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Compositional diversity of Eocene–Oligocene basaltic magmatism in the Eastern Rhodopes, SE Bulgaria: implications for genesis and tectonic setting

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Abstract

Basaltic magmatism occurred only rarely within the extensive Eocene–Oligocene volcanic activity in the Eastern Rhodope Mts., SE Bulgaria. The earliest mafic volcanism started at ca. 34 Ma with K-rich trachybasalts strongly enriched in large ion lithophile elements (LILE), particularly Ba, Sr, Pb, Th, and light rare earth elements (REE) relative to the high field strength elements (HFSE). They have high $^{87}Sr/^{86}Sr$ ratios (0.70688–0.70756), low $^{144}Nd/^{144}Nd$ (0.51252–0.51243), and very high $^{207}Pb/^{204}Pb$ (15.74–15.76) and $^{208}Pb/^{204}Pb$ (39.07–39.14) at low $^{206}Pb/^{204}Pb$ (18.72–18.73) ratios, reflecting high degrees of crustal contamination. Shoshonitic basalts and absarokites and calc-alkaline and high-K calc-alkaline magmas, which erupted between 33 and 31 Ma, have decreasing Sr isotope initial ratios from west (0.70825) to east (0.70647) at approximately constant $^{143}Nd/^{144}Nd$ isotopic compositions (0.51252–0.51243) and slightly decreasing $^{207}Pb/^{204}Pb$ (15.66–15.72) and $^{208}Pb/^{204}Pb$ (38.80–38.96) and increasing $^{206}Pb/^{204}Pb$ (18.73–18.90) in comparison with the trachybasalts. All these rocks are characterized by negative Nb–Ti and Eu anomalies. They resulted from different degrees of partial melting of enriched asthenosphere, and the magmas were later contaminated by the Rhodopian crust. The end of the magmatic activity (28–26 Ma) was marked by emplacement of alkaline dykes, spatially associated with metamorphic core complexes. They are characterized by low $^{87}Sr/^{86}Sr$ (0.70323–0.70338), high $^{144}Nd/^{144}Nd$ (0.51290–0.51289), and $^{206}Pb/^{204}Pb$ (18.91–19.02) at lower $^{207}Pb/^{204}Pb$ (15.52–15.64) and $^{208}Pb/^{204}Pb$ (38.59–38.87) ratios, consistent with an origin from a source similar to OIB-like European Asthenospheric reservoir contaminated by depleted mantle lithosphere.

The Eastern Rhodope Eo-Oligocene mafic magmatism formed as part of the prolonged extensional tectonics of the whole Rhodope region in Late Cretaceous–Paleogene time, similar to those in the U.S. Cordillera and Menderes Massif (Turkey). Initiation of extension is constrained by the formation of metamorphic core complexes, low-angle detachment faults, and supradetachment Maastrichtian–Paleocene sedimentary basins, intimately associated with 70–42 Ma granitoids and

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metamorphism which record mantle perturbation. The Eo-Oligocene stage started with block faulting, sedimentary basin formation, and extensive acid-intermediate and basic volcanism over the entire Eastern Rhodope area. The order of emplacement of the basalts from high-Ba trachybasalts through shoshonites, calc-alkaline and high-K calc-alkaline basalts, and finally to purely asthenospheric-derived alkaline basalts, with progressively decreasing amount of crustal component, reflects upwelling asthenospheric mantle. Most of the models proposed in the literature to explain extension and magmagenesis in the Rhodopes and the Mediterranean region cannot be applied directly. Critical evaluation of these models suggest that some form of convective removal of the lithosphere and mantle diapirism provide the most satisfactory explanation for the Paleogene structural, metamorphic, and magmatic evolution of the Rhodopes.

Keywords: Eastern Rhodopes; Bulgaria; Basalts; Extension; Metamorphic core complex

1. Introduction

Convergence between the Eurasian and African plates played a key role in controlling magmatism in the Balkan Peninsula. Since the Late Cretaceous, collision resulted in the formation of several subparallel southward-migrating magmatic belts with the youngest one being the present-day Aegean arc (Fyticas et al., 1984). During the Late Eocene–Oligocene, magmatic activity occurred in the Macedonian–Rhodope–North Aegean region (Harkovska et al., 1989; Marchev and Shanov, 1991) (Fig. 1). The magmatic belt extends to the NW into Macedonia and Serbia, crossing the Vardar zone (Bonchev, 1980; Cvetkovic et al., 1995), and continues to the SE in the Thracian Basin and Western Anatolia (Yilmaz and Polat, 1998; Aldanmaz et al., 2000). K–Ar dating of the volcanic rocks in the northern Dinarides (Pamic et

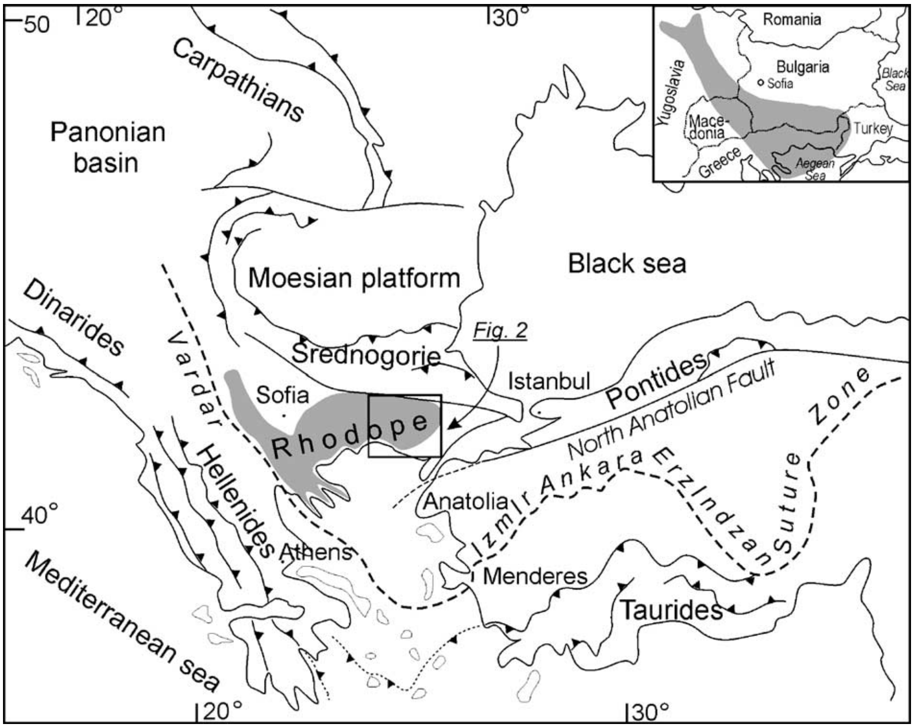

Fig. 1. The position of the Rhodope Massif with respect to the main tectonic units of southeastern Europe. Shaded area in the inset shows distribution of Eo-Oligocene magmatic rocks.

al., 2000) suggests that this volcanism extends even further to the NW and is connected with the Periadriatic tonalite suite (von Blanckenburg and Davies, 1998) and dyke swarms of the NW Alps (Venturelli et al., 1984; von Blanckenburg and Davies, 1995).

In Bulgaria, Late Eocene–Oligocene magmatism was confined mostly to the Rhodope Massif, a metamorphic complex forming a significant part of the Balkan Peninsula (Fig. 1). The Rhodope Massif has been regarded as a Precambrian or Variscan stable continental block (Bonchev, 1971, 1988), but recent work has demonstrated that it was actively involved in Alpine convergent tectonic processes (Burchfiel, 1980; Ivanov, 1989; Burg et al., 1990, 1995). The Rhodope Massif is bounded to the east by the Circum Rhodope Belt and to the north by the Maritsa fault which separates the massif from the Srednogorie Zone. To the west, the Rhodope Massif is separated from the Serbo-Macedonian Massif by a tectonic contact interpreted as a middle Miocene–late Pliocene Strymon detachment fault (Dinter and Royden, 1993). In recent works, the two massifs have been combined and broadly termed the "Rhodope Massif" (Burg et al., 1995; 1996; Ricou et al., 1998; Jones et al., 1992; Lips et al., 2000).

Petrological studies in the past two decades have documented that mafic igneous rocks are relatively common in the eastern Rhodope Massif (Yanev et al., 1989; 1998a; Marchev et al., 1998a, b). Mafic (<53 SiO2) magmas have been identified by Marchev et al. (1998a) as: (1) orogenic Mg-rich (absarokites) and high-Al basalts with variable enrichment of K belonging to the calc-alkaline, high-K calc-alkaline, and shoshonitic series; and (2) as within-plate basalts. In this paper, we add a previously unrecognised group of rocks, with subduction-related trace element signatures, but with extremely high Ba and Sr contents, which increases the compositional diversity of the orogenic magmas.

Major issues of the petrogenesis of the orogenic rocks in the region are the nature of the sources of the mafic magmas and the role of crustal contamination in controlling their compositions. Two different hypotheses have been suggested. Most workers (Pe-Piper et al., 1998; Nedialkov and Pe-Piper, 1998; Francalanci et al., 1990; Yanev et al., 1998a) consider that enriched lithospheric mantle is the source. Other authors (Marchev et al., 1998a) suggest that they originated by contamination of asthenospheric-derived mantle melts by crustal material.

The tectonic significance of the Paleogene Rhodope volcanism is also the subject of debates. Some authors suggest a relationship with subduction (Barr et al., 1999), whereas others (Jones and Robertson, 1991; Clift, 1992; Yanev and Bakhneva, 1980; Yanev et al., 1989, 1998a) argue that the volcanic activity was collision-related. Most workers, however, agree that the orogenic magmas associated with dyke swarms (Kharkovska, 1984; Harkovska et al., 1998; Marchev et al., 1998a) reflect post-collisional tectonic extension. The coexistence of Late Cretaceous–Early Eocene (70–42 Ma) granitoids (Peytcheva et al., 1998a; Christofides et al., 2001), Late Eocene–Early Miocene calc-alkaline (39–19 Ma, Lilov et al., 1987; Yanev et al., 1998b; Christofides et al., 2002), and Middle Oligocene within-plate magmatic products (Marchev et al., 1998b) defines the Rhodopes as a key area for understanding the relationships between magmatism, crustal extension, and metamorphic core complex formation.

The purpose of this work is to provide an updated review of the mafic volcanic rocks from the Eastern Rhodopes. We describe their chemistry and spatial and temporal distribution in order to address the origin and particularly the source region characteristics of the basalts, using published and newly obtained trace element and isotopic data. We integrate these data with the Late Alpine tectono-metamorphic and Late Cretaceous–Early Eocene magmatic history in order to understand the tectonic setting of the magmatism. We suggest that magmatic, metamorphic, and tectonic processes in this period require a long-lasting asthenospheric upwelling in the Rhodope region.

2. Geology of the Eastern Rhodope zone

2.1. Basement rocks

Basement rocks of the Eastern Rhodopes crop out in the Kesebir and Biala Reka metamorphic dome complexes (Fig. 2). Two major tectonostratigraphic complexes have been recognised on the basis of composition and tectonic setting of the metamorphic rocks: a Gneiss–migmatite complex and Variegated complex (Haydoutov et al., 2001), which correspond

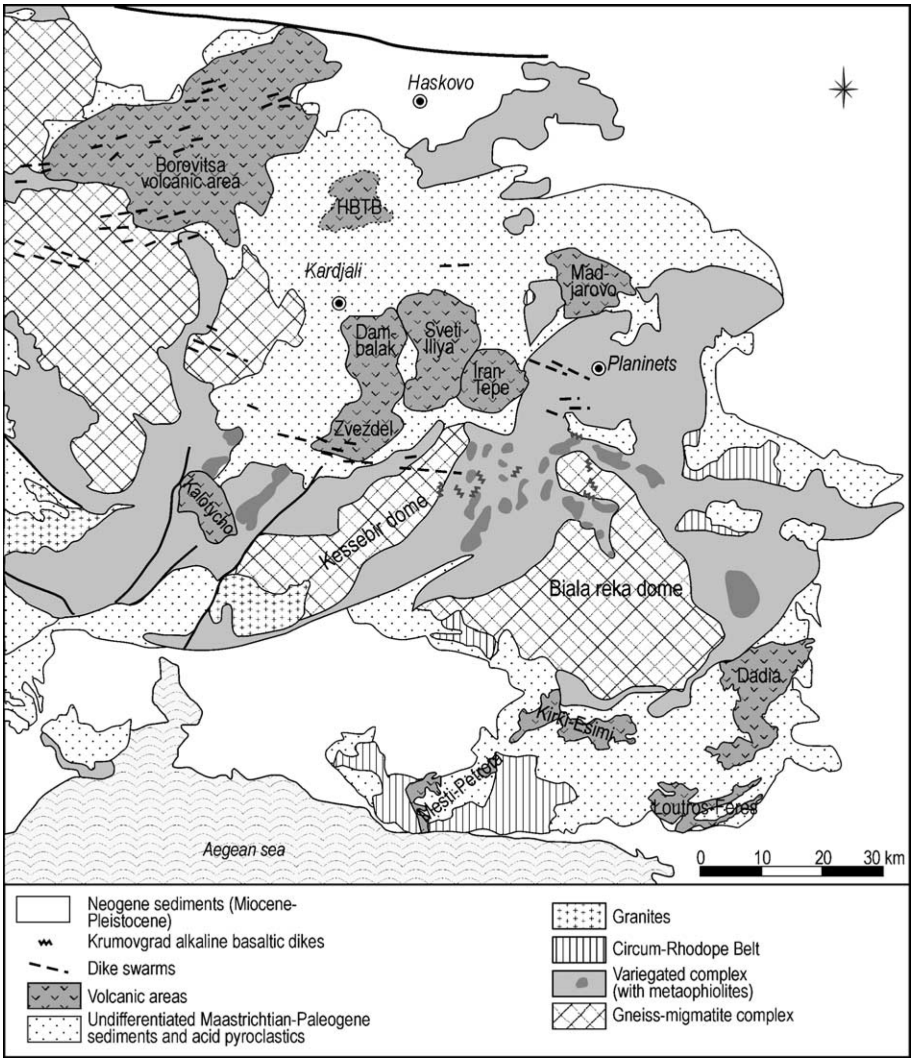

Fig. 2. Schematic geological map of the Eastern Rhodope showing the metamorphic dome structures and the major volcanic areas and dyke swarms. Compiled from Ricou et al. (1998), Yanev et al. (1998a), and Marchev et al. (1998a,b).

to the continental and mixed units of Ricou et al. (1998), respectively. In Greece, the Variegated complex is known as the Kimi complex (Mposkos and Krohe, 2000; Krohe and Mposkos, 2002). Stratigraphically, the lower Gneiss–migmatite complex is regarded as the “core” of the Kesebir and Biala Reka metamorphic domes (Burg et al., 1996; Ricou et al., 1998; Bonev, 2002). In Greece, these are known as the Kardamos and Kechros complexes, respectively (Mposkos and Krohe, 2000; Krohe and Mposkos, 2002). The Gneiss–migmatite complex is dominated by metagranites, migmatites, and migmatized gneisses overlain by a series of pelitic gneisses, marbles, and amphibolites. Eclogites and eclogite amphibolites have been described in the Kechros complex in Greece (Mposkos and Krohe, 2000). Zircons from metagranites from Biala Reka yield Late Paleozoic ages (ca. 320–305 Ma, Peytcheva and Von Quadt, 1995; 301±4 Ma, Carrigan et al., 2003). Similar Rb–Sr ages (334±3 and 328±25 Ma) have been obtained from metapegmatites in Greece (Mposkos and Wawrzenitz, 1995) and from metagranites in the Kesebir dome, Bulgaria (Peytcheva et al., 1998b). However, Carrigan et al. (2003) reported ages ranging 660–2500

Ma for inherited zircon cores in the Biala Reka metagranites, suggesting derivation of the Gneiss–migmatite complex from Variscan or even Proterozoic continental basement.

The overlying Variegated complex consists of a heterogeneous assemblage of metasedimentary rocks and ophiolite bodies (Kozhoukharova, 1984; Kolcheva and Eskenazy, 1988). Metamorphosed ophiolitic peridotites and amphibolitised eclogites are intruded by metamorphic gabbros, gabbronorites, plagiogranites, and diorites of boninite and arc-tholeiitic affinities. The Variegated complex probably originated in an island-arc setting (Kolcheva and Eskenazy, 1988; Haydoutov et al., 2001). U–Pb zircon dating of a gabbro from Biala Reka yields a Late Neoproterozoic age (572±5 Ma) for the core and Hercynian age (~300–350 Ma) for the outer zone (Carrigan et al., 2003). In the Eastern Rhodopes, the Variegated complex is tectonically overlain by phyllites, albite gneisses, marbles, and mafic and ultramafic metaigneous rocks of Jurassic–Early Cretaceous age, traditionally assigned to the so-called Circum–Rhodope Belt (Kockel et al., 1977).

Crustal thickness beneath the Rhodope Massif has been examined in many studies (Dachev and Volvovsky, 1985; Shanov and Kostadinov, 1992; Riazkov, 1992; Boykova, 1999; Papazachos and Skordilis, 1998). Seismic data indicate a marked reduction of crustal thickness from >50 km in the NW part down to 25 km under the dome structures and thickening to 32–35 km under the Variegated complex of the Eastern Rhodopes.

2.2. Alpine tectonomagmatic features

Several authors have suggested that the Rhodope Massif suffered Variscan and even Precambrian metamorphism (Kozhoukharov et al., 1988; Zagortchev, 1993), whereas others deny the pre-Alpine metamorphic evolution of the massif (Burg et al., 1990, Dinter, 1998; Barr et al., 1999). However, all workers agree that in Alpine times, the Rhodope Massif was characterized by a complicated tectono-metamorphic evolution. Ivanov (1989) and Burg et al. (1990) distinguish two phases in the evolution of the Rhodopes. The first compressional phase caused large-scale, south-vergent thrusting and amphibolite-facies metamorphism. The following extensional phase involved tectonic erosion of the thrust complex and formation of detachment and synthetic faults.

Available age data suggest the existence of Early Cretaceous or older high-pressure metamorphism in the Eastern Rhodopes. Migmatites from the Variegated complex have been dated at 159±19 Ma, using Rb–Sr whole-rock methods (Peytcheva et al., 1998b). A Sm–Nd isochron age of 119±3.5 Ma on a spinel-garnet pyroxenite of the Kimi complex in Greece has been interpreted to reflect an Early Cretaceous high-pressure (16 kbar and 750–800 °C) subduction-related metamorphic event (Wawrzenitz and Mposkos, 1997; Mposkos and Krohe, 2000). The presence of relict diamond and coesite in eclogitic and metapelitic garnets from the UHP Kimi metamorphic complex suggests prior subduction to depths of ca. 220 km (70 kbar) (Mposkos and Kostopoulos, 2001). A Rb–Sr isochron age of 65.4±0.7 Ma from an undeformed metamorphic pegmatite was interpreted by Mposkos and Wawrzenitz (1995) as a minimum age of migmatization. Similar ages but a different metamorphic evolution in the same area were presented by Liati et al. (2002) on the basis of U–Pb SHRIMP studies. These authors interpreted an age of 117.4±1.9 Ma on a garnet-rich mafic rock to reflect the age of magmatic crystallization of the protolith, which was later metamorphosed at 73.5±3.4 Ma. According to these authors, the last retrograde stage of this metamorphism (ca. 500 °C, 5 kb) occurred at 61.9±1.9 Ma.

The onset of extension seems to have occurred in the Late Cretaceous (ca. 70 Ma). Metamorphism in the Kimi complex (73.5±3.4 Ma) closely coincides with the age (70 Ma) of granitoids intruded in the Variegated complex in Bulgaria. Extension led to the formation of the Biala Reka and Kessebir metamorphic core complexes (Burg et al., 1996; Bonev, 2002), low-angle detachment faulting (Mposkos and Krohe, 2000; Krohe and Mposkos, 2002; Bonev, 2002), and sedimentary basins (Boyanov and Goranov, 2001). The upward transition from Maastrichtian–Paleocene colluvial–proluvial to marine sediments in the area north of the two dome structures and the presence of several unconformities in the overlying Late Eocene strata (Goranov and Atanasov, 1992; Boyanov and Goranov, 1994, 2001) suggest a supradetachment evolution of these basins, associated with surface uplift and exhumation of UHP metamorphic lithologies. The timing of unroofing of the Biala Reka and Kessebir core complexes—65–42 Ma for the upper (Variegated) complex and 42–30 Ma for the lower (Gneiss–migmatite) unit (Krohe and Mposkos, 2002)—is consistent with the sedimentary record, new data suggesting revised timing of the beginning and end of exhumation, respectively, to ca. 70 and 25 Ma. Several ages between 42 and 35 Ma, obtained from Rb–Sr isochrons (Peytcheva, 1997) and $^{40}Ar/^{39}Ar$ dating of mica and biotite (Lips et al., 2000; Marchev et al., 2002b; Mukasa et al., 2003), suggest that unroofing of the Variegated and gneiss–migmatite complexes in Biala Reka and Kessebir was accompanied by cooling below 350 °C. A similar Rb–Sr age (37±1.0) of white mica from a mylonitic orthogneiss from Biala Reka (Kechros) has been interpreted by Wawrzenitz and Mposkos (1997) as a minimum age of high-pressure metamorphism.

2.3. Eo-Oligocene Eastern Rhodope magmatic zone

During Late Eocene–Oligocene time, the Eastern Rhodope zone was the locus of extensive magmatic activity which followed development of the Late Eocene–Oligocene extensional basins (Harkovska et al., 1989; Boyanov and Goranov, 2001). The locations of discrete volcanic centres in the Rhodope Massif are shown in Fig. 2. Magmatic rocks from the Central and Eastern Rhodopes differ significantly in composition, demonstrating a strong dependence on crustal thickness (Marchev et al., 1989, 1994; Marchev and Shanov, 1991). The Central Rhodopes magmatic rocks show considerable compositional variation, intermediate, and acid magmas predominating over basic types (Marchev and Shanov, 1991; Marchev et al., 1998a; Yanev et al., 1989, 1998a).

Volcanic products in the Eastern Rhodopes appear at Loutros-Fere, Dadia, Kirki-Esimi, Mesti-Petrota, Kaloticho, Iran Tepe, Zvezdel-Dambuluk, Sveti Ilia, Madjarovo, and Borovitsa. In the exhumed metamorphic domes, igneous activity is represented by dyke swarms of predominantly felsic, intermediate, or rarely bimodal (basalt-rhyolite) compositions. A large number of intrusive bodies hosted in the metamorphic and volcanic rocks occur throughout the whole area (Del Moro et al., 1988; Mavroudchiev et al., 1993).

Mafic volcanic rocks are reported from several volcanic centres in the Bulgarian part of the Eastern Rhodopes and also from Kotili-Zlatograd in Greece (Elefteriadis, 1995; Yanev et al., 1989). While lacking modern chemical and isotopic data, mafic intrusive rocks are more common and better studied (Christofides et al., 1998; Del Moro et al., 1988). Below, we summarise the geology and petrology of volcanic centres from the Bulgarian Eastern Rhodopes from which we have studied samples of mafic rocks.

3. Age and occurrence of basaltic rocks

3.1. High-Ba trachybasalts (HBTB)

These rocks are described here for the first time in the Eastern Rhodopes. They are predominantly epiclastic overlying Priabonian flysch-like marls and coal-bearing formations north–northwest of Kardjali (Fig. 2). The HBTB are located in an area of about 35–40 km^2 east of the Borovitsa volcanic area and seem to correspond to the lowermost part of the Borovitsa precaldera complex (see below). We have obtained 3 K–Ar dates on whole-rock samples from these rocks that range from 33.1±1.3 to 29.2±1.1 Ma. Although the ages are rather scattered, the epiclastic nature and position above the Priabonian sediments suggest that they are among the oldest known volcanic rocks in the Eastern Rhodopes.

3.2. Borovitsa volcanic area

The Borovitsa volcanic area is located in the NE part of the Eastern Rhodopes and is a large volcanic complex (1150 km^2; Fig. 2) with a large (30×15 km) caldera structure (Ivanov, 1972; Yanev et al., 1998a; Singer and Marchev, 2000). K/Ar and $^{40}Ar/^{39}Ar$ age determinations indicate that volcanism was active between 34.5 and 31.7 Ma (Singer and Marchev, 2000). Precaldera lavas erupted through three major E–W-trending regional dyke swarms (Ivanov, 1972). Volcanic rocks in the area comprise a typical shoshonitic suite (Marchev, 1985), dominated by intermediate (shoshonites, latites) to silicic (quartz-latites and rhyolites) lavas. The overall silica contents range from 49 to 78 wt.%. $^{87}Sr/^{86}Sr$ initial ratios for Borovitsa volcanic rocks range from 0.70790 to 0.71199 and $^{143}Nd/^{144}Nd$ range from 0.51244 to 0.51234, with the postcaldera rocks exhibiting higher

$^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ (Marchev et al., unpublished data). The range of Pb isotopes values is very narrow (lower $^{206}Pb/^{204}Pb$=18.91–18.63, $^{207}Pb/^{204}Pb$=15.68–15.56, $^{208}Pb/^{204}Pb$=38.97–38.86). Mafic lavas and dykes are comparatively rare and are represented by shoshonitic and ultrapotassic basalts and absarokites. For this study, we have chosen an absarokite (B84/1) and a high-Al shoshonitic basalt (2260) from the precaldera stage and an ultrapotassic basalt (41a) from the postcaldera stage. Previously, the absarokite was studied by Ivanov (1978), Yanev et al. (1989), and Marchev et al. (1998a), and data for the ultrapotassic basalt was reported by Marchev et al. (1998a).

3.3. Madjarovo volcano

The Madjarovo volcano is situated about 30 km E–SE of Borovitsa. It covers an area of about 120 km^2 in an E–W-trending sedimentary basin (Fig. 2). $^{40}Ar/^{39}Ar$ dating places the volcanism in the Oligocene between 32.7 and 32.2 Ma (Marchev and Singer, 2002). The volcanic activity was fed by a radial dyke swarm producing a 600–700 m thick shield volcano (Ivanov, 1960). The volcanic rocks are high-K calc-alkaline to shoshonitic basic-intermediate to silicic varieties (51.2–67.0 wt.% SiO_2), with K content increasing during the evolution of the volcano. $^{87}Sr/^{86}Sr$ initial ratios range from 0.70775 to 0.70861, and $^{143}Nd/^{144}Nd$ ranges from 0.512454 to 0.512376, with decreasing $^{87}Sr/^{86}Sr$ and increasing $^{143}Nd/^{144}Nd$ ratios in the last stage of volcanic evolution (Marchev et al., 2002a and unpublished data). The range of Pb isotopes values of the intermediate and acid volcanics is very narrow ($^{206}Pb/^{204}Pb$=18.78–18.68, $^{207}Pb/^{204}Pb$=15.68–15.67, $^{208}Pb/^{204}Pb$=38.88–38.77). Basaltic samples chosen for this study (M89-107, M88-251, and M86-2) are representative of different stratigraphic units.

3.4. Zvezdel volcano

The Zvezdel volcano lies SE of the Borovitsa volcanic area and southwest of the Madjarovo volcano. $^{40}Ar/^{39}Ar$ dates from two pyroclastic flows (32.0 and 31.17 Ma) reveal that volcanism occurred for ca. 1 Ma (Singer and Marchev, 2000 and unpublished data). An E–W-trending dyke swarm in the middle of the volcano seems to be the main feeder structure. Volcanic rocks are mostly high-K calc-alkaline with subordinate calc-alkaline and shoshonite varieties ranging from basalts to andesites and latites (SiO_2 49–65 wt.%). Basalts are comparatively rare (Yanev et al., 1989; Nedialkov and Pe-Piper, 1998) and tend to occur at the end of volcanic activity. Initial $^{87}Sr/^{86}Sr$ ratios of Zveldel lavas range from 0.70713 to 0.70737, and $^{143}Nd/^{144}Nd$ range from 0.512410 to 0.512397. Two analyses of Pb isotopes give a range $^{206}Pb/^{204}Pb$=18.86–18.84, $^{207}Pb/^{204}Pb$=15.69–15.67, and $^{208}Pb/^{204}Pb$=38.95–38.88. Here we include four basalts from the lava flows (samples 2014, 25G, and Zd95-4) and one from the late dykes (Zd01-12).

3.5. Bimodal rhyolite-mafic dyke field from the north edge of Biala Reka dome

Rhyolitic to rhyodacitic dykes oriented WNW–ESE and NE–SW are exposed to the south of Madjarovo and east of the Iran Tepe volcanoes (Fig. 2). In the east, these intrude the Biala Reka dome metamorphic rocks, while to the west, the dykes are emplaced in Paleogene sedimentary cover. Rare mafic dykes and circular bodies intrude the largest, easternmost rhyolite body in the vicinity of Planinets (Fig. 2) (Mavroudchiev, 1964). We have analysed sample Bz18 from a circular body, which we consider to be an absarokite (Marchev et al., 1998a). The $^{40}Ar/^{39}Ar$ age date for the host rhyolite is 32.88±0.23 (Marchev et al., 2002b).

3.6. Krumovgrad alkaline basalts (KAB)

Basic subvolcanic bodies (Mavroudchiev, 1964; Marchev et al., 1997, 1998b) crop out in the Biala Reka and Kessebir domes over an E–W-oriented area of 1000 km^2. Three major fault systems, striking E–W, N–S, and NW–SE, accommodate the dykes (Marchev et al., 1997). In general, they are subvertical, but some are subparallel to the foliation of the host metamorphic rocks. Hereafter, dykes are referred to as Krumovgrad alkaline basalts (KAB). They are of Middle Oligocene age (28–26 Ma, Marchev et al., 1997) and are the most primitive Paleogene magmatic rocks found in the Rhodope region.

4. Analytical techniques

Major elements of most samples were determined by XRF at Washington State University (Department of Geology) and the University of Lausanne (samples TO1-1, T01-3, ZD01-12). Major elements of samples 2260a, M98-107, and M88-251 were determined by standard wet chemical analyses at the Geological Institute of the Bulgarian Academy of Sciences (GI BAS). Trace-element (Sc, V, Ni, Cr, Zr, Y, Nb, Cu, Zn, Pb, Th, Ba, Rb, and Sr) analyses were determined on the fused glass discs by XRF, and REE, Ba, Th, Nb, Y, Hf, Ta, U, Pb, and Rb were determined by ICP-MS. Co was measured by AA at GI BAS, XRF (University of Lausanne), and ICP-MS (sample 96-26a).

$^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic analyses from Borovitsa samples were undertaken at the USGS, Denver, and samples M86-2 and M88-251 were analysed at SURRC, East Kilbride. Sr isotopes were measured with a VG Micromass 54R mass spectrometer in Denver and VG Isomass 54E mass spectrometer in East Kilbride. HBTB and Zvezdel samples, and M89-107 and Bz18, were analysed at Royal Holloway University of London with a VG354 thermal ionisation mass spectrometer. An appropriate age correction has been applied depending on the age of the magmatic centres. Sr isotopic results for the KAB are those reported in Marchev et al. (1998b).

Lead isotopic compositions of HCl–HNO_3-leached and HNO_3–HF-dissolved whole rock samples were determined on a Finnigan MAT 262 mass spectrometer at the University of Geneva. Procedural Pb blanks during this study were less than 120 pg and were therefore negligible compared to values measured in the samples.

5. Petrography of the Eastern Rhodope basic rocks

5.1. High-Ba trachybasalts (HBTB)

The HBTB carry abundant large phenocrysts of clinopyroxene and subordinate olivine (entirely altered), biotite, plagioclase, anorthoclase, biotite, Ti-magnetite, and apatite. The groundmass may contain large poikilitic sanidine that encloses plagioclase, olivine, clinopyroxene, and Ti-magnetite. Petrographic character is generally comparable to that of orogenic lamproites from Central Turkey (Francalanci et al., 2000), the Betic Cordillera, SE Spain (Turner et al., 1999), and the Tibetan plateau (Chung et al., 2001; Miller et al., 1999), but differ from those in that phenocrysts of plagioclase are present.

5.2. Shoshonitic basalts (Borovitsa and Madjarovo)

The shoshonitic (SHO) basic rocks contain phenocrysts of plagioclase, clinopyroxene, olivine, orthopyroxene (usually coated by clinopyroxene), and Ti-magnetite. Rare xenocrysts of biotite occur in the lava flows that show mixing phenomena. The groundmass consists of microlites of plagioclase, clinopyroxene, orthopyroxene, pigeonite, olivine and sanidine, and accessory apatite and Ti-magnetite.

5.3. Absarokites (ABS) and ultrapotassic (UK) basalt

Absarokites (ABS) are characterized by abundant clinopyroxene and olivine phenocrysts (sometimes completely altered), and Ti-magnetite, apatite, and anorthoclase as groundmass. Minor biotite in the groundmass of Bz18 is the only water-bearing mineral. Although fresh leucite is not present, subhedral leucite pseudomorphs have been observed in the groundmass of B84/1. In the well-crystallized Bz18, the groundmass consists of poikilitic sanidine which encloses clinopyroxene, olivine, titanomagnetite, and biotite. UK basalt 41a has similar phenocrysts with a larger amount of biotite and sanidine in the groundmass.

5.4. High-K calc-alkaline (HKCA) and calc-alkaline (CA) basalts (Zvezdel volcano)

The high-K calc-alkaline (HKCA) and calk-alkaline (CA) basalts from Zvezdel are mainly porphyritic, with phenocrysts of plagioclase, clinopyroxene and olivine, and rare orthopyroxene. The groundmass is holocrystalline to glassy with abundant plagioclase, ortho- and clinopyroxene, Ti-magnetite, and rare olivine.

5.5. Krumovgrad alkaline basalts (KAB)

The KAB contain phenocrysts of olivine, Mg-rich and Fe–Na-rich clinopyroxenes, biotite, and microphenocrysts of plagioclase. Megacrysts of sanidine, olivine, clinopyroxene, and kaersutite are also present. The groundmass is composed of clinopyroxene and amphibole microlites and Ti-magnetite grains with abundant interstitial analcite and sanidine. The alkaline basalts contain small xenoliths of spinel lherzolites and xenocrysts of high-Mg olivine. A large number of cumulus clinopyroxenites, olivine and hornblende-bearing clinopyroxenites and hornblendites, and crustally derived xenoliths are also present.

6. Geochemical characteristics

6.1. Major and compatible trace elements

The HKCA, SHO, and HBTB are characterized by elevated SiO_2 contents (49.6–54 wt.%) and lower MgO (5.5–3.2 wt.%) in comparison to the KAB (46–48 wt.% SiO_2 and 10.1–8.4 wt.% MgO). Similarly, their Ni (34–3 ppm) and Cr (132–8 ppm) are lower than those in the KAB (105–40 ppm Ni; 170–30 ppm Cr). The ABS and the UK basalts contain about 50 wt.% SiO_2 and have higher MgO (8.8–6.0 wt.%), Ni (110–20), and Cr (460–110 ppm) than the HKCA, SHO, and HBTB, but lower MgO, comparable Ni, and higher Cr content compared to KAB, reflecting clinopyroxene accumulation. HBTB are the most Mg-poor (4.7–3.1 wt.%) basaltic rocks in the Eastern Rhodopes (Table 1).

The HKCA, SHO, HBTB, and KAB rocks have higher Al_2O_3 (16.5–19.5 wt.%) in comparison with the ABS-UK group (11.5–13.7 wt.%). The K_2O contents (Fig. 3) of SHO, HBTB, and ABS are in the range 2.7–4.6 wt.%. In the Zvezdel HKCA basalts, K_2O is lower (1.3–2.8 wt.%), whereas the UK Borovitsa basalt (41a) has the highest K_2O content (5.3 wt.%). K_2O/Na_2O ratios vary between 0.4 and 3.4.

6.2. Incompatible trace elements

Incompatible trace element characteristics are illustrated in mantle-normalised multielement diagrams in Fig. 4. The HKCA, SHO, HBTB, and ABS-UK basalts show similar incompatible trace element patterns with enrichment in large-ion lithophile elements (LILE) (K, Rb, Ba, and Pb) and depletion of high field strength elements (HFSE) (Nb and Ti). These are typical characteristics for magmas from convergent margin tectonic settings (Saunders et al., 1980). An increase in K from Zvezdel CA-HKCA basalts to Madjarovo and Borovitsa SHO and ABS-UK basalts is accompanied by a general enrichment in Rb, Ba, Th, as well as in Nb, La, Pb, and P.

The HBTB are distinguished from the other orogenic lavas by far more extreme enrichment in incompatible trace elements, particularly Ba, Sr, Pb, U, and Th at comparable contents of K and Rb to the shoshonitic basalts.

Incompatible trace element patterns of the KAB differ from those of the other volcanic rocks of the Eastern Rhodopes (Marchev et al., 1998b). They have high Nb abundances and a negative anomaly in Pb, which are typical features of within-plate basalts (Sun and McDonough, 1989). Compared to typical oceanic island basalts (Fig. 5), however, they show elevated incompatible elements (e.g. K, Rb, Ba, Th, Nb, and La) at similar HFSE.

The REE patterns of Eastern Rhodope basalts are shown in Fig. 6. The KAB and HBTB have steeper REE patterns than HKCA, SHO, and ABS-UK basalts, although HREE of all groups are similar, suggesting a genetic relationship. An important feature of all the basalts, except the KAB group, is that they show negative Eu anomalies.

6.3. Sr and Nd isotopic data

New Sr and Nd isotopic analyses for various mafic rocks from the Eastern Rhodopes, along with the results from Marchev et al. (1998a,b) are given in Table 2 and illustrated in Fig. 7. The HBTB show the largest range ($^{87}Sr/^{86}Sr$=0.70688–0.70756; $^{143}Nd/^{144}Nd$=0.51252–0.51243). $^{87}Sr/^{86}Sr$ value of the SHO basalt 2260 from Borovitsa is slightly higher (0.70823), but its Nd isotope ratio is identical to the lowest value of the HBTB (0.51244). The Madjarovo HKCA and SHO basalts have $^{87}Sr/^{86}Sr$ ratios intermediate (0.70789–0.70792) between HBTB and the Borovitsa SHO and overlapping values for $^{144}Nd/^{144}Nd$ (0.51246–0.51245). Zvezdel HKCA basalts are displaced to lower $^{87}Sr/^{86}Sr$ ratios (0.70713–0.70722) and slightly lower $^{143}Nd/^{144}Nd$ (0.51244–0.51243) ratios compared to Madjarovo basalts, but they are still in the range of the HBTB.

Table 1
Major and trace element data of mafic volcanic rocks from the Eastern Rhodopes

Sample no.	HBTB						ABS		UK basalt	SHO basalts	
	96-26a	96-26b	T01-1	T01-3	96-28	96-27b	Bz-18 Planinets	B84/1 Borovitsa	41a Borovitsa	2260a Borovitsa	M89-107 Madjarovo
SiO_2	50.98	51.76	51.79	52.43	53.68	53.76	48.24	50.22	52.16	49.60	51.21
TiO_2	0.91	0.78	0.79	0.74	0.71	0.72	0.89	0.87	0.78	0.69	1.09
Al_2O_3	18.47	18.01	16.65	19.44	17.78	17.82	11.96	11.36	14.15	16.70	18.37
Fe_2O_3			7.75^T	5.97^T						4.23	3.11
FeO	7.10^T	7.33^T			6.60^T	6.93^T	8.21^T	27.05^T	7.18^T	4.80	4.71
MnO	0.20	0.16	0.12	0.11	0.12	0.19	0.19	0.18	0.14	0.14	0.18
MgO	4.74	4.01	3.14	3.15	3.36	3.42	8.75	7.46	6.06	5.17	5.06
CaO	8.64	8.58	6.14	6.80	6.04	7.53	10.82	11.15	7.97	8.31	9.00
Na_2O	2.90	3.27	5.42	3.22	3.33	3.74	1.78	3.00	1.60	3.15	3.06
K_2O	3.34	3.74	3.33	4.38	4.63	3.97	3.80	3.36	5.48	3.10	2.66
P_2O_5	0.58	0.87	0.79	0.61	0.70	0.70	0.75	1.39	0.99	0.73	0.38
H_2O^-										0.28	
LOI	2.14	1.49	3.63	2.18	3.06	1.22	4.61	3.95	3.50	2.92	0.88
Total	100.01	100.00	99.54	99.02	100.01	100.00	91.79	92.94	92.84	99.82	99.71
Sc	23	24	8	7	19.7	24.3	47	41	33	20	30
V	219	202	227	174	181	314	216	198	195	211	201
Cr	28	13	21	19	16	23	313	460	110	20	132
Co	28		16	17				30	24	24.6	29.0
Ni	14	8	11	11	10	10	51	107	22	6	34
Zn	76	71	64	91	72	71	89	64	68	69	69
Cu	44	42	44	70	50	128	124	82	40	24	36
Pb	76	115	62	86	92	95.73	22	22.53	39	29	29
Zr	177	257	180	262	290	278	104	144	151	145	161
Hf	4	5	7	12	5.93	5.57	2.60	3.04	3.05	3.35	3.74
Nb	15.31	12.78	10	21	12.88	12.04	5.78	11.29	9.65	11.75	13.8
Ta	1.3	0.75			0.8	0.74	0.39	0.55	0.50	0.65	0.94
U	6.2	7.38	6	2	10.12	9.79	3.21	1.14	2.61	1.96	4.7
Y	20	27	27	27	26.63	26.08	17.72	21.85	22.47	29.13	24
Th	28.7	29.38	20	27	29.78	26.83	6.63	6.41	12.24	6.87	11.15
Rb	88	91	111	187	152	114	344	85	251	230	108
Cs		3.2			6.79	3.52	6.5	271	1.99	12.99	143
Sr	1221	1721	890	1364	1901	1603	400	633	647	637	568
Ba	3723	4767	2725	2348	4090	4043	1581	1475	2850	1778	1218
La	54.6	75.56	44	60	62.23	54.07	13.10	20.42	29.31	26.78	23.40
Ce	88.6	127.12	72	82	95.69	89.64	28.33	40.74	54.11	53.19	45.41
Pr	8.7	13.53			10.33	9.56	3.67	5.04	5.93	6.33	5.37
Nd	37.2	51.74	35	40	39.73	36.71	16.74	21.40	23.52	26.65	22.48
Sm	7.2	9.7			7.85	7.68	4.46	5.34	5.42	6.28	5.72
Eu	1.83	2.47			2.18	2.07	1.23	1.39	1.44	1.72	1.46
Gd	5.5	7.66			6.51	6.6	4.00	4.52	4.53	5.84	4.84
Tb	0.8	1.03			0.91	0.92	0.61	0.71	0.70	0.92	0.80
Dy	4.4	5.39			5.02	5.04	3.45	3.99	3.92	5.26	4.66
Ho	0.94	0.99			0.96	0.95	0.67	0.76	0.74	1.01	0.90
Er	2.4	2.56			2.56	2.54	1.68	2.14	2.08	2.74	2.50
Tm	0.3	0.36			0.37	0.36	0.23	0.29	0.29	0.39	0.34
Yb	2.2	2.2			2.3	2.2	1.43	1.82	1.81	2.50	2.05
Lu	0.25	0.37			0.38	0.36	0.22	0.27	0.29	0.38	0.32

		HKCA basalts				CA basalts	KAB			
M86/2 Madjarovo	M88-251 Madjarovo	2014 Zvezdel	Zd01-12 Zvezdel	Zd95-4 Zvezdel	2050 Zvezdel	25-G Zvezdel	IEG-01	IIEG-01	GJ-17	STR-10
51.80	52.17	51.35	52.83	53.35	53.35	53.84	46.44	46.74	46.75	48.03
0.62	1.07	1.11	0.82	0.97	1.25	1.01	2.01	2.02	2.23	2.20
18.50	18.40	17.90	17.21	16.90	17.82	17.00	16.48	16.68	16.86	18.04
4.95	3.01		8.47^{T}		3.87					
3.00	4.80	8.77^{T}		8.15^{T}	4.35	8.07^{T}	9.24^{T}	8.72^{T}	8.75^{T}	8.42^{T}
0.13	0.14	0.16	0.14	0.16	0.15	0.16	0.15	0.16	0.18	0.20
3.80	4.19	5.23	4.88	4.28	4.50	4.64	7.61	7.38	5.96	4.51
7.80	8.94	9.91	9.22	9.08	8.37	8.25	10.10	9.35	8.49	8.39
2.98	2.64	2.66	2.52	2.61	2.67	3.31	2.11	3.75	3.91	3.61
3.84	2.44	1.74	1.95	2.12	2.09	1.31	2.30	2.34	2.94	3.24
0.40	0.37	0.34	0.33	0.29	0.28	0.19	0.57	0.69	0.75	0.88
1.25					0.19					
0.76	1.06	0.84	1.36	2.10	0.91	2.23	3.13	2.08	2.91	2.54
99.83	99.23	100.00	99.74	91.85	99.80	91.94	90.89	91.19	90.98	91.63
8		24.2	22.0	30.0	22.3	30	20.9	18.6	15.6	8.1
168		256	245	229	207	228	194	174	158	133
15	45	22	62	58	8	23	159	166	83	32
19.7			28		22	20	34	33	30	24
12	26.7	8	22	12	3	3	104	102	78	42
57	81.4	78	73	65	76	80	71	72	77	76
42	25.1	30	70	18	22	15	53	43	34	25
24	49.4	12	23	22	16	10	3.2	4.3	4.2	3.9
138	155.7	126	114	150	156	149	187	224	262	307
4.2		2.78	4.0	3.88	3.7	3.44	3.3	4.1	5.0	5.3
18.3	14.2	7.4	8	7.73		8.27	74	83	94	104
1.16		0.5		0.6	0.4	0.54	3.7	4.3	5.6	6.1
7.49		2.42	6.00	3.31	1.80	1.35	2.2	2.1	3.0	2.4
24.3	23.5	23.9	23.0	24.2		27.27	25	26	25	25
14.9		7.0	10.0	9.0	9.8	6.02	8.3	10.2	9.3	9.6
214	85.7	94	81	99	100	57	59	75	66	72
12.11		3.90		3.99	2.80	2.65				
484	714.4	582	547	528	571	493	637	801	839	934
1198	1104.6	722	966	836	962	574	699	695	744	802
23.45		21.36	26	25.20	26.00	23.07	42.5	47.7	48.7	53.3
46.27		42.17	38	48.15	53.00	44.90	73	76	80	88
5.65		5.01		5.56		5.28				
23.53		21.36	22	22.51		21.40	27	29	31	29
5.59		5.24		5.27	4.40	5.09	6.4	7.4	7.0	7.5
1.49		1.42		1.40	1.50	1.38	2.0	2.1	2.1	2.2
4.78		4.64		5.01		4.70				
0.77		0.75		0.79		0.84	0.9	0.8	0.7	0.8
4.48		4.51		4.54		4.99				
0.85		0.91		0.92		1.00				
2.29		2.51		2.47		2.90				
0.33		0.32		0.35		0.41				
2.14		2.03		2.11	2.40	2.59	2.0	1.9	1.7	2.0
0.32		0.31		0.34	0.24	0.40	0.33	0.31	0.37	0.28

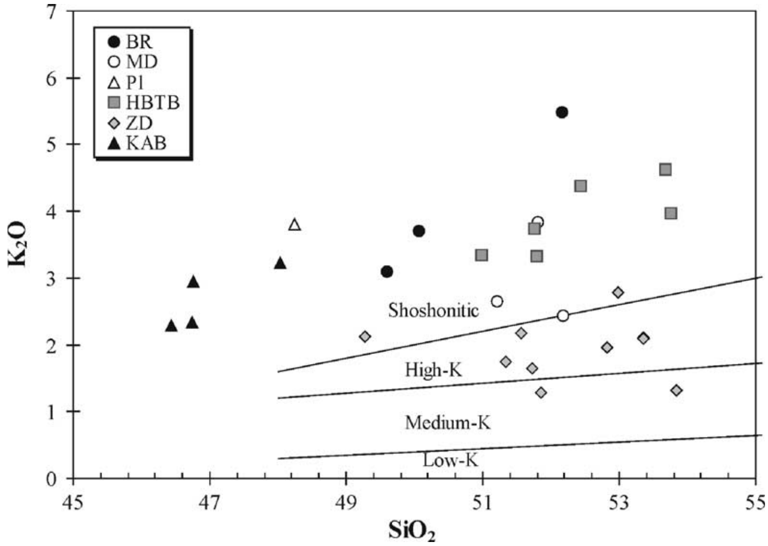

Fig. 3. K_2O–SiO_2 classification diagram after Peccerillo and Taylor (1976). Additional analyses for the Zvezdel volcano are included from Nedialkov and Pe-Piper (1998) and Yanev et al. (1989).

Comparison between the two ABS samples (B84/1 and Bz18) shows that despite their similar major element chemistry, Sr, Ba, and Th contents, the Borovitsa ABS has distinctly higher $^{87}Sr/^{86}Sr$ (0.70825) relative to the absarokite from Planinets (0.70647). The isotopic compositions of the ABS are similar to the isotopic compositions of the more evolved volcanic rocks from the two areas. In summary, the orogenic basalts have rather variable $^{87}Sr/^{86}Sr$ ratios (0.70688–0.70825), but relatively restricted $^{143}Nd/^{144}Nd$ isotopes (0.51252–0.51243).

Sr and Nd isotopic compositions of the KAB include very low $^{87}Sr/^{86}Sr$ ratios (0.70323–0.70338) and high $^{143}Nd/^{144}Nd$ ratios (0.51290–0.51289). An amphibole megacryst has Sr and Nd isotope ratios indistinguishable from its host basalt, supporting a cognate relationship, and arguing against significant crustal contamination after amphibole crystallization.

6.4. Pb isotopic composition

The HKCA and SHO lavas display a relatively restricted range of $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ ratios (18.90–18.84, 15.69–15.67, and 38.94–38.80, respectively; Table 2). Compared to them, HBTB have lower $^{206}Pb/^{204}Pb$ (18.73–18.72) but much more radiogenic $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (15.76–15.74 and 39.14–39.07, respectively). The ABS from Planinets is slightly less radiogenic than the HBTB ($^{206}Pb/^{204}Pb$=18.78; $^{207}Pb/^{204}Pb$=15.72; $^{208}Pb/^{204}Pb$=38.96). KAB show higher $^{206}Pb/^{204}Pb$ (19.02–18.91) at lower $^{207}Pb/^{204}Pb$ (15.64–15.52) and $^{208}Pb/^{204}Pb$ (38.87–38.59) ratios. On Pb isotope diagrams (Fig. 8), the data form a nearly vertical array from KAB, lying on the Northern Hemisphere Reference Line (NHRL; Hart, 1984) to the HBTB at the other extreme.

7. Petrogenesis of the Late Eocene–Oligocene Eastern Rhodope magmatic rocks

7.1. Krumovgrad alkaline basalts

The presence of spinel lherzolite xenoliths in some of the KAB (Marchev et al., 1997; 1998b) confirms the upper mantle origin of these alkaline rocks. However, based on their low Mg#, Ni, and Cr contents, KAB are not primary magmas and have probably undergone of olivine and clinopyroxene fractionation. From their high LILE abundances and high MREE/HREE ratios, Marchev et al. (1997, 1998b) suggested that the KAB were generated in a garnet-bearing LREE-enriched peridotite mantle source containing metasomatic biotite. KAB bulk rocks and their amphibole megacrysts have Sr–Nd

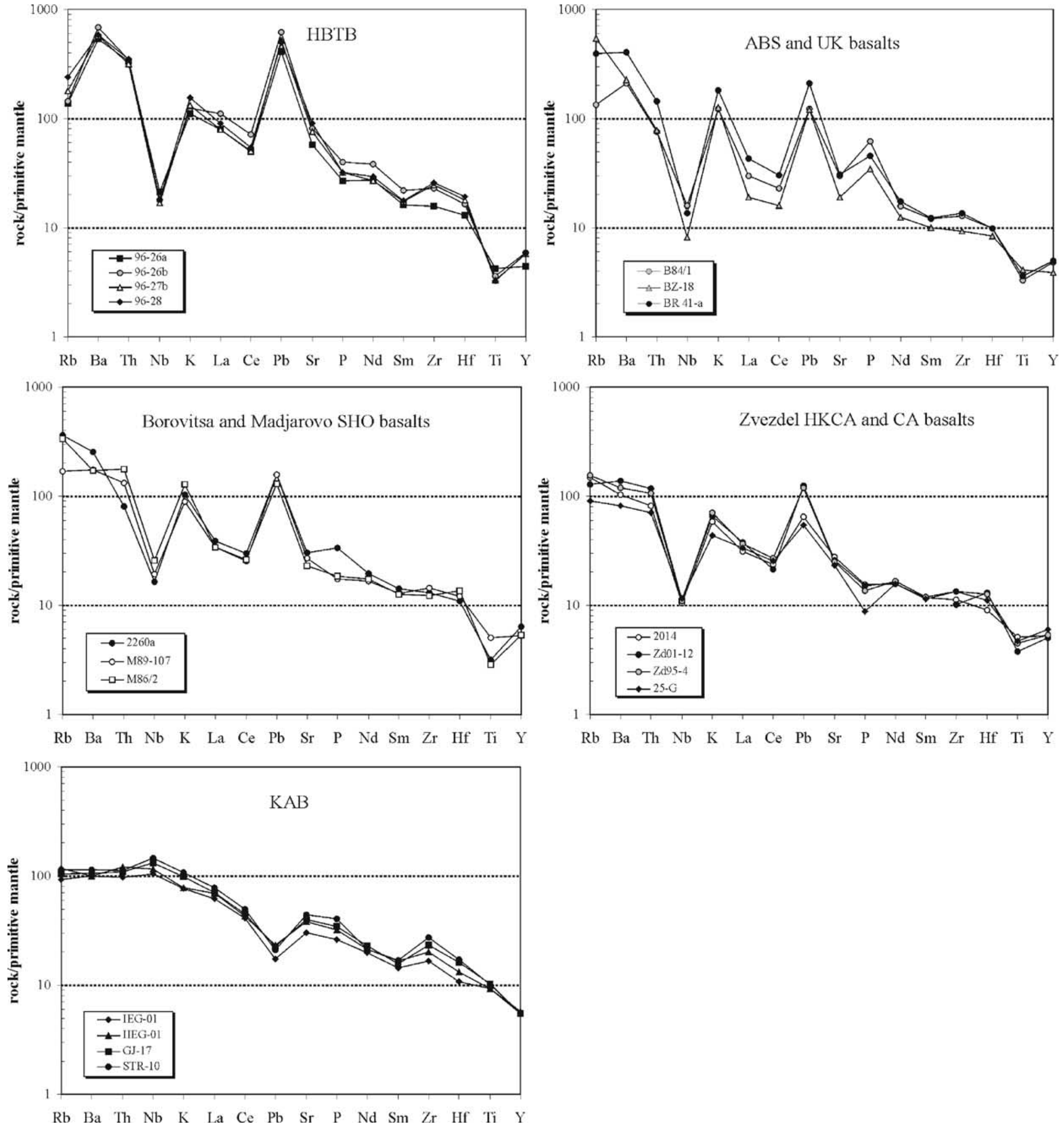

Fig. 4. Primitive mantle-normalised incompatible trace element diagrams for the different mafic rocks in the Eastern Rhodopes. Normalisation values from Sun and McDonough (1989).

isotope compositions that overlap the European Asthenospheric Reservoir (EAR) inferred to be present in the shallow asthenospheric mantle throughout central and western Europe (Granet et al., 1995; Cebria and Wilson, 1995). Compared to the EAR values, KAB have lower $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ at similar $^{207}Pb/^{204}Pb$ (Fig. 8). Marchev et al. (1998b) showed that lower $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ isotope ratios are typical features in Eastern European alkali basalts reflecting the presence of a depleted

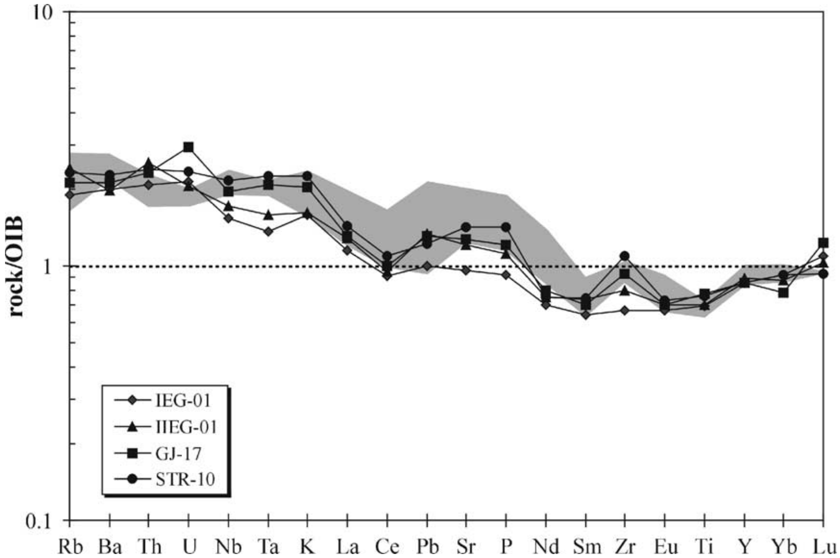

Fig. 5. OIB-normalised incompatible trace element patterns for representative samples from the KAB and the Kula lavas (Alici et al., 2002). Normalisation values from Sun and McDonough (1989).

mantle lithospheric component rather than the effects of crustal contamination. Most of the European mantle has a composition similar to that of the most depleted spinel lherzolite xenoliths from the Pannonian–Carpathian region and the Srednogorie clinopyroxene megacrysts. Sr and Nd isotopic compositions of this lithosphere (Fig. 7) are similar to EAR. Hence, these isotopic systems are insensitive to lithospheric contamination. However, the depleted lithosphere is characterized by far lower $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios than the EAR (Fig. 8), and Pb isotopes can therefore be very sensitive indicators of contamination of asthenosphere-derived magma. The compositions of the cumulates, in addition to the Fe–Na-rich (jadeitic) clinopyroxene xenocrysts, provide evidence for prolonged high-pressure fractionation of originally more Mg-rich parental magmas. Interaction of primitive within-plate basalts with this depleted mantle might result in the observed shift of $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios of the KAB.

7.2. The origin of the orogenic basalts

The origin of the Eastern Rhodope orogenic magmas is contentious. In an attempt to explain the high $^{207}Pb/^{204}Pb$ ratios of the Oligocene–Miocene igneous rocks from the Greek part of the Rhodopes, Pe-Piper et al. (1998) suggested that they were derived by melting of metasomatised mantle, enriched in LILE and HFSE during ancient subduction. Similarly, following the model proposed by Foley et al. (1987), Francalanci et al. (1990), Nedialkov and Pe-Piper (1998), and Yanev et al. (1998a) suggested that parental magmas were generated from mantle sources heterogeneously enriched by fluids and melts derived from previous subduction. These components formed hydrous veins in the mantle. Different degrees of partial melting of the veined mantle (Foley, 1988) can produce magmas variously enriched in K, incompatible elements, and isotopic composition. An old enrichment of the source for these basalts is also supported by the high $^{87}Sr/^{86}Sr$ (0.70688–0.70825), low $^{143}Nd/^{144}Nd$ (0.51252–0.51243), and high $^{207}Pb/^{204}Pb$ (15.76–15.66) ratios of the Eastern Rhodope basalts, which require a mantle source with a time-integrated history of enrichment in Rb, LREE, and U.

Plots of Nb / La versus Ba / Rb (Fig. 9) are useful as a tool for exploring the influence of the subduction-related component (e.g. Wang et al., 1999). Fluid and sediments released from the subducting slab cause decreases in Ba / Rb ratios due to preferential partitioning of Rb over Ba in this fluid (Tatsumi et al., 1986; Peccerillo, 1999). Despite large variations in

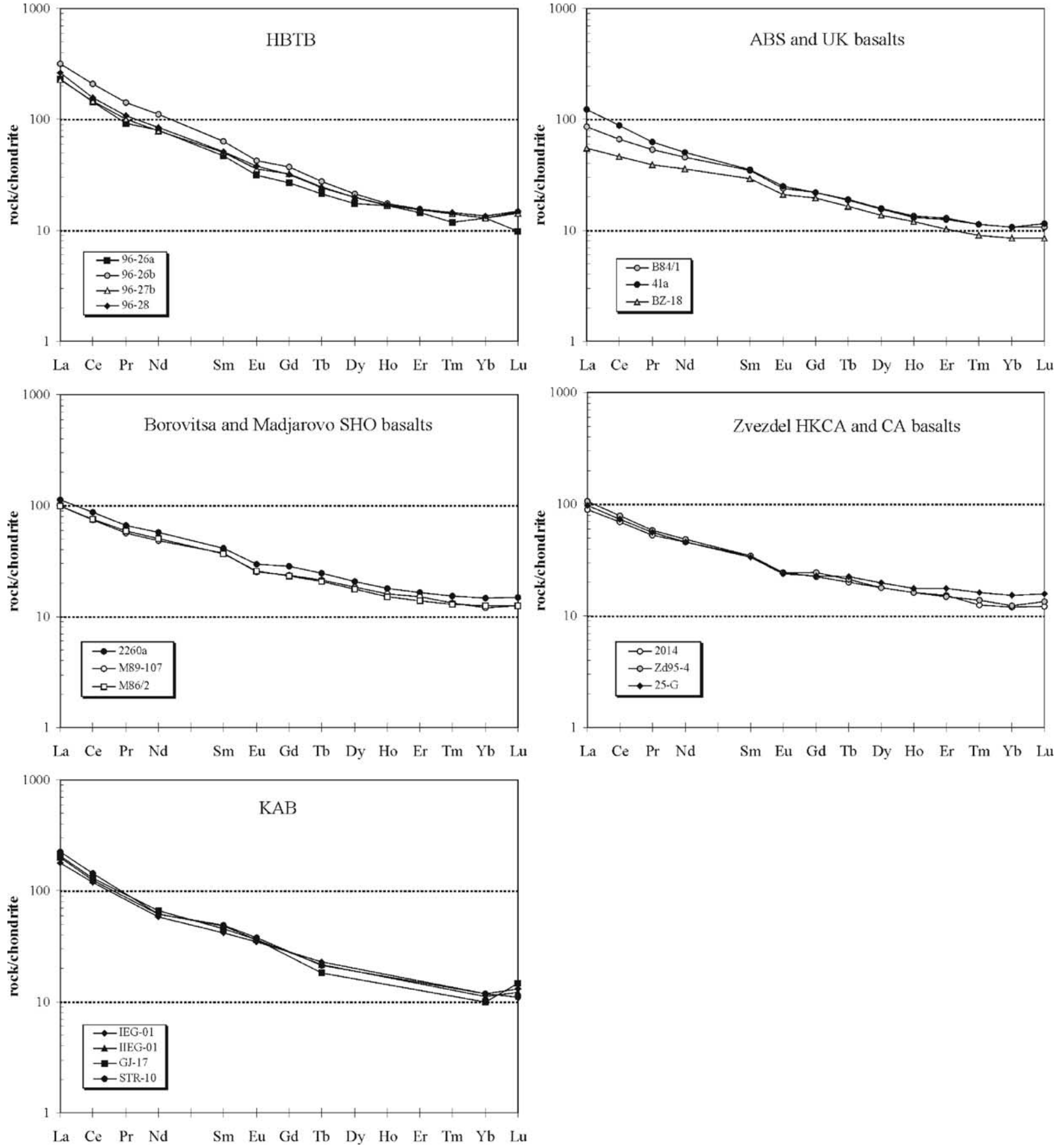

Fig. 6. Chondrite-normalised REE diagrams (Boynton, 1984). The overall similarity of the REE patterns in all studied rocks and the development of a negative Eu anomaly in the orogenic rocks should be noted.

Nb/La, the Ba/Rb ratios of the CA, SHO basalts (5.6–17.4), and KAB (7.7–11.8) are close to those of non-orogenic lavas (~11; Sun and McDonough, 1989). The vertical continental crust-N-MORB-OIB mantle trend (Fig. 8) suggests mixing of a within-plate magma with a crustal component rather than the influence of a subduction-related fluid. The HBTB show a shift towards much higher Ba/Rb (26.9–52.4)

Table 2
Sr, Nd, and Pb isotope compositions of representative mafic rocks from the Eastern Rhodopes

	HBTB				ABS		SHO basalts				HKCA basalts		KAB			
	96-26a	96-26b	96-28	96-27b	Bz-18 Planinets	B84/1 Borovitsa	2260a Borovitsa	M89-107 Madjarovo	M86/2 Madjarovo	M88-251 Madjarovo	2014 Zvezdel	Zd95-4 Zvezdel	IEG-01	IIEG-01	GJ-17	STR-10
$(^{87}Sr/^{86}Sr)_I$	0.70688	0.70756	0.70736	0.70727	0.70647	0.70825	0.70823	0.70798	0.70789	0.70792	0.70713	0.70722	0.70335	0.70338	0.70323	0.70324
$(^{143}Nd/^{144}Nd)_i$	0.512523	0.512430	0.512439	0.512455	0.512514		0.512436	0.512450	0.512461		0.512441	0.512434	0.512894	0.512902	0.512896	0.512903
$^{206}Pb/^{204}Pb$	18.728	18.724			18.775	18.730	18.905	18.862	18.838		18.857	18.839	18.905	18.921	19.020	18.910
$^{207}Pb/^{204}Pb$	15.738	15.760			15.720	15.665	15.685	15.694	15.668		15.672	15.691	15.637	15.585	15.624	15.518
$^{208}Pb/^{204}Pb$	39.067	39.142			38.961	38.804	38.865	38.939	38.823		38.879	38.950	38.852	38.730	38.872	38.592

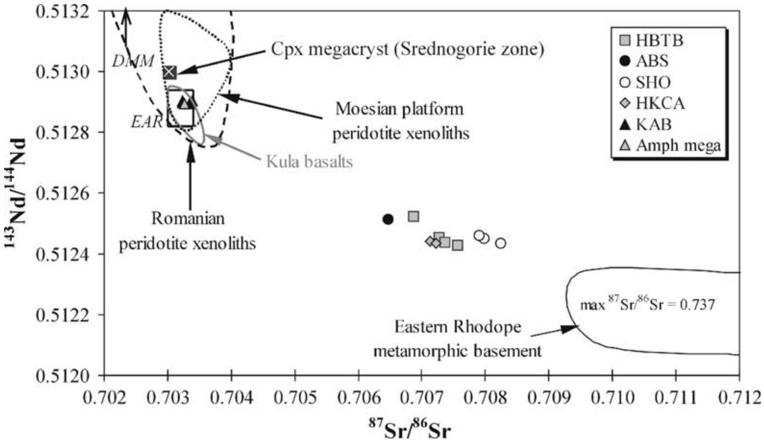

Fig. 7. $^{143}Nd/^{144}Nd$ versus $^{87}Sr/^{86}Sr$ isotope variation diagram of the basaltic rocks from ERMZ. Data for the Kula basalts after Alici et al. (2002). The field for the metamorphic basement rocks from the Eastern Rhodopes is constructed using unpublished data of von Quadt and I. Peytcheva and P. Marchev. The fields for the peridotite xenoliths from the Moesian Platform and Romania are after Vaselli et al. (1995, 1997). EAR after Granet et al. (1995) and Cebria and Wilson (1995).

and lower Nb/La (0.17–0.28), which is completely opposite of what would be expected if subduction processes had operated. Thus, we conclude that the enrichment process in the Eastern Rhodopes basalts is unrelated to slab-released hydrous fluids or contamination by sediment subduction. Given the high-Ba regional character of asthenospheric-derived magmas in the Eastern Rhodopes (see also next section), an alternative explanation invokes a decrease in the amount of partial melting of a Ba-enriched source.

The most convincing evidence that orogenic magmas in the Rhodopes were not derived from enriched subcontinental lithospheric mantle is the lack of enriched or veined mantle xenoliths. Spinel lherzolite xenoliths in the KAB and the alkali basalts in the neighbouring Srednogorie zone and Moesian Platform have a predominantly depleted composition (Fig. 7). Xenoliths from the Moesian Platform have $^{87}Sr/^{86}Sr$ values of 0.70297–0.70387 and $^{143}Nd/^{144}Nd$ of 0.51350–0.512827 (Vaselli et al., 1997). These values are within the range of the Sr–Nd isotopic composition of xenoliths from Romania ($^{87}Sr/^{86}Sr$=0.7018–0.7044 and $^{143}Nd/^{144}Nd$=0.51355–0.51275) (Vaselli et al., 1995) and ultramafic xenoliths from the European subcontinental lithospheric mantle (Downes, 2001). For the Moesian Platform xenoliths, Marchev et al. (2001) demonstrated that enriched samples show evidence for considerable interaction with their host alkaline basalts. Depleted spinel lherzolites from the Pannonian Basin and Romania have very low $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (Rosenbaum et al., 1997) (Fig. 8). The suggestion of a depleted lithosphere beneath the Rhodopes is substantiated by the identical isotope compositions of a clinopyroxene megacryst, derived from disaggregated spinel lherzolites from the Srednogorie zone, 50 km north of Borovitsa area.

Given their high SiO_2 and low Mg, Ni, and Cr contents, the ERMZ orogenic basalts, except probably the absarokites, do not represent primary mantle melts and probably suffered olivine and clinopyroxene fractionation (Marchev et al., 1998a). The observed NW–SE change in isotopic composition in Central and Eastern Rhodopes orogenic magmas (Marchev et al., 1989, 1994) correlates clearly with crustal thickness, consistent with the possibility of variable crustal contamination. However, Marchev et al. (1998a) argued that fractionation of olivine and clinopyroxene and even contamination with Rhodopian upper crustal rocks would not significantly affect the incompatible element patterns of the basaltic rocks. These authors used the rare earth elements (e.g. La/Yb) and HFSE (Nb/Zr) ratios as monitors of the degree of partial melting (e.g. Thirlwall et al., 1994). HBTB broadly straddle the previously recognised partial melting vector (in Fig. 10), suggesting that they are genetically indistinguishable from the other orogenic rocks and their enrichment is probably due to the very small

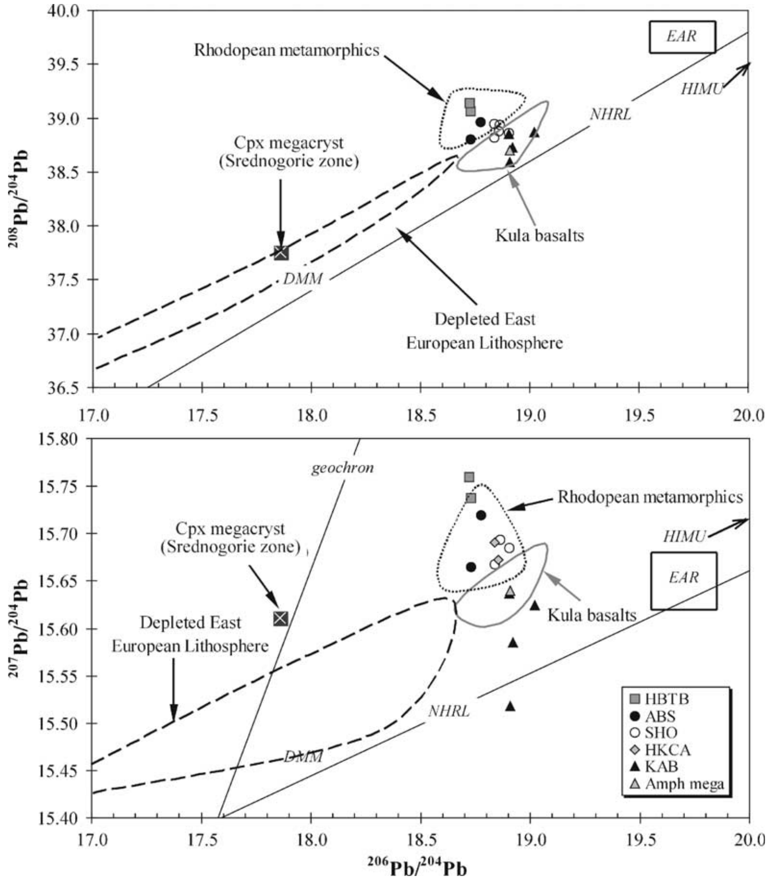

Fig. 8. ^{206}Pb/^{204}Pb versus ^{208}Pb/^{204}Pb (upper panel) and ^{207}Pb/^{204}Pb (lower panel) for the ERMZ basaltic rocks. Kula basalts after Alici et al. (2002). The field for depleted SE Europe peridotite xenoliths after Rosenbaum et al. (1997). Data for the metamorphic basement from the Chalkidiki part of the Serbomacedonian Massif, which is analogous to the Variegated complex of the Eastern Rhodopes, are from Frei (1995).

degree of partial melting. As a whole, the geochemical variations exhibited by the Eastern Rhodope mafic magmas indicate that they are related to a common mantle source, which is characterized by a relative enrichment of K, Ba, and water contents. Similarities of the MREE/HREE ratios in the orogenic basalts (Tb/Lu=2.32–3.20) to those of the KAB (2.61–2.86) suggest that both magma types tap garnet-bearing source regions.

Crustal contamination can explain also the elevated Sr and lower Nd isotopes and particularly the high ^{207}Pb/^{204}Pb and ^{208}Pb/^{204}Pb ratios. In Fig. 8, the Pb isotope values of the HBTB overlap those of the metamorphic rocks of the western part of the Rhodope Massif. This is not surprising since Pb is commonly cited as an element typically introduced in magma by selective contamination (Taylor et al., 1980; Dickin, 1981).

An increase of ^{87}Sr/^{86}Sr combined with decreasing Eu/Eu* ratios (Fig. 11) provides good evidence for the increasing role of crustal contamination from the HKCA rocks in the Greek part of the Rhodopes through Borovitsa and Madjarovo shoshonites to HBTB. Negative Eu anomalies can be produced by plagioclase fractionation, residual plagioclase in the source, or source contamination by slab-derived fluids and sediments (Ellam and Hawkesworth, 1988). Plagioclase is a liquidus phase in the high-Al basalts, but their high-Al contents mean that they have not undergone significant plagioclase fractionation. Moreover, there is no difference in the Eu anomalies between plagioclase-bearing basalts and plagioclase-

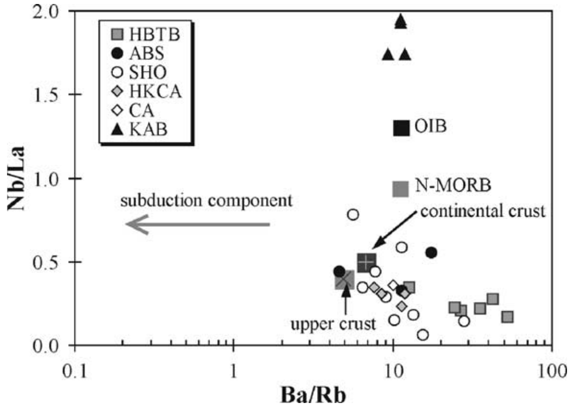

Fig. 9. Nb / La versus Ba / Rb diagram for basaltic rocks from the ERMZ. Data for OIB and MORB are from Sun and McDonough (1989). The continental crust values of (Ba / Rb=6.76; Nb / La=0.5; upper crust Ba / Rb=4.9; Nb / La=0.4) are from McLennan (2001).

free ABS. Thus, contamination by the Rhodope metamorphic lithologies, which have large Eu anomalies (Cherneva and Daieva, 1986), is the most reasonable explanation for the Eu anomaly of the

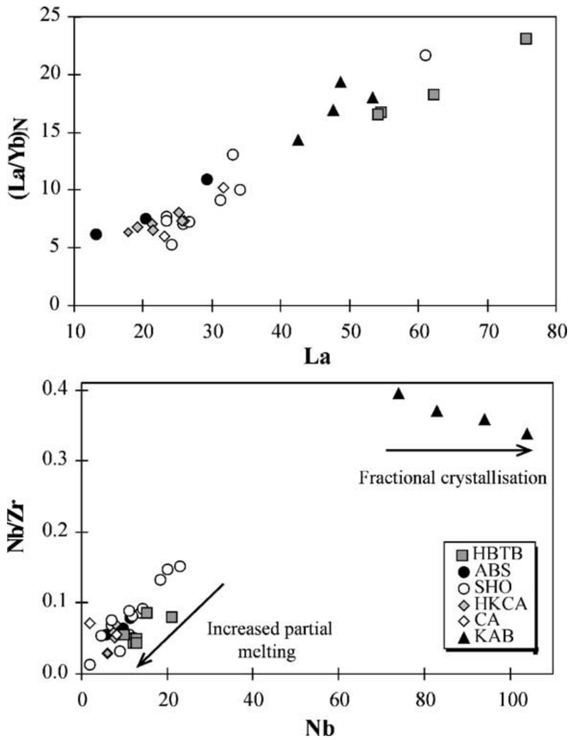

Fig. 10. $(La/Yb)_N$ versus La (ppm) and Nb / Zr versus Nb (ppm), illustrating effects of partial melting and fractionation.

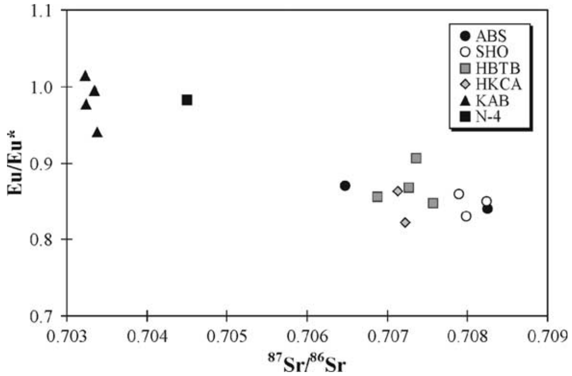

Fig. 11. Eu / Eu* versus $^{87}Sr/^{86}Sr$ ratios showing the consistent correlation between the two. Sample N-4 is from unpublished data of G. Christofides.

basalts. Finally, we also consider that the variable depletion in Nb in the Eastern Rhodope basalts is consistent with the combined effects of different degree of partial melting and crustal contamination. The latter process was suggested by Turner et al. (1999) to explain Ta–Nb anomaly in the Miocene tholeiitic and calc-alkaline volcanic rocks of SE Spain.

8. Discussion

8.1. Relationship between tectonics, magmatism, and metamorphism

After a period of crustal thickening (160–115 Ma?), an extensional regime was established in the Eastern Rhodopes. The process started with exhumation of the Variegated complex of Biala Reka and Kesebir metamorphic core complexes and deposition of Maastrichtian–Paleocene subaerial to submarine sediments in basins north of these domes (Goranov and Atanasov, 1992; Boyanov and Goranov, 2001). The onset of extension coincided with the intrusion of granitoid plutons within the Variegated complex, which at that time was probably at middle to upper crustal depths. The plutonic activity ranged from 70 Ma; cf. Quadt, personal communication, 2003) to 42 Ma (Peytcheva et al., 1998a, Ovcharova, personal communication, 2003), coinciding with metamorphism (73.5±3.4 Ma) of garnet-rich rocks in the Greek part of the complex (Liati et al., 2002), thereby defining a significant Late Cretaceous magmatic-metamorphic event.

In Late Eocene time, another major episode of extension started with block faulting and formation of E–W to NNW–SSE sedimentary basins. The oldest sediments are nonvolcanic (Harkovska et al., 1989; Boyanov and Goranov, 2001) and give way upwards to voluminous dyke-fed intermediate to basaltic volcanic rocks (Harkovska et al., 1989; Marchev, 1985; this study). This activity is dated between 39 and 30 Ma, peaking at 33–31 Ma (Yanev et al., 1998b; Marchev et al., 1998a; Lilov et al., 1987), with younger activity reported in Greece (Christofides et al., 2001). Tectonic extension in this region partly coincides with a 42–35 Ma thermal event and exhumation of the lower Gneiss–migmatite complex. The emplacement of OIB-like KAB dykes at 28–26 Ma appears to have succeeded exhumation of the core complexes and magmatism in the Bulgarian Eastern Rhodopes.

8.1.1. Subduction zone model

A subduction mechanism has been proposed to explain the Late Cretaceous magmatism in the Srednogorie zone (e.g. Dabovski, 1991). In this model, the Rhodopes are considered to be a frontal part of the Srednogorie arc (Fig. 1; Kamenov et al., 2000). High-precision U–Pb zircon and rutile age dating in the Central Srednogorie (Peytcheva et al., 2002) document a southward shift of this magmatism from 92 to 78 Ma. Therefore, the 70–42 Ma Rhodope granitoid intrusions may be a continuation of the Srednogorie magmatism into the Rhodope region. The available age and geochemical data for Late Cretaceous–Early Eocene granitoid magmatism are still insufficient to clarify this period of the evolution of the Rhodopes. However, it is clear that an isolated subduction event can hardly explain 45 million years' continuous magmatism in the Rhodopes unless the subduction process persisted. Indeed, progressive southward migration of magmatic activity in the Aegean region (Fyticas et al., 1984), commenced in the Rhodopes in the Late Eocene (Yanev et al., 1998b), has been confirmed by seismic tomography (Spakman et al., 1988), implying that present-day north-vergent subduction in the Aegean region had begun by at least 40 Ma.

8.1.2. Rollback and slab detachment (slab break off)

Rollback is considered to play a primary role in the modern tectonic evolution of the Aegean region (Wortel and Spakman, 2000 and references therein). It leads to extension in the cold back-arc lithosphere and its replacement by hot asthenosphere to shallow mantle levels and explains the southward trench and magmatic migration in the region.

Slab break off was suggested as a mechanism to explain the magmatism in the Alps and Aegean region (Davies and von Blanckenburg, 1995; De Boorder et al., 1998). It leads to heating of the overriding lithospheric mantle by upwelling asthenosphere and melting of its enriched layers. Different degrees of partial melting of the enriched lithosphere produce magmas ranging from alkaline to ultrapotassic, whereas slightly higher degrees of melting of more fertile peridotite layers produce calc-alkaline melts. Decompression melting of dry asthenosphere is possible only in case of detachment at depths of <50 km (Davies and von Blanckenburg, 1995). Slab break off can explain the variations of the Rhodope Eo-Oligocene magmatism from more alkaline to less alkaline varieties, but the duration and distribution of this magmatism are larger than those predicted by this model. In the case of the Eo-Oligocene basalts, we do not see isotopic variations that would confirm that melting of differently enriched lithosphere has occurred since the isotopic compositions display a comparatively narrow range and variations show correlation with crustal thickness (Marchev et al., 1989, 1994) rather than with degree of enrichment. In addition, melting of the asthenosphere in the region implies break off at <50 km and removal of the entire lithosphere and may be part of the crust (see Fig. 2 of De Boorder et al., 1998). This should promote massive crustal melting and large volumes of felsic magma, but the similarity in isotopic composition of the felsic and mafic rocks suggests that they were derived by fractionation of a common melt rather than melting of different sources. Finally, it is obvious that at least some of the lithosphere mantle had remained to provide the source for the spinel lherzolite xenoliths in the KAB.

8.1.3. Delamination

Delamination of thickened crust after the Europe–Apulia continental collision has been proposed by

Yanev et al. (1998a) to explain Priabonian–Oligocene magmatism in the Eastern Rhodopes. In many respects, this model is indistinguishable from slab break off. It predicts a rapid intrusion of the asthenosphere in the place of delaminated mantle and eclogitized lower crust resulting in uplift and extensional collapse. Rapid asthenospheric uplift would produce a thermal anomaly leading to partial melting of the asthenosphere and lower lithosphere and underplating of these magmas beneath the lower crust. This would cause widespread lower crustal melting and hybridization as indicated by the bimodal volcanic compositions. The southward migration of the activity could reflect a delamination process which would have started under the Eastern Rhodopes and then migrated southward under the North Aegean region. This model has the same shortcomings as the previous one.

8.1.4. Convective removal

Convective thinning of the unstable thickened lithospheric mantle (Houseman et al., 1981; Turner et al., 1992; Platt and England, 1994; Houseman and Molnar, 1997) may result in removal of the whole mantle lithosphere and its replacement by hot asthenosphere. Rapid replacement of the lithosphere by hot asthenosphere causes fast uplift of the overlying crust. According to Turner et al. (1999), if thermal thinning raises the asthenosphere to within 50 km of the surface, then decompression melting of the asthenosphere could occur, with increasing amounts of crustal contamination of the asthenospheric melts. Distinguishing between convective removal and other processes leading to removal of the lithosphere (e.g. delamination) is difficult since they will have similar effects on extension and magmatism. According to Houseman and Molnar (1997), convective thinning differs from delamination in: (1) symmetry of the convective thinning with no migrating delamination front; and (2) removal of the lower part of the mantle only, which excludes direct contact between the asthenosphere and the crust. While Eo-Oligocene magmatism in the Rhodopes and associated geochemical features of Eastern Rhodope basalts could be linked to convective thinning, the observation that some Paleocene sediments in the Eastern Rhodopes are of submarine provenance suggests the activity was not accompanied by rapid plateau uplift. Moreover, the model does not account for the general southward migration of the magmatism.

8.2. Comparison with the North American Cordillera and Menderes Massif

With its prolonged history of extension and magmatism, the Rhodope region shows many similarities with the Tertiary evolution of the North American Cordillera. Initiation of extension in the North American Cordillera correlates with intrusion of abundant plutons and formation of core complexes (Armstrong, 1982; Wernicke et al., 1987). This stage of extension was followed by eruptions of widespread intermediate-silicic calc-alkaline magmatism, accompanied by limited extension (Best and Christiansen, 1991). The most important extensional phase, which led to the formation of the Basin and Range province, started in middle Miocene times (ca. 17 Ma) with block faulting, and basin sedimentation, followed by bimodal basaltic–rhyolitic or basaltic volcanism (see Wernicke et al., 1987; Liu and Shen, 1998 and references therein).

Most workers recognise the lack of a single cause for extensional tectonics and magmatism in the North American Cordillera. Liu and Furlong (1993) and Ranalli et al. (1989) have observed that extension and plutonism that closely followed crustal shortening in the Canadian Cordillera required mantle thermal perturbations. In the case of Late Cretaceous–Paleocene Rhodopian granitoids, the geochemical and isotopic character reflects involvement of an unradiogenic component from an enriched mantle or lower crustal source (Peytcheva et al., 1998a; Christofides et al., 2001). Liu and Shen (1998) argued that thermal energy lost through conduction during Late Miocene mantle upwelling, lithosphere extension, and magmatism in the basin and range required some 20-m.y. replenishment. This process can also explain the change of basalt isotopic signatures from older enriched lithosphere-derived to younger asthenosphere-derived ones (Harry and Leeman, 1995; Hawkesworth et al., 1995). Other authors (e.g. Wernicke, 1992, Wenrich et al., 1995) argue that older basalts also formed from an asthenospheric mantle which underwent crustal contamination. In the Eastern Rhodopes, older orogenic basalts have REE patterns that suggest an origin in the garnet stability field, similar to asthenospheric-derived KAB.

In the Eastern Rhodopes, the KAB show close spatial relationships with the core complexes suggesting some role in the last stage of the core complex evolution. Such a relationship seems to be a rare but not a unique feature in the Eastern Mediterranean region. A most relevant modern analogue to the Eastern Rhodope core complexes is the Menderes Massif in SW Turkey. The latter is a large, NE–SW elongated metamorphic core complex (Bozkurt and Park, 1994; Bozkurt, 2001; Hetzel et al., 1995; Lips et al., 2001) which is crosscut by E–W-trending graben structures, subdividing the massif into three submassifs. In Eocene–Miocene time, the region was affected by collapse, which resulted in a series of exhumation events at 40–35, 20–28, and 7–6 Ma (Lips et al., 2001). Spatially related to the northern submassif are the Quaternary alkaline basalts from Kula, which erupted between 1.7 and 0.025 Ma (Richardson-Bunbury, 1996), preceded by Middle-Late Miocene basalts in the western edge of the massif similar to those in the Eastern Rhodopes (Robert et al., 1992). In pursuance of the idea of a common tectonic and magmatic evolution of the Eastern Rhodope and Menderes core complexes, we compare the isotopic compositions of the two OIB basalt localities. Figs. 8 and 9 clearly demonstrate that alkaline basalts from the Eastern Rhodopes and Kula are isotopically indistinguishable. They exhibit similar low $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios like the KAB and other Eastern European alkali basalts, extending this tendency to the SE of this region. The Kula basalts also show higher Rb, Ba, Sr, and Nb (Fig. 5) and generally lower Ni and Cr contents than typical OIB. In practice, the duration of the formation of the Menderes and both Eastern Rhodope core complexes is similar (ca. 35–40 Ma) and the evolution of the magmatism is identical. Most importantly, recent tomographic images of the lithosphere and mantle beneath the Aegean region (Spakman et al., 1993; Wortel and Spakman, 1992; De Boorder et al., 1998) show that Menderes Massif is underlain by a deep vertical low-velocity zone, which appears to provide heat and asthenospheric magmas for the OIB-like Kula basalts, probably playing an important role in the core complex formation.

Seismic studies in the Eastern Rhodopes (Babuška et al., 1987) indicate a regionally thinned lithosphere about 80 km thick. Although the mantle and crust structure in the geologic past is more difficult to constrain, we can argue that the lithosphere at Oligocene time was considerably thinner. Tomographic images for the southern part of the Rhodope Massif (Papazachos and Skordilis, 1998) show very strong variations of the crustal thickness. The most important result is the thinning of the crust down to 25 km under the Tassos, Pangaio, and Biala Reka dome structures as opposed to crustal thickening of 32–35 km beneath the upper unit.

Our results cannot help the selection of a single simple model that can explain the extension and magmatism in the Rhodope region. The main conclusion from the above discussion is that the ca. 45 Ma history of extension and magmatism requires a continuous supply of heat in the whole region. Isotopic signatures of basaltic rocks indicate an origin in the mantle asthenosphere, suggesting an anomalously hot mantle, which can be attributed to upwelling of the asthenosphere under the Rhodope region. This is confirmed by the deep low-velocity zone which has been imaged by seismic tomography under the modern Menderes Massif, suggesting the raising of the asthenosphere to very shallow depths at the last stage of evolution of these areas. In addition, the mantle uplifting under the core complexes caused thinning of the crust that was imaged by the seismic tomography in the southern part of the Rhodopes. Thus, some form of convective thinning of the lithosphere and of mantle diapirism could be appropriate processes explaining the magmatism and extension in the Rhodope region. Future structural and age studies of magmatism and metamorphism are necessary to clarify the details of these processes.

9. Conclusions

Late Eocene–Oligocene (34–26 Ma) mafic magmatism in the Eastern Rhodopes formed as part of the extensional geodynamic evolution of the whole Rhodope region beginning in Late Cretaceous–Paleogene time. Initiation of extension is constrained by the formation of metamorphic core complexes, low-angle detachment faults, and supradetachment sedimentary basins during Maastrichtian–Paleocene time, accompanied by ca. 70–38 Ma metamorphism and granitoid intrusions. Emplacement of HBTB followed by SHO, HKCA, and CA basalts and finally by purely asthenospheric-derived within-plate basalts, with progressively decreasing amount of crustal component, reflects upwelling asthenospheric mantle. The asthenospheric basalts originated from isotopically depleted mantle enriched in LILE by fluids stabilising phlogopite. Older orogenic lavas and dykes also originated from a similar asthenospheric source, but their trace element and isotopic signatures require different degrees of partial melting followed by different degrees of crustal contamination.

Most of the models proposed in the literature to explain the cause of extension and magma genesis in the Mediterranean region cannot be applied directly to the Rhodopes. Critical evaluation of these models suggests that some form of convective removal of the lithosphere and mantle diapirism provides the most satisfactory explanation for the Paleogene structural, metamorphic, and magmatic evolution of the Rhodopes.

The protracted extension and magmatism show surprising similarity with those in the Western U.S. Cordillera and particularly with the Menderes Massif, SW Turkey. Both Eastern Rhodopes and Menderes core complexes exhibit exhumation histories of surprisingly similar duration (about 35–45 Ma) and evolution of the mafic magmatism, and similar mechanisms likely explain these processes in the two regions.

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