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Geodynamic and petrogenetic evolution of Alpine ophiolites from the central and NW Dinarides: an overview

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Abstract

Dismembered ophiolites occur in the Dinaride Ophiolite Zone (DOZ) that is related to the open-ocean Tethyan realm, whereas highly dismembered ophiolites occur in the Vardar Zone (VZ) related to a back-arc basin. The ophiolites of DOZ are associated with a Jurassic olistostrome mélange (DOZM), the youngest component of which are Tithonian limestone exotics and with the Mesozoic bed-to-bed Radiolarite Formation. Late Jurassic/Early Cretaceous to Late Cretaceous clastic sequences, comprising redeposited fragments of ophiolites, disconformably overlie the DOZM. Ophiolites of VZ are associated with tectonized ophiolite mélange (VZM), the youngest component of which are Late Cretaceous–Paleogene limestone exotics. The VZM is associated with the Late Cretaceous–Paleogene flysch formation. Ophiolites of both the DOZ and the VZ are predominantly peridotite tectonites, represented mainly by fertile spinel lherzolite in the western and central part of DOZ and VZ, and by depleted harzburgites in their southeastern parts. Cumulate ultramafics and gabbros are subordinate and are in some places overlain by massive or sheeted dyke complexes, capped by metabasaltic pillow lavas. Metamorphic soles of ophiolites are represented by varieties of amphibolites with subordinate pyroxenite schists and scarce eclogites with ultramafic interlayers, which were progressively metamorphosed under P–T conditions of eclogite (?), granulite, amphibolite and greenschist facies. The according protoliths are cumulate gabbros in the DOZ, medium-grade bimineralic epidote–amphibolite facies amphibolites derived from diabase–dolerites, and low- to medium-grade metapelites and metapsammites. K–Ar and Sm–Nd measurements yield ages of 174 ± 14–136 ± 15 Ma on ophiolites from DOZ and 109.6 ± 6.6–62.2 ± 2.5 Ma on ophiolites from VZ. Basic petrological and geochemical features for all Dinaridic ophiolites and associated amphibolites are correlatively presented both for DOZ and VZ. Dinaridic ophiolites were generated in the Dinaridic Tethys over the period of about 150 Ma. The bulk of oceanic crust was generated during the Late Triassic to pre-Late Jurassic/Early Cretaceous when oceanic subduction processes, accompanied by DOZ ophiolite obduction onto the Apulian margin, started. Generation of the oceanic crust continued during the Cretaceous–Early Paleogene in a reduced Dinaridic Tethys under back-arc setting. Eocene closure of the Dinaridic Tethys was accompanied by the second emplacement of VZ ophiolites and the final structuration of the Dinarides and their uplift. At the end, geological and petrological similarities and dissimilarities of ophiolites from both DOZ and VZ are presented.

Keywords: Mesozoic ophiolites; Dinarides; Dinaride Ophiolite Zone; Vardar Zone; Geodynamics; Petrology

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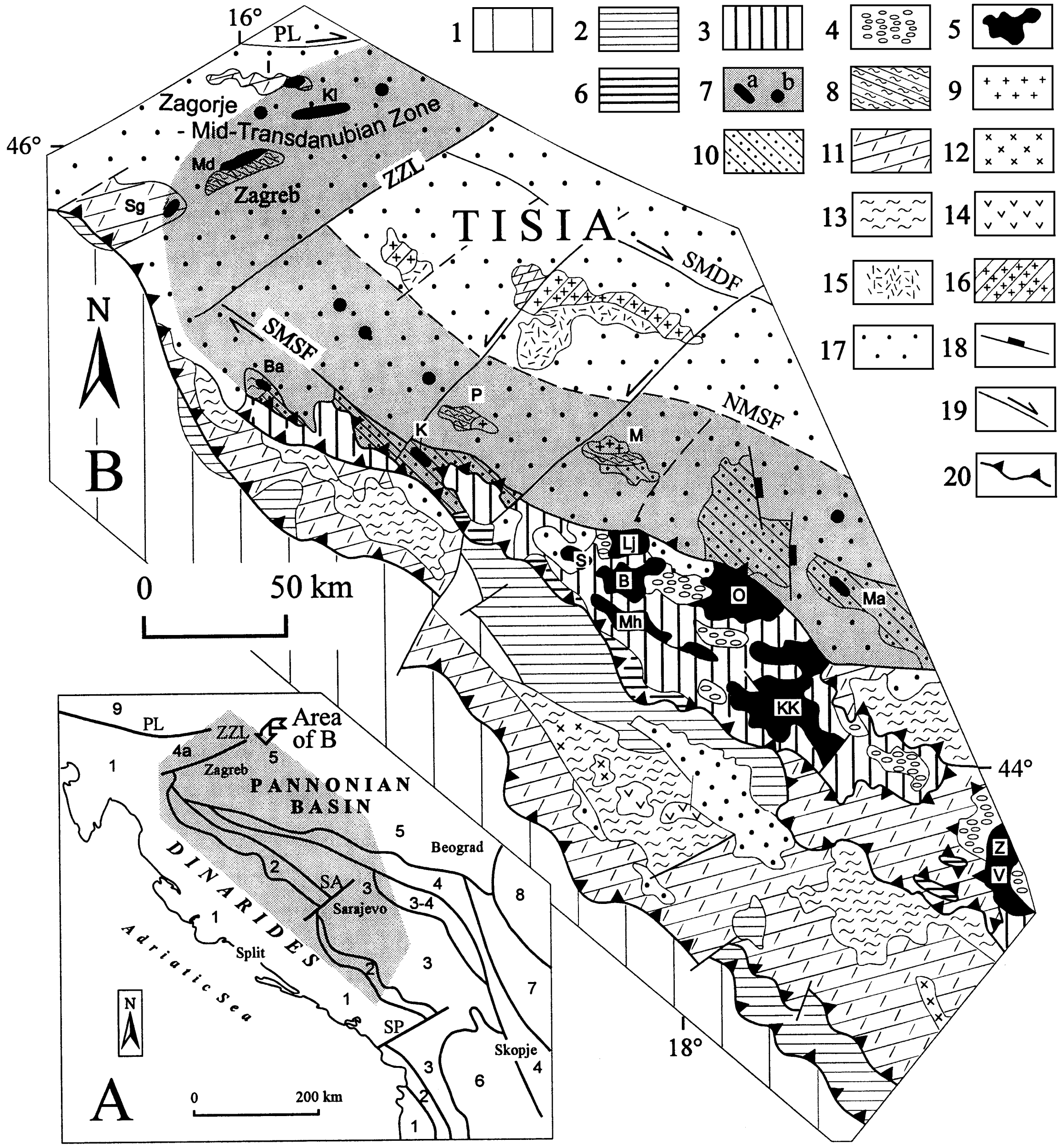
1. Introduction

The Dinarides, which can be traced along strike for about 700 km, merge in the northwest with the Southern Alps and in the southwest with the Hellenides. Fold, thrust and imbricate structures strike NW–SE and verge to the SW over most part of the Dinarides (Fig. 1A).

The Dinarides include in their internal parts large masses of ophiolites that are associated with genetically related sedimentary formations. These are among the largest ophiolite complexes built into the western parts of the Alpine–Himalaya orogenic belt. The bibliography of the Dinaridic ophiolites is voluminous and most of the papers, addressing separate smaller areas in the central and northwestern Dinar-



ides, were published in local languages and journals and thus are not readily accessible to the international geological community. For a review, the reader is referred to Pamić and Majer (1977), Karamata et al. (1980), Pamić (1982), Lugović et al. (1991) and Trubelja et al. (1995). This overview paper is based on all available but selected analytical data, presents a new and original two-stage geodynamic interpretation of the Dinaridic ophiolites and evaluates geological and petrological similarities and dissimilarities of the Dinaride Ophiolite Zone (DOZ) and Vardar Zone (VZ) ophiolites.

2. Regional geological and tectonic setting

The central Dinarides, which developed during the Alpine evolution in the Dinaridic Tethys, are characterized by a regular zonal pattern in the spatial distribution of characteristic Mesozoic–Paleogene tectonostratigraphic units (Pamić et al., 1998). From southwest to northeast, that is, from the Adriatic microplate toward the Pannonian Basin and Tisia, respectively, the following tectonostratigraphic units, which originated in different environments, can be distinguished (Fig. 1A,B): (1) Adriatic–Dinaridic carbonate platform (ADCP) of the External Dinarides; (2) Carbonate-clastic sedimentary rocks of the Apulian passive margin, corresponding to the "flysch bosniaque" (Aubouin et al., 1970); (3) Ophiolites associated with genetically related sedimentary formations forming the Dinaride Ophiolite Zone; (4) Sedimentary, igneous and metamorphic units, including ophiolites of the North Tethyan active continental margin of the Vardar Zone sensu lato (Aubouin, 1974). The tectonostratigraphic units 2–4 define the Internal Dinarides, also referred to as the Supradinaric (Herak, 1986); (5) Nappes involving Paleozoic–Triassic rock sequences, that is, the Golija Zone of Rampnoux (1970), which are thrust onto the internal units of the eastern Dinarides; in their southwestern, frontal parts, these nappes directly overlie the northeastern margin of the External Dinarides. According to this geodynamic model, the VZ represents the most internal unit of the Dinarides, which originated in the Dinaridic Tethys.

The northwesternmost and southeasternmost parts of the Dinarides, close to Apulia and Moesia, do not exhibit the same regular structural pattern. In the area west of the Zagreb–Zemplin Line (Fig. 1A,B), structures deflect into generally E–W strike, controlled by the Periadriatic Line. In the easternmost Dinarides, adjacent to the Carpathians and Balkan, the main structures deflect to a NNW–SSE strike (Fig. 1A). Here, the internal Dinaridic units are mainly covered by Paleozoic–Triassic nappes as in the NW Dinarides adjacent to the Alps.

In the central and northwestern Dinarides, ophiolites and associated sedimentary formations occur both in the DOZ and VZ. However, the ophiolites of these two zones differ in their origin, structure and ages, as well as in their relation to associated syngenetic sedimentary formations and overstepping sedimentary sequences. These two ophiolite zones continue without a break southeastward into the Hellenides, whereas they die out northwestward toward the Periadriatic Lineament area (Fig. 1B; Pamić, 1999).

Fig. 1. Major tectonic units of the Dinarides. (A) Index map showing major tectonic units of the Dinarides and surrounding regions (simplified after Aubouin, 1974). Legend: 1 External Dinarides and Southern Alps; 2 Carbonate platform margin formations; 3 Dinaride Ophiolite and Mirdita Zones; 3–4 Golija Zone; 4 Vardar Zone; 4a Zagorje–Mid-Transdanubian Zone; 5 Pannonian Basin; 6 Pelagonide; 7 Serbo-Macedonian Massif; 8 Carpathians and Balkan; 9 Eastern Alps; Major faults: PL Periadriatic Lineament; ZZL Zagreb–Zemplin Line; SA Sarajevo Line; SP Scutari-Peć Line. (B) Geological map of the central and northwestern Dinarides (modified after Pamić et al., 1998) Legend: 1 Adriatic-Dinaridic carbonate platform; 2 Jurassic–Cretaceous passive continental margin sequences ("flysch bosniaque" and "zone prekarstique"; Aubouin et al. (1970); 3 Dinaride Ophiolite Zone; 4 Late Jurassic–Cretaceous sequences unconformably overlying ophiolites; 5 large ultramafic massifs; 6 Radiolarite Formation; 7 Vardar Zone with ophiolites exposed on surface (a) and in deep wells (b); 8 Late Paleogene synkinematic metamorphic rocks; 9 Late Paleogene synkinematic granitoids; 10 Late Cretaceous–Early Paleogene flysch; *Paleozoic–Triassic Nappes*: 11 allochthonous Triassic sequences; 12 Triassic igneous rocks; 13 allochthonous Paleozoic sequences; 14 Paleozoic volcanics; *Tisia*: 15 Hercynian progressively metamorphosed sequences; 16 Hercynian granitoids and migmatites; 17 Neogene overstep sequences; 18 fault with downthrown block; 19 strike-slip fault; 20 major thrust. Large faults: PL Periadriatic Lineament; ZZL Zagreb–Zemplin Line; SMDF Southern marginal Drava fault; NMSF Northern marginal Sava fault; SMSF Southern marginal Sava fault. Mountains: B Borje, Ba Banovina, I Ivanščica, K Kozara, Kl Kalnik, KK Krivaja–Konjuh, Lj Ljubić, M Motajica, Ma Majevica, Md Medvednica, Mh Mahnjača, O Ozren, P Prosara, S Skatavica, Sg Samoborsko gorje, V Varda, Z Zlatibor.

3. Ophiolites

3.1. Basic geological data

3.1.1. Geology of the Dinaride Ophiolite Zone ophiolites

These ophiolites are associated with an ophiolite mélange and the Radiolarite Formation (Dimitrijević and Dimitrijević, 1973; Pamić, 1982). The mélange is composed of a shaly–silty matrix, embedding the fragments of graywacke, ultramafics, gabbro, diabase, basalt, tuff, amphibolite, chert, schist, and exotic limestone blocks of different ages, originating in different environments. The youngest so far found limestone exotic is Tithonian in age. In some areas, the original interlayering of shales and graywackes with occasional slumps and basalt flows is preserved. For convenience, we shall call it the Dinaride Ophiolite Zone mélange (DOZM). In some areas, particularly those with abundant limestone exotics, the DOZM is an olistostrome.

The petrography of individual ophiolite fragments varies. Small fragments, up to a kilometer in size, are composed only of one rock-type, for example, basalt or diabase or gabbro or peridotite. Larger blocks are also commonly composed only of one rock type, but in some areas, there are larger mappable fragments, on which parts of the normal undisturbed ophiolite sequence are preserved, for example, tectonic peridotites and cumulate gabbro–peridotites or sheeted diabase overlain by pillow lavas.

Ultramafic rocks predominantly occur as small, centimeter- to decimeter-sized fragments or hectometer- to kilometer-sized bodies included in the mélange, commonly completely serpentinized and cataclasized, and as large massifs (100–1000 km^2) corresponding to thrust sheets that overlay the mélange. The thickness of the ultramafic thrust sheets varies from a few hundreds up to 2000 m, as indicated by geophysical prospecting data (Roksandić, 1971).

Gabbros and diabase–dolerites form small bodies that in some places intrude larger ultramafic massifs (e.g. in the Krivaja–Konjuh massif, Fig. 2). Diabase–dolerites occur mainly as massive bodies, some of them sheeted. In preserved ophiolite sections, the combined thickness of diabase–dolerite and gabbro masses is estimated to reach some 1000–2000 m each.

Larger ultramafic bodies, such as the 2000-m-thick Krivaja–Konjuh massif (Fig. 2), form concave sheets, that are separated into several blocks. Each block is characterized by structural homogeneity as shown by the strike and dip of foliation: the largest, eastern block represents a NE–SW-trending synform. This ultramafic body is conformably underlain by a metamorphic sole, composed of a variety of amphibolites. In the northwestern part, the massif is intruded by gabbro (Pamić et al., 1977). Some smaller ultramafic massifs (e.g. Ozren) display more complex internal structures (Pamić, 1964).

In shales and graywackes, no index-fossils have been found yet. Platy limestones, rarely interlayered with shales, contain stratigraphically uncharacteristic Jurassic microfossils. The youngest limestone exotics of Tithonian age, however, document the upper boundary of the DOZM (Pamić, 1982).

Radiometric measurements were carried out on spinel lherzolites from larger ultramafic massifs and on rocks from metamorphic soles (Table 1). K–Ar ages obtained on hornblende from the Krivaja–Konjuh and Zlatibor amphibolite metamorphic soles range from 174 ± 14 to 157 ± 4 Ma (Doggerian; Lanphere et al., 1975). From the metamorphic sole of the Brezovica ultramafic body, positioned in the southeasternmost DOZ adjoining the northernmost Mirdita Zone of the North Hellenides, whole-rock isotopic ages range from 192 ± 8 to 161 ± 6 Ma on whole-rock metasediments and from 168 ± 5 to 159 ± 5 Ma on muscovite concentrates. K–Ar ages of 173 ± 7–149 ± 3 Ma were obtained on whole-rock amphibolites and 176 ± 6–161 ± 6 Ma on hornblende concentrates (Karamata and Lovrić, 1978). Four out of five spinel lherzolites from the Mts. Ljubić–Borje–Konjuh ultramafic massifs yielded an apparent Sm–Nd isochron age of 136 ± 15 Ma (Valanginian; Lugović et al., 1991).

In conclusion, radiometric ages and the fact that the youngest limestone exotics found in the mélange are of Tithonian age, indicate Early Jurassic to Early Cretaceous age of the DOZ ophiolites.

Scarce, few-hundred meter-sized fragments of plagiogranites are included in the DOZ mélange and occur also as dykes along fault zones of some larger ultramafic bodies (e.g. Mts. Ozren and Borje; Pamić and Tojerkauf, 1970). These granitoids have not been dated radiometrically yet.

The DOZM including the large thrust sheets of ultramafic rocks are disconformably overlain by some 1000–2000-m-thick, bed-to-bed Late Jurassic/Early

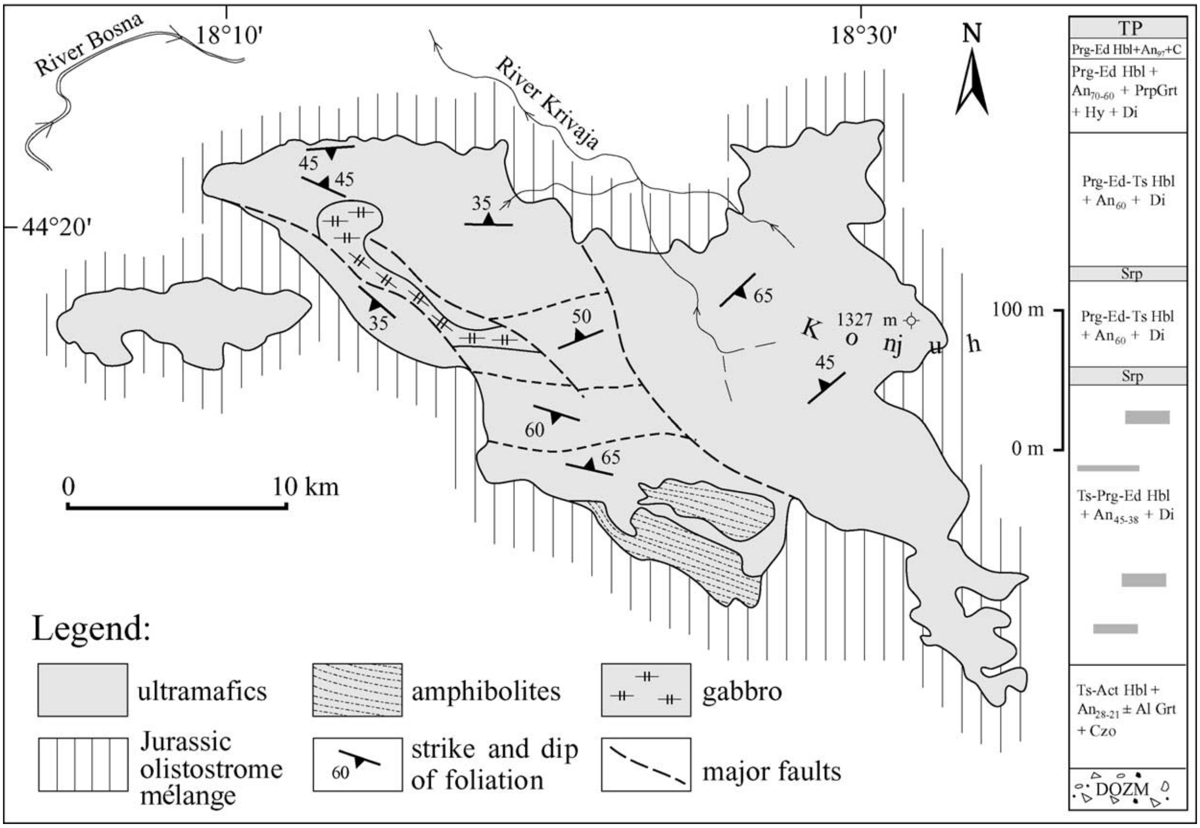


Fig. 2. Simplified geological map of the Krivaja–Konjuh ultramafic massif with geological column for amphibolite metamorphic sole (Pamić et al., 1977). Geological column legend: Act actinolite; Alm almandine; An anorthite; C corundum; Czo clinozoisite; Di diopside; DOZM Dinaride ophiolite mélange; Ed edenite; Grt garnet; Hbl hornblende; Hy hypersthene; Prg pargasite; Prp pyrope; Srp antigorite serpentine; Ts tschermakite; TP tectonic peridotites.

Cretaceous to Late Cretaceous Urgon-type (?) sequences. These sequences are mainly composed of unsorted shallow-marine conglomerates and breccias (the Pogari Series; Jovanović, 1957) and grade into lithic sandstones with subordinate shales and fossiliferous reefal limestones. They occur within kilometer- to tens of kilometer-long synclines. Clasts in these sediments include ophiolites as well as abundant coarse-grained, red granitoids of presumably Variscan age and uncertain derivation because they do not outcrop in the Dinarides (Varićak, 1955). Other synclines, in which mainly Upper Cretaceous limestones are preserved, contain bauxite and Ni-bearing iron ores at their base, representing a redeposited crust of weathered ophiolites (Gušić and Slišković, 1976; Pamić et al., 1998).

The Radiolarite Formation forms a narrow discontinuous zone, a few hundred kilometers long, along the southwestern margin of the DOZ (Fig. 1B). This is a bed-to-bed sequence, up to 800 m thick, which was deposited in an open-oceanic environment. The sequence is characterized by predominant radiolarites interlayered with shales, micrites and rare basalt flows. The radiolarite sequence covers a large stratigraphic interval, spanning Middle/Late Triassic to Early Cretaceous times (Pamić, 1982). To the southeast, the radiolarites crop out in tectonic windows beneath the Paleozoic–Triassic nappe of the western parts of the Golija Zone (Fig. 1B).

Jurassic mélange of the DOZ continues southwestward into the Mirdita zone, whereas the Radiolarite Formation correlates with the Budva Zone and Krasta–Cukali–Subpelagonian zones of the Hellenides (Aubouin et al., 1970; Papanikolau, 1984; Robertson and Shallo, 2000).

Table 1
Radiometric ages of Alpine ophiolites from the central and NW Dinarides

Locality	Petrography	Dated minerals/rocks	Method	Age (Ma ± δ)	References
Dinaride Ophiolite Zone					
Krivaja–Konjuh	Corundum-plagioclase-pargasite-spinel amphibolite	Pargasite	K–Ar	157 ± 4	
Krivaja–Konjuh	Corundum-plagioclase-pargasite amphibolite	Pargasite	K–Ar	170 ± 11	Lanphere et al. (1975)
Zlatibor	Garnet–plagioclase amphibolite	Amphibole-1	K–Ar	174 ± 14	
		Amphibole-2	K–Ar	168 ± 8	
Ljubić-Borje-Krivaja-Konjuh	Spinel lherzolite (n = 5)	Whole rocks Clinopyroxene	Sm–Nd	136 ± 15	Lugović et al. (1991)
Brezovica, southeasternmost DOZ adjoining the North Mirdita Zone	Mica schist and metasandstone	Whole rock	K–Ar	192 ± 8–161 ± 6	Karamata and Lovrić (1978)
		Muscovite	K–Ar	168 ± 5–159 ± 5	
	Amphibolite	Whole rock	K–Ar	173 ± 7–149 ± 3	
		Hornblende	K–Ar	176 ± 6–161 ± 6	
Vardar Zone					
Banovina	Bimineral hornblende–plagioclase amphibolite	Hornblende	K–Ar	166 ± 10	Majer and Lugović (1985)
	Metadiabase	Whole rock	K–Ar	160 ± 10	
Mt. Kalnik	Hornblende-augite-diabase-dolerite	Hornblende	K–Ar	189 ± 6.7	Pamić (1997a)
	Hornblende-augite-dolerite-ophitic gabbro	Hornblende	K–Ar	185 ± 6	
	Ophitic basalt	Whole rock	K–Ar	86.8 ± 8	
Mt. Medvednica	Ophitic basalt	Whole rock	K–Ar	85.4 ± 3.5	Pamić (1997a,b)
	Amphibole dolerite	Hornblende	K–Ar	94.3 ± 4.5	
Borehole cores from the South Pannonian Basin (Sava Depression)	Metadolerite	Whole rock	K–Ar	66.8 ± 2.9	Pamić (1997a,b)
	Gabbro	Whole rock	K–Ar	80.3 ± 3.4	
	Hornblende diabase	Hornblende	K–Ar	109.6 ± 6.6	
	Hornblende basalt	Whole rock	K–Ar	63.7 ± 2	
	Slightly altered diabase	Whole rock	K–Ar	62.2 ± 2.5	
Mt. Prosara	Alkali–feldspar granite	Whole rock	Rb–Sr	71.2 ± 2.5	
Mt. Požeška Gora	Bimodal basalt–rhyolite, ignimbrite	Whole rock	K–Ar	72.8 ± 2.1–62.1 ± 1.8	Pamić et al. (2000) and Pamić and Lanphere (1991)

3.1.2. Geology of the Vardar Zone ophiolites

The VZ ophiolites are also included in an ophiolite mélange (VZM). The VZ was first defined by Kossmat (1924) at the locus typicus, for example, Macedonia and South Serbia. Subsequently, it was extended to the north via Belgrade and with the WNW deflection along the northwesternmost Dinarides adjoining the South Pannonian Basin up to the area south of Zagreb (Aubouin et al., 1970). Most recently, Pamić and Tomljenović (1998) proposed its continuation into the Zagorje–Mid-Transdanubian Zone (4a in Fig. 1A) composed of mixed Alpine and Dinaridic lithologies.

The VZ of the northwestern and central Dinarides is composed mainly of fossiliferous Cretaceous–Early Paleogene flysch and ophiolite mélange. The flysch is composed of conglomerate, arenite, siltstone, sandstone, shale, marly-shale and limestone (Jelaska, 1978; Dimitrijević, 1995). In some places, the flysch sediments include rare blueschist blocks. The Senonian parts of flysch sequences are interlayered in some places with subduction-related basalts and rhyolites with ignimbrites (K–Ar ages = 72.8 ± 2.1–62.1 ± 2.5 Ma), which are intruded by penecontemporaneous alkali–feldspar granites (Rb/Sr isochron = 71.2 ± 2.5 Ma; Pamić et al., 2000—Table 1). In some parts of the

VZ, flysch formations are intruded by Eocene syncollisional granitoids (Rb/Sr isochrons = 56 and 48.7 Ma; Lanphere and Pamić, 1992) and Oligocene post-syncollisional granitoids (K/Ar ages mainly 34–30 Ma; Delaloye et al., 1989; Knežević et al., 1994). In the eastern parts of the VZ (not included on the map of Fig. 1), the granitoids are accompanied by large masses of penecontemporaneous andesites and dacites with latites (Karamata et al., 1992). Late Paleogene granitoids are more common in the subsurface than on the surface, as indicated by magnetic data (Vukašinović and Stefanović, 2000).

The Upper Cretaceous–Paleogene rocks laterally grade into regionally metamorphosed sequences, showing progressive zonation from unmetamorphosed sedimentary and igneous rocks to very low-, low- and medium-grade metamorphic rocks, as indicated by several lines of evidence (Pamić et al., 1992). The slates and phyllites of the metamorphic sequences yield a Late Cretaceous and Paleogene microflora (Pantić and Jovanović, 1976); K–Ar ages of 48 ± 8–38 ± 5 Ma were obtained on hornblende and biotite from medium-grade paragneisses (Lanphere and Pamić, 1992).

Generally, the ophiolite mélange of the VZ is almost the same in appearance as the mélange of the DOZ, that is, in a shaly–silty and rarely mylonitized serpentinite matrix ophiolite fragments and other rocks comprising limestone exotics are included, the youngest ones being of Late Cretaceous/Early Paleogene age (Pamić et al., 1998). However, the matrix of the VZM is strongly sheared and tectonized and ophiolites are much more dismembered. Hence, the DOZM is characterized mainly by olistostrome signatures, whereas the VZM represents tectonic mélange (Dimitrijević, 1995).

In matrix of the VZM, only a few Early Cretaceous (?) angiosperms have been found so far, preserved on bedding planes of graywackes and shales (Pamić, 1993).

K–Ar whole-rock and mineral measurements indicate two geochronological groups of ophiolites (Table 1). Most of isotopic ages obtained on diabase, dolerite and gabbro fragments and basalts interlayered with shales and graywackes range between 109.6 ± 6.6 and 62.2 ± 2.5 Ma, that is, spanning the interval between Albian and Danian. In the second geochronological group, K–Ar ages ranging between 189 ± 6.7 and 160 ± 10 Ma were obtained on fragments of dolerite, metadiabase and amphibolite also included in the VZM (Pamić, 1997a,b). The first group of radiometric ages fits with the age of youngest limestone blocks, whereas the second group of ages indicates that in some parts of the VZM ophiolite, fragments from the pristine DOZM are preserved.

Fragments of plagiogranites are also included in the VZM; however, they are here much more subordinate than in the DOZM (Majer, 1993).

3.2. Petrography

Ophiolites, represented by ultramafics, gabbro and diabase–basalt, are commonly dismembered; the degree of dismembering is much higher in the VZ than in the DOZ. In the Mt. Varda area of the DOZ, an undisturbed fragment of oceanic crust, about 4–5 km thick, is preserved (Fig. 3). It includes a complete sequence from peridotite tectonites through ultramafic cumulates, gabbro cumulates, sheeted ophitic gabbro–dolerite–diabases to basaltic lavas capped by sedimentary rocks (Pamić and Desmons, 1989). Unfortunately, a petrological–geochemical comparison of the ophiolites from the DOZ and VZ has not yet been carried out despite their distinct geological differences. For that reason, we tried to analyze petrological–geochemical similarities and dissimilarities of the DOZ and VZ ophiolites which will be discussed and presented in the following text.

3.2.1. Peridotite tectonites

In the central and northwestern Dinarides, lherzolites distinctly predominate over harzburgites and they are accompanied by subordinate dunites and quite sparse pyroxenites. Maksimović (1975), Maksimović and Majer (1981) and Maksimović and Kolomejceva-Jovanović (1987) recognized two peridotite zones within the Internal Dinarides, namely an eastern harzburgite zone, and a western lherzolite zone. According to this idea, geographically and geotectonically, the lherzolite subprovince is included in the DOZ, whereas the harzburgite subprovince is included in the VZ. However, in the Bosnian part of the DOZ, a lateral gradation from lherzolites to harzburgites has been discovered within the lherzolite zone, as documented by detailed examinations of systematically sampled ultramafics (Pamić, 1983). Namely, in the

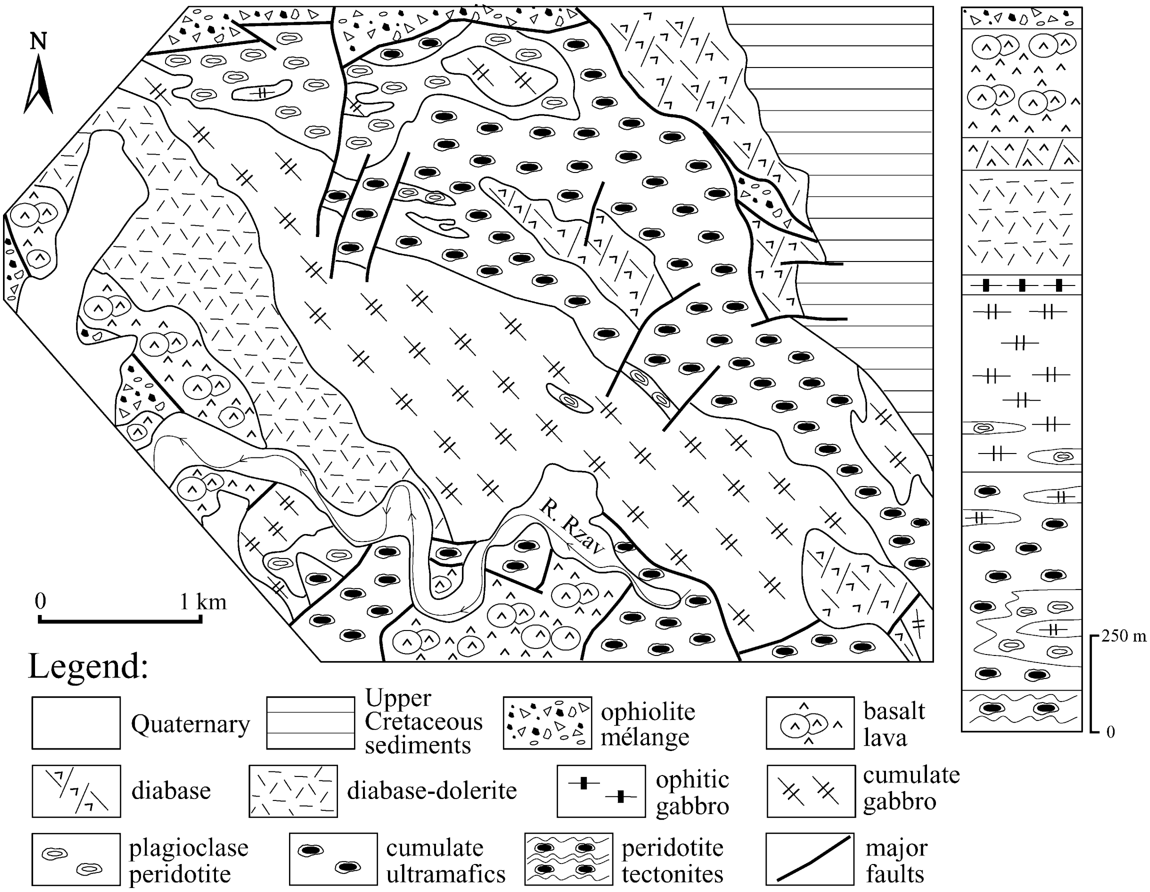


Fig. 3. Simplified geological map and geological column of the Mt. Varda oceanic crust fragment (after Pamić and Desmons, 1989).

northwestern and central parts of the DOZ, all ultramafic massifs are composed only of lherzolites, as already recognized by Kišpatić (1897). The southeasternmost outcrops of the lherzolites are found in the Varda massif (Fig. 1B) while in the neighboring ultramafic Zlatibor massif, harzburgites slightly predominate over lherzolites and transitional clinopyroxene-bearing harzburgite (Popević and Karamata, 1996). Further southeastward along the DOZ, ultramafic massifs are composed of harzburgites, although in its extension into the Mirdita Zone in Albania, some ultramafic massifs, despite the predominance of harzburgites, also contain lherzolites (Shallo, 1994). Also in the VZ, lherzolites prevail in its western and central parts, but the southeastern parts are characterized by a predominance of harzburgites. This difference could reflect the fact that in the western parts of both zones, the predominant lherzolites were derived from continental mantle–lithosphere according to Pamić (1983), whereas the predominant harzburgites of their southeastern parts were derived from oceanic upper mantle.

Major rock-forming minerals of both the DOZ and the VZ peridotite tectonites are characterized by chemical homogeneity. Their average compositions are as follows: the most abundant olivine is $Fo_{89.9}$ in lherzolites and $Fo_{91.1}$ in harzburgites, enstatite is $Fs_{10}En_{87.8}Wo_{2.2}$ in lherzolites and $Fs_{8.9}En_{89.1}Wo_{2.0}$ in harzburgites, clinopyroxene is $Fs_{4.9}En_{49.4}Wo_{45.7}$ in lherzolites and $Fs_{3.3}En_{50.0}Wo_{46.7}$ in harzburgites (Maksimović and Kolomejceva-Jovanović, 1987). These slight differences in chemical composition of major rock-forming minerals are due either to the fertile character of the lherzolites and the depleted character of the harzburgites or to analytical errors. Spinels are represented mainly by chromium spinel in lherzolites and alumochromite in harzburgites.

The predominant lherzolite tectonites from both the DOZ and the VZ have a typical metamorphic fabric that was already recognized and described in detail by Kišpatić (1897). These rocks are mainly porphyroblastic in texture. Subordinate granoblastic varieties are homogeneous in structure, whereas the porphyroblastic, particularly lherzolite varieties, are commonly foliated (Pamić and Majer, 1977). Subordinate harzburgites are mainly coarse-grained granoblastic in texture and homogeneous in structure.

The primary rock-forming minerals are transformed to various degrees: olivine into lizardite, clinochrysotile and antigorite, orthopyroxene mainly into bastite and talc, clinopyroxene into uralite and chlorite and spinel into magnetite. In some places, these secondary minerals are accompanied by quartz, chalcedony, opal, magnesite, breunnerite and dolomite. Along larger fault zones, particularly those located along the marginal parts of larger ultramafic bodies, comparatively large masses of quartz–carbonate rocks (listvenites) were generated (Sijarić and Šćavničar, 1972; Pamić and Olujić, 1974).

3.2.2. Cumulate ultramafics

As distinguished from peridotite tectonites, the cumulate peridotites form smaller bodies found either in partially or completely preserved ophiolite sections, or, more commonly, they occur as smaller dismembered fragments in the ophiolite mélange. The complete thickness of the cumulate peridotites varies from 500 to 1000 m.

These rocks, as distinguished from the peridotite tectonites, have typical magmatic anhedral-granular textures and homogeneous structures. The cumulate ultramafic rocks are represented by dunite, olivine-enriched harzburgite, lherzolite and wehrlite, grading into plagioclase wehrlite and plagioclase dunite, with rare lherzolite and pyroxenite ± plagioclase (Pamić et al., 1977; Lugović, 1986; Pamić and Desmons, 1989).

The cumulate ultramafics contain the same mineral assemblage as the peridotite tectonites. The most common cumulus phase is olivine, the rounded shape of which is due to magmatic resorption. Orthopyroxene occurs as single grains or rarely lamellae. The clinopyroxene is clearly an intercumulus phase which occurs either as reaction rims around olivine and orthopyroxene, or, more frequently, as interstitial anhedral grains, fresh or altered to uralite and chlorite. Accessory interstitial plagioclase is largely replaced by hydrogarnet (Pamić and Desmons, 1989).

As compared with the peridotite tectonites, the chemical composition of rock-forming minerals, as exemplified by the Varda area cumulates, is as follows: olivine is Fe-richer (Fo_{86}), orthopyroxene is slightly Fe-richer (En_{85}), clinopyroxene contains Wo_{31}, whereas among accessories, magnetitized spinel predominates over pyrite (Trubelja, 1960; Pamić and Desmons, 1989).

3.2.3. Cumulate gabbros

On undisturbed normal sections, the cumulate ultramafics are conformably overlain by cumulate gabbro zones, which are also interstratified with a variety of the cumulate ultramafics. The gabbro masses are composed largely of olivine gabbro and troctolite with subordinate normal gabbro and gabbro–norite, all of them showing a great variation from highly melanocratic varieties (grading into plagioclase peridotite) to extremely leucocratic varieties (grading into anorthosite). An example is given in Fig. 4 from the Krivaja–Konjuh massif; their maximum thickness is about 1000 m.

Despite their comparatively simple mineral compositions (plagioclase, olivine and pyroxene), the cumulate gabbros show great variations, both in structure and texture. The texture is hypeuhedral- to anhedral-granular, most commonly medium- to coarse-grained. The structure is commonly homogeneous but many outcrops also show a distinct banding that parallels layers of alternately different modal and mineral composition of different grain size. The parallel structure can also be displayed within individual gabbro layers by the foliation and lineation of rock-forming minerals. It is probable that much of the layering was secondary imposed during deformation accompanying sea-floor spreading away from the MORB axis (Sinton and Detrick, 1992).

The cumulate gabbros originated by in situ gravitational crystallization from a primitive basic magma as demonstrated in detail in the Krivaja–Konjuh and Varda gabbro masses by Pamić et al. (1977) and Pamić and Desmons (1989). All three major minerals can crystallize, both as cumulus and intercumulus

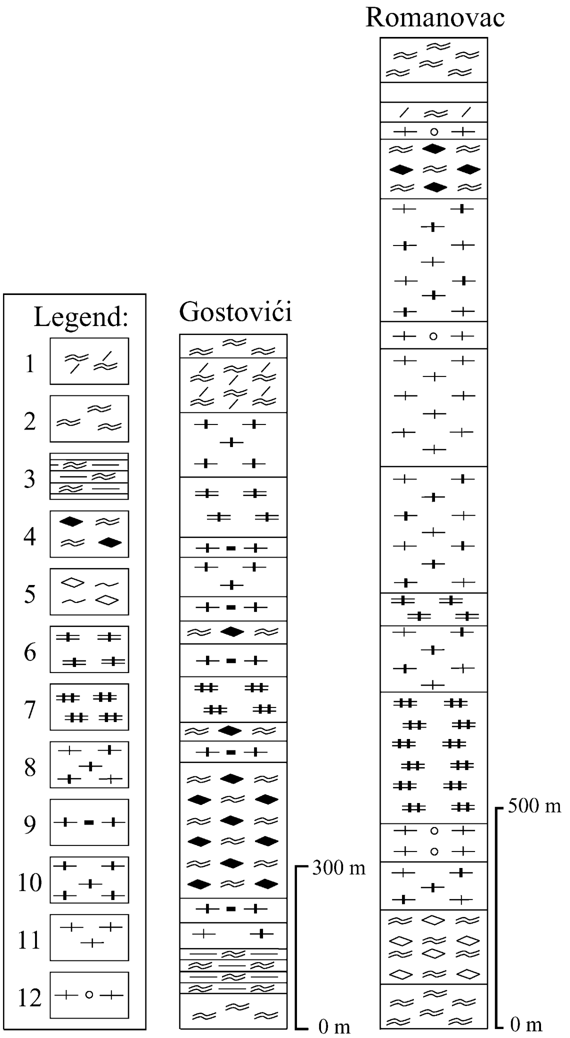


Fig. 4. Columnar sections in the Gostovići and Romanovac gabbro masses in the Krivaja–Konjuh ophiolitic complex (Pamić et al., 1977). Legend: 1 serpentinite; 2 lherzolite tectonite; 3 banded cumulate peridotite; 4 cumulate plagioclase wehrlite; 5 cumulate plagioclase dunite; 6 troctolite; 7 olivine gabbro; 8 uralite–olivine gabbro; 9 uralite–olivine gabbro, femic; 10 uralite–olivine gabbro, salic; 11 normal gabbro; 12 uralite gabbro.

phase, depending on the composition of those portions of the melt from which they crystallized. Plagioclase is a cumulus phase in leucocratic varieties and olivine in the melanocratic ones, and vice versa.

The chemical composition of the cumulate gabbro rock-forming minerals, as exemplified by the Varda area, is as follows: olivine is similar to that contained in the cumulate peridotites, clinopyroxene contains Wo_{42-43} and plagioclase is labradorite or bytownite (Pamić and Desmons, 1989).

3.2.4. Diabase–dolerites

They show mainly ophitic to subophitic textures with varying grain size: (1) the finest grained (0.05–0.5 mm) are referred as diabase varieties; (2) the medium-grained (1–2 mm) as dolerite varieties, and (3) the coarse-grained (1–6 mm, rarely to 10 mm) as ophitic gabbro varieties. All varieties have a massive structure. Compositional variations in diabase–dolerite bodies are presented in Fig. 5; their maximum thickness is about 1000 m.

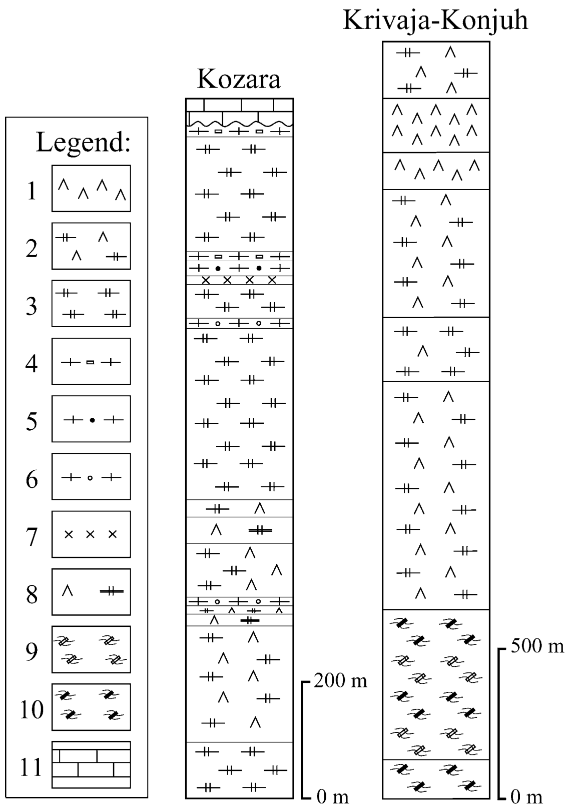


Fig. 5. Columnar sections in Kozara and Krivaja–Konjuh diabase masses (Pamić, 1996; Pamić et al., 1977). Legend: 1 diabase; 2 dolerite; 3 ophitic amphibole gabbro; 4 chloritized gabbro; 5 albitized gabbro; 6 saussurite gabbro; 7 amphibole norite; 8 amphibole dolerite; 9 amphibolitized diabase; 10 amphibolite; 11 Eocene sediments.

Mineral assemblages comprise plagioclase, clinopyroxene, amphibole and olivine, and accessory skeletal ilmenite, zircon and apatite. The plagioclase is commonly labradorite (e.g. An_{62-60} in the area of Varda) and rarely calcic andesine. Clinopyroxene is augite ($En_{55}Fs_{14}Wo_{29}$), amphibole is commonly actinolite, rarely primary hornblende, and olivine is Fo_{70} and thus more enriched in Fe than in the underlying cumulate gabbros (Pamić and Desmons, 1989; Lugović et al., 1991).

Dolerite–diabase rocks are slightly to moderately affected by hydrothermal metamorphism, as shown in albitization, prehnitization and calcitization of plagioclase, uralitization of augite, chloritization of amphibole and augite, and serpentinization of olivine. The dolerite–diabase bodies are cross-cut by veins and veinlets filled by laumontite, stilbite, epidote, albite, chlorite, calcite, tremolite and quartz (Pamić et al., 1979).

3.2.5. Basalts (metabasalts)

On undisturbed normal sections, the basalts are conformably underlain by diabase–dolerites. However, more commonly, they occur as small and large fragments included in the mélange. Rarely, they occur as flows interlayered within undisturbed graywackes and shales and the Radiolarite Formation. The basalts occur as pillow lavas, brecciated pillow lavas and massive flows. The basalts are commonly fine-grained ophitic, frequently with divergent arrangement of minerals. The structure is massive commonly with moderate quantities of amygdales.

Mineral assemblages are essentially characterized by mineral pair feldspar + clinopyroxene (augite), which has higher abundance of Ti and Al than those from diabase–dolerites (Lugović et al., 1991). They are to various degrees altered to chlorite and/or epidote. Olivine is rare, commonly almost completely serpentinized. Feldspar is plagioclase, most commonly completely transformed into pure albite which is filled by fine secondary prehnite, calcite, chlorite (?) and analcite.

The groundmass of subordinate porphyritic metabasalt varieties consists mostly of smectite and chlorite in which are included plagioclase and augite phenocrysts. The amygdales and secondary veins, cross-cutting volcanic bodies, contain calcite, chlorite with subordinate epidote, quartz and zeolites (Pamić et al., 1979).

3.2.6. Non-ophiolitic associates

These rocks are represented by (a) plagiogranites included both in DOZM and VZM, and (b) bimodal basalt–rhyolites interlayered with Late Cretaceous/Early Paleogene flysch sequences of the VZ.

(a) Plagiogranites are extremely leucocratic and massive rocks composed largely of quartz, albite–oligoclase with rare muscovite, quite scarce biotite and accessory zircon. Some granitoid bodies are cross-cut by quartz veins, 10–15 m long and up to 2 m thick (Pamić and Tojerkauf, 1970).

(b) Rocks of bimodal basalt–rhyolite association, forming volcanic bodies with the surface area up to 30 km^2, are represented by approximately equal proportions of amygdaloidal ophitic basalts, mostly transformed into metabasalts (mainly spilites) and alkali–feldspar rhyolites. The basalts are interlayered with tuffs and agglomerates and cross-cut by veins and veinlets filled with analcite, laumontite, chlorite, pumpellyite (?) and prehnite, whereas the rhyolites are interlayered with ignimbrites (Pamić et al., 2000).

Rocks of the bimodal basalt–rhyolite volcanic association as well as the rocks of the surrounding Late Cretaceous/Early Paleogene flysch are intruded by dykes, a few meters thick, and rare stocks, up to 5–6 km^2, composed largely of extremely leucocratic alkali–feldspar granites with subordinate alkali–feldspar syenites and quite scarce diorites. The predominant granite is largely composed of quartz, albite, orthoclase and microperthite, subordinate oligoclase, rare annite-enriched greenish biotite, frequently transformed into bauerite and muscovite. Among accessories, zircon distinctly predominates over apatite, saegenite and tourmaline (Pamić and Lanphere, 1991).

3.3. Geochemistry, geothermometry and geobarometry

3.3.1. Ultramafic rocks

Predominant spinel lherzolites are characterized by rather fertile composition as indicated by low clinopyroxene content averaging 12 vol.% and average concentrations of $Al_2O_3 = 3.16$ wt.% (DOZ) and 3.43 wt.% (VZ), $CaO = 3.32$ wt.% (DOZ) and 3.41 wt.% (VZ), and $TiO_2 = 0.19$ wt.% (DOZ) and 0.12 wt.% (VZ)—Table 2 (ans. 1 and 3) which are correlative to primitive upper mantle composition (McDonough and Sun, 1995). Volatile-free MgO variation diagrams of

Table 2
Selected average major (wt.%) and trace element content (ppm) of ophiolites, metamorphic sole rocks and some associated non-ophiolitic igneous rocks from the Dinaridic Ophiolite Zone and Vardar Zone

	Ultramafic tectonite				Ultramafic – mafic cumulate				Diabase – dolerite		Basalt – spilite				Metamorphic sole				Non-ophiolitic associates			
	1— DOZ	2— DOZ	3— VZ	4— VZ	5— DOZ	6— DOZ	7— VZ	8— VZ	9— DOZ	10— VZ	11— DOZ	12— DOZ	13— VZ	14— VZ	15— DOZ	16— DOZ	17— DOZ	18— VZ	19— DOZVZ	20— VZ	21— VZ	22— VZ
	l	h	l	h	uc	mc	uc	mc	d	d	b	s	b	s	ac	ae	ad	ad	pg	bb	brh	ag
SiO_2	42.83	45.47	44.64	43.99	44.75	45.78	45.27	46.52	49.31	49.51	48.59	49.84	46.20	48.99	45.84	39.43 – 45.86	49.19	49.58	73.64	47.29	73.06	71.55
TiO_2	0.19	0.06	0.12	0.09	0.16	0.17	0.14	0.26	1.38	1.50	1.16	1.62	1.10	1.70	0.18	0.18 – 0.45	1.26	1.54	0.07	1.54	0.36	0.17
Al_2O_3	3.16	1.07	3.43	2.25	3.58	17.48	2.52	18.05	15.81	16.18	15.28	14.52	15.91	15.90	19.40	14.51 – 17.55	14.28	14.48	13.37	18.27	13.00	14.25
FeO *	9.15	9.63	8.47	9.95	8.55	5.77	7.98	5.97	9.69	8.84	9.28	9.79	10.61	14.23	4.69	4.66 – 16.10	10.07	10.72	2.24	9.90	2.64	3.34
MnO	0.11	0.15	0.14	0.08	0.11	0.12	0.11	0.10	0.19	0.24	0.16	0.16	0.12	0.15	0.09	0.06 – 0.22	0.16	0.18	0.01	0.12	0.02	0.01
MgO	40.82	43.10	39.52	42.46	32.14	12.59	31.54	12.18	8.26	7.64	8.59	6.67	7.20	5.13	11.00	7.42 – 14.73	7.90	7.70	0.78	6.72	0.51	0.37
CaO	3.32	0.43	3.41	1.07	5.12	11.90	6.51	11.54	9.76	9.23	10.71	9.03	12.14	6.40	14.89	11.65 – 18.04	10.59	10.41	1.34	9.04	0.64	0.54
Na_2O	0.35	0.08	0.26	0.10	0.43	2.02	0.17	2.28	3.41	3.30	3.16	4.99	2.18	5.35	1.59	0.74 – 3.27	2.89	3.29	5.43	3.11	4.34	4.59
K_2O	0.07	0.01	0.01	0.01	0.02	0.24	0.02	0.38	0.21	0.44	0.36	0.32	0.06	0.31	0.17	0.04 – 0.34	0.16	0.21	1.57	0.97	4.13	3.41
P_2O_5	0.01	0.00	nd	0.00	0.04	0.02	0.02	0.03	0.13	0.10	0.15	0.23	0.08	0.33	0.03	0.04 – 0.29	0.12	0.09	0.18	0.19	0.18	0.12
H_2O					4.90	3.97	5.48	2.74	2.15	2.25	2.07	3.68	3.66	2.22	1.98	0.57 – 1.84	2.95	2.50	1.20	2.30	1.04	1.50
Total	100.01	100.00	100.00	100.00	99.80	100.06	99.76	100.05	100.30	99.23	99.51	100.85	99.26	100.71	99.86		99.57	100.70	99.83	99.45	99.92	99.85
M-value	88.83	88.86	89.27	88.38	87.01	79.55	87.57	78.43	60.31	60.64	62.26	54.84	54.74	39.12	80.70		58.31	56.15	38.30	54.75	25.62	16.49
Ba	7	1	5		4	5	5	11	134	296	–	132	100	196	31		21		140	111	262	230
Sr	4	1	24		37	144	9	164	175	202	156	156	127	170	125		139		85	413	76	57
Nb	–	–	1		–	–	1	–	5	3	–	11	3	7	–		–		6	7	–	11
Zr	8	3	30		6	8	10	9	85	84	101	138	80	116	11		78		81	87	391	332
V	78	46	–		75	144	123	–	267	–	282	311	240	–	112		288		13	139	12	–
Cr	2426	2652	2940		1900	696	2447	330	303	162	333	154	117	150	934		265		8	169	12	–
Co	107	129	48		83	47	114	–	27	37	42	25	45	41	36		43		3	33	4	–
Ni	2198	2614	2265		1088	472	1080	262	117	65	83	68	107	59	295		113		4	36	6	–
Cu	42	12	28		–	82	< 10	–	51	58	44	27	5	45	46		44		1	74	4	–
Zn	55	62	59		46	24	63	–	59	22	64	114	31	88	29		65		10	135	4	–

Sc	15	11			22	15	–		41		–	42	–		29		42		1	20	5	4
La	0.004	0.005			0.3	0.3	0.54		4.6		3.7	1.9	3.4		0.58		3.65		–	13.7	29	25.8
Ce	0.058	0.036			0.9	0.58	1.5		10.23		12.5	8.53	13.7		1.32		10.98		–	33.0	55.3	43.8
Nd	0.263	0.108			0.54	0.34	1.32		9.67		7.72	10.72	3.12		1.20		8.09		–	20.75	26.0	21.6
Sm	0.188	0.072			0.22	0.11	0.57		3.14		3.03	4.02	1.14		0.69		3.24		–	5.88	5.7	5.73
Eu	0.086	0.033			0.09	0.10	0.20		1.11		1.04	0.93	3.66		0.20		1.10		–	1.68	0.8	0.8
Gd	0.381	0.144			0.22	0.11	0.78		2.33		3.13	3.94	0.95		0.41		2.94		–	5.48	6.2	4.62
Tb	–	–			0.04	0.04	0.16		0.66		0.48	0.42	0.59		0.28		0.58		–	0.91	0.9	0.82
Dy	0.578	0.221			0.43	0.15	0.14		4.65		3.44	5.56	1.11		0.72		4.52		–	6.50	6.0	–
Ho	–	–			0.08	0.06	0.25		0.82		0.83	1.23	3.59		0.24		0.95		–	0.93	1.1	–
Tm	–	–			0.04	0.02	0.14		0.40		0.33	0.53	0.32		0.08		0.43		–	0.55	0.62	0.60
Yb	0.402	0.172			0.29	0.14	0.73		2.81		2.30	3.61	3.19		0.40		3.17		–	4.10	4.56	3.83
Lu	–	–			0.03	0.02	0.18		0.43		0.33	0.30	0.46		0.08		0.46		–	0.60	0.66	0.58
Y	4	2	3		4	5	7		32	27	25	33	24		7		32		20	23	31	29
K/Ba	83.0	83.0	16.6		41.5	398.5	33.2	286.8	56.2	130.4		83.0	5.0	14.3	45.5		63.2		93.1	72.5	130.9	123.1
Ba/Sr	1.8	1.0	0.2		0.1	0.0	0.6	0.1	0.2	0.2		0.2	3.7	1.1	0.2		0.2		1.6	0.3	3.4	4.0
Ba/La	1750	200.0			13.3	16.7	9.3		67.4			16.8	29.4		53.4		5.8			8.1	9.0	8.9
La/Nb							0.5		0.1			0.2	1.1							2.0		2.3
La/Y					0.1	0.1	0.1				0.1	0.1	0.1		0.1		0.1			0.6	0.9	0.9
Zr/Y	2.0	1.5	10.0		1.5	1.6	1.4		2.7		4.0	4.2	3.3		1.6		2.4		4.1	3.8	12.6	11.4
Sr/Nd	15.2	9.3			68.5	423.5	6.8		18.1		20.2	14.6	8.7		104.2		17.2			19.9	2.9	2.6
Sr/Zr	0.5	0.3	0.8		6.2	18.0	0.9	18.2	2.1	1.7	1.5	1.1	0.3	1.5	11.4		1.8		1.0	4.7	0.2	0.2

M-value = 100 × molar MgO/(MgO + FeO_{tot}).

l lherzolite tectonite; h harzburgite tectonite; uc ultramafic cumulate; mc mafic cumulate; d diabase – dolerite; b basalt; ac amphibolite related to ultramafic – mafic cumulates; ae eclogite from cumulate-related amphibolites; ad amphibolite related to diabase – dolerite; pg plagiogranite; bb bimodal basalt; brh bimodal rhyolite; ag alkali – feldspar granite.

1 Krivaja – Konjuh lherzolite (*n* = 20); 2 Varda-Zlatibor harzburgite (*n* = 7); 3 Medvednica-Banovina-Ljubić lherzolite (*n* = 7); 4 Medvednica-Banovina-Ljubić harzburgite (*n* = 10); 5 Krivaja-Konjuh-Varda-Zlatibor ultramafic cumulate (*n* = 6); 6 Krivaja-Konjuh-Varda-Zlatibor cumulate (*n* = 15); 7 Medvednica ultramafic cumulate (*n* = 3); 8 Banovina-Kozara-Ljubić mafic cumulate (*n* = 7); 9 Borje-Krivaja-Konjuh-Varda diabase – dolerite (*n* = 12); 10 Banovina-Kozara-Ljubić diabase – dolerite (*n* = 14); 11 Krivaja – Konjuh – Varda basalt (*n* = 5); 12 Krivaja-Konjuh-Borje-Varda spilite (*n* = 9); 13 Medvednica-Banovina-Ljubić basalt (*n* = 3); 14 Medvednica-Banovina-Ljubić spilite (*n* = 24); 15 Krivaja – Konjuh amphibolite related to ultramafic – mafic cumulates (*n* = 24); 16 variation intervals for Krivaja-Konjuh-Borje-Skatavica eclogites associated with cumulate gabbro-related amphibolites (*n* = 8); 17 Krivaja – Konjuh amphibolite related to diabase – dolerites (*n* = 13); 18 Banovina-Ljubić amphibolite related to diabase – dolerites (*n* = 15); 19 Krivaja-Konjuh-Borje-Banovina plagiogranite (*n* = 5); 20 BARB-related basalts from the NW Vardar Zone (*n* = 4); 21 BARB-related bimodal rhyolite including ignimbrite from the NW Vardar Zone (*n* = 20); 22 BARB-related alkali – feldspar granite from the NW Vardar Zone (*n* = 5).

Data taken from Pamić et al. (1973, 1977, 2000), Pamić (1976, 1985, 1997a,b), Pamić and Majer (1977), Majer (1984), Pamić and Desmons (1989), Lugović et al. (1991), Pamić and Lanphere (1991), Majer (1993), Trubelja et al. (1995), Majer and Garašić (2001) and Slovenec (in preparation).

the spinel lherzolite tectonites show that Al_2O_3, CaO and TiO_2 decrease and Ni increases relative to MgO (Fig. 6A–D; Lugović et al., 1991).

There is a positive correlation of MgO vs. Ca/Al (weight ratio) in the Dinaridic lherzolites and samples at low MgO end have chondritic Ca/Al ratios thus documenting their primitive mantle composition. In the Ligurian ophiolites, Rampone et al. (1995) also documented the primordial character of predominant spinel lherzolites. In the spinel lherzolites from both the DOZ and VZ, LREEs predominate over HREEs (Fig. 7). The uniform increase from La to Yb, together with the strong steepness of the pattern in the La to Nd area provides strong evidence for the absence of any metasomatic re-enrichment subsequent to depletion. Such a REE pattern suggests that the lherzolites represent residua produced by fractional melting of very small increments rather than by batch melting (Lugović et al., 1991). The Dinaridic lherzolites could be probably of subcontinental origin as indicated by their low-Cr# of spinel ranging mainly between 0.11 and 0.14 and Na_2O content of clinopyroxene averaging 0.52 wt.% (Arai, 1994; Maksimović and Kolomejceva-Jovanović, 1987).

All these subcontinental mantle signatures probably represent the reason why the Dinaridic fertile spinel lherzolites show some distinct differences from the peridotites of typical ophiolite complexes. Namely, whereas the REE patterns of most ophiolite peridotites show evidence of metasomatic enrichment, the extremely depleted patterns of the Dinaridic spinel lherzolites indicate that they were not metasomatically enriched in LREE (Lugović et al., 1991).

Based on geobarometric and geothermometric calculations proposed by Roeder et al. (1979) and Mercier (1980), Maksimović and Kolomejceva-Jovanović (1987) obtained that lherzolite equilibrated at 1050–1210 °C for olivine–spinel geothermometer, and up to 1100–1220 °C and at about 17 kbar for enstatite geothermometer and geobarometer, respectively. Pamić and Desmons (1989) calculated that spinel lherzolite equilibrated at slightly higher temperatures of 1250 °C, using the clinopyroxene geothermometer. All these data for the Dinaridic spinel–facies assemblage record high (asthenosperic) equilibration temperatures (Rampone et al., 1996).

Subordinate harzburgites are characterized by rather depleted composition as indicated by average concentrations of $Al_2O_3 = 1.07$ wt.% (DOZ) and 2.25 wt.% (VZ)—Table 2 (ans. 2 and 4). Maksimović and Kolomejceva-Jovanović (1987) using enstatite geothermobarometer (Mercier, 1980) calculated that harz-

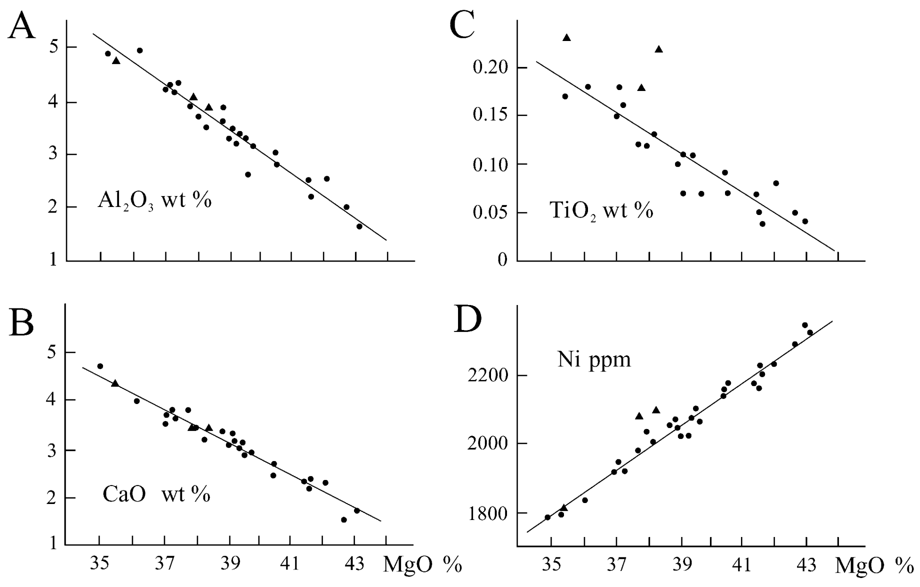


Fig. 6. Volatile-free MgO variation diagrams for Al_2O_3 (A), CaO (B), TiO_2 (C) in weight percentage and Ni (D) in parts per million (marked by full circles) for peridotite tectonites with primitive mantle estimates (data marked by full triangle from Lugović et al., 1991).

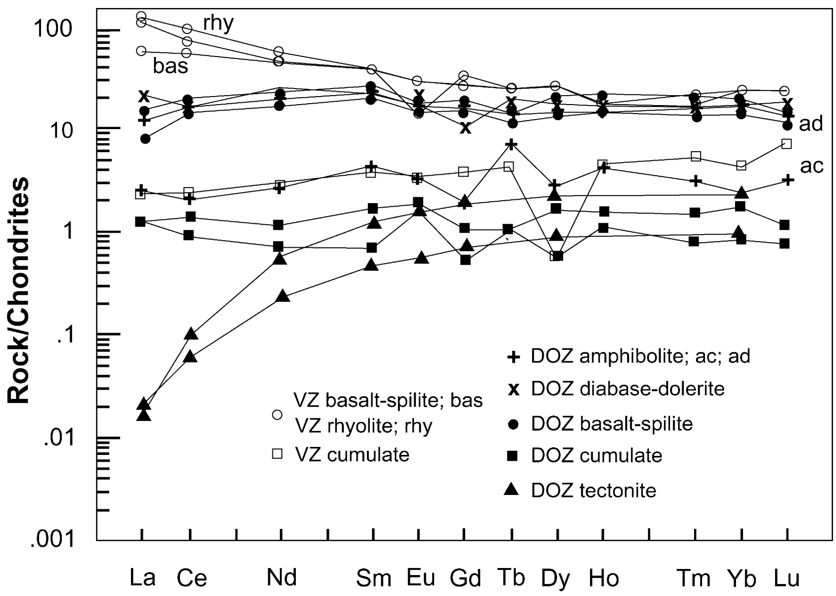


Fig. 7. Chondrite normalized REE pattern for samples from Dinaride Ophiolite Zone (full symbols) and Vardar Zone (empty symbols); ac amphibolite related to DOZ ultramafic–mafic cumulates; ad amphibolite related to DOZ diabase–dolerites; bimodal bas—basalt; rhy—rhyolite from the VZ. Normalization after Sun and McDonough (1989).

burgites from the Raška ultramafic body from the southeasternmost parts of DOZ (not presented in the map of Fig. 1) equilibrated at 975 °C and 14.5 kbar.

3.3.2. Cumulate rocks

These range in chemical composition from highly magnesian (12–34 wt.%) in melanocratic gabbros and magmatic (cumulate) peridotites to less magnesian (3–10 wt.%) in more evolved leucocratic gabbros whose extremely leucocratic varieties are characterized by high CaO (15–18 wt.%) and Al_2O_3 (17–30 wt.%) — Pamić et al. (1977); Pamić and Desmons (1989). Table 2 (ans. 5–8) demonstrates that average major element contents are the same both in the DOZ and VZ cumulates. Average MgO and FeO* contents in cumulate gabbros amounting to 12.59 and 12.18 wt.% vs. 5.77 and 5.97 wt.% fall within generally acceptable interval of primitive tholeiitic gabbro melts ranging between 10 and 18 wt.% (Sinton and Detrick, 1992).

The overall trend from Mg-rich to Ca- and Al_2O_3-rich varieties is governed by fractional crystallization and the formation of layered sequences composed of gabbro layers, ranging in thickness from a few centimeters to a few meters. For that reason, cumulate rocks do not provide unequivocal information on magma types. Based on CIPW norms, cumulate rocks from the Krivaja–Konjuh massif display mainly olivine tholeiite affinity, suggesting primitive magma types (Pamić et al., 1977).

This is also reflected by trace element abundances, which correlate well with the fractionation of main rock-forming minerals. Compatible trace elements are fairly equal, as exemplified by Ni (190–400 ppm) and Cr (270–330 ppm), whereas incompatible trace elements (Zr, Y, Ba, Rb, Nb) are rather low (Trubelja et al., 1995; see Table 2, ans. 5–8). Chondrite-normalized REE pattern is characterized by fairly equal concentrations of LREEs and HREEs with slight positive Eu anomaly (Fig. 7).

Based on TiO_2 vs. $FeO_{tot}/FeO_{tot}+MgO$ (wt.%) classification diagram proposed by Serri (1981), cumulate rocks from both the Krivaja–Konjuh and the Varda ultramafic massifs within the DOZ fit with cumulate rocks from Mid-Atlantic Ridge, suggesting that the DOZ cumulates might have originated at accretionary plate margin (Trubelja et al., 1995). By contrast, cumulate peridotites from Mt. Medvednica from the northwesternmost parts of the VZ contain abundant titaniferous edenite–pargasite suggesting evolved volatile enriched melts (Slovenec and Lugović, 2000). In this aspect, the Mt. Medvednica ultramafic cumulates are correlative with plutonic rocks from immature volcanic arcs (DeBari and Coleman, 1989; Slovenec and Lugović, 2000).

3.3.3. Mafic hypabyssal and extrusive rocks

As distinguished from cumulate rocks, these rocks, which do not show significant variations both in major and trace element contents, give much better information on magma types and their evolution. Dolerites display average *M*-values of 60 (DOZ) and 61 (VZ), lower average Ni abundance of 117 (DOZ) and 109 ppm (VZ), Cr averages of 303 (DOZ) and 272 ppm (VZ), much higher TiO_2 averages of 1.38 (DOZ) and 1.50 wt.% (VZ), and Na_2O averages of 3.41 (DOZ) and 3.30 wt.% (VZ). These data indicate that the diabase–dolerites are more evolved rocks than the underlying gabbro cumulates (Table 2, ans. 9–10).

Diabase–dolerites contain fairly equal concentrations of LREEs and HREEs and their REE pattern is essentially flat with slightly negative Eu anomaly (Fig. 7).

Despite the fact that basalts experienced strong to complete hydrothermal ocean-floor metamorphism, they display positive correlation in most major and trace element concentrations among themselves and with diabase–dolerites. Besides the most significant difference in Na_2O content (2.18–3.16 wt.% vs. 4.99–5.45 wt.%), there is exception in FeO* and MgO averages ranging from 8.84 to 9.69 wt.% vs. 7.64 to 8.26 wt.% in diabase–dolerites and from 9.19 to 14.23 wt.% vs. 5.13 to 7.20 wt.% in basalts and spilites (Table 2, ans. 9–14). The trace elements Zr, Y and Nb are lower in diabase–dolerites than in basalts and spilites.

The basalts and spilites from both the DOZ and VZ are hypersthene normative, TiO_2 averages range from 1.10 to 1.70 wt.%, MgO averages mainly between 5.13 and 7.20 wt.% with low K_2O (mainly about 0.30 wt.%) and P_2O_5 (mainly 0.08–0.15 wt.%) indicating their MORB signatures (Table 2, ans. 11–14). Basalts and spilites have almost the same REE pattern as diabase–dolerites (Fig. 7).

Trubelja et al. (1995), by using different geochemical discrimination diagrams, concluded that they have geochemical characteristics of N-MORB type (Fig. 8B). However, on the Ti/Cr vs. Ni diagram (Beccaluva et al., 1979), averages of DOZ and VZ basalts plot along the boundary line dividing ocean-floor tholeiite and island arc tholeiite (Fig. 8A).

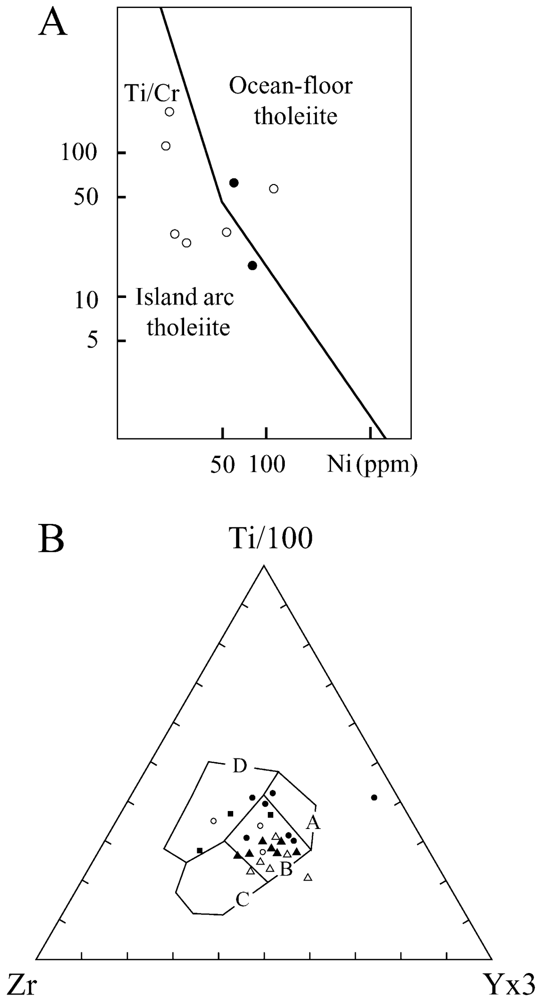


Fig. 8. (A) Ti/Cr vs. Ni diagram (Beccaluva et al., 1979). Vardar Zone basalts (empty circles); Dinaride Ophiolite Zone basalt and spilite averages (full circles); data from Pamić et al. (2000) and Table 2. (B) Zr–Ti–Y triangle for basalt–spilites (full circle—data from Trubelja et al., 1995; open circle—data from Lugović et al., 1991) and diabase–dolerites (full triangles—data from Trubelja et al., 1995; open triangles—data from Lugović et al., 1991; full squares—data from Pamić et al., 2000). D within plate basalts; B ocean-floor basalts; A and B island-arc basalts; B and C calc-alkaline basalts.

3.3.4. Non-ophiolitic associates

(a) Plagiogranites associated with ophiolites, commonly included in the M-type family (Coleman and

Peterman, 1975; Pitcher, 1987), represent a genetic puzzle within ophiolite complexes composed entirely of ultramafic and mafic rocks. The Dinaridic plagiogranites are extremely leucocratic rocks characterized by high SiO_2 (74 wt.%), increased quantities of Na_2O (5–6 wt.%) and low contents of FeO*, MgO and CaO. Incompatible trace elements predominate (c. 200 ppm) over the compatible ones (less than 30 ppm); Zr averages 81 ppm and Y is 20 ppm (Table 2, an. 19).

In some other ophiolite terranes, for example, Semail ophiolite complex, plagiogranites were explained by crystal fractionation of 70–80% of the mean diabase dyke liquids by separation of clinopyroxene + hornblende and plagioclase and late-stage segregation by filter pressing (Pallister and Hopson, 1981).

(b) Rocks of bimodal volcanic association, interlayered in Late Cretaceous/Early Paleogene flysch of the VZ, are represented by alkalic to subalkalic basalts and alkali–feldspar rhyolites with ignimbrites (Pamić et al., 2000). The fresh, more primitive basalts represent primary or slightly fractioned rocks, probably affected by olivine fractionation. They plot in the field of island arc tholeiites (Fig. 8A; Beccaluva et al., 1979) but on some other discrimination diagrams, basalts are scattered in different geotectonic fields. However, in some trace elements, particularly Ba, Sr, Cr, Ni, Nb, Zr and Y, these basalts can be correlated with BARB basalts (Saunders and Tarney, 1984).

Trace element contents of alkali–feldspar rhyolites are characterized by high Zr concentrations averaging 391 ppm which are similar to the Zr average of 332 ppm in the accompanied alkali–feldspar granites. In the Nb versus Y diagram (Pearce et al., 1984), both rhyolites and granites plot in the field of syncollisional and volcanic arc granitoids and thus can be correlated with a subduction environment (Pamić et al., 2000). In some characteristic element ratios, the alkali–feldspar rhyolites are very similar to ignimbrites associated with basalts from modern active continental margins and back-arc basins (Ikeda and Yuasa, 1989).

Late Cretaceous basalts and alkali–feldspar rhyolites are enriched in REE content (Table 2, ans. 20 and 21), while LREE predominates over HREE in both groups of rocks (Fig. 7). Unlike the basalts, the rhyolites have a distinct negative anomaly indicating either fractional crystallization of feldspars or partial melting of source rocks that contain residual plagioclase.

Recalculated initial $^{87}Sr/^{86}Sr$ ratios for basalts range between 0.70382 and 0.70432 and whole-rock $\delta^{18}O$ from 5.3 to 6.3‰ for fresh basalts, indicating upper mantle origin. $^{87}Sr/^{86}Sr$ ratio obtained from a Rb/Sr isochron for alkali–feldspar rhyolites and cogenetic alkali–feldspar granites is 0.7073 indicating an apparent continental crust origin. However, when these data are combined with the $\delta^{18}O$ ranging between 8.8 and 9.2‰, they fit with the Adamello tonalites, suggesting that the alkali–feldspar rhyolites and associated alkali–feldspar granites might have been fractionated from a primitive basalt melt (Pamić et al., 2000).

4. Metamorphic soles

Varieties of amphibolites and related crystalline rocks are so frequently associated with the Dinaridic ophiolites that one can speak of a "quaternity" instead of Steinmann's (1926) trinity (Pamić et al., 1973).

In the central and northwestern Dinarides, compositionally and genetically different groups of metamorphic sole rocks can be distinguished: (1) the amphibolites originated from cumulate rocks; (2) the amphibolites genetically related to diabase–dolerites; and (3) metasedimentary sequences ± amphibolites originated from pristine sedimentary rocks ± diadiabase–basalts.

4.1. Amphibolites originated from cumulate rocks

In tectonically less disturbed areas, the amphibolites build up the metamorphic sole of larger ultramafic massifs (Fig. 2). However, there are fragments (10 km^2) included in the DOZM in which amphibolites predominate over associated ultramafic rocks (Fig. 9). There are also smaller fragments completely dismembered and composed only of amphibolites. These amphibolites have not been found yet in the VZ.

Disregarding the size and mutual proportions, the contact of these amphibolite–ultramafic fragments are, as a rule, gradational and marked by fine alternating bands of amphibolites, garnet–plagioclase pyroxenites, garnet hornblendites and peridotites. Foliation, both in amphibolites and predominant spinel lherzo-

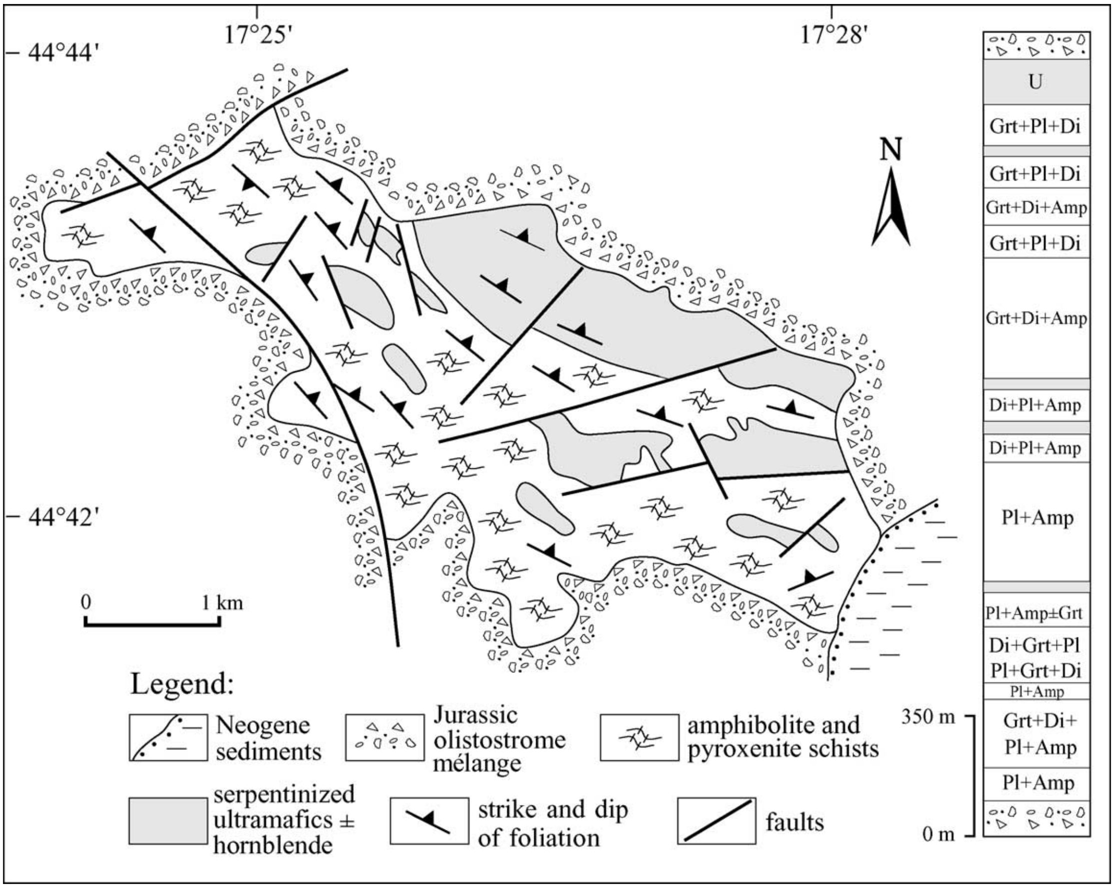


Fig. 9. Simplified geological map of the Skatavica ultramafic–amphibolite mass with the geological column (Pamić, 1972). Column legend: Amp amphibole; Di diopside; Grt garnet; Pl plagioclase; U ultramafic interlayers.

lites, is conform and their mutual contact is parallel to the strike of foliation. In some areas, the interlayered amphibolites are folded together with predominating spinel lherzolites (Popević and Pamić, 1973).

The amphibolites have granoblastic, porphyroblastic and nematoblastic textures of various grain size. The structure is most commonly shown in foliation, rarely in layering. The most important varieties of the amphibolites are: (1) bimineralic (plagioclase + amphibole) massive amphibolite and amphibolite schist, (2) diopside amphibolite schist, (3) garnet amphibolite schist, (4) garnet–diopside amphibolite schist, rarely with eclogite lenses up to a few meters long, (5) rare monomineralic amphibole schist, (6) garnet–plagioclase–pyroxenites, and (7) extremely rare corundum amphibolite (edenite–pargasite) schist. This large group of crystalline rocks hereafter will be called amphibolites for convenience.

Each of the single rocks mentioned above shows progressive cryptic variations in bulk composition, particularly amphibole and plagioclase and thus belong to different metamorphic facies. The amphiboles range from pargasite, edenite, and their isomorphic mixture through transitional varieties enriched in tschermakite to ‘‘common’’ hornblende (Fig. 10A,B). Only some of them contain up to 10% of glaucophane molecule. Plagioclase varies from anorthite to sodic oligoclase, and garnet also shows variations of the main end-members: pyrope 59–22%, almandine 49–28%, grossularite 23–16% and andradite 7–3% (Fig. 10C). By contrast, the chemical composition of clinopyroxene is fairly monotonous; in the highest grade

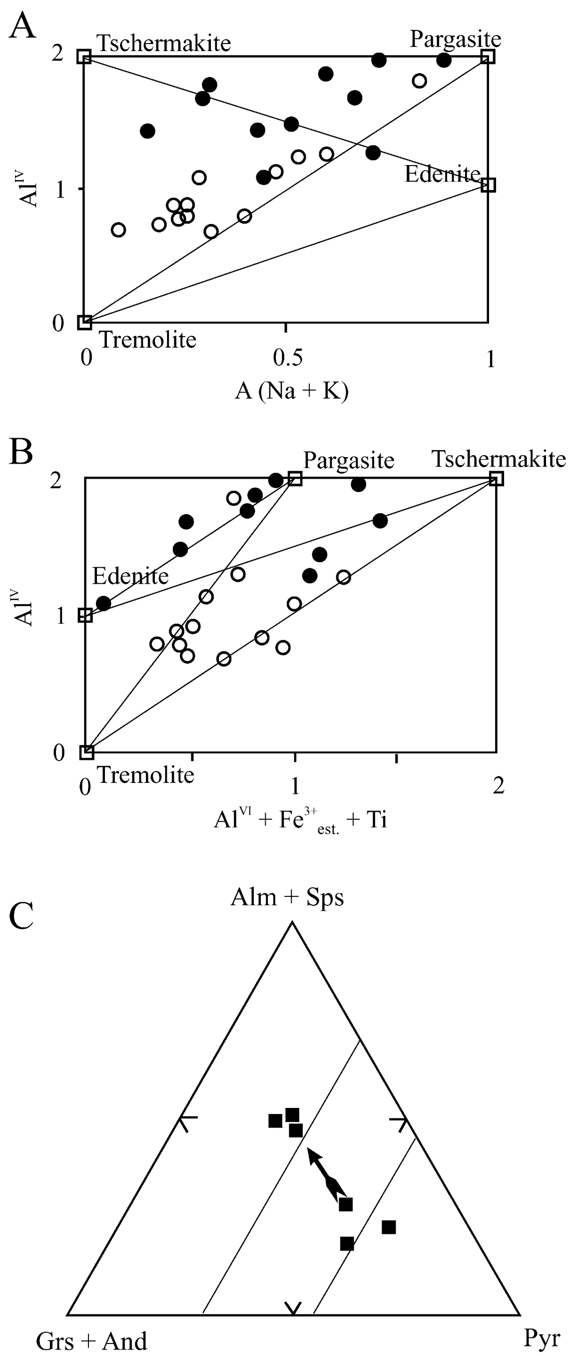


Fig. 10. Diagrams A and B showing isomorphic substitutions of amphiboles from Vardar Zone (empty circles) and Dinaride Ophiolite Zone (full circles). (C) Triangle diagram showing compositional variations of garnets from the Dinaride Ophiolite Zone based on mole percentage of garnet end members. Alm almandine; And andradite; Grs grossularite; Pyr pyrope; Sps spessartite.

amphibolites, it is diopside with admixtures of hedenbergite and jadeite ($Di_{62-56}Hd_{31-24}Jd_{20-10}$) as shown by Pamić et al. (1973). Subordinate minerals of the amphibolites are hypersthene, corundum and clinozoisite or epidote and biotite (?).

Mutual variation of the rock-forming minerals shows a regular pattern. Putting aside any rock terminology, the mineral assemblages of amphibolites can be grouped as follows (Pamić et al., 1973).

(1) Pargasite–edenite amphibole + jadeitic diopside (omphacite) + anorthite–bytownite ± garnet ± corun-corundum. Based on published mineral chemical analysis (Pamić et al., 1973; Pamić, 1976), we calculated *P–T* conditions for these highest grade sole amphibolites from the southwestern margin of the Zlatibor ultramafic massif using geothermometers and geobarometers (Graham and Powell, 1984; Powell, 1985; Plyusnina, 1986; Blundy and Holland, 1990; Holland and Blundy, 1994). Obtained temperature and pressure values range between 670 and 750 °C and 740 and 930 MPa, respectively, indicating granulite–amphibolite facies *P–T* conditions (Bucher and Frey, 1994). However, it can be presumed that *P–T* conditions of eclogite facies must have existed along direct contact with overlying lherzolites as indicated by occurrences of eclogites of variable chemical composition within progressively metamorphosed amphibolite sequence (Pamić, 1976).

The mineral assemblage edenite–pargasite + bytownite–anorthite is in equilibrium with the mineral assemblage of the associated predominant spinel lherzolite. These two mineral associations were experimentally produced at 800–900 °C and 10–20 kbar (Gilbert, 1969) and at 1015 °C and 15 kbar (Kushiro, 1970), respectively, that is, under *P–T* conditions of the upper mantle. On the other hand, most recent experiments (Niida and Green, 1999) demonstrated that pargasite is stable in MORB pyrolite. This interpretation is strongly supported by field data, as shown in fine-scale, millimeter to centimeter interlayering of pargasite–edenite amphibole schists and olivine + pyroxene bands, their conform structural relations and in common folding.

(2) Transitional greenish pargasite–edenite-rich hornblende + labradorite–calcic andesine ± diopside ± pyrope-rich garnet ± hypersthene. To illustrate this paragenesis, we calculated *P–T* conditions for the Skatavica amphibolite–ultramafic fragment body

using different geothermobarometers (Graham and Powell, 1984; Powell, 1985; Plyusnina, 1986) on available parageneses. We obtained average temperature of 670 ± 50 °C and average pressure of 720 ± 100 MPa indicating amphibolite–granulite transition (Bucher and Frey, 1994).

(3) Brownish tschermakite-rich hornblende ± andesine ± diopside ± almandine-rich (c. 30–45%) garnet. This paragenesis belongs to the amphibolite facies.

(4) "Normal" hornblende ± oligoclase ± clinozoi-clinozoisite ± almandine-rich (50–70%) garnet which belong to epidote–amphibolite facies.

Due to different degrees of dismembering, most commonly in amphibolite fragments included in mélange, only some parts of the mineral assemblages are preserved. Only in the progressive amphibolite metamorphic sole from the southern part of the Krivaja–Konjuh ultramafic massif is a complete metamorphic sequence preserved; it is tentatively presented in the geological column in Fig. 2 (Pamić et al., 1977).

In conclusion, all varieties of amphibolites including the most common bimineralic amphibolites and the rarest eclogites could crystallize under HT, MT and LT conditions (Bucher and Frey, 1994), as the result of very slow exhumation rate related to common ophiolite–amphibolite obduction.

The genetic relation between ultramafic–mafic cumulates and these metamorphic sole amphibolites is supported by several lines of evidence: (a) The most complete profiles of both complexes have almost the same thickness of about 1000 m; (b) In both of them, ultramafic interlayers are very common along contact areas with overlying ultramafic massifs; (c) Rocks of both complexes have the same geochemical signatures as shown in *M*-values (c. 80), low TiO_2 (0.17 vs. 0.18 wt.%) and particularly similar trace element concentrations (Sr = 144 vs. 125 ppm, Ni = 472 vs. 295 ppm, Zr = 8 vs. 11 ppm) including REE patterns (Table 2, ans. 6 and 15); Fig. 7).

4.2. Amphibolites related to diabase–dolerites

These rocks are also very common, both in the DOZ and the VZ. Majer and Lugović (1985) studied these rocks in Banovina in the northwestern part of VZ, where they build up a fragment of about 10 km^2 included in VZM. The amphibolites are granonematoblastic to nematoblastic; relict blastoophitic and ophitic textures are common. They are commonly distinctly foliated but never layered.

As a rule, they are bimineralic and composed of amphibole and plagioclase with subordinate zoisite or epidote. Predominant amphibole is commonly actinolitic hornblende and magnesian hornblende with low Al^{VI} and generally low Ti, suggesting low-pressure conditions; in some places, amphiboles are slightly to moderately transformed into chlorite. Plagioclase ranges in composition from An_{46} to An_{21}, but relict transitional dolerite–amphibolite varieties contain more calcic plagioclase (Majer and Lugović, 1985). As the protolithic diabase–dolerites, these amphibolites are cross-cut by veins filled by laumontite, stilbite, epidote, albite, chlorite, calcite, tremolite and quartz (Pamić et al., 1979). This paragenesis, together with the occurrence of secondary actinolite, chlorite and sodic plagioclase indicate that the amphibolites were influenced by a retrograde greenschist event as were the surrounding ophiolites.

Based on published mineral chemical analyses (Majer and Lugović, 1985) using geothermometers and geobarometers after Spear (1981), Plyusnina (1982), Fershtater (1990), Blundy and Holland (1990) and Holland and Blundy (1994) for hornblende–plagioclase paragenesis, we calculated *P–T* values for three amphibolite localities from Banovina in the northwestern VZ. Obtained *P–T* values are in a range between 600 and 650 °C and 520 and 250 MPa. This result fits quite well with *P–T* values of 580–645 °C and 3.5–4.5 kbar calculated for the surrounding metasedimentary metamorphic sole around adjacent ultramafic body (Majer, 1984). These data indicate *P–T* conditions of amphibolite facies (Bucher and Frey, 1994).

Genetical relation between this genetical group of amphibolites and diabase–dolerites is supported by several lines of evidence: (a) Within the amphibolites, relict blastoophitic and ophitic textures are very common; (b) Uneven amphibolitization of diabase–dolerite bodies on a larger scale; (c) Rocks of both groups are characterized by the same geochemical signatures as shown in *M*-values (60 vs. 58), high TiO_2 contents (1.40 vs. 1.50 wt.%) and particularly in trace element concentrations (Sr = 160–170 vs. 140; Zr = 90 vs. 80;

Ni = 117 vs. 113 ppm) including REE patterns (Table 2, ans. 9–10 and 17; Fig. 7).

4.3. Metamorphosed sedimentary sequences

These are found along the margins of some larger ultramafic massifs (e.g. Mt. Zlatibor), but also as dismembered fragments together with genetically related ultramafics scattered within the DOZM and VZM and rarely as isolated and exotic fragments without peridotites. Karamata and Pamić (1972) concluded that this metamorphism might have taken place under the influence of solid but still heated ultramafic blocks during their emplacement and dismembering over neighbouring shales and graywackes.

Along the eastern margin of the Zlatibor ultramafic massif, the most complete profile of the contact aureole, about 150–250 m thick, is preserved (Korikovsky et al., 2000). Protolithic sequences consist of graywackes, metasandstones, shales, tuffs, basalts and diabases. Within the aureole zone, three progressive metamorphic groups can be distinguished: (1) The peripheral zone includes a mineral assemblage (pumpellyite, chlorite, epidote, actinolite, albite and quartz) that originated at 300–350 °C; (2) The intermediate zone with chlorite, epidote, albite and prograde zoned amphibole (actinolite–hornblende) originated at 450–550 °C; (3) The internal zone, characterized by hornblende, clinopyroxene (with 2–3% jadeite), plagioclase (up to 50% anorthite) and quartz, originated at 550–650 °C. The total pressure of the zonal metamorphism was estimated at 3–3.5 kbar with the obduction and cooling of the ultramafic massif at 12–13 km.

Much more complicated relations were established in the inverted contact aureole, 50–200 m thick, beneath the Brezovica harzburgite massif from the southeasternmost DOZ (not presented in Fig. 1B). Based on detailed examinations of phase equilibria, Karamata et al. (1996, 2000) concluded that the Brezovica ultramafic massif was emplaced to depths of 30–60 km into LT (300–350 °C)–HP sequence. Based on petrological evidence, the inverted metamorphism took place at 8–9 kbar within the temperature range from 350–500 °C to a maximum of 700–750 °C.

In the northwesternmost parts of the VZ, southeast of Zagreb, a comparatively small ultramafic body is in tectonic contact with mappable Abukuma-like metamorphic sequence, composed mainly of phyllite, varieties of mica schist and gneiss, and amphibolite which yielded K–Ar ages of 168–165 Ma (Majer, 1976; Majer and Winkler, 1979; Majer and Lugović, 1985). The metamorphic sequence, together with the ultramafic body, forms an exotic block included in the VZM. Majer (1984) calculated *P–T* conditions of 580–645 °C and 3.5–4.5 kbar for this metasedimentary metamorphic sole. In the southeastern VZ in Serbia, *P–T* conditions of 435–450 to 515–550 °C and about 5 kbar were calculated in metamorphic sole mica schists and gneisses beneath the Povlen harzburgite body (Srećković-Batoćanin and Vasković, 2000).

5. Discussion

5.1. Geodynamics

The Dinaridic ophiolites, which occur in the DOZ and the VZ, show some similarities but also distinct dissimilarities in their ages, stratigraphy and facies of the associated genetically related sedimentary, igneous and metamorphic formations, as summarized in Table 3. The origin of the Dinaridic ophiolites is related to the evolution of the Dinaridic part of the Tethyan oceanic system. Its history can be divided into several phases.

I. Rifting phase. The earliest steps in the evolution of the Dinarides coincide with the break-up of Pangea and commenced during the Middle Permian with a rifting episode, accompanied by magmatic activity, that affected Variscan basement and lasted for some 40–50 Ma (Pamić, 1984). On the Adriatic–Dinaridic carbonate platform, and within the Paleozoic–Triassic nappes of the Internal Dinarides, some discrete zones of Triassic rift-related igneous rocks occur. These mark the superficial traces of rifts transecting the Paleozoic basement (“aborted rifts”; Fig. 11A).

During the Late Permian, the rift arches started to subside giving rise to the development of shelf area accompanied by the deposition of clastic sediments. Late Permian rifting was accompanied only in a few places by very weak magmatic activity. Later, during the Scythian, accelerated graben subsidence was accompanied by increased influx of terrigeneous material and only in a few places by coeval volcanic activity.

Table 3
Similarities and dissimilarities of the DOZ ophiolites and VZ ophiolites and associated genetically related sedimentary, igneous and metamorphic units

Dinaride Ophiolite Zone	Vardar Zone
Ophiolite mélange with lower degree of dismembering and tectonization—olistostrome mélange. In shaly–silty matrix are included fragments of ophiolites and genetically related sediments and exotics.	Ophiolite mélange with higher degree of dismembering and tectonization—strongly tectonized mélange. In shaly–silty, rarely mylonitized serpentinite matrix are included fragments of ophiolites and genetically related sediments and exotics.
Large ultramafic massifs are thrust onto mélange	Subordinate large ultramafic massifs are thrust onto mélange
In both DOZ and VZ lherzolites are predominant in their northwestern and central parts, whereas depleted harzburgites predominate in their southeastern parts	
The youngest limestone exotics are of Tithonian age	The youngest limestone exotics are of Late Cretaceous/Paleocene age
In matrix no index fossils; in very rare limestone interlayers, uncharacteristic Jurassic fossils occur	No characteristic fossils; preserved angiosperm leafs indicate minimal Early Cretaceous age
Radiometric ages of ophiolites and metamorphic sole rocks: 174 ± 14–136 ± 15 Ma	In remains of DOZM 189 ± 6.7–160 ± 10 Ma plus 109.6 ± 6.6–62.2 ± 2.5 Ma in VZM
Metamorphic soles: (a) eclogite (?)–granulite–amphibolite facies amphibolites related to cumulates, partly retrograded; (b) medium-grade amphibolites related to diabase–dolerites, partly retrograded; (c) low- and medium-grade metasedimentary sequences	Metamorphic soles: (a) medium-grade amphibolites related to diabase–dolerites, partly retrograded; (b) low- and medium-grade metasedimentary sequences; sparse glaucophane schists as olistoliths in Cretaceous flysch
Associated sedimentary and igneous formations: (a) Lower Cretaceous transgressive clastics with redeposited ophiolites and related sediments and Upper Cretaceous limestones underlain by crust of weathering of ophiolites (b) Radiolarite Formation with basalt interlayers forms an interrupted, more than a 100-km-long zone, built into present structure of the DOZ	Associated sedimentary, igneous and metamorphic formations: (a) Late Cretaceous – Early Paleogene flysch, at the base interlayered with subduction-related basalts and rhyolites, underlain by Lower Cretaceous paraflysch (?) (b) Mappable fragments of radiolarites and Paleozoic and Triassic formations included into VZM (c) Late Paleogene very low-, low- and medium-grade metamorphic sequences originating from Cretaceous–Early Paleogene flysch (d) Synkinematic S-type and alkali–feldspar granitoids (56–47.8 Ma)
Postorogenic Oligocene volcanics rarely occur in northern marginal parts of the DOZ adjoining the VZ	Very common are postorogenic Oligocene volcanics ± penecontemporaneous granitoids correlative with the Periadriatic tonalites
Intramontane Neogene freshwater coal-bearing basins	Transpressional Oligocene marine to freshwater sediments ± volcanics disconformably overlain by marine to freshwater fill of the Pannonian Basin ± contemporaneous volcanics
Environments: open-ocean (MORB), pelagic and near-continental margin	Environments: BARB; secondary accretion of MORB-like ophiolites and arc-related magmatism influenced by metasomatized upper mantle wedge peridotites
First Late Jurassic/Early Cretaceous obduction of ophiolites related to intra-oceanic subduction	Second Late Paleogene obduction of ophiolites related to the termination of subduction and continent–continent collision

The onset of the Anisian was marked by a regional (eustatic?) transgression; mixed carbonate and clastic sedimentation on shelves was only locally accompanied by weak rift-related volcanic activity. Strong reduction in the terrigeneous material input, and its subsequent complete termination, is the result of a regional transgression.

Magmatic activity peaked during the Ladinian, reflecting accelerating rifting activity. This was accompanied by the subsidence of probably isolated, narrow basins in which deep-water cherts, pelites and limestones were deposited. Synsedimentary rift-related volcanic activity varied in intensity along reactivated rift faults. On rift shoulders between the basins, carbonate sedimentation continued. Along the distal margin of the Adriatic-Dinaridic carbonate platform (ADCP), strong volcanic activity was accompanied by the intrusion of plutonic rocks (Pamić, 1984).

In the western part of the central Dinarides, volcanic activity continued up into the Early Norian. There are no traces of magmatic activity in ADCP sequences after the Norian. With this, a passive margin, facing the Internal Dinarides tectonostratigraphic units was established. With that, the ADCP formed part of the extensive carbonate shelf and/or platform system, which fringed the gradually opening Western Tethys embayment (Marcoux et al., 1993).

II. Opening of the oceanic Dinaridic Tethys may have started during the Middle/Late Triassic and/or at the beginning of the Early Jurassic. Some of the Ladinian/Late Triassic isolated rift basins, located along the outer margin of the ADCP may predispose the location of a future sea-floor spreading center, whereas rift zones located in their interior parts aborted simultaneously. Along the latter, the oldest radiolarite sequences coeval with basalts of Middle and Late Triassic age (Halamić and Goričan, 1995) indicate that in the Dinaridic Tethys, pelagic deep-water conditions were established already before onset of sea-floor spreading (Fig. 11B).

The open-ocean regime with pelagic sedimentation and sea-floor spreading may have lasted in the Dinaridic Tethys for a period of 70–80 Ma, that is, from the Late Triassic until the Late Jurassic/Early Cretaceous. The complete ophiolite profile preserved in the Mt. Varda area, with an average thickness of about 4–5 km is compositionally correlative with the recent oceanic crust (Coleman, 1977), thus indicating obduction of the entire oceanic crust onto the Apulian margin.

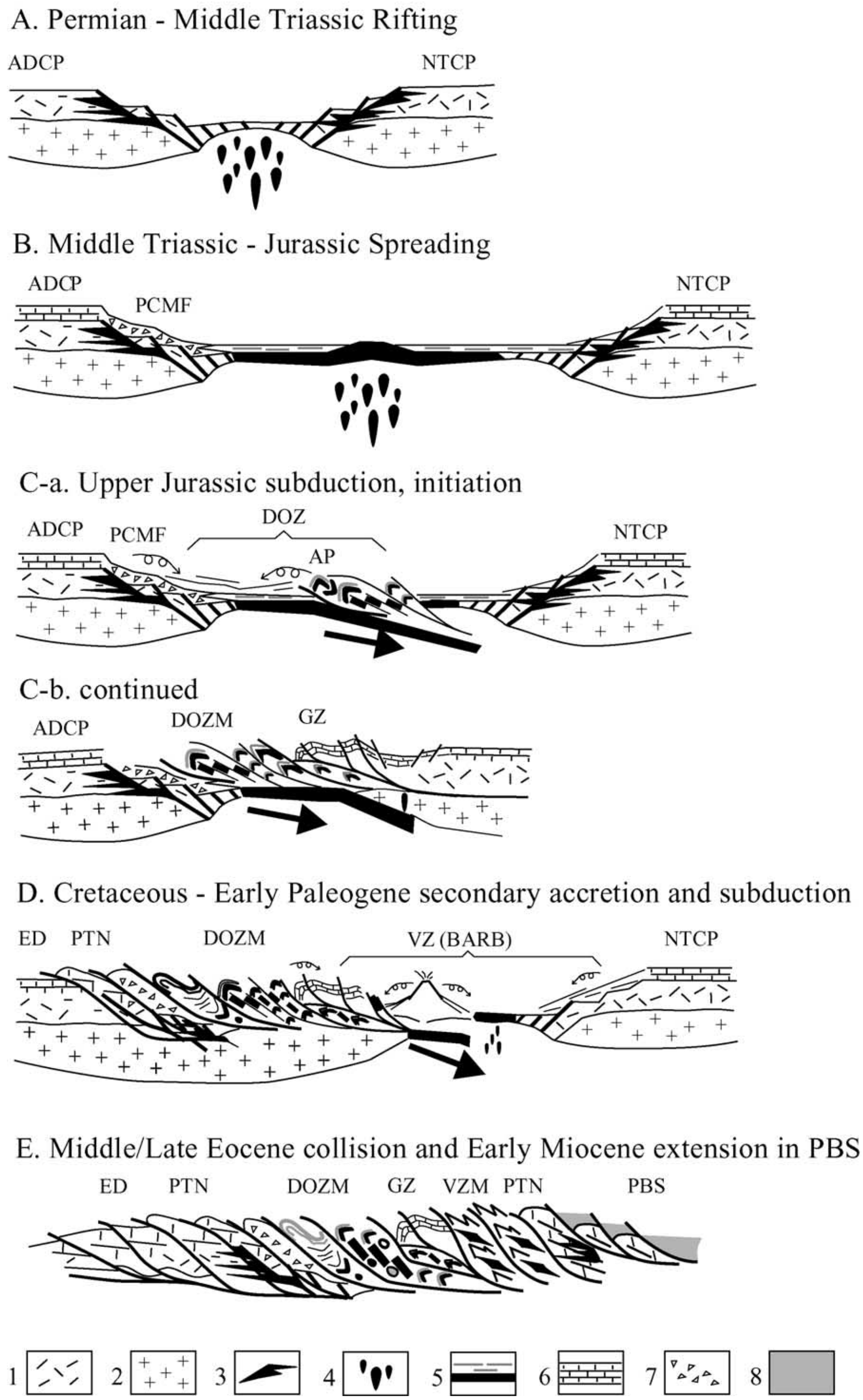
This Triassic–Jurassic Dinaridic Tethys should correspond to the Maliak–Meliata ocean proposed by Stampfli et al. (1991).

III. Convergence and the first emplacement of ophiolites. It is difficult to model the active continental margin of the Dinaridic Tethys because the Mesozoic–Paleogene formations of the northern Dinarides are mainly covered by the Tertiary sediments of the South Pannonian Basin which conceal the contact with the Eurasian margin. It could be visualized that graywackes and shales with coeval basalt flows originated during the Late Triassic (?) and Jurassic/Early Cretaceous along the northern margin of the Dinaridic Tethys where several rises and troughs facilitated the development of the olistostrome mélange. However, there is no evidence for the existence of a magmatic arc.

Processes of intra-oceanic NE-dipping subduction must have started during the Late Jurassic/Early Cretaceous as indicated by the first ophiolite obduction onto the Apulian marginal units (Fig. 11C-a). This was accompanied at the base of subducting oceanic crust by the generation of metamorphic sole composed of higher graded eclogite (?)–granulite–amphibolite facies amphibolites, the protoliths of which are cumulate gabbros, and lower graded amphibolites, the protoliths of which are diabase–dolerites. With the initiation of obduction processes, the exhumation of the higher grade amphibolite masses started but under changed geothermal gradient giving rise to gradual metamorphic zonation of the metamorphic sole amphibolites. Subducting diabase–dolerite masses and surrounding graywackes and shales equilibrated with obducting peridotites at higher levels, that is, the P–T conditions of amphibolite facies and during the subsequent exhumation were affected by greenschist facies retrogression. Breaking and dismembering of obducting oceanic crust with its metamorphic sole started at near-surface parts of the obducting system.

In this geodynamic interpretation, it is difficult to explain the original paleogeographic position of the allochthonous Paleozoic–Triassic units of the eastern parts of the DOZ. Robertson and Karamata (1994) proposed that, during the Mesozoic, these units formed a continental block, referred to as the Drina–Ivanjica–Pelagonide microcontinent (Dimitrijević, 1982) which separated the Vardar Ocean in the east from the Dinaridic Ocean in the west.

However, an alternative interpretation (Aubouin et al., 1970; Pamić, 1993) suggests that this alleged Drina–Ivanjica microcontinent represents an allochthonous mass, the Golija Zone (Rampnoux, 1970), which was thrust over the DOZ units. This is indicated by numerous tectonic windows exposing ophiolites and genetically related sedimentary rocks beneath the Golija nappe (Fig. 11C-b).


A. Permian - Middle Triassic Rifting
ADCP
NTCP
B. Middle Triassic - Jurassic Spreading
ADCP
PCMF
NTCP
C-a. Upper Jurassic subduction, initiation
ADCP
PCMF
DOZ
AP
NTCP
C-b. continued
ADCP
DOZM
GZ
D. Cretaceous - Early Paleogene secondary accretion and subduction
ED
PTN
DOZM
VZ (BARB)
NTCP
E. Middle/Late Eocene collision and Early Miocene extension in PBS
ED
PTN
DOZM
GZ
VZM
PTN
PBS
1
2
3
4
5
6
7
8

The obducted ophiolite complex partly emerged, underwent weathering and erosion. The erosion products were redeposited during the Late Jurassic/Early Cretaceous to Late Cretaceous in shoals and depressions located between the emerging ridges. The occurrence of blueschist olistoliths and pebbles at the base of the Late Cretaceous flysch sequence of the northern Dinarides suggests that rapid exhumation must have occurred during this first post-emplacement period (Michard et al., 1994). Late Jurassic/Early Cretaceous overstep clastic sequences of the northern marginal parts of the DOZ were not only charged by detritus from adjacent ophiolite terranes but also by detritus and exotic blocks of reddish Paleozoic (?) granites from an unknown source area (Varićak, 1955), probably somewhere from the southern Eurasian margin. Such granites are nowhere exposed in the present structure of the Dinarides.

With the onset of Late Jurassic/Early Cretaceous subduction, gradual closure of the Dinaridic Tethys and the development of a magmatic arc commenced (Fig. 11D). This arc was located north of the obducted ophiolites along the north Dinaridic Tethyan margin. In the trench associated with this magmatic arc, Cretaceous–Early Paleogene flysch sequences accumulated. The sedimentary records of the southeastern VZ indicates that trench sedimentation persisted during the Early and Middle Cretaceous (Dimitrijević, 1995). Below the trench, the upper mantle diapir was probably uplifted by a mechanism of secondary accretion (Karig, 1971; Crawford et al., 1981) as indicated by the ophiolite fragments (110–62 Ma) included in the VZM. Apart from a continuing generation of oceanic crust, the persistent subduction processes along this arc–trench system also gave rise to a bimodal basalt–rhyolite volcanism and granite plutonism influenced by upper mantle wedge, as indicated by geochemical data (Pamić et al., 2000). This arc–trench system can be conceived as a back-arc basin, presumably spatially related to the southern European active margin, and formed the westernmost part of the huge north Tethyan subduction zone which extends southeastward to Greece, Zagros and Afghanistan (Camoin et al., 1993). According to Stampfli et al. (1991), the back-arc basin of the Dinaridic Tethys corresponds to the Vardar Ocean.

IV. Collision and second emplacement of ophiolites. The collision which took place during the Early Eocene (56–48 Ma) had the strongest impact on the present-day architecture of the Dinarides, resulting in their final uplift (Fig. 11E). This deformational episode was characterized by: (1) Tectonization of the pristine DOZM, plus incorporation of Cretaceous–Paleogene ophiolites and sedimentary fragments into the VZM and their emplacement on top of the DOZM that has been obducted already during the initial Late Jurassic/Early Cretaceous subduction; (2) Medium-grade metamorphism of the Late Cretaceous–Paleogene trench sediments with accompanying igneous rocks; and (3) Synkinematic granite plutonism. With this Eocene final orogenic pulse, Tethyan subduction processes terminated in the Dinaridic domain and structuration of the Dinarides was completed.

Consequently, the Dinaridic ophiolites represent remnants of the Mesozoic oceanic crust which was generated in the Dinaridic part of the Tethys during a period of about 150 Ma. The bulk of oceanic crust was generated in the progressively opening Dinaridic Tethys during the Late Triassic to Late Jurassic/Early Cretaceous at which times subduction processes, accompanied by the first obduction of ophiolites onto the Apulian margin, commenced. Subsequently, generation of the oceanic crust continued by secondary accretion during the Cretaceous and Early Paleogene in reduced Tethyan basin under a back-arc setting. Final closure of this BARB during the Late Eocene was accompanied by the second emplacement of ophiolites and collisional deformation, which gave rise to the final structuration of the Dinarides fold-thrust belt and their uplift.

Fig. 11. Tentative graphic presentation of the geodynamic evolution of the central Dinarides (see text for explanation). Legend: 1 upper continental crust; 2 lower continental crust; 3 rifting related magmatism; 4 magma diapirs; 5 oceanic crust with radiolarites; 6 carbonate platform sediments; 7 passive margin talus; 8 Neogene fill; ADCP Adriatic-Dinaridic carbonate platform; AP accretionary prism; BARB back-arc basin; DOZ Dinaride Ophiolite Zone; DOZM Dinaride Ophiolite Zone mélange; ED External Dinarides; GZ Golija Zone; NTCP North Tethyan carbonate platform; PBS Pannonian Basin System; PCMF passive continental margin formations; PTN nappes mainly composed of Paleozoic–Triassic formations; VZ Vardar Zone; VZM Vardar Zone mélange.

5.2. Petrological and geochemical similarities and dissimilarities of the DOZ and VZ ophiolites

There is a discrepancy in the consideration of geotectonic setting of the Dinaridic ophiolites. Trubelja et al. (1995) are of the opinion that they have MORB signatures, whereas Pamić and Desmons (1989) and Lugović et al. (1991) cautiously emphasize that they have both MORB and subduction BARB signatures and thus did not bring an unequivocal geodynamic conclusion.

According to the geodynamic model presented (Fig. 11), the DOZ ophiolites represent fragments of the oceanic crust which originated during the Late Triassic (?)/Jurassic period along the constructive plate margin (MORB), that is, in an open ocean geotectonic setting. On the other hand, the VZ ophiolites originated during the Cretaceous/Early Paleogene in the more reduced Tethys along the consuming plate margin, that is, in a back-arc basin. Hence, the Dinaridic ophiolites give a unique chance to be considered as products of two quite different geotectonic settings.

The most predominant fertile spinel lherzolites and subordinate-depleted harzburgites have the same fabric in both DOZ and VZ; they have slight and quite neglectable differences in chemical composition of rock-forming minerals. This is reflected in their average major and trace element contents (Table 2, ans. 1–4). Predominant lherzolites grade laterally southwestward in predominant harzburgites both in DOZ and VZ. The predominance of the lherzolites in the central and northwestern parts of the two zones extends further northwestward in the Ligurian ophiolite belt also characterized by the predominance of fertile spinel lherzolites (Rampone et al., 1995). Consequently, the westernmost wedge parts of the Tethys are genetically related to subcontinental-upper mantle which is similar to the MORB upper mantle.

Because the VZ ultramafics have the same geochemical signatures as those from the DOZ, the upper mantle diapir which generated oceanic crust in the Cretaceous/Early Paleogene back-arc basin setting must also have had the same MORB-type signatures.

Ultramafic and mafic cumulates from both the DOZ and VZ have the same petrographic and geochemical characteristics as indicated by their major and trace element averages (Table 2, ans. 5–8). The cumulate rocks crystallized from tholeiitic primitive melts which originated by partial melting of the underlying peridotites along the constructive plate margin during the Jurassic as well as during the Cretaceous/Early Paleogene along consuming plate margin. However, strongly metasomatized cumulate peridotites from Mt. Medvednica in the northwesternmost VZ suggest that they must have been included in upper mantle wedge before their exhumation, and subsequent erosion and incorporation in VZM.

Diabase–dolerites and basalt–spilites from both the DOZ and VZ have the same fabric, petrography and geochemistry, as indicated by their average major trace element concentrations (Table 2, ans. 9–14). It means that the upper parts of the Dinaride ophiolite sequences were generated both along the constructive and consuming plate margins during the Jurassic and Cretaceous time, respectively.

The geochemical signature of both ultramafic–mafic cumulates and diabase–dolerites is also reflected in the derivative metamorphic sole amphibolites. However, the cumulate-related amphibolites have been so far registered only in DOZ.

Fig. 11D demonstrates the evolution from incipient ocean–continent subduction setting through the onset of rifting and the development of an embryonic BARB, generated as a consequence of the diapiric upwelling of deep asthenospheric mantle beneath the arc axis. This is "secondary accretion" (Karig, 1971) and the upper mantle diapir partially melts as a consequence of adiabatic decompression to produce MORB-like back-arc basin.

The rising mantle diapir interferes with the process of the arc magmatism, isolating (?) the arc volcanoes and their plumbing system from the arc magma sources represented by metasomatized pargasite-bearing peridotites from the upper mantle wedge (Crawford et al., 1981).

Consequently, final stages of the generation of the Dinaridic ophiolites, which took place in a BARB setting during the late Cretaceous/Early Paleogene, were accompanied by penecontemporaneous subduction-related arc magmatism. It gave the bimodal basalt–rhyolite association with cogenetic alkali–feldspar granites, which originated by fractional crystallization of primitive basalt melt of the upper mantle wedge source. Beside this, occurrences of fragments of edenite–pargasite bearing ultramafic cumulates in

VZM (Slovenec and Lugović, 2000) indicate that metasomatized upper mantle wedge rocks must also have been detached from depth, subsequently affected by erosion and incorporated in the VZM.

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