THE BAJGAN COMPLEX REVEALED AS EARLY CRETACEOUS SUBDUCTION COMPLEX: A NEW KEY TO UNRAVEL THE GEODYNAMICS OF MAKRAN (SOUTHEAST IRAN)

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ABSTRACT

1. INTRODUCTION

Subduction complexes represent fossil accretionary wedges developed at convergent margins where ocean-plate material has accumulated by off-scaping and underplating adjacent to the leading edge of the upper plate (e.g. Kusky et al., 2013, Wakita, 2015). The subduction complexes consist of assemblages of deep-sea pelagic deposits and minor ophiolites, variably deformed and metamorphosed at different depths during the accretionary processes. The subduction complexes are thus characterized by rock assemblages deformed under HP/LT metamorphic conditions, from blueschist to eclogite facies, where informations about the processes active during their accretion at deep in the subduction zone can be collected. These rock assemblages are able to supply a set of useful informations for the geodynamic reconstructions, as the location and the onset age of the subduction zones, the age of the oceanic lithosphere involved into the convergence, the accretion mechanisms, the timing of the subduction-related metamorphism, the exhumation processes, etc…. The identification of the fossil subduction complexes is thus crucial for the reconstruction of the geodynamic history in the worldwide orogenic belts.

In the Makran area, a wide accretionary wedge orginated by the north-dipping subduction of the Neo-Tethyan basin beneath the Central Iran and Afghan blocks is exposed on-land (McCall and Kidd, 1982; McCall, 1985; 1997; Glennie, 2000; Burg et al., 2008; 2013; Burg, 2018; Saccani et al., 2015; 2018; Monsef et al., 2018). The rear of the accretionary wedge is represented by the North Makran Domain that consists of an imbricate stack of tectonic units. These units are weakly deformed and metamorphosed and mainly representd by ophiolite units of Early Cretaceous age associated to Late Cretaceous tectonic and sedimentary mélanges (McCall, 1985; 2002; Ghazi et al., 2004; Shaker-Ardakani et al., 2009; Hunziker et al., 2015; Moslempour et al., 2015; Delavari et al., 2016; Saccani et al., 2018; Burg, 2018; Monsef et al., 2018). In the North Makran Domain is included also the Bajgan Complex, that in the available literature is interpreted as a metamorphic continental basement regarded as of Early Paleozoic age or older (e.g., Dorani et al., 2017). The occurrence of continental basement in the North Makran Domain represent a first-order constrain for all the geodynamics reconstruction so far proposed (Hunziker et al., 2015; Saccani et al., 2018; Burg, 2018; Monsef et al., 2018).

This paper provide a complete set of lithostratigraphic, geochemical, structural, metamorphic and geochronological data able to support the interpretation of the Bajgan Complex as a subduction Complex of Mesozoic age. The collected data indicate that the Bajgan Complex consists of Late Jurassic to Early Cretaceous metaophiolites and related pelagic metasediments, both affected by HP/LT metamorphism during the Late Cretaceous and then by green-schist metamorphism during their progressive exhumation up to shallow structural levels, that were reached in the Early Eocene time. The implication of these data for the geodynamic history of the Makran area is also discussed.

2. GEODYNAMIC FRAMEWORK OF MAKRAN

In the Makran region, SE Iran (Fig.1), one of the largest worldwide accretionary wedge is exposed for ~1000 km along strike (from west to east) between the Minab-Sabzevaran-Nayband dextral fault system and the Chaman-Ornach-Nal sinistral fault system (McCall and Kidd, 1982; Bayer et al., 2006; Burg et al., 2013). The southern boundary of the Makran accretionary wedge is represented by north-dipping, low-angle subduction zone originated by the underthrusting of the Neo-Tethys oceanic lithosphere of the Oman Plate, whereas its northern boundary with the Lut continental block is masked by the Quaternary cover of the Jaz Murian and Mashkel basins (Fig. 1). These basins are placed within a thinned continental crust of the Lut block (Abdetedal et al., 2015) characterized in its northward rim by an active volcanic arc resulting from the present-day northward subduction. The Makran accretionary wedge includes an active, southernmost off-shore portion and a northernmost on-shore portion that is well exposed on-land. The latter portion is in turn divided, from north to south, into North, Inner, Outer and Coastal Makran Domains (McCall, 1997; Burg et al., 2013; Burg, 2018). The Inner, Outer and Coastal Makran Domains consist trench and thrust-top deposits that document the southward (i.e., oceanward) progressive migration of the compressive front starting from the Middle Eocene (McCall, 1997; Burg et al., 2013; Burg, 2018). The rear of the Makran accretionary wedge is represented by the North Makran Domain that consists a stack of different tectonic units, all deformed and metamorphosed during the pre-Eocene tectonics. The reconstruction of geodynamic history of the North Makran Domain has been recently reconsidered in several papers (Hunziker et al., 2015; Saccani et al., 2018; Burg, 2018; Monsef et al., 2018). All these reconstructions indicate that the north-dipping subduction of the Neo-Tethys oceanic lithosphere below the southern margin of the Lut block was already active before the Early Cretaceous (e.g., Barrier et al., 2018). This subduction has been responsible during the Early Cretaceous for the development of a volcanic arc located in the southern rim of the Lut continental block (Saccani et al., 2018). Furthermore, the slab retreat of the subduction produced the opening of a supra-subduction oceanic basin (i.e. the North Makran Ocean) leading to the separation of a small continental microplate from the Lut block (Hunziker et al., 2015; Saccani et al., 2018). According to Saccani et al. (2018), the convergence of North Makran Ocean started in the uppermost Late Cretaceous as consequence the collision between an oceanic plateau built-up on the Neo-Tethys crust and the subduction zone. This event resulted in a subduction jump toward the south, but also in the closure of the North Makran Ocean that occurred before the Early Eocene. Recently, Barbero et al (2020) have furthermore suggested that the closure of the North Makran Ocean occurred by a new north-dipping subduction located at the rim of this ocean close to the Lut Block continental margin. Despite the different interpretations, all the available papers agree that the closure of the North Makran Ocean produced the deformation of the oceanic lithosphere and the subsequent involvement in the convergence of the continental microplate, leading to the present-day stack of tectonic units recognized in the North Makran Domain.

All the proposed reconstructions are based on the geodynamic interpretation of the different tectonic units from the North Makran Domain. From south to north and from bottom to top, six tectonic units have been identified in the North Makran Domain (Fig. 2): 1) the Coloured Mélange Complex (McCall and Kidd, 1982; McCall, 1985; Saccani et al., 2018), also known as the Imbricate Zone of Burg et al. (2013); 2) the southern ophiolites; 3) the Bajgan-Durkan Complex (McCall, 1985; 2002); 4) the northern ophiolites (also known as North Makran ophiolites; 5) the Deyader metamorphic Complex; 6) the Ganj Complex (Shaker-Ardakani et al., 2009; Barbero et al.,2020). The main deformations in these units are sealed by the Early Eocene sedimentary deposits that clearly indicates that the units of the North Maktan Domain has been extensively deformed during the pre-Eocene tectonic evolution of the North Makran (see also McCall, 1985, 2002). However, in several shear zones at the boundaries of the tectonic units, slices of Early Eocene deposits have been identified, thus suggesting that also the post-Eocene tectonics has played an important role.

The occurrence of the North Makran Ocean (e.g., McCall and Kidd, 1982; McCall, 2002; Burg, 2018) is testified by the North Makran ophiolites that are represented by three distinct tectonic units, which are: 1) the Band-e-Zeyarat/Dar Anar (Ghazi et al., 2004); 2) the

Remeshk/Mokhtarabad ophiolite (Moslempour et al., 2015; Monsef et al., 2018); 3) the Fanuj-Maskutan ophiolite (Desmons and Beccaluva, 1983). According to Barbero et al. (in press), the Band-e-Zeyarat/Dar Anar ophiolites represent an upper oceanic crustal section, including a well developed sequence of volcanic rocks showing showing either normal-type (N) or enriched-type (E) mid-ocean ridge basalt affinities (MORB). 40Ar-39Ar ages obtained from intrusive rocks yield ages of about 140–143 Ma (Ghazi et al., 2004) whereas U-Pb geochronological data on zircons from the same rocks gave ages ranging from 122 to 129 Ma. (Barbero et al., in press). In contrast, the Fanuj-Maskutan ophiolites show a complete sequence from mantle peridotites to pillow-lava flows and the sedimentary cover (Moslempour et al., 2015). The Remeshk/Mokhtarabad ophiolites include a mantle section consisting of harzburgites and impregnated lherzolites topped by gabbros, as well as basalts and andesites interlayered with deep-sea sedimentary rocks (Monsef et al., 2018; Burg, 2018). The Remeshk/Mokhtarabad and Fanuj-Maskutan ophiolites are regarded as remnants of a supra-subduction oceanic lithosphere.

The North Makran ophiolites are associated to the Deyader Metamorphic Complex that consists of metaophiolites and its metasedimentary sequences affected by HP-LT metamorphism (McCall, 1985; Hunziker et al., 2015; Omrani et al., 2017), whose age has been determined by K/Ar method as ranging from 100 to 81 Ma (Delaloye and Desmons, 1980).

These North Makran ophiolites are thrust by the the Ganj Complex, previously regarded as ophiolites (Shaker-Ardakani et al., 2009) but recently re-interpreted as a Turonian-Coniacian volcanic arc formed in an extensional intra-arc setting or, eventually in a proximal forearc environment, located in the southern margin of the Lut block (Barbero et al., 2020).

The North Makran Ophiolites as well as the Ganj Complex and the Bajgan-Durkan Complex are thrust on the Coloured Mélange Complex (Gansser, 1955, 1959; McCall, 1983; Saccani et al., 2018) consisting of metric- to decametric blocks of volcanic rocks, cherts, limestones, serpentinites, gabbros, and shales, as well as HP-LT metabasalts. The development of the Coloured Mélange Complex occurred in the late Paleocene, probably during the final stage of the collision of the oceanic plateau with the southern rim of the continental microplate (Saccani et al., 2018; Burg, 2018).

At the top of the Coloured Mélange Complex several slices of ophiolites known as southern ophiolites (Sorkhband and Rudan ophiolites) have been recognized (McCall, 2002; Delavari et al., 2016). Petrographic and geochemical data on the Sorkhband ophiolites indicate that they consist of two different tectonic slices, one derived from a MORB setting and the other from a supra-subduction zone setting (Delavari et al., 2016).

In this framework, the Bajgan-Durkan Complex plays an important role mainly because is regarded as representative of the remants of the continent microplate located south of the North Makran ocean. This complex occur at the base of the northern ophiolites and has been till now consisdered as a single tectonic unit but can be divided in two different complexes, the Bajgan and the Durkan Complexes, according to the occurrence of a main shear zones between them. The Durkan Complex consists of granitoids, alkaline lavas and associated shallow and deep marine sedimentary rocks of Middle Jurassic - Late Cretaceous age (McCall, 1985; Hunziker et al., 2015, Burg, 2018). However, Barbero et al (2020) have demonstrated that the Durkan Complex is indeed consisting of multiple slices derived from the same highly deformed Early Cretaceous – Paleocene carbonatic and volcanic successions interpreted as the remnants of disrupted seamount rather than continental margin successions, as it was previously described. The most important element for the geodynamic reconstructions is however represented by the Bajgan Complex, i.e. the topic of this paper, that has been till now interpreted as a metamorphic basement of the continental microplate, over which the sequence of the Durkan Complex has been deposited (e.g., McCall, 1985; Hunziker et al., 2015). The continental microplate from wich the Bajgan Complex is believed to be responsible in many reconstructions of the collision with the continental Lut block and the consequence closure of the North Makran oceanic basin (…..). The available literature indicate that the Bajgan Complex includes an assemblage of metamorphic rocks of Paleozoic age or older, (McCall, 1985; Dorani et al., 2017).

3. METHODS

To provide a complete picture of the features of the Bajgan Complex, an integrate, multidisciplinary approach has been carried out. Firstly, the geological mapping in selected key-areas has been performed in association to mesocale structural analyses on the different lithologies identified in the field. During the geological mapping, several rock samples for the different laboratory analyses has been collected. The laboratory studies include firstly the geochimical analyses of the sampled meta-igneous rocks coupled with the identification of their tectono-magmatic signficance. In addition, the ages of the meta-igneous rocks have been determined by U-Pb geochronology. The microscale features of the different deformation phases identified in the field have been stuided in thin section. The metamoprhism have been studied using different petrological thermobarometers in association to petrographic thermometers that have allowed a reconstruction of a P-T path for the metasedimentary rocks. The dating of the main foliation has been performed too by Ar-Ar geochronology. Finally, the age of the last magmatic event identified in the field has been performed by U-Pb geochronology. Datailed informations on the methods are reported in the supplementary material.

4. THE BAJGAN COMPLEX

The Bajgan Complex is exposed in the triangular area of about 2400 km^2. This complex consists of an assemblage of metamorphic rocks aligned along a NW-SE trending main foliation and includes different slices separated by a NW-SE trending mylonitic shear zones. The different slices show the same lithotypes with the same deformation history and the same metamorphic imprint. The main foliation and the shear zones in the Bajgan Complex are cut by swarm of not metamorphosed felsic dykes and covered by continental deposits quaternary in age.

The area where the Bajgan Complex crops out is bounded by the Sabzevaran Fault to the E, the Fanhuj-Nurabad line to W and the Rudan Thrust to the S. The Sabzevaran Fault and the Fanhuj-Nurabad line correspond to still active N-S trending destral strike-slip faults, whereas the Rudan Thrust is an important km-thick brittle shear zone along which the Bajgan Complex is thrust over the Sorkhband ophiolites and the Coloured Mélange Complex (Delavari et al., 2016). The involvement of slices of nummulite-bearing sandstones in the Rudan Thrust shear zone indicate its post-Middle Eocene age.

4.1.Lithostratigraphy

The Bajgan Complex shows a wide range of metamorphic rocks spanning from meta-serpentinites to meta-igneous and meta-sedimentary rocks.

The meta-serpentinites, mainly located in the NE area of the Bajgan Complex, are represented by huge body with massive fabric sorrounded by highly foliated zones. In the massive body the meta-serpentinites preserves relics of the primary assemblage represented by pyroxene minerals. Bands of chromitites are also recognized. The foliated meta-serpentinites are instead devoid of relics of primary mineral assemblage and the foliation is represented by anastomozing slip surfaces where the serpentine groups minerals are recrystallized.

The meta-igneous rocks are represented by both meta-volcanic and meta-intrusive rocks. The meta-intrusive mainly consists of meta-gabbros characterized by a strong partioning of the deformation with poorly-deformed meta--gabbros alternanting with foliated meta-gabbros and mylonitic meta-gabbros. The poorly-deformed meta-gabbros shows a well-preserved magmatic fabric with plagioclase and pyroxene as magmatic relics. The meta-gabbros show different grain-size and different mineralogical composition from gabbro, gabbronorite, melagabbro up to anortosite. A magmatic layering due to different mineralogical composion and/or different grain size has been also observed. The foliated meta-gabbros are instead characterized by a metamorphic structure made without relict magmatic structures consisting of layers of green minerals, mainly elongated grains of amphibole, alternating with leucocratic layers consisting of plagioclase. These bands are strongly folded and boudinaged. The only relic of the magmatic minerals is represented by scattered pyroxenes in the green layers. The mylonitic meta-gabbro are fine-grained strongly foliated rocks with rare porphyroclasts of magmatic pyroxene. In association with meta-gabbros, also small stocks of meta-plagiogranites also occur.

The meta-volcanic rocks occur as thick lenticular bodies of banded, well foliated and lineated amphibolite. The bands show different mineralogical composition with amphibole- and plagioclase-rich mm-thick layers. No magmatic relics have been identified in the meta-volcanic rocks.

The meta-sedimentary rocks includes meta-volcanoclastites, quarzites, micaschists, paragneisses, calcschists and impure marbles. The meta-volcanoclastites occur as foliated epidote- and albite-rich schists with variable percentage of mica-rich schists. The meta-volcanoclastites occur in association with quarzites, represented by quartz-rich layers alternanting with thin layers of well-foliated micaschists. The impure marbles are instead composed of alternating fine-grained marble layers and calcschists layers. Fine- to medium-grained impure massive marbles are also recognized. In some places, probably at the top of the marbles, also thin layers of micascists also occur. The micaschists are the most represented lithology. They consists of monotonous thick layers of fine- up medium-grained, well foliated quartz-rich to quartz-free micaschists. The micaschists show a transition to thick layers of fine-grained mica-rich and fine- up coarse-grained, quartz-rich paragneisses.

4.2. Geochemistry of the magmatic rocks

A total of thirty-one samples of metamorphic rocks derived from magmatic protoliths were taken for petrographic and whole rock geochemical studies with the aim of assessing the geochemical nature and the tectono-magmatic setting of the magmatic events that are recorded into the Bajgan Complex. We will therefore focus our discussion on the magmatic processes that can be inferred from these rocks, regardless of the late metamorphic processes that affected the Bajgan Complex rocks.

In following sections we will focus our discussion on the chemical features of the magmatic protoliths, regardless of their stratigraphic position. From a chemical point of view, we generally recognized different types of magmatic protoliths, from mafic intrusive, to mafic volcanic/subvolcanic rocks up to acidic intrusive rocks. Since the purpose of this section is to use the geochemical features of these rocks for assessing their tectono-magmatic setting of formation, the studied rocks will be further subdivided into three different chemical types that will be described in detail in the next sections (sections 6.2.2, 6.2.3, and 6.2.4).

4.2.1. Group 1 magmatic protoliths (N-MORB) (CONTROLLARE SE ACRONIMI Già DEFINITI PRIMA)

The protoliths of Group 1 rocks include basalt, ferrobasalt, troctolite, gabbro, and plagiogranite (Table P1). All these rocks display a clear sub-alkaline affinity, with Nb/Y ratio < 0.19 (Fig. P2). The meta-anorthosite MK704 shows high SiO_2, Al_2O_3, CaO, Na_2O, and Sr contents (Table P1) and low contents in trace elements, such as Th, Ta, Nb (Fig. P1c). The Chondrite-normalized REE pattern show moderate LREE/MREE and LREE/HREE enrichments (e.g., La_N/Sm_N = 1.93; La_N/Yb_N = 4.67), coupled with rather low HREE contents (e.g., Y_N = 4 times chondrite abundance, Sun and McDonough, 1989), as well as a marked Eu positive anomaly (Fig. P1d). Due to its cumulitic nature, a definition of the magmatic affinity of this anorthosite cannot be straightforwardly determined. Nonetheless, the LREE/HREE enrichment and Th (0.03 ppm), Ta (0.06 ppm), and Nb (0.65 ppm) contents are very low compared to those of anorthosites derived from enriched-type magmas (La_N/Yb_N = 10 - 35; Th = 0.2 - 3 ppm; Ta = 0.15 – 1.4 ppm; Nb = 1.7 – 14.1 ppm; Mukherjee et al., 2005; Ghose et al., 2008; Shellnutt et al. 2020). The marked depletion of these elements suggests that the magmatic protolith of this meta-anorthosite was most likely formed from a depleted N-MORB type parental liquid. In addition, the (Th/Tb)/(Th/Ta) ratio (i.e., ratio of hygromagmatophile elements ratios) is 0.24, which is in the range of variation for Group 1 metagabbros and metabasalts (0.18 – 0.26), whereas it is much lower than those of anorthosites derived from enriched-type magmas (1.8 – 3.5; Shellnutt et al. 2020).

Metagabbros are characterized by high MgO contents (10.9-14.8 wt%) and Mg# (84-69), as well as by high values of incompatible elements (e.g., Cr = 439-1069 ppm). By contrast, they show variable, but generally low abundances of incompatible elements (Table P1). These elements are particularly low in sample MK254 (e.g., TiO_2 = 0.36 wt%, P_2O_5 = 0.02 wt%, Zr=19.3 ppm, Y=6.68pp), which shows relicts of magmatic cumulitic texture, whereas they are comparatively higher in sample MK705A (e.g., TiO_2 = 0.67 wt%, Zr = 85.5 ppm, Y = 34.2ppm), which is a meta-isotropic gabbro.

Compared to metagabbros, metabasalts show relatively lower MgO (7.88-10.5 wt%), and incompatible elements (e.g., Cr = 118-336 ppm) contents, as well as Mg# (67-54) coupled with generally higher abundance of incompatible elements (e.g., TiO_2 = 1.28-1.44 wt%, P_2O_5 = 0.11-0.15 wt%, Zr = 78.7-109 ppm, Y = 21.8-40.0 ppm; see also Table P1). The meta-ferrobasalt is characterized by very high TiO_2 (3.85 wt%), FeO_t (16.73 wt%), and V (453 ppm) coupled with low Mg# (54). These values are in agreement with the occurrence of abundant Fe-Ti oxides, as observed in thin section. The incompatible elements spider diagrams show that Group 1 rocks share affinity with N-MORB basalts and gabbros, as suggested by flat N-MORB normalized patterns (Fig. P1c). High field strength elements (HFSE) contents range from ~0.2 to ~0.6 N-MORB abundance (Sun and McDonough, 1989) in the meta-cumulitic gabbro, whereas in meta-isotropic gabbros and meta-basaltic rocks they range from ~0.9 to ~1.6 times N-MORB abundance (Fig. P1c). The chondrite-normalized REE patterns further suggest an N-MORB affinity for Group 1 metagabbros and metabasalts. All these rocks show almost parallel patterns characterized by depletion of light REE (L-REE) with respect to medium (M-REE) and heavy (H-REE), with $(La/Sm)_N$ = 0.53 – 0.79 and $(La/Yb)_N$ = 0.51 – 0.81 (Fig. P1d). An N-MORB affinity for Group 1 rocks is further suggested by the discrimination diagram in Figure 3a, in which these rocks plot in the field for subduction-unrelated settings, within the compositional field for typical N-MORB (Saccani, 2015). An N-MORB affinity is also suggested by the co-variation of Zr/Nb and Zr/Y ratios as shown in Figure 3b, in which these rock plot in the field for N-MORB compositions.

The meta-plagiogranite is characterized by high SiO_2 content (72.58 wt%) and very low contents of TiO_2, P_2O_5, MgO, Cr, Co, Ni, and many other trace elements (Table P1). Given the very fractionated nature of this rock, Zr (83 ppm), Y (29 ppm), and Nb (3 ppm) contents are very low, thus suggesting an N-MORB affinity of its magmatic protolith.

4.2.2. Group 2 magmatic protoliths (E-MORB)

The protoliths of Group 2 rocks include basalt, gabbro, and plagiogranite (Table P1). All these rocks display a clear sub-alkaline affinity, with Nb/Y ratio ranging in mafic rocks from 0.24 to 0.37 (Fig. P2). Significantly, these values are intermediate between those of Group 1 and those of Group 3 rocks (see Fig. P2 and section 6.2.4).

No significant chemical differences can be observed between the metagabbro MK272 and meta-basalts. Therefore, the composition of this rock will be described together with those of meta-basalts. Group 2 mafic protoliths rocks show a small range of MgO (7.70- 10.55 wt%) and Mg# (74-63). Though fairly variable, TiO_2 (0.88-1.61 wt%), P_2O_5 (0.09-0.31 wt%), Nb (5.62-11.4 ppm), Y (15.5-30.9 ppm), and Zr (65.3-137 ppm) are generally high. Compatible elements contents are also variable, depending on the degree of fractionations of the different rocks and they show good positive correlations with MgO contents (Table P1). The only exception is represented by the meta-gabbro MK272 in which Cr content (752 ppm) is fairly high in relation to the MgO value, possibly reflecting a small amount of Cr-spinel fractionation in the magmatic protolith. In the N-MORB normalized (Sun and McDonough, 1989) incompatible elements spider diagram for Group 2 meta-basalts and meta-gabbro show regularly decreasing pattern from Th (Th_N = 4-9) to Yb (YbN = 0.55-1.1) (Fig. P1e). These patterns are very similar to that of the typical enriched-type MORB (E-MORB) of Sun and McDonough (1989). Chondrite-normalized REE patterns are characterized by slight LREE enrichment with respect to MREE and HREE (Fig. P1f), as testified by the $(La/Sm)_N$ and $(La/Yb)_N$ ratios that are in the range 1.38-1.67 and 2.18-3.20, respectively. These patterns and REE ratios are very similar to that of the typical E-MORB (Sun and McDonough, 1989).

In summary, the overall geochemical features of Group 2 meta-gabbro and meta-basalts point out for an E-MORB geochemical affinity of the magmatic protoliths, as also suggested by the diagrams in Figures 3a, b, where these rocks plot in the fields of typical E-MORB compositions (Sun and McDonough, 1989; Saccani, 2015).

The meta-plagiogranite is characterized by high SiO_2 content (71.19 wt%) and very low contents of TiO_2, P_2O_5, MgO, Cr, Co, Ni (Table P1). In contrast to the meta-plagiogranite of Group 1, this rock show high values of Y (115 ppm), Zr (374 ppm), Nb (35.1 ppm), Th (2.81 ppm), and Ta (1.56 ppm) at comparable degree of fractionation of the magmatic protoliths. The relatively high content in incompatible elements is exemplified in the N-MORB normalized incompatible elements pattern shown in Fig. P1e. Significantly, this pattern is characterized by negative anomalies of Ti and P, which reflect the crystallization of Fe-Ti oxides and apatite from a rather differentiated melt. The crystallization of these minerals is indeed observed in differentiated melt with ferrobasaltic-andesitic compositions in MORB-type magmatic series (e.g., Beccaluva et al.1983).

Though the overall REE content is significantly higher than those of meta-basalts and meta-gabbros, the meta-plagiogranite MK701 displays an REE pattern that is almost parallel to those of the meta-mafic rocks with slight LREE/MREE (La_N/Sm_N = 2.31) and LREE/HREE (La_N/Yb_N = 3.54) enrichments (Fig. P1f). However, this rock is characterized by a significant Eu negative anomaly, which reflect early crystallization and removal of plagioclase from the parental melt. The REE patterns (Fig. P1f) and the moderately high Nb, Ta, and Th abundances suggest that the magmatic protolith was formed as a consequence of extreme fractionation form an E-MORB type parental liquid.

4.2.3. Group 3 protoliths (alkaline basalts)

This group includes five amphibolites derived from basaltic protoliths, which show alkaline affinity, as suggested by the high (i.e., > 0.79) Nb/Y ratio (Table P1, Fig. P2). Group 3 basalts display rather homogeneous compositions. MgO contents are in the range 7.30-10.9 wt% and Mg# is generally around 60, with the only exception of sample MK247 (MgO = 7.30 wt%) that has Mg# = 53, likely suggesting a moderately fractionated nature of its magmatic protolith. (Table P1). TiO_2 (1.84-2.33 wt%) and P_2O_5 (0.16-0.56 wt%) contents are relatively high in agreement with the alkaline affinity of these rocks and generally higher than those of Group 1 and 2 rocks. Similarly, these rocks show higher contents of Nb (19.0 - 65.2 ppm) and Zr (148-324 ppm) compared to the other groups, whereas Y (19.6-37.3 ppm) abundance is comparable with Group 2 rocks. Quite different contents of Cr (66-350 ppm) and Ni (27-141 ppm) are observed in the different rock types (Table P1). The incompatible elements spider diagrams are characterized by marked enrichment in LILE with respect to HFSE and regularly decreasing patterns from Rb to Yb (Fig. P1g). Significantly, these patterns are well comparable with those of OIB-type basalts (Sun and McDonough, 1989). The REE patterns are characterized by marked enrichment of LREE with respect to MREE and HREE (Fig. P1f), as exemplified by the $(La/Sm)_N$

and $(La/Yb)_N$ ratios that are in the range 2.29-3.33 and 7.24-20.1, respectively. In addition, a distinctive feature of the Group 3 rocks is the high MREE and HREE ratios (Sm_N/Yb_N = 3.05 – 6.63). As shown in Figure P1h, these features are collectively well comparable with those of the typical OIB of Sun and McDonough, 1989). Accordingly, in the tectonic discrimination diagram of Figure P3a, Group 3 rocks plot in the field for subduction-unrelated settings, namely within the compositional field for typical P-MORB and OIB (Saccani, 2015). A similar conclusion can be found out from the Zr/Nb vs. Zr/Y diagram (Fig. 3b) in which the rocks of Group 3 plot close to the composition of the typical alkaline OIB (Sun and McDonough, 1989).

4.2.3. Tectono-magmatic significance

In the previous sections we have shown that, in contrast to previous interpretations, the metamorphic rocks of the Bajgan Complex were derived from magmatic protoliths belonging to an oceanic lithopshere displaying subduction-unrelated oceanic affinity (see Pearce, 2008, Dilek and Furnes, 2011; Saccani, 2015). Some trace-element ratios (e.g. Zr/Nb, Ce/Y, Th/Ta, Th/Tb) are little affected by fractional crystallization of predominantly olivine + clinopyroxene + plagioclase. Therefore, even in the presence of significant amounts of fractionation, they are believed to represent the elemental ratios in the source (e.g. Allègre and Minster, 1978). Ratios of incompatible elements (e.g., Zr/Nb, Th/Ta), ratios of hygromagmatophile element ratios ((Th/Ta)/(Th/Tb)) (Table 1), as well as distinct normalized multi-element and REE patterns (Fig. P1) suggest that the different Groups of magmatic protoliths from the Bajgan Complex have been originated from chemically distinct mantle sources. Group 1 rocks show high Zr/Nb (23.6 -47.9) and (Th/Ta)/(Th/Tb) (3.6-5.6) ratios (Table P1, Fig. P3b) suggesting that Group 1 basaltic protoliths were generated from a depleted-type mantle source. In contrast, Group 3 basaltic protoliths show low Zr/Nb (5.0 - 8.8) and (Th/Ta)/(Th/Tb) (0.2-0.5) ratios, suggesting that they were generated from an enriched-type mantle source (see Saccani et al., 2015, and references therein). Group 2 mafic protoliths show Zr/Nb (8.1 - 16.4) and (Th/Ta)/(Th/Tb) (0.80-1.29) ratios, which are generally low but intermediate between those of Group 1 and Group 3 rocks (Table P1, Fig. P3b). These elemental ratios suggest that Group 2 mafic rocks are compatible with a genesis from primary magmas originating from oceanic mantle source slightly enriched with respect to N-MORB sources. The Nb-Th co-variation (Fig. P3a) and the very low Th/Ta ratios (< 2, Table P1) displayed by all rock groups indicate that they were originated from sub-oceanic mantle sources, with no influence of subduction-related or continental crust chemical components.

The high and very high LREE/HREE ratios (La_N/Yb_N = 7.24-20.1) displayed by the alkaline basalts of Group 3 rocks (Fig. P1h, Table P1) suggest an involvement of a garnet-facies peridotite source and imply a mantle source much more enriched in LREEs than the depleted N-MORB mantle (DMM). The Nb/Yb-Ti/Yb co-variation highlights depth of melting, as the variance of Ti/Yb values is almost entirely depending on garnet residues during melting, whereas the Nb/Yb variation mainly depends on source composition (i.e., depletion vs. enrichment) and melting degree (Pearce, 2008). In addition, according to Pearce (2008) and Saccani (2015), the Th-Nb co-variation (Fig. P3a) may be applied to determine whether or not basaltic rocks are truly oceanic, and then the TiO_2/Yb –Nb/Yb diagram (Fig. P4) to determine the type of oceanic setting. In fact, on the TiO_2/Yb –Nb/Yb diagram (Fig. P4), Group 3 rocks have higher TiO_2/Yb ratios than MORB reflecting an alkaline OIB composition and deeper melting with involvement of a garnet-bearing mantle source. In addition, Group 3 rocks show the higher Nb/Yb ratios, which clearly point out for a “plume-type” enriched mantle source. Group 1 and Group 2 mafic rocks fall within the MORB array suggesting thus shallow melting in the spinel-facies mantle. Group 2 rocks have Nb/Yb ratios higher than Group 1 rocks and therefore we suggest that these rocks were generated from partial melting of an E-MORB type mantle source. Interestingly, the co-variation of TiO_2/Yb and Nb/Yb ratios of Group 2 rocks reflects plume–ridge interaction (Fig. P4). In fact, Figure P3b shows that Group 2 compositions conform extremely well to the mixing curve calculated using the OIB and N-MORB end-members. Such mixing relationships are consistent with either magma mixing or source region mixing (or eventually, a combination of both). Finally, the TiO_2/Yb and Nb/Yb ratios of Group 1 rocks suggest a genesis from a depleted N-MORB type mantle source without any influence of plume-type chemical components (Fig. P4).

It has been demonstrated that different oceanic settings are characterized by distinct mantle source compositions, as well as distinctive petrogenetic processes (i.e., partial melting degrees, type and depth of melting, etc.) (see Pearce, 1982, 2008). In order to better constrain the type of oceanic setting in which the different Bajgan rocks were generated, we performed non-modal, batch partial melting models (Fig. P5). In these models, we use plots of LREE/HREE (i.e. La/Yb) vs. MREE/HREE (i.e. Dy/Yb) ratios (Fig. P5a) and plots of Th vs. Nb/Yb ratio (Fig. P5b). These plots are particularly useful for distinguishing between melting in the spinel and garnet stability fields (e.g., Thirlwall et al., 1994), whereas the abundance of Th and Nb (Fig. P5b) is particularly useful for evaluating the enrichment of the source (Saccani, 2015). In this Figures, three compositionally different mantle sources are considered based on the diagram in Figure P4: (1) a DMM source (Workman and Hart, 2005) melting in the spinel-facies; (2) a theoretical DMM source enriched in LREE and incompatible elements by a plume-type chemical component (i.e., plume-influenced source of Saccani, 2015; or plume-proximal of Pearce, 2008) that melts in the spinel-facies. The composition of this source was assumed based on modelling, as presented by Saccani et al. (2013a, 2014, 2015); (3) an OIB-type source (Lustrino et al., 2002) that starts melting in the garnet-facies and continues to larger degrees into the spinel facies. In addition, it should be noted that Th and Nb are sensitive of fractional crystallization in shallow level magma chambers. Therefore, fractional crystallization trends for primary melts generated from each mantle source are shown in Fig. P5b. They are calculated assuming the crystallization of olivine, plagioclase, clinopyroxene, and spinel in the proportions shown in Figure.

Group 1 mafic protoliths with N-MORB chemistry are compatible with melt generation from a depleted MORB-type mantle at shallow levels. In particular, both the REE ratios and Th-Nb-Yb composition of these rocks can be explained by high degrees (generally, 15 - 18 %) of partial melting of a DMM source in the spinel facies (P5). Both diagrams in Figure P5 show that primitive alkaline basalts of Group 3 rocks are compatible with low degree partial melting (5 – 7%) of an OIB-type mantle source that starts to melt in the garnet-facies mantle (20-30% of the total melt) and it continues to melt to greater extent (70-80% of the total melt) in the spinel-facies (polybaric melting).

The modellings in Figures P5a and P5b shows that E-MORB protoliths (Group 2) is compatible with moderate degrees (~8 - 10 %) of partial melting in the spinel facies of a DMM source enriched in Th, Nb, and LREE by plume-type components. Alternatively, the Th-Nb-Yb composition of Group 2 E-MORBs may have been derived from very high degrees (>20%) of polybaric melting of an alkaline OIB-type mantle source (Fig. P5b). However, the REE composition of these rocks cannot be explained by this melting process (Fig. P5a) and such a very high degree of melting is unreasonable in an oceanic subduction-unrelated setting.

In summary, the overall chemical composition of the metamorphic rocks from the Bajgan Complex (Figs. P3, P4) and the melting models shown in Figs. P5a, b allow the following conclusions to be drawn: (1) the magmatic protoliths of these rocks were formed in an subduction-unrelated oceanic setting with no influence from either subduction or continental crust chemical components; (2) the chemically distinct groups of magmatic protoliths include N-MORBs, E-MORBs, and alkaline OIBs that are related to different mantle source compositions, which are associated, in turn, to distinct oceanic tectonic settings, such as mid-ocean ridge, mid-ocean ridge influenced by plume-type components, and within-plate seamounts, respectively.

4.3. Deformation history

In the Bajgan Complex, four deformation phases, from D1 to D4, have been identified in the field according to mesoscale overprinting relationships. The overall structural setting of this complex is represented by several slices bounded by shear zones developed during the late stage of the D2 phase. In the slices the relics of the D1 phase as well the structures of the early D2 phase can be recognized. The D1 and D2 phase, including the shear zones, are deformed by the subsequent D3 and D4 phases.

The oldest deformations recognized in the field belong the first deformation phase (D1) whose relic are mainly represented by the S1 continuous foliation. Scattered isoclinal F1 folds with acute hinges have also identified in the field. The S1 foliation and the F1 folds are everywhere deformed by the structures of the D2 phase.

The second deformation phase (D2) is characterized by the S2 foliation that is the structural elements recognized in the field. The S2 foliation occurs as continuous and pervasive foliation everywhere represented by a composite layering defined by overprinting of S2 foliation on the S1 one. Only in the F2 hinge-zone, the S2 foliation can be classified as crenulation cleavage shsowing the microlithons where the S1 foliation is still preserved. Tight to isoclinal F2 folds are well developed within metavolcanoclastics, fine-grained micaschists and fine-grained paragneisses. The F2 folds show symmetric profile with thickened hinge, thinned and delaminated limbs and sub-rounded up to sub-angular hinges profile (FIGURE PIEGHE). A2 fold axes mostly trends from NW-SE to N-S with a plunge ranging from 0 to 40° (FIGURE NETS). The main D2 structure documented in the field is the S2 foliation. It develops heterogeneously as axial plane and mylonitic foliation and shows a trend ranging from NW-SE to NNE-SSW with variable dip mainly toward northeast (FIGURE NETS). The S2 foliation bears L2 mineralogical lineations represented by elongated grains of quartz, calcite and amphibole showing variable orientation from NE-SW to N-S with variable plunge. Along the limbs of the F2 folds, the parallelism between S1 and S2 foliation leads to a composite Sp foliation whereas in the hinge zone, the S2 axial plane foliation wraps relics of S1. In marble, S2 foliation is highlighted by spaced (from 0.2 to 0.5 cm) pressure solution surfaces.

During the late stage of the D2 phase mylonitic shear zones oriented as the main foliation developed mainly at the boundaries of lithologies with contrasting rheology. The mylonitic display a continuous and pervasive foliation that bears NE-SW to N-S trending lineations. These shear zones thus divide the Bajgan Complex in several slices, each consisting of meta-ophiolites and the related meta-sedimentary rocks.

D3 phase is mainly represented by cylindrical F3 folds and by top-to-the-west thrusts. F3 folds show interlimb angle ranging from 30° to 120° and sub-angular to sub-rounded hinge zones (FIGURE). The $L_{(S0-S3)}$ intersection lineation is mainly represented by S3 foliation-S0 bedding intersection and mullion structures. A3 fold axis trend mainly from NW-SE to NW-SE with low plunge (FIGURE NET). S3 foliation is a spaced axial plane crenulation cleavage with subhorizontal attitude (FIGURE NET). No evidence of mineralogical lineation has been detected on the S3 foliation. F3 folds are associated with NW-SE-trending extensional shear zones dipping from 20° to 60° toward north and showing a top-to-the S ssense of shear.

The D4 phase produced parallel folds (type 3, Ramsay 1967) with interlimb angles ranging from 160° to 80° and rounded hinges. They show sub-vertical axial planes and A4 axis with NW-SE trend. The axial plane foliation, showing a NW-SE to N-S trend with subvertical dip, is represented by a disjunctive cleavage without sin-kinematic recrystallizations. Associated with F4 folds, extensional brittle shear zones reactivated the pre-existing D3 thrusts indicating a top-to-the N sense of shear.

Finally, high-angles dextral strike-slip or oblique faults probably belonging to the Sabzevaran Fault system dissect all the structures of the Bajgan Complex.

4.4. Microstructural features

In meta-serpentinites, rare relicts of cumulite texture are testified by cumulus clinopyroxene, olivine and red-brown spinel, and occasionally intercumulus clinopyroxene (generally replaced by fine-grained aggregate of actinolite + tremolite + chlorite + Fe-Ti oxides) preserved in low-strain domains. These domains aresurrounded by the S2 foliation consisting of fine-grained aggregate of chlorite + serpentine + opaque minerals. Generally, the In meta-serpentinites have a mylonitic to ultramylonitic texture with the S2 mylonitic foliation marked by chlorite + serpentine + opaque minerals wrapping rounded aggregates of serpentine and/or tremolite, and olivine crystals affected by microfaults.

In the meta-plagiogranites, the main foliation is marked by lenses of quartz ± albite ± garnet (maximum thickness c. 500 μm) and 550-800 μm thick layers of quartz + white mica + chlorite + epidote (± green hornblende ± chloritoid ± opaque minerals). Quartz shows grain size varying from 80 to 500μm, strong crystal preferred orientation, weak undulatory extinction and microstructures indicative of grain boundary migration recrystallization (i.e., dragging structures and lobate grain boundary). It results partly overprinted by grain boundary area reduction mechanisms that produced straight boundaries and 120° triple points. The quartz-rich layers wrap 0.7-1.3 mm sized feldspar crystals showing a well-developed shape preferred orientation with the long axis parallel to the main foliation and inclusions of quartz and white micas oriented sub-parallel or at high-angle respect to the external foliation. They have weak undulatory extinction, lobate boundary (where in contact with quartz crystals), rare deformation twins and a variable grade of recrystallization. Large (400-700 μm) white mica crystals are folded and show evidence of undulatory extinction

In meta-volcanic rocks, the S2 foliation) is highlighted by a mm-thick layers of green hornblende/actinolite + chlorite ± opaque minerals, and fine-grained quartz + albite + epidote ± calcite locally wrapping coarse-grained (2-3 mm) plagioclase porphyroclasts.

In meta-volcanoclastites and in meta-sedimentary rocks, mainly in the micaschists, relicts of S1 foliation are represented by aggregates of chlorite and phengite crystals oriented orthogonal to the external S2 foliation (see Fig. XX and section XXX for a detailed description of their mineral chemistry) and by the trails of white mica included in the core of the feldspar crystals. In the micaschists and paragneisses, the S2 foliation is marked by a compositional layering consisting of maximum 3 mm thick layers of quartz (± white mica) and thinner layers (0.6 -1 mm) of quartz + white mica + chlorite±epidote or by irregular lenses of quartz±garnet±white mica±epidote and discontinuous layer of white mica + chlorite ± epidote. These layers wrap feldspar crystals (400-700 μm) showing lobate grain boundaries with quartz crystals, weak undulatory extinction and a strong shape preferred orientation with long axis parallel to the main foliation. Locally, they have thin deformation twins and a thin and irregular ribbon of feldspar neoblasts. The core of the biggest feldspar crystals shows small inclusions of quartz and white mica oriented from high angle to sub-parallel to the external foliation. White mica crystals have a maximum size of 200 μm. The bigger white mica crystals show micro-folding and undulatory extinction. Their big size and the microstructures suggest medium deformation temperature. Garnets are euhedral, 150-200 μm in size with homogeneous composition and no evidence of replacing. These features suggest that they grew simultaneously to the development of the S2 foliation. In the granoblastic layers, quartz grains (80-350 μm grain size) shows few evidences of intracrystalline deformation (weak undulatory extinction), moderate to weak shape preferred orientation and locally, dragging structures (Passchier and Trouw, 2005), lobate boundaries and seriate grain size. These features may suggest that recovery involved both subgrain rotation and grain boundary migration recrystallization. In the quartz + white mica (+ chlorite) layers, phyllosilicate crystals have prevented quartz grain growth producing a strong preferred alignment of quartz grain boundaries parallel to foliation and a consequent moderate to weak shape preferred orientation of quartz crystals. Grain boundary are also at high angle to the main foliation indicating high mobility grain boundaries. Quartz often has polygonal grains and straight grain boundaries that meet in 120° triple points indicating that grain boundary area-reduction mechanisms locally occur. The mineral association along the S2 of quartz + garnet + epidote may indicate medium to high grade temperature. In impure marble, the S2 axial plane foliation is highlighted by mm- to cm- thick layers of calcite ± green amphibole ± chlorite and monomineralic layers of quartz showing lobate grain boundaries, seriate grain size, well-defined crystal preferred orientation, and weak undulatory extinction. Calcite crystals show thick and irregular deformation twins (i.e., Type IV of Ferrill et al., 2004). Evidences of the old S1 foliation is highlighted by chlorite crystals oriented at high angle respect to the compositional banding.

D2 shear zones where documented both in metasediments and metagabbros (sample 707 and 325, respectively). In the metasediments, D2 shear zones are highlighted by less than 1 cm-thick S-C-C' structures and σ-type porphyroclasts of feldspar crystals with asymmetric tails of chlorite + white mica + fine grained quartz ± epidote. C and C' planes are highlighted by white mica + fine grained quartz. In the metagabbros, the S2 mylonitic foliation is marked by amphibole + epidote + quartz ± albite ± white mica. The syn-kynematic amphibole crystals are small (30-50 μm), euhedral and characterized by a strong preferred orientation with long axis parallel to the mylonitic foliation. The S2 mylonitic foliation wrap large (from 0.2 to 4 mm) amphibole porphyroclasts showing weak shape preferred orientation, asymmetric tails of small syn-kynematic amphibole + epidote, synthetic and antithetic microfaults (i.e. bookshelf structures) and locally fishoid shape (i.e. amphibole fish). All these features are coherent with the same kinematics. Quartz show lobate grain boundary suggesting that dynamic recrystallization occurred by grain boundary migration mechanism.

D3 phase??).

4.5. Petrological thermobarometers

A detailed microscale study of the deformation history was carried out on different lithotypes of the Bajgan Unit. Basing on the metamorphic recrystallization observed in the different rock types, the metapelites were chosen to estimate the P-T conditions of the D1 and D2 phases. Selected samples of the metapelites have been collected in different slices and three samples (MK701, MK706 and MK707) have been chosen as representative. The description of the mineral chemistry of these three samples is provided in the Supplementary material (SIGLA SUPPLEMENTARY MATERIAL). Temperature and pressure conditions were estimated using the Chl-Qz-wt (Vidal et al., 2006) and the Ph-Qz-wt (Dubacq et al., 2010) methods, respectively. The results of these two methods allowed to identify different areas in the P-T space representing the stability fields of the D1 and the D2 phases. The Chl and Ph analysis employed for these calculations were also used to calculate the equilibria of single Chl-Ph couples through the Chl-Ph-Qz-wt method (Vidal and Parra, 2000). The sample in which the Chl and the Ph phases were not present, the pressure conditions were instead calculated using the Al-in hornblende barometer (Schmidt, 1992).

Chl-Qz-wt method (Vidal et al., 2006): this method was applied on samples MK707 (Fig.Ea) and MK707 (Fig.F). In the sample MK707 the calculations were performed fixing pressures at 0.60 GPa and water activity at 0.8. The single generation of Chl (Chl D2) identified (Fig.Ca and Tab.B) turns out to be stable in a T range of 280-450°C (Fig.Ea). In the sample MK706 three selected groups of Chl were distinguished: the first group (Chl D1) were sampled within the S1 foliation and shows a composition enriched in Dph end-member. The second (Chl D2-peak) and the third (Chl late-D2) groups are related to the S2 foliation and show a predominance of Ame and Sud end-members, respectively (Fig.Aa). The calculations were performed fixing pressures (0.60 GPa, 1.20 GPa and 0.40 GPa for the first, second and third groups respectively) and water activity (0.8) parameters. The T ranges of stability calculated for the Chl D1 are 150-320°C and 375-400°C (Fig.Fa), that for the Chl D2-peak are 250-330°C and 370-450°C (Fig.Fb)and that for the Chl late-D2 are 150-350°C and 375-400°C (Fig.Fc).

Ph-Qz-wt method (Dubacq et al., 2010)***:*** this method was used to calculate the P ranges through the Ph compositions. In the sample MK707 one generation of Ph developed parallel to the S2 foliation (Ph D2) were analyzed (Fig.Cb, Tab.B). The calculations were performed within the T range obtained with the Chl-Qz-wt method (i.e. 280-450°C), and the % Fe^{3+} to 30, which results to be the best optimized value. All the calculations with the Ph-Qz-wt method have been performed fixing the water activity to 0.8. The P estimate for the D2 phase of sample MK707 ranges between 0.40 and 1.00 GPa. The Ph observed in the sample MK706 have been divided into three groups: Ph grown along the S1 foliation (Ph D1) that shows a Prl affinity and the Ph grown along the S2 foliation, distinguishing those enriched in the Cel end-member (Ph D2-peak) from those wth Ms predominance (Ph late-D2, Fig.Ab). To apply the Ph-Qz-wt method, different T ranges of stability has been assigned to the three groups of Ph on the base of the Chl-Qz-wt method's results (i.e., the T range of Chl D1 for the Ph D1, those of Chl D2-peak for the Ph D2-peak and those of Chl late-D2 for the Ph late-D2), as well as the % Fe^{3+} content. All the calculations with the Ph-Qz-wt method have been performed fixing the water activity to 0.8. The results show that the P range associated to the Ph D1 is 0.33-0.96 GPa, that of the Ph D2-peak is 0.60-1.35 GPa and that of the Ph late-D2 is 0.22-0.78 GPa (Fig.F).

Al-in-hornblende method (Schmidt, 1992)***:*** the estimate of the P conditions of the sample MK701 was performed through the calibration based on the Al_{tot} content in calcic amphibole (Fig.Db). The analyzed Ca-Amp belong to the metamorphic paragenesis which constitute the S2 foliation. Therefore the P range calculated for the sample MK701 of 0.80-1.10 GPa regards the P condition registered during the D2 phase.

Chl-Ph-Qz-wt method (Vidal & Parra, 2000)***:*** one Chl-Ph pairs have been identified within the same microstructure, this method allows to find points in the P/T space which represent their optimized equilibrium conditions in presence of Qz and wt. Only the Chl-Ph couples the equilibrium of which is reached in the T and P ranges calculated with the Chl-Qz-wt and Ph-Qz-wt were considered in this work. Sample MK707 is characterized by a single population of Chl and Ph (Chl-Ph D2) that reached the equilibrium condition at 0.60-1.00 GPa and 280-320°C. (Fig.Ec). In the sample MK706 the equilibrium conditions related to the first generation of Chl and Ph (Chl-Ph D1) are 0.52-0.94 GPa and 250-295°C, those related to the second generation (Chl-Ph D2-peak) are 1.05-1.30 GPa and 260-310°C and those related to the third generation (Chl-Ph late-D2) are 0.28-0.57 GPa, 315-350°C (Fig.F). The P-T conditions calculated with the Chl-Qz-wt, Ph-Qz-wt and Chl-Ph-Qz-wt methods of the sample MK706 were compared with those estimated with classical thermobarometry (Tab.D).

4.6. Microstructure-based thermometers

Quartz microstructures: Quartz microstructures documented on metasediments and on meta-intrusive rocks (i.e., weak undulatory extinction, lobate boundary, dragging structures) suggest deformation temperature of at least (450-)500°C. Analogous deformation temperature can be deduced by the microstructure documented on feldspar crystals (i.e., undulatory extinction, lobate boundary, rare deformation twins, recrystallization and the lack of microfaulting). Microstructures on the larger porphyroclasts of white mica suggest medium grade deformation temperature.

Calcite microstructures:

4.7.U-Pb geochronology of the meta-intrusive rocks

Several samples of plagiogranites have been collected in order to achieve reliable ages by U-Pb geochronology. These samples have been collected in different slices of the Bajgan Complex with the aim to obtain a complete picture of the ages for the meta-intrusive rocks. Three sample, namely MK701,704 and 705, have been analyzed.

Large part of the U-Pb data from MK704 and MK701 samples resulted severely discordant.

The zircon grains from MK701 sample are generally euhedral and small (<100μm) with low aspect ratios. They are characterised by large homogeneous core surrounded by a brighter thin rim and rarely a faint oscillatory zoning is recognisable. Inclusions of apatite are common (CL features of: zrc55, 67, 50, 52). Nineteen U-Pb analyses on seventeen zircon grains were collected for the MK701 sample. Only 7 data provided a concordance <8% (Fig.) and a weighted average $^{206}Pb/^{238}U$ age of 156±6Ma.

The zircon grains from MK704 sample are generally small (<75μm) and stubby. They are characterised by darker core surrounded by brighter rims (Zrc 87, 88, 92, 104). Thirteen U-Pb data were collected for U-Pb analyses on fifteen zircon grains were performed for sample MK704. Nine U-Pb ratios with a concordance <9% provided a weighted average $^{206}Pb/^{238}U$ age of 136±3Ma.

The zircon grains from MK705 sample are euhedral, small (<100μm) and with low aspect ratios. Oscillatory zoning Zrc78, homogeneous large core (72, 76) 83. They may contain inclusions of apatite. Twelve U-Pb analyses on eleven zircon grains were collected for sample MK705. U-Pb data showed a concordance better than that observed within other samples, generally <2%, with a weighted average $^{206}Pb/^{238}U$ age of eight data at 112±4Ma.

On the whole, the obtained ages range from 156±6Ma to 112±4Ma to, i.e. from Late Jurassic up uppermost Early Cretaceous.

4.8. U-Pb geochronology of the felsic dyke FF

In order to constrain the ages of the main deformation phases, a sample of a not metamorphosed felsic dyke has been collected. The sampled dyke belongs to a swarm that cuts all the deformations belonging to D1, D2 and D3 phases, including the shear zones that subdivide the Bajgan Complex into several slices.

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4.9. Ar-Ar- geochronology of the micaschists FC

Age uncertainties are reported to 2σ; uncertainties on the isotopic measurements are reported to 1σ.

5. DISCUSSION

5.1. The Bajgan Complex re-interpreted as remnant of an Early Cretaceous oceanic lithosphere

Results from the present study proving that the so far proposed interpretations for the Bajgan Complex as the pre-Mesozoic continental basement of a microcontinent (the so-called Bajgan-Durkan microcontinent (McCall, 1985; Hunziker et al., 2015; Burg, 2018; Saccani et al., 2018) must be discarded. Several lines of evidence clearly suggest that the Bajgan Complex consists of the remnants of an Early Cretaceous oceanic lithosphere.

The first line of evidence is provided by the geochemistry of the meta-igneous rocks. As well assessed in literature (…..), the geochemistry of the igneous rocks can be used to constraint the geodynamic setting of their origin. The collected data about the meta-igneous rocks from Bajgan Complex indicate that they belong to three different types: group 1 made up of meta-basalts, meta-gabbros and meta-plagiogranites with N-MOR affinity; group 2 made up of meta-basalts, meta-gabbros and meta-plagiogranites with E-MOR affinity and group 3 made up only of meta-basalts with OIB affinity. These groups can be associated to distinct oceanic settings, such as mid-ocean ridge, mid-ocean ridge influenced by plume-type components, and within-plate seamounts, respectively.

A further line of evidence is represented by the age of the meta-intrusive rocks that has been determined by U/PB method on zircons. The collected data indicate an age ranging from 161 to 114 Ma, i.e. from Oxfordian to Aptian. These data exclude a Paleozoic origin for the meta-igneous rocks of the Bajgan Complex. The different ages detected in the meta-ophiolites are related to different slices, each bounded by the mylonitic shear zones developed during the D2 phase. These slices can be thus regarded as derived from different oceanic settings with different ages of the same basin.

To summarize, the geochemical characteristics of meta-ophiolites from Bajgan Complex imply the occurrence of remnants of a well-developed and thick oceanic crust that developed for a long time (more than 40 Ma) and produced MOR-type intrusive and effusive rocks. The presence of OIB and E-MOR meta-basalts point to a plume influenced magmatism in the Neo-Tethys oceanic Basin (cfr. plume proximal oceanic lithosphere of Pearce 19…). Evidence of this Cretaceous oceanic within-plate basaltic magmatism in the Neo-Tethys oceanic basin have been indicated by several authors from Armenia to Himalaya (……….) and recently proposed for the Cretaceous Durkan and Band-e-Zeyarat Complexes by Barbero et al. (2020a, b).

The third line of evidence is provided by the lithostratigraphy. If the geochemistry suggests an interpretation of the meta-igneous rocks as fragments of oceanic lithosphere, the Bajgan Complex can be regarded as an assemblage of slices of both meta-ophiolites and meta-sedimentary rocks. The meta-ophiolites firstly include the meta-serpentinites, that can be interpreted as representative of a mantle section and/or derived from a lower part of the intrusive sequence. In the latter interpretation, the meta-serpentinites can be regarded as cumulates forming the lower part of an intrusive sequence dominated by peridotites, dunites and melagabbros. The lower part of the intrusive magmatic sequence is instead represented by the meta-intrusive rocks derived from metamorphism and deformation of a primary sequence of gabbro, gabbronorite, melagabbro and anortosite but also including more felsic and evolved intrusive rocks as plagiogranites. The upper part of oceanic crust is instead represented by the meta-volcanic rocks, probably derived by massive and/or pillow-lava basalts. Associated to meta-volcanic rocks, the widespread meta-volcanoclastics are representing the transition to the meta-sedimentary rocks. The metasedimentary rocks derive from a succession consisting of quartzites, impure marbles, calcschists, fine - up medium-grained micaschists and fine- up coarse-grained paragneisses. These metamorphic rocks likely derived from a pristine succession including cherts, limestone and siliciclastic to mixed turbidites. All these lithologies are generally found in the sedimentary deposits at the top of the oceanic crust. The age of this assemblage cannot be detected by fossils because they have been totally recrystallized during the metamorphism, but an Early Cretaceous age can be proposed according to the Late Jurassic-Early Cretaceous age of meta-igneous rocks and the Early to Late Cretaceous age of the metamorphism.

On the whole, all the lithologies detected in the Bajgan Complex are coherent with its origin from deformation and metamorphism of oceanic lithosphere. No metamorphic rocks clearly derived from continental crust, as meta-granites, meta-rhyolites, meta-conglomerates or meta-dolostones, have been detected during the field survey.

5.3. The tectono-metamorphic history of the Bajgan Complex as representative of a subduction-exhumation path in a subduction setting

The interpretation of the Bajgan Complex as consisting of remnants of an Early Cretaceous oceanic lithosphere is confirmed by the tectono-metamorphic evolution reconstructed in the meta-sedimentary rocks.

As previously described, the Bajgan Complex is represented by an assemblage of slices, each bounded by mylonitic shear zones developed during the D2 phase. In each slice, the same polyphase deformation history has been reconstructed. The scattered relics of the D1 phase are represented by the rare isoclinal F1 folds with acute hinge as well as the remnants of S1 foliation. The D2 phase is characterized by a pervasive and widespread deformation that include tight to isoclinal F2 folds, mylonitic shear zones and a continuous S2 foliation bearing well-developed mineral lineations. The main structures are represented by meso-to-map scale open to tight parallel F2 folds with sub-horizontal axial plane and sub-horizontal axes showing a N-S trend. The subsequent D3 phase consists of a composite fabric acquired during semi-brittle to brittle deformation regimes.

The significance of these phases can be hypothesized only taking into account the P-T conditions estimated using the metamorphic mineral assemblage grew along the S1 and S2 foliations. Only in one sample the relics of the D1 phase have been identified. P and T estimates indicate that D1 phase were acquired during blueschist facies conditions (T = 250-295°C, P = 0.52-0.94 GPa) at depth of ~17-31 km. Even if the available data do not allow a detailed reconstruction of the metamorphic history for the whole Bajgan Complex, the identification of the high-pressure metamorphism represents a useful constraint for the interpretation of the D1 phase.

More data are instead available for the D2 phase. The data acquired by the thermodynamic calculations conducted by Chl-Qz-wt, Ph-Qz-wt and Chl-Ph-Qz-wt methods indicate the occurrence of two groups of mineral assemblage grown along the S2 main foliation.

The first group is grown during the early stage of D2 phase, whose P and T conditions can be framed between 0.30-1.35 GPa and 250-450°C. These metamorphic conditions can be better constrained by Al-in-hornblende method that indicate for the early stage of the D2 phase a P of 0.80-1.10 GPa. In addition, the quartz microstructures suggest a T during the deformation in a range of 400 – 600°C. The P and T conditions during the early D2 phase can be thus regarded as developed at 0.80-1.10 GPa and about 400-450°C.

The thermodynamic calculations conducted by Chl-Qz-wt, Ph-Qz-wt and Chl-Ph-Qz-wt methods indicate that the second group of mineral assemblage has grown during the late stage of D2 phase within a wide range of P and T conditions, i.e. 0.22-0.78 GPa and 150-400°C. The data acquired by the calcite twinning indicate a T higher than 250-300°C, thus constraining the P and T conditions during the late stage of the D2 phase within the greenschist metamorphic facies. This result is also suggested by combining the result of the Chl-Ph-Qz-wt multi-equilibrium approach (Vidal & Parra, 2000) and Chl-Qz-wt method (Vidal et al., 2006) that indicate P and T conditions of 0.28-0.57 GPa and 315-350°C.

On the whole, these data indicate that the D1 phase was developed at T = 250-295°C and P = 0.52-0.94 GPa, corresponding to blueschists facies metamorphic conditions. During the D2 phase two different mineral assemblages were registered. The early stage of the D2 phase occurs at T= about 400-450°C and P=0.80-1.10 GPa at conditions corresponding to the boundary between amphibolite and eclogite metamorphic facies. The late stage of the D2 phase instead occurs at P and T, corresponding to greenschist facies metamorphic conditions.

The estimated P-T conditions are in agreement with the finding of glaucophane reported by Mc Call (1985). However, McCall (1985) has also defined within the meta-volcanic rocks and micaschists of the Bajgan Complex three zones of progressive metamorphism from greenschist to amphibolite facies with P ranging from moderate up high. The boundaries of these zones cut at high angle the main foliation as well as the main shear zones suggesting their origin by a later metamorphic imprint. So, the significance of the metamorphic zones reported by McCall (1985) remain to be clarified. Our data can be also compared with those provided by Dorani et al

(2019). These data are collected in a single area and indicate for the first tectono-metamorphic phase in garnet-bearing schists a pressure of more than 9 kbars and temperatures between 560 and 675 °C, at conditions corresponding to the boundary between amphibolite and eclogite metamorphic facies. This phase is followed by a later one developed under greenschist facies metamorphism with a decrease in temperature and pressure (370–450◦C and 3–6 kbars). A comparison with our microstructural data suggests the first tectono-metamorphic phase from Dorani et al (2019) corresponds to the early stage of the D2 phase, whereas the second one can be compared with the late stage of the D2 phase. This interpretation is also suggested by the comparison (Fig…) with the phengites grown along the S2 foliation and those belonging to the metamorphic assemblage analyzed by Dorani et al. (2017). In this assemblage, the phengites show a composition corresponding to that of the phengites developed during the thermic peak identified in this paper.

A further important constraint is also provided by the age of the D2 phase that has been determined using Ar-Ar dating on withe micas grown along the D2 foliation and indicate that this event occurred during the late Albian-Turonian (101-90 Ma). The Ar-Ar white mica closure temperature of ca. 350°C (citare) support the interpretation of this radiometric age during the D2 phase.

The sequence of deformation phases, from D1 to D3, coupled with the P-T conditions of the metamorphism that shows a clockwise trajectory from blueschists to greenschists facies, is coherent with a transfer of fragments of oceanic lithosphere at the base of an accretionary wedge and their subsequent exhumation up to shallow structural levels (….). The D1 phase can be thus regarded as the result of a coherent underplating at the base of an accretionary wedge within a subduction zone. More puzzling is the interpretation of the early stage of the D2 phase that developed at the P and T conditions corresponding to the boundary of the amphibolite and eclogite metamorphic facies. Compared to P and T conditions of the D1 phase, the metamorphism developed during the early stage of the D2 phase indicate a relevant increase of T, whereas the P conditions seems to be overlapping with those estimated for the D1 phase. Even if more data are required for the interpretation of early stage of the D2 phase, its development occurs after the underplating at the same depth or slightly deeper than D1 phase but during an increase of T. The early stage of the D2 phase can be thus regarded as a post-underplating deformation probably acquired within the subduction channel, i.e. the wedge-shaped zone at the interface between the down-going plate and the accretionary wedge. The subduction channel is the site of the exhumation of high- and ultrahigh-pressure metamorphic slices up to shallow crustal levels, and eventually to the surface, from depths of more than 100 km. However, the subduction channel is characterized by complex dynamics that produced a flow of the underplated material not only upwards but also downwards, and always in presence of ductile deformations. The early stage of the D2 phase can be thus regarded as acquired within the subduction channel after the underplating and before the inception of the exhumation. Consequently, the deformations related to the late stage of the D2 phase can be regarded as acquired when the process of the exhumation was in an advanced stage, as indicate by the strong decrease of P conditions, coherent with a depth of 9-19 Km. The exhumation continued also during the D3 phase, when semi-brittle to brittle deformations developed probably at very shallow levels.

The data provided by Dorani et al (2019) indicate a T during the main deformation of the Bajgan Complex higher than that established in this paper for the early stage of the D2 phase. Probably this difference in the T value is detected in two different slices that experienced deformation at different T conditions. In a subduction setting, the thermal regime can change in response to geodynamic events like a subduction of a volcanic seamount or a decrease in the convergence rate. These events are able to explain the higher T detected by Dorani et al (2019).

This tectono-metamorphic history of the Bajgan Complex is analogous to that reconstructed for the oceanic lithosphere in many world-wide fossil accretionary wedge (……..) where the deformations acquired during the accretion are followed by those developed during the exhumation that resulted into transfer to the meta-ophiolites and related meta-sedimentary deposits from the deep in the accretionary wedge up to shallow structural levels.

To sum up, the Bajgan Complex experienced buried and underplating reaching the blueschist facies metamorphism conditions and then exhumed from deep to shallow crustal levels.

This evolution is constrained by the emplacement of the Middle Eocene (41,95 +/- 0,07 Ma) calk-alkaline dykes that crosscut the D1-D3 structures. The calk-alkaline dykes are slightly deformed by the late tectonic event as indicated by deformation of the phenocrystals. The collected data indicate that the Bajgan Complex was exhumed from ~30 km to shallow level in the time spam from 90-101 Ma to 42 Ma.

5.3. Implications for the geodynamic history of the Makran

Most of the reconstructions so far proposed for the geodynamic history for the Makran area are based on the occurrence since the Early Cretaceous of two different oceanic basins, namely the North Makran basin located close to the Eurasian continental margin and the wide Neo-Tethys basin, whose remnant is still subducting below the Eurasian continental plate. These two basins are interpreted as developed in different geodynamic settings with different times and affected by different geodynamic evolutions. The North Makran ocean is regarded as opened as a suprasubduction basin in the Late Jurassic-Early Cretaceous and completely closed before the Eocene, according to the occurrence of Early Eocene sedimentary deposits unconformably lying at the top of the tectonic units derived from the North Makran basin. The location of these tectonic units north of the present-day subduction zone suggest a pristine location of the North Makran basin between the Eurasian margin and a microcontinent, the so-called Bajgan-Durkan microcontinent. The proofs of the existence of this microcontinent are searched in two different units of North Makran, i.e. the Durkan and the Bajgan Complexes. The first one is regarded as consisting of a highly deformed assemblage mainly consisting of shallow water deposits associated to abundant volcanic rocks (McCall, 1985; Hunziker, et al., 2015). This complex is commonly interpreted as the sedimentary cover of the Bajgan complex that is regarded as a continental basement older than Paleozoic (e.g., McCall and Kidd, 1982; Samimi Namin, 1983; McCall, 1985; Hunziker, 2014; Hunziker, et al., 2015; Burg, 2018). This reconstruction must be modified according the evidence provided by Barbero et al. (2020) and by this paper. The data provided by Barbero et al. (2020) indicate that the Durkan Complex represents fragments of seamounts tectonically incorporated in the Makran accretionary wedge during the latest Late Cretaceous - Paleocene. In addition, we have provided the evidence that the Bajgan Complex cannot be regarded as a continental basement of Paleozoic age but instead it represents a fossil accretionary wedge of Late Cretaceous age where Late Jurassic – Early Cretaceous ophiolites are involved.

Thus, the mechanism of closure for the North Makran oceanic basin by a collision of a microcontinent with the Eurasian margin seems to be unlikely and different geodynamic mechanism must be proposed. A valuable suggestion is provided by the occurrence of the remnants of volcanic seamounts in the North Makran. These remnants are preserved within the Durkan Complex by Barbero et al. (2020), but also in the Band-e-Zeyarat ophiolites where the remnants of an oceanic crust derived from Early Cretaceous chemically composite oceanic crust formed in a mid-ocean ridge setting by partial melting of a depleted suboceanic mantle variably metasomatized by plume-type components has been testified by Barbero et al (2020).

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6. CONCLUSION

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