THE METABASITES FROM THE KOPAONIK METAMORPHIC COMPLEX, VARDAR ZONE, SOUTHERN SERBIA: REMNANTS OF THE RIFTING-RELATED MAGMATISM OF THE MESOTETHYAN DOMAIN OR EVIDENCE FOR PALEOTETHYS CLOSURE IN THE DINARIC-HELLENIC BELT?

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ABSTRACT
The Kopaonik area belongs to the Vardar Zone, regarded as the easternmost terrane of the Dinaric-Hellenic belt. This area is characterized by a pile of tectonic units that includes, from bottom to top: the Kopaonik metamorphic complex, the Brzece Unit, the Ophiolite Unit and the Brus Unit. During Early Oligocene, the nappe pile was intruded by the I-type Kopaonik intrusive complex. The Kopaonik metamorphic complex mainly includes metasedimentary rocks consisting of Upper Triassic metalmelstones, metarenites, metapelites and metadolomites. In addition, this complex is characterized by metabasites that occur as up to 200 m thick bodies folded together with the metasedimentary rocks of Late Triassic age. Their original stratigraphic position is here interpreted at the base of the metasedimentary succession. These rocks show a complex deformation history that includes two phases developed under upper greenschist-upper amphibolite conditions followed by subsequent two phases characterized by very-low grade metamorphism. The geochemical affinity of the igneous protoliths of the studied rocks, evaluated using discrimination diagrams based on the relative distributions of several immobile elements, indicates basaltic magmas generated at continental arc settings. Two different, alternative geodynamic hypotheses about the origin of these rocks are proposed in this paper. The first hypothesis suggests an origin during the Lower Triassic rifting phase by partial melting from small areas of enriched hydrous mantle, probably a consequence of preexisting subduction. By contrast, an origin of the metabasites from Kopaonik metamorphic complex as remnants of the Upper Permian-Middle Triassic calc-alkaline magmatism related to Cimmerian orogenesis is provided by the second hypothesis.

INTRODUCTION
The Vardar Zone, belonging to the Dinaric-Hellenic belt of Alpine age, is constituted by an assemblage of oceanic and continental-derived units which runs from Greece to Macedonia and Serbia. In the frame of the eastern Tethys geodynamic evolution, the history of the Vardar Zone is still an open problem, mainly owing to the paucity of reliable geological data. In this domain, mafic igneous rocks of different ages, generally belonging to the Jurassic ophiolite sequences, are well represented. However, metamorphic basic rocks have been found in several occurrences, as, for instance, in the Kopaonik area. The deformation and metamorphic features as well as the geochemical affinity of these metamorphic rocks can provide very useful constraints in the reconstruction of the geological history of the Vardar Zone. However, modern geological data on these metabasites are still lacking.

In this paper, we report data about the structural geology, petrography and geochemistry of the metabasites recognized in the Kopaonik area, Vardar Zone, discussing the implications for the tectonic evolution of the Dinaric-Hellenic belt.

GEOLOGICAL SETTING
The Dinaric-Hellenic belt is an orogenic chain of Alpine age derived from the Mesozoic to Tertiary convergence between the Adria and the Eurasia Plates. In the classical reconstructions, the evolution of the Dinaric-Hellenic belt started with a rifting stage mainly developed in Early Triassic time (Dimitrijević, 1982; Pamić et al., 2002, Bortolotti et al., 2004), along the northern margin of Gondwanaland. The rifting process, characterized by syn-rift deposits and calc-alkaline magmatism, evolved in the Middle to Late Triassic to oceanic spreading and drifting with development of wide basin characterized by mid-ocean ridge (MOR) oceanic lithosphere (Collaku et al., 1992; Pamić et al., 2002, Bortolotti et al., 2004; Saccani et al., 2004). This oceanic basin, carrying (Karamata et al., 1994; Dimitrijević, 2001) or not (Pamić et al., 1998, 2002) a microcontinent inside, was located between the Adria and Eurasia continental margins. The convergence began during the Early Jurassic, with development of an intraoceanic subduction followed by the formation of new oceanic lithosphere in the suprasubduction basin. As a consequence of convergence, obduction process developed, resulting in the emplacement of oceanic lithosphere slices onto the continental margins of the Adria Plate during the Middle to Late Jurassic time (Dimitrijević, 2001). The convergence between Adria and Eurasia finally led to the continental collision during Late Jurassic - Early Cretaceous time span. After the continental collision and up to Neogene, a continuous convergence, still active today, affected the continental margin of the Adria Plate, that was progressively deformed in westward-vergent, thick-thinned thrust sheets. The continental collision was also characterized by emplacement of calc-alkaline granitoids, mainly of Late Eocene - Early Oligocene age (Pamić, 1998).

This long-lived geodynamic evolution produced the structural pattern of the Dinaric-Hellenic belt, represented by an assemblage of northwest-southeast to north-south
trending zones, corresponding to the modern concept of terranes (Bortolotti et al., 2004 and quoted references). Each zone consists of an assemblage of variably deformed and metamorphosed tectonic units of oceanic and/or continental origin. These zones, from west to east, are 1- Deformed Adria Zone, 2- External ophiolite belt, 3- Pelagonian-Korab-Drina-Ivanjica Zone and, 4- Vardar Zone. These zones are bounded to the west by the undeformed Adria Zone and to the east by the Serbo-Macedonian-Rhodope Massif, generally considered as the stable margin of the Eurasia Plate (Fig. 1).

The Deformed Adria Zone consists of a west-verging imbricate stack of tectonic units derived from the continental margin of the Adria Plate. These units are thrust onto the undeformed Adria margin. From west to east, this deformed zone is represented by the Ionian, Gavrovo (Kruja in Albania), Pindos (Krasta-Cukali in Albania) and Parnassos Units. All these units are characterized by unmetamorphosed sequences, each including Triassic to Paleocene neritic and pelagic carbonate sequences topped by widespread Upper Cretaceous to Miocene siliciclastic turbidite deposits. The age of inception of the flysch deposition ranging from Late Cretaceous in the Pindos Unit to Late Oligocene in the Ionian Unit, is related to the westward migration of the deformation across the continental margin of the Adria Plate.

In Montenegro, Bosnia, Croatia and Serbia, the Deformed Adria Zone is represented, from west to east, by the Budva, high-Karst and pre-Karst Units. Whereas the Budva Unit can be regarded as the northward counterpart of the Pindos Unit of Greece and Krasta-Cukali of Albania, the high-Karst and pre-Karst Units can be probably correlated with the Par- nassos Unit in Greece.

Eastward, the Deformed Adria Zone is thrust by the External ophiolite belt, represented by an huge oceanic nappe. This nappe is characterized by the occurrence of ophiolites ranging in age from Triassic to Jurassic, representative of the oceanic basin located eastward to the Adria Plate. This nappe consists of a stack of ophiolite units showing at their base a sub-ophiolite mélange, formed by an assemblage of continental- and oceanic-derived units. The External ophiolite belt is recognized as continuous nappe from Argolis, Othrys, Pindos, Vourinous in Greece, to Mirdita in Albania, Bistrica and Zlatibor in Serbia up to Krivaja in Croatia (e.g. Bortolotti et al. 2004).

By contrast, the Pelagonian-Korab-Drina-Ivanjica Zone, hereafter simply reported as Pelagonian, is represented by
an assemblage of tectonic units consisting of a pre-Alpine basement intruded by Upper Paleozoic granitoids and covered by a Permian to Lower Triassic siliciclastic deposits followed by Middle Triassic to Upper Jurassic carbonates. The Pelagonian Units are thrust by the units belonging to the Vardar Zone.

The easternmost Zone is represented by the Vardar one, located close to the Serbian-Macedonian Massif: the Vardar Zone is represented by a composite assemblage of continental and oceanic-derived units, including also both Triassic and Jurassic ophiolites. The latters represent the internal ophiolite belt of the Dinaric-Hellenic chain.

The study area is located in the Kopaonik area, Southern Serbia, where a pile of continental- and ocean-derived units belonging to the Vardar Zone crop out (Dimitrijević, 1997). The tectonic stack of the studied area includes, from bottom to top (Fig. 2): the Kopaonik metamorphic complex (KMC), the Brzece Unit, the Ophiolite Unit and the Brus Unit (Marroń et al., 2004). During Early Oligocene, the nappe pile was intruded by the I-type Kopaonik intrusive complex (Dimitrijević, 1997), producing high-temperature contact metamorphism, represented by skarns and hornfels, along its margins, mainly in the KMC. The KMC, also known as “The Central Kopaonik Series” (Urosević et al., 1973), consists of metabasites, metarenites, metapelites, metadolomites and metalimestones, showing a complex deformation and metamorphic history. The recognized conodont fauna (Suder, 1986; Mijić et al., 1968) point to a Carnian age for the metalimestones. Although the relationships among the metabasites and the metasedimentary rocks are still unclear, this succession has been regarded as a single rifting-related Triassic volcano-sedimentary formation by Dimitrijević (2001), according to the same deformation and metamorphic features in both metabasites and metasedimentary rocks. However, the features of the metabasites belonging to the KMC, as well as their geodynamic significance are still unknown. To highlight this problem, selected samples of metabasites form the Kopaonik area have been analyzed in order to determine their geological, structural, metamorphic and geochemical features.

**SAMPLING AND ANALYTICAL METHODS**

About 40 samples of metabasites have been collected during the field mapping, mainly in the western area. All the different types of metabasites have been sampled in order to provide a complete range of these lithologies. Eleven samples have been selected for whole-rock and mineral chemistry analysis.

Mineral compositions (Tables 1 and 2) were determined with a JEOL JXA 8600 electron microprobe fitted with four wavelength-dispersive spectrometers at the CNR Istituto di Geoscienze e Georisorse in Florence. Running conditions were 15 kV accelerating voltage and 10 nA beam current on a Faraday cage. Counting time for the determined elements ranged from 10 to 60 s at both peak and background. Nominal beam spot size was 1 mm for the analyses of all minerals but plagioclases, for which a defocused beam, from 5 to 15 mm in diameter according to grain size, was employed to reduce the undesirable loss of volatile elements. The Bence and Albee (1968) method was employed for the correction...
### Table 1 - Analyses of amphiboles from the Kopaonik metabasites.

<table>
<thead>
<tr>
<th>Sample</th>
<th>KP-12</th>
<th>KP-12</th>
<th>KP-12</th>
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<td>SiO₂ (wt.%)</td>
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<td>44.4</td>
<td>44.4</td>
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<td>55.2</td>
<td>55.3</td>
<td>53.8</td>
<td>53.1</td>
<td>50.7</td>
<td>48.9</td>
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<td>TiO₂</td>
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<td>0.54</td>
<td>0.76</td>
<td>1.97</td>
<td>-</td>
<td>-</td>
<td>0.11</td>
<td>0.14</td>
<td>0.27</td>
<td>0.40</td>
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<tr>
<td>Al₂O₃</td>
<td>10.7</td>
<td>10.6</td>
<td>9.52</td>
<td>9.52</td>
<td>12.7</td>
<td>1.14</td>
<td>0.83</td>
<td>2.89</td>
<td>4.25</td>
<td>5.59</td>
<td>6.48</td>
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<td>FeO</td>
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<td>20.2</td>
<td>19.3</td>
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<td>6.99</td>
<td>9.22</td>
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<td>0.38</td>
<td>0.26</td>
<td>0.54</td>
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<td>9.65</td>
<td>10.9</td>
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<td>20.3</td>
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<tr>
<td>CaO</td>
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<td>11.9</td>
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<td>11.7</td>
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<tr>
<td>Na₂O</td>
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<td>0.25</td>
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<td>0.94</td>
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<td>1.34</td>
<td>-</td>
<td>-</td>
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<td>0.01</td>
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<td>97.2</td>
<td>97.4</td>
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</tbody>
</table>

Structural formulae calculated on the basis of 23 oxygen atoms

### Table 2 - Analyses of feldspars from the Kopaonik metabasites.

| Sample | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 | KP-12 |
|--------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
|        | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   | rim   |
| SiO₂ (wt.%) | 59.8  | 58.9  | 58.7  | 59.5  | 60.7  | 57.8  | 59.2  | 58.0  | 61.8  | 62.6  | 55.2  | 64.9  |      |      |      |      |
| Al₂O₃   | 25.7  | 25.6  | 26.8  | 25.8  | 25.0  | 27.7  | 25.0  | 27.2  | 24.7  | 23.6  | 28.9  | 18.9  |      |      |      |      |
| Fe₂O₃   | 0.39  | 0.38  | 0.39  | 0.39  | 0.33  | 0.34  | 0.30  | 0.33  | 0.34  | 0.06  | -     | 0.09  |      |      |      |      |
| CaO     | 7.50  | 7.82  | 8.41  | 7.53  | 6.53  | 9.19  | 7.63  | 9.09  | 5.53  | 5.30  | 10.9  | 0.14  |      |      |      |      |
| Na₂O    | 6.45  | 6.83  | 6.52  | 7.08  | 7.58  | 6.31  | 6.98  | 5.29  | 8.01  | 8.57  | 5.16  | 0.49  |      |      |      |      |
| K₂O     | 0.15  | 0.17  | 0.15  | 0.13  | 0.09  | 0.14  | 0.11  | 0.14  | -     | -     | -     | 0.03  | 15.5 |      |      |      |      |
| Total   | 100.0 | 99.7  | 100.9 | 100.3 | 100.2 | 101.6 | 100.3 | 100.7 | 100.0 | 99.1  | 100.4 | 99.8 |      |      |      |      |

Structural formulae calculated on the basis of 8 oxygen atoms

# Structural formulae calculated following Leake et al. (1997).
of all data. A number of synthetic and mineral standards were used for instrumental calibration.

Major elements were determined by X-ray fluorescence (XRF) (Philips PW 1480) analysis on pressed powder pellets (Table 3). Matrix effect were corrected for by applying the procedure of Franzini et al. (1975). Estimated precision (relative standard deviation) is about 1% for SiO$_2$ and about 2% for the other major elements except for low concentrations (~ < 0.50 wt.%) for which the absolute standard deviation is about ± 0.01%.

The concentration of a set of thirty-four trace elements (Table 3) was determined by inductively coupled plasma-mass spectrometry (Fisons PQII Plus). Powder aliquots of 50-70 mg were dissolved in PFA vessels on a hot plate at 120°C using HF + HNO$_3$. The sample solutions, spiked with Rh, Re and Bi as internal standards, were measured in replicates by external calibration. Elemental sensitivities were obtained by measuring solutions of dissolved geochemical reference samples with basaltic composition. Analytical precisions, evaluated by repeated analyses of the in-house standard HE-1 (Mt. Etna hawaiite), are between 2 and 5% relative standard deviation, except for Gd, Tm, Be, Sc, Pb (6-8%).

THE KOPAONIK METABASITES

Field and structural geology

The Kopaonik metabasites consist of rootless, up to 200 m thick bodies of metabasites that occur as scattered lenses folded with the other metamorphic rocks. Where observed, the relationships between metabasites and the other lithologies of the KMC are represented by shear zones. The main bodies are found in the northwestern side of the Kopaonik Mts., mainly near Josanicka Banja. The metabasites are intruded by the Oligocene Kopaonik intrusive complex with development of m-thick hornfels in the contact area.

![Fig. 3 - Field occurrence of metabasites from KMC. The outcrop is characterized by cm-thick layers represented by epidote-rich levels that are deformed by the isoclinal, rootless F2 folds.](image)

![Fig. 4 - Photomicrographs of the Kopaonik metabasites. a) typical metabasite from Kopaonik (sample KP-13) showing nematoblastic texture (plane polarised light). b) particular of actinolite crystals showing a discontinuous rim (arrow) of green Mg-hornblende/pargasite (sample KP-38; plane polarised light).](image)
The metabasites occur as well-foliated bodies, characterized by green colour and epidote-rich layers parallel to main foliation. A complex deformation history of the KMC, marked by four deformation phases (from D1 to D4), has been recognized both at the meso- and microscale. The oldest structure identified in the metabasites is represented by the S1 relict foliation, mainly visible at the micro-scale. In the field, the relics of the S1 foliation have been detected mainly within the hinge zones of the F1 isoclinal folds. Where observed, the S2 foliation occurs as a continuous schistosity characterized by aligned, medium to coarse-grained minerals. No evidence of folds related to D1 phase have been identified. The S1 foliation was folded during the D2 deformation phase into rootless, isoclinal F2 folds (Fig. 3). Generally, these folds are tight to isoclinal, showing thickened hinge zone with interlimb angles between 10° and 20°. The F2 folds are generally symmetric and strongly non-cylindrical, showing subrounded, rounded up to subangular hinges. The low-dipping A2 axes show a N/S trend, even if these structural elements are very scattered owing to the geometry of the F2 folds and the later deformation phases. However, the most widespread structure of the D2 phase recognized at the micro- and meso-scale is represented by the S2 foliation (Fig. 3), interpreted as a pervasive and, generally, fine-grained continuous schistosity. At the map scale the metabasites are folded by the D3 deformation phase represented by open to close, cylindrical F3 folds with subvertical axial plane. The A3 display a NNW-SSE trend and very low dipping in the whole investigated area. The associated foliation is represented by a disjunctive cleavage. The structures related to the D3 phase are deformed during the subsequent D4 phase. This phase is characterized by F4 symmetrical and cylindrical folds with subhorizontal axial plane. The F4 folds, very open to close, show subrounded hinge zone with an axial plane foliation represented by a disjunctive cleavage. The trend of the A4 axes is subparallel to the margin of the intrusive complex with a N/S trend along the western and eastern margins and E-W trend in the northern area. The F4 folds are associated to low-angle normal faults characterized by meter thick cataclastic shear zones. Also the hornfels are deformed by the D4 phase structures.

**Petrography and mineral chemistry**

The Kopaonik metabasites are fine- to medium-grained foliated amphibolite rocks with mineral assemblages including hornblende, plagioclase, actinolite, biotite, Fe-Ti oxides, clinopyroxene, quartz, epidote, chlorite, sphene, apatite and...
calcite. The modal proportions of these phases are widely variable, especially for epidote, chlorite and biotite. Where present, epidote and chlorite are retrograde phases indicative of a partial re-equilibration in the lower greenschist facies. The main structural element is represented by the S\(_2\) foliation, that occurs as schistosity defined by oriented grains of amphibole and plagioclase and by mineral layering (Fig. 4a). Quartz ± biotite ± chlorite minerals are also recognized as mineral grains aligned along the S\(_2\) foliation. In thin section, the S\(_2\) foliation has been recognized only within microlithons and/or inside the hinge zone of the F\(_2\) folds (Fig. 4b). The S\(_1\) foliation is a coarse grained schistosity characterized by amphibole + plagioclase ± quartz ± biotite ± chlorite. Overall the S\(_1\) and S\(_2\) foliation are characterized by similar metamorphic mineral assemblages, that can be regarded as characteristic of upper greenschist - upper amphibolite metamorphic facies.

Two samples (KP-41 and KP-12) have been selected for EMP analyses (Tables 1 and 2). These samples, characterized by different grain size or mineral assemblage, have been collected far from the Oligocene granitoids in order to minimize the effects of the contact metamorphism.

In the lower grade sample KP-41 we observed colourless actinolite cores with epitaxial overgrowths of green Mg-hornblende/pargasite (Table 1; Figs. 5 and 6), whereas in the same sample the plagioclase has very variable composition (An\(_{32-44}\); Table 2). Amphibole crystals in the more equilibrated amphibolite KP-12 are quite homogeneous Mg-hastingsite/pargasite/edenite (Table 1; Figs. 5 and 6). The plagioclase crystals of the amphibolite KP-12 are homogeneous andesine (An\(_{32-44}\); Table 2). The brown mica crystals analysed in sample KP-41 are biotite with Mg/(Mg+Fe\(^{2+}\)) ratio of 0.67 (total iron calculated as Fe\(^{2+}\)) and 0.6 apfu of Al\(^{VI}\). Sample KP-12 also contains granoblastic layers bearing metamorphic clinopyroxene (diopside, En\(_{27.8-37.4}\)Wo\(_{47.3-49.4}\); Table 3), suggesting that this unique sample reached the upper amphibolite facies conditions.

Phase equilibria and mineral chemistry data suggest that the Kopaonik metabasites formed at medium pressure regional metamorphism under upper greenschist to upper amphibolite facies (just for sample KP-12), with a final retrogression to lower greenschist facies.

Geochemistry
Nine representative samples from the Kopaonik metabasites were selected for the geochemical characterization (Table 4). The compositional variability of these samples is reasonably wide though some common features can be recognized. All samples have relatively high Mg\# values (58 - 64; Mg\# = molar MgO/(MgO + FeO\(_{tot}\)) and low TiO\(_2\) and P\(_2\)O\(_5\) contents (<1.5 wt.% and <0.20 wt.% respectively). On the other hand, their absolute contents of Na\(_2\)O and K\(_2\)O, as well as their Na\(_2\)O/K\(_2\)O ratios are extremely variable. Most samples except KP-42 have a low K\(_2\)O content (0.11 – 0.79 wt.%) while the Na\(_2\)O content varies between 2.2 and 6.0 wt.%. The trace-element distributions of the studied samples attest that their igneous protoliths had consanguineous rela-

![Fig. 6](image_url) - Hornblende and actinolite composition in the Kopaonik metabasites. (a) plotted in terms of (Na + K) vs. Mg/(Mg + Fe\(^{2+}\)); (b) plotted in terms of Al\(^{VI}\) vs. Mg/(Mg + Fe\(^{2+}\)).

![Fig. 7](image_url) - N-MORB-normalized incompatible element patterns (a) and CI chondrite-normalized rare earth elements patterns (b) for the Kopaonik metabasites. Normalizing values after Sun and McDonough (1989) and McDonough and Sun (1995), respectively.
tionships. In the diagrams of Fig. 7, where the incompatible element concentrations are normalized to the average N-MORB composition (Sun and McDonough, 1989), most samples (grey field of Fig. 7a) follow a similar distribution characterized by: i) HREE at N-MORB levels, or slightly lower; ii) slight negative Ti anomalies; iii) roughly progressive enrichments of the most incompatible elements (2 - 30 times the N-MORB level); iv) slight negative Ta and Nb anomalies.

A larger variability characterizes the abundances of Sr, K, Ba, Rb and Cs as expected for these highly mobile elements. The most remarkable difference between sample KP-42 and the other samples, is represented by its strong enrichment in the LIL elements K, Ba, Rb and Cs and lower contents of U, Th, Ta, Nb and LREE (Fig. 7). The REE distributions of the studied samples are better depicted in Fig. 7b where the REE concentrations are normalized to the CI chondrite of McDonough and Sun (1995). Most samples (grey field of Fig. 7b) have very similar REE patterns characterized by: i) moderate LREE/HREE fractionation ([La/Yb]N = 3.0 - 4.4); ii) low REE concentrations (La = 9.2 - 13.9 ppm); iii) very slight or absent negative Eu anomalies; iv) roughly rectilinear distribution.

The classification of the igneous protholiths of the metabasite rocks of this study was accomplished by using the diagrams proposed by Winchester and Floyd (1977). These diagrams are based on the abundance and distribution of the immobile elements Si, Ti, Zr, Nb and Y and are well suited to classify altered or metamorphosed rocks whose Na and K concentrations could have been severely modified by post-igneous processes. Combining the information derived from the SiO2 vs. Zr/TiO2, SiO2 vs. Nb/Y and Zr/TiO2 vs. Nb/Y diagrams (not shown), we conclude that the igneous protoliths of the studied samples can be classified as follows: KP-46, sub-alkaline basalt; KP-37, 38, 39, 40, 41, 44, 45, basaltic andesite/andesite. The clas-

Fig. 8 - (a) Zr-Nb-Y tectonomagmatic discrimination diagram (Meschede, 1986). (b) Zr-Ti-Y tectonomagmatic discrimination diagram (Pearce and Cann, 1973). (c) Th/Yb vs. Ta/Yb diagram used to differentiate subduction-related basalts from mid-ocean ridges and intraplate basalts (Pearce, 1983). Rocks from oceanic and continental arcs usually plot in different areas of the same diagram. (d) Zr/Y vs. Y diagram used to discriminate arc-related rocks in continental versus oceanic settings (Pearce, 1983). Abbreviations: VAB, volcanic arc basalts; CAB, calc-alkali basalt; LKT, low-potassium tholeiites; OFB, ocean-floor basalts; WPB, within-plate basalts; WPA, within-plate alkali basalts; WPT, within-plate tholeiites; P-MORB, plume-type mid-ocean ridge basalts; N-MORB, normal-type mid-ocean ridge basalts. Symbols: circles, Kopaonik metabasites (this work); squares, Triassic metabasites from Greece (from Pe-Piper, 1998).
sification of the igneous protolith of KP-42 is more contro-
versial: using the immobile elements diagrams of Winches-
ter and Floyd (1977) it would classify as a subalkaline 
basalts, nonetheless this rock contains about 7 wt.% of 
Na2O + K2O. Probably, this discrepancy is due to an intro-
duction of alkalies during metamorphism, possibly related 
to the upper Oligocene emplacement of the Kopaonik I-
type granitoids.

The geochemical affinity of the igneous protoliths of the 
studied rocks, evaluated using discrimination diagrams 
based on the relative distributions of several immobile ele-
ments, is typical of igneous rocks produced at destructive 
plate margins (Fig. 8a and b). In particular, the Zr/Y and 
Ta/Yb ratios are indicative of a geochemical affinity compa-
rable to basaltic magmas generated at continental arc set-
tings (Fig. 8c and d).

**DISCUSSION**

The Kopaonik metabasites occur in the KMC as up to 
200 m thick folded and boudinaged bodies inside the 
metapelites, metadolomites and metastones of Late Tri-
assic age. Although the contacts between the lithologies 
in the KMC are marked by shear zones, the same deformation 
and metamorphic history provide indicate that all the 
metabasites represented the base of the metasedimentary 
successions, as suggested by the lacking in these rocks of 
dykes or stocks of magmatic rocks. In this picture, the age 
of the metabasites can be regarded as pre-Late Triassic in 
age, probably spanning from Permian to Middle Triassic. 
The structural analyses performed in the metabasites 
detects a deformation history consisting of four phases. The 
first two phases were characterized by folds associated to 
continuous foliation developed under upper greenschist-up-
per amphibolite facies metamorphism. The deformation his-
tory, as well as the metamorphic grade, in the metabasites 
are analogous to that detected in the others metamorphic 
rocks from KMC. In addition, the geochemistry of the 
metabasites indicates a geochemical affinity comparable to 
basaltic magmas generated at continental arc settings.

Fig. 9 - Sketch showing the different models proposed for the Permian-Triassic boundary. 1 Classical model (Collaku et al., 1992; Pamć et al., 1998; 2002; 
Dimitrijević, 2001; Karamata et al., 2000; Bortolotti et al., 2004; Saccani et al., 2004 and many others) where the calc-alkaline magmatism found in the 
Kopaonik area is interpreted as rifting-related by partial melting from small areas of enriched hydrous mantle, according to Pe-Piper (1998). 2) Models where 
the same calc-alkaline magmatism is connected with the Cimmerian orogenesis a) model proposed by Sengor et al., (1984) where Anatolian Cimmerian struc-
tures correspond with the peri-Rhodopan belt, within the western margin of the Serbo-Macedonian massif in the Eurasia Plate; b) model proposed by 
Stampfli et al. (2003) that suggests a location of the Cimmerian suture zone west of the Pindos Zone.
Their significance in the geodynamic history of the Eastern Tethys is problematic and different hypothesis can be proposed. In some reconstructions of the paleogeography of eastern Tethys, the Triassic time is characterized by a Lower Triassic rifting phase followed by a Middle to Upper Triassic phase of opening of an oceanic basin between the Adria and Eurasia continental Plates (e.g. Bortolotti et al., 2004 and quoted references). Rifting is associated to alkaline magmatism as recognized in the sub-ophiolite mélangé from Albania and Greece (e.g. Saccani et al., 2003). However the Triassic magmatism in Greece is characterized by a wide range of magmatic rocks showing contrasting geochemical affinity (Pe-Piper, 1998 and quoted references). Between them, rare shoshonites and calc-alkaline intermediate rocks have been found associated to the Pindos and Parnassos Units, both derived from the Adria continental margin. These rocks have been interpreted (Pe-Piper, 1998) as originated during the Triassic rifting phase by partial melting from small areas of enriched hydrous mantle, probably a consequence of preexisting subduction (Fig. 9.1). Therefore, the metabasites from KMC can be interpreted as originated during the rifting phase in the Triassic time, similar to analogous rocks found in Greece.

However, an alternative geodynamic interpretation can be proposed. In the Eastern Mediterranean, the paleogeographic reconstructions proposed for the Permo-Triassic time (e.g. Stampfli and Borel, 2004) includes a large oceanic area (PaleoTethys) characterized by a subduction zone in its northwestern side. This oceanic area closed in the Early Triassic leading to continental collision and rapid slab detachment producing subduction-related magmatism and related tectonics (Cimmerian orogenic event). At present, the Cimmerian structures are well exposed in the Turkey, but the outcrop continuity is interrupted northwestward by the northern Aegean sea. However, some authors (e.g. Sengor et al., 1984 and quoted references) correlated the Anatolian Cimmerian structures with those of the peri-Rhodopian belt, within the western margin of the Serbo-Macedonian massif located east of the Vardar Zone (Fig. 9.2a). By contrast, Stampfli et al. (2003) suggest a different location of the Cimmerian suture zone. According to the evidences provided by Vavassisi et al. (2000), these authors suggest a pristine location of the subduction zone related to Paleotethyan closure west of the Pindos Zone (Fig. 9.2b). Despite the different interpretations, all the reconstructions suggest that the western margin of the PaleoTethys basin was affected in Late Permain-Middle Triassic time by subduction of oceanic lithosphere with related magmatic activity. For instance, in the Menderes Massif from western Turkey, Lower-Middle Triassic calc-alkaline magmatism is reported as result of subduction connected with the closure of the PaleoTethys oceanic Domain (Koralay et al., 2001). A Middle Triassic subduction-related magmatism is also found in the Naxos Island (Reischmann, 1998), which is located near the on-land outcrops of the Vardar Zone. In this frame, the metabasites of the KMC can be regarded as related to the subduction in the northern corner of the Cimmerian subduction zone.

CONCLUSIONS

The Kopaonik area is characterized by metabasites that occur as up to 200 m thick bodies folded together with the metasedimentary rocks of Late Triassic age. Their original stratigraphic position was probably at the base of the metasedimentary succession. These rocks show a complex deformation history that includes two phases developed under upper greenschist-upper amphibolite P/T conditions followed by subsequent two phases characterized by very-low grade metamorphism. The geochemistry of the igneous protoliths of the studied rocks, evaluated using discrimination diagrams based on the relative distributions of several immobile elements, is indicative of an affinity comparable to basaltic magmas generated at continental arc settings. Two different, alternative geodynamic hypotheses of these rocks are proposed in this paper. The first hypothesis suggests an origin of these rocks during the Lower Triassic rifting phase by partial melting from small areas of enriched hydrous mantle, probably a consequence of preexisting subduction. By contrast, an origin of the metabasites from KMC of Kopaonik area as remnants of the Upper Permian-Middle Triassic calc-alkaline magmatism related to Cimmerian orogenesis is provided by the second hypothesis. At the present day, the available geological, geochemical and geochronological data cannot further discriminate between these two hypotheses.

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