al., 2018). The chemo-stratigraphic relationships of the lavas from these ophiolites, integrated with mantle geochemistry and paleomagnetic data, support the idea of arc infancy at the southern margin of Eurasia during the Late Cretaceous (Dilek et al., 2007; Maffione et al., 2017; Moghadam and Stern, 2011; Monsef et al., 2018; Pearce and Robinson, 2010). It is also suggested that the earliest subduction stage after the SI coincided with the transient transfer of high-pressure rocks from the top of the subducting slab to the overriding plate (van Hinsbergen et al., 2015). Late Cretaceous high-pressure rocks are common within the Oman and Zagros ophiolites (Guilmette et al., 2018; Moghadam et al., 2017a), and further attest to the transfer of deep-seated rocks during the earliest stage of subduction.

Because of their setting far to the north of the Late Cretaceous Zagros SI plate margin, STHO ophiolites must have formed in a back-arc basin. They are easily related to the Late Cretaceous (~104-98 Ma) formation of a new subduction zone along the southern margin of Iran, where seafloor spreading formed new oceanic lithosphere (Zagros inner and outer belts ophiolites) (Moghadam and Stern, 2015). Subduction initiation was followed quickly by arc magmatism to form the Urumieh-Dokhtar Magmatic Belt (e.g., (Chiu et al., 2013; Ghasemi and Manesh, 2015; Ghorbani et al., 2014; Honarmand et al., 2014)). Opening of the STHO BAB probably reflected strong regional extension of the overriding plate during subduction initiation. Our zircon U-Pb ages range from ~99 to 92 Ma, reflecting a period of ~7 Myr for formation of the THO ophiolite. A similar age range is also indicated by Cenomanian to Turonian (~ 99-90 Ma) carbonates deposited over the massive and pillow lavas.

Such a broad zone of extension accompanying SI is consistent with IODP 351 drilling results west of the IBM arc, where scientists were surprised to find Eocene basalts far west of where subduction began at ~51 Ma (Hickey-Vargas et al., 2018). The dominance of Late Cretaceous extension in NE Iran is also confirmed by the presence of a sheeted dike complex in the Sabzevar ophiolites with U-Pb ages of ~90 Ma obtained from the felsic dikes (Moghadam et al., 2014b). This extension was also associated with basement-involved low-angle normal faulting, core complex exhumation, block tilting and sedimentary basin formation in some parts of Iran (e.g., in Saghand, Fig. 1 (Verdel et al., 2007), in Golpayegan (Moritz and Ghazban, 1996) and in Torud (Malekpour-Alamdari et al., 2017), attested by ^{39}Ar-^{40}Ar and K-Ar ages. These results show that extensional deformation prevailed during Late Cretaceous in what is now the Iranian plateau. Extension in NE Iran also generated a volcano-sedimentary basin (Southern Sabzevar basin, N-NW of Oryan) which was filled by Late Cretaceous pelagic sediments, green siliceous tuffs and submarine volcanic rocks (Kazemi et al., 2019). Late Cretaceous extension also allowed asthenospheric melts with radiogenic isotopic compositions to invade the continental crust and generate granitoids and lavas with radiogenic Nd and zircon Hf isotopic compositions (e.g., (Alaminia et al., 2013; Kazemi et al., 2019)).

Tectonic setting of the Torbat-e-Heydarieh ophiolite

Three key questions concerning the Tobat-e-Heydarieh ophiolite are addressed here: 1) What is the relationship between the Sabzevar and Torbat-e-Heydarieh ophiolites? 2) What is the relationship between STHO ophiolites and other ophiolites of similar age in eastern Iran? 3) What is the relationship between the Zagros fore-arc ophiolites and the STHO?

The ages obtained here agree with U-Pb zircon ages of 100 to 78 Ma for the Sabzevar ophiolite (Moghadam et al., 2014a). THO ages are also similar to zircon U-Pb ages obtained for dacitic-andesitic lavas from the Cheshmeshir and Oryan ophiolites (~102-76 Ma, Fig. 3; *Moghadam et al., unpublished data*). Some of the younger ages may reflect arc magmatism. Together, results from the THO, Sabzevar and Cheshmehshir-Oryan ophiolites indicate a Late Cretaceous oceanic basin that was magmatically active and perhaps open for ~26 Ma. Our age results indicate that the Zagros SI may be somewhat older than the formation of the NE Iran back-arc basins (104-98 Ma *vs* 99-92).

Late Cretaceous mantle upwelling accompanying extension led to the emplacement of several juvenile intrusions and associated extrusive rocks (Late Cretaceous-Eocene magmatic belt in Fig. 3). These Late Cretaceous igneous rocks are abundant in NE Iran and are mostly juvenile additions from the mantle to the crust. Xenocrystic zircons (with ages of 1.1-1.5 Ga) from sample TH11-29 (diabasic dike within peridotites) suggest that recycling of ancient subducted materials accompanied back-arc opening. These xenocrytic zircons are found in many Neotethyan ophiolites in China and the Mediterranean realm (e.g., (Gong et al., 2016; Robinson et al., 2015)).

The Torbat-e-Heydarieh ophiolite probably represents an extension of the Sabzevar ophiolite. Although the formation age of these ophiolites (Sabzevar and THO) is nearly synchronous, there are striking differences in the tectono-metamorphic evolution of the two ophiolite belts. The main differences include: 1- mantle peridotites are more important in the THO and spinels from Sabzevar mantle peridotites have SSZ- or fore-arc-type signatures, characterized by higher Cr# (Shafaii Moghadam et al., 2015) (Fig.

10B). 2- Podiform chromitites are rare in the THO and abundant in the Sabzevar ophiolite; 3 -Acidic volcanic rocks are abundant in Sabzevar, whereas mafic volcanics are more important in the THO (Fig. 13A). Sabzevar volcanic rocks show more pronounced arc signatures than do those of the THO, indicated by higher Th/Yb, Th/Nb and Ce/Nb ratios (Figs. 15A-B). 4- OIB-type lavas are present in Sabzevar (Moghadam et al., 2014a) but are absent in the THO, although the latter contains minor E-MORB. 5- The Sabzevar ophiolite marks an orogenic suture, with a Paleocene HP/LT metamorphic core (Omrani et al., 2013; Rossetti et al., 2014) and an external thrust-and-fold belt, showing evidence of a ductile-to-brittle top-to-the-SSE sense of tectonic transport (Rossetti et al., 2014). Moreover, a HP granulite event, Albian in age, may exist in the Sabzevar zone (Nasrabady et al., 2011; Rossetti et al., 2010). Conversely, the Torbat-e-Heydarieh ophiolite realm does not show evidence of orogenic metamorphism, and the post-Cretaceous tectonic evolution has been controlled by polyphase strike-slip shearing (Tadayon et al., 2017).

The observation that all Sabzevar lavas and mantle peridotites have SSZ signatures whereas THO lavas and peridotites have mostly MORB-like signatures, as well as the presence of older high-P rocks in the Sabzevar and their absence in the Torbat-e-Heydarieh, may indicate that the Torbat-e-Heydarieh and Sabzevar ophiolites represent distinct back-arc basins, differing proximities to the arc, or different stages in BAB opening. It is probably not realistic to consider two neighboring oceanic basins so close together in NE Iran during Late Cretaceous time, although distinct rift basins existed due to strike-slip faulting accompanying break-up of the continental lithosphere in NE Iran. We prefer to interpret STHO ophiolite fragments as having formed in a single back-arc basin, perhaps in distinct extensional basins (grabens), developed during Late Cretaceous hyperextension in the region.

Regarding the second question, what is the relationship of STHO to other Late Cretaceous oceanic remnants in eastern Iran? We believe that back-arc opening experienced distinct spatio-temporal and geodynamic evolution, but nearly all formed due to the extensional regime which prevailed in the region during the latest Early to Late Cretaceous (~110-70 Ma). The Birjand-Zahedan ophiolites in eastern Iran (Fig. 1) are similar in age to the STHO. SHRIMP U-Pb zircon dating of metafelsic rocks and eclogites from the Birjand-Zahedan ophiolites gave ages of ~ 86-89 Ma (Bröcker et al., 2013) (Fig. 13). In addition, zircon U-Pb ages from Birjand-Zahedan gabbros are somewhat older than STHO ophiolites; 113-107 Ma (Zarrinkoub et al., 2012). These may be related to a different episode of Early Cretaceous subduction in the region. Zircon and titanite from felsic segregations in mafic granulites from the Sabzevar ophiolite yield U-Pb ages of 107.4±2.4 and 105.9±2.3 Ma, respectively (Fig. 13; (Rossetti et al., 2010)). This evidence might suggest the formation of Early Cretaceous (i.e. pre-Albian) Sabzevar-Birjand-Zahedan single oceanic realm in the upper plate above a Neothethys subduction zone. In fact, the formation of these high HP granulites in a subduction zone is questionable, as these rocks contain no high-P minerals and its zircon U-Pb age (ca. 107 Ma) is similar to its titanite age (ca. 105 Ma) and both are similar to the zircon U-Pb ages reported from the felsic to mafic magmatic rocks from the Sabzevar area (Fig. 3). Therefore, we believe these metamorphic rocks (with amphibole-plagioclase-garnet ± titanite as rock-forming minerals) had a mafic protolith and were not metamorphosed in a subduction zone. It seems that a MORB precursor related to 105 Ma granulites existed in the Sabzevar area during the Lower Cretaceous (Nasrabady et al., 2011). This may imply that an oceanic crust older than 105 Ma was present in NE Iran, before the Torbat-e-Heydarieh back-arc basin start to open.

Finally, what is the relationship between Zagros ophiolites and the STHO? There are geochemical similarities, but Zagros ophiolites are mostly slightly older (*ca* 104-98 Ma) than the STHO (*ca* 99-80 Ma). This may reflect the migration of extension accompanying subduction initiation; first in what becomes the fore-arc, then as the slab descends the locus of extension migrates to the back-arc region (Fig. 17A). In this scenario, the STHO represents a back-arc basin that formed behind the Zagros fore-arc and the nascent Urumieh-Dokhtar magmatic arc (Fig. 17B). Back-arc opening would have disrupted Cadomian continental crust, including Cadomian subcontinental lithospheric mantle (SCLM; the Lu-Hf *vs* Hf plot, *not shown*, indicates an isochron with a Cadomian mantle-depletion age of ~585 Ma). Involvement of such SCLM may explain the origin of OIB-type lavas in the Sabzevar ophiolite and E-MORB in THO. How far west this back-arc basin can be traced remains unclear, because the region to the SW between the STHO and the Nain ophiolites of similar age is covered by younger deposits for ~400 km (Fig. 1). Geophysical studies of this region including magnetics and gravity may help answer this question.

It is also unresolved when and why the STHO BAB collapsed. U-Pb ages for THO indicate a lifespan of ~7 million years, slightly less than the lifespan of 12.5±4.7 Ma for global extinct BABs (Stern and Dickinson, 2010). One possibility is that the Iranian continental lithosphere had been weakened by prolonged arc igneous activity, allowing the basin to collapse when regional stress changed from extension to compression as the subduction zone evolved. There is abundant evidence for prolonged arc igneous activity from Late Cretaceous time onwards (e.g., (Chiu et al., 2013; Honarmand et al., 2014; Hosseini et al., 2017; Verdel et al., 2011)). The volcano-pelagic series above the STHO includes Cenomanian to Maastrichtian deep-sea pelagic sediments interbedded with pyroclastic and andesitic to dacitic lavas. Dacitic-andesitic lavas from the Cheshmehshir and Oryan ophiolites have Late Cretaceous (~102-76 Ma) U-Pb ages (*Moghadam et al., unpublished data*). This sequence grades upward into a series of Maastrichtian to Paleocene shallow water sediments and then into the Oryan marine sediments (lower and middle Oryan sediments) from early Eocene to early middle Eocene. Moreover, zircon U-Pb ages for plutonic rocks from south of Sabzevar and Neyshabour (Fig. 3, between the Sabzevar ophiolite in the north and the Cheshmeshir ophiolite in the south) are of Cenomanian-Maastrichtian (97.0± 0.2 Ma; 67.5±0.5 Ma) to Oligo-Miocene (29.8 ± 0.2 Ma) (*Moghadam et al., unpublished data*; (Alaminia et al., 2013)). The Late Cretaceous ages are similar to the U-Pb zircon ages of the Sabzevar plagiogranites (100 to 78 Ma).

CONCLUSIONS

Our study of the THO confirms that it formed in Late Cretaceous time. Dikes intruding THO mantle peridotites and plagiogranites have U-Pb ages of 99-92 Ma. THO igneous rocks have a range of εNd(t) between +5.7 - +8.2 and their εHf(t) values range from +14.9 to +21.5; THO zircons have εHf(t) values of +8.1 - +18.5. Like all NE Iran ophiolites, the THO is found in a tectonic position well to the north of slightly older Zagros fore-arc ophiolites and the Urumieh-Dokhtar magmatic belt, and thus appears to have formed as a continental back-arc basin. Petrological, geochemical, and isotopic compositions are consistent with this interpretation. THO peridotites contain spinels and clinopyroxenes with compositions like those in peridotites from mid-ocean ridges and back-arc basins. THO igneous rocks have Hf-Nd isotopic compositions that are similar to Indian Ocean MORBs, and do not show clear evidence for addition of sediment-derived melts or fluids from a subducted slab. Magmatic rocks in the Torbat-e-Heydarieh ophiolite show both MORB-like and SSZ geochemical signatures. We conclude that formation of the Sabzevar- Torbat-e-Heydarieh ophiolite reflects opening of a Late Cretaceous back-arc basin as a result of regional hyper-extension accompanying subduction initiation along southern Iran.

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FIGURE CAPTIONS

Fig. 1- Simplified geological map of Iran emphasizing Cadomian rocks, the main ophiolitic belts (thick dashed lines) and Cenozoic magmatic rocks. A-B line shows the cross-section position in Fig. 2C.

Fig. 2- (A) and (B) Schematic model showing the opening and evolution of the Sabzevar-Torbat-e-Heydarieh back-arc basin relative to the Zagros, Neotethyan Ocean (modified after (Dercourt et al., 1986; Kazmin et al., 1986).

Fig. 3- Geological map of the Sabzevar-Torbat-e-Heydarieh region, north of the Dorouneh Fault, with emphasis on the distribution of ophiolitic and arc-related rocks. (Abbreviations are M et al.= *Moghadam et al., unpublished data*; K et al.= (Kazemi et al., 2019); M et al., 2014= (Moghadam et al., 2014a)).

Fig. 4- Simplified geological map of the Torbat-e-Heydarieh ophiolite (modified after (Vaezi Pour et al., 1992)).

Fig. 5- Simplified stratigraphic column displaying idealized internal lithologic successions in THO.

Fig. 6- Outcrop photos of the THO rock units. (A) outcrops of mantle peridotites. (B) Rodingitized dikes within the THO mantle peridotites. (C) Outcrop of interlayered cumulate gabbros (layered gabbros) and clinopyroxenites. (D) Small intrusions of cumulate (layered) gabbros within the THO mantle sequence. (E) Plagiogranitic dikes within the mantle gabbroic intrusions. (F) Angular gabbroic xenoliths within the plagiogranitic pockets. (G) Outcrops of crustal massive and pillow lavas. (H-I) Late Cretaceous (Cenomanian to Turonian, ~ 99-90 Ma) pelagic sediments are interlayered with and conformably cover the lavas.

Fig. 7- Photomicrographs of the THO mantle peridotites and gabbros. (A) Opx, Cpx and serpentinized olivine in lherzolites. (B) Serpentinized olivines, Opx and Cpx grains as embayment within the Opx in Cpx-harzburgites. (C) Serpentinized olivine, Opx and Cpx aggregates within the impregnated lherzolites. (D) Coarse-grained Cpx crystals and altered plagioclase in plagioclase-bearing lherzolites. (E) and (F) Opx, Cpx and plagioclase in mantle cumulate gabbros.

Fig. 8- Back-scattered images of THO mantle rocks. (A) Association of clinopyroxene and vermicular spinels at the embayments of orthopyroxene porphyroclasts in Cpx-harzburgites. (B) Altered plagioclase enclosed by large Cpx grains in plagioclase lherzolites. (C) Olivine, Opx, Cpx and plagioclase in plagioclase lherzolites. (D) Cpx and chloritized amphibole inclusions in the chromite of dunites. (E) and (F) Cumulate texture in mantle cumulate gabbros with early crystallized Opx and Cpx and late-stage, interstitial plagioclase crystals.

Fig. 9- Modal composition of THO mantle peridotites (A) and gabbros (B).

Fig. 10- (A) Relationship between Cr# and TiO_2 contents of spinel in mantle peridotites of the Torbat-e-Heydarieh ophiolite. The thick gray arrow shows the effect of melts (with composition of Troodos boninites) on refractory subduction-zone peridotites. Compositions of Tonga and Izu-Bonin-Mariana island arc tholeiites and abyssal basalts are from (Pearce et al., 2000a) and the composition of Troodos boninite is from (Dick and Bullen, 1984). (B) Relationship between Cr# and Mg# of spinels in Torbat-e-Heydarieh mantle peridotites. Abyssal-peridotite fields are from (Dick and Bullen, 1984), fore-arc-peridotite field is from (Pearce et al., 2000a). Data on Sabzevar peridotites are from (Moghadam et al., 2014a). (C) Mg# *vs* TiO_2 contents of clinopyroxene from

peridotite and mantle gabbro (abyssal and fore-arc peridotite fields are from (Bedard et al., 2009). (D)- Anorthite content *vs* clinopyroxene Mg# (modified after (Sanfilippo et al., 2013)) in Torbat-e-Heydarieh gabbroic rocks. The Kizildag ophiolite data are from (Bagci et al., 2005). Fields of MORB and arc gabbro are from (Burns, 1985).

Fig. 11- Abundances of Rare Earth Elements and other trace elements in clinopyroxene in lherzolites and cumulate gabbros (A-B) and impregnated and plagioclase lherzolites (C-D), from the Torbat-e-Heydarieh ophiolite. Chondrite normalization values are from (Mcdonough and Sun, 1995). Fields for abyssal and supra-subduction zone (SSZ-) peridotite clinopyroxenes are from (Bizimis et al., 2000; Johnson et al., 1990). The composition of MOR cumulate Cpx is from (Ross and Elthon, 1993). Data on Sabzevar ophiolite rocks are from (Shafaii Moghadam et al., 2015).

Fig. 12- Chondrite-normalized REE patterns (chondritic abundances from (Mcdonough and Sun, 1995)) and primary mantle and N-MORB normalized multi-element patterns (N-MORB and primary mantle concentrations from (Mcdonough and Sun, 1995)) for Torbat-e-Heydarieh peridotites, mantle gabbros, plagiogranites and crustal magmatic rocks.

Fig. 13- (A) FeO^*/MgO *vs* SiO_2 (Miyashiro, 1974) and (B) Ti *vs* V diagrams (Shervais, 1982) for THO rocks. Data for the Sabzevar ophiolite are from (Moghadam et al., 2014a).

Fig. 14- εHf *vs* εNd for the Torbat-e-Heydarieh magmatic rocks, recalculated at 100 Ma (modified after (Chauvel et al., 2009)). In panel (A), our samples are compared to the MORB and OIB-type lavas, whereas in panel (B), the mixing trend between depleted mantle reservoir and different-types of sediments is shown. Data for Oman plagiogranites are from (Haase et al., 2015) and data from Dehshir ophiolites are from (Moghadam et

al., 2012). MORB and OIB data are from (Chauvel and Blichert-Toft, 2001a; Nowell et al., 1998; Pearce et al., 1999; Woodhead et al., 2001). Mantle array data are after (Vervoort and Blichert-Toft, 1999a).

Fig. 15- Zircon SHRIMP and LA-ICPMS U-Pb data from the Torbat-e-Heydarieh plagiogranite and diabasic-gabbroic dikes.

Fig. 16- (A) Th/Yb *vs* Ta/Yb (Pearce and Peate, 1995) and (B) Ce/Nb *vs* Th/Nb (Sandeman et al., 2006) diagrams for the Torbat-e-Heydarieh magmatic rocks. Hf/Nb *vs* Zr/Nb (C) and Th/Nb *vs* ^{143}Nd/^{144}Nd plots (D) (Sorbadere et al., 2013) for THO magmatic rocks, Okinawa Trough, Central Lau basin and Sabzevar ophiolitic igneous rocks, comparing the effect of enrichment and depletion in the mantle sources of these magmas, with respect to the N-MORB source. Pale red field shows bulk mixing between N-MORB and sediment and between N-MORB and fluid and thus represents the slab contribution (modified after (Sorbadere et al., 2013)). Data from the Sabzevar ophiolite are from (Moghadam et al., 2014a). Data from the Lau back-arc basin and the Okinawa Trough are from (Tian et al., 2011) and (Shinjo, 1999) respectively.

Fig. 17- Schematic model showing the generation of the Sabzevar-Torbat-e-Heydarieh ophiolites. (A) Subduction initiation and Neotethys sinking caused strong extension in the region above the sinking lithosphere, leading to seafloor spreading and forming the proto-fore-arc crust. Strong extension in the Iranian continental crust lead to extensional basins opening and nucleation of the STHO back-arc basin. (B) Beginning of true subduction (~100 Ma), and formation of the Urumieh-Dokhtar arc and the STHO back-arc basin.

TABLE CAPTION

Table 1- Provenance of the mantle-crust-lava units of THO.

SUPPLEMENTARY DATA

Supplementary Table 1- Representative compositions of minerals in the Torbat-e-Heydarieh ophiolites.

Supplementary Table 2- Trace elements composition of minerals from THO mantle peridotites and cumulate gabbros.

Supplementary Table 3- Whole rock data of the THO mantle peridotites and gabbros.

Supplementary Table 4- Whole rock analysis of the THO magmatic rocks.

Supplementary Table 5- Whole rock Nd-Hf isotope data for the Torbat-e-Heydarieh ophiolitic rocks.

Supplementary Table 6- SHRIMP U-Pb data for zircon from magmatic rocks of the Torbat-e-Heydarieh ophiolites.

Supplementary Table 7- LA-ICPMS U-Pb data for zircon from magmatic rocks of the Torbat-e-Heydarieh ophiolites.

Supplementary Table 8- Lu-Hf isotope data for zircon from magmatic rocks of the Torbat-e-Heydarieh ophiolites.

ELECTRONIC APPENDIX A

Major element compositions of minerals were analyzed using JEOL wavelength dispersive electron probe X-ray micro-analyzer (JXA 8800R) at Kanazawa University. Accelerating voltage, beam current, and beam diameter for the analyses were 20 kV, 20 nA, and 3 μm, respectively. Representative mineral compositions are reported in Supplementary Table 1.

Trace-element contents of minerals from peridotites and gabbros were analyzed on polished thick-sections in the Geochemical Analysis Unit (GAU) at CCFS/GEMOC, Macquarie University; Sydney, Australia. An Agilent 7700 laser ablation system has been used to analyze trace element abundances in minerals. The Agilent 7700 was coupled with a New Wave UP-266 nm Nd:YAG laser microprobe. Data were collected and processed using the GLITTER software, which allows for the cleanest part of the time-resolved spectrum to be selected, avoiding inclusion phases and host silicate phases. NIST-612 and BCR-2 glasses were used as an external calibration. The trace-element compositions of minerals are reported in Supplementary Table 2.

Major and trace element analyses of pillow lavas, massive lavas, plagiogranites and dikes were carried out using ICP-AES and ICP-MS at CNRS-SARM, Nancy University (France), using BR, DR-N, UB-N, AN-G and GH standards. For major elements, the uncertainty (1 sigma) is better than 2% for concentrations higher than 5 wt.% and better than 5% in the range 0.1-5 wt.%. For trace elements and REEs, the precision is 5% in the range 1-100 ppm and 10% in the range 0.1-1 ppm. The bulk rock major and trace elements analyses for pillow lavas, massive lavas, plagiogranites and dikes are shown in Supplementary Table 4.

Major elements of the Torbat-e-Heydarieh mantle gabbroids and peridotites were analyzed using ICP-AES and ICP-MS at ACME Analytical Laboratories Ltd, Canada. Concentrations of trace elements in Torbat-e-Heydarieh gabbroids and peridotites were determined by Inductively Coupled Plasma Mass Spectroscopy (ICP-MS) using a Thermo Scientific X-Series 2 in the Department of Earth Sciences at the University of Durham, following a standard nitric and hydrofluoric acid digestion (Ottley et al., 2003).

Sample preparation was undertaken in clean air laminar flow hoods. Briefly the procedure is as follows; into a Teflon vial 4ml HF and 1ml HNO_3 (SPA, ROMIL Cambridge) is added to 100 mg of powdered sample, the vial is sealed and left on a hot plate at 150 °C for 48 h. The acid mixture was evaporated to near dryness, the moist residue has 1 ml HNO_3 added and evaporated again to near dryness. 1 ml HNO_3 was again added and evaporated to near dryness. These steps convert insoluble fluoride species into soluble nitrate species. Finally, 2.5 ml HNO_3 was added and diluted to 50 ml after the addition of an internal standard giving a final concentration of 20 ppb Re and Rh. The internal standard was used to compensate for analytical drift and matrix suppression effects. Calibration of the ICP-MS was via international rock standards (BHVO-1, AGV-1, W-2, and NBS688) with the addition of an in-house standard (GP13) (Ottley et al., 2003). These standards and analytical blanks were prepared by the same techniques as for the THO samples. To improve the signal-to-noise threshold for low abundances of incompatible trace elements in ultramafic rocks, instrument dwell times were increased (Ottley et al., 2003). The composition of the reference samples (W-2, AGV-1, BHVO-1, BE-N, NBS688) was analyzed as unknowns during the same analytical runs. For the analyzed elements, reproducibility of these reference samples is generally better than 2% and the measured composition compares favorably with that published information in (Potts et al., 1992). The bulk rock major and trace elements analyses for cumulate gabbroids and peridotites are shown in Supplementary Table 3.

Eight whole-rock samples were analyzed for their Nd and Hf isotope compositions (Supplementary Table 5). Chemical separation of Nd and Hf used c. 100 mg of sample powder that was spiked with mixed ^{176}Lu-^{180}Hf and ^{149}Sm-^{150}Nd tracers and subsequently digested on a hotplate in a 1:1 mixture of concentrated HF-HNO_3 (24 hours, 120°C). Subsequently, the acid mixture was dried and the samples were digested in a 3:2 mixture of concentrated HF-HNO_3 in steel-jacketed Parr pressure vessels (3 days, 180°C). Subsequently, fluorides were decomposed by drying with 1 ml of perchloric acid and three additional dry-down steps with concentrated HNO_3. The chemical separation of Hf and Lu from the matrix follows the procedure of (Munker et al., 2001). Samarium and Nd were separated from the residue after Hf separation, using Bio-Rad AG50W-X8 cation resin (200-400 mesh) and Ln-Spec resin (Pin and Zalduegui, 1997).

The Nd and Hf isotope compositions as well as Lu-Hf and Sm-Nd concentrations were determined using a Thermo-Finnigan Neptune multi-collector ICP-MS at the joint clean lab facilities in Cologne/Bonn. Data was collected in static multi-collection mode. Values of $^{143}Nd/^{144}Nd$ and $^{176}Hf/^{177}Hf$ were corrected for mass fractionation using the exponential law and $^{146}Nd/^{144}Nd$=0.7219 and $^{179}Hf/^{177}Hf$ =0.7325, respectively. Repeated analyses of the standards La Jolla Nd and Hf AMES (isotopically identical to JMC-475) yield mean values of $^{143}Nd/^{144}Nd$= 0.511835 (n=2) and $^{176}Hf/^{177}Hf$=0.282156 (n=18). The external long-term reproducibility is c. ±40 ppm for Nd and Hf isotopes (2 RSD). Reported values are given relative to 0.511859 for La Jolla and 0.282160 for Hf AMES. Procedural blanks were typically below 60 pg for Hf and Nd. The external precision of the Nd and Hf measurements was further assessed by multiple digestions of six samples (see Supplementary Table 5). External precision (2 RSD) for the replicates was better than 30 ppm for $^{143}Nd/^{144}Nd$ and 40 ppm for $^{176}Hf/^{177}Hf$.

In order to have precise ages for the Torbat-e-Heydarieh ophiolite, zircons from plagiogranites (4 samples) and diabasic (1 sample)- rodingitized gabbroic dike (1 sample) within the mantle harzburgite were analyzed by SHRIMP at the Korea Basic Science Institute, Ochang, South Korea and LA-ICPMS at Geochemical Analysis Unit (GAU), CCFS/GEMOC, Macquarie University. For SHRIMP analysis, zircon grains were mechanically separated using conventional mineral separation techniques that employed crushing, grinding, sieving, and magnetic separation steps, followed by handpicking of zircons under a binocular microscope. The internal structure of these zircon grains was studied using transmitted and reflected light optical microscopy, and by scanning electron microscope-cathodoluminescence (CL) imaging. Analytical procedures including instrumental set-up, and the acquisition and treatment of data employed in this SHRIMP U-Pb zircon dating study, are outlined in (Compston et al., 1984). For the reduction of raw data and in making the final age calculations, we used the programs SQUID and Isoplot/Ex (Ludwig, 2003, 2009). The SHRIMP U-Pb zircon analytical data are summarized in Supplementary Table 6.

For LA-ICPMS analysis, zircons were separated following electrostatic disaggregation (selFrag) of the rock sample, then using standard gravimetric and magnetic techniques; grains were picked under a binocular microscope and mounted in epoxy discs for analysis. All grains were imaged by CL and BSE to provide maps to guide the choice of analytical spots. Zircon U-Pb ages were obtained using a 193 nm ArF EXCIMER laser with an Agilent 7700 ICP-MS system. Detailed method descriptions have been given by (Jackson et al., 2004). The ablation conditions included beam size (30 µm), pulse rate (5Hz) and energy density (7.59 J/cm^2). Analytical runs comprised 16 analyses with 12 analyses of unknowns bracketed by two analyses of a standard zircon GJ-1 at the beginning and end of each run, using the established TIMS values

(^{207}Pb/^{206}Pb age= 608.5 Ma, (Jackson et al., 2004)). U-Pb ages were calculated from the raw signal data using the on-line software package GLITTER (Griffin et al., 2008). U-Pb age data were subjected to a common-lead correction, except for those with common-Pb concentrations lower than detection limits. The results were processed using the ISOPLOT program of (Ludwig, 2003). The external standards, zircons 91500 and Mud Tank, gave mean ^{206}Pb/^{238}U ages of 1063.5±1.8 Ma (MSWD=1.3) and 731.1±1.2 Ma (MSWD=0.77), respectively, which are similar to the recommended ^{206}Pb/^{238}U ages of 1062.4±0.4 Ma and 731.9±3.4 Ma respectively (Chang et al., 2006; Woodhead and Hergt, 2005; Yuan et al., 2008). LA-ICPMS U-Pb zircon analytical data is summarized in Supplementary Table 7.

In situ zircon Lu-Hf isotopic analyses were performed using a Nu Plasma multi-collector ICP-MS, coupled to a Photon Machines 193 nm ArF excimer laser system at CCFS (Macquarie University). The analyses were carried out using the Nu Plasma time-resolved analysis software. The methods, including calibration and correction for mass bias, are described by (Griffin et al., 2004; Griffin et al., 2000). The ablation spots (55 µm) for the Hf isotope analyses were situated close to the U-Pb analysis positions on each grain. The accuracy of the Yb and Lu corrections during LA-MC-ICPMS analysis of zircon has been demonstrated by repeated analysis of standard zircons with a range in ^{176}Yb/^{177}Hf and ^{176}Lu/^{177}Hf. Four secondary standards (Mud Tank and Temora) were analyzed between every ten unknowns to check instrumental stability. ^{176}Hf/^{177}Hf ratios of the Mud Tank zircon gave an average of 0.2825355±0.0000041 (2SD; n=122); those of Temora gave 0.2826971±0.0000078 (2SD; n=42). These values are identical to those recommended for Mud Tank (0.282507±0.000003) and Temora (0.282693±0.000052)

(Fisher et al., 2014). The isobaric interferences of ^{176}Lu and ^{176}Yb on ^{176}Hf are very limited, because of the extremely low ratios of Lu/Hf and Yb/Hf in the measured standard zircons. The interference of ^{176}Yb on ^{176}Hf was corrected by measuring the interference-free ^{172}Yb isotope and using ^{176}Yb/^{172}Yb to calculate ^{176}Yb/^{177}Hf. The appropriate value of ^{176}Yb/^{172}Yb was determined by successive spiking the JMC475 Hf standard (1 ppm solution) with Yb, and iteratively finding the value of ^{176}Yb/^{172}Yb required to yield the value of ^{176}Hf/^{177}Hf obtained on the pure Hf solution (Griffin et al., 2004; Griffin et al., 2000). Zircon Hf isotope data are presented in Supplementary Table 8.

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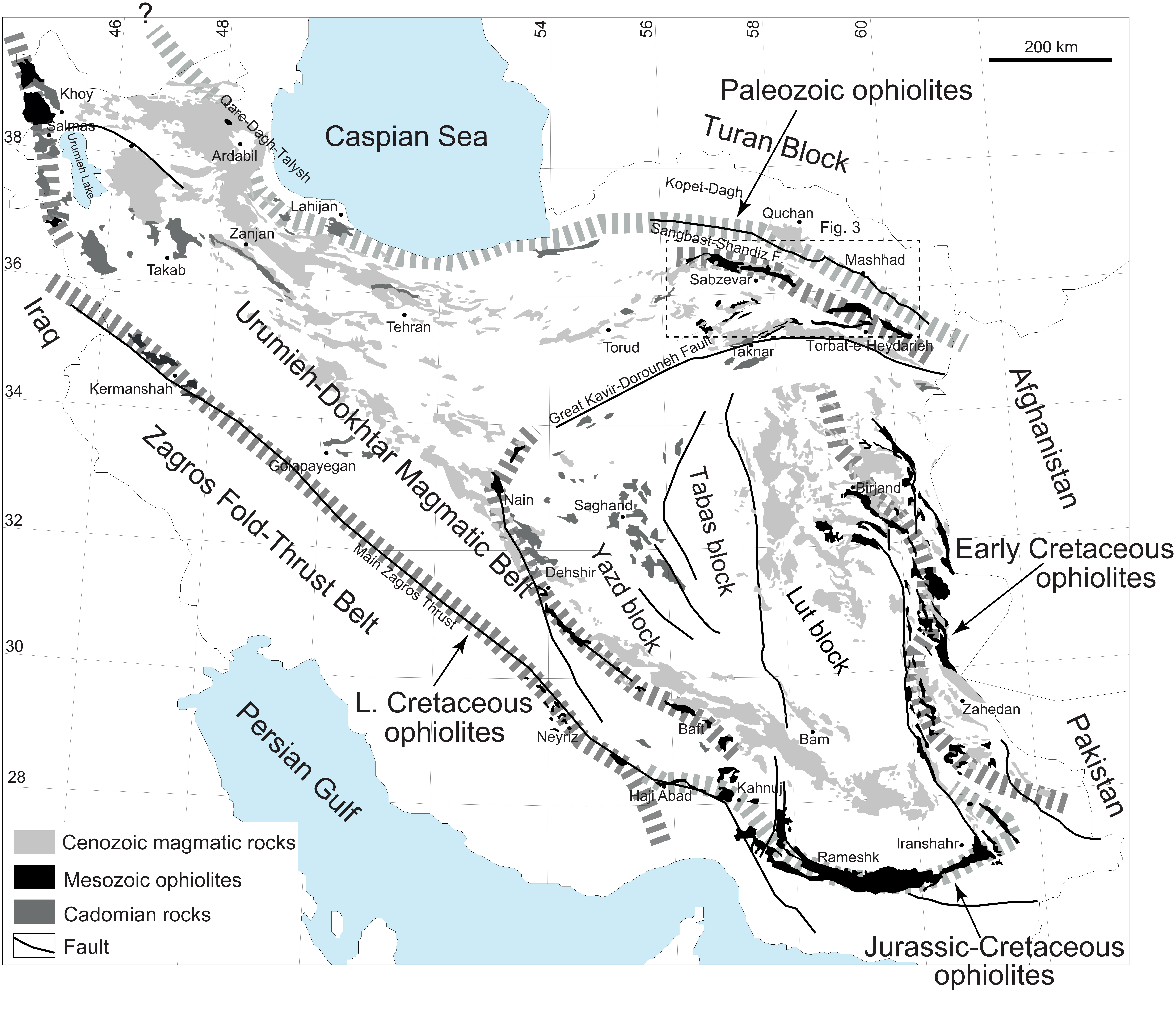
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Paleozoic ophiolites
Turan Block
Caspian Sea
Qare-Dagh-Talysh
Khoy
Salmas
Urumieh Lake
Ardabil
Lahijan
Zanjan
Takab
Kopet-Dagh
Quchan
Fig. 3
Sangbast-Shandiz F.
Mashhad
Sabzevar
Tehran
Torud
Taknar
Torbat-e-Heydarieh
Great Kavir-Dorouneh Fault
Afghanistan
Iraq
Kermanshah
Urumieh-Dokhtar Magmatic Belt
Golpayegan
Nain
Saghand
Tabas block
Birjand
Early Cretaceous ophiolites
Yazd block
Dehshir
Lut block
Zagros Fold-Thrust Belt
Main Zagros Thrust
L. Cretaceous ophiolites
Neyriz
Baft
Bam
Zahedan
Pakistan
Persian Gulf
Haji Abad
Kahnuj
Iranshahr
Rameshk
Jurassic-Cretaceous ophiolites
Cenozoic magmatic rocks
Mesozoic ophiolites
Cadomian rocks
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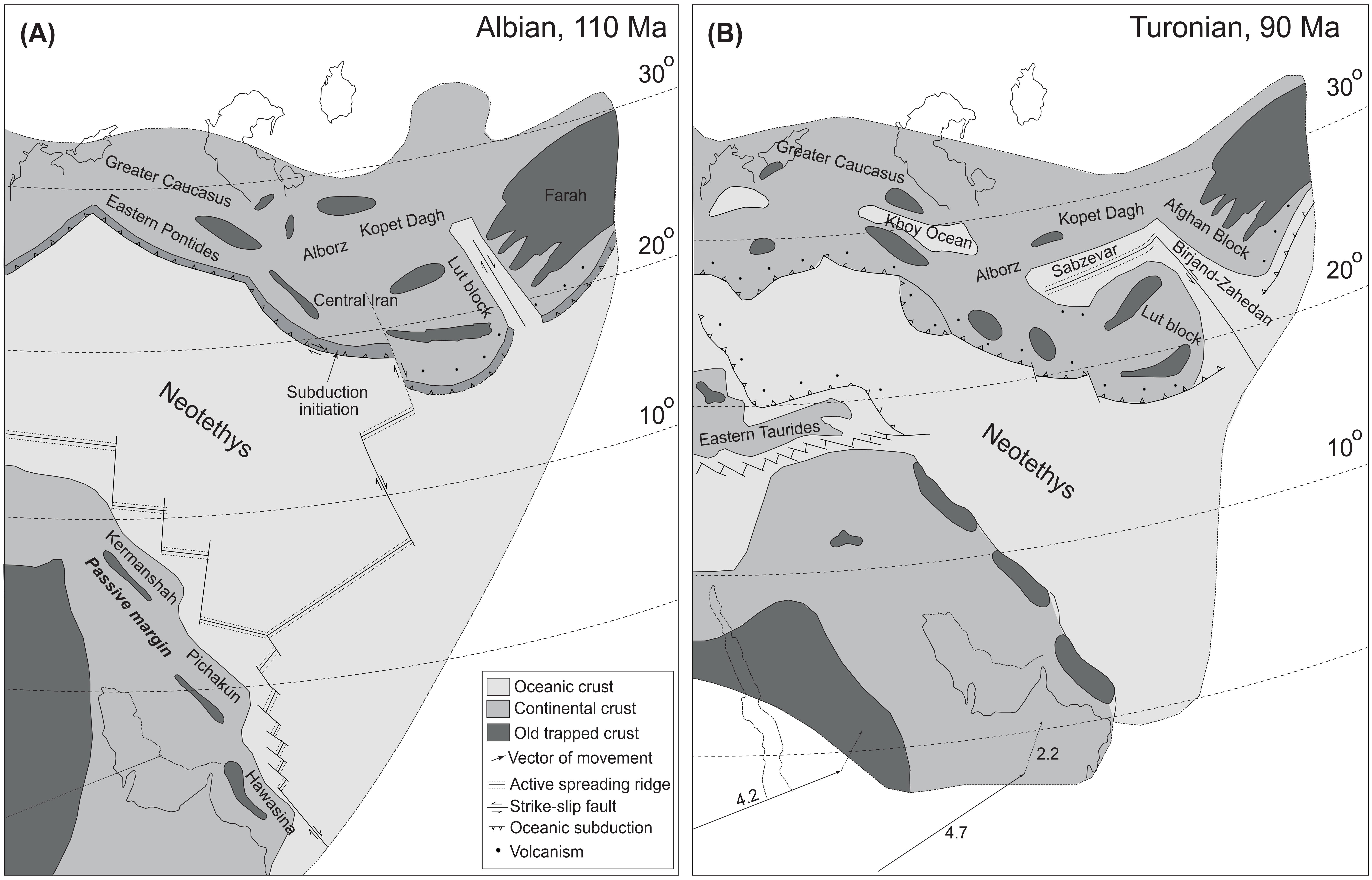
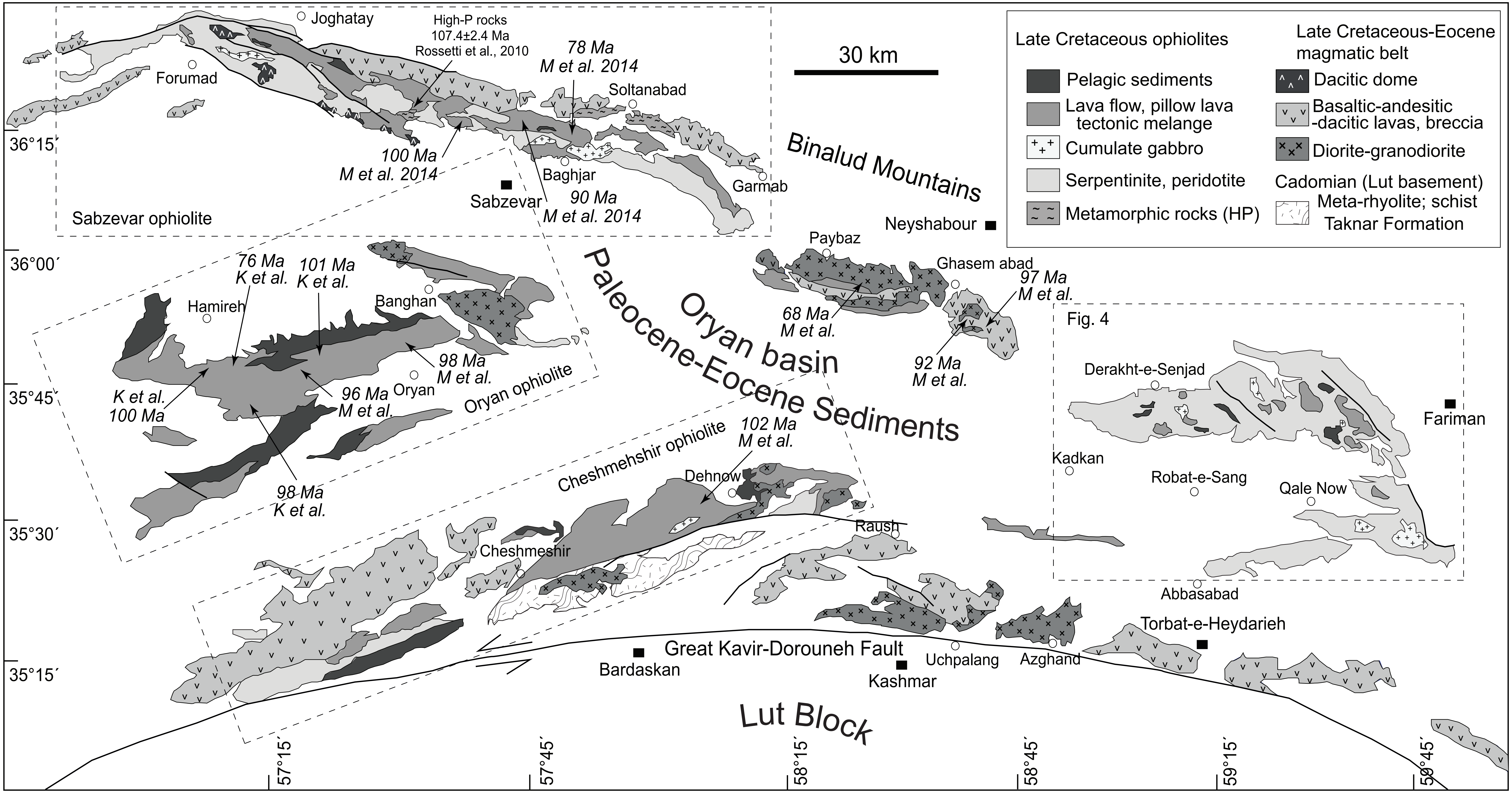
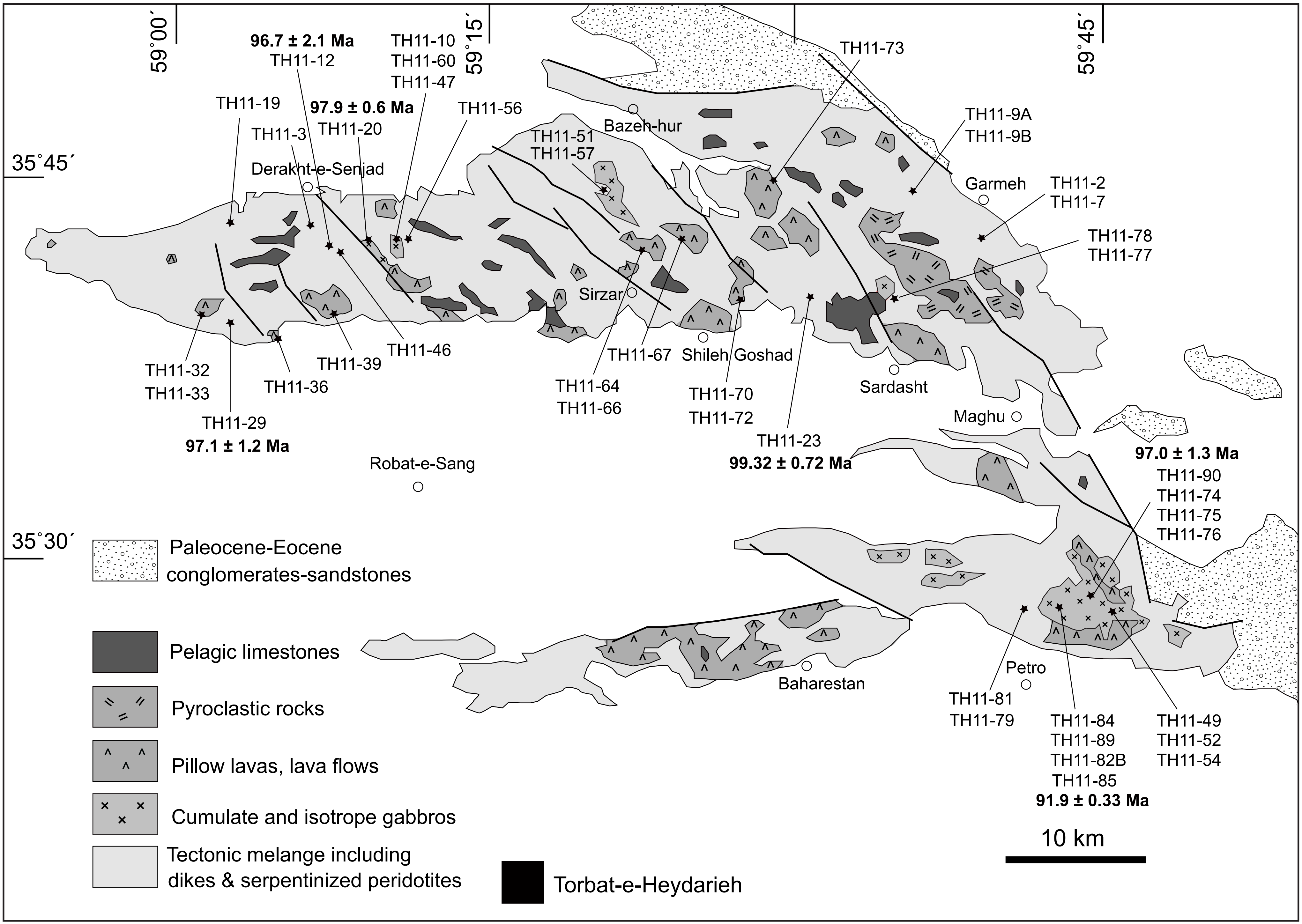


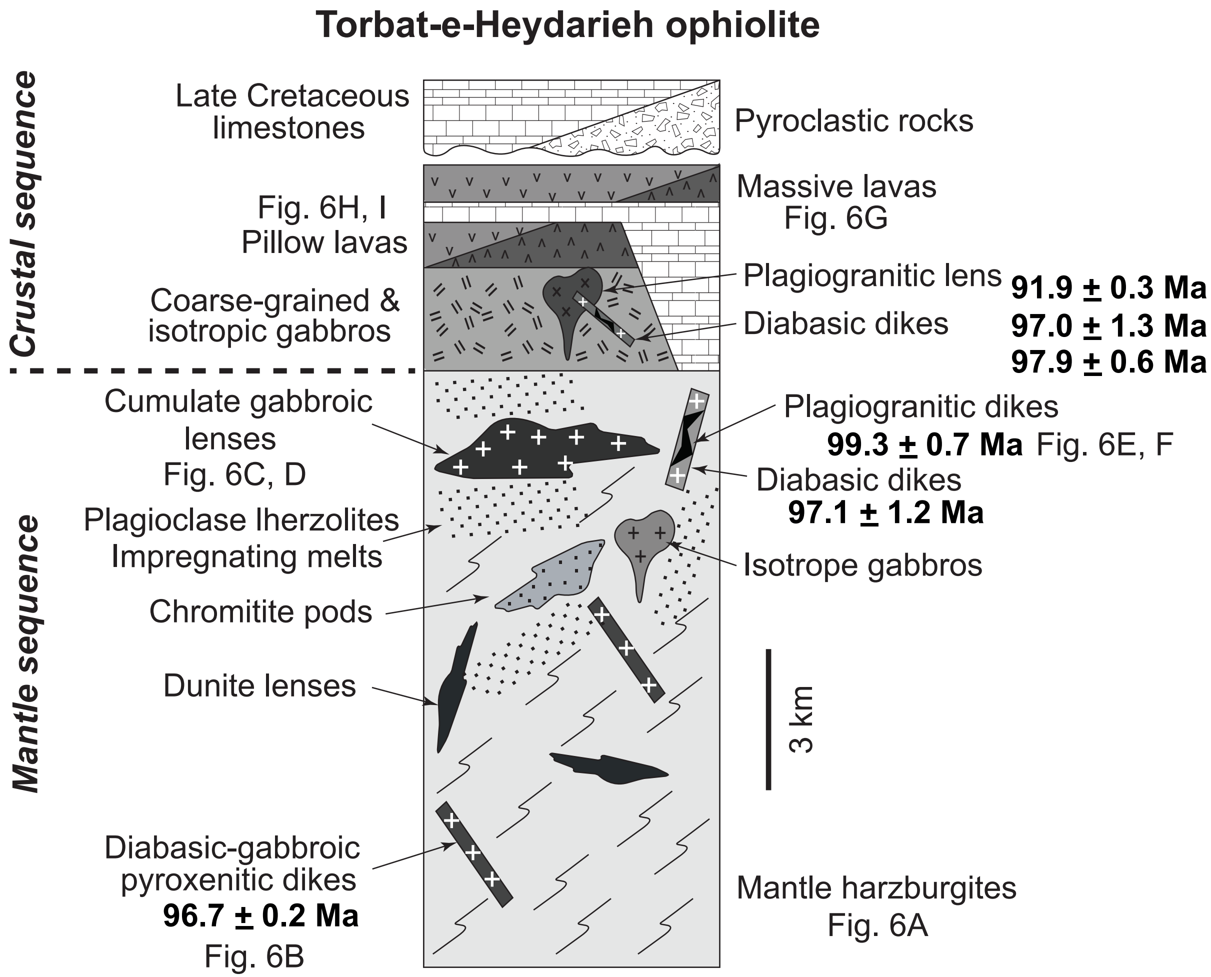
(A)
Albian, 110 Ma
30°
20°
10°
Greater Caucasus
Eastern Pontides
Alborz
Kopet Dagh
Farah
Lut block
Central Iran
Subduction initiation
Neotethys
Kermanshah
Passive margin
Pichakun
Hawasina
Oceanic crust
Continental crust
Old trapped crust
Vector of movement
Active spreading ridge
Strike-slip fault
Oceanic subduction
Volcanism
(B)
Turonian, 90 Ma
30°
20°
10°
Greater Caucasus
Khoy Ocean
Alborz
Kopet Dagh
Afghan Block
Sabzevar
Birjand-Zahedan
Lut block
Eastern Taurides
Neotethys
4.2
2.2
4.7

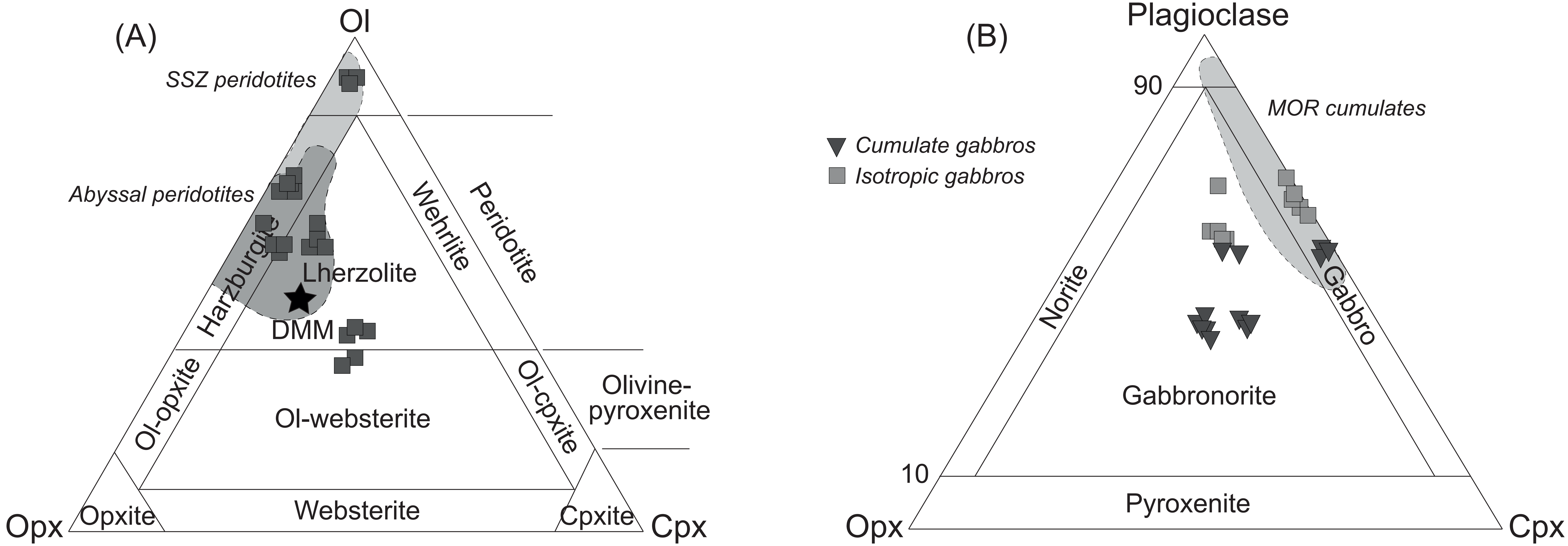


Late Cretaceous ophiolites
Pelagic sediments
Lava flow, pillow lava tectonic melange
Cumulate gabbro
Serpentinite, peridotite
Metamorphic rocks (HP)
Late Cretaceous-Eocene magmatic belt
Dacitic dome
Basaltic-andesitic -dacitic lavas, breccia
Diorite-granodiorite
Cadomian (Lut basement)
Meta-rhyolite; schist Taknar Formation
30 km
Joghatay
Forumad
High-P rocks 107.4±2.4 Ma Rossetti et al., 2010
78 Ma M et al. 2014
Soltanabad
100 Ma M et al. 2014
90 Ma M et al. 2014
Baghjar
Sabzevar
Garmab
Sabzevar ophiolite
Binalud Mountains
Neyshabour
Paybaz
Ghasem abad
68 Ma M et al.
97 Ma M et al.
92 Ma M et al.
76 Ma K et al.
101 Ma K et al.
Hamireh
Banghan
98 Ma M et al.
96 Ma M et al.
K et al. 100 Ma
Oryan
Oryan ophiolite
98 Ma K et al.
Oryan basin
Paleocene-Eocene Sediments
Fig. 4
Derakht-e-Senjad
Fariman
Kadkan
Robat-e-Sang
Qale Now
Cheshmehshir ophiolite
102 Ma M et al.
Dehnow
Raush
Cheshmeshir
Abbasabad
Torbat-e-Heydarieh
Great Kavir-Dorouneh Fault
Bardaskan
Kashmar
Uchpalang
Azghand
Lut Block
36°15´
36°00´
35°45´
35°30´
35°15´
57°15´
57°45´
58°15´
58°45´
59°15´
59°45´

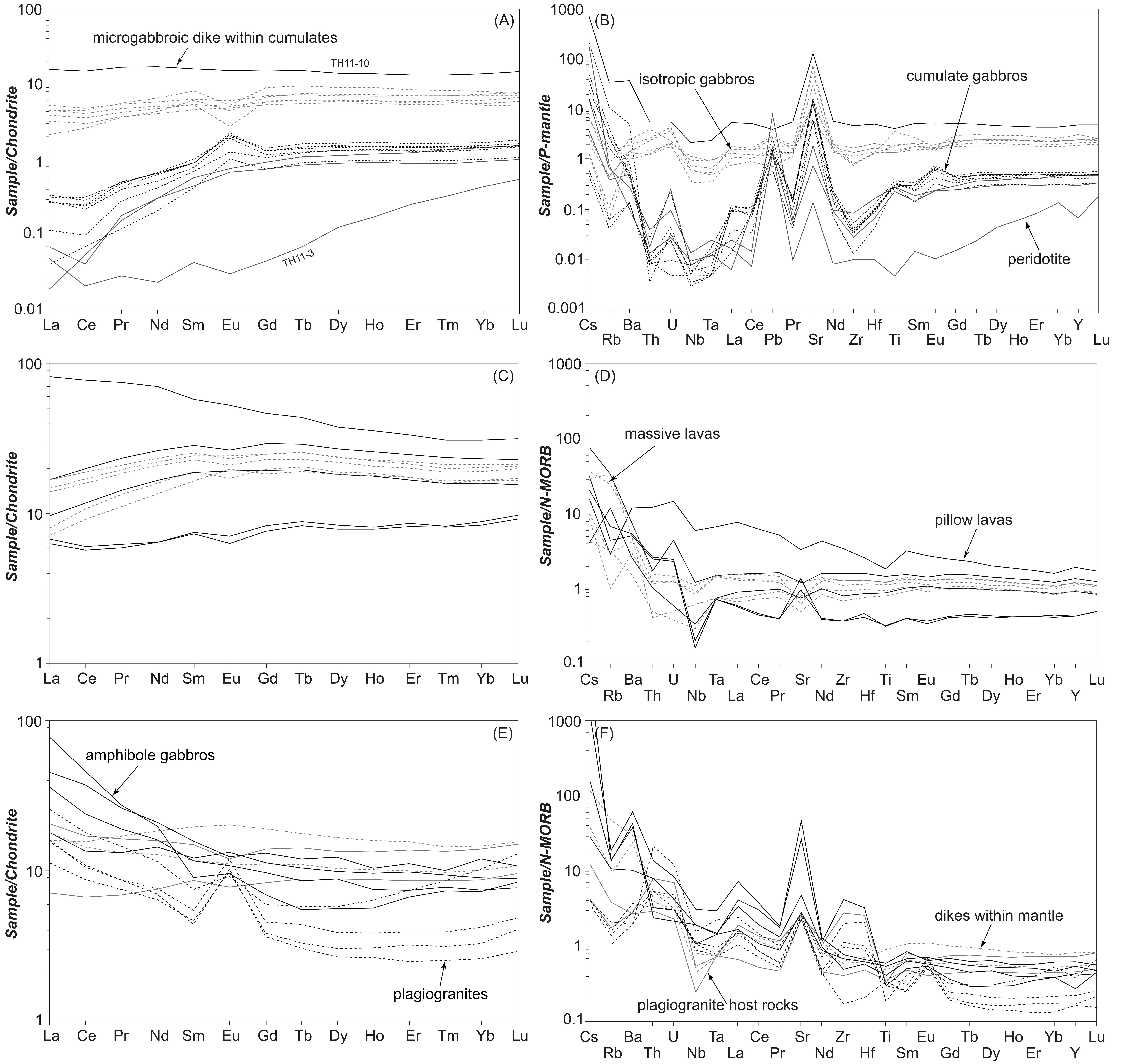


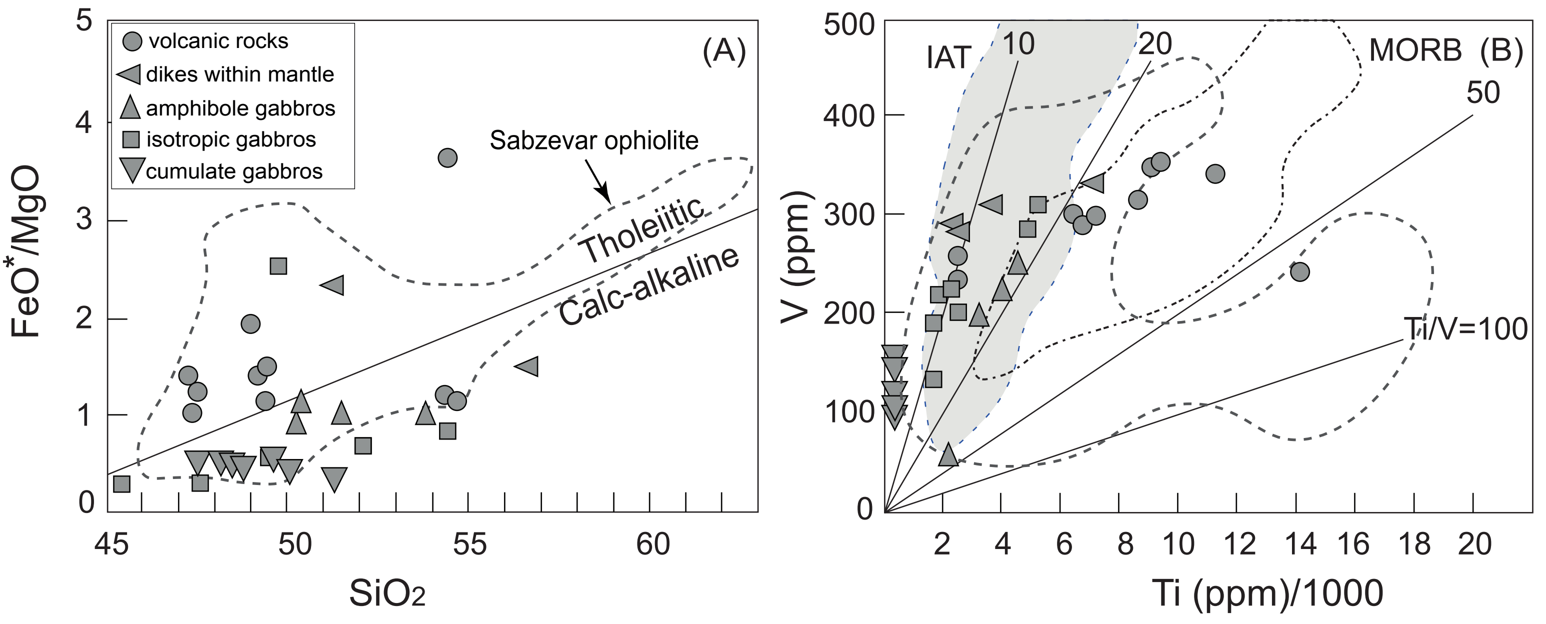
59°00´
59°15´
59°45´
35°45´
35°30´
96.7 ± 2.1 Ma
TH11-12
TH11-10
TH11-60
TH11-47
97.9 ± 0.6 Ma
TH11-20
TH11-19
TH11-3
TH11-56
TH11-73
TH11-51
TH11-57
Bazeh-hur
Derakht-e-Senjad
TH11-9A
TH11-9B
Garmeh
TH11-2
TH11-7
TH11-78
TH11-77
Sirzar
TH11-32
TH11-33
TH11-36
TH11-39
TH11-46
TH11-67
Shileh Goshad
TH11-64
TH11-66
TH11-70
TH11-72
Sardasht
Maghu
TH11-29
97.1 ± 1.2 Ma
TH11-23
99.32 ± 0.72 Ma
Robat-e-Sang
97.0 ± 1.3 Ma
TH11-90
TH11-74
TH11-75
TH11-76
Paleocene-Eocene conglomerates-sandstones
Pelagic limestones
Pyroclastic rocks
Pillow lavas, lava flows
Cumulate and isotrope gabbros
Tectonic melange including dikes & serpentinized peridotites
Torbat-e-Heydarieh
Baharestan
Petro
TH11-81
TH11-79
TH11-84
TH11-89
TH11-82B
TH11-85
91.9 ± 0.33 Ma
TH11-49
TH11-52
TH11-54
10 km


Torbat-e-Heydarieh ophiolite
Crustal sequence
Mantle sequence
Late Cretaceous limestones
Pyroclastic rocks
Massive lavas
Fig. 6G
Fig. 6H, I
Pillow lavas
Coarse-grained & isotropic gabbros
Plagiogranitic lens
Diabasic dikes
91.9 ± 0.3 Ma
97.0 ± 1.3 Ma
97.9 ± 0.6 Ma
Cumulate gabbroic lenses
Fig. 6C, D
Plagiogranitic dikes
99.3 ± 0.7 Ma Fig. 6E, F
Diabasic dikes
97.1 ± 1.2 Ma
Plagioclase lherzolites
Impregnating melts
Isotrope gabbros
Chromitite pods
Dunite lenses
3 km
Diabasic-gabbroic pyroxenitic dikes
96.7 ± 0.2 Ma
Fig. 6B
Mantle harzburgites
Fig. 6A

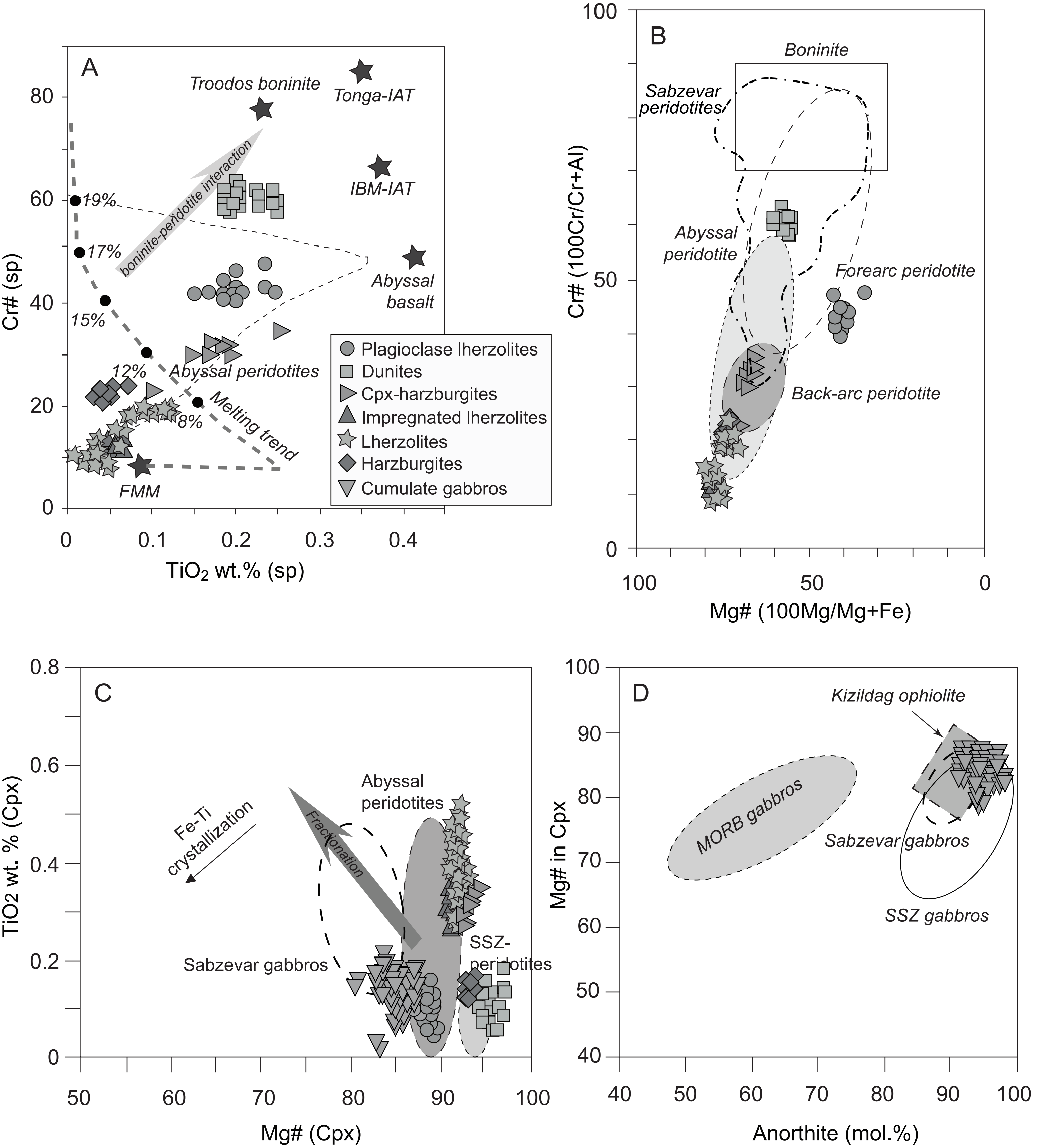


(A)
Ol
SSZ peridotites
Abyssal peridotites
Harzburgite
Lherzolite
DMM
Wehrlite
Peridotite
Ol-opxite
Ol-websterite
Ol-cpxite
Olivine-pyroxenite
Opxite
Websterite
Cpxite
Opx
Cpx
(B)
Plagioclase
90
MOR cumulates
Cumulate gabbros
Isotropic gabbros
Norite
Gabbro
Gabbronorite
10
Pyroxenite
Opx
Cpx

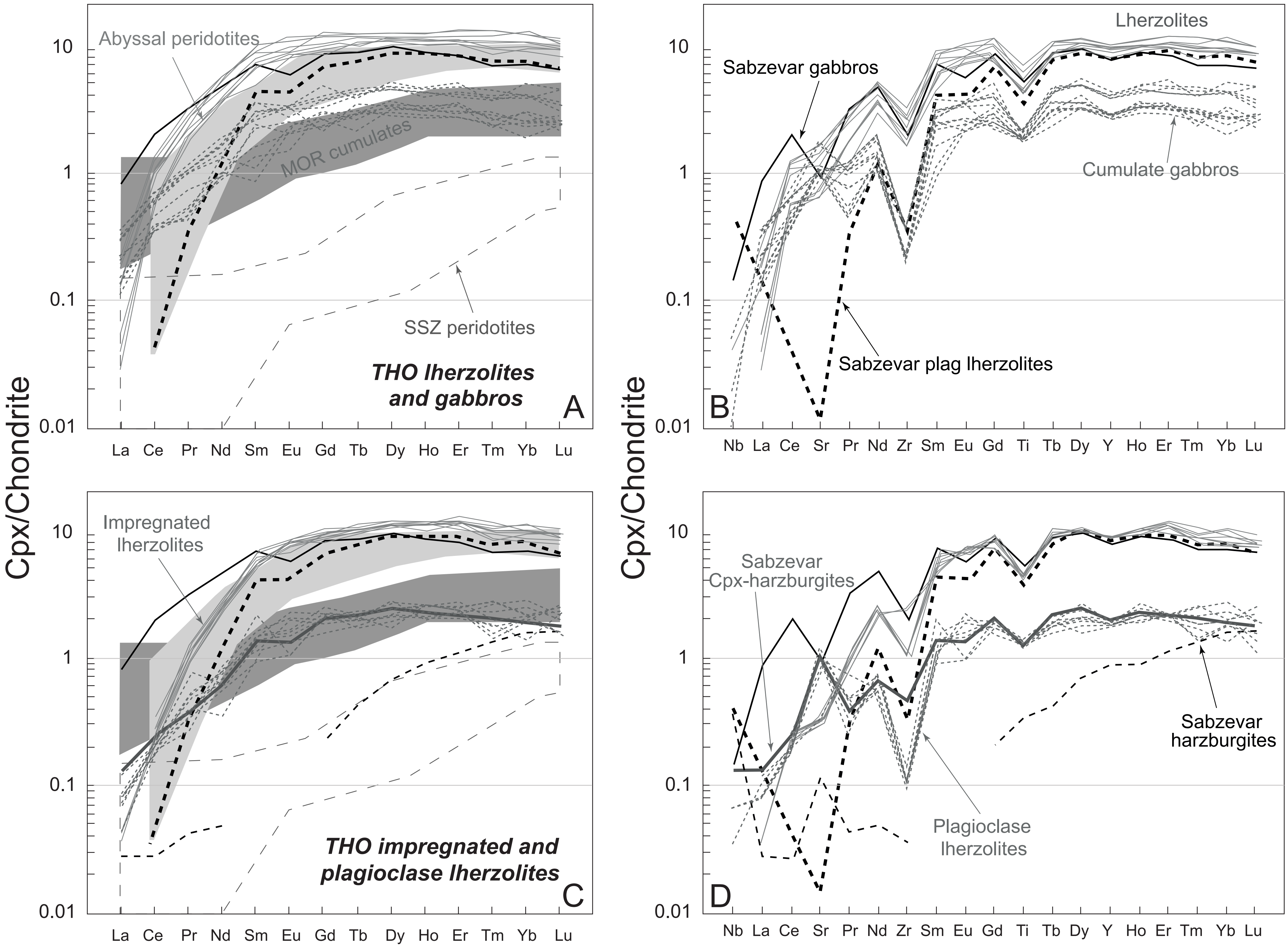

(A)
microgabbroic dike within cumulates
TH11-10
TH11-3
Sample/Chondrite
La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
(B)
isotropic gabbros
cumulate gabbros
peridotite
Sample/P-mantle
Cs Rb Ba Th U Nb Ta La Ce Pb Pr Sr Nd Zr Hf Ti Sm Eu Gd Tb Dy Ho Er Yb Y Lu
(C)
Sample/Chondrite
La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
(D)
massive lavas
pillow lavas
Sample/N-MORB
Cs Rb Ba Th U Nb Ta La Ce Pr Sr Nd Zr Hf Ti Sm Eu Gd Tb Dy Ho Er Yb Y Lu
(E)
amphibole gabbros
plagiogranites
Sample/Chondrite
La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
(F)
dikes within mantle
plagiogranite host rocks
Sample/N-MORB
Cs Rb Ba Th U Nb Ta La Ce Pr Sr Nd Zr Hf Ti Sm Eu Gd Tb Dy Ho Er Yb Y Lu

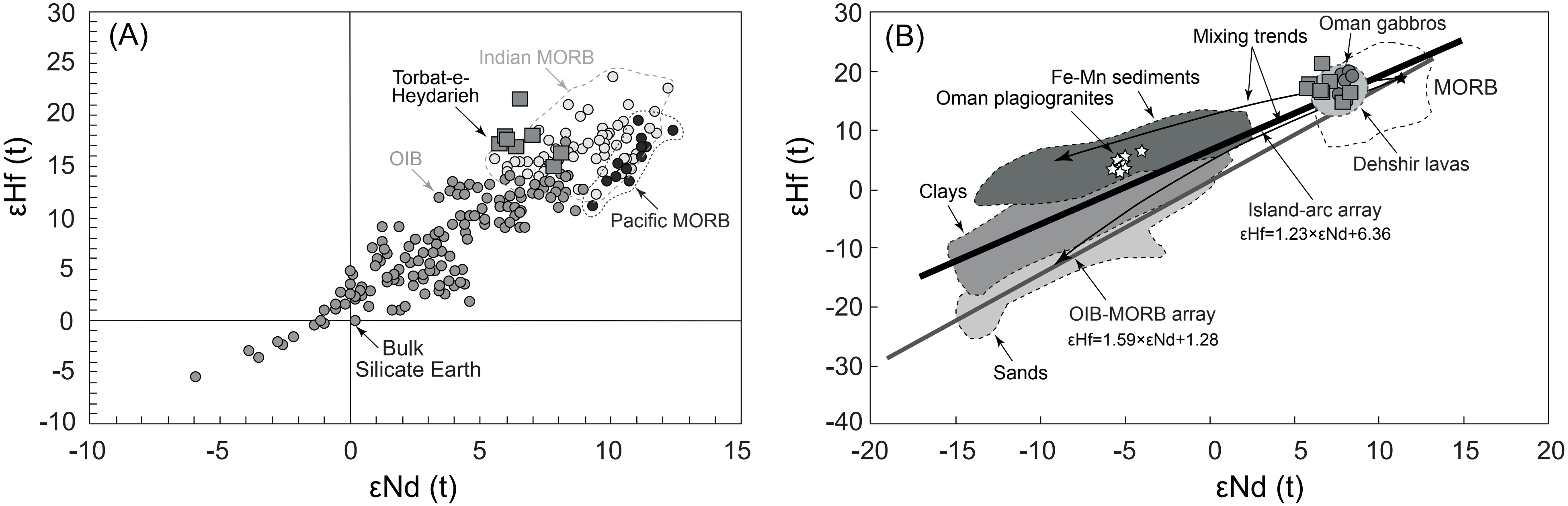


volcanic rocks
dikes within mantle
amphibole gabbros
isotropic gabbros
cumulate gabbros
(A)
Sabzevar ophiolite
Tholeiitic
Calc-alkaline
FeO*/MgO
SiO2
IAT
10
20
MORB
(B)
50
Ti/V=100
V (ppm)
Ti (ppm)/1000

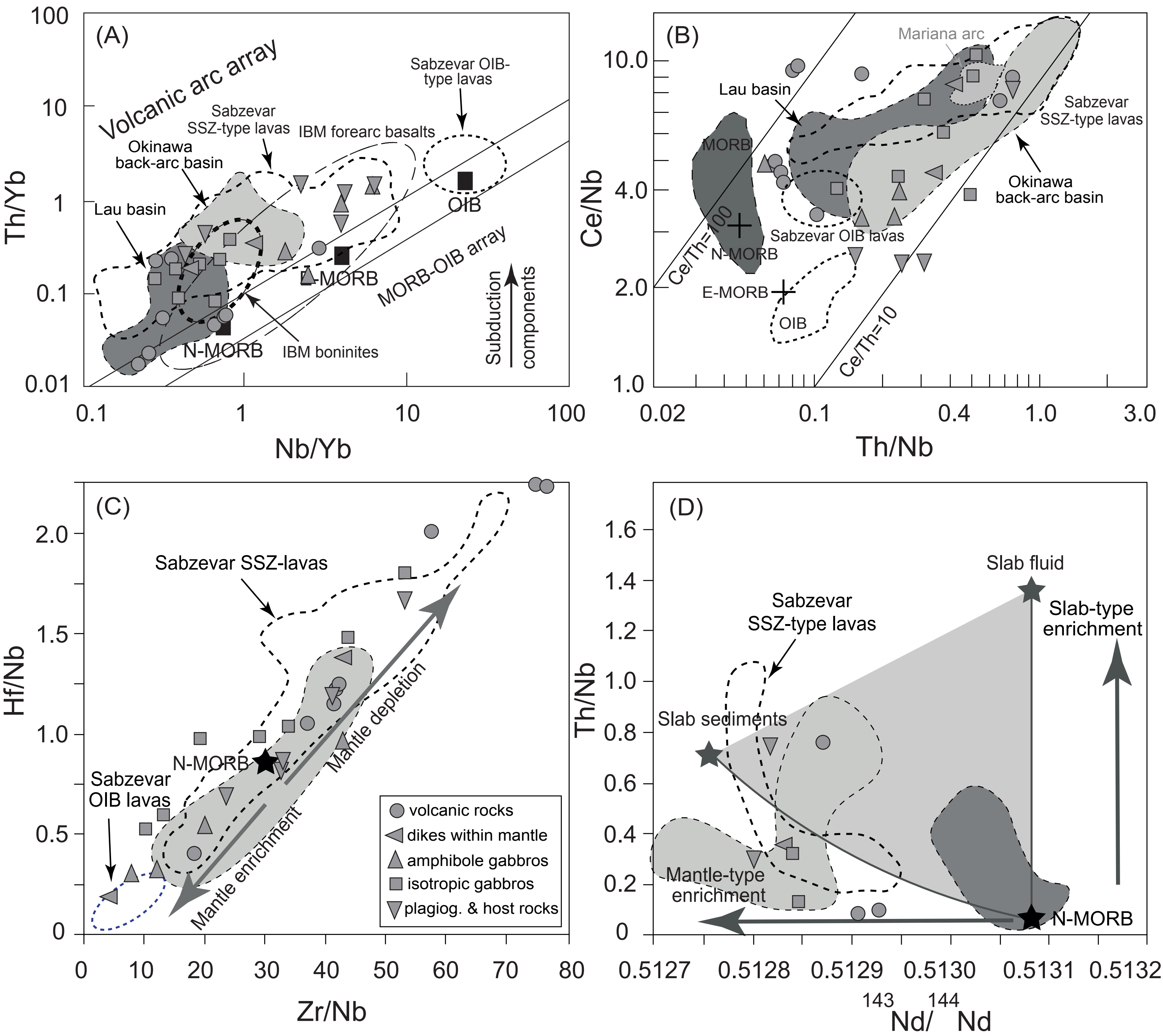


A
Troodos boninite
Tonga-IAT
IBM-IAT
Abyssal basalt
boninite-peridotite interaction
19%
17%
15%
12%
8%
Abyssal peridotites
Melting trend
FMM
Cr# (sp)
TiO2 wt.% (sp)
Plagioclase lherzolites
Dunites
Cpx-harzburgites
Impregnated lherzolites
Lherzolites
Harzburgites
Cumulate gabbros
B
Boninite
Sabzevar peridotites
Abyssal peridotite
Forearc peridotite
Back-arc peridotite
Cr# (100Cr/Cr+Al)
Mg# (100Mg/Mg+Fe)
C
Fe-Ti crystallization
Fractionation
Abyssal peridotites
Sabzevar gabbros
SSZ-peridotites
TiO2 wt. % (Cpx)
Mg# (Cpx)
D
Kizildag ophiolite
MORB gabbros
Sabzevar gabbros
SSZ gabbros
Mg# in Cpx
Anorthite (mol.%)

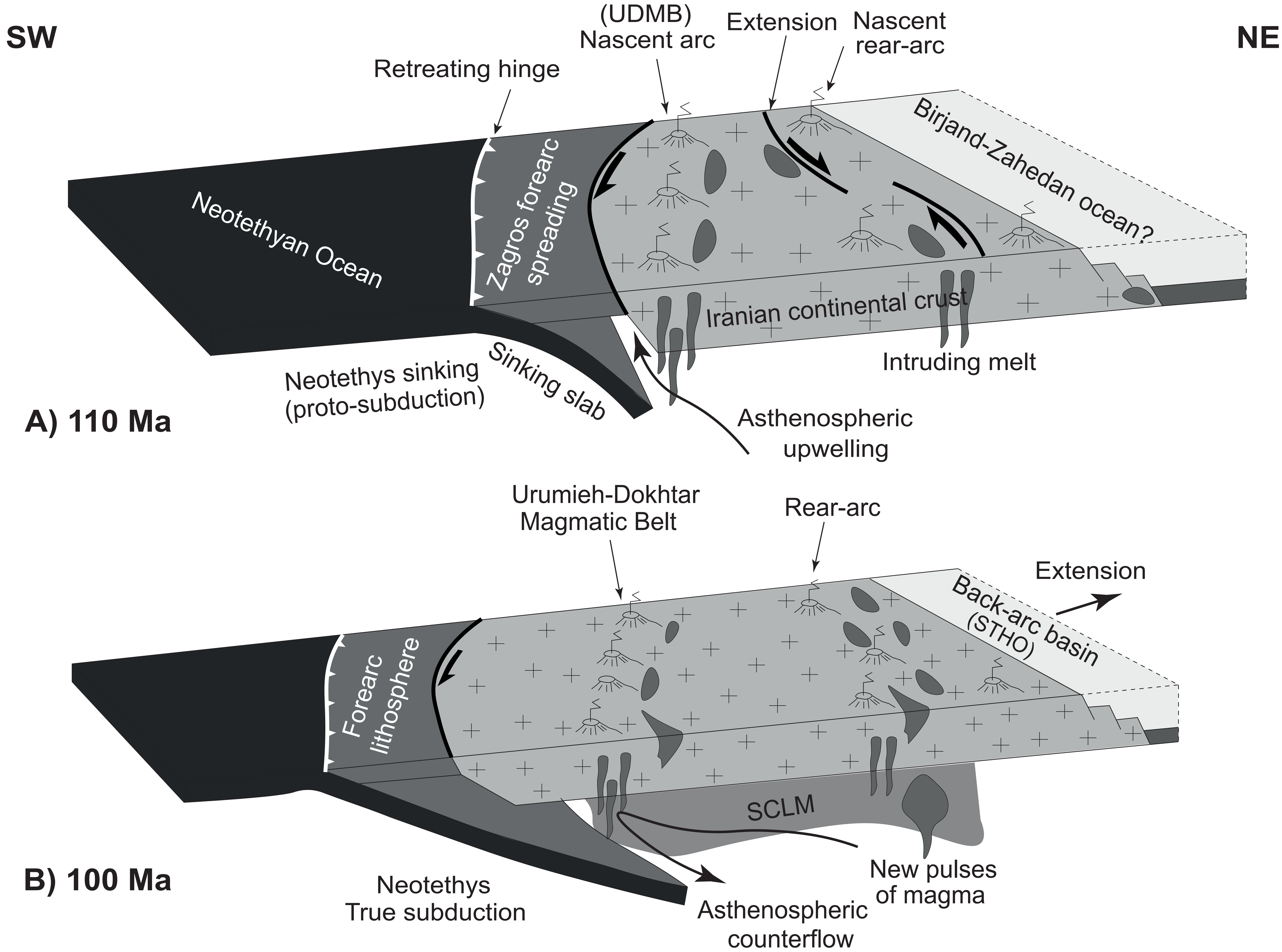

Cpx/Chondrite
Abyssal peridotites
MOR cumulates
SSZ peridotites
THO lherzolites and gabbros
A
B
Lherzolites
Sabzevar gabbros
Cumulate gabbros
Sabzevar plag lherzolites
Impregnated lherzolites
THO impregnated and plagioclase lherzolites
C
D
Sabzevar Cpx-harzburgites
Sabzevar harzburgites
Plagioclase lherzolites
La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
Nb La Ce Sr Pr Nd Zr Sm Eu Gd Ti Tb Dy Y Ho Er Tm Yb Lu
0.01
0.1
1
10



(A)
Indian MORB
Torbat-e-
Heydarieh
OIB
Pacific MORB
Bulk
Silicate Earth
εHf (t)
εNd (t)
(B)
Mixing trends
Oman gabbros
Fe-Mn sediments
Oman plagiogranites
MORB
Dehshir lavas
Clays
Island-arc array
εHf=1.23×εNd+6.36
OIB-MORB array
εHf=1.59×εNd+1.28
Sands
εHf (t)
εNd (t)



(A)
Volcanic arc array
Sabzevar OIB-type lavas
Sabzevar SSZ-type lavas
IBM forearc basalts
Okinawa back-arc basin
Lau basin
OIB
E-MORB
MORB-OIB array
N-MORB
IBM boninites
Subduction components
Th/Yb
Nb/Yb
(B)
Mariana arc
Lau basin
MORB
Sabzevar SSZ-type lavas
Okinawa back-arc basin
Sabzevar OIB lavas
N-MORB
E-MORB
OIB
Ce/Th=100
Ce/Th=10
Ce/Nb
Th/Nb
(C)
Sabzevar SSZ-lavas
N-MORB
Mantle depletion
Mantle enrichment
Sabzevar OIB lavas
volcanic rocks
dikes within mantle
amphibole gabbros
isotropic gabbros
plagiog. & host rocks
Hf/Nb
Zr/Nb
(D)
Slab fluid
Slab-type enrichment
Sabzevar SSZ-type lavas
Slab sediments
Mantle-type enrichment
N-MORB
Th/Nb
$^{143}Nd/^{144}Nd$



SW
NE
(UDMB)
Nascent arc
Extension
Nascent
rear-arc
Retreating hinge
Birjand-Zahedan ocean?
Neotethyan Ocean
Zagros forearc
spreading
Iranian continental crust
Intruding melt
Neotethys sinking
(proto-subduction)
Sinking slab
Asthenospheric
upwelling
A) 110 Ma
Urumieh-Dokhtar
Magmatic Belt
Rear-arc
Extension
Back-arc basin
(STHO)
Forearc
lithosphere
SCLM
New pulses
of magma
B) 100 Ma
Neotethys
True subduction
Asthenospheric
counterflow