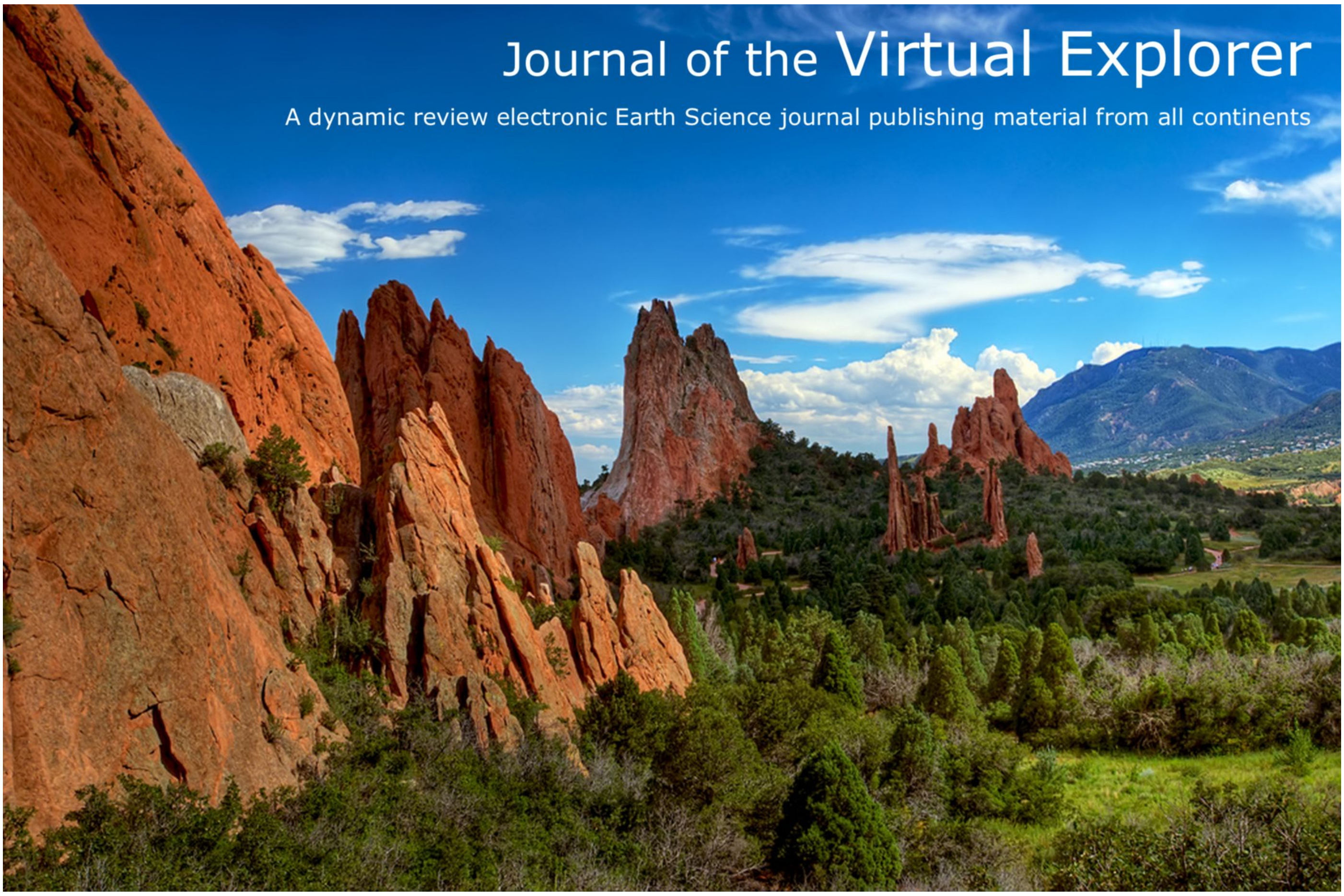

Rhodope: From Mesozoic convergence to Cenozoic extension. Review of petro-structural data in the geochronological frame

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Rhodope: From Mesozoic convergence to Cenozoic extension. Review of petro-structural data in the geochronological frame

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Abstract: Mylonitic gneisses of the Bulgarian and Greek Rhodope were deformed under amphibolite-facies conditions of medium pressure type metamorphism. The kinematic information contained on the strain regime and histories of these gneisses shows that ductile, shear-deformation occurred during development of a nappe complex. The nappe complex is characterised by south to southwestward (forelandward) piling-up and both coeval and subsequent extension. Different lithologies, deformation and metamorphic histories discriminate lower (footwall) and upper (hangingwall) continental terranes that define a crustal-scale duplex. Ultrahigh-Pressure metamorphic rocks, eclogites, ophiolitic and magmatic arc protoliths are found in various units of the crustal-scale duplex structure. These rocks delineate a suture zone between the hanging wall and footwall continental units. Synmetamorphic suturing and thrusting imply crustal thickening during the Cretaceous, which implies that the Rhodope massif is a complex of synmetamorphic nappes stacked in a Tethyan active margin environment. The two blocks involved in the collision are the Moesian part of the European continent to the north, and the Lower-Rhodope Terrane to the south, which was a migrating block detached from Pangea during breakup times of this supercontinent.

Regional inversions of synmetamorphic sense-of-shear indicate that exhumation tectonics began in Cretaceous times, possibly linked to upward-forward expulsion of low density arc and continental rocks. A Late Eocene marine transgression separates the early, late-orogenic extension/exhumation phase from another extension event accompanied by a major thermal and magmatic event and followed by the Miocene Aegean extension responsible for late grabens over the Rhodope Massif.

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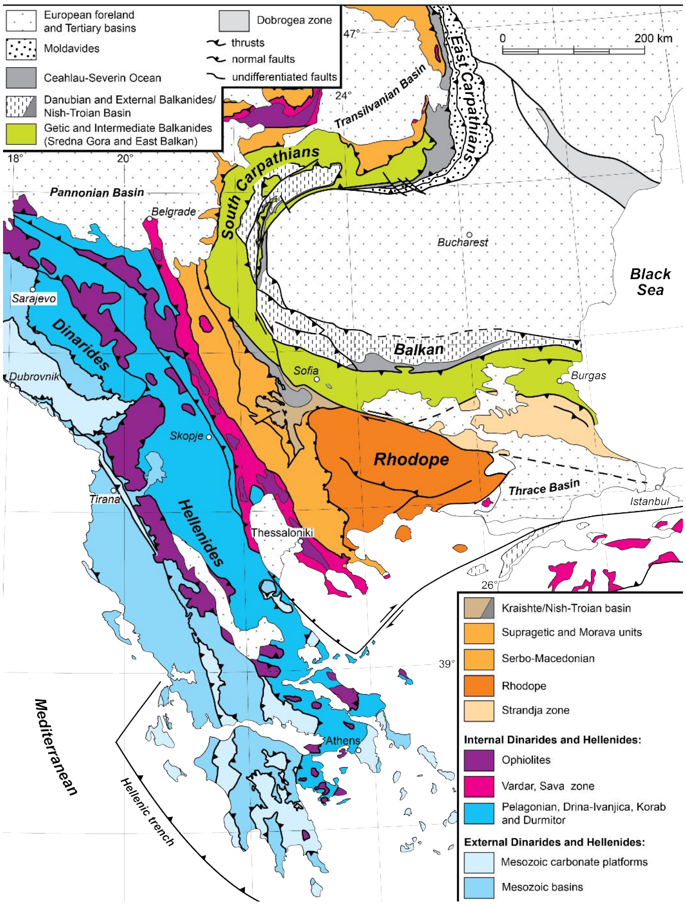
Introduction

This review paper deals with the mylonitic gneisses of the Rhodope Massif. They were deformed under medium-pressure amphibolite-facies conditions (biotite-garnet-staurolite parageneses are common in metapelites) and are examined in the light of results and concepts developed since the 1980. Although this special volume is dedicated to Greece, it is impossible to ignore the work carried out in the Bulgarian Rhodope. It is also impossible to cite all references -many are considered to be obsolete. Consequently, this review unavoidably reflects the author's opinion. For instance, we follow Ricou *et al.* [1998] to integrate the Greek and much of the Bulgarian parts of the so-called Serbo-Macedonian Massif [Kockel and Walther, 1965] to the Rhodope (Fig. 1). This attribution is a return to initial definitions of units, which bordered the Rhodope Massif along its southwestern boundary with the Vardar Zone [Kossmat, 1924].

We first summarize the evolution of concepts and discussions on the Rhodope since the earliest geological explorations in the area. We then place emphasis on recent discoveries and progress, in particular those related to geochronology, to definitively assert that the Rhodope is a deformed segment of the Alpine-Himalayan suture and collision system. Reviewing these data yields timing of deformational and thermal events that controlled the geodynamic evolution of the area. A synmetamorphic duplex system includes ultrabasic and basic rocks traditionally lumped under the term "ophiolites" and a Jurassic magmatic arc. These rocks were partly subducted to Ultra-High-Pressure (UHP) metamorphic conditions and retrogressed to amphibolite-facies during Cretaceous times. Subsequent crustal thickening led to syn-orogenic extension and extrusion/exhumation of the metamorphic rocks already brought back to the surface in Latest Cretaceous-Paleocene, at least in the northeastern Rhodope. The present-day structure results strongly from these combined large-scale thrusting and pervasive exhumation tectonics. A Late Eocene [Priabonian; Černjavska, 1977] marine transgression marks isostatic equilibrium of the belt in Early Tertiary, until a major extensional event reworked the Rhodope Metamorphic Complex during the Eocene and the Oligocene. This event, associated with voluminous magmatism, is responsible for synmetamorphic reworking of the older gneiss sytem in major, low-angle detachments. A second extensional event produced grabens related to later Aegean extension.

Figure 1. Location of the Rhodope in the Alpine Mediterranean chains.



The high-grade Rhodope metamorphic complex is limited upward by roof greenschists below the European, Carpatho-Balkanic units. It is limited at the Rhodope-Vardar boundary zone by the different, western greenschists. Adapted from Kounov *et al.* [2010].

The Rhodope Metamorphic Complex: A part of the Alpine orogenic system

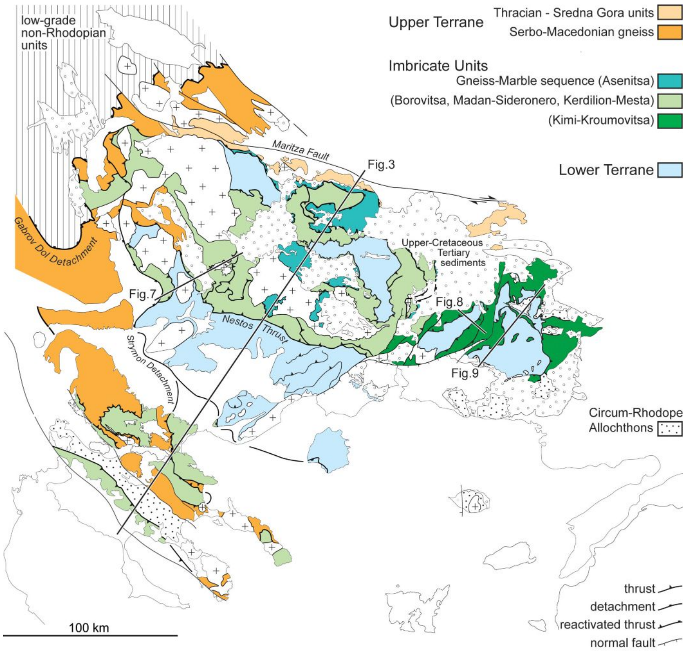
Historical perspective

A rosy-cheeked (etymology of Rhodope) Naiad nymph, queen of Thrace, and her husband, king Haimos, were transfigured into mountains after they dared compare themselves to, and so offend the celestial and almighty couple Hera and Zeus. Yet, Zeus showed some indulgence since he did not separate them: Haimos was metamorphosed into the neighbouring Balkan Mountains (Ovid, Metamorphoses, book 6. 87-89). The mythology narrates that Rhodope was the daughter of the river-god Hebros, hence the grand-daughter of Okeanos and his sister and wife, Tethys.

Much geological work since the late 1980’ has re-established the affiliation between Rhodope and Tethys, the long-recognized oceanic basin from which the Alpine-Himalayan orogenic system rose [Suess, 1888].

The oldest description of gneissic and plutonic rocks in the Rhodope is due to Boué [1836] and the name "Massif de Rhodope" to Viquesnel [1853]. The belief of the time was that such rocks should be old, such that crystalline rocks of the Balkan and Rhodope Mountains were considered as pre-Alpine basement

Figure 2. Sketch map of the Rhodope Metamorphic Complex.



Denudation of the northern part, along the Gabrov Dol Detachment, is older than the crosscutting, Turonian (ca 75 Ma) pluton (Table 7 and Fig. 10). The Strymon Detachment is the larger identified Cenozoic denudation fault. Several Rhodope thrust sheets have been further displaced during extensional tectonics and coeval formation of the Cenozoic basins.

[Cvijič, 1904]. Some authors compared these gneisses to those of the eastern Alps and suggested that they "received" their tectonic character from the "movements" that formed the whole Alpine fold belt [Kossmat, 1924]. This interpretation was possibly too daring for the time and Kober [1921; 1928; 1929] suggested that the Rhodope was a rigid "Zwischengebirge" between two branches of the Alpine chains: the Dinarides-Hellenides on the one side, and the Carpathians-Balkanides on the other. Kober considered the Vardar ophiolitic zone [his Narbe root zone; Kober, 1952]) as the major boundary between the two Alpine branches and attributed the Rhodope to the South-Balkan realm, extending from the Vardar Suture zone northeastwards. Since then, the Rhodope as a geological entity has been placed between the Vardar (Axios in Greece) valley to the west, the Aegean Sea to the south, and the Maritza Fault to the north (Figs. 1 and 2). The old-basement interpretation was reiterated for some years [Jaranoff, 1938; Vergilov *et al.*, 1963; Boncev, 1971; Foose and Manheim, 1975; Pal'shin *et al.*, 1975; Kozhoukharov *et al.*, 1988] and accepted for early plate tectonics descriptions [Dewey *et al.*, 1973; Hsü *et al.*, 1977; Burchfiel, 1980], despite clear statements on Alpine tectonics and metamorphism by few authors who followed Kossmatt's ideas [Petraschek, 1931; Janichevsky, 1937; Gâlâbov, 1938]. "Dinaride"-type orogen (i.e. southwestward thrusting of Alpine orogeny) was even emphasised for the eastern Rhodope [Jaranoff, 1938]. Admittedly, such emphasis was mostly based on feelings rather than on data, and the same authors accepted that most Rhodopean granites and gneisses were Precambrian or Variscan. A stable Rhodope continental block during the Alpine orogeny was questioned in Greece [Meyer, 1968] and Bulgaria [Ivanov, 1988]. Its ancient origin was brought into dispute from:

- (1) the lack of Palaeozoic and/or Mesozoic sedimentary cover as already noted by Viquesnel [1853] but the point was forgotten by those who worked after him, although this cover is typical of basement regions in southern Europe, even in the Strandja Massif, next to and northeast of the Rhodope Metamorphic Complex [e.g. Görür *et al.*, 1997];

- (2) early geochronological determinations pointing to Mesozoic [Zagorčev and Moorbath, 1983; 1986; Soldatos *et al.*, 2008] to Eocene-Oligocene [Borsi *et al.*, 1965; Meyer, 1968; Pal'shin *et al.*, 1975] protolith ages of gneiss and "synkinematic" granitoids;

- (3) the consistency of synmetamorphic structures and kinematics with the bulk Mesozoic convergence between Europe and Africa and

- (4) a still thick crust with the Moho lying at ca. 50 km depth [Dačev and Petkov, 1978; Geiss, 1987].

The alternative was that the Rhodope is a complex of Alpine synmetamorphic nappes formed during closure of the Tethys [Burg *et al.*, 1990; Koukouvelas and Doutsos, 1990]. Multiphase recumbent folding had been established by several structural studies of the Rhodope gneissic complex, both in Bulgaria [Ivanov *et al.*, 1984; Ivanov *et al.*, 1985] and in Greece [Papanikolaou and Panagopoulos, 1981]. Polyphase deformation was a hint as to the correctness of the present-day structural interpretations whereby early thrusting and thickening of the crust has been largely overprinted by extensional metamorphic core complexes and associated low-angle detachment faults [Kolocotroni and Dixon, 1991; Dinter and Royden, 1993; Sokoutis *et al.*, 1993; Brun and Sokoutis, 2007]. Ductile extension makes a link with brittle extension that controlled formation of the Cenozoic sedimentary basins widely distributed over the Rhodope [e.g. Tzankov *et al.*, 1996]. The Rhodope nappe stack was overlain transgressively by Lutetian/ Priabonian (48–42 Ma) deep- to shallow-water sediments [Krohe and Mposkos, 2002] and was intruded by large-scale Tertiary granitoids, which led to local migmatisation of the host rocks [e.g. Peytcheva *et al.*, 2004; Liati, 2005].

Boundaries of the Rhodope

Kossmat [1924] first proposed that the Vardar Zone separates the Rhodope Massif, to the east from the Pelagonian Massif, to the west. He extended the Rhodope, through the Aegean Sea, to Anatolia (Asia Minor).

The Serbo-Macedonian Massif was separated from the Rhodope Massif by Kockel and Walther [1965] based on the observation of different metamorphic grades on either side of the Strymon Valley and apparent differences in lithological contents and inferred ages. Kockel and Walther [1965] placed the boundary on the eastern border of the Tertiary Strymon Basin along their west-dipping Strymon Thrust, later revealed to be a Miocene normal fault [Dinter and Royden, 1993]. Similar lithologies with similar protolith and Cretaceous metamorphic ages on both sides of the Srymon basin [Himmerkus *et al.* 2007 and 2009a] show that distinction between the Serbo-Macedonian and the Rhodope is not necessary.

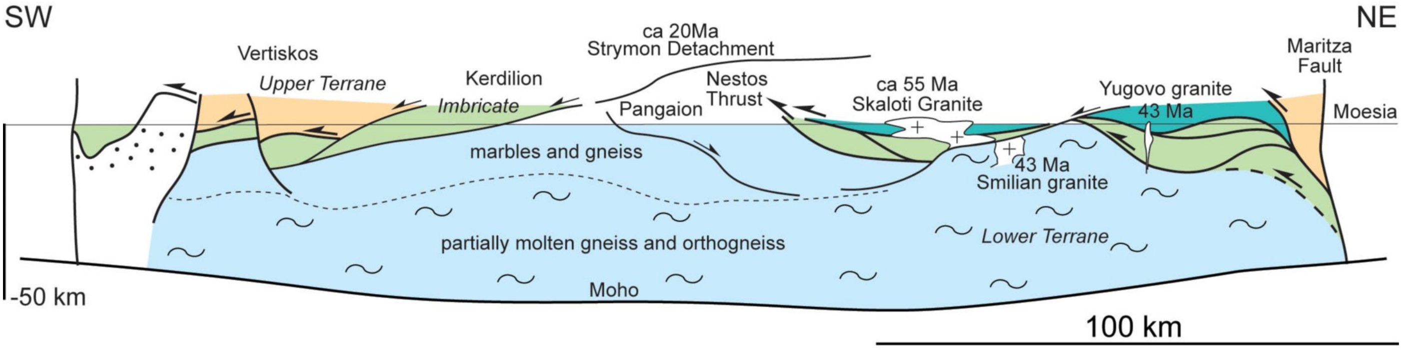
As envisaged nowadays, the Rhodope thrust system incorporates the Serbo-Macedonian Massif such that the Rhodope Massif extends to where early authors placed its western boundary: along the Vardar Suture Zone [Ricou *et al.*, 1998]. The northern boundary of the Rhodope Massif is the dextral Maritza strike-slip fault (Fig. 2), which deforms late Jurassic granitoids [Naydenov *et al.*, 2009] and developed Late Cretaceous (ca. 100 Ma, $^{40}Ar/^{39}Ar$) syn-metamorphic shear fabrics [Velichkova *et al.*, 2004; Rieser *et al.*, 2008].

To the northwest, in Bulgaria, the top-to-NW Gabrov Dol Detachment (Fig. 2) separates amphibolite-facies rocks now attributed to the Rhodope Massif, from lower grade sequences (Struma Diorite, Vlasina and Frolosh formations in the literature) and non- or weakly-metamorphosed and fossiliferous Palaeozoic sequences unconformably overlain by Permian detrital sediments [Bonev *et al.*, 1995]. This detachment is older than the crosscutting 73 Ma Plana pluton [Boyadjiev, 1981]. Although this K-Ar age constraint needs confirmation from a more robust isotopic system, it is supported by the 82.25 ± 0.22 Ma U/Pb zircon age of the Varshilo Granite, which belongs to the same plutonic system as the Plana Granite [Von Quadt and Peytcheva, 2005], and by the ca. 90 Ma zircon fission-track ages in both the footwall and hanging wall of the detachment [Kounov *et al.*, 2010]. Importantly, these ages demonstrate that at least the northwestern Rhodope was already tectonically unroofed by Late Cretaceous times. To the east, gneiss with similar characteristics as those reviewed below might be a lateral extension of the Rhodope Metamrophic Complex, in Turkey [Bonev and Beccaletto, 2007].

General structure, strain and kinematics

The overall structure of the Rhodope is the 300 x 300 km, open antiform identified in early work (Fig. 3). This NW-SE antiform bends lithological contacts transposed into the regional main-phase foliation during multiphase, tight to isoclinal recumbent folding. Multiphase designates a sequence of structures overprinting older structures that pertain to several generations of mostly coaxial folds [e.g. Meyer, 1969; Papanikolaou and Panagopoulos, 1981; Ivanov *et al.*, 1985; Burg *et al.*, 1996b]. Strain gradients from protomylonites to ultramylonites and ubiquitous sense-of-shear criteria indicate that foliation and folds result from intense, non-coaxial ductile deformation [Burg *et al.*, 1990; Burg *et al.*, 1993]. Ductile shear zones were active under amphibolite-facies conditions and delineate tectonic contacts between terranes with distinct structural-metamorphic histories. Therefore, the Rhodope metamorphic complex is viewed as a region of large-scale nappe tectonics (Figs. 2 and 3).

Figure 3. Rhodope Metamorphic Complex.



Synthetic cross section of the Rhodope Metamorphic Complex (approximate trace on Figure 2). Moho depths after Velchev *et al.* [1971], Dačev and Petkov [1978] and Geiss [1987].

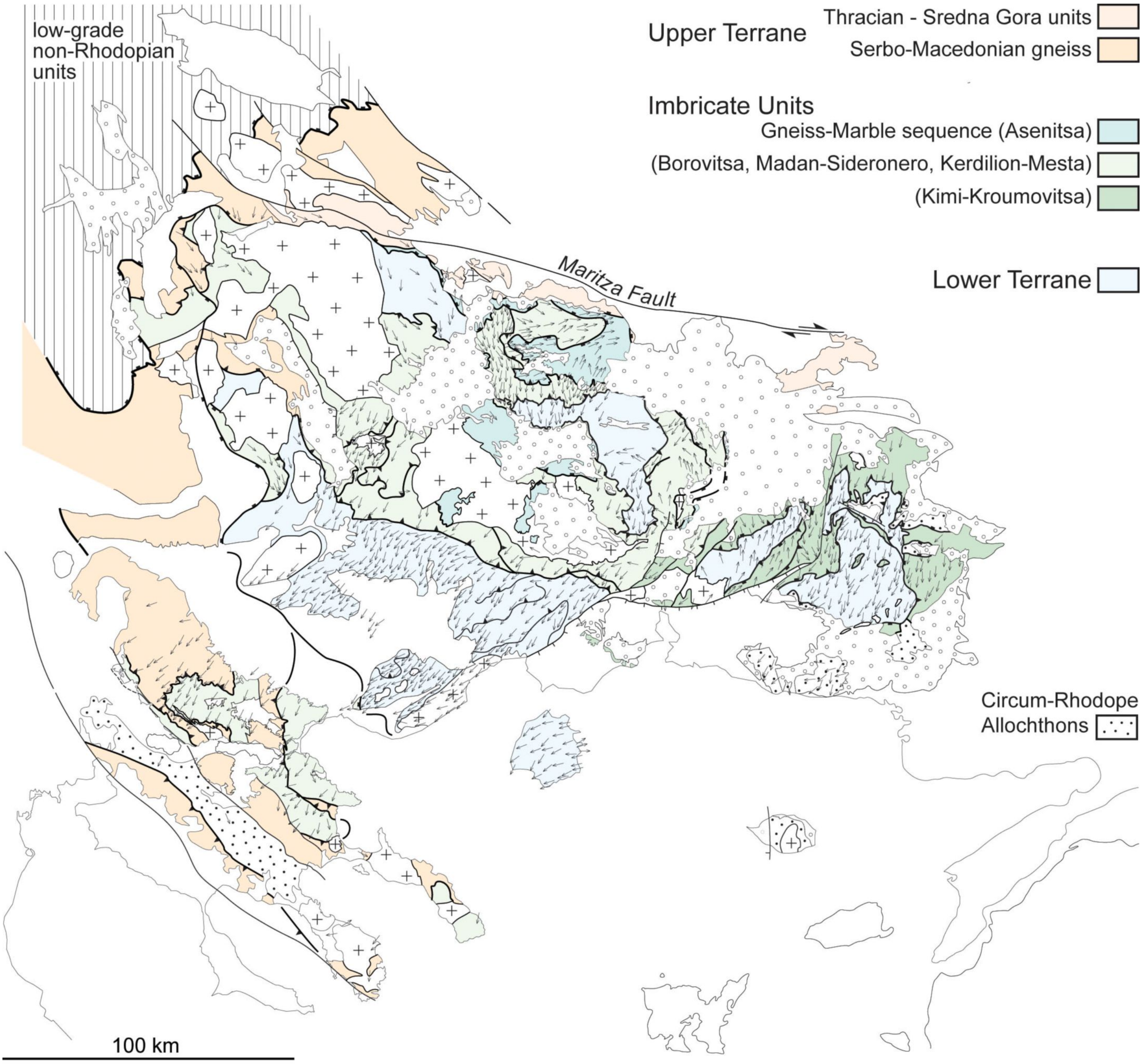
The bulk structure is simplified as a crustal-scale, syn-metamorphic, amphibolite-facies duplex [Ricou *et al.*, 1998]: the top and bottom units are different associations of paragneiss, orthogneiss and marbles. They are the hanging wall and footwall of a complex imbrication including meta-ophiolites and relicts of a Jurassic magmatic arc. These imbricates, previously lumped under the term of intermediate units [Ricou *et al.*, 1998], define a dismembered suture. Thrusts placed higher grade rocks onto lower grade rocks during intermediate- to high-pressure metamorphic conditions, before pervasive, syn-kinematic equilibration in amphibolite- and greenschist-facies conditions [Burg *et al.*, 1996a; Burg *et al.*, 1996b]. The main-phase foliation and shear zones contain the dominantly, almost homogeneous north-northeast lineation pattern defined by isoclinal and exceptionally sheath-fold axes, mineral lineations, and boudinage [Burg *et al.*, 1996a; Burg *et al.*, 1996b]. The general attitude of mylonites, foliations and stretching lineations (Fig. 4) demonstrate a regionally consistent, bulk southwestward thrusting [Burg *et al.*, 1990; Kilias and Mountrakis, 1990; Burg *et al.*, 1996a; Barr *et al.*, 1999]. Finite strain measurements and quartz fabrics indicate that deformation was close to plane strain [Burg *et al.*, 1996b]. In Greece, the intermediate and lower units were identified as Upper and Lower Units [Papanikolaou and Panagopoulos, 1981], respectively. At variance with earlier descriptions, and in the light of more recent work referred to hereafter, the hanging-wall terrane of this review is continental (the Serbo-Macedonian of previous authors), while the ophiolitic unit defined as roof unit in Burg *et al.* [1993] is now considered as one of the intermediate imbricates.

Erosion, tectonic denudation and deposition of colluvial - proluvial sediments unconformable on the metamorphic rocks occurred as early as Maastrichtian - Paleocene times [Boyanov *et al.*, 1982; Goranov and Atanasov, 1992]. This was followed by a period of apparent quiescence (a question that needs clarification) before widespread, graben-forming extension and intermediate to acid volcanism in Eocene-Oligocene times. Some thrust zones were then occasionally reworked as low-angle normal faults. The structural overprint leads to disputes where folded thrusts are tilted to attitudes with apparent normal sense of shear. Yet, there are distinct, low-dip shear zones and brittle normal faults that additionally displaced parts of the thrust system [Burg *et al.*, 1990; Krohe and Mposkos, 2002].

The amount of Cenozoic volcanism above a subduction zone, the unusual high topography and large crustal thickness today hint at possible similarities between the Andes and the Rhodope mountains that have been discussed, up to now, in terms of collision only. This question refers to first-order features that will be examined in the interpretation paragraph of this review.

A second phase of general extension began in the late Miocene. It is still active and related to the Aegean extension in the hanging wall plate of the rolling back Aegean subduction [e.g. Zagorčev, 1992].

Figure 4. Map of stretching lineations over the Rhodope metamorphic Complex.



All arrows point towards the local shear direction. Note that the pattern does not allow separating lineations of different age clusters. Same symbols as Figure 2.

Plate tectonic setting - geophysical constraints

More than 2500 km of anticlockwise, rotational convergence between Africa and Europe since about 140 Ma has produced the Hellenides-Dinarides-Rhodope orogenic system [Savostin *et al.*, 1986; Schettino and Scotese, 2005]. Paleogeographic reconstructions agree on the implication of several continental blocks derived from Gondwana and separated by Tethys-related oceanic basins [Ricou *et al.*, 1998; Stampfli and Borel, 2002]. Seismic tomography images a continuous, slab-type high-velocity anomaly to about 1500 km depth, with a 300 km flat segment lying on the 660 km discontinuity [Bijwaard *et al.*, 1998]. The length of the slab fits the inferred amount of post-Jurassic convergence, whilst the 300 km long flat slab would measure the amount of slab rollback responsible for the late Cenozoic extension in the Aegean [Van Hinsbergen *et al.*, 2005]. These results imply that no slab breakoff has occurred on the Vardar-Aegean side since the Jurassic. They also suggest that the Mediterranean slab did not begin to subduct under the Aegean continental plate in Miocene times but instead was continuously subducted at an average convergence rate of 2-2.5 cm/a over the last 100 Ma [Hafkenscheid *et al.*, 2006]. Subduction possibly involved several continental and oceanic lithospheres [e.g. Jolivet and Brun, 2010].

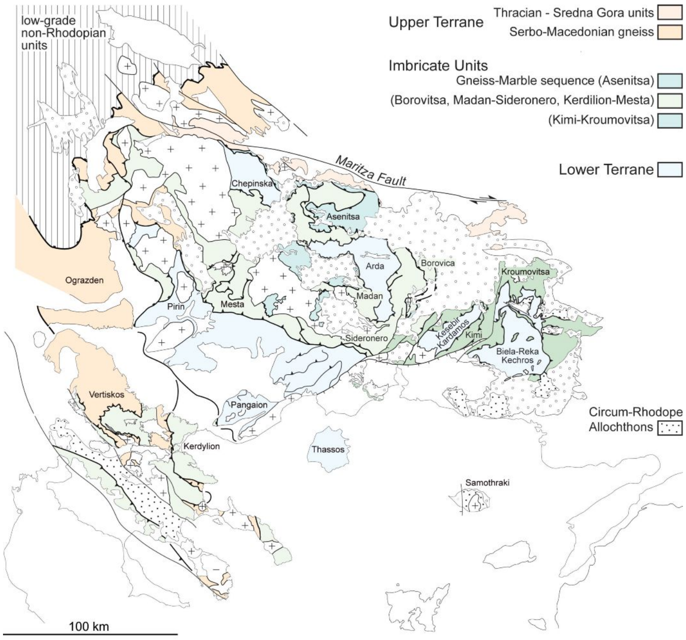
Paleomagnetic measurements document a clockwise rotation of much of the Rhodope by > 12° since the mid-Oligocene [Dimitriadis *et al.*, 1998]. Therefore, it is almost due to unfortunate coincidence that post-Oligocene, extension-related, NE-SW lineations are parallel to the older, thrust-related lineations. The latter should be turned back towards more southerly-directed directions to integrate corresponding kinematics into the Tethys collisional framework.

Supportive evidence for the Mesozoic age of the tectonic and metamorphic events in the Rhodope Metamorphic Complex has been granted in the recent years, as reviewed in the following paragraphs. Therefore, they must pertain to this geophysically documented, long-lived subduction system.

High-pressure rocks in imbricates; Jurassic-Cretaceous subduction

The multiplicity of local terms for the intermediate, imbricate units reflects some variability in lithological content, which in turn may reflect different crustal fragments. The extensive terminology is, in this review, simplified to the most common names reported in figure 5. Two main types of lithological subunits are distinguished, from bottom to top:

Figure 5. Location of the principal unit names found in the literature and used in this review.



Same shade colours and symbols as Figure 4.

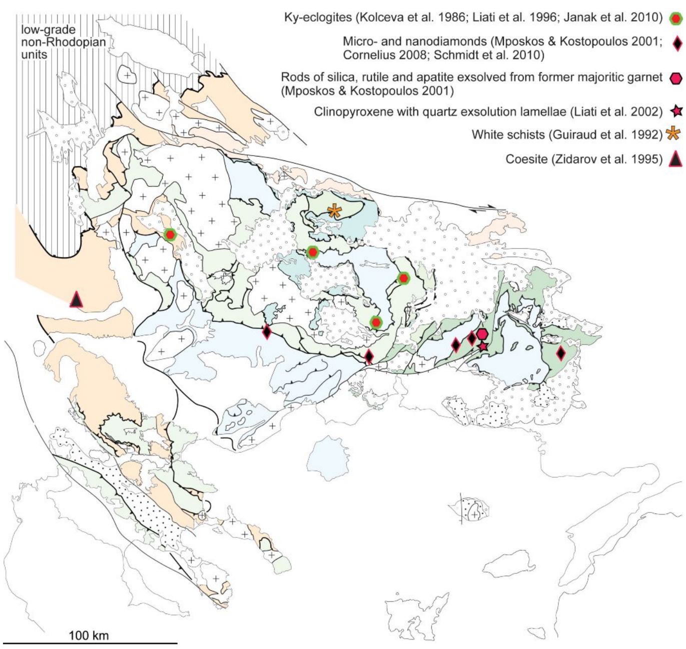
- 1) Eclogite-metabasic-gneiss sequence: The imbricates exposing this sequence include the Kerdylion, Sideronero, Kimi formations, in Greece, which find their equivalence in the Mesta, Madan (also Arda2), and Kroumovitsa formations in Bulgaria (Fig. 5). Paragneiss, minor graphitic marbles, and subordinate micaschists screen bodies of metamorphosed gabbros, diorites and granitoids. These rocks contain metamorphosed ultramafic and mafic bodies that locally preserved eclogite-facies parageneses [Kozhoukharova, 1984a; 1984b; Kolceva *et al.*, 1986; Kolčeva and Eskenazy, 1988; Liati and Mposkos, 1990; Sapountzis *et al.*, 1990]. High-pressure mineral assemblages are also preserved in pelitic rocks [Guiraud *et al.*, 1992]. Granulite-facies parageneses formed during retrogression from eclogite to amphibolite-facies have been described in places [Kolceva *et al.*, 1986; Liati and Seidel, 1996; Liati *et al.*, 2002]. The scattered high-pressure rocks were overprinted by regional, amphibolite-facies metamorphism [680-560°C at 0.6-0.3 GPa, e.g. Georgieva *et al.*, 2002] while the strongly mylonitic fabric of the country gneiss tended to be reset by partial melting. Ages of clastic zircons in paragneiss and marbles of this unit, to the north of Xanthi, attest for sedimentation younger than 300-280 Ma and a likely Gondwana source [Liati *et al.*, 2011].

- 2) Gneiss-marble sequences: Structurally higher, lower grade units (Asenitsa and Borovitsa in Bulgaria, Fig. 2 and 5) contain thin marble sequences interlayered with para- and orthogneiss and minor amphibolites. The Asenitsa sequence is overlain by massive and coarse-grained marbles [Ivanov *et al.*, 1984]. No convincing high-pressure and ultra-high-pressure relict has been documented in this essentially metasedimentary unit.

High-pressure metamorphic rocks

Eclogites have recorded various metamorphic histories according to their location and retrogression paths. Highest metamorphic pressures are ca 2 GPa at 700-800°C both in Bulgaria [Kolceva *et al.*, 1986; Kolčeva and Eskenazy, 1988; Janák *et al.*, 2011] and Greece [Liati and Seidel, 1996]. There are typically as many publications as there are outcrops because each rock has its own petrological specificity, including evidence for early, ultrahigh metamorphic pressures (Fig. 6). However, the regional information can be simplified. All high-pressure parageneses and retrogression paths generally document isothermal decompression to about 1 GPa followed by nearly isobaric cooling to the regional amphibolite-facies [0.8-1.1 GPa, 580-750°C, Mposkos, 1989; Machev and Kolcheva, 2008]. While some eclogites went through high-pressure granulite-facies [800°C - 1.5 GPa, Liati and Seidel, 1996; 700ºC at 1.26 GPa, Carrigan *et al.*, 2002] others, as in Eastern Rhodope, went through blueschist-facies metamorphism [Tzontcheff-Bonev, 1992].

Figure 6. Location and references of the ultra-high to high pressure parageneses reported in the Rhodope Metamorphic Complex.



Same shade colours and symbols as Figure 4.

Ultrahigh pressure conditions inferred from quartz exsolution lamellae in clinopyroxene [Liati *et al.*, 2002] typically reflect crystallization of magmatic pyroxene in the mantle and cannot be extended to the whole region. Indeed, these rocks crop out in the Kimi area in association with ultramafic rocks that are mainly mantle lherzolites and peridotitic cumulates with garnet and clinopyroxene [Mposkos, 2001]. They could be related to the 160 Ma arc. Ultrahigh pressure metamorphism is more convincingly documented by coesite in kyanite-eclogites [Zidarov *et al.*, 1995]. Microdiamond inclusions in garnet from paragneisses [Mposkos and Kostopoulos, 2001; Perraki *et al.*, 2006; Schmidt *et al.*, 2010] indicate that some of these rocks recrystallized within the microdiamond stability field, which is however very sensitive to fluid compositions [e.g. Simakov *et al.*, 2008].

Ages

Many ages have been produced, often without clear description of the tectonic and structural context of the sampled rocks and often with disputable relationships between ages obtained from mineral domains and regional geology / metamorphic history.

Protoliths

Protolith ages define two main groups: Palaeozoic and Jurassic-Cretaceous (Table 1).

The older group is Carboniferous to Permian, from ca. 300 to ca. 250 Ma. Zircon cores of two eclogitic gabbros are 245.6 ± 3.9 [Liati, 2005] and 255.8 ± 2.1 Ma [Liati *et al.*, 2011]. Cores of monazites between 265 and 295 Ma [Bosse *et al.*, 2009] may also witness this magmatic event.

Two gabbros of the Kroumovitsa unit are an exception showing protolith ages >500 Ma and metamorphic rims of 350-300 Ma [Carrigan *et al.*, 2003]. They fall in a time bracket identified also in the Upper Terrane [e.g. Himmerkus *et al.*, 2006], which raises the question as to whether this imbricate is a "Serbo-Macedonian" thrust sheet.

Concordant zircons from orthogneisses indicate Late Jurassic-Early Cretaceous intrusions from ca. 160 to ca. 130 Ma (Table 1). Forty zircon grains dated between 121 and 159 Ma [Bosse *et al.*, 2009] may represent this magmatic event. The 117 Ma oscillatory zircon domain of a garnet-mafic rock (Table 1) may also reflect the protolith age [Liati *et al.*, 2011].

Table 1. Geochronological data: Protolith ages from Intermediate Units

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb zircon	
Gabbro *(Bubino*)*	572 ± 5		[Carrigan *et al.*, 2003]
Metaplagiogranite *(S-Kesebir*)*	511 ± 5	mean age	[Bonev *et al.*, 2010a]
Amphibolite *(S-Kesebir*)*	459-434	core	[Bonev *et al.*, 2010a]
Metagabbro *(S-Kesebir*)*	474 ± 6	core	[Bonev *et al.*, 2010a]
Orthogneiss *(Sidironero)*	294 ± 8		[Liati & Gebauer, 1999]
Migmatitic orthogneiss	294.3 ± 2.4		[Liati, 2005]
Migmatitic orthogneiss *(Thermes)*	291.4 ± 3.4	inherited core	[Turpaud & Reischmann, 2010]
Augengneiss *(Siroko)*	275.8 ± 3.9		[Turpaud & Reischmann, 2010]
Garnet-gneiss *(Kimi)*	290-247	core	[Liati *et al.*, 2011]
Eclogite *(NE Komotini)*	255.8 ± 2.1		[Liati *et al.*, 2011]
Metagabbro *(Drama-Sideronero)*	245.6 ± 3.9		[Liati, 2005]
Biotite-gneiss *(Sminthi)*	164.4 ± 7.1	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]
Biotite-gneiss *(N-Drama)*	163.4 ± 2.1		[Turpaud & Reischmann, 2010]
Metadiorite *(Thermes)*	158.7 ± 1.7	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]
Orthogneiss *(Bachkovo)*	153.5 ± 4.1		[Von Quadt *et al.*, 2006]
Orthogneiss *(Zlatograd*)*	151.9 ± 2.2	Concordant zircons	[Ovtcharova *et al.*, 2004]
Orthogneiss *(Kimi)*	151.5 ± 2.0	Oscillatory domain	[Liati *et al.*, 2011]
Orthogneiss *(General Geshevo*)*	149.0 ± 0.66	Concordant zircons	[Ovtcharova *et al.*, 2004]
Orthogneiss *(Thermes)*	148.7 ± 5.6	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]
Biotite-gneiss *(Echinos)*	137.8 ± 5.1	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
Biotite-gneiss *(Paranesti)*	136.5 ± 4.3	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]
Orthogneiss *(Paranesti)*	134.0 ± 3.5	Pb-Pb Evaporation	[Turpaud & Reischmann, 2010]
40 zircons *(W-Xanthi)*	159 to 121		[Bosse *et al.*, 2009]
		Rb-Sr	
Orthogneiss *(Kechros)*	334 ± 5	Muscovite	[Mposkos & Wawrzenitz, 1995]

High-pressure metamorphism

Ages obtained in western, central and eastern Rhodope are similar, which fits a first-order process that goes beyond regional variations. The UHP metamorphism had probably started during the Early Jurassic [older than 170 Ma, Reischmann and Kostopoulos, 2002; Bauer *et al.*, 2007]. The Sm-Nd method applied to a 1.5-1.6 GPa pyroxenite from Eastern Rhodope yielded 119 ± 3.5 Ma, interpreted as the age of the HP metamorphism [Wawrzenitz and Mposkos, 1997]. U-Pb SHRIMP analyses on oscillatory zoned zircon of a garnet-bearing basic rock from Central Rhodope yielded an equivalent, weighted mean age (117.4 ± 1.9 Ma), which was considered as dating the protolith [Liati *et al.*, 2002] but may very well date early metamorphic conditions (no clear core). Metamorphic, zircon rims from a garnet-kyanite paragneiss for which high-pressure conditions have been suggested are dated at 148.8 ± 2.2 Ma, complemented by ages of 147.2 ± 4.7 Ma (paragneiss) and 143.4 ± 3.3 Ma (strongly amphibolitized eclogite) zircon domain ages [Liati, 2005].

Amphibolite-facies recrystallization

Further constraints are provided, in Central Rhodope, by oscillatory zircon domains at 73.5 ± 3.4 Ma interpreted to reflect HP/UHP metamorphism [Liati, 2005] but possibly dating regional, amphibolite-facies metamorphism. This weighted mean age for metamorphism is further supported by the 61.9 ± 1.9 Ma pegmatite vein that intruded these rocks [Liati *et al.*, 2002] and the 65-63 Ma trondhjemitic veins cutting amphibolitized eclogites in eastern Rhodope [Baziotis *et al.*, 2007] where an additional Rb-Sr isochron age of 65.4 ± 0.7 Ma is given for another cross-cutting pegmatite [Liati *et al.*, 2002]. In eastern Rhodope, two syn- to post-deformation granites were dated at ca. 70 Ma [Marchev *et al.*, 2006]. In western Rhodope, zircon rims of a Palaeozoic gabbro are 51.0 ± 1.0 Ma [Liati, 2005]. These dated veins provide solid evidence that amphibolite-facies deformation was waning by ca. 50 Ma. The number of ages distributed with no obvious hiatus between this upper bound and ca 170 Ma (Tables 2 and 3) reflects a protracted residence under evolving eclogite-, granulite- and amphibolite-facies conditions.

Table 2. Geochronological data for (ultra-) high-pressure metamorphism in the Rhodope massif.

Rock type *(location)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb zircon	
Metapelite *(Kimi)*	171 ± 1		[Bauer *et al.*, 2007]
Eclogite *(Kimi)*	> ca 160		[Bauer *et al.*, 2007]
Paragneiss *(Siroko)*	148.8 ± 2.2		[Liati, 2005]
Paragneiss *(Siroko)*	147.2 ± 4.7	domain	[Liati, 2005]
Eclogite *(Siroko)*	143.4 ± 3.3	domain	[Liati, 2005]
Garnet-gneiss *(Kimi)*	153-139	inner rim	[Liati *et al.*, 2011]

Rock type *(location)*	Age (Ma)	Method	Reference
Metapelite *(Chepelare*)*	142-137	monazite cores	[Bosse *et al.*, 2010]
Garnet-amphibolite *(Kimi)*	117.4 ± 1.9	domain	[Liati *et al.*, 2002]
Metapelite *(Kimi)*	160 ± 1	HT overprint	[Bauer *et al.*, 2007]
Metapelite *(W-Xanthi)*	148 to 121	granulitic overprint?	[Krenn *et al.*, 2010]
Eclogite *(Kimi)*	ca 115	HT overprint	[Bauer *et al.*, 2007]
		Sm-Nd	
Amphibolite *(Volvi)*	153±13		Kostopoulos, pers.com, 2010
Paragneiss *(NW-Xanthi)*	140 ± 4		[Reischmann & Kostopoulos, 2002]
Garnet-pyroxenite *(Kimi)*	119 ± 3.5		[Wawrzenitz & Mposkos, 1997]
		$^{40}Ar/^{39}Ar$	
Mylonite *(Chalkidiki)*	142.98 ± 4.89	white mica	[Lips *et al.*, 2000]

Table 3. Metamorphic overprint in high-pressure rocks and their country rocks and ages of post deformational granites and pegmatite veins.

Rock type *(location, *= Bulgaria)*	Age (Ma)	Method	Reference
		U-Pb zircon	
Paragneiss *(Siroko)*	82.8 ± 1.3		[Liati, 2005]
Pegmatites *(Chepelare)*	around 77		[Bosse *et al.*, 2009]
Eclogite *(Kimi)*	79 ± 3		[Bauer *et al.*, 2007]
Garnet-gneiss *(Kimi)*	73.9 ± 0.8		[Liati *et al.*, 2011]
Garnet-amphibolite *(Kimi)*	73.5 ± 3.4		[Liati *et al.*, 2002]
Pyroxenite *(Kimi)*	72.9 ± 1.1		[Liati *et al.*, 2011]
Orthogneiss *(Kimi)*	71.4 ± 1.1		[Liati *et al.*, 2011]
Orthogneiss *(Bachkovo*)*	55.9 ± 7.2		[Von Quadt *et al.*, 2006]
Garnet-amphibolite *(Sideronero)*	51.0 ± 1.0		[Liati, 2005]
Granite *(Chuchuliga)*	68.94 ± 0.4		[Marchev *et al.*, 2006]
Granite *(Rozino)*	68 ± 15		[Marchev *et al.*, 2006]
Discordant pegmatite *(Kimi)*	61.9 ± 1.9		[Liati *et al.*, 2002]
		Rb-Sr	
Undeformed pegmatite *(Kimi)*	65.4 ± 0.7	White mica	[Mposkos & Wawrzenitz, 1995]

Tertiary recrystallization

42 Ma zircons in a retrogressed eclogite were interpreted as dating the eclogite-facies event [Liati and Gebauer, 1999] while zircon rims between ca. 40 to ca. 38 Ma in migmatites containing the eclogites and in an adjacent amphibolite have been interpreted as dating the regional, amphibolite-facies metamorphism [Liati, 2005]. These ages (Table 4) are within the range of the numerous K-Ar and $^{40}Ar/^{39}Ar$ amphibole and mica ages between 50 and 35 Ma reported for many rocks of the

intermediate units [Liati and Kreuzer, 1990; Kaiser-Rohrmeier *et al.*, 2004] and with the Eocene age of zircons from discordant leucosomes [Ovtcharova *et al.*, 2002; Peytcheva *et al.*, 2004]. They are also coincident with the many cooling ages measured within this time span all over the Rhodope Metamorphic Complex (Tables 4, 6 and 7) and the ca. 35 Ma age of hydrothermal deposits and volcanic rocks in Eastern Rhodope [Márton *et al.*, 2010]. Since K-Ar and $^{40}Ar/^{39}Ar$ are reportedly low temperature systems, one cannot exclude that the 42-38 zircons also registered retrogression down to hydrothermally influenced metamorphic conditions of ca. 300°C - 0.3 GPa [Liati and Seidel, 1996]. As such, the many Tertiary ages may simply record cooling from amphibolite to greenschist-facies. Zircon and apatite fission-track ages overlapping K-Ar and $^{40}Ar/^{39}Ar$ dates (Table 4) demonstrate very fast cooling in Oligocene times [Wüthrich, 2009; Márton *et al.*, 2010].

Table 4. Geochronological data for Tertiary thermal event and cooling in the Intermediate units of the Rhodope massif.

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb zircon	
Synfolial pegmatite *(W-Xanthi)*	49.6 ± 3.9		[Bosse *et al.*, 2009]
Metagabbro *(S-Kesebir*)*	49.1 ± 6	rim	[Bonev *et al.*, 2010a]
Synfolial pegmatite *(W-Xanthi)*	48.2 ± 2.2		[Bosse *et al.*, 2009]
Quartz vein *(Sminthi)*	45.3 ± 0.9		[Liati & Gebauer, 1999]
Eclogite *(Thermes)*	42.2 ± 0.9		[Liati & Gebauer, 1999]
Migmatitic orthogneiss *(Thermes)*	42.1 ± 1.0		[Liati & Gebauer, 1999]
Orthogneiss *(Thermes)*	42.0 ± 1.1		[Liati & Gebauer, 1999]
Leucosome *(Thermes)*	40.0 ± 1.1		[Liati & Gebauer, 1999]
Leucosome *(Sideronero)*	39.7 ± 1.2		[Liati, 2005]
Leucosome *(Sideronero)*	38.1 ± 0.8		[Liati, 2005]
Garnet-amphibolite *(Sideronero)*	38.1 ± 1.2		[Liati, 2005]
Leucosome	37.08 ± 0.38		[Ovtcharova *et al.*, 2002]
Pegmatite *(Sminthi)*	36.1 ± 1.2		[Liati & Gebauer, 1999]
		Monazite	
Synfolial pegmatite *(W-Xanthi)*	54.9 ± 1.7	core	[Bosse *et al.*, 2009]
Orthogneiss *(Zlatograd*)*	47.4 ± 0.66		[Ovtcharova *et al.*, 2004]
Pegmatite *(Tchepelare*)*	42.1 ± 1.2		[Bosse *et al.*, 2009]
Metapelite *(Chepelare*)*	42-38	rims	[Bosse *et al.*, 2010]
Synfolial pegmatite *(W-Xanthi)*	38.6 ± 1.1	rim	[Bosse *et al.*, 2009]
Leucosome	37.8 ± 1.5		[Ovtcharova *et al.*, 2002]
Orthogneiss *(Banite-Gulubovo*)*	35.83 ± 0.4		[Peytcheva *et al.*, 2004]

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		$^{40}Ar/^{39}Ar$	
Eclogite *(Belopolci*)*	45 ± 2	amphibole	[Mukasa *et al.*, 2003]
Orthogneiss *(Kazak*)*	39 ± 1	muscovite	[Mukasa *et al.*, 2003]
Gneiss *(Pelevun*)*	39.28 ± 0.24	muscovite	[Márton *et al.*, 2010]
Gneiss *(Kremenitz*)*	37.28 ± 0.19	muscovite	[Márton *et al.*, 2010]
Amphibolite *(Ada Tepe*)*	36.9 ± 0.16	muscovite	[Márton *et al.*, 2010]
Adularia *(Kuklitza*)*	35.94 ± 0.36	adularia	[Márton *et al.*, 2010]
Amphibolite *(Ada Tepe*)*	36.9 ± 0.16	muscovite	[Márton *et al.*, 2010]
Orthogneiss *(NW-Pilima)*	35.4 ± 0.4	biotite	[Moriceau, 2000]
Orthogneiss *(Banite-Gulubovo*)*	35.35 ± 0.22	biotite	[Peytcheva *et al.*, 2004]
Orthogneiss *(Pilima)*	35.3 ± 0.4	muscovite	[Moriceau, 2000]
Gneiss *(Davidkovo*)*	35.25 ± 0.36	muscovite	[Kaiser-Rohrmeier *et al.*, 2004]
Orthogneiss *(NW-Pilima)*	35.0 ± 0.4	muscovite	[Moriceau, 2000]
Adularia *(Ada Tepe*)*	34.95 ± 0.36	adularia	[Márton *et al.*, 2010]
Pegmatite *(Tchepelare*)*	34.9 ± 0.1	muscovite	[Bosse *et al.*, 2009]
Volcanite *(Iran Tepe*)*	34.62 ± 0.46	amphibole	[Márton *et al.*, 2010]
Synfolial pegmatite *(W-Xanthi)*	34.3 ± 0.2	biotite	[Bosse *et al.*, 2009]
Synfolial pegmatite *(W-Xanthi)*	33.2 ± 0.3	muscovite	[Bosse *et al.*, 2009]
Gneiss *(Imera)*	32.0 ± 0.3	muscovite	[Moriceau, 2000]
		Rb-Sr	
Paragneiss *(NW-Xanthi)*	37	muscovite	[Reischmann & Kostopoulos, 2002]
Pegmatite *(Banite-Gulubovo*)*	35.31 ± 0.25		[Peytcheva *et al.*, 2004]
Paragneiss *(NW-Xanthi)*	34	biotite	[Reischmann & Kostopoulos, 2002]
		K/Ar	
Amphibolites *(NW-Xanthi)*	95-57	hornblende	[Liati & Kreuzer, 1990]
Orthogneiss *(E-Kardamos)*	42.1 ± 1	muscovite	[Krohe & Mposkos, 2002]
Orthogneiss *(E-Kardamos)*	39.4 ± 1	biotite	[Krohe & Mposkos, 2002]
		Fission-track	
Gneisses *(Central Rhodope*)*	ca. 35	several zircons	[Wüthrich, 2009]
Migmatites *(Starcevo*)*	33.2 ± 3.8	apatite	[Wüthrich, 2009]
Migmatites *(Borovica*)*	33.0 ± 4.4	apatite	[Wüthrich, 2009]
Gneisses *(Central Rhodope*)*	35-20	several apatites	[Wüthrich, 2009]
Amphibolite *(Pelevun*)*	25.0 ± 1.5	apatite	[Márton *et al.*, 2010]
Gneiss *(Kremenitz*)*	18.3 ± 1.9	apatite	[Márton *et al.*, 2010]

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
Amphibolite *(Ada Tepe*)*	14.8 ± 5.1	apatite	[Márton *et al.*, 2010]

Protolith compositions

Geochemical analyses document a variety of protoliths. Trace-element and REE ratios of eclogites and associated ultrabasic and basic rocks [Kozhoukharova, 1984a; 1984b; Kolčeva and Eskenazy, 1988; Liati *et al.*, 1990] were suggested to have been derived from mid-ocean ridge basalts with a tholeiitic trend of differentiation and a close relationship to ocean-type lithosphere (dominance of harzburgites and wehrlites with orthopyroxenite veins and dunite). However, the given analyses are mostly T-MORBs (Kostopoulos, pers. comm, 2010). Therefore, they may represent the oceanic floor of attenuated lithosphere, such as a marginal basin that would have retained some continental crust [Barr *et al.*, 1999]. Supporting this hypothesis, trace-element geochemistry and igneous zircon U-Pb ages (SHRIMP II) for some eclogite/amphibolite strata also suggest intrusion of basaltic sills and dykes, of T-MORB affinity, into attenuated continental basement during the Early Triassic [Liati, 2005].

In contrast, amphibole-bearing and biotite orthogneisses are evolved volcanic-arc intrusions [Cherneva and Georgieva, 2005] while some amphibolites derived from basalt to basaltic andesites also define volcanic arc affinity [Liati and Seidel, 1996].

The anatectic para- and orthogneiss, marbles and eclogitic to amphibolitic metabasites of the Kerdylion unit (the lowest Serbo-Macedonian, in Greece, Fig. 5) are comparable to those of the Mesta and Sideronero imbricates.

These protolith compositions demonstrate that the Rhodope metamorphic pile has involved an active margin environment before and during synmetamorphic thrusting. The question is whether subduction occurred below an island arc or a continental margin. This question has importance for any geodynamic reconstruction and will be discussed in the relevant paragraph of this review.

Lower terrane: Footwall continental plate

The lower terrane is the Pangaion (also called Boz Dağ Unit) of the Greek literature [e.g. Kronberg *et al.*, 1970; Kronberg and Raith, 1977]. It extends towards the NW, into Bulgaria, where it is called Pirin Unit. This terrane represents a microcontinent with a carbonate platform that cores the large antiform from the Rila Mountains to the island of Thassos through the Pirin and Pangaion mountains. Synmetamorphic thrusts, such as the pervasive and much studied Nestos (Meso-Rhodopean) Thrust Zone (Figs. 2 and 3) [Papanikolaou and Panagopoulos, 1981; Zachos and Dimadis, 1983; Gerdjikov and Milev, 2005], are responsible for regional metamorphic inversion, placing higher amphibolite-facies intermediate terranes (Mesta, Sideronero, Kerdylion) onto upper-greenschist to lower amphibolite-facies rocks of the structurally lower terrane [Papanikolaou and Panagopoulos, 1981; Mposkos, 1989]. The activity of such thrust zones is likely older than, and lasted until, the ca. 55 Ma, syn-folial yet late-deformation pegmatite veins, while a spread of younger ages, derived from a variety of geochronological methods, refers to lower-grade reactivation and/or fluid circulation [Bosse *et al.*, 2009].

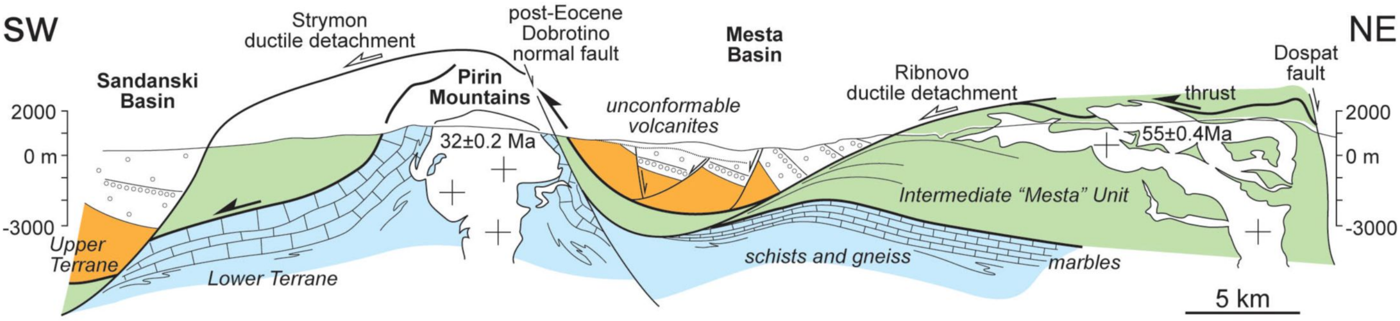
In this review, and in the light of protolith ages, we additionally attribute to the lower terrane the string of four separated domes. Those are, from northwest to southeast, the Chepinska, Arda, Kesebir and Biela Reka "units" in Bulgaria; the last two are called Kardamos and Kechros in Greece (Fig. 5). These domes expose monotonous, quartzo-feldspatic, strongly deformed gneiss of dioritic composition intruded by metagranitoids, some of which are presumably syntectonic. This interpretation would define the Nestos Thrust Zone (Fig. 2) as the ramp where the imbricates climb up the footwall sequence from the deeper orthogneissic levels, to the north, over the carbonate platform, to the south.

Lithological content - Metamorphism

The tectonostratigraphy of the lower terrane defined by the Thassos-Pangaion-Pirin half-window comprises from bottom to top i) a unit of schists underlying ii) pre-eminent marbles, iii) leucocratic orthogneiss and paragneisses, and iv) an upper unit of micaschists, amphibolites and thin intercalations of marbles [Meyer *et al.*, 1963; Birk *et al.*, 1970; Jacobshagen, 1986]. The lower-terrane is particularly well identified in the Pangaion area where marbles are involved in a hinterland-dipping thrust system [Kilias and Mountrakis, 1990]. Inverted, intermediate-pressure metamorphism evolves from greenschist-facies conditions in the Pangaion [Fig. 3, Zachos and Dimadis, 1983] to sillimanite-bearing migmatites to the north, against the Nestos Thrust [Mposkos *et al.*, 1989]. Peak metamorphic conditions recorded in metapelites of the lower schists, in the Thassos Island, reached ca. 600-680°C for 0.6-0.8 GPa [Dimitriadis, 1989; Schulz, 1992].

The migmatite-gneiss sequence that cores the four northern domes is placed, in this review, below the tectonostratigraphy listed for the Thassos-Pangaion-Pirin antiform (Figs. 3 and 7). Consistent with this interpretation, minor marble and amphibolites occur at the top of this deeper migmatite-gneiss sequence. Evidence for high-pressure metamorphism [Mposkos and Liati, 1993; e.g. 450°C-1.3 GPa, Macheva, 1998] is rare. Leucosomes due to partial melting of the gneiss are common and more pervasive northwestward. In effect, the regional amphibolite-facies metamorphic grade is usually lower in the Biela-Reka--Kechros dome [lower amphibolite-facies, ca 550°C-0.6 GPa, Macheva, 1998] than in the central (Arda) and western (Chepinska) regions [higher amphibolite-facies and migmatites at ca. 650°C-0.7 GPa, Cherneva and Georgieva, 2007].

Figure 7. Cross section across the Pirin Mountains, located in Fig. 2.



See also Moriceau [2000], Burchfiel *et al.* [2003] and Georgiev *et al.* [2010].

Ages

Protolith

Tubular features found in the lower section of the marbles [Meyer *et al.*, 1963] have been tentatively determined as coral forms of Silurian to Carboniferous age [Rugosa according to R. Wolfart in Jordan, 1969]. Drilled marbles in Bulgaria yielded a Mid-Ordovician to Early Carboniferous brachiopod [Atrypida? according to O.V. Bogoyavlenskaya, in Ancirev *et al.*, 1980]. These faunas exclude Precambrian lithological and metamorphic ages and support the interpretation of non-layered marble bodies being reef structures within the sequence [Kronberg, 1966; Jordan, 1969]. A reef-platform environment would explain strong thickness variations that are attributed to sedimentary features rather than to pervasive isoclinal folding [Jordan, 1969]. However, the description is not sufficiently informative to know whether the drilled marbles belong to the Lower terrane or top the Vertiskos Upper Terrane.

Zircon U-Pb ages from orthogneisses point to Palaeozoic granitoids to be the main protoliths (from ca 300 to ca 270 Ma, Table 5). These magmatic ages demonstrate that the continental block placed at the bottom of the Rhodope thrust system had been in the realm of the Variscan orogen before the assembly of Pangea. Orthogneisses of the Arda dome (Fig. 5) have chemistry typical of syn-collisional peraluminous leucogranites to late-collisional granites [Cherneva and Georgieva, 2005].

Table 5. Geochronological data: Protolith ages from the Lower terrane.

Rock type *(location, *= Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb zircon	

Rock type *(location, *= Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
Orthogneiss *(Kesebir*)*	334 ± 5		[Peytcheva & Von Quadt, 1995]
Orthogneiss *(Stoyanov Bridge*)*	310.7 ± 4.6		[Ovtcharova *et al.*, 2002]
Metagranodiorite *(Banite*)*	310 ± 11		[Peytcheva *et al.*, 2004]
Orthogneiss *(Biela Reka*)*	301 ± 4		[C.W. Carrigan *et al.*, 2003]
Augengneiss *(Pilima)*	291.2 ± 8.8		[Turpaud & Reischmann, 2010]
Augengneiss *(Pilima)*	289.5 ± 7.6		[Turpaud & Reischmann, 2010]
Augengneiss *(N-Drama)*	286.4 ± 4.0		[Turpaud & Reischmann, 2010]
Augengneiss *(Thassos)*	282.9 ± 4.8		[Turpaud & Reischmann, 2010]
Biotite gneiss *(N-Drama)*	282.7 ± 3.0		[Turpaud & Reischmann, 2010]
Leucocratic gneiss *(Kavala)*	281.1 ± 6.4		[Turpaud & Reischmann, 2010]
Biotite gneiss *(Kato Nevrokopi)*	278.7 ± 7.7		[Turpaud & Reischmann, 2010]
Leucocratic gneiss *(W-Kavala)*	276.6 ± 9.5		[Turpaud & Reischmann, 2010]
Augengneiss *(Siroko)*	275.8 ± 3.9		[Turpaud & Reischmann, 2010]
Augengneiss *(W-Paranesti)*	275.5 ± 3.6		[Turpaud & Reischmann, 2010]
Leucocratic gneiss *(N-Kavala)*	269.7 ± 9.0		[Turpaud & Reischmann, 2010]

Table 6. Geochronological data covering Tertiary thermal event in the Lower terrane of the Rhodope massif *(see references for more ages)*.

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		Rb-Sr	
Pegmatoid *(Thassos)*	51.4 ± 0.8	white mica	[Wawrzenitz & Krohe, 1998]
Pegmatoid *(Thassos)*	51.3 ± 0.7	white mica	[Wawrzenitz & Krohe, 1998]
Pegmatoid *(Thassos)*	51.2 ± 0.5	white mica	[Wawrzenitz & Krohe, 1998]
Pegmatoid *(Thassos)*	40.3 ± 0.4	biotite	[Wawrzenitz & Krohe, 1998]
Pegmatoid *(Thassos)*	39.0 ± 0.4	biotite	[Wawrzenitz & Krohe, 1998]
Orthogneiss *(Kechros)*	37.2 ± 0.3	white mica	[Wawrzenitz & Mposkos, 1997]
Various gneiss *(Thassos)*	27.4 to 12	many micas	[Wawrzenitz & Krohe, 1998]
Paragneiss *(Pangaion)*	22.6 ± 0.7	muscovite	[Del Moro *et al.*, 1990]
Paragneiss *(Pangaion)*	22.3 ± 0.7	muscovite	[Del Moro *et al.*, 1990]
Paragneiss *(Pangaion)*	18.3 ± 0.6	muscovite	[Del Moro *et al.*, 1990]
Paragneiss *(Pangaion)*	12.6 ± 0.4	biotite	[Del Moro *et al.*, 1990]
Paragneiss *(Pangaion)*	12.2 ± 0.4	biotite	[Del Moro *et al.*, 1990]
		$^{40}Ar/^{39}Ar$	

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
Orthogneisses *(several places)*	42-19	white micas	[Lips *et al.*, 2000]
Gneiss *(Kato Nevrokopi)*	34.0 ± 0.3	muscovite	[Moriceau, 2000]
Gneiss *(Kato Nevrokopi)*	30.9 ± 0.4	biotite	[Moriceau, 2000]
Granodiorite *(Kavala)*	15/11/11	biotite-K-feldspar	[Dinter *et al.*, 1995]
		Fission-track	
Various rocks *(Thassos, Kavala)*	< 10	apatite	[Hejl *et al.*, 1998]
Gneiss+schists *(Drama, Serres)*	11.8 to 17.8	apatite	[Hejl *et al.*, 1998]

Metamorphism

Metamorphic ages mostly obtained from micas refer to a set of cooling ages between ca. 50 and <15 Ma for temperatures higher than ca 250°C (Table 6). These ages, similar to those reported for the Intermediate Units (Table 4), are unevenly distributed and thus refer to a regional thermal system that affected all tectonic units, disregarding tectonic contacts. Youngest ages centered on the Pangaion denote the influence of the Strymon Detachment.

Upper terrane: hanging wall continental plate

The hanging wall continent exhibits a Cadomian to Variscan basement [Carrigan *et al.*, 2005; Carrigan *et al.*, 2006] with Early Permian granitoid intrusions and an Early Triassic to Middle Jurassic sedimentary cover metamorphosed to greenschist-facies and deformed by thick-skinned, north-directed thrusting during the Late Jurassic [Okay *et al.*, 2001]. These features typify the European continent and find equivalence within the "Serbo-Macedonian" part of west Rhodope, namely the Vertiskos in Greece [Kockel *et al.*, 1977] and Ograzden in Bulgaria [Zagorčev, 1976]. In this review, the highest, high-grade structural unit of East Rhodope, the so-called Kroumovitsa, is one of the Rhodope intermediate imbricates.

Lithological content

Regional investigations and correlations based on map continuity, similarities of lithologies, in particular that of basic and ultrabasic rocks, structures, geochronological ages and strain regime reveal the wide extension of the upper terrane. Rocks are mostly quartzo-feldspathic migmatites containing bodies of basic and ultrabasic rocks, some of which are eclogites retrogressed into the regional amphibolite-facies [Dimitriadis and Godelitsas, 1991; Zidarov *et al.*, 1995; Zidarov and Nenova, 1995]. In Greece, the Upper continental terrane is essentially the "Serbo-Macedonian", which has been subdivided into the lower Kerdylion Unit and the higher Vertiskos Unit [Kockel *et al.*, 1971].

The Kerdylion Unit consists of gneisses and migmatites with both Permo-Carboniferous and Late Jurassic protolith ages [Himmerkus *et al.*, 2007]. Such protolith ages offer equivalence with the eclogite-metabasic-gneiss intermediate terrane of the Rhodope (Sideronero in Greece) and further demonstrate that the concept of "Serbo-Macedonian", under its original definition, should be discarded. Metamorphic mafic and ultramafic rocks, the so-called Volvi Ophiolites with supra-subduction marginal basin chemistry, occur at the boundary between the Vertiskos Unit, and the underlying Kerdylion unit [Dixon and Dimitriadis, 1984]. These mafic and ultramafic rocks are Late-Permian to Triassic in age [252 ± 13 Ma, Liati *et al.*, 2011] and may have been the basement of the Rhodope Arc at the continent-ophiolite transition. Alternatively, they may represent a distinct, accreted oceanic fragment.

The Vertiskos/Ograzden Unit is part of the roof of the Rhodope Metamorphic Complex. It is a composite unit comprising a metaophiolite-bearing mylonite zone (ultrabasic rocks are mostly serpentinized harzburgites) between a lower metaturbiditic and orthogneissic sequence and an upper migmatitic para- and orthogneissic sequence [Burg *et al.*, 1995]. The ca. 250 Ma metaophiolites (Table 8) bear the geochemical signature of a marginal basin [Dixon and Dimitriadis, 1984]. Ages and chemistry of the gneisses of the upper sequence demonstrate a distinct, Gondwana-derived microcontinent with its complex tectonic, magmatic and metamorphic history.

Table 7. Geochronological data for granitoid intrusions in the Rhodope Massif

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb	
Granitoid *(Rila*)*	69.3 ± 0.3		[Von Quadt & Peytcheva, 2005]
Granitoid *(Rila*)*	66.8 ± 0.3		[Von Quadt & Peytcheva, 2005]
Granitoid *(Spanchevo*)*	56 ± 0.5		[Jahn-Awe *et al.*, 2010]
Granodiorite *(Skaloti)*	55.93 ± 0.28		[Soldatos *et al.*, 2008]
Granodiorite *(Dolno Dryanovo*)*	55 ± 0.4		[Jahn-Awe *et al.*, 2010]
Granite *(Ierissos)*	53.6 ± 6.2	uranothorite	[Frei, 1996]
Granitoid *(Pripek*)*	52.89 ± 0.89		[Ovtcharova *et al.*, 2004]
Granite *(Kalin)*	ca 46		[Arnaudov *et al.*, 1989]
Granitoid *(Smilian*)*	43.4 ± 1.41		[Ovtcharova *et al.*, 2003]
Granitoid *(Yugovo*)*	42.3 ± 0.54		[Ovtcharova *et al.*, 2003]
Granodiorite *(Rila-Pirin)*	ca 42		[Peytcheva *et al.*, 1998]
Pegmatite *(Sminthi)*	36.1 ± 1.2		[Liati & Gebauer, 1999]
Granitoid *(Teshovo*)*	32 ± 0.2		[Jahn-Awe *et al.*, 2010]
Granodiorite *(Stratoni)*	27.1 ± 1.1	average	[Frei, 1992]
Granodiorite *(Kavala)*	21.1 ± 0.8	titanite	[Dinter *et al.*, 1995]
		Rb-Sr	
Granitoid *(Kresna*)*	82 ± 22	whole rock	[Zagorčev & Moorbath, 1983]
9 Granitoids *(Leptokaria)*	34.9 to 28.4	biotite	[Del Moro *et al.*, 1988]
Granite *(Central Pirin*)*	37 ± 2	whole rock	[Zagorčev *et al.*, 1987]
Pegmatite *(Banite*)*	35.31 ± 0.25	biotite	[Peytcheva *et al.*, 2004]
		$^{40}Ar/^{39}Ar$	
Granitoid *(Ouranopolis)*	47 ± 0.7	muscovite	[De Wet *et al.*, 1989]
Granodiorite *(Sithonia)*	43 ± 0.6	biotite	[De Wet *et al.*, 1989]
Pegmatite *(SW-Sminthi)*	28.4 ± 0.4	biotite	[Moriceau, 2000]
Granodiorite *(Mesoropi)*	21.7 ± 0.5	hornblende	[Eleftheriadis *et al.*, 2001]
Granodiorite *(Mesoropi)*	13.8 ± 0.5	biotite	[Eleftheriadis *et al.*, 2001]
		K/Ar	
Diorite *(Plana*)*	ca 73	biotite	[Boyadjiev, 1981]
Granitoid *(Dautov*)*	30-41	biotite	[Boyadjiev & Lilov, 1976]
Pegmatite dyke *(Tholos)*	38.3 ± 1.1	muscovite	[Meyer, 1968]
Monzodiorite *(Kentavros)*	38	hornblende	[Liati & Kreuzer, 1990]

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
Granitoid *(Vrondou)*	30 ± 3	biotite	[Durr *et al.*, 1978]
Granodiorite *(Xanthi)*	30 ± 1	hornblende	[Liati & Kreuzer, 1990]
Granitoid *(Stratoni)*	29.6 ± 1.4		Papadakis in [Kilias *et al.*, 1999]
Granodiorite *(Xanthi)*	27.9 ± 0.5	biotite	[Meyer, 1968]
Granodiorite *(Xanthi)*	27.1 ± 0.4	biotite	[Meyer, 1968]
Granodiorite *(Granitis)*	28.2 ± 0.5	biotite	[Meyer, 1968]
Granodiorite *(Panorama)*	26.8 ± 0.5	biotite	[Meyer, 1968]
Granodiorite *(Krinides)*	26.0 ± 0.5	biotite	[Meyer, 1968]
Granodiorite *(Kavala)*	17.8 ± 0.8	biotite	[Kokkinakis, 1980]
Granodiorite *(Kavala)*	15.5 ± 0.5	biotite	[Kokkinakis, 1980]
Granodiorite *(Mesolakkia)*	15.0 ± 0.3	biotite	[Harre *et al.*, 1968]
Granodiorite *(Mesolakkia)*	13.8 ± 0.2	biotite	[Harre *et al.*, 1968]

Table 8. Geochronological data for protolith ages in the Upper Terrane (more ages in references).

Rock type *(location, *=Bulgaria)*	**Age (Ma)**	**Method**	**Reference**
		U-Pb or Pb-Pb	
Orthogneiss *(Pirgadikia)*	587.6 ± 3.4	evaporation	[Himmerkus *et al.*, 2006]
Orthogneiss *(Pirgadikia)*	570.0 ± 7.0	evaporation	[Himmerkus *et al.*, 2006]
Paragneiss *(Taxiarchis)*	555.8 ± 2.6	evaporation	[Himmerkus *et al.*, 2006]
Orthogneiss *(Pirgadikia)*	433.0 ± 2.1	evaporation	[Himmerkus *et al.*, 2006]
Orthogneiss *(Pirgadikia)*	428.2 ± 1.2	evaporation	[Himmerkus *et al.*, 2006]
20 orthogneisses *(Vertiskos)*	426 to 444	evaporation	[Himmerkus *et al.*, 2009a]
Metaophiolite *(Volvi)*	252 ± 13		[Liati *et al.*, 2010]
Granitoid *(Skrut*)*	248.85 ± 0.70		[N Zidarov *et al.*, 2007]
Granite *(Kerkini)*	247 ± 2		[Christofidies *et al.*, 2006]
Granite *(Chortiatis)*	240.7 ± 2.6	evaporation	[Himmerkus *et al.*, 2009b]
Granite *(Arnea)*	228.8 ± 5.6	evaporation	[Himmerkus *et al.*, 2009b]
Granite *(Serres)*	221.7 ± 1.9	evaporation	[Himmerkus *et al.*, 2009b]

The Sredna Gora Zone belongs to the Upper terrane along the northern border of the Rhodope. Like the Vertiskos/Ograzden Unit, a composite "basement" of metasediments with amphibolites, eclogites and orthogneisses [e.g. Zagorčev *et al.*, 1973] bears evidence of Palaeozoic (Variscan) metamorphic and magmatic activity [Carrigan *et al.*, 2006]. Lithological and age comparisons can be made with the Strandja Massif, to the east [e.g. Okay *et al.*, 2001]. These rocks are therefore tentatively ascribed to the Upper Rhodope Terrane and associated in this review with the Serbo-Macedonian, for the sake of simplification.

Ages

Protolith

Late-Proterozoic to Silurian [ca 590-430 Ma, Himmerkus *et al.*, 2006; Meinhold *et al.*, 2010a] Vertiskos gneisses have been intruded by Triassic (241-222 Ma) granites (Table 8). The Precambrian age of most of these gneisses was inferred from, unconformably overlying early Palaeozoic sediments [Kockel *et al.*, 1971; Zagorchev, 2001]. The Triassic magmatism, recognized in the Vertiskos only [Himmerkus *et al.*, 2009b], is attributed to the global rifting event that led to the opening of the Tethys Ocean [Himmerkus *et al.*, 2009b]. As everywhere in the Rhodope, a series of Tertiary granites intruded the Vertiskos (Table 7).

Metamorphism

The oldest orthogneiss were first metamorphosed during the Palaeozoic Variscan orogeny [Borsi *et al.*, 1965; Kockel *et al.*, 1977] and later again during the Early Cretaceous under lower amphibolite-facies conditions [Rb-Sr and K-Ar ages on Vertiskos hornblende and muscovite are between 116 and 90 Ma, Harre *et al.*, 1968; Papadopoulos and Kilias, 1985; and $^{40}Ar/^{39}Ar$ mica ages are ca. 135 Ma, De Wet *et al.*, 1989]. Cretaceous metamorphism and erosion are supported by a 71.9 ± 9.4 zircon fission-track age of Vertiskos gneiss, which yielded an apatite fission-track age of 43.0 ± 6.8 Ma [Wüthrich, 2009].

The Volvi basic rocks reached metamorphic temperatures of ca. 750°C [Dixon and Dimitriadis, 1984] whereas the lower unit of gneiss and metapelites did not reach more than 600°C at less than 0.7 GPa [Papadopoulos and Kilias, 1985; Kilias *et al.*, 1999]. Kostopoulos *et al.* [2000] reported graphitised microdiamonds, taken as evidence for UHP metamorphism, in rocks of the Vertiskos Unit and the Circum-Rhodope Belt. With its early Palaeozoic and Variscan signatures, the Vertiskos displays a clear European affinity. The abundance of Triassic magmatism in the Variscan basement points to an attenuated crust during opening of the Tethys Ocean or a branch of it.

Circum-Rhodope Allochthons

Several units of low-grade Mesozoic rock sequences were termed “Circum-Rhodope Belt” by Kauffmann *et al.* [1976]. This definition refers to the concept of the Rhodope being an old microcontinent island surrounded (hence the prefix “circum”) and stratigraphically covered by younger sediments [Jaranoff, 1960; Kockel *et al.*, 1977]. However, all contacts are fault zones, and different rock types and associations were lumped under the term. Detrital heavy minerals demonstrate that indeed there are different units with different source areas [Meinhold *et al.*, 2009; Meinhold *et al.*, 2010b]. The fact that they occur along the western and southern margins of the high-grade Rhodope is their unifying character.

To the southwest, there are greenschist- and blueschist-facies metasediments, including deep marine Triassic sediments and metabasalts [Michard *et al.*, 1994]. They were initially interpreted as the cover of the Vertiskos Unit [Kockel *et al.*, 1971] but the contact is demonstrably tectonic [Meinhold *et al.*, 2009]. The blueschist-facies [and possibly local UHP, Kostopoulos *et al.*, 2000] metamorphism must be at least Jurassic in age because similar metamorphic rocks are found as pebbles in non-metamorphosed Tithonian to Lower Cretaceous conglomerates [Kockel *et al.*, 1971]. Like many other blueschists, they may have been exhumed in an accretionary wedge, before any collisional event. Greenschist-facies metamorphism is Eocene [Kockel *et al.*, 1977].

Low-grade, Late-Permian and Mesozoic sediments are associated with Jurassic arc and back-arc magmatic rocks in Eastern Rhodope [Magganas, 2002; Bonev *et al.*, 2010a]. There, greenschist-facies metabasalts and phyllites of prehnite-pumpellyite-facies [Maratos and Andronopoulos, 1964] tectonically overlie the high-grade, eclogite-bearing Kroumovitza unit [Bonev and Stampfli, 2008]. These dominantly Jurassic rocks [see review of fossiliferous and absolute ages in Bonev and Stampfli, 2008] were emplaced by northward thrusting during the Early Cretaceous, and represent a volcanic island arc and backarc basin. Interestingly, the 155 Ma Samothraki diorite [Tsikouras *et al.*, 1990] and the nearly 160 Ma Samothraki mafic suite [Koglin *et al.*, 2009] could belong to this arc-backarc unit. Bonev and Stampfli [2008] argue, after Magganas [2002], that this backarc basin was linked to the Vardar system above a south-dipping subduction zone that preceded arc-continent collision. Ages of detrital zircons suggest that these sediments were deposited in front of an eroded source with Rhodopean affinity [Meinhold *et al.*, 2010b]. Importantly, these rocks were never buried deeper than low-grade metamorphic conditions and the ca. 150 Ma fission-track ages [Bigazzi *et al.*, 1987] show that they escaped thermal and tectonic disturbance since mid-Jurassic times. As such, they are part of the Upper Terrane(s) preserved in the Rhodope.

Crustal extension

The collisional nappe pile is overprinted by extensional detachment faulting, and intruded by pre-, syn-, and post-tectonic granitoid plutons. Sense-of-shear criteria indicate bulk top-to-south-southwest shear, consistent with magmatic fabrics of late, often calc-alkaline granitoids [Kolocotroni and Dixon, 1991; Zananiri *et al.*, 2004] that deform the nappe system into large dome and basin structures (Fig. 1). SW-NE trending folds and lineations (Fig. 4) conventionally related to the thrusting event have been attributed to crustal extension [Dinter and Royden, 1993; Sokoutis *et al.*, 1993; Dinter, 1994]. The distinction between thrusting- and extension-related shearing is, in many places, structurally difficult. Extension-related ductile deformation seems coeval with thrusting structures because both evolved under similar metamorphic conditions. Such interacting features are easier to interpret as gravitational adjustment of an unstable orogenic wedge during its tectonic accretion. Late Eocene marine sediments which unconformably overlie the metamorphic Rhodope Complex markedly separate previous structures from normal faults active since the Oligocene throughout the Aegean realm [e.g. Angelier *et al.*, 1982; Lister *et al.*, 1984; Gautier and Brun, 1994].

Backward crustal stretching

The first evidence for crustal stretching and extension was inferred from both structural considerations and kinematic indicators opposed to (i.e. northeastward to subparallel to the strike of the orogen) and overprinting those denoting southwestward thrust tectonics. Evidence for backward shear with respect to the bulk regional shear was particularly discussed for some of the imbricate units. It is attributed to syn-orogenic extension.

Gneiss-marble sequence and eclogite-metabasic-gneiss sequence

In the structurally high gneiss-marble imbricate (so-called Asenitsa, Fig. 5) the foliation contains a stretching lineation related to top-to-east-northeast shear-sense criteria [Burg *et al.*, 1990; Burg *et al.*, 1996a]. This backward (top to NE) shear is linked with the exhumation from depths greater than 30 km, at a rate fast enough to prevent significant retrogression of the white-schist parageneses at the basal contact of the gneiss-marble imbricate [Guiraud *et al.*, 1992], and the eclogites in the underlying imbricates (Fig. 6).

Foliation-parallel scars

The shear-inversion zones between SW-directed and NE- or NW-directed sheared units are a few metres thick and show no attitude change of the low-dip foliation across the zone in question. Such zones were reported at several levels [Ricou *et al.*, 1998] where they may represent the foliation-parallel scar left by a missing unit displaced south-westward with respect to both the footwall and hanging wall units. Such contacts have locally been reduced to thin brittle zones where thick crustal segments are missing, for example between the high-grade allochthonous Kroumovitza hanging wall and medium-grade Biela-Reka-Kardamos footwall gneisses in Eastern Rhodope [Ricou *et al.*, 1998; Krohe and Mposkos, 2002]. The metamorphic gap is less pronounced but also exists between the Asenitsa and the underlying Arda2 (Madan) and Borovitsa units (Fig. 5). Forward extrusion of high grade units [Chemenda *et al.*, 1995] involves thinning in the scar area coeval with thickening further southwest, towards the foreland, where the extruded unit was transferred.

Low angle normal faults and basins

Semi-ductile shear zones active under greenschist-facies and overprinting, very low-grade brittle fault zones mark the late Cretaceous Gabrov Dol Detachment, which is conveniently taken here as the roof boundary of the Serbo-Macedonian-Rhodope Metamorphic Complex [Bonev *et al.*, 1995]. Other low-dip faults cut the foliation of the high-grade metamorphic rocks and show low-grade and brittle conditions [e.g. Ivanov, 1988]. They are particularly associated with Eocene-Oligocene half grabens and basins into which large olistoliths of metamorphic rocks have glided [Ivanov *et al.*, 1979; Burchfiel *et al.*, 2000; Burchfiel *et al.*, 2003].

The Strymon low-angle detachment is the most cited one. Dinter *et al.* [1995] noted that a major ductile shear zone overprints ductile structures associated with thrust tectonics along the eastern side of the Strymon Valley. Observing that the extensional fabric is pervasive in the Kavala Granodiorite and in the southwestern part of the Vrondou Granodiorite (Fig. 5), they distinguished a

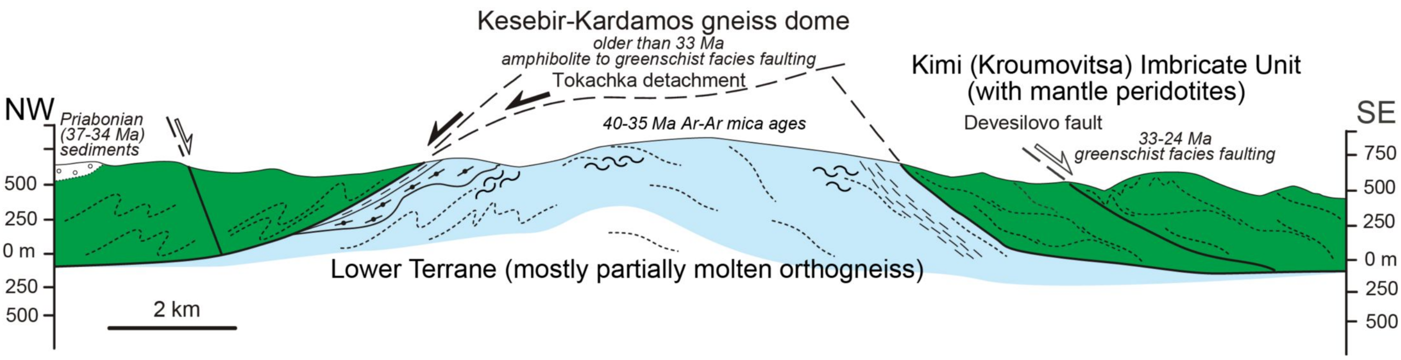
21-22 Ma ductile mid-crustal extension stage from a ca. 16 to 3.5 Ma ductile to brittle extensional movement [see also Dinter, 1998]. The Tertiary activity of the Strymon Detachment is documented by the contrasting 50 versus <26 Ma cooling ages between the hanging wall and the footwall in Thasos [Wawrzenitz and Krohe, 1998]. Contrasting cooling ages (K-Ar and $^{40}Ar/^{39}Ar$ on micas) between younger (Eocene-early Oligocene) footwall and older (mostly Cretaceous) hanging wall were also obtained for the west-dipping shear zone between the Vertiskos and Kerdylion units [Harre *et al.*, 1968; De Wet *et al.*, 1989; Frei, 1996]. Zircon and apatite fission-track ages on both sides of the Strymon Detachment on the eastern side of the Strymon Valley are also younger than those on both sides of the Vertiskos/Kerdylion contact [Wüthrich, 2009]. Ages do not allow correlating these two west-dipping fault zones.

Gneiss domes

Most of the large antiforms described in the early literature and ascribed to late compressional folding have now turned out to be (for the most part) extensional core complexes [Dinter and Royden, 1993; Sokoutis *et al.*, 1993; Brun and Sokoutis, 2004; Kounov *et al.*, 2004; Bonev *et al.*, 2006a]. Core granitoids and low-pressure anatexites hint at crustal melting during extension-related decompression of the gneiss exhumed below ductile, normal shear zones.

The Kesebir-Kardamos is one of the best-documented extensional gneiss domes [Fig. 8, Bonev *et al.*, 2006a; Krenn *et al.*, 2010]. A low-angle (Tokachka) detachment separates the core of intermediate-pressure, amphibolite-facies orthogneisses and migmatites of the Lower Terrane [0.3-0.9 GPa, 550-620°C, Mposkos *et al.*, 1989; Krohe and Mposkos, 2002] from the eclogite-metabasic-gneiss sequence of the Kimi-Kroumovitsa imbricate. Maastrichtian–Paleocene sediments [Goranov and Atanasov, 1992] rest directly on the fault contact. 40-35 Ma $^{40}Ar/^{39}Ar$ mica ages date cooling below 350-300°C of the core migmatites [Bonev *et al.*, 2006b; Márton *et al.*, 2010]. Zircon fission-track ages (all from 39.2 ± 4.2 to 35.6 ± 5.2 Ma) and apatite fission-track ages (35.8 ± 8.6 to 28.1 ± 6.2 Ma) show that both the hanging and the footwall of the Tokachka detachment cooled rapidly together from ~300°C down to 60°C without significant displacement after ca. 33 Ma [Wüthrich, 2009]. Reset zircon and apatite FT ages from both overlying sediments and subjacent basement rocks consistently indicate that the onset of high-angle normal faulting is placed between 33 and 24 Ma [Wüthrich, 2009; Márton *et al.*, 2010].

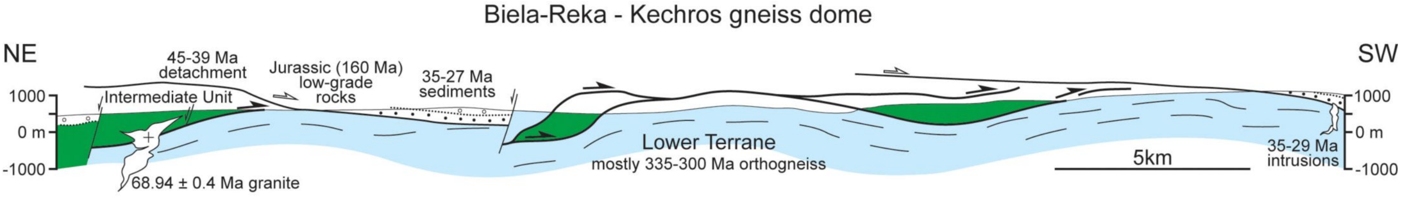
Figure 8. Cross section across the Kesebir-Kardamos extensional dome.



Adapted from Bonev *et al.* [2006a]. Trace on Figure 2.

Further east, the very flat Biela Reka-Kechros Dome (Fig. 9) is also interpreted as an extensional feature [Krohe and Mposkos, 2002; Bonev, 2006]. Likewise, most of the antiforms exposing the lower terrane are tectonic windows surrounded by ductile detachments, often reworking previous thrusts. The general structure of very flat domes such as the Biela-Reka-Kechros raises the question as to whether there should not be more caution in identifying extensional core complexes based on few normal senses of shear, since those are possible in any dome [Burg *et al.*, 2004] and a bent thrust may appear like a normal fault. In most cases only the contrasting thermal history of the upper plate in comparison to the lower plate will reveal whether any particular structure is related to extensional detachments or to thrusting.

Figure 9. Cross section across the Biela-Reka-Kechros dome.



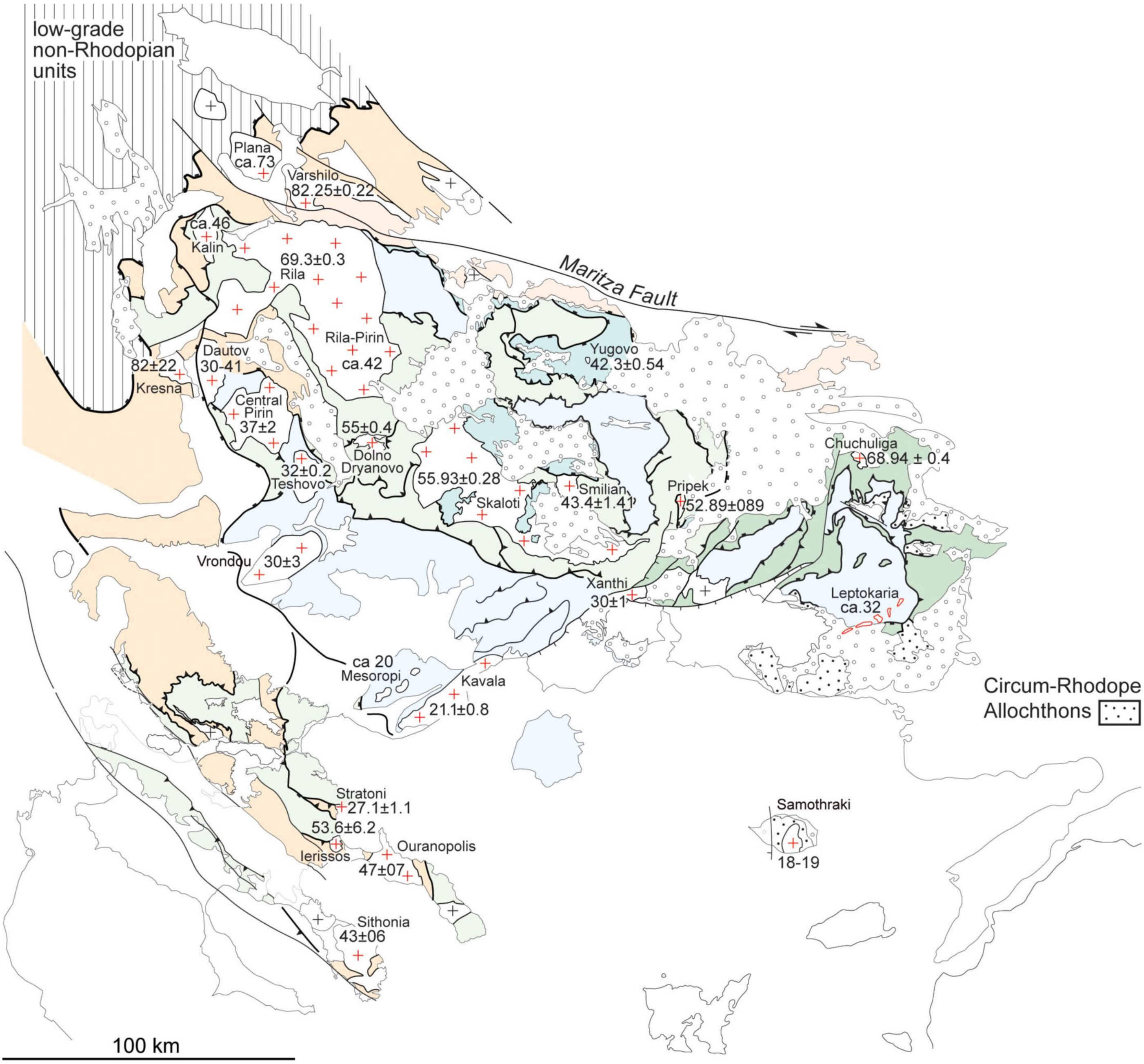
Trace on Fig. 2. See also Kozhoukharov [1987] and Bonev [2006].

Granitoids

Voluminous plutonism is one of the Rhodope characteristics noted in the earliest work [Viquesnel, 1853] and was used to infer an old basement at times when granitoids were professed to be rare in Alpine orogens. Geochronology (Table 7) has been the key to establishing the Rhodope peculiarity.

The Kavala [Dinter and Royden, 1993] and Vrondou [Kolocotroni and Dixon, 1991] plutons are dated at 21 and ca. 30 Ma, respectively (Fig. 10; Table 7). They display pervasive C/S type fabrics that generally indicate top to the SW or WSW, normal shearing [Dinter and Royden, 1993; Sokoutis *et al.*, 1993]. Owing to their Oligocene to Early Miocene intrusion age, they are likely the most convincing argument to demonstrate ductile, extensional deformation in the intruded middle crust at times when sediments were being deposited on the Rhodope surface.

Figure 10. Map of dated granitoids.



Ages and references in Table 7. Same shade colours and symbols as Figure 4.

The early to mid-Tertiary granitic intrusions in the Rhodope mostly represent calc-alkaline, deep crustal melts attributed to elevated temperatures in the mountain root and emplaced during extensional collapse of the thickened crust [Jones *et al.*, 1992]. The upwelling, decompressing asthenosphere would have been the heat source and produced melts with strong mantle signature [Koukouvelas and Pe-Piper, 1991]. Hydrothermal base- and precious-metal deposits are mostly related to the Oligocene magmatism [Singer and Marchev, 2000; Kaiser-Rohrmeier *et al.*, 2004; Marchev *et al.*, 2005].

Timing of extensional events

The sedimentary record of basins and grabens suggests several extensional episodes, which are coeval with sets of absolute ages and volcanic events. Although grabens, which are limited in size, record local events that can be heterogeneously distributed during a wider continuum, three regional events stand out.

Sedimentary information

Late Cretaceous-Paleocene sedimentary cover

The oldest post-metamorphic cover of the Rhodope is Maastrichtian-Paleocene [Athanasov and Goranov, 1984; Goranov and Atanasov, 1992], perhaps even Upper Santonian-Campanian [Boyanov *et al.*, 1982]. It consists of colluvial and proluvial clastic deposits with olistoliths of country gneiss and marbles [Kozhoukharov *et al.*, 1991; Goranov and Atanasov, 1992; Boyanov and Goranov, 1994; Zagorchev, 1998]. Campanian tuffs cover both high-grade gneiss and "circum-Rhodope" low-grade sequences in the eastern Rhodope [Boyanov and Russeva, 1989]. These sediments demonstrate that much of the metamorphic Rhodope was eroded to near sea-level by ca. 60 Ma and remained a shallow sedimentation site until at least 40 Ma. This early sedimentation history spans ca. 20 Ma and seems in conflict with the many cooling ages younger than 40 reported in tables 4 and 7. Twenty million years of apparent quiescence is a long time span in this very "mobile" zone, which raises the question of polyorogeny for the upper tectonic levels of the Rhodope.

Eocene-Oligocene basins

Early Eocene, marine sediments (Nummulite-bearing limestones) are preserved in several basins in Bulgaria and Greece [Von Braun, 1993; Zagorchev, 2001]. These sediments unconformably transgressed an erosional surface in the Priabonian (Late Eocene, ca. 35 Ma) and quickly gave way to Late Oligocene and Early Miocene continental sedimentation. Extensional structures involve the European continent beyond the Rhodope Metamorphic Complex, into northern Bulgaria and the Balkan Mountains [Tzankov *et al.*, 1996; Zagorchev *et al.*, 1999; Burchfiel *et al.*, 2000; Kounov *et al.*, 2004; Tueckmantel *et al.*, 2008; Schefer *et al.*, 2011]. Sedimentation continued into Early Miocene times to the northeast of the Rhodope (in Thrace). The Middle and early Late Miocene witnessed a general sedimentary break possibly coeval with general erosion [Burchfiel *et al.*, 2000].

Late Miocene to Present

A second extensional event started in the Middle Miocene. Late Miocene, alluvial and proluvial sediments rest unconformably on older rocks [Zagorčev, 1992; Dinter and Royden, 1993; Georgiev *et al.*, 2010]. Grabens formed at that time over a broad region that extends southward to the Aegean Sea [Mascle and Martin, 1990]. This event is generally related to clockwise rotation of the Greek Peninsula with respect to Europe during southward retreat of the Hellenic trench [McKenzie, 1978; Le Pichon and Angelier, 1981; Van Hinsbergen *et al.*, 2008].

Geochronological constraints

The decrease of ages of metamorphic rocks toward dome cores and deeper structural levels indicates distinct Cretaceous, Eocene, and Oligo-Miocene tectonic-metamorphic pulses that successively caused cooling and exhumation of gneiss complexes situated at deeper levels. The structural information from dated minerals, in particular micas, indicates that during Eocene to Miocene sedimentation on the surface, gneissic foliations and shear zones continued to form deeper in the crust.

The ca. 100 Ma Rb/Sr age [Zagorčev and Moorbath, 1986; Arnaudov *et al.*, 1990b] from a metamorphosed granitoid with the NE-directed shear fabric would date the earliest crustal stretching. However, NE-directed extensional shearing is constrained to ca. 155 Ma by $^{40}Ar/^{39}Ar$ mica ages in allochthonous units of Eastern Rhodope [Bonev *et al.*, 2010a]. Such ages suggest either heterogeneously distributed, protracted events, or polyorogeny already evoked, which will be discussed in the relevant paragraph later in this contribution.

For younger extensional events, important constraints come from deformed granites such as the 52.8±0.89 Ma Pripek laccolith [Gerdjikov, 2005, Table 7]. $^{40}Ar/^{39}Ar$ white mica ages of ca. 40 Ma distributed throughout the Rhodope constrain the Middle Eocene cooling history of Central [Kaiser-Rohrmeier *et al.*, 2004] and Eastern Rhodope [Bonev *et al.*, 2010a; Bonev and Stampfli, 2011] below ca. 350 °C. Rapid cooling of the Rhodope Metamorphic Complex is further documented by one titanite fission-track age at 55.9 ± 6.2 Ma in Central Rhodope [Wüthrich, 2009] and several zircon and apatite fission-track ages between 48 and 18 Ma in Central [Wüthrich, 2009] and Eastern [Márton *et al.*, 2010] Rhodope. Younger (15-6 Ma) fission-track ages occur in Western

Rhodope, in the direct footwall of the Strymon Detachment [Hejl *et al.*, 1998; Wüthrich, 2009].

Magmatism

Magmatism has been very variable in composition, space and time over the Aegean and North Aegean region since the Eocene [e.g. Pe-Piper and Piper, 2006]. This magmatism has been related to either hydrous melting of the asthenospheric mantle wedge during subduction of the Pelagonian or Vardar oceanic lithosphere [Boccaletti *et al.*, 1974b; Burchfiel *et al.*, 2000] or to a collisional - post-collisional event [Yanev, 2003]. In details, the magmatic history is more complex. Subduction-related, calc-alkaline magmatism was dominant during the Middle to Late Eocene [35-37 Ma, Lilov *et al.*, 1987]. That would link the coeval extension to arc--back-arc extension. During the Early Oligocene, massive calc-alkaline to shoshonitic volcanism accompanied sedimentation in fault-bounded basins [Harkovska *et al.*, 1989; Dabovski *et al.*, 1991; Pecskay *et al.*, 2000]. The abundance of rhyolites and ignimbrites points to crustal melting, hence a major thermal/decompressional event at 40-30 Ma [Marchev *et al.*, 2005]. Seismic tomography shows that there was no slab breakoff since the Mid Jurassic [Bijwaard *et al.*, 1998]. Therefore, this magmatic event requires another interpretation. Bimodal volcanism with rhyolitic and latitic–andesitic rocks associated with few basalts [Yanev *et al.*, 1998] was active along a 1600 km long belt, from the Eastern Alps through the Rhodope to Eastern Thrace in Turkey [Harkovska *et al.*, 1989; Schefer *et al.*, 2011]. The length of this magmatic belt supports a lithospheric-scale origin. For this reason, it is possible that the thermal event is lithospheric delamination and subsequent asthenopheric rise that could have melted a heterogeneously enriched subcontinental lithospheric mantle [Pe-Piper and Piper, 2006]. In any case, the cause would have been rather short-lived because magmatism ceased by the end of Oligocene times, after the effusion of alkali basalts between 28 and 26 Ma [Marchev *et al.*, 1998].

Magmatism shifted further south during the Miocene, with very local magmatic activity [Jones *et al.*, 1992].

Interpretation of the Rhodope Metamorphic Complex

The combination of four characteristics of the Rhodope massif lead to a geodynamical model that accounts for simultaneous thrusting and exhumation over a long period of time:

1) A long-lasting subduction is indicated by the large volume of Jurassic, arc-type magmatic protoliths within the Rhodope Metamorphic Complex, and by the Late Cretaceous opening of the Sredna Gora back-arc basin on its northern side [e.g. Boccaletti *et al.*, 1974a; von Quadt *et al.*, 2005]. This subduction was dipping below Eurasia as indicated by the position of the Europe-type remnants of the Upper continental Terrane upon the metamorphic complex and by the S-SSW-directed senses of shear found in pre-Maastrichtian thrust zones. This geological conclusion is reinforced by the tomographic image of a single, northward-dipping slab [Bijwaard *et al.*, 1998], which excludes the Rhodope from being a segment of the European-subducting side of the Balkan-Carpathian system.

2) The metamorphic complex contains a large amount of low-density continental material (marbles, orthogneiss and paragneiss) and a comparatively very small volume of basic and ultrabasic rocks. As argued by Ricou *et al.* [1998], the buoyancy of the arc and continental material favours decoupling and upward extrusion of such subducted rocks, which then migrate upward with respect to both the hanging wall and footwall tectonic units [Chemenda *et al.*, 1996]. At that stage, thrusting of high-grade metamorphic units over lower grade rocks can be responsible for inverse metamorphic zonation and inversion of synmetamorphic senses-of-shear from floor to roof contacts.

3) The metamorphic Rhodope was eroded to near sea-level by ca. 60 Ma and remained comparatively quiet until ca. 40 Ma. Then, ubiquitous extension began and reached a paroxysm between 35 and 30 Ma.

Island-arc or active continental margin?

The Late-Jurassic granitoids dated in the intermediate units of the Rhodope are products of a magmatic arc. Two questions then arise: (1) was subduction taking place beneath an island arc or a continental margin? (2) Was subduction south- or north-dipping below the Rhodope magmatic arc of that time?

Answering the first question makes use of modelling results on arc-continent collision [Chemenda *et al.*, 2001b; Boutelier *et al.*, 2003]. The accepted interpretation implies an active margin on the northern, European side [e.g. Ricou *et al.*, 1998], and this was based on the intimate association of presumed arc-derived and continental rocks. Now, many more arc rocks have been identified so that the island-arc possibility cannot be discarded without further discussion. Since the majority of the arc-rocks are plutonic, collision systems that would lead to complete arc subduction can be excluded. The lack or scarcity of volcanic and sedimentary sequences also excludes systems leading to accretion of the entire island arc such as the Kohistan [e.g. Bard, 1983; Treloar *et al.*, 1996]. Partial accretion of the island arc addresses the tectonic significance given to the upper crust of the island-arc recognized in the Circum-Rhodope Allochthons of Eastern Rhodope [Bonev and Stampfli, 2008]. One may also wonder whether the gneiss-marble sequences on the northern slope of the arc (Asenitsa-Borovitsa, Figs. 2 and 5) may represent island-arc volcaniclastic and carbonate rocks similar to those seen in today's Bismarck Sea [e.g. Hoffmann *et al.*, 2009]. Separation of the upper crust from the deeper, plutonic crust of the arc now found in the imbricate units involves decoupling within the arc, at the time it became involved in the trench. The possibility of an island arc in the Tethys is tectonically plausible. Collision would have scraped off the upper crust, which remained in low-grade metamorphic conditions. The middle and lower arc crust would have been, at the same time, deeply subducted to sublithospheric "ultrahigh" pressure conditions where they might undergo relatively low temperatures below the thermal shield provided by the subducted arc plate [Chemenda *et al.*, 2001b]. Involving an intra-oceanic arc allows resorption of large amounts of fore-, arc, and back-arc lithospheres, hence more convergence than along a continental margin.

This lack of confident determination between island-arc and continental margin at this stage of the discussion leads us to the question of subduction polarity.

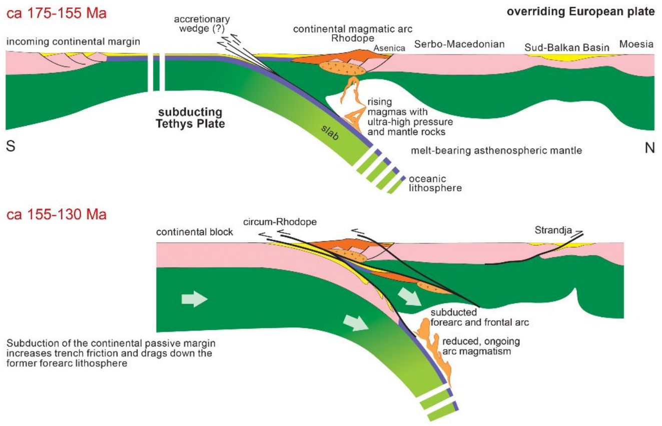
Subduction polarity

The northward sense of shear found in Circum-Rhodope Allochthons of Eastern Rhodope is inconclusive with regard to early orogenic kinematics over a S-dipping slab of the Meliata-Maliac oceanic lithosphere beneath the Vardar oceanic lithosphere [Bonev *et al.*, 2010a]. Since the basal contact is a "scar" with a large metamorphic contrast between high-grade footwall and low-grade hanging wall, the structural argument cannot exclude that the possible island arc has been obducted southward before extrusion of the footwall (hence northward, "scar" shearing in the contact zone). Moreover, and still referring to models [Chemenda *et al.*, 2001a], obduction of an island arc over a north-dipping subduction remains possible. The150-160 Ma age of northward-shearing would date arrival of the continental Lower Terrane in the trench where some back-thrusting could be active. Subduction of that continent would trigger the collisional forces needed to subduct parts of the arc and forearc, and close the backarc [Boutelier *et al.*, 2003; Boutelier and Chemenda, 2008]. This scenario would also be consistent with north-dipping subduction generally inferred in the literature. Consistently north-dipping subduction is also in better agreement with deep tomography, which does not show remnants of a south-dipping slab in the asthenosphere of the region [Bijwaard *et al.*, 1998; Piromallo and Morelli, 2003]. In that case, the Rhodope arc became an active continental margin at the southwestern border of Moesia, the neighbouring part of the European continent [e.g. Ricou *et al.*, 1998; Van Hinsbergen *et al.*, 2005], and this discussion comes back to the point where, concerning the Alpine history, the Rhodope was the active continental margin of Europe, at least during Mid-Jurassic to Late Cretaceous times. Northward subduction is further consistent with marginal basins that were inverted before Albian times in the European upper plate [see discussion and review in Ricou *et al.*, 1998]. The later, Turonian-Santonian rifting in the Sredna Gora back-arc basin indicates the re-establishment of extensional forces in the upper plate, which may indicate slab roll back at that time. In such a reconstruction, the lower terrane was either one of the continental fragments derived from Triassic rifting and migrating away from Africa towards Europe (the Upper terrane) or a continental fragment separated from the southern Eurasian margin during the Permo-Triassic breakup of Gondwana. Like for the Briançonnais in the Alps, closure of Tethys would have rewelded continental fragments of the Early Mesozoic passive margin of Eurasia [e.g. Pleuger *et al.*, 2007].

We are left with some uncertainty concerning the pre-Mid-Jurassic history. For this review, and as far as the Rhodope is concerned, the simplest solution is to consider northward subduction below an active continental margin. The author uses as argument the scarcity of N-MORB ophiolites and the predominance of lherzolite as protolith of most metaperidotites to prefer a continental setting rather than an intra-oceanic setting of the Jurassic arc (Fig. 11). Arc-related magmas of Jurassic age in the

Rhodope are evidence for pre-Cretaceous subduction in this part of the Tethys collisional system. This evidence offers a possible correlation with the western contact of the Serbo-Macedonian of the Former Yugoslav Republic of Macedonia, to the northwest [Šarić *et al.*, 2009] and with the Jurassic Pontides, Crimea and Lesser Caucasus arcs known further east [Kazmin *et al.*, 1986; Sengör *et al.*, 1993].

Figure 11. Tectonic interpretation across the Rhodope island arc during the Late Jurassic.



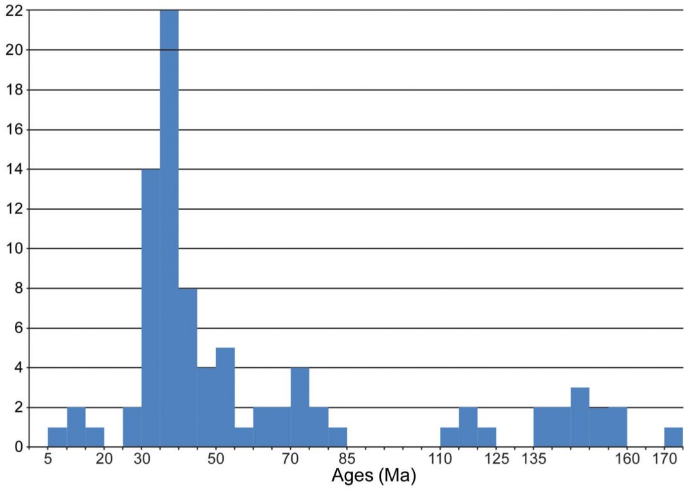
Subsequent subduction of the active margin requires an incoming continent that will be pulled down into the trench. The increasing buoyancy forces of the subducted continental margin can trigger the required torque to produce failure of the overriding active margin followed by subduction of the arc plate [Chemenda *et al.*, 2001b].

Age clusters

Four age clusters (ca 150, 75, 35, <20 Ma) with a possible additional one at about 120 Ma (Fig. 12) are an outcome of the geochronological effort put into the Rhodope over the last decade [see also Liati *et al.*, 2011]. These are, in essence, thermal events recorded by various isotopic systems. The question is whether each corresponds to distinct subduction-collision events thus denoting several subductions of separate terranes [e.g. Liati, 2005]. Such an interpretation neglects the risk that several, often not well understood processes disturb the isotopic systems [Villa, 1998]. This is even true for U-Pb in eclogite zircons [Rubatto and Hermann, 2003]. In this review, the author prefers to understand age clusters in geological units that are linked, as map continuity demonstrates. The author admits hesitation in correlating age cluster and terrane when it comes, in particular, to the Tertiary events of the Rhodope Metamorphic Complex. Several ages taken in the literature as evidence for high-grade, even high-pressure metamorphism are younger than intrusive granites that did not record such metamorphic conditions. Tertiary high-pressure rocks imply extremely fast exhumation rates to be brought to the surface at the time the unconformable sediments were deposited. If exhumation rates are several centimetres per year, one must find the corresponding voluminous amount of erosional products in the proximal sedimentary basins coeval with exhumation. The balance cannot be made in the Rhodope-Aegean region.

The oldest- age cluster is obtained from the ultra-high to high-pressure - high temperature conditions of rocks with older protolith age and of both continental and arc-origin in the imbricate units. Ages span from > 170 to ca 120 Ma (Table 2); 50 Ma is too long a time for a tectono-metamorphic event but we note an overlap with magma emplacement (protolith ages) between 160 and 130 Ma in the same imbricate units (Table 1). These two remarks open the possibility that some and oldest ultra-high and high pressure rocks, including melt-bearing sediments and hydrated peridotites, have been carried into the Rhodope Massif by arc magma ascending from an active slab and the mantle wedge. Such a long-lasting, trans-lithospheric process is simulated numerically [Gerya and Meilick, 2010]. Subsequent arc collision and subduction below Europe (Fig. 11) triggered a high-stress collisional regime, which might be responsible for the Late Jurassic-Early Cretaceous (155-130 Ma), intra-continental shortening in the Strandja orogenic belt [Sunal *et al.*, 2011]. The 130-115 Ma sub-cluster may include some high-pressure metamorphic conditions and is therefore considered as the time during which the arc system reached its deepest subduction (Fig. 13). That was immediately followed by the fast upward return of the Rhodope arc to regional, amphibolite-facies metamorphic climax coeval with early, backward crustal stretching at about 110Ma (Fig. 13). The collisional orogen entered then in its waning stage; the 85-60 Ma cluster (Fig. 12, Tables 3 and 7) suits decompression melting, minor plutonism and cooling below metamorphic temperatures of the amphibolite-facies within the orogen, coeval with back-arc magmatism on the European side [e.g. Von Quadt *et al.*, 2003; Georgiev *et al.*, 2009]. The mountain belt was eroded by Late Cretaceous times and earliest sediments were deposited at about 60 Ma (Fig. 14). With this history, we have a simple subduction/collision system, placing extrusion - exhumation of the deeply subducted arc in the Late-Cretaceous and a 110-60 Ma longest-possible erosion period, an acceptable duration in the light of existing mountain belts such as the Alps [e.g. Schlunegger, 1999] and Himalayas [e.g. Najman and Garzanti, 2000]. The orogenic period apparently vanishes until ca. 40 Ma with subordinate magmatism and fault activity in the Rhodopean crust. Although the tectonic picture should be three-dimensional, one may conjecture that subduction and closure of the Vardar Ocean, to the “southwest” of the Rhodope Massif in two-dimensional reconstruction, absorbed plate convergence while mantle delamination and subduction continued. Complexity is met with the younger age clusters. Either geochronological interpretations are correct, and there was a second subduction during the Eocene, or isotopic systems were more or less reset by the large thermal overprint expressed in widespread magmatism. In the first case, the distribution of Eocene-Oligocene ages should display a regional subduction polarity that is not found. Additionally, Late-Cretaceous-Paleocene sediments were not buried and there is no trace of any suture between the basins they represent. The Rhodope massif was deeply eroded and covered by a Late-Eocene shallow-marine cover, which is further evidence for return to isostatic equilibrium of the orogenic area at that time. A potential Eocene subduction and/or resumed subduction zone is at odd with this information.

Figure 12. Histogram of "metamorphic" ages.



Histogram of "metamorphic" ages reported in the literature for the Rhodope Metamorphic Complex (Tables 2, 3, 4 and 6). Clusters have more significance than number of ages, which largely depends on methods.

Figure 13. Tectonic interpretation across the Rhodope island arc in the Early Cretaceous.

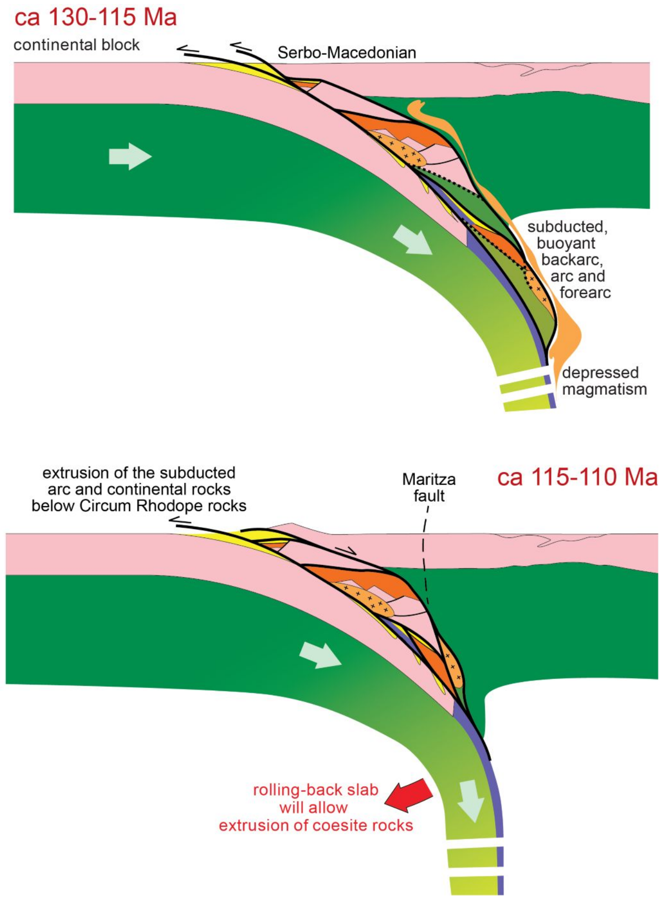
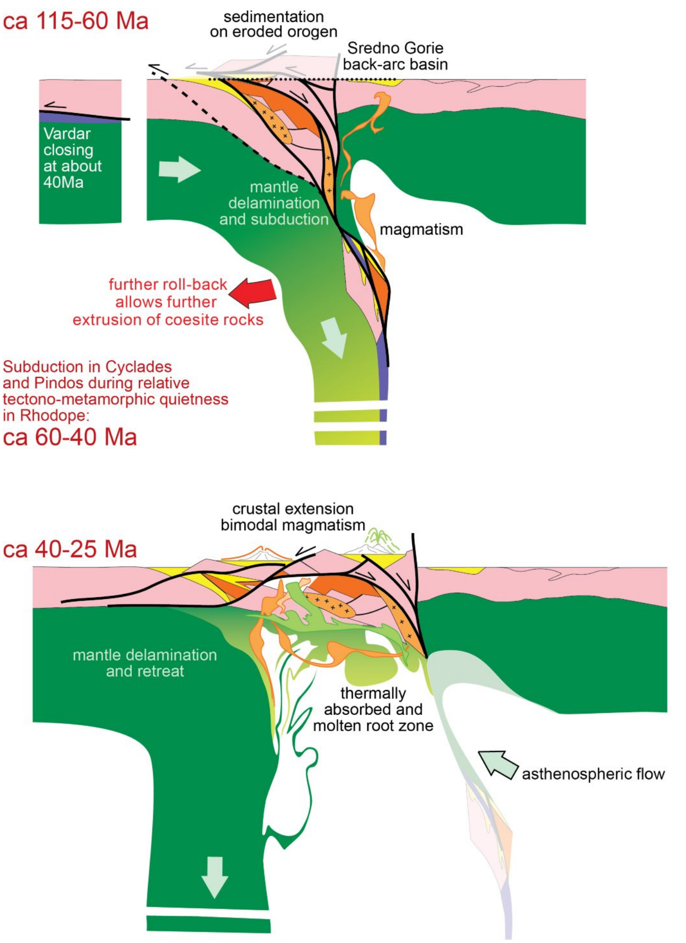


Figure 14. Tectonic interpretation across the Rhodope island arc in Late Cretaceous to Oligocene times.



The Late Eocene marks a turning point in the magmatic and tectonic history of the Rhodope. Low-angle detachments demonstrate localized ductile deformation while basins were forming at the surface. The new magmatic cycle that produced acid to intermediate volcanites between 37 and 25 Ma implies melting of the continental crust material. At the same time uplift is documented by the marine beds followed by continental deposits in the related grabens, which indicate crustal extension [Goranov and Atanasov, 1992]. Yet, this event should not have produced much topography as the basins were virtually not eroded since they were formed, which brings suspicion to orogen-forming, collision event. Uplift and crustal extension is a paradoxical association if an isostatically equilibrated crust is accepted for the Late Eocene. Looking for a lithospheric-scale interpretation, as discussed previously, and combining crustal-melting, supra-subduction magmatism, rock uplift, extension and slab retreat since the Eocene-Oligocene, mantle delamination is a plausible solution that has been put forward in other orogens [e.g. the Carpathians, Chalot-Prat and Girbacea, 2000; the Sevier-Laramide, Wells and Hoisch, 2008; the neighbouring Anatolian segment of the Alps-Himalaya collision system, Göğüş and Pysklywec, 2008]. Whilst mantle lithosphere progressively peels away from the Rhodopean crust as a coherent sheet, the delamination point migrates away from its starting point, across the thinned lithospheric region, hence allowing ductile extension of the crust (Fig. 14). Hot asthenosphere flows into the gap between mantle and crust as it opens. Subsequent heating would cause partial melting and magma generation. Magma intrudes the upper crust as post-orogenic plutons and supplies regional, bimodal volcanism while the hot asthenosphere replacing the heaviest lithosphere triggers isostatic uplift and initiates the current topography. Renewed extension in the Miocene may signal the time when mantle peeling reached the oceanic lithosphere behind the collided continental block and started the active Aegean subduction system. This "minimalistic" interpretation in terms of number of subduction-exhumation cycles brings us to the question concerning the present-day crustal structure of the Rhodope.

Crustal structure of the Rhodope

An unresolved structural problem addresses the reconciliation between the large crustal thickness of the central Rhodope, which is over 50 km [Velchev *et al.*, 1971; Dačev and Petkov, 1978; Geiss, 1987] and the large amount of crustal extension expressed in the multiplicity of detachment faults reported in the literature. The 30-50 km deep crust cannot be treated as an old root if we accept that the shallow-marine Eocene cover is evidence for isostatic equilibrium at that time and that mantle delamination generated widespread magmatism in Oligocene times. Abundant magmatic underplating and volcanic “overplating” during the Oligocene is a possible [Isacks, 1988; Kono *et al.*, 1989] yet conjectural process of crustal thickening. Mantle delamination as sketched in figure 14 would result in the thermally weakened and eroded lithosphere as imaged on tomographic profiles [Van Hinsbergen *et al.*, 2005]. Yet, since regional extension reigned in the Rhodope-Aegean region since then, it is unlikely that the present-day 50km thick crust results from compressional shortening of the weak lithosphere, as suggested in the Andes [Beck *et al.*, 1996]. In the quest for an explanation, the dimensions of the Rhodope are too small to invoke lower crustal flow to add material to the crust [Bird, 1991]. One faces the same uncertainty in the interpretation of the thick crust below the Andes where shortening is estimated to be relatively small [e.g. Kley and Monaldi, 1998]. Therefore, the formation of the Rhodope Mountains, with their tectonic specificity, remains an important question to students of orogenic processes.

Conclusion

Both stratigraphic ages of the sedimentary cover and radiometric dating of crystalline rocks substantiate synmetamorphic thrusting and exhumation in Early Cretaceous times, and additional extensional exhumation in two steps starting with the Late Eocene transgression.

The imbricate units represent an Alpine suture zone in which remnants of a partially subducted magmatic arc, foundered on the attenuated margin of Europe, are preserved. Subduction and collision of the arc with the incoming Lower-Terrane continent produced decoupling within the arc and subduction of its deeper parts along with the frontal parts of the Lower Terrane, one of the intra-Tethys continental blocks derived from Gondwana. This is when the complex north-dipping stack of synmetamorphic nappes developed in the Rhodope. Imbrication involved crustal extrusion that generated east-north-east-verging, flat-lying and synmetamorphic shear zones coeval with, and as a consequence of, terminal continental collision. Gravitational collapse contributed to lateral spreading of the upward rising high-grade terranes.

Another extensional episode started in the early Eocene in the Rhodope, 10–15-Ma earlier than in the Cyclades and might be due to mantle delamination. Aegean extension, which started in the Miocene, is still shaping the geology of the Rhodope Metamorphic Complex.

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