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Continental margin ophiolites of Neotethys: Remnants of Ancient Ocean–Continent Transition Zone (OCTZ) lithosphere and their geochemistry, mantle sources and melt evolution patterns

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We present an overview of the geology, geochemistry and petrogenesis of continental margin ophiolites (CMO), which represent the lithospheric remnants of rift-generated paleo ocean – continent transition zones (OCTZ) in orogenic belts. The igneous stratigraphy and geochemical signatures of Neotethyan CMOs reflect the extent of geochemical heterogeneity, partial melting degrees, and melt evolution patterns in the continental lithospheric mantle prior to the onset of seafloor spreading in rifted margins. Basaltic rocks of the Jurassic CMOs in the External Ligurian units of the Northern Apennines have N-MORB and G-MORB affinities with strong HREE/MREE depletion, and represent the products of partial melting of a heterogeneous sub-continental lithospheric mantle containing small volumes of garnet pyroxenite layers. These extrusive rocks were erupted directly on the exhumed fertile spinel lherzolites of Adria during its OCTZ evolution. Volcanic rocks of the Triassic CMOs in the Albanide-Hellenide orogenic belt are represented by calc-alkaline suites; alkaline basalts and subordinate trachybasalts, trachyandesites, and trachytes; transitional to sub-alkaline plume-type P-MORB basalts; sub-alkaline enriched, E-MORB basalts; and, sub-alkaline N-MORB basalts. Upper mantle peridotites are not exposed. Magmas of these extrusive rock associations were derived from compositionally distinct mantle sources, which were affected by previous subduction and plume events in the geological history of the region. The CMOs in the Zagros orogenic belt include metamorphosed lherzolites with gabbro and mafic dike intrusions, which show N-MORB and G-MORB affinities. Basalts and basaltic andesites making up the majority of the Zagros volcanic sequences have E-MORB and P-MORB affinities, whereas minor alkaline rocks that are composed of basalts, trachybasalts and trachytes display OIB signatures. The mantle sources of the Zagros CMOs were progressively enriched in Th and Nb. The OIB component of the mantle beneath the Zagros OCTZ was derived from previous plume events during the early Carboniferous, when Paleotethys was undergoing its rift-drift tectonics. The observed differences in the igneous stratigraphy and geochemical affinities of these Neotethyan CMOs are a result of extreme mantle heterogeneity caused by previous subduction and plume events during the Wilson Cycle evolution of the older Paleotethys.

Introduction

Ophiolites represent fragments of ancient oceanic lithosphere (Dewey and Bird, 1971; Coleman, 1971) that were incorporated into continental margins during a variety of plate interactions (see Dilek and Furnes, 2011). They are commonly found along suture zones, which mark major boundaries between amalgamated plates or accreted terranes (Dilek, 2003a; Lister and Forster, 2009). Ophiolites record magmatic activities and tectonic processes associated with the construction and consumption of ancient oceanic basins from their rift–drift and sea floor spreading stages to subduction initiation and final closure phases. Magmatism during each of these phases produces ultramafic to mafic and evolved rock assemblages, which have distinct internal structures, geochemical afûnities, and age ranges, depending on their original tectonic setting of formation. The geological record of the Wilson cycle evolution of ocean basins is, therefore, partially preserved in most ophiolite complexes along suture zones.

The Penrose definition of an ophiolite is a "*distinctive assemblage of ultramafic and mafic rocks*" that includes, from bottom to top, tectonized peridotites, cumulate peridotites, and pyroxenites overlain by layered gabbros, sheeted basaltic dikes, a volcanic sequence, and a sedimentary cover ('Penrose-type igneous sequence'; Anonymous, 1972, p. 24). An ophiolite may be incomplete or tectonically dismembered, and it can also be metamorphosed. Dilek and Furnes (2011, 2014) have recently argued that this original Penrose definition of ophiolites (Anonymous, 1972) is inadequate to explain the significant variations in the internal structures, compositions, geochemical signatures, and emplacement mechanisms of ophiolites. They have introduced a more detailed classification based on the characteristic internal structures, geochemical fingerprints, and regional tectonics of ophiolites. Among the most important parameters controlling the magmatic evolution of different ophiolite types are: (1) the rate, geometry and nature of seafloor spreading, (2) the proximity of a spreading center to a plume or trench, (3) mantle composition, temperature and fertility, and (4) the availability of fluids (Dilek and Furnes, 2014).

In their new classification, Dilek and Furnes (2011, 2014) have defined the Continental Margin (CM) ophiolites as mafic-ultramafic and sedimentary rock assemblages that formed during the continental breakup and embryonic oceanic crust generation at ocean-continent transition zones (OCTZ). These ophiolites do not have the ideal Penrose-type igneous sequence and generally include fertile peridotites of continental lithospheric mantle that are directly overlain by basaltic lavas and hemi-pelagic sedimentary rocks. Basaltic rocks display a normal (N-) MORB geochemistry characterized by marked garnet signatures commonly typified by depletion in heavy rare earth elements (HREE) with respect to middle rare earth elements (MREE). Saccani (2015) used the term G-MORB (garnet-influenced MORB) for identifying this basaltic sub-type, which can easily be distinguished from typical N-MORBs using Ce-Dy-Yb systematics (see Fig. 8 in Saccani, 2015). The petrogenesis of CM ophiolites involves small degrees of partial melting of little depleted lithospheric mantle and slowly uprising asthenosphere in magma-poor, "cold rift" tectonic settings (Dilek et al., 1998). Thus, CM ophiolites constitute an important end-member of the entire ophiolite spectrum (Dilek, 2003a).

Some of the best examples of CM ophiolites include the Jurassic ophiolites in the Ligurian units of the Northern Apennines (e.g., Montanini et al., 2008; Marroni and Pandolfi, 2007), Alpine Corsica (Durand-Delga, 1984; Malavieille et al., 1998; Saccani et al., 2008a), and the Western Alps (Lagabrielle and Cannat, 1990; Desmurs et al., 2002; Rampone et al., 2005; Manatschal and Müntener, 2009; Festa

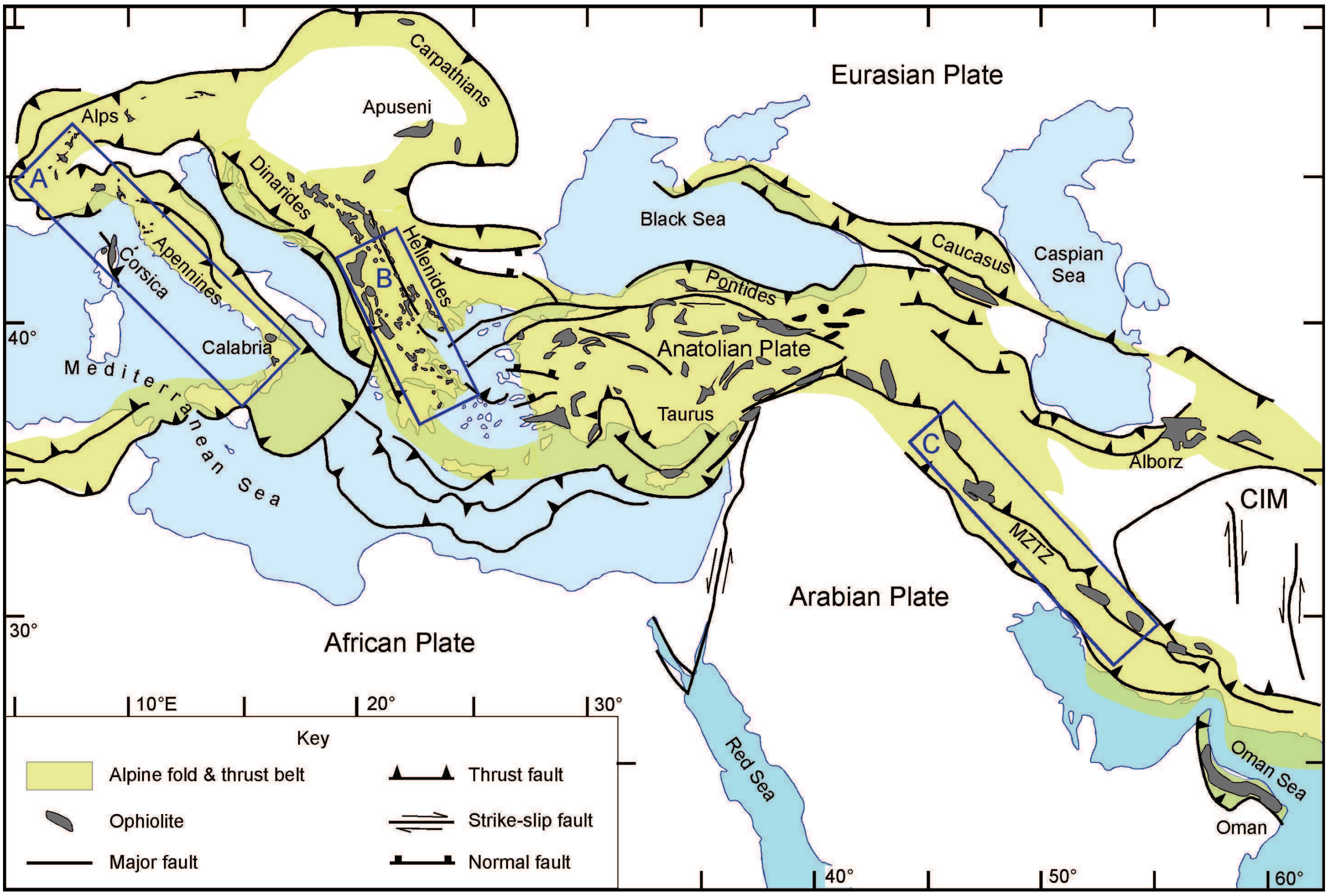

Figure 1. Distribution of major ophiolitic complexes in the Alpine orogenic belts from the central Mediterranean area to the Oman Sea. The major tectonic lineaments are also shown. Boxes indicate the investigated areas. A: Western Tethys ophiolites (Alps-Corsica-Apennine-Calabria); B: Albanide-Hellenide ophiolites; C: Zagros ophiolites. MZTZ: Main Zagros Thrust Zone, CIM: Central Iranian Microcontinent.

et al., 2015; Balestro et al. 2015;), and the early Cretaceous Dongba-Purang ophiolites in southern Tibet (Liu et al., 2015; Yang and Dilek, 2015). However, many more CM ophiolites are yet to be documented in the Phanerozoic and Precambrian orogenic belts through systematic structural field and geochemical studies. Most CM ophiolites are commonly multiply deformed and metamorphosed, undergone subduction-exhumation processes, and/or incorporated into sub-ophiolitic mélanges as blocks or thrust sheets (Dilek, 2003a; Dilek and Robinson, 2003). Thus, recognition of CM ophiolites and associated tectonostratigraphic units in continental collisional zones may be particularly difficult at best.

In this paper we examine the geological and geochemical characteristics of continental margin ophiolites of Neotethyan origin (Fig. 1), and discuss their melt evolution in OCTZs. We first review the geological and magmatic features of rifted continental margins. We then describe the geology and geochemistry of three continental margin ophiolites in different domains of the Neotethyan realm, in the Western Alps–Alpine Corsica–Northern Apennines (i.e., the Western or Alpine Tethys), in the Albanide-Hellenide belt (Pindos-Mirdita ocean basin), and in the Zagros orogenic belt in Iran. We next compare and contrast various petrogenetic models for the formation of these ophiolites, focusing on their mantle sources and melt evolution. We think that this overview of continental margin ophiolites will help us differentiate similar mafic-ultramafic rock associations of ancient OCTZs that have been otherwise unrecognized in collisional orogenic belts.

Rifted Continental Margins

Rifted continental margins and OCTZs have been divided into "non-volcanic" (Iberia-type) and "volcanic" (East Greenland-type) types, based largely on the study of the Central and North Atlantic regions by the deep Ocean Drilling Program (e.g., Whitmarsh et al. 2001). Non-volcanic rifted margins are invariably magma-poor and are considered to have undergone extension that varied vertically and laterally both in space and time. Simple shear (Wernicke, 1981) and detachment (Lister et al., 1991) models have been proposed to interpret this depth-dependent extension (Royden and Keen, 1980; Kusznir and Karner, 2007), and to explain the differences observed in the tectonic architecture of modern magma-poor rifted continental margins. The first step in the evolutionary path in these models (Fig. 2a) is the formation of offset rift basins via distributed deformation (Huismans and Beaumont, 2007; Beaumont and Ings, 2012; Chenin and Beaumont, 2013). The subsequent extensional deformation becomes focused leading to the development of a major basin bounded by listric faults. This rifting process is characterized by a profound asymmetric geometry and formation of transitional zones with variable width where the thickness of the continental crust decreases abruptly. Little surface magmatism occurs during continued rifting. The resulting magma-poor continental margins and the related OCTZs are thus characterized by exhumation of serpentinized continental mantle lithosphere as well as high- grade metamorphic rocks of lower crust (Fig. 2b). The best modern analogue for such magma-poor rifted margins is the Iberia–Newfoundland conjugate margins pair (e.g. Péron-Pinvidic and Manatschal, 2009; Van Avendonk et al., 2009; Sibuet and Tucholke, 2013). A very thin oceanic crust may form in the initial stages of spreading, following the continental breakup.

The majority of rifted continental margins worldwide are

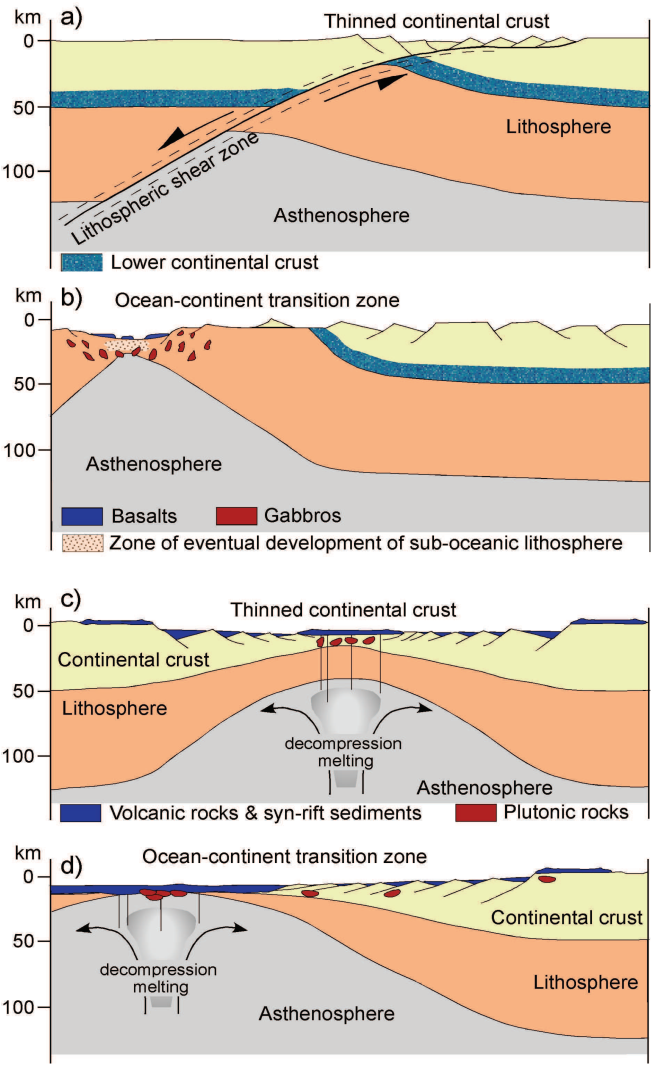

Figure 2. Simplified rift-drift models for magma-poor (a, b) and volcanic (c, d) continental rifted margins.

represented by volcanic types, which are generally characterized by thick wedges (up to 15 km) of volcanic flows (Hinz, 1981; Mutter et al., 1982; Menzies et al., 2002; Skogseid, 2001) that form seaward dipping seismic reflectors. The rifting mechanism produces a crustal architecture that shows an overall symmetry (Fig. 2c). Crustal thinning typically occurs over a comparably short distance in the order of 50 to 100 km. As a consequence, the volcanic rifted continental margins are typically devoid of exposures of subcontinental mantle or high-grade lower crust. The lack of strong passive margin subsidence during and after breakup and the presence at depth of a dense lower crustal body with anomalously high seismic velocities (Vp>7.1 km/s) are also characteristic features of volcanic rifted margins (Planke et al., 1991; Eldholm et al., 2000). Volcanic rifted margins are commonly associated with large igneous provinces (LIPs) onshore (Fig. 2c), and major regional dike swarms and sills (Coffin and Eldholm, 1994; Dilek, 2003b). Examples of LIPs associated with volcanic rifted margins include the North Atlantic Igneous Province (e.g., Kerr, 1994; Storey et al., 2007) and the Kerguelen–India–Antarctica–southwestern Australia area (e.g., Coffin et al., 2002). As the regional extensional stress regime remains active and rifting continues, a seafloor spreading system becomes established following the crustal breakup whereby normal oceanic crust is produced (Fig. 2d).

Continental Margin Ophiolites of Western Tethys

Regional Geology

The ophiolites in the Northern Apennines, Alpine Corsica, and Western Alps (Fig. 1) represent the remnants of the Western Tethys Ocean (also known as Liguria-Piemonte Ocean or Alpine Tethys). This oceanic basin developed in a western seaway of the Tethyan realm between Europe–Corsica and Adria continental domains after the Triassic to middle Jurassic rifting and the middle to late Jurassic seafloor spreading episodes (Manatschal and Müntener, 2009 and quoted references). The Western Tethys oceanic basin has been interpreted as a narrow (<600 km-wide) seaway, developed as a result of ultra slow spreading (e.g., Marroni and Pandolfi, 2007). Major reorganizations in plate motions and plate boundary configurations within the Tethyan system in the late Cretaceous led to oblique convergence between Europe-Corsica and Adria that resulted in the development of an intra-oceanic subduction zone within the Western Tethys oceanic basin (e.g., Marroni et al., 2010 for the Northern Apennines and Compagnoni, 2003 for the Western Alps). Following the terminal closure of the Western Tethys oceanic basin in the early Tertiary, Adria and Europe collided leading to the development of a thick, doubly-verging orogenic wedge in the Alpine-Apennine mountain system. Continental margin ophiolites were thrust eastwards in the Northern Apennines and westward in the Western Alps and Alpine Corsica during this continental collision stage.

The ophiolites in the Western Alps and the Schistes Lustrés of Corsica display eclogite to blueschist facies metamorphic overprint as a record of the subduction events in the Western Tethys oceanic basin. In contrast, the ophiolites in the Internal Ligurian (IL) units of the Northern Apennines, and Alpine Corsica (Pineto, Balagne and Rio Magno) are only weakly metamorphosed but strongly deformed. These features are interpreted to have resulted from shallow level, low-P accretion (Marroni and Pandolfi, 2003; Marroni et al., 2004; Meneghini et al., 2009). The upper mantle peridotites, gabbros and basalts in the External Ligurian (EL) units occur, however, in a different structural position. They occur as blocks embedded in a Santonian-Campanian sedimentary mélange, which also includes fragments of granitoids and mafic granulites of the Hercynian continental crust (e.g. Casanova and Ragola Complexes; Marroni et al., 2001).

Geochemistry

Mantle lherzolites

Studies of the ophiolites in the Western Alps and the Northern Apennines have revealed high degrees of chemical and isotopic heterogeneity in their upper mantle peridotites (Rampone et al., 1995, 1998, 2005, Montanini et al., 2006; Rampone and Hofmann, 2012). These findings led to the recognition of two major types of upper mantle sequences in the Ligurian units of the Northern Apennines: the mantle peridotites of the EL units, and the mantle peridotites of the IL units.

The upper mantle peridotites of the EL units consist mainly of fertile spinel lherzolites, representing the subcontinental lithospheric mantle exhumed during the early stages of continental rifting, which led to the opening of the Jurassic Western Tethys oceanic basin. The isotopic ages of these peridotites range between 2.4 Ga and 780 Ma, and their Sr and Nd isotopic compositions are also similar to those of the subcontinental orogenic spinel lherzolites documented from the western Mediterranean region (Rampone et al., 1995). The fertile nature of these upper mantle peridotites is demonstrated by relatively high REE concentrations coupled with a moderate LREE/HREE depletion (Fig.3a), as well as high Al_2O_3 (Fig.3b) and CaO contents (Rampone et al., 1995). Several meters-thick garnet-bearing clinopyroxenite to websterite layers locally occur within the mantle sequences of the EL units (Montanini et al., 2006).

The upper mantle peridotites of the IL units consist of clinopyroxene-poor spinel lherzolites, which represent a rare example of the depleted mantle lithosphere of the Jurassic Ligurian Tethys (Rampone et al., 1998). These rocks exhibit severe depletion in highly incompatible elements, such as LREE (Fig. 3a), and less pronounced depletion in Al_2O_3 (Fig. 3b), CaO, Sc, and V. These geochemical features are consistent with those of a residual mantle, which

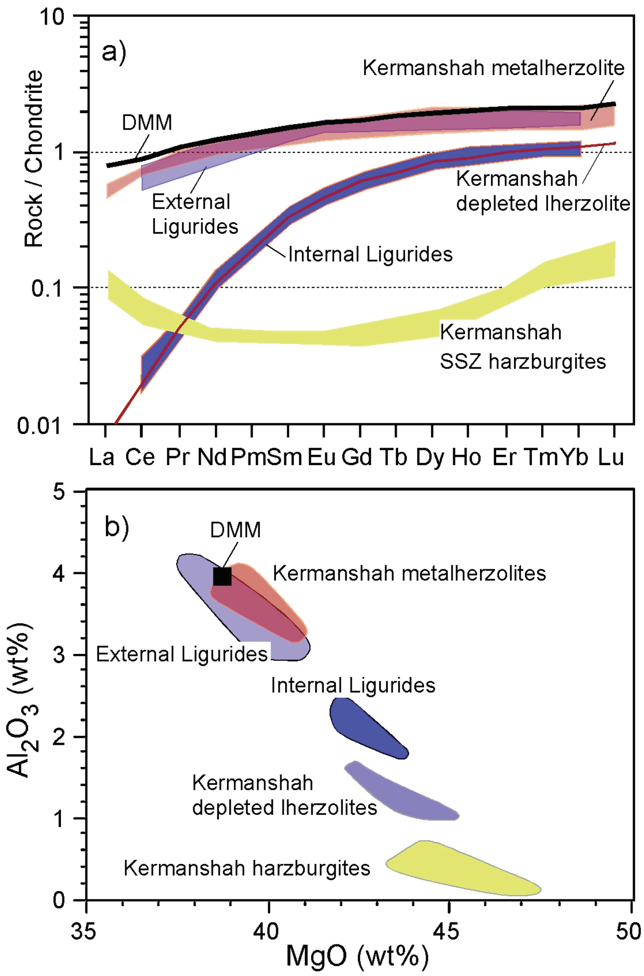

Figure 3. (a) Chondrite–normalized rare earth element (REE) compositional variation, and (b) variation of Al_2O_3 vs. MgO for mantle peridotites from the Western Tethys and Zagros ophiolites. The average composition of the depleted MORB mantle (DMM) is shown for comparison (data from Workmann and Hart, 2005). Normalizing values are from Sun and McDonough (1989). For data source, see Table 1, as well as Allahyari et al. (2010) for Kermanshah depleted lherzolites and harzburgites.

underwent low-degrees (<10%) of repeated episodes of partial melting, initiated in the garnet stability field (Rampone et al., 1998). The available Sr and Nd isotopic compositions of these peridotites indicate an extremely depleted signature, and the Sm/Nd model ages suggest their partial melting experience during the Permian (Rampone et al., 1998). These observations and data suggest that the mantle rocks in the IL units represent the products of upwelling and melting of a MORB-type asthenosphere that formed in response to the onset of regional extensional tectonics, which led to the opening of the Western Tethys Ocean.

Basaltic and metabasaltic rocks

Although limited in volume, basaltic rocks are widespread in the Western Tethyan ophiolites. They mainly occur as blueschists and eclogites in the Western Alps, Alpine Corsica, and Calabria, and as unmetamorphosed basaltic lavas in the IL and EL units, as well as in some units in Alpine Corsica (Balagne, Nebbio, Pineto, Rio Magno). Two varieties of basaltic rocks are identified in the Western Tethyan ophiolites, regardless of their metamorphic grade: (1) basaltic rocks showing the typical geochemical features of N-MORB, and (2) N-MORB lavas displaying marked garnet influence in the melt source; these basaltic rocks have been defined as G-MORB (garnet-influenced MORB) by Saccani (2015).

N-MORB rocks crop out in Alpine Corsica and the EL units of the Northern Apennines, and are represented mainly by basalts and minor basaltic andesites. These rocks show a clear sub-alkaline nature, as evidenced by their very low Nb/Y ratios (0.04-0.09). Their TiO_2 (0.98 – 1.78 wt%), P_2O_5 (0.13 – 0.40 wt%), Zr (53-192 ppm) and Y (20-48 ppm) contents, as well as Ti/V ratios (29-48), are highly similar to those of modern MORBs. The Nb and Th values plot along the MORB-OIB array and cluster towards relatively depleted compositions (Fig. 4a). These rocks show LREE depletion with respect to both MREE (La_N/Sm_N = 0.47-0.67) and HREE (La_N/Yb_N = 0.53-0.91). In the discrimination diagram of Saccani (2015), these basaltic rocks plot in the typical N-MORB field (Fig. 4b).

Basaltic rocks with G-MORB affinities are volumetrically predominant in the Western Tethys ophiolites with respect to N-MORB (Table 1). These rocks generally show a wide range of fractionation degrees (Mg# = 81.7 - 50.5). Basaltic compositions are volumetrically predominant, whereas basaltic andesitic and ferrobasaltic rocks are subordinate. These basaltic rocks have overall chemical compositions that are highly similar to that of N-MORBs, as evidenced by their TiO_2 (1.07 – 2.33 wt%), P_2O_5 (0.10 – 0.44 wt%), Zr (37 - 281 ppm) and Y (15 - 57 ppm) contents, as well as by their Nb/Y (0.04 – 0.20) and Ti/V ratios (28-56). Although G-MORBs largely overlap with N-MORB compositions in Figure 4a, there is a tendency for G-MORBs to plot towards compositions that are less depleted in Th and Nb. Because of their comparatively higher contents in these elements as well as in their LREE/HREE ratios, some authors have interpreted these rocks as transitional-type MORBs (T-MORBs), generated from slightly enriched mantle sources during the onset of oceanic spreading (e.g., Venturelli et al., 1979). In their re-evaluation of this interpretation, Montanini et al. (2008) and Saccani et al. (2008a) have observed that these rocks typically show significant depletion in HREE with respect to MREE, and that their $(Dy/Yb)_N$ ratios (1.15 to 1.59; Fig. 4b) are higher than that of typical N-MORB (Dy_N/Yb_N = 1; Sun and McDonough, 1989). They have suggested, therefore, that the LREE/HREE enrichment was related to depletion in HREE rather than enrichment in LREE. They hence interpreted the HREE/MREE depletion as a clear garnet signature of their mantle sources.

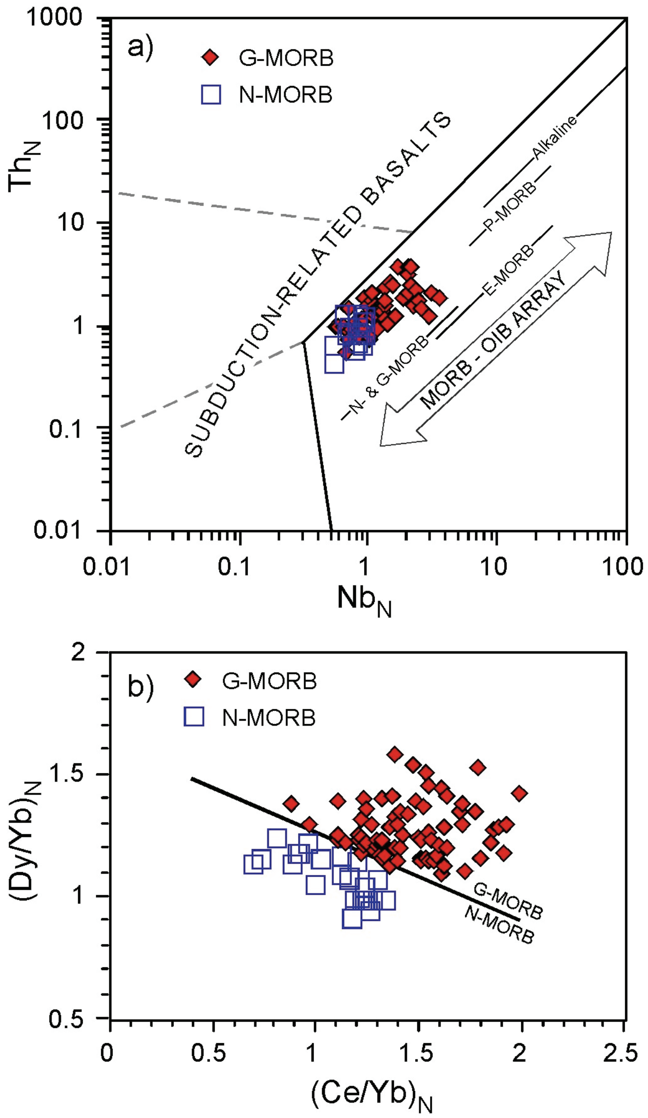

Figure 4. (a) N-MORB-normalized Th_N vs. Nb_N diagram (Saccani, 2015) for basalts and metabasalts from the Western Tethys (Alps, Alpine Corsica, Apennine, and Calabria) ophiolites. Abbreviations, MORB: mid-ocean ridge basalt, N-: normal type, E-: enriched type, P-: plume type, G-: garnet-influenced type. Bars showing the compositional variation of N-, E-, P-MORBs and alkaline basalt from worldwide ophiolites are from Saccani (2015). (b) Chondrite-normalized $(Dy/Yb)_N$ vs. $(Ce/Yb)_N$ diagram (Saccani, 2015) used for discriminating between G-MORB (garnet-influenced MORB) and N-MORB. Compared to N-MORBs, G-MORBs show a marked garnet signatures testified by HREE/MREE depletion (Saccani, 2015). Normalization values in both panels are from Sun and McDonough (1989). For data source, see Table 1.

Mantle sources and petrogenesis of basaltic and metabasaltic rocks of the Western Tethys ophiolites

Compositional variations observed in mafic rock suites commonly reflect the effects of fractional crystallization of olivine, plagioclase, and clinopyroxene in relatively evolved rocks (see Saccani et al.,

Table 1. Age, regional distribution, geochemical type, and related references of Jurassic and Triassic rocks in the key areas examined in this paper. [(1)]: Age constrained on geological bases (see Saccani et al., 2008b for the W Vardar zone ophiolites and Saccani et al., 2013a for the Kermanshah ophiolites)

Region	Unit	Age	Rock-type	References
		Western Tethys ophiolites		
		Mantle lherzolites		
N. Apennine	External Liguride			Rampone et al., 1995
	Internal Liguride			Rampone et al., 1998
		Basalts		
W Alps	Platta	Jurassic	G	Desmurs et al., 2002
Corsica	Pineto, Rio Magno, Balagne	Jurassic	G, N	Saccani et al., 2000; Saccani et al., 2008a
	Schistes Lustres	Jurassic	G, N	Saccani et al., 2008a
N Apennine	External Ligurides	Jurassic	G, N	Marroni et al., 1998; Vannucci et al., 1993
	Internal Ligurides	Jurassic	G	Cortesogno & Gaggero, 1992; Rampone et al., 1998
Calabria	Pollino	Jurassic	G	E. Saccani's unpublished data
		Albanide-Hellenide ophiolites		
		Basalts		
Mirdita	W & E Belt mélanges	L-Triassic	N, E	Bortolotti et al., 2004; Tashko et al., 2007
	Porava (S Mirdita)	M-L-Triassic	N, AB	Bortolotti et al., 2006
Hellenides	Othrys mélange	M-L-Triassic	N, E, P, AB, CAB	Photiades at al., 2003; Bortolotti et al., 2008; Monjoie et al., 2008; Capedri et al., 1997
	Koziakas mélange	M-L-Triassic	N, E, P, AB	Capedri et al., 1997; Saccani et al., 2003a; Chiari et al., 2012
	Argolis mélange	Triassic[(1)]	N, E, P, AB	Capedri et al., 1997; Saccani et al., 2003b; Saccani & Photiades, 2005
	W Vardar zone mélange	Triassic[(1)]	N, E, AB, CAB	Saccani et al., 2008b
	Vardoussia mélange	M-L-Triassic	N, P, CAB	Capedri et al., 1997; Bortolotti et al., 2009
	Avdella mélange (Pindos)	L-Triassic	N, P, AB	Capedri et al., 1997; Jones & Robertson, 1991
	Kerassies	L-Triassic	E, AB	Pe-Piper, 1998
	Evvia & Samos	E-M-Triassic	AB	Pe-Piper & Panagos, 1989; Pe-Piper & Kotopouli, 1991
	Atalanti	Triassic	CAB	Capedri et al., 1997
	Edipsos, Kremasta, Giona	M-Triassic	CAB	Pe-Piper, 1983; Pe-Piper & Panagos, 1989, Pe-Piper & Piper, 1991
	Formation à blocs	L-Triassic	CAB	Pe-Piper, & Mavroniki 1990
	Metalliko-Kilkis	E-Triassic	CAB	Asvesta, 1992
		Zagros ophiolites		
		Metalherzolites		
Zagros	Kermanshah	Pre-Jurassic[(1)]		Saccani et al., 2013a
		Metagabbros and metadykes		
	Kermanshah	Pre-Jurassic[(1)]	G, N	Saccani et al., 2013a
		Basalts		
	Kermanshah	Pre-Jurassic[(1)]	E, P, AB	Ghazi Hassanipak, 1999; Saccani et al., 2013a
	Sarve-Abad	Pre-Jurassic[(1)]	E, P	Saccani et al., 2014

2008a). Most of the rocks examined in this study have Mg# less than 70, that is, lower than the Mg# of > 70 inferred for primary melts of MORB-type lherzolites (Sinton and Detrick, 1992). For this reason, our discussion on the petrogenesis of these rocks is based only on the most primitive samples. The petrogenesis of basalts and metabasalts in the Western Tethys ophiolites and the characteristic features of their mantle sources have been discussed extensively by Montanini et al. (2008) for the EL units, and by Saccani et al. (2008a) for the basalt and metabasalt occurrences in Alpine Corsica. We summarize the salient points and conclusions of these studies here, using the LREE/HREE vs. MREE/HREE ratios (Fig. 5) that are particularly useful for highlighting the garnet influence in mantle melt sources.

The main distinctive feature between N-MORBs and G-MORBs lies in different fractionation trends among LREE, MREE, and HREE that cannot be explained by fractional crystallization alone; however, it can be explained by variable influence of a garnet signature in their mantle sources (Montanini et al., 2008). The garnet signature can be related either to deep initiation of melting in the garnet peridotite stability field, or to the melting of a heterogeneous mantle source characterized by garnet-bearing mafic/ultramafic layers. The diagram in Figure 5 shows that primitive N-MORB compositions are compatible with 10 – 20% partial melting of a depleted MORB-type mantle (DMM) source in the spinel stability field. In contrast, the compositions of primitive G-MORBs are compatible with either partial melting of a DMM source bearing garnet pyroxenites, or partial

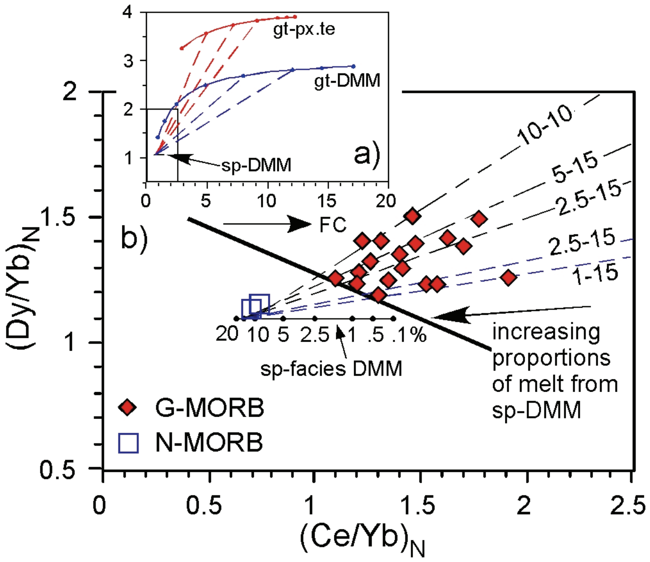

Figure 5. (a) Batch melting curves on $(Dy/Yb)_N$ vs. $(Ce/Yb)_N$ diagram for a garnet pyroxenite (gt-px.te) and a depleted MORB mantle (DMM) source in both garnet (gt) and spinel (sp) stability fields. Dashed lines represent the mixing lines of various melt fractions from different sources. DMM composition is from Workman and Hart (2005) garnet-pyroxenite composition is from Liu et al. (2005); Source modes, melting proportions, and partition coefficient are given in Appendix A. Box indicates the area expanded in panel (b). (b) Plot of the most primitive basalts and metabasalts from Western Tethys ophiolites on the close up of the melting model in panel (a). Red dashed lines represent the mixing lines of various melt fractions between DMM in the spinel stability field + garnet-pyroxenite relics; blue dashed lines represent mixing lines of various melt fractions between DMM in the spinel stability field + DMM in the garnet stability field. The percentages of melt fractions from each source are indicated on the mixing line (e.g., 5-15 indicates 5% partial melt of sp-DMM that mix with 15% partial melt of gt-pyroxenite). Mixing proportions between different partial melts are not shown. However, samples plot in the range from 0.9 – 0.6 melt from sp-DMM + 0.1 – 0.4 melt from other sources. Abbreviations, N-MORB: normal-type mid-ocean ridge basalt; G-MORB: garnet influenced MORB. See text for further explanations.

melting of a DMM source that started to melt in the garnet lherzolite field (~1 – 2.5% melting) and that it continued to higher degrees (~15% melting) in the spinel lherzolite field. All samples from the Western Tethys basalts and metabasalts fall within the mantle array (Fig. 4a), suggesting that a chemical influence of a lower crust component was limited or absent.

Basaltic lavas in the EL units of the Northern Apennines were likely the products of partial melting of a heterogeneous sub-continental lithospheric mantle bearing small volumes of garnet pyroxenite layers (Montanini et al., 2008). There exist several different interpretations for the inferred origin of garnet in the mantle melt source of these lavas. It might have been produced by relict garnet-pyroxenite material left in the depleted mantle source after the delamination and sinking of parts of the deep garnet-pyroxenite-bearing lithospheric mantle (Piccardo, 2008). Alternatively, it might have resulted from partial melting of a depleted MORB-type mantle source that initially started in the garnet-facies mantle and then continued in the spinel-facies mantle (Saccani et al., 2008a).

Continental Margin Ophiolites in the Albanide-Hellenide Segment of the Alpine Orogen

Geology

The NW-SE-trending Albanide-Hellenide mountain belt (Fig. 1) is part of the Alpine orogenic system and stretches from Slovenia and Serbia in the north to the mainland Greece in the south. Its late Mesozoic-Cenozoic tectonics was controlled mainly by the oblique convergence and continental collision between Adria and Europe, whereas its early Mesozoic tectonics was dominated by continental rifting, rift-drift tectonics and opening of an oceanic domain within the Neotethyan realm (e.g., Shallo and Dilek, 2003; Dilek et al., 2007, 2008; Bortolotti et al., 2013 and quoted references). Whether there was one or were two separate basins in this oceanic domain is still a subject of debate in the literature. Regardless of the paleogeography of the inferred ocean basin(s), final continental break-up and initial seafloor spreading in the Albanide-Hellenide segment of the Neotethyan system appear to have occurred in the Carnian–Norian (Bortolotti et al., 2013).

The main rift-drift events that lead to the opening of the Neotethys basin(s) in the region took place in the early–middle Triassic (Robertson et al., 1991). Upper Scythian and lower Anisian sub-marine lavas, locally associated with red limestones and cherts, are overlain by transgressive mid-Triassic limestones in the Tripolitsa, Subpelagonian and Pelagonian tectonic zones (Pe-Piper, 1998). Submarine volcanic rocks that are stratigraphically associated with late Anisian radiolarian cherts occur in the sub-ophiolitic mélanges in the Albanide-Hellenide belt (Jones et al., 1992, Chiari et al., 1996; Ozsvart et al., 2011). Rift-related volcanism appears to have continued into the late Triassic, as documented in numerous localities in the Albanide-Hellenide belt (see Table 1). Similar Triassic rift-related volcanic rocks intercalated with hemipelagic sedimentary rock sequences are widespread elsewhere in the Dolomite Alps of northern Italy (Winterer and Bosellini, 1981), in the Dinarides (Bébien et al., 1978, Pamiç, 1984), and in the Taurides of southern Turkey (Dilek and Rowland, 1993; Dilek et al., 1999).

Geochemistry of Triassic mafic volcanic rocks

Five main types of Triassic mafic to intermediate volcanic rock associations are distinguished in the Albanide-Hellenide belt (see Table 1 for localities and references): (1) calc-alkaline (CAB) and minor shoshonitic basalts, basaltic andesites and andesites; (2) alkaline basalts and subordinate trachybasalts, trachyandesites, and trachytes (e.g., Saccani et al., 2003a); (3) transitional to sub-alkaline basalts showing marked LREE/HREE enrichment, resembling plume-type MORB (P-MORB) compositions; (4) sub-alkaline basalts showing moderate LREE/HREE enrichment, resembling enriched-type MORB (E-MORB) compositions; and (5) sub-alkaline basalts showing N-MORB chemistry. We briefly summarize below the geochemical features of these main types of Triassic basaltic rock associations based on the existing literature (see Table 1 for regional distribution and references). Most of these volcanic rock associations are incorporated into sub-ophiolitic mélange units as discrete tectonic slices or blocks. However, only volcanic rocks that are spatially associated with Triassic sedimentary rocks (either carbonate rocks or radiolarian cherts) are described below.

Calc-alkaline volcanic rocks

Triassic calc-alkaline volcanic rocks are found in several extrusive units of the Hellenide belt, and rarely in the sub-ophiolitic mélanges of the Othris, Vardoussia, and Almopias ophiolites (Table 1). The calc-alkaline rocks in the Metalliko-Kilkis area are Lower Triassic in age (Asvesta, 1992), whereas all the other rock units are Middle to Upper Triassic in age. The main rock types include basalts, basaltic andesites, andesites, dacites and rhyolites, although basalts and basaltic andesites are more extensive by volume. Basaltic and andesitic rocks have low to moderate TiO_2 (0.7–1.5 wt%), Zr (64 - 218 ppm), and Y (16 – 47 ppm). They are sub-alkaline in nature (Fig. 6a), show significant Th enrichment with respect to Nb (Fig. 6b), and plot in the compositional field for calc-alkaline basaltic rocks. These rocks are characterized by incompatible element patterns with positive anomalies of Th, U, La, Ce and marked negative anomalies of Ta, Nb, Hf, and Ti (e.g., Pe-Piper and Piper, 1991; Capedri et al., 1997; Pe-Piper, 1998; Bortolotti et al., 2009). They show highly variable LREE/HREE enrichments, with La_N/Yb_N ratios ranging from ~2 to ~11.

Alkaline basaltic rocks

Triassic alkaline rocks occur as hundreds of meters-thick successions in the volcanic sequences of the Porava unit in South Albania (Bortolotti et al., 2006), and on the islands of Evvia and Samos in Greece (Pe-Piper and Panagos, 1989; Pe-Piper and Kotopouli, 1991). They are also widespread in sub-ophiolitic mélanges of the Subpelagonian zone, where they are mainly represented by basalts, and subordinately by trachybasalts, trachyandesites, and trachytes (Capedri et al., 1997; Saccani et al., 2003a; Saccani and Photiades, 2005; Bortolotti et al., 2009). In the Vardar zone to the east, however, they are represented by metabasaltic rocks showing intense deformation (Saccani et al., 2008b). The alkaline nature of these rocks is marked by their Nb/Y ratios >1.4 (Fig. 6a). Basaltic rocks have relatively high TiO_2 (1.4 – 3.4 wt%) and P_2O_5 (0.2 – 1 wt%) contents, and display significant enrichment in Th and Nb (Fig. 6b), as well as marked LREE/HREE enrichments (La_N/Yb_N = 3.8-26.0). These elements and many incompatible element ratios (e.g., Th/Yb = 0.8-4.3; Ta/Yb = 0.4-3.2) are highly similar to those observed in typical ocean island basalts (OIB) (Sun and McDonough, 1989), suggesting that magmas of these rocks in the Vardar Zone may have originated from an OIB-like mantle source. Pb and Nd isotope compositions of the Evvia and Samos alkali basalts resemble those of Saint Helena or Afar OIB, although their Pb isotopic compositions are similar to those of the Aegean granitoids suggesting a hint of crustal contamination (Pe-Piper, 1998). However, all transitional to alkaline basalts plot within the MORB - OIB array in Figure 6b, indicating that any possible chemical contamination from the lower crust should have been very limited.

Highly enriched (plume-type) mid-ocean ridge basalts (P-MORBs)

Middle to Late Triassic P-MORBs consisting of pillowed and massive lavas occur as thrust sheets within sub-ophiolitic mélanges beneath the Hellenide ophiolites (Table 1). These rocks are

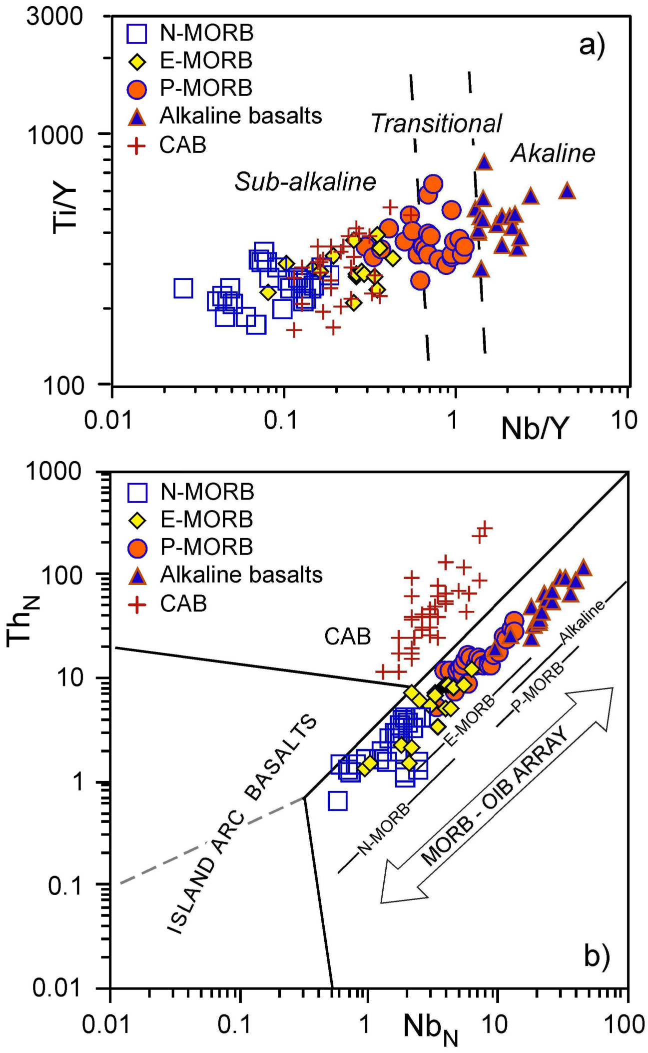

Figure 6. (a) Ti/Y vs. Nb/Y diagram (Pearce, 1982), and (b) N-MORB-normalized Th_N vs. Nb_N diagram (Saccani, 2015) for Triassic basalts from the Albanide-Hellenide ophiolites. Abbreviations, MORB: mid-ocean ridge basalt, N-: normal type, E-: enriched type, P-: plume type, CAB: calc-alkaline basalt. Bars showing the compositional variation of N-, E-, P-MORBs and alkaline basalt from worldwide ophiolites are from Saccani (2015). Normalization values are from Sun and McDonough (1989). See Table 1 for data source.

characterized by rather variable chemical compositions and fractionation degrees (Mg#=80-52). Their compositions range from basalt to ferrobasalt (FeO_t = ~11–14 wt%), indicating that they are sub-alkaline to transitional in nature (Fig. 6a). They have TiO_2 = 0.83-2.28 wt%, P_2O_5 = 0.15-0.43 wt%, Zr = 45-196 ppm, and Y = 13-40 ppm, reminiscent of the geochemical features of plume-type MORBs (e.g., Schilling et al., 1983) generated at plume-proximal mid-ocean ridge settings. These basalts show marked enrichment in Th and Nb (Fig. 6b), as well as in Ta, Hf, and LREE. Their LREE abundance is generally lower than 100 x chondrite (e.g., Saccani and Photiades, 2005) with La_N/Yb_N=2.71-7.82.

Enriched mid-ocean ridge basalts (E-MORBs)

Middle to Upper Triassic E-MORBs are widespread in the sub-ophiolitic mélanges of both the Mirdita-Subpelagonian and Vardar tectonic zones in the Albanide-Hellenide belt (Table 1). The E-MORB rocks consist of pillowed and massive lavas of basaltic and ferrobasaltic compositions, showing a clear sub-alkaline affinity (Fig. 6a). The E-MORBs examined in this study have rather variable chemical compositions and fractionation degrees (Mg#=75-52). Their TiO_2 (0.61-1.70 wt%) values are generally lower than those in P-MORBs, whereas their P_2O_5 (0.08-0.42 wt%), Zr (30-125 ppm), and Y (14-42 ppm) contents are comparable to those of P-MORBs. The examined E-MORBs are characterized by a moderate Th and Nb (Fig. 6b) enrichment coupled with moderate LREE/HREE (La_N/Yb_N=1.4-3.3) and LREE/MREE (La_N/Sm_N=1.0-1.7) enrichments. The overall geochemistry of these rocks is highly similar to that of modern E-MORBs (Sun and McDonough, 1989).

Normal mid-ocean ridge basalts

Middle to Upper Triassic rocks showing N-MORB chemistry are mainly found in sub-ophiolitic mélanges (Table 1). Nonetheless, they are also common in the Middle to Upper Triassic volcanic successions of the Porava unit (South Albania), where they are associated with alkaline basalts (Bortolotti et al., 2006). N-MORBs are represented by basalts, basaltic andesites, and ferrobasalts in the Mirdita-Subpelagonian zone, as well as by intensely deformed metabasalts in the sub-ophiolitic mélanges of the West Vardar zone (Saccani and Photiades, 2005; Saccani et al., 2008b).

N-MORBs display a clear sub-alkaline affinity (Fig. 6a) and are characterized by variable chemical compositions and fractionation degrees (Mg# = 70.0 - 52.0 in basalts and basaltic andesites and 58.6 - 37.0 in ferrobasalts). However, their overall geochemical features are similar to those of basalts generated at mid-ocean ridges. They have TiO_2 = 0.75 - 2.43 wt%, P_2O_5 = 0.03 - 0.30 wt%, Zr = 45 - 125 ppm, and Y = 21 - 49 ppm, and their FeO_t contents ranges from 7.19 to 11.19 wt%. They display little Th, and Nb depletion or enrichment relative to other incompatible elements (e.g., Pe-Piper, 1998; Saccani and Photiades, 2005; Bortolotti et al., 2006, 2008). In Figure 6b their Th and Nb compositions plot within the typical N-MORB values (Saccani, 2015). Although they show a wide range of compositional variations, most of the samples cluster towards more enriched compositions and partly overlap with E-MORB samples. Basalts show variable depletions in LREE with respect to MREE and HREE, as evidenced by their La_N/Sm_N (0.46 - 0.99) and La_N/Yb_N (0.38 - 1.00) ratios, in line with their N-MORB affinity.

Mantle sources and petrogenesis of basaltic rocks

Calc-alkaline basalts (CAB) display Ta, Nb, and Ti depletion, suggesting that their mantle source was already depleted as a result of previous melt extraction events (Pearce, 1982). The Th/Nb relative enrichment (Fig. 6b) suggests that these mantle sources were significantly influenced by an arc-type geochemical component (Pearce, 1982; Saccani, 2015). CAB magmatism peaked in the middle late Triassic, and was spatially and temporally associated with rift-related igneous activity producing alkaline basalts and highly enriched to depleted MORBs in the absence of an active subduction (Pe-Piper, 1998). Isotopic data suggest that CABs formed under extensional conditions from partial melting of hydrated, melt-depleted peridotites in the sub-continental mantle along the northern margin of Gondwana (Pe-Piper, 1998). Depletion and further hydration of this mantle source probably took place earlier, during a Hercynian subduction event.

Several authors have suggested, based on isotopic and trace element data, that Triassic alkaline basalts of the Albanide-Hellenide ophiolites were derived from partial melting of a plume-type source; variably enriched, subalkaline and transitional MORB rocks were derived, on the other hand, from partial melting of MORB-type mantle sources, which were influenced by different OIB-type chemical contribution (e.g., Pe-Piper, 1998; Saccani and Photiades, 2005). Partial melting that generated enriched basalts may have occurred either in the deep (garnet-facies) or in the shallow (spinel-facies) mantle (Saccani and Photiades, 2005).

To constrain the nature of the mantle source and the depth of melting during the rift-drift phases is fundamental for better reconstructing the nature of geodynamic processes occurring during the initial phases of an oceanic basin evolution. However, a rigorous quantification of the melting processes is not possible because the composition of the mantle sources can only be postulated. Saccani and Photiades (2005) have shown that small variations in the hygromagmatophile element ratios within each rock group can be observed suggesting that magmas of different rocks within each group may have been derived from mantle sources that were slightly different in composition. However, a semi-quantitative modeling of the incompatible elements can place some constraints on the mode and nature of the mantle processes at a regional scale.

We present in Figure 7 a non-modal, batch partial melting model, using Th and Nb/Yb ratio. This diagram has the advantage to combine two types of information in a single plot. The abundance of Th and Nb is used to evaluate the enrichment of the source, whereas the Nb/Yb ratio is sensitive to the presence of residual garnet in the source. In this Figure, four compositionally different mantle sources that undergo partial melting in both garnet- and spinel-facies are considered: (1) DMM source (Workman and Hart, 2005); (2) OIB-type source (Lustrino et al., 2002) with Nb = 1.5 ppm, Th = 0.18 ppm, Yb = 0.353 ppm; (3) a theoretical, slightly enriched DMM source with Nb = 0.63 ppm, Th = 0.08 ppm, Yb = 0.353 ppm; and (4) a theoretical, enriched DMM source with Nb = 0.79 ppm, Th = 0.10 ppm, Yb = 0.353 ppm. The compositions of the sources 3 and 4 were assumed based on modeling, as recently presented by Saccani et al. (2013a, 2014).

Primitive alkaline basalts are compatible with low degree partial melting (0.5 – 1%) of an OIB-type mantle source starting in the garnet-facies mantle and continuing to greater extent (2.5 – 5%) in the spinel-facies (polybaric melting). Alternatively, the chemistry of these basalts can be explained by very low degree of melting (< 0.1%) of enriched DMM sources in the garnet facies (Fig. 7). However, such a very low degree of melting is unreasonable, and therefore this hypothesis is ruled out.

The chemistry of N-MORB rocks generally points to melt generation from a depleted MORB-type mantle at shallow levels. In fact, the Th-Nb-Yb composition of primary N-MORBs can be explained by variable degrees (generally, 8 - 20 %) of partial melting of a DMM source in the spinel facies (Fig. 7).

E-MORBs have Th and Nb compositions, which are slightly enriched compared to those of N-MORBs (Fig. 6b). Therefore, it can be postulated that E-MORBs represent melts derived from more enriched sources compared to N-MORBs. The modeling in Figure 7 shows that E-MORB compositions are compatible with variable degrees (~5 - 20 %) of partial melting of a slightly enriched DMM

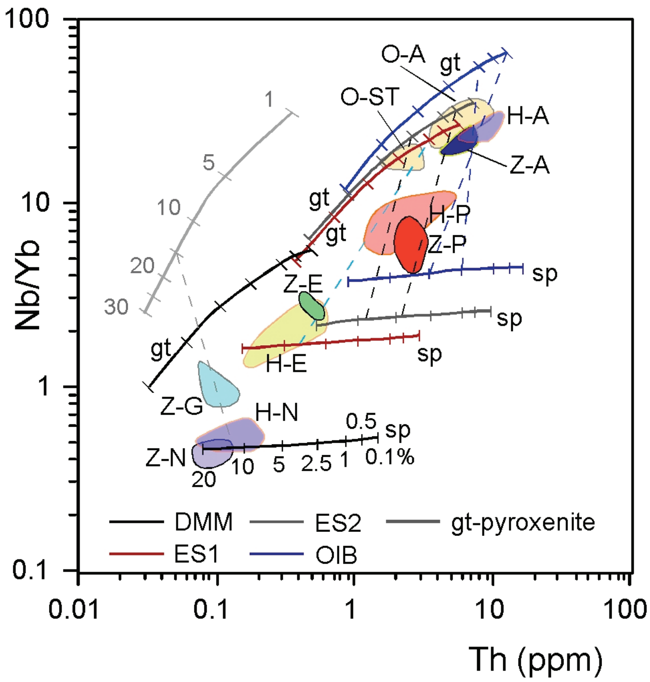

Figure 7. Plot of the Th vs. Nb/Yb compositional variations of the most primitive basalts from the Albanide-Hellenide and Zagros ophiolites, as well as batch melting curves for garnet-pyroxenite, depleted MORB mantle (DMM), and variably enriched theoretical mantle sources (ES1, ES2, OIB) in both garnet (gt) and spinel (sp) stability fields. ES1: slightly enriched source, ES2: moderately enriched source; OIB: ocean island basalt-type enriched source. Dashed lines represent the mixing lines of various melt fractions from different sources. DMM composition is from Workman and Hart (2005). Source modes, melting proportions, and partition coefficient are given in Appendix A. Ticks on the melting curves indicate the same percentages of melt fractions as shown for the sp-DMM. See text for further explanations. Abbreviations, O: Oman; H: Albanide-Hellenide; Z: Zagros; A: alkaline; ST: subalkaline-transitional; P: plume-type mid-ocean ridge (MOR); E: enriched-type MOR; N: normal-type MOR; G: garnet-influenced MOR.

source (ES1) in the spinel facies. Alternatively, E-MORBs may have been derived from polybaric melting of a DMM source, starting in the garnet facies. However, the REE compositions of these rocks (not shown) cannot be explained by this melting process alone.

P-MORBs have Th and Nb contents that plot toward the enriched values within the MORB–OIB array (Fig. 6b). Although definitely enriched in many incompatible elements, the overall chemical composition of these rocks points to a MORB-type composition. Therefore, we can rule out the possibility of their genesis from an OIB-type source. However, it can be postulated that these rocks represent melts derived from more enriched sources compared to N-MORBs and E-MORBs. The diagram in Figure 7 shows that the Th-Nb-Yb composition of primitive P-MORBs is compatible with polybaric melting of the theoretical enriched source (ES2) starting in the garnet-facies (0.5 – 2.5% melt) and continuing into the spinel facies (5 – 10% melt). Alternatively, primitive P-MORBs may have been derived from polybaric melting of a slightly enriched source (ES1). However, this hypothesis would require implausible melting proportions, that is: very low degree of melting in the garnet-facies (<0.5%), followed by a much greater extent of melting in the spinel-facies (e≥20%). In addition, the difference between hygro-magmatophile element ratios of E-MORBs and P-MORBs (not shown) suggests that they were derived from compositionally distinct mantle sources.

Continental margin ophiolites in the Zagros Belt

Regional Geology

The Zagros ophiolites crop out along the Main Zagros Thrust Zone (MZTZ) in western Iran and display a complete record of the rift-drift, seafloor spreading and subduction zone tectonic evolution of the Southern Neotethyan Ocean. The Southern Neotethyan seaway opened in the Iranian–Oman segment during the Early–Middle Triassic (Agard et al., 2005; Robertson, 2007) and evolved between Arabia and the Sanandaj-Sirjan continental block (Berberian and King, 1981; Alavi, 1994; Agard et al., 2005; Mouthereau et al., 2012). The MZTZ ophiolites include, from NW to SE, the Penjween ophiolites in Iraq (Aswad et al., 2011) and the Sarve-Abad, Kermanshah, Neyriz, and Baft ophiolites in Iran (Ghazi and Hassanipak, 1999; Babaie et al., 2006; Allahyari et al., 2010, 2014; Rajabzadeh and Dehkordi, 2013; Shafaii Moghadam et al., 2013; Saccani et al., 2013a, 2014). They are temporally equivalent to the Oman ophiolite in the south (Glennie, 2000).

From the southwest to the northeast and structurally upward, three main tectonic units occur along the MZTZ: (1) the Zagros Fold Belt, (2) the Zagros Crush zone (or High Zagros) with ophiolites, and (3) the Sanandaj-Sirjan zone (SSZ). The Zagros fold belt consists mainly of Triassic to Upper Cretaceous shelf carbonates overlain by a Paleocene to Pliocene sedimentary succession. These shelf deposits have been interpreted to represent the Arabian passive margin (Stöcklin, 1968).

The Sanandaj-Sirjan zone consists of metamorphic and sedimentary sequences (Stöcklin, 1968; Berberian and King, 1981) that are composed of a Precambrian–Paleozoic continental basement, unconformably overlain by Permian-Triassic limestones, Jurassic phyllites with metavolcanic rocks, and Barremian-Aptian limestones (Stöcklin, 1968). The Jurassic sequence is intruded by late Jurassic–late Cretaceous calc-alkaline plutons. From the Middle Jurassic to the Cretaceous, the Sanandaj-Sirjan zone represented an Andean-type margin with robust calc-alkaline magmatism (e.g., Stöcklin, 1968; Berberian and King, 1981; Ghasemi and Talbot, 2006).

The Zagros Crush zone is composed of imbricated tectonic slices including: (1) Upper Triassic–Cretaceous limestones (Bisotun unit) and radiolarites deposited on the subsiding peri-Arabian shelf (see Stöcklin, 1968 for a detailed review); (2) ophiolitic units, (3) Upper Cretaceous sedimentary rocks and calc-alkaline andesitic lavas, and (4) Eocene–Miocene sedimentary deposits and volcanic rocks. Although Sabzehei and others (1968) described the Upper Cretaceous sedimentary-volcanic sequences as part of the Crush zone, it is likely that these units belong to the Andean-type volcanic arc that developed on the southern margin of the Sanandaj-Sirjan continental block (Berberian and King, 1981) because they show a clear calc-alkaline affinity.

Ophiolitic remnants commonly occur in a tectonic mélange (Stöcklin, 1968), which includes ophiolitic subunits as isolated and dismembered blocks. These ophiolitic subunits are represented by: (1) upper mantle peridotites; (2) ultramafic cumulates, (3) gabbroic sequences; (4) dike rocks; and (4) pillow basalts. Also included in this melange are metamorphosed lherzolites and gabbros that are intruded by numerous basaltic dikes.

Upper mantle peridotites are volumetrically the most abundant ophiolitic subunit, and are represented mainly by harzburgites and subordinate depleted lherzolites, locally intruded by pyroxenite dikes and dunite–chromitite bands (Allahyari et al., 2010; Rajabzadeh and Dehkordi, 2013; Shafaii Moghadam et al., 2013). Ultramafic cumulates have been described from the NW part of the Zagros Belt, namely in the Penjween area (Aswad et al., 2011) and in the Sarve-Abad ophiolites (Allahyari et al., 2014). These upper mantle peridotites and ultramafic cumulates are interpreted as the mantle residua and the cumulate sequence, respectively (Allahyari et al., 2010, 2014; Aswad et al., 2011; Rajabzadeh and Dehkordi, 2013; Shafaii Moghadam et al., 2013). Gabbros crop out in the Sarve-Abad, Kermanshah, Neyriz and Baft ophiolites, and are locally associated with troctolites and/or diorites and plagiogranites, as well as basaltic dikes and pillow lavas (Ghazi and Hassanipak, 1999; Babaie et al., 2006; Allahyari et al., 2010, 2014; Rajabzadeh and Dehkordi, 2013; Saccani et al., 2013a, 2014; Shafaii Moghadam et al., 2013). Gabbros and basalts mainly occur as isolated blocks in the Kermanshah, Neyriz and Baft areas, whereas an incomplete ophiolite sequence of cumulate and isotropic gabbros, microgabbros, a dike complex, a dike/pillow transition zone and pillow lavas exist in the Sarve-Abad area. Gabbroic and volcanic rocks from the MZTZ ophiolites display a wide range of geochemical affinities, ranging from E-MORB to P-MORB, and alkaline basalts and trachytes (Ghazi and Hassanipak, 1999; Babaie et al., 2006; Allahyari et al., 2010, 2014; Rajabzadeh and Dehkordi, 2013; Saccani et al., 2013a, 2014; Shafaii Moghadam et al., 2013).

Metamorphosed ophiolites have so far been described only from the Kermanshah area, where they mainly consist of foliated metagabbros crosscut by numerous metabasaltic dikes. Strongly serpentinized and foliated meta-peridotites occur below these metagabbros and consist of sub-continental meta-lherzolites, analogous to those exposed in the EL units of the Northern Apennines (Saccani et al., 2013a). Metagabbros and the associated mafic dikes have a general N-MORB affinity, but two sub-groups of rocks are also recognized: (1) rocks showing no garnet influence, and (2) rocks showing a marked garnet influence (Saccani et al., 2013a).

The Zagros ophiolites were emplaced onto the Arabian continental margin in the Campanian–Maastrichtian as a result of an arc–continent collision (Desmons and Beccaluva, 1983; Ricou, 1994; Agard et al., 2005; Dilek and Sandvol, 2009; Aswad et al., 2011; Saccani et al., 2013a). Andean-type, active margin magmatism continued along the southern edge of the Sanandaj-Sirjan block between the late Cretaceous and the Eocene, and produced calc-alkaline volcanic and plutonic rocks (Berberian and King, 1981; Agard et al., 2005; Ghasemi and Talbot, 2006). The collision of Arabia with Eurasia in the middle Miocene caused thrusting of the Eocene magmatic rocks over the ophiolite complexes.

Geochemistry

Metalherzolites, metagabbros and metabasaltic dikes

Metalherzolites of the Kermanshah ophiolite can readily be distinguished from the lherzolites representing MORB residual mantle and harzburgites, representing SSZ residual mantle in other MZTZ ophiolites (e.g., Penjween: Aswad et al., 2011; Neyriz: Babaie et al., 2006; Rajabzadeh and Dehkordi, 2013; Baft: Shafaii Moghadam et al., 2013). The Zagros metalherzolites have higher HFSE (high field strength elements) and REE concentrations (Fig. 3a) compared to the other upper mantle peridotites (Saccani et al., 2013a). Their REE values (Fig. 3a) and TiO_2, Al_2O_3 (Fig. 3b), CaO, MgO, Ni, and Cr contents are highly similar to those of the sub-continental lherzolites in the EL units of the Northern Apennines (Saccani et al., 2013a). Hence, we have interpreted these Zagros metalherzolites to represent the sub-continental mantle of the Arabian margin (Saccani et al., 2013a).

Magmatic protoliths of the metagabbros were composed of olivine-gabbros and both cumulate and isotropic gabbros, whereas those of dike intrusions in the metagabbros were made of basalts (Saccani et al., 2013a). Both isotropic metagabbros and metabasaltic dikes have relatively high contents of TiO_2, Zr, Y, and V, and relatively low Al_2O_3. These rocks are subalkaline in nature (Fig. 8a), and plot along the MORB-OIB array towards relatively depleted compositions in the discrimination diagrams (Fig. 8b). They can be subdivided, however, into distinct, N-MORB and G-MORB groups on this discrimination diagram (Fig. 8b).

The Zagros N-MORB rocks include some of the metabasaltic dikes, which display smooth MREE/HREE enrichments (Dy_N/Yb_N = 1.07 to 1.11) and marked LREE/MREE depletion (Ce_N/Dy_N = 0.68 - 0.87). In contrast, G-MORB rocks include metagabbros and some metabasaltic dikes. These G-MORB rocks display flat N-MORB normalized HFSE patterns, and chondrite-normalized REE compositions showing LREE/MREE depletion (Ce_N/Dy_N = 0.77 – 1.22), coupled with marked MREE/HREE enrichment (Dy_N/Yb_N = 1.31 – 1.44) (Saccani et al., 2013a). Such MREE/HREE enrichment is significantly higher than that of typical N-MORB (e.g., Dy_N/Yb_N=0.99, Sun and McDonough, 1989). These values are comparable to those of equivalent basalts from the EL units in the Northern Apennines.

Volcanic rocks

The Zagros E-MORBs are represented by basalts and basaltic andesites showing a sub-alkaline nature (Fig. 8a). These rocks are characterized by slightly enriched LILE/HFSE and LREE/HREE patterns. They also have a generally lower concentration of incompatible elements with respect to those of P-MORBs and alkaline rocks (e.g., Saccani et al., 2013a).

P-MORBs in the Zagros ophiolites are represented by basalts, basaltic andesites and rare andesites that generally have a sub-alkaline to transitional nature (Fig. 8a) These rocks display multi-element patterns significantly enriched in LILE and LREE compared to HFSE and HREE, respectively, with $(La/Yb)_N$ ratios ranging from 2.46 to 4.37 (see Ghazi and Hassanipak, 1999; Saccani et al., 2013a, 2014). These patterns are similar to those of typical P-MORB from modern oceanic settings (e.g., Schilling et al., 1983).

Alkaline rocks in the Zagros are represented mainly by basalts and subordinately by trachybasalts and trachytes, displaying a clear alkaline nature (Fig. 8a). Alkaline basalts show high abundance in HFSE with respect to N-MORB and display LREE/MREE and LREE/HREE enriched patterns (see Ghazi and Hassanipak, 1999; Saccani et al., 2013a, 2014). In the discrimination diagram in Figure 8b, these

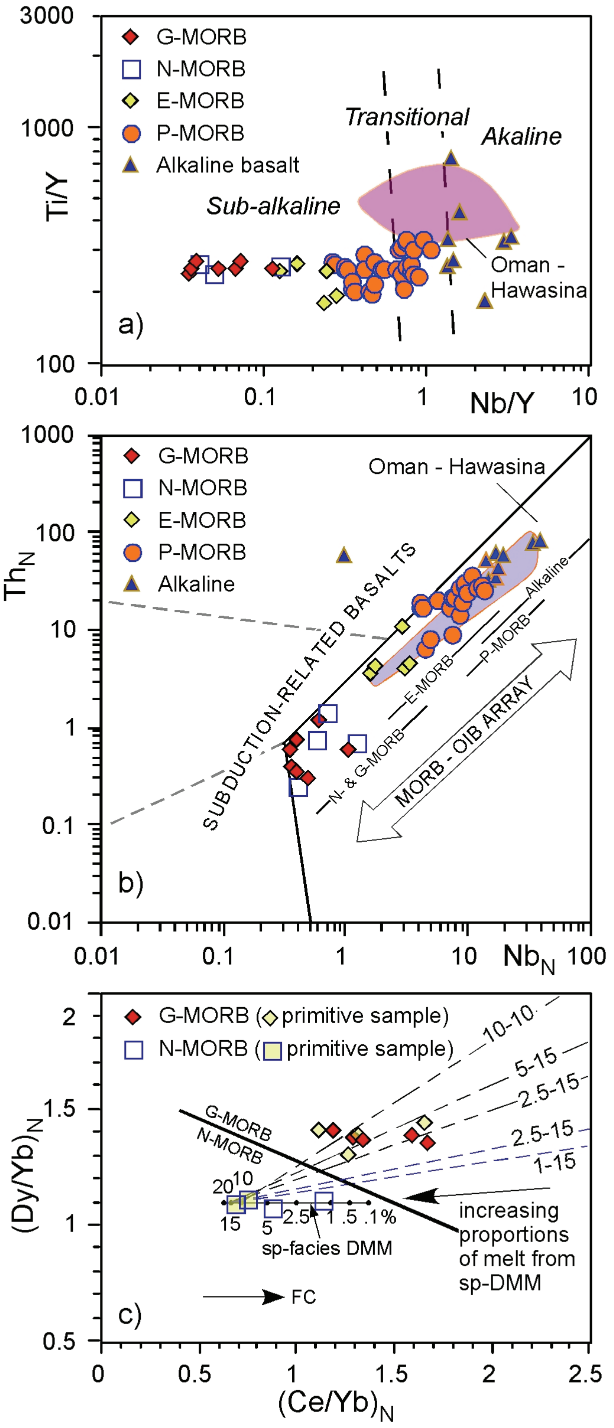

Figure 8. (a) Ti/Y vs. Nb/Y diagram (Pearce, 1982), and (b) N-MORB-normalized Th_N vs. Nb_N diagram (Saccani, 2015) for basalts and metabasalts from the Zagros ophiolites. Abbreviations, MORB: mid-ocean ridge basalt, N-: normal type, E-: enriched type, P-: plume type, G-: garnet-influenced type. Bars showing the compositional variation of N-, G-, E-, P-MORBs and alkaline basalt from worldwide ophiolites are from Saccani (2015). The compositional variation of basalts from the Hawasina Nappe from Oman ophiolites (Chauvet et al., 2011) is shown for comparison. (c) Chondrite-normalized $(Dy/Yb)_N$ vs. $(Ce/Yb)_N$ diagram (Saccani, 2015) used for discriminating between G-MORB and N-MORB. Normalization values in both panels (b) and (c) are from Sun and McDonough (1989). See Table 1 for data source.

volcanic units from the MZTZ ophiolites show a continuous compositional variation from the less enriched to the more enriched rocks. Their overall geochemistry resembles that of alkaline basalts generated in within-plate ocean island settings (OIB-type).

Mantle sources and petrogenesis of basaltic rocks

Our modeling in Figure 7 shows that N-MORB primitive metabasalts may have been derived from ~15–20% non-modal batch partial melting of a DMM source (Workman and Hart, 2005) in the spinel-facies. In contrast, the LREE/HREE (Ce/Yb) and MREE/HREE (Dy/Yb) ratios of G-MORB metagabbros and primitive metabasalts are compatible with ~10–15% partial melting of DMM with garnet-pyroxenite layers in the spinel-facies (Figs. 7 and 8c). Based on these figures we can exclude a possible genesis of G-MORBs from polybaric melting of a DMM source starting in the garnet-facies and then continuing into larger degrees of partial melting in the spinel-facies. Such melting process would result in comparatively lower MREE/HREE ratios. Alternatively, the LREE/HREE and MREE/HREE ratios of G-MORB could theoretically be generated by large degree (10–15%) of partial melting of a DMM source in the garnet-facies, followed by a lower extent of melting in the spinel facies (Fig. 7). However, this scenario is also implausible.

Many geochemical indicators (basically, highly incompatible and hygromagmatophile elements) suggest that the E-MORBs, P-MORBs, and alkaline basalts in the Sarve-Abad and Kermanshah ophiolites were originated from chemically distinct mantle sources, which started melting at different depths (see Saccani et al., 2013a, 2014). Primary E-MORBs are compatible with ~10% partial melting of a DMM source slightly enriched in LREE and incompatible elements (e.g., Th, Nb, and Ta), whereas the primary P-MORB may result from ~8% partial melting of a DMM source, significantly enriched in LREE and incompatible elements (Saccani et al., 2013a, 2014). Both these rock-types were generated from partial melting in the spinel-facies mantle. However, a very limited contribution from the garnet-facies mantle cannot be excluded. In contrast, the REE systematics displayed by the alkaline basalts suggests that primary melts of these rocks were produced by the mixing of small melt fractions (~1%) from garnet-facies with larger melt fractions (~4%) from spinel-facies, derived from partial melting of an OIB-type lherzolite mantle. The results of our modeling, using Th vs. Nb/Yb in Figure 7, are consistent with these conclusions. We can see in this figure that mantle sources that are progressively enriched in Th and Nb are required to explain the compositional variations observed from E-MORBs to P-MORBs and alkaline basalts in the Zagros ophiolites. It is worth noting that the oldest melts in the production of the Middle to Late Triassic basaltic rocks in the conjugate Oman margin show isotopic evidence for lower crustal contamination.

Discussion

Tectono-magmatic model for the rift-drift evolution of the Alpine Tethys

The rifting model proposed for the Western Tethys oceanic basin is mainly based on the reconstruction of the architecture of a deformed and metamorphosed continental margin pair, currently exposed in the Northern Apennines and Alpine Corsica. The first rifting stage started in the Middle Triassic (Fig. 9a) and was preceded by a longlived extensional deformation episode in the Permo-Triassic that involved the transition from a gravitational collapse of the Variscan orogenic crust to the inception of a true rift-drift phase. In the Northern Apennines, evidence for this rifting phase can be found at Punta Bianca, where a middle Triassic extensional basin sequence consisting of marine deposits intercalated with alkaline basaltic flows crops out (Stoppa, 1985). Rifting-related middle Triassic faults also occur in the continental domain of Alpine Corsica and in the Western Alps (Durand-Delga; 1984; Froitzheim and Manatschal, 1996). The geometry of these normal faults, which are west-facing in the Southalpine and Austroalpine domains (Bernoulli et al., 1979; Bertotti et al., 1993) and east-facing in the Briançonnais and Dauphinois domains (Lemoine and Trumpy, 1987), suggests that this first rifting phase was dominated by lithospheric-scale stretching by pure shear extension (Fig. 9a).

The second stage of rifting (Fig. 9b) via an asymmetric, simple-shear kinematics developed in the Middle Jurassic (e.g., Marroni et al., 1998). However, the main episode of normal faulting associated with dismemberment of the carbonate platforms took place in the early Jurassic, as documented from the Adria continental margin sequences in the Northern Apennines (Bernoulli et al., 1979), Alpine Corsica (Durand-Delga, 1984) and Western Alps (Bertotti et al., 1993). The asymmetric crustal architecture of these continental margins can be deduced from the structure of the Corsica – Northern Apennine conjugate margin pair.

Reconstruction of the OCTZ at the European margin indicates the existence of upper continental crustal rocks, displaying prominent topographic escarpments produced by high-angle normal faults (Fig. 9b1). Slide blocks in the late Cretaceous sedimentary mélanges (e.g., Casanova Complex; Marroni et al., 1998) of the EL units show, on the other hand, that the Adria rifted margin was characterized by a wide OCTZ with subcontinental lithospheric mantle and lower continental crust exhumed on the seafloor, and tectonically overlain by extensional allochthons (or extensional rafts) (Fig. 9b1). Adria represented the lower plate in this asymmetric extension scenario, whereas Europe marked the upper plate (Marroni and Pandolfi, 2007 and quoted references). The middle to late Jurassic oceanic crust formation was characterized by a magma-poor, slow-spreading mid-ocean ridge system. The more internal oceanic areas experienced limited volcanism, which took place directly on the exhumed, serpentinized mantle peridotites, gabbros, and/or ophiolitic breccias, as deduced from the IL units and the Schistes Lustrés units (Fig. 9b3). The middle to late Jurassic rift-drift stage was associated with upwelling of asthenospheric mantle in response to lithospheric extension and continental rifting. This extensional phase was associated with limited partial melting of heterogeneous mantle

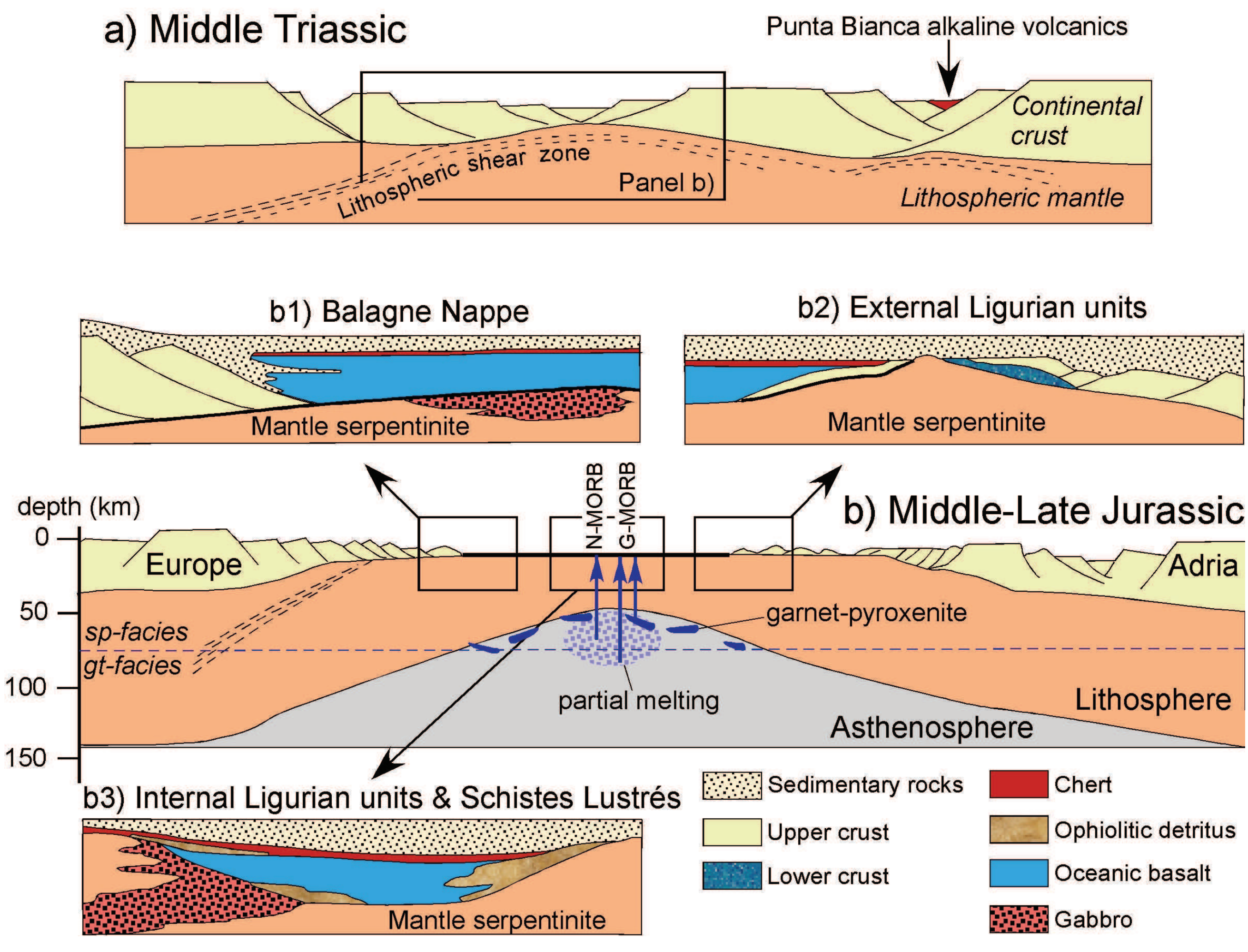

Figure 9. 2-D cartoons showing the rift-drift, tectono-magmatic evolution of the Western Tethys ocean basin (modified from Marroni and Pandolfi, 2007). Generalized stratigraphic and magmatic settings in different sections of the Western Tethys are also shown. (b1) European margin ocean-continent transition zone (OCTZ); (b2) Adria margin OCTZ; (b3) internal oceanic area.

sources, locally bearing garnet-pyroxenite relics (Fig.9b). Piccardo (2008) suggested that garnet-pyroxenite relics were left in the DMM melting source after the delamination and sinking of portions of the deep garnet-pyroxenite-bearing lithospheric mantle. This partial melting process resulted in the formation of G-MORB type rocks now exposed in the Western Tethyan ophiolites. However, a minor amount of G-MORBs may have been generated from polybaric partial melting, which started in the garnet lherzolite field and continued to higher degrees in the spinel lherzolite field. The model in Figure 9b can also explain the occurrence of volumetrically minor basalts with typical N-MORB compositions in several ophiolitic units in Alpine Corsica (Saccani et al., 2008a), as well as in the Ligurian ophiolites (Rampone et al., 2005). N-MORB primary melts could also be produced by partial melting of a pure DMM source, which did not contain garnet-bearing rocks.

Tectono-magmatic model for the rift-drift evolution of the Albanide-Hellenide belt

The Triassic rift-drift phase in the Albanide-Hellenide segment of Neotethys involved the formation of alkaline and variably enriched rocks with plume-like chemical signatures, and with calc-alkaline and shoshonitic rocks displaying SSZ-type chemical features. However, the extant geological evidence suggests that the Triassic extension and the early phase of oceanic crust formation at the OCTZ within the Albanide-Hellenide segment of Neotethys were not associated with a major mantle plume event or a contemporaneous active subduction (Pe-Piper, 1998). There is no physical evidence, such as regional doming, anomalous thermal regime, and basaltic plateaus, supporting the existence of a mantle plume in this area. In contrast, Wooler et al. (1992) have suggested that subsidence rates were small, and that there was hardly any uplift, with the rift shoulders remaining near the sea-level. The lack of magmatic evolution from more depleted to more enriched rocks that is commonly observed in plume-related magmatism, the absence of basaltic plateaus, and a relatively small volume of plume-related volcanic rocks collectively argue against the existence of a well established, long-lasting mantle plume in the region.

Calc-alkaline and shoshonitic rocks were produced, contemporaneously with the formation of alkaline basalts, between the early and late Triassic in both continental (e.g., Kremasta, Evia) and oceanic settings (e.g., Edipsos, Othrys mélange, and west Vardar zone mélange (see Table 1 for references). Magmas of these calc-alkaline rocks may have been produced by partial melting of hydrated, melt-depleted peridotite in the sub-continental mantle. Pe-Piper (1998) has suggested that a previous, Hercynian subduction event that produced an Andean-type arc on the northeastern margin of Gondwana may have played an important role in hydrating the subcontinental lithospheric mantle. Subsequent partial melting of this previously metasomatized lithospheric mantle during the Triassic extension and rifting episode would have then produced calc-alkaline and shoshonitic melts. The occurrence of similar extensional, Triassic calc-alkaline and shoshonitic rocks with similar petrogenesis has also been documented from the Southern Alps (e.g., Bonadiman et al., 1994).

We present in Figure 10 an interpretive model depicting the development of the rift-drift phases of the opening of the Albanide-Hellenide Neotethys. This model implies a heterogeneous sub-continental mantle, in which the hydrated-depleted (SSZ-type) mantle portions left by the earlier Hercynian subduction events occur in the lithospheric mantle. In contrast, the OIB-type metasomatized portions are likely to be prominent in the asthenospheric mantle (Fig. 10a). We infer that convective thinning of the lithosphere might have brought SSZ-type mantle portions of the lithospheric mantle into direct contact with asthenospheric temperatures. These SSZ mantle portions, which had their solidus lowered by the addition of volatiles, underwent partial melting leading to the eruption of calc-alkaline and shoshonitic lavas starting in the early Triassic. Meanwhile, the uprising OIB-type metasomatized asthenospheric mantle underwent polybaric partial melting, which generated alkaline basalts associated in space and time with the calc-alkaline rocks (Fig. 10b). Subsequently, during the middle and late Triassic variably enriched (E-MORB and P-MORB) and depleted (N-MORB) magmas were also generated by partial melting in the spinel-facies mantle (Fig. 10b) and were erupted at the OCTZ. We deduce that N-MORBs were generated from uprising primitive asthenospheric sources, whereas E-MORBs and P-MORBs may have been generated either from variably metasomatized mantle portions, or by complex interactions (e.g., source contamination, magma mixing, etc.) between OIB-type metasomatized mantle and primitive asthenospheric mantle. The abundance of calc-alkaline rocks

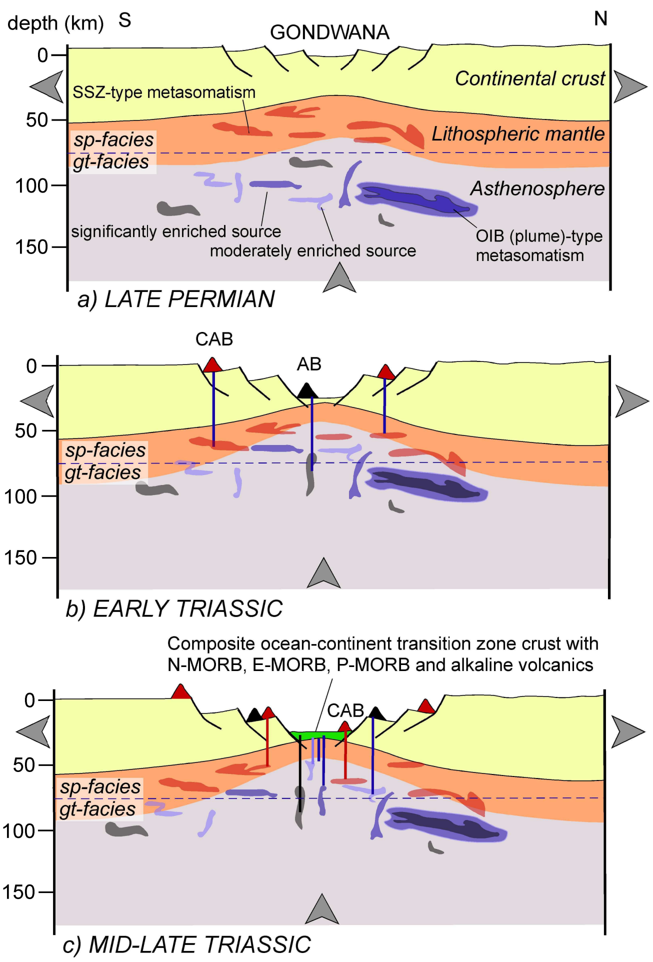

Figure 10. 2-D cartoons depicting the rift-drift, tectono-magmatic evolution of the Neotethyan Ocean in the Albanide-Hellenide sector. Abbreviations, SSZ: supra-subduction zone; OIB: ocean island basalt; CAB: calc-alkaline basalt; AB: alkaline basalt; MORB: mid-ocean ridge basalt; N-: normal-type; E-: enriched-type; P-: plume-type; sp: spinel; gt: garnet.

seems to have decreased from the middle to the late Triassic, and their occurrence appears to be more scattered compared to other rock types. This observation supports the hypothesis that SSZ-type metasomatized mantle portions were likely concentrated in the lithospheric mantle (Fig. 10a). By the end of the Triassic, the rift-drift phase gave way into seafloor spreading, and then the involvement of the sub-continental heterogeneous mantle sources in magmatism ceased. Consequently, the production of CABs, alkaline basalts, E-MORBs and P-MORBs ended in the latest Triassic. From the early Jurassic onwards, the oceanic crust was generated entirely by the eruption of N-MORBs in response to uprising of sub-oceanic primitive mantle.

Tectono-magmatic model for the rift-drift evolution of the Southern Neotethys

Triassic ophiolitic rocks found in the northwestern section of the MZTZ include: (1) metalherzolites and metagabbros, crosscut by metabasaltic dikes. Structurally and compositionally, all these rocks resemble those similar ophiolitic subunits in the Alpine-Apennine system; (2) gabbros, dikes and pillow lavas ranging in composition from E-MORB to P-MORB and alkaline basalts. Similar basalts are also abundant in the Oman – UAE ophiolites south of the Persian Gulf (Chauvet et al., 2011).

A tectono-magmatic model for explaining the tectonic evolution of the Kermanshah and Sarve-Abad ophiolites has been proposed earlier by Saccani et al. (2013a, 2014). Here, we present a refined version of this model and summarize it, with a particular focus on the rift-drift development of the Southern Neotethys (Fig. 11). Prior to the Triassic, this region was part of a coherent Arabian-Iranian continental platform (Fig. 11a). The Sanandaj-Sirjan continental block started rifting away from Arabia in the early Triassic, or perhaps earlier, in the late Permian (Ghasemi and Talbot, 2006; Azizi and Moinevaziri, 2009; Saccani et al., 2013a; Whitechurch et al., 2013). As in the Western Tethys, the Sanandaj-Sirjan block was rifted from the northern margin of Arabia through asymmetrical passive extension (Fig. 11b), as also postulated by Dilek et al. (1991) and Dilek and Thy (1998) for the Kizildag ophiolite in southern Turkey that is interpreted to be the westward continuation of the Zagros ophiolites.

This simple shear extension led to the exhumation of the sub-continental mantle of Afro-Arabia, which is now represented by the Kermanshah metalherzolites. In the meantime, upwelling of the underlying asthenosphere in response to lithospheric extension and continental rifting was associated with limited partial melting of heterogeneous mantle sources, locally bearing garnet-pyroxenite relics (Fig. 11b). Similar to the Western Tethys, this magmatic episode led to the emplacement of volumetrically small quantities of gabbros and dikes with either N-MORB or G-MORB chemistry, originated from partial melting of a pure DMM source or a DMM source bearing garnet-pyroxenite relics, respectively. However, in contrast to the Western Tethys, rift-drift and seafloor spreading stages in the Southern Neotethys were also characterized by the marked influence of OIB-type (plume-type?) components associated with the uprising of asthenospheric mantle (Fig. 11c). The polybaric partial melting of an OIB-type enriched source led to the production of alkaline basalts (Fig. 11d1), whereas the partial melting of a MORB-type mantle, which was heterogeneously modified by OIB-type components, resulted in the production of both P- and E-MORBs (Figs. 11d2, d3).

The MZTZ ophiolites represent a particular case of a Continental Margin-type ophiolite, as introduced by Dilek and Furnes (2011,

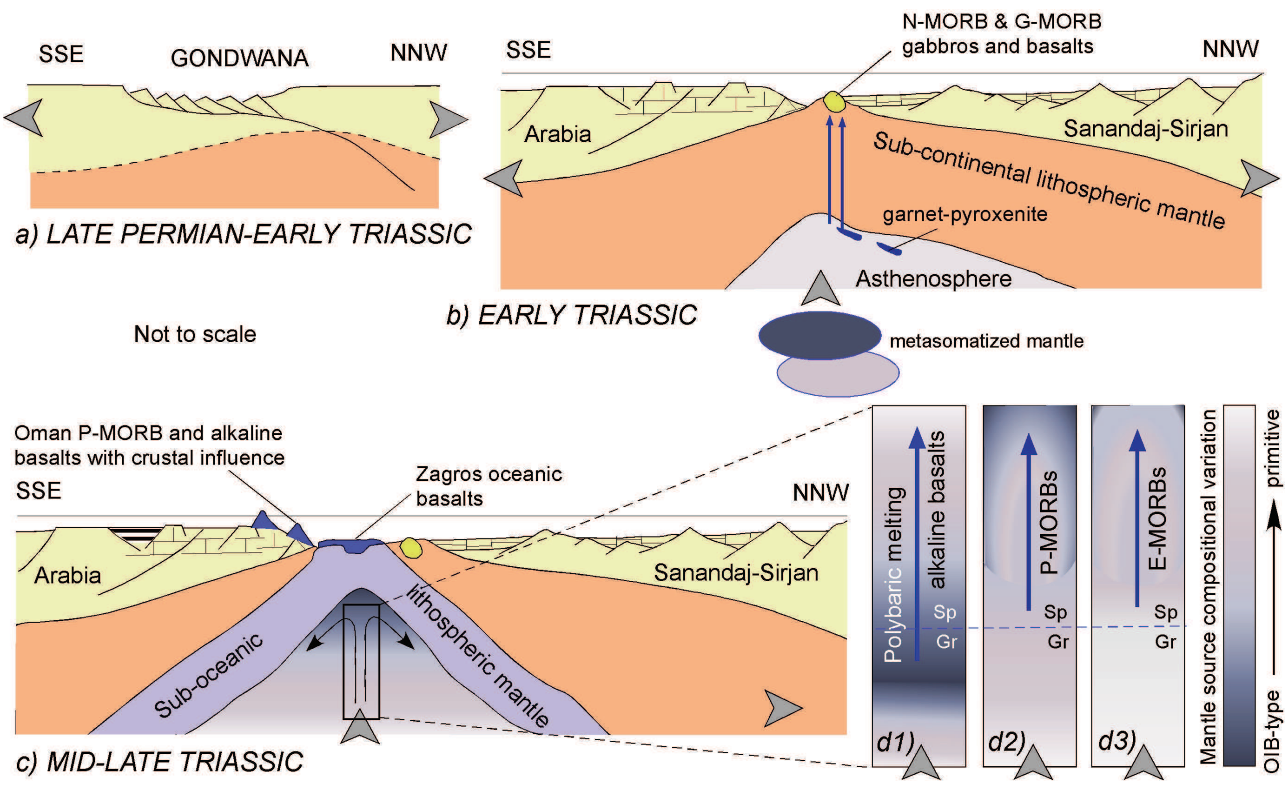

***Figure 11.** 2-D cartoons (modified from Saccani et al., 2013a) depicting the rift-drift, tectono-magmatic evolution of the Southern Neotethyan Ocean. Abbreviations, OIB: ocean island basalt; MORB: mid-ocean ridge basalt; N-: normal-type; E-: enriched-type; P-: plume-type; G: garnet-influenced type; Sp: spinel-facies mantle; Gt: garnet-facies mantle.*

2014). The mineral chemistry, geochemistry and isotopic data (Saccani et al., 2013a, 2014) suggest that these ophiolites may in fact represent the combination of Continental Margin- and Plume-type ophiolites. However, whether chemically enriched rocks represent derivation from OIB-like asthenosphere, or from a metasomatized mantle that was previously enriched by OIB-type components is unknown. In other words, the plume-type geochemical signature displayed by the middle Triassic MZTZ and Oman basalts does not necessarily provide evidence for the concomitant existence of an active mantle plume in the northern Gondwana area. Similar to our observations in the Triassic Neotethys section of the Albanide-Hellenide orogenic belt to the west, no clear evidence has been documented supporting the existence of a mantle plume in the Arabian-Iranian area during this time. Rather, a relatively small volume of plume-related volcanic rocks argues against the existence of a mantle plume. Chauvet et al. (2011) have suggested that the plume-type signature of the conjugate Oman lithospheric mantle was produced by pervasive metasomatism that was caused by the percolation of Permian plume-related melts through it (Lapierre et al., 2004). Mantle plume activities appear to have characterized the opening of Paleotethys during the Devonian-Carboniferous (see Xiao et al., 2008 and references therein; Xu et al., 2015). Therefore, in the lack of any direct evidence, we postulate that the OIB-type geochemical signature of the MZTZ ophiolites and Oman basalts was most likely inherited from previous mantle plume activities that occurred in the same area during the early Carboniferous opening of Paleotethys (Saccani et al., 2013b).

Conclusions

Continental Margin (CM) ophiolites consist of mafic-ultramafic and sedimentary rock assemblages developed during the continental breakup and embryonic oceanic crust generation at ocean-continent transition zones (OCTZ). These ophiolites are commonly highly dismembered and occur as tectonic slices in sub-ophiolitic mélanges. Their petrogenesis generally involves small degrees of partial melting of less depleted, continental lithospheric mantle and slowly upwelling asthenospheric mantle. Compositional heterogeneities in the sub-continental mantle strongly control the igneous stratigraphy and geochemical signatures of CM ophiolites. This heterogeneity is mainly a result of previous subduction events and/or earlier plume melt interactions with host peridotites, creating mantle contingency. This phenomenon is well illustrated by the occurrence of coeval calc-alkaline basalts and shoshonites, N-MORBs, alkaline basalts, E-MORBs, and P-MORBs in the Triassic CM ophiolites in the Albanide-Hellenide belt. The inferred hydrated and depleted (SSZ-type) mantle residues in the Adria continental lithospheric mantle were the products of earlier Hercynian subduction events in the late Paleozoic. PMORBs and E-MORBs in the Zagros Continental Margin ophiolites were produced by partial melting of a MORB-type continental lithospheric mantle, which was heterogeneously modified by previous mantle plume events that took place during the early Carboniferous opening of Paleotethys.

Continental margin ophiolites are also diverse in terms of their internal structure and igneous stratigraphy. Some CM ophiolites are devoid of mantle peridotites, analogous to volcanic rifted margins. The Triassic Albanide-Hellenide ophiolites are a good example of this type, although they have volumetrically limited volcanic products and extensional calc-alkaline rocks, atypical of volcanic rifted margins. The CM ophiolites in the Zagros belt (e.g., Kermanshah ophiolite) contain continental lithospheric mantle lherzolites and gabbros crosscut by basaltic dikes and overlain by basaltic pillow lavas. This igneous stratigraphy is similar to the one displayed by the EL ophiolites (Western Tethys ophiolites) and documented from non-volcanic (or magma-poor) rifted margins. However, upwelling of an asthenospheric mantle with plume-influenced components resulted in the eruption of alkaline basalts and enriched sub-alkaline basalts in the Kermanshah CM ophiolites, a phenomenon that is unlike of magma-poor, rifted margins. These findings from the three cases of CM ophiolites we report in this study indicate that the fossil examples of the OCTZ crustal architecture and magmatic record do not invariably match with those of the two end-members of rifted margins (volcanic vs. non-volcanic). The rates and kinematics of rifting, the existence (or not) of an active plume, upwelling rates of the asthenospheric mantle in response to lithospheric thinning, and the mode and scale of mantle heterogeneity caused by previous subduction and plume events strongly affect the rift architecture and the geochemical character of an OCTZ lithosphere.

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APPENDIX - A

Initial compositions and melting modes of different sources, as well as the partition coefficients (KD) used in the melting models in Figs. 5, 7, 8c.

Source composition	Depleted MORB mantle	Enriched source 1 (ES1)	Enriched source 2 (ES2)	Ocean island basalt	Garnet pyroxenite
Th (ppm)	0.0068	0.08	0.1	0.18	0.017
Nb (ppm)	0.1277	0.6	0.8	1.5	0.48
Yb (ppm)	0.353	0.353	0.353	0.353	0.84

	ol	opx	cpx	gt	sp
Source Mode					
garnet-facies[a]	0.57	0.21	0.13	0.09	
spinel-facies[b]	0.578	0.27	0.119		0.033
garnet-pyroxenite			0.7	0.3	
Melting Mode					
gt-facies[a]	0.04	-0.19	1.05	0.11	
sp-facies[b]	0.1	0.27	0.5		0.13
garnet-pyroxenite			0.7	0.3	
KD					
garnet-facies					
Yb	0.0015[1]	0.049[1]	0.28[1]	5.73[2]	
Th	0.0001[1]	0.058[3]	0.00026[1]	0.0001[1]	
Nb	0.063[4]	0.01[5]	0.05[1]	0.0003[6]	
spinel-facies					
Yb	0.0015[1]	0.049[1]	0.28[1]		0.01[1]
Th	0.0001[1]	0.058[3]	0.00026[1]		0.0024[7]
Nb	0.063[4]	0.003[5]	0.05[1]		0.001[8]
garnet-pyroxenite					
Yb			0.43[9]	7.86[10]	
Th			0.0083[11]	0.001[10]	
Nb			0.05[1]	0.0008[10]	

Abbreviations, ol: olivine; opx: orthopyroxene; cpx: clinopyroxene; gt: garnet; sp: spinel.

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